

# Earthquakes — Converging at Cascadia

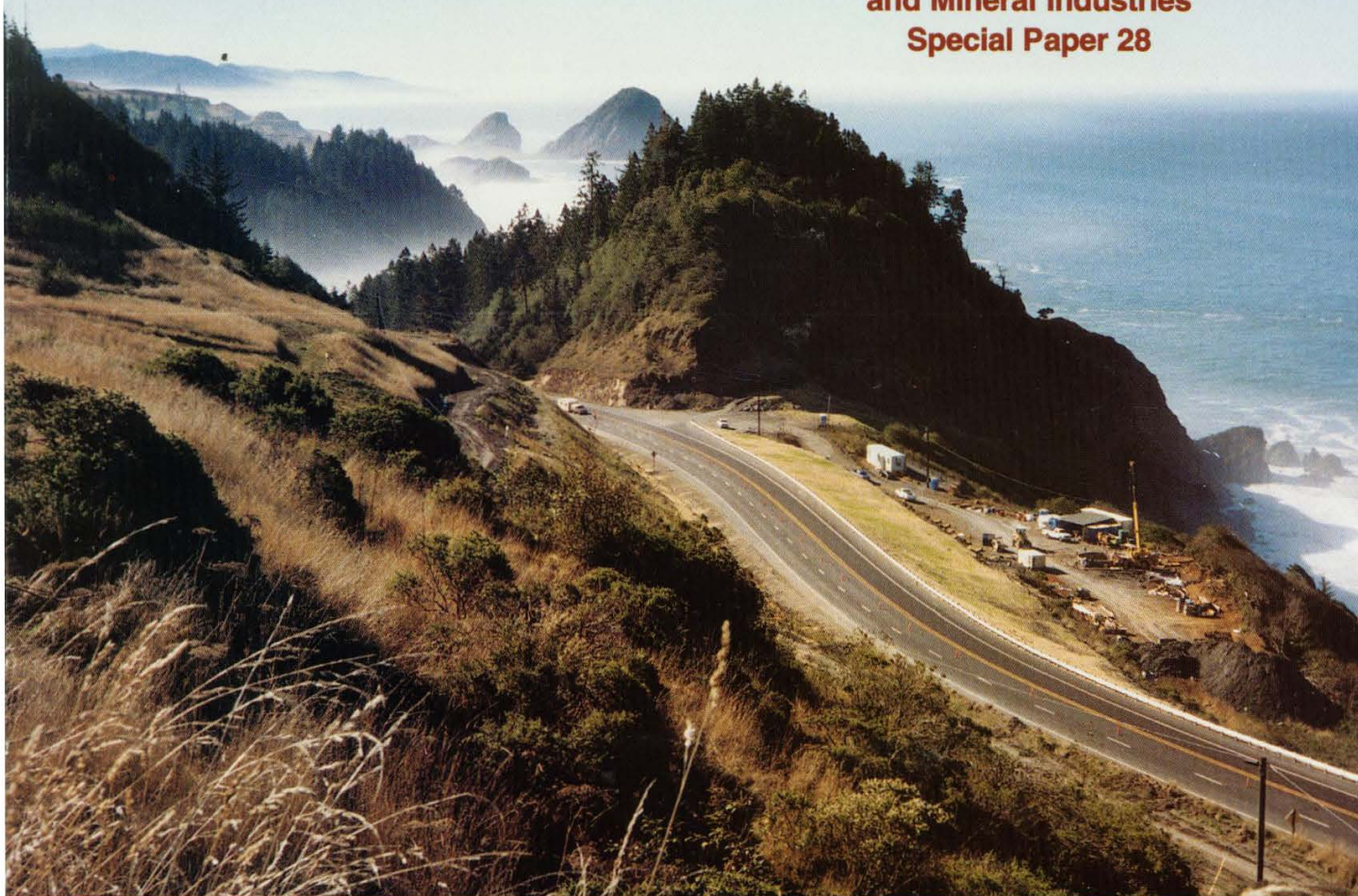
## Symposium Proceedings

Association of Engineering Geologists

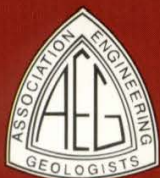
40th Annual Meeting September 30 - October 4, 1997, Portland, Oregon

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and Mineral Industries  
Special Paper 28**



Hosted by the Oregon Department of Geology and Mineral Industries  
and the Oregon Section of the Association of Engineering Geologists



***Converging at  
Cascadia***



STATE OF OREGON  
DEPARTMENT OF GEOLOGY AND MINERAL INDUSTRIES

**SPECIAL PAPER 28**

ASSOCIATION  
OF ENGINEERING GEOLOGISTS

**SPECIAL PUBLICATION 10**

**1997**

# **Earthquakes — Converging at Cascadia**

**Symposium Proceedings**

**Association of Engineering Geologists**

**40th Annual Meeting, September 30 – October 4, 1997**

**Portland, Oregon**

**Edited by**

**Yumei Wang and Klaus K.E. Neuendorf**

**Oregon Department of Geology and Mineral Industries**

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## PREFACE

The Association of Engineering Geologists held its 40th annual meeting in Portland, Oregon, from September 30 through October 4, 1997. A special event of the meeting was the symposium "Earthquakes—Converging at Cascadia." The present volume contains the invited papers given at this symposium.

During the planning of the symposium program, the participating geologists and engineers had agreed that they needed a better understanding of current research findings in the Pacific Northwest. Consequently, the program reflected a process from scientific findings to their application. The focus of the symposium was to broaden the understanding of earthquake research issues centering on the Cascadia region earthquake setting.

Since the late 1980s, the Pacific Northwest has been an exciting region for earthquake professionals. There has been a rapid growth of scientific findings on a wide range of topics. In the Pacific Northwest, earthquakes are relatively uncommon, and there is only a small chance for a devastating event. Most studies and mitigation efforts have been conducted in relatively recent times, and findings indicate that our communities are severely underprepared. Consequently, the professional earthquake community is striving for more information. This includes identifying and better identifying potential earthquake sources, understanding the characteristics of ground shaking and propagation, and determining hazards (e.g., by studying paleoseismic records for clues and estimating future damage). Over the last decade, much fundamental knowledge has been obtained, but much more still needs to be discovered.

The proceedings volume includes contributions by researchers, listed in speaking order, presenting the most recent results of their work. The sequence begins with an overview of seismic hazards, then moves into basic research, and concludes with topics in applied research.

At the symposium, in the order of appearance, **Bruce A. Bolt**, University of California at Berkeley, delivered the keynote address, sharing his perspective in "An Assessment of Seismic Hazard Assessments." **Ray E. Wells**, U.S. Geological Survey, reviewed the geologic framework and ongoing research in the Pacific Northwest, focusing on kinematic models and geologic mapping. **Ivan G. Wong**, Woodward-Clyde Federal Services, reviewed the historic earthquake record in the Pacific Northwest and how it is used to assess seismic hazards. **Silvio K. Pezzopane**, U.S. Geological Survey, provided a synopsis of late Quaternary faulting and characterization of earthquake sources in Oregon. **David K. Keefer**, U.S. Geological Survey, and **Yumei Wang**, Oregon Department of Geology and Mineral Industries, presented preliminary methods for predicting earthquake-induced landslides in making hazard maps. **Stephen F. Obermeier**, U.S. Geological Survey, and **Stephen E. Dickenson**, Oregon State University, discussed paleoliquefaction findings and their implication in the Pacific Northwest. Finally, **Kaye M. Shedlock**, U.S. Geological Survey, and **Norman A. Abrahamson**, Pacific Gas and Electric Co., provided an overview of the current understanding of ground motion attenuation relationships including relationships appropriate for the Cascadia subduction zone.

This symposium was an opportunity for scientists and practitioners to converge, share ideas, and learn about each other's work in progress. Research is vital and allows us to gain an improved understanding of the earthquake environment and assorted risks.

Equally important is utilizing the research findings. Acting on the basis of what has been learned is as valuable as the basic research, because without risk reduction there is no true progress in dealing with earthquakes.

Today, we have improved tools, techniques, and coordination that allow more effective design, construction, mitigation, and planning. For example, we can evaluate the performance of sites and estimate ground shaking at various frequencies to design better structures; we can estimate damage and loss from future earthquakes for a wide region; we can reduce the shaking of structures by incorporating viscous dampers and base isolation systems for new or existing structures; and we have statewide earthquake drills to maximize our readiness during actual disasters.

Although it is hard to make present sacrifices to avoid future earthquake losses, it should not take an earthquake to stir us into action. We already have identified a clear and present danger. We need smart decisions on difficult political questions, such as, "What level of risk should we as a community be willing to accept?" We must think of cost-effective ways to reduce earthquake risks, then prioritize the necessary efforts to get us there, and lastly, apply them to our community. We do not know when the next earthquake will occur. If we act now by taking manageable, bite-size pieces, we can make a positive impact.

Risk reduction, after all, does not happen on its own, nor all at once. We need to take a "systems approach" and account for the lifelines serving the community. Not only should we build better, we need to strengthen what we have. We must rehabilitate old, weak buildings and harden lifelines, so that water to fight the fires and electricity to operate businesses will be available when needed.

Engineering geologists are eminently qualified to serve their communities—and the communities' children and grandchildren—in reducing earthquake risks. It is hoped that this symposium has brought their qualifications yet another step forward.

My thanks go to the organizers of the annual meeting, especially to chair Gary Peterson, and the Association of Engineering Geologists for providing the opportunity to convene this symposium. I gratefully acknowledge the contributors of this volume, both for their presentations at the symposium as well as for their papers. Lastly, my sincerest appreciation to my colleagues at DOGAMI, especially John Beaulieu and Klaus Neuendorf, for their support.

Yumei (Mei Mei) Wang, Symposium Chair  
Oregon Department of Geology and Mineral Industries

# An Assessment of Seismic Hazard Assessments

by

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## Abstract

Seismic hazard in a technical sense arises from the earthquake-related dangers themselves and not the vulnerability and risk associated with structures, people and the economy. The parameters needed to define adequately and predict the occurrence of the various seismic hazards are diverse; the shaking hazard requires seismic source characterization (moment, mechanism tensor, interoccurrence time, etc.), wave-path structural geometry, soil characterization, marks for maximum ground acceleration, velocity, displacement, duration, spectral content, attenuation, and intensity. The estimates available are often rather uncertain and assumption laden.

Seismic hazard maps are the most used products for engineering, insurance, and planning purposes. Recent examples, in terms of peak accelerations with certain probabilities in a given time, illustrate the uneven quality of geological seismic source maps and available strong-motion recordings. Additional needs for improvement are immediately apparent: more robust characterization of geological structures, denser geophysical sampling of the ground motions, and use of memory-dependent hazard functions. Vulnerability and risk demands now emphasize the seismic ground velocity and displacement as key hazard factors. Regional and national maps of these and alternate wave hazard parameters still need to be estimated. Seismic zonation is an important aspect of hazard assessment. The requirements are for more digital accelerometers sited on the surface and downhole as planned strategies to quantify specific vulnerable urban regions and common geological conditions, such as alluvial basins. More reliable seismic hazard studies, whether deterministic or probabilistic, require separate ground shaking attenuation functions and seismic source characteristics for compressional and extensional tectonics. An overview of seismic attenuation studies illustrates the substantial amount of interpolation, smoothing, extrapolation, and aggregation still being used.

On the geological side, there remain confusing variations in the definition of fault "activity" or "capability". Fault trenching, tsunami run-up, and other recent tools of paleoseismology are developing strongly, but the seismogenic implications of low slip-rates and blind faults present unsolved geological challenges.



## 1. Definitions

The societal interest is in future earthquake dangers not historical ones (Bolt, 1991). But hazard estimation requires extrapolation from the past so that we are required to predict future catastrophic changes using a form of the geological principle of uninformatarianism. It is surely curious that earthquake prediction has been commonly limited to mean the forecasting of the time, place and size of an earthquake. Such endeavors, always with low prior probability of success, have proved futile, while the more useful program of forecasting ground shaking, liquefaction, etc. (i.e., the seismic hazard) has had considerable success.

A recent International Association of Seismology publication (McGuire, 1993) summarizes seismic hazard assessment using various methodologies with very different geological bases in 88 countries of the world. The principal seismic hazards are (1) ground motions, including shaking, differential settlement, liquefaction, slides, ground lurching, and avalanches; (2) ground displacement along the ruptured fault; (3) floods from dam failures; and (4) fires. In this paper, I consider only the first hazard category.

## 2. Seismicity

An explanation for the uneven geographical pattern of significant earthquake around the world is given by the theory of plate tectonics. First, most earthquakes occur along the edges of the interacting tectonic plates (*interplate earthquakes*), but a few, including some of large magnitude (such as the 1811 and 1812 New Madrid earthquakes in North America) occur within a plate (*intraplate earthquakes*). In some seismically active areas, such as along the Pacific margins of Oregon, Washington and Alaska (*Benioff zones*), plate convergence results in crustal rocks plunging down (*subducting*) into the Earth. These convergent plate boundaries contribute more than 90% of the Earth's release of seismic energy for shallow earthquakes, as well as most of the energy for intermediate and deep-focus earthquakes (down to 680 km depth). Most of Earth's largest earthquakes, such as the 1960 and 1985 Chile earthquakes, the 1964 Alaskan earthquake, and the 1985 Mexico earthquake, originate in subduction zones (Bolt, 1993). A high rate of seismicity also occurs along the mid-oceanic ridges where the tectonic plates are created by volcanic processes along undersea faults. These diverging plate margins involve both dip-slip faulting (*normal faults*) and horizontal slip (*transform faults*). A much studied example of the latter is the San Andreas fault system, which connects the ocean ridges in the Gulf of California with the Gorda ridge under the Pacific Ocean of Oregon. Plate margins where continents collide, such as the Himalayas and Caucasus, also generate energetic earthquakes.

This broad global classification of seismogenesis caused by convergence and divergence does not accommodate the many earthquakes, including major damaging ones, that occur far from the plate boundaries, under the abyssal plains of the deep oceans and within the continental crust; significant extension must be made to plate tectonic kinematics to provide a ready explanation for these intraplate earthquakes. Hinterland seismic activity has been found in all continents except Greenland (see Table 1). The available earthquake catalogues (dating back to the 16th century) cite at least 15 major earthquakes in portions of the Earth's crust that on the

simple plate tectonic theory would be regarded as stable. A striking example is the catastrophic earthquake that struck deep inside continental China in 1556 on January 23 in Shensi Province near the old capital city of Xian. This earthquake produced the greatest loss of life ever recorded from seismic activity. The official Chinese catalogue estimates that 830,000 people died from all causes. Strong earthquakes in the European intraplate region have also been described over the historical centuries. Another intraplate earthquake of exceptional size occurred in the Rann of Kutch, in northeast India in 1819. In the Kutch earthquake, a 3 m high scarp appeared striking east to west for about 250 km.

Table 1		
Sample large intraplate earthquakes		
location	year	M <sub>s</sub>
Basel, Switzerland	1356	(7.4)
Shensi, China	1556	(8.0)
Quebec (St. Lawrence river), Canada	1663	(7.5)
Shantung (Tanlu fault), China	1668	(8.5)
New Madrid, U.S.A (two similar earthquakes 1811, 1812)	1812	(8.3)
Kutch, India	1819	(7.8)
Charleston, U.S.A.	1886	(7.6)
Libya, North Africa	1935	(7.1)
Meeberrie, Australia	1941	(7.0)
San Juan, Argentina	1944	(6.5)
Meckering, Australia	1968	(6.8)
Nahanni, Canada	1985	(6.8)
Tennant Creek, Australia	1988	(6.7)
Latur, India	1993	(6.4)

In North America east of the Rocky Mountains, hazardous earthquakes are infrequent, with a few notable exceptions. The most active zone runs from southern Missouri southward along the Mississippi River. In the fall and winter of 1811-1812, three principal earthquakes and numerous aftershocks occurred near the town of New Madrid, Missouri. These earthquakes were of high intensity and were felt over distances of 1000 km; there were intensity reports from as far away as Washington, D.C. These intraplate earthquakes are probably the most energetic recorded in the contiguous United States, yet the closest plate boundary is more than 4000 km away. Another significant seismically-active zone lies along the St. Lawrence River in Quebec province, where large earthquakes were reported in the seventeenth century (see Table 1). The largest historical earthquake along the eastern continental margin occurred in North Carolina in 1886, damaging Charleston. Despite extensive subsequent geological field studies, (Dewey et al., 1989) no specific surface faulting associated with any of these large earthquakes has been discovered, although the rift fault zone of the New Madrid earthquakes has now been well mapped from hypocenters of the frequent small-magnitude seismicity. Indeed, globally, in stable continental regions only 11 earthquakes are known historically to be associated with causative surface fault rupture.

Quantification of the strength of an earthquake has become quite sophisticated for hazard assessment purposes. The oldest measure is seismic *intensity*. Intensity is the measure of damage to the works of man, to the ground surface, and of human reaction to the shaking. Intensity maps prepared from calibrated scales, such as the Modified Mercalli Scale developed in California, provide crude, but valuable information on the distribution of strong ground shaking, the effect of surficial soil and underlying geological state, the extent of the source, and other matters pertinent to insurance and engineering problems (Bolt, 1993). Some work is now being done to make intensity scales more specific to structural type. Because intensity scales are subjective and depend on social and construction conditions of the country, the size of the earthquake is better tabulated in terms of earthquake magnitude, an instrumental scale.

The most widely used magnitude is the Richter magnitude ( $M_L$ ), defined for local earthquakes in California as the logarithm to the base 10 of the maximum seismic wave amplitude in microns ( $10^{-4}$  cm) recorded on the Wood-Anderson seismograph located at a distance of 100 km from the earthquake's epicenter. The definition of magnitude entails that there are no theoretical upper or lower limits. The limitations are set by the physical limitations on the amount of strain energy that could be stored in the strained rock around a rupture of a given area in the Earth's crust. It became common in the last decades to use another magnitude measure called the surface wave magnitude ( $M_S$ ); many earthquakes in the strong earthquake data base are given in terms of  $M_S$ . This magnitude scale is based on measuring the amplitude of seismic surface waves with a period of 20 seconds and is more suitable than the Richter scale for remote measurements of very large earthquakes.

A more robust procedure has superseded the above quantities in hazard studies. Seismologists favor a size measure (Bullen and Bolt, 1985) based on seismic moment ( $M_0$ ). It yields a consistent scale of earthquake size from small earthquakes to the largest known and thus

provides earthquake hazard assessments. The physical basis is the equivalency of the force couples (moment tensor) of the stresses that produce the fault rupture. Such couples have mechanical moments which, unlike magnitude, have a firm basis in dynamics. Seismic moments range over many orders of magnitude from the smallest to the largest earthquake. Thus, between magnitude 2 and magnitude 8, the seismic moment typically spans over 6 orders of magnitude. The moment of the 1906 San Francisco earthquake is estimated to be over 10 times that of the 1989 Loma Prieta earthquake.

Moment values are correlated with the magnitude scale to produce the moment magnitude, written  $M_w$ . For comparison, the 1989 Loma Prieta earthquake has an estimated magnitude  $7.1 M_s$  and  $6.9 M_w$ . Approximately (Kanamori and Anderson, 1975)

$$M_w = 0.67 \log M_0 - 10.7 \quad (1)$$

In addition to historical reports, knowledge of earthquake occurrence depends mainly on the instrumental record of seismicity and geodetic and geological studies. Resulting catalogues and maps of locations and magnitudes of earthquakes down to quite small sizes provide the basis for broader statistical treatment of earthquake frequency and help define earthquake source regions. The occurrence is modelled in terms of the total number  $N$  of earthquakes of magnitude  $M$  during  $T$  years in an area  $S$  square kilometers. For a uniform distribution of sources (i.e., the assumption of geological fault ignorance),

$$\log N = a - b M + \log S + \log T \quad (2)$$

where  $a$  and  $b$  are observational parameters.

In seismic areas, there is usually need to truncate the form (1) at lower magnitude  $m_0$  and a maximum credible magnitude  $m_1$ :  $m_0 < M < m_1$ . These threshold values are often rather arbitrary but sometimes dominate.

In order to make predictions of the probability  $P(r|H)$  of future seismic hazard  $r$  with any confidence, the regressed data  $H$  on the limited historical seismicity should be augmented by information on the seismogenic framework provided by structural geology. The zones of high stress and current dislocation of the Earth's crust must be mapped and rates of deformation estimated by stratigraphic methods and geodetic surveys.

The long interoccurrence time of major earthquakes produces two basic difficulties in hazard estimation. First, there is a general ignorance of the rate of occurrence, and secondly the location of the most stressed fault segment is unclear. In the case of intraplate earthquakes, the rarity of surface faulting makes special field study especially valuable. Paleoseismologic techniques have been recently developed to determine the average interoccurrence rate of earthquakes produced by slip on such faults.



Recently, there has been a major effort in the United States to map and substantiate Quaternary active faults across the continent (Dewey et al., 1989). In dramatic contrast to the seismically-active western states, the only fault in the eastern region with recurrence intervals of seismic surface rupture is the Meers fault in southwest Oklahoma (Crone and Luza, 1990). The fault scarp occurs in Quaternary deposits. Stratigraphic work indicates movement occurred on this fault during the past 3,000 years and perhaps as recently as 1,200 years ago. Despite such recent slip, seismographs detect no earthquakes along the fault at the present time.

A key question in specifications of strong ground motion is what is the relevant maximum earthquake; in other words, what is the realistic maximum limit to the fault length that will rupture? In attempting to estimate the probability of rupture lengths, geologists rely on the identification of fault segments that are separated by such mechanical barriers as changes in fault strike or lateral fault offsets. The importance of this segmentation method requires more checks of its validity as major earthquakes occur.

The second basic problem of seismology in areas of intraplate seismicity is constructing robust empirical models of seismic wave attenuation. Generally, the attenuation of seismic waves is lower in older stable continental shield areas than in younger crustal rocks. For example, Modified Mercalli intensity maps in the eastern United States indicate that within 100 km of a large earthquake the attenuation is similar to that in the west, but thereafter the distance required for the peak ground velocity to be reduced by one half, almost doubles. A set of papers giving comparisons between the wave attenuation laws in the eastern and western continental regions has recently been published (see Abrahamson and Shedlock and following papers, 1997). Few large earthquakes necessarily mean few relevant ground motion records. Thus while the strong-motion attenuation laws used for peak motions and for spectral attenuations come mainly from strong motion recordings in California, in the eastern states they are derived from historical intensity recordings and the few small earthquakes that have been recorded (Atkinson and Boore, 1997). Note also that variation in crustal properties changes the seismogram wave patterns in different frequency bands and distance ranges. For example, in the continental shield certain higher-mode surface waves, such as the seismic Lg phase, propagate efficiently; in California they do not.

### **3. Hazard Estimation Methodologies**

All statements of hazard contain, either explicitly or implicitly, elements of probability (see Figure 1). Applications of the chance of earthquakes for building codes, for engineering design, and for risk reduction policy decisions are becoming common.

Numerical statements of specific hazard or risk odds are sometimes difficult to interpret unless they are compared with probabilities for other hazards. Thus, the risk of death per year to an individual from a motor vehicle accident is about 1 in 4,000; from earthquakes in the most exposed metropolitan areas, the risk is perhaps 1 in 50,000. But much more is involved than these simple propositions. The individual risk clearly varies with individual situations. There is also a collective or societal risk. Recent refinements are largely improved geological

databases, but limited attention has been given to the explanation of the probability statements.

A recent development has been the assessment of the odds of large earthquakes along segments of the San Andreas and other major active faults in California (see Yeats et al, 1997 for a discussion). Benefits from such assessments require careful application and consideration of the societal context. After the Loma Prieta earthquake, probability evaluations of this type were given much publicity. The combination of different bodies of observations indicates that there is a better than 1 in 2 chance of a major magnitude earthquake occurring in the San Francisco Bay Area in the next 20 years. An earlier, more specific study gave the chance of a 6.5 to 7 magnitude earthquake along a 30-km-long segment of the San Andreas fault in the southern Santa Cruz Mountains as 30% in 30 years. This value was higher than for the adjacent San Francisco Peninsula segment of the San Andreas fault to the north and led to the mistaken impression that the 1989 earthquake was predicted.

If probability assessments are to be adopted widely as a basis for hazard evaluation, considerable thought is needed in formulating and describing the statistical conclusions. Among the explanations required are: What is the range of earthquake size involved rather than the specification of a particular magnitude? What are the overall uncertainties in the calculations? And are such statements predictions at all or only summary accounts of past events? An attempt to provide such explanations has been made in recent reassessments of occurrence probabilities of damaging earthquakes in the San Francisco Bay Area, but the full uncertainties are still probably underrated.

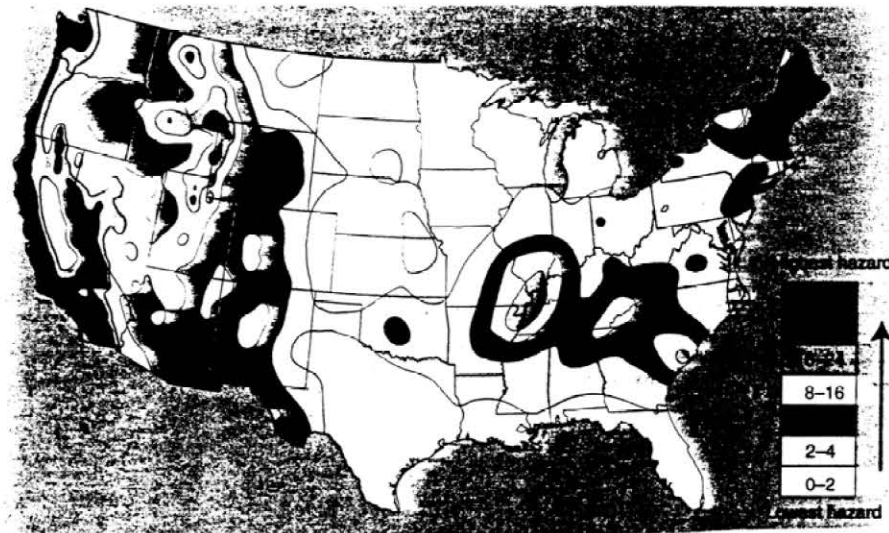


Figure 1. Seismic hazard map for the contiguous United States (Frankel et al., 1996)

Probability models have also been used to prepare ground shaking hazard maps (see Frankel et al., 1996) for the whole United States and for specific regions (see Fig. 1). These maps give the expectation in a given time (such as 100 years) of exceedance of seismic intensity parameters (such as acceleration). In computing the expectation of these parameters, the older concept of discrete hazard zones, drawn mainly on the basis of the historical seismicity, was abandoned and replaced by the rate of occurrence of earthquakes of various magnitudes weighted by geological evidence of active fault systems.

#### **4. Seismic Hazard Mapping within Tectonic Plates**

The most common methodology for the construction of seismic hazard maps now uses a composite pseudo-Bayesian analysis in which the evidence  $H$  in  $P(r|H)$  is culled from geology, fault-slip rates, historical seismicity, plate tectonic models, and various measurements of the shaking parameters. The model variability comes from the (usually subjective) assignation of odds to a finite number of plausible alternatives at each branching of a logic tree. The weaknesses of the method include the non-exhaustiveness of alternatives, the non-commutation (non-independence) of probability factors, and the lack of objective experimental tests of the random process assumed. Some of these difficulties are obviated by extensive sensitivity analysis with many numerical repetitions to determine if the central tendencies are convergent. In any new application, there needs to be concern of the robustness of the process and checks against observed seismicity catalogues are vital.

A more fundamental seismological criticism of many hazard studies is that the computational models do not encompass the complexity of strong earthquake motions (see next section for two examples). Ground motion hazard for representative sites in a continental province are usually specified in terms of seismic intensities, peak (or effective) ground accelerations, and sometimes peak velocities, and/or specific damped spectral acceleration amplitudes at a finite number of periods. All seismic hazard is a function of wave frequency and hazard mapping can be extended to the spacial variation of equal hazard spectra, i.e., of curves of the response spectral values which have an equal (given) likelihood of being exceeded at any (given) response period.

Many hazard zonation procedures for intraplate earthquakes characterize the intensity of ground shaking by the effective values of the maximum wave amplitudes. Yet supremum amplitude spikes, of say peak ground acceleration, are often erratic and not representative of the maximum sustained wave energy. The key role of other measures, such as strong-motion velocity and displacement pulses, has been emphasized more strongly in recent studies of damage caused in the near field by such earthquakes as the 1989 Loma Prieta and 1994 Northridge earthquake in California. Also the importance of wave duration was highlighted in the 1988 Mexico earthquake that produced heavy damage in Mexico City. One aspect of low attenuation of seismic waves in stable continental regions is that the surface wave trains become more prominent as distance from the causative fault increases and the S wave portion and the surface wave codas lengthen. The reason is dispersion of the seismic wave trains with distance,

particularly across large alluvial basins.

The above type of hazard mapping, intended as the foundation of prediction of damage potential or vulnerability, excludes two key variables: duration and near-fault wave behavior. First, the duration of shaking is known to affect the capacity for elastic response of engineered structures. Collapse of 10- to 14-story buildings in the Lake Zone in Mexico City is a clear illustration. The 10 or more cycles of this 1985 earthquake (centered over 350 km away), led to progressive structural weakening. In general, shaking duration needs specification not only for ground acceleration but also for ground velocity and displacement, particularly when considering the input to large structures, such as long bridges, highrise buildings, and base-isolated structures.

Second, a difficulty with broad, smooth mapping of hazard from intraplate earthquakes is the occurrence of a large amplitude wave pulse or "fling" near to the rupturing fault (Bolt, 1996). Such a phase pattern has now been observed in a number of interplate earthquakes, such as the 1971 San Fernando and 1994 Northridge earthquakes (Bolt, 1971); their generation is predicted by seismic wave theory generally for fault sources. These velocity and displacement pulses are not usually represented in general seismic hazard mapping for continents, yet may have serious consequences for building response.

Because many interplate earthquakes up to magnitude 7 have now been recorded by accelerometers, prediction of the future ground shaking by interpolation and extrapolation from actual measured motions has become reasonably reliable. The exceptions are for the very largest earthquakes ( $M \geq 7.5$ ) and for earthquakes with normal fault sources. The lack of the necessary ground motion measurements for intraplate earthquakes has been compensated for in a number of ways. Until shown otherwise, a basic assumption is that the physics of earthquake fault rupture genesis governs for *both interplate and intraplate earthquakes*; allowance for different wave attenuation laws, crustal structures, and source mechanism must, however, be made in computing synthetic seismograms. A variety of methods is now described in the literature (e.g., Cohee et al., 1991; Somerville et al., 1991).

A recent extensive case-history illustrates the procedure: Bureau of Reclamation studies of the seismic hazard at Hoover Dam (Nevada). Hoover Dam is a 221 m concrete gravity arch structure located near two types of seismogenic faults: a normal fault within 3 km of the dam of sufficient dimension to produce a  $M_w = 6.5$  earthquake and a nearby strike-slip fault with similar capability. An immediate problem is that few ground motions from normal-slip mechanism earthquakes have been recorded, even for interplate earthquakes. Therefore, the ground motion input synthetics must be computed with little guidance from observation and no direct verification by means of comparisons with relevant accelerograms. Such computations usually assume plausible values for the area of fault-slip, the distribution of fault rupture velocities, slip rise-times, stress drops, or other dislocation parameters. The resulting synthetics must satisfy general constraints concerning the energy partition between the main seismic wave types, namely, P, S and surface waves, the relative smoothness of the rock structures, appropriate durations, and so on.



For intraplate earthquakes where previous earthquakes are either unknown, unrecorded or small, some assistance can be attained from recordings of very small earthquakes or explosions in the same tectonic region. These sources provide time histories from which wave scattering functions and attenuation laws appropriate for the local rock conditions can be estimated. At one stage of the Hoover Dam studies, such scattering functions were derived and convolved with a simple source function to yield input motions.

## 5. Hazards in the Pacific Northwest

Seismic hazards from earthquakes along subducting margins, are also present in some seismic regions of the United States, namely, in Alaska, Oregon, Washington, and far northern California. Along the Cascadia Subduction Zone, the Pacific and North American plates are converging, with the Juan de Fuca plate being overridden by the North American plate at an average rate of about 4 cm per year. This tectonic model led to special studies that changed attitudes towards seismic hazards in the northwestern United States. A lack of significant historical high seismicity in most of the region had previously led to the inference that the zone was essentially aseismic. The recent direct evidence includes observed deformation of water-saturated ocean-floor sediments of Holocene age along the fold and thrust belt. Many fold structures in the southern part of Cascadia zone in the upper most crust are relatively young and the lateral folding by shortening is consistent with a continued pushing together of the two plates. There is also evidence of cycles of subsidence of the ground surface along the coastline. Borings along estuaries and tidal inlets into water-saturated muds show mud layers interspersed with thinner fossiliferous layers containing tree stumps, driftwood, decayed plant matter, and peat. The organic material in these peaty beds can be dated by radioactive carbon techniques sufficiently precisely to indicate burial of the extensive vegetative lowlands in at least six subsidence episodes of 0.5 to 2 m in the last seven thousand years. The most recent subsidence reported, age-dated from a peaty layer, occurred about 300 years ago. The more recent interoccurrence dates have also correlated with historical evidence of large tsunamis in the Pacific Ocean which ran up on the continental shore.

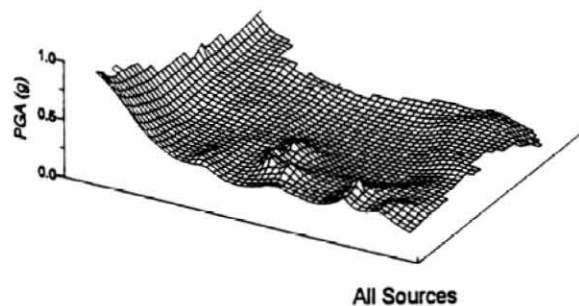


Figure 2. Total seismic hazard from both crustal and subduction sources in Oregon (5 percent damped with 1,000 year return period) (Geomatrix, 1995)

A seminal case history which incorporates evidence on subduction earthquakes was made by Geomatrix Consultants of San Francisco, California, for the Oregon Department of Transportation (Geomatrix Consultants, 1995). This hazard study provided contour maps showing levels of ground shaking with return periods of 500, 1000, 2500 years (Fig. 2). These maps, which are defined for a typical rock site, were constructed for three ground motion marks (i.e., peak ground motion acceleration, and 5% damped spectral acceleration for periods of 0.3 and 1.0 seconds). The report states that the very low levels of earthquake activity that have been measured in historical times in Oregon, from both instrumental and historical records, are consistent with relatively low rates of activity on the seismogenic faults in that state. "As a result, the historic record does not provide an adequate basis for characterizing any of the important elements of source description, including source location and geometry, maximum earthquake magnitudes, and recurrence rates." For this reason, the study concentrated on available evidence from geological field studies and used the methodology of probabilistic hazard analysis, including logic trees. The latter allow subjective weights to be assigned to various hypotheses. These weights reflect the degree of belief in the hypothesis, given the available evidence. This Oregon report illustrates the state-of-the-art of hazard analysis for general application at a regional scale. The resulting maps did not present any surprise. The main hazard surface contours for the state indicate that the highest hazard is along the coast, arising from assumptions concerning very large subduction zone earthquakes. There is attenuation of the hazard inland, interrupted by protrusions of hazard around the historically moderate seismic areas of Portland, Klamath Falls, etc. A criticism related to the application of such maps is that the basis of hazard used is peak acceleration, which taken alone is not adequate for many risk and design demands.

A comparison case is the study for California by geologists and seismologists at the California Division of Mines and Geology and the U.S. Geological Survey (Peterson et al., 1996). The main products were similar to those in the Oregon study and again are not meant to be site-specific. Seismic hazard maps (see Fig. 3) were produced for peak horizontal accelerations for firm rock sites at a hazard level of 10% probability of exceedence in fifty years (500-year return period). The report included summaries of geodetic and historical damage data and comparisons and seismic hazard for specially important populated regions across the state. The hazard analysis incorporated both the historical seismicity and the geologic information within the fault zones that display evidence of displacement during late Pliocene and Holocene times. The uncertain presence in certain areas of the State of blind thrust faults such as that which slipped in the 1994 Northridge earthquake prevents an exhaustive assessment.

Two aspects of the California report are worthy of critical comment. First, estimates of the time variable depend heavily on estimates of the slip rate along known active faults. This technique provides numbers which, while uncertain, give an overall indication of the rate of strain release, although being averages they cannot yield reliable estimates of the history of occurrence of specific magnitude ranges. There is a particular difficulty in judging the implications of low slip rates. Secondly, the work places great emphasis on consensus. "The parameters in this report are not the work of any individual scientist but denote the efforts of many scientists, engineers, and public policy officials that participated in developing the

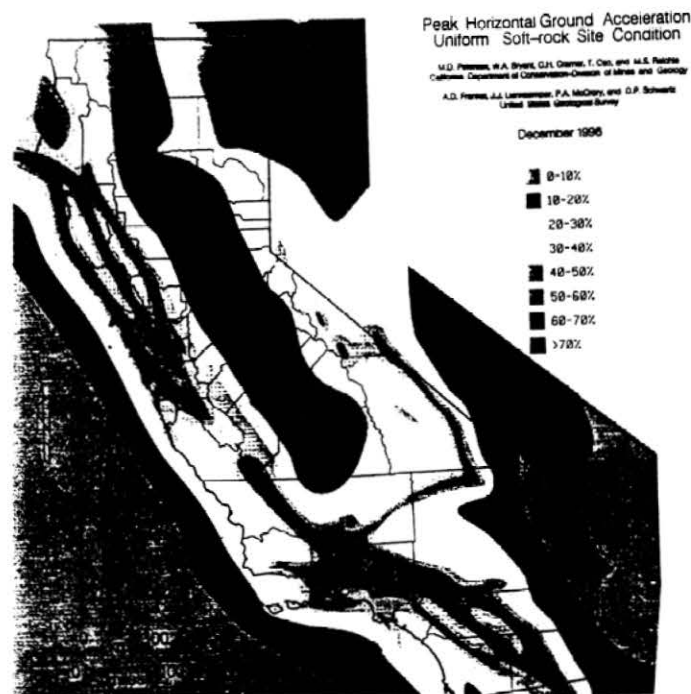


Figure 3. Seismic hazard map for California (Petersen et al., 1996)

statistical distribution used in the analysis." This approach to hazard analysis by aggregation of expertise has limited objectivity. Experience on such assessment teams gives rise to doubts because the assessing of weights of logic trees ("systemic probabilities") and the selection of hypotheses can be dominated by forceful rhetoric. Nevertheless, for many broad purposes such hazard maps are of practical use. An objectionable application is determining quantitative hazard over short distances because the maps do not reflect the variability of seismic ground motions due to local geological, topographic and soil conditions.

## 6. Improvements for Regional Seismic Hazard Assessment

Future efforts must be focused on developing regional zonation maps for earthquake ground shaking and liquefaction potential in urban areas where the seismic hazard is moderate to high. These maps can be used in more accurate loss estimation than is currently possible for most urban areas. To validate and calibrate some of the assumptions and models used in the development of the present generation of regional hazard maps, strong-motion data recorded in high risk areas are a vital need. Reliable seismic hazard values need an understanding of the seismic source, path, and site effects for which strong motion records are essential. For example, most of the moderate-to-large metropolitan areas (e.g., Los Angeles, San Francisco Bay Area, Seattle, Salt Lake City, Boston, etc.) with relatively high seismic hazard are situated in sedimentary basins where site response of unconsolidated sediments will be significant (see Somerville et al., 1991). Seismographic station distribution to sample both weak and strong ground motions is still not adequate.

At a recent workshop, the U.S. Committee for Advancement of Strong Motion Programs considered ways in which improvements should occur. The main points are as follows (Dr. R.D. Borchardt, Chairman):

- Earthquake Source Characterization

- The major seismicity across the U.S. is now being determined reasonably well by a National Seismic Network of about 150 sensitive, broad frequency band, digital earthquake observatories.
- The more frequent smaller earthquakes are now located in an inhomogeneous manner by several Regional Seismic Networks. Only a few instruments in these networks now are designed to serve both strong-motion and seismicity observation purposes.
- An enlarged national distribution of strong-motion instrumentation is required to determine the earthquake rupture process using stations distributed along the source. Only then can near-source motions and the nature of seismogenic failure be documented, such as effects related to the type of faulting, the stress regime in crustal rocks, seismic stress drop, rupture directivity, and distribution of fault slip and asperities.

- Path and Attenuation Characterization

- The regional characterizations of seismic-wave attenuation and other path effects are needed. These properties include the effects of anelastic damping in geologic materials, critical reflections from within the crust, alluvial basin effects, and the extended duration of surface wave trains.
- Uncertainty in existing wave attenuation relations needs reduction. The strong-motion measurements essential for estimating the variation in ground shaking amplitude and spectral energy with increasing distance are sparse for earthquake magnitudes and distances of greatest interest. Outside of coastal California, there are very few such observations to provide a reliable basis for engineering design and risk minimization plans.

- Local Site Characterization

- Local site response and near-surface site amplification and damping require characterization, such as the effects on seismic shaking of surficial geology, shear-wave velocity structure, lateral heterogeneity, and nonlinear soil effects. Geophysical and geotechnical measurements are needed at all instrument sites with a "significant" recording in order that such recordings can be properly



interpreted.

- Further instrumental recordings of ground shaking patterns much be secured to validate and calibrate assumptions and models used in the development of new maps of regional hazard being developed in the U.S.

## 6. Concluding Remarks

The above sections have stressed that, because of the relative observational paucity of seismic strong-motion measurements for earthquakes within the tectonic plates, both in stable oceanic and continental regions, the assessment of seismic hazard, and consequently seismic risk, therein is more uncertain than for interplate earthquakes along plate margins, such as transform fault and subduction zones. Nevertheless, research and application developments in recent years have strengthened many key theoretical and estimation aspects of both earthquake zoning and ground motion prediction. Promising algorithms, which are mathematically powerful and physically plausible, have now been published for seismic regionalization, seismic hazard mapping, and computation of seismic wave synthetics. The appropriate probabilistic properties have begun to be addressed.

Because verification and validity checks on hazard assessment algorithms are still necessarily limited, the conclusions, inferences, and computations on seismic risk based on them are critically dependent on model assumptions, uncertain selections of mathematical forms, and uncertain analogies between intraplate and interplate seismicity. It is important that these assumptions not become entrenched and continued criticism and debate are needed. Above all, it is crucial that seismographic instrumentation capable of recording strong ground motions not be restricted to the plate margins, but that resources be found to distribute judiciously strong motion instruments at locations that will eventually suffer intraplate earthquakes. As a start, at least one strong motion accelerometer with digital channels should be operated at every permanent seismographic observatory (such as stations of the IRIS network). In this way, accelerometers at such *Continental Reference Stations* (CRS) will inexorably provide the critical database for seismic hazard prediction.

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**Tectonics and earthquake potential of Cascadia:  
Effects of rotation and northward transport  
of forearc crustal blocks**

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The distribution of earthquakes, volcanoes, and young deformation in the Pacific Northwest may be explained if northward transport and clockwise rotation of forearc crustal blocks is occurring along the obliquely subducting Cascadia convergent margin. Although low historical rates of seismicity and the lack of a regional geodetic network presently precludes measurement of coastwise motion, geophysical and geological evidence for Neogene clockwise rotation and northward motion of large forearc blocks is abundant.

Deformation occurs mostly around the margins of a large, relatively aseismic Oregon coastal block composed of thick, accreted seamount crust that is moving slowly northward and rotating out over the trench at Cape Blanco. A robust extensional volcanic arc is built on the trailing edge of the rotating Oregon block. In the populated Puget-Willamette lowland, small, seismically active blocks are compressed against the Vancouver Island buttress by the northward moving Oregon block. Arc-parallel transport of Coast Range blocks is inferred to be about 0.5m per century, similar to that observed onshore in SW Japan, and sufficient to produce large earthquakes on a plexus of faults forming a broad deformation zone along the northern and eastern edge of the coastal blocks. The crustal earthquake hazard from this motion is in addition to that from subduction zone and deeper slab earthquakes, and it depends on the number, location and size of active faults along the block boundaries. Work is presently underway to map potentially active structures inferred from aeromagnetic and seismic data.

The offshore extension of the thick mafic Oregon block coincides with local minima in modern coastal uplift and horizontal shortening rates that are inferred to result from elastic bending above the locked subduction zone, thus suggesting that the block limits the area of strong interplate coupling. The composition of accreted terranes in contact with the downgoing slab (mafic seamount or accreted sediments) may affect seismic coupling between the plates by controlling the flow of heat and/or fluids from the fault zone and the size of the contact patch.

Finally, rotation of the forearc blocks causes the convergence rate to vary along strike, which may affect earthquake recurrence intervals along the subduction zone; perhaps not unexpected given the long term volcanic and seismic segmentation of the upper plate.

**Reference:**

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Thus the question arises as to what extent is the historical record accurately portraying the earthquake potential in the Pacific Northwest and how can it be used in seismic hazard evaluations given its shortcomings. (In this paper, I confine my discussion of the Pacific Northwest to principally Washington and Oregon.)

A very significant field of geologic studies taken in the past two decades to push the earthquake record back in time, particularly in the Pacific Northwest, has involved paleoseismology. Paleoseismic investigations along the coast of the Pacific Northwest have been the key to our understanding of the earthquake processes within the Cascadia subduction zone. Geologic studies to investigate crustal faults are, however, constrained in the region because few late Quaternary faults have been identified in western Washington and western Oregon. West of the Cascades, dense vegetation and relatively rapid erosion rates make it difficult to find evidence of young faulting (e.g., Pezzopane and Weldon, 1993). Alternative approaches such as subsurface imaging are now being carried out in the Puget Sound (e.g., Johnson *et al.*, 1996) and the Portland area (Blakely *et al.*, 1995). Because of resolution and age date limitations, however, such techniques yield at best, only approximate information on earthquake recurrence.

Based on the paleoseismic and geophysical investigations performed in the past decade, it is obvious that the historical record does not adequately portray the potential seismic hazards in the Pacific Northwest. In particular, the recent recognition that the megathrust within the Cascadia subduction zone and numerous crustal faults (e.g., Seattle fault) can produce earthquakes as large as moment magnitude ( $M_w$ ) 9 and  $M_w$  7+, respectively, are examples of such a deficiency. Despite these limitations, the historical record in general provides the only approach to characterizing most seismic sources that are beyond the visibility of geologists.

In addition to the historical record itself, the general lack of instrumental recordings of significant pre-1970s events has hindered our understanding of the earthquake processes and the identification of seismic sources in the Pacific Northwest. Also, with the exception of the few records of the 1949 and 1965 Puget Sound earthquakes, strong motion data to quantify crustal attenuation and geologic site effects are generally nonexistent. The available records do indicate site effects are important in regions such as the Puget Sound and most likely, the Willamette Valley (e.g., Silva *et al.*, 1997).

In this paper, I review the approaches used to incorporate the historical earthquake record into seismic hazard evaluations, discuss its limitations, and implications to hazard in the Pacific Northwest. Issues relevant to Oregon are emphasized, such as the potential for intraslab (Wadati-Benioff zone) events and 1872-sized crustal earthquakes, because the record is even more limited than in Washington. I also discuss the implications of the available strong motion data.

## **HISTORICAL SEISMICITY**

More than 33,000 earthquakes are contained in the 176-year-old historical record of the Pacific Northwest (1820-1996) of which about 26,000 events occurred in Washington and 7,000 events in Oregon. The vast majority of these earthquakes are smaller than  $M_L$  3.0 in size (microearthquakes). A total of 95 earthquakes of approximate  $M_L$  5.0 (Modified Mercalli [MM] VI) and larger are known; 83 events in Washington but only 12 in Oregon. The largest known earthquake is the 14 December 1872 North Cascades event in Washington, which has been the subject of extensive studies (e.g., Malone and Bor, 1979). Despite these studies, only an approximate location for this earthquake has been estimated, somewhere in the North Cascades east of Seattle, and its size is still somewhat controversial (Figure 1). The most definitive study performed to date, by Malone and Bor (1979), places the event near Ross Lake and they assign a  $M$  7.4 based on an

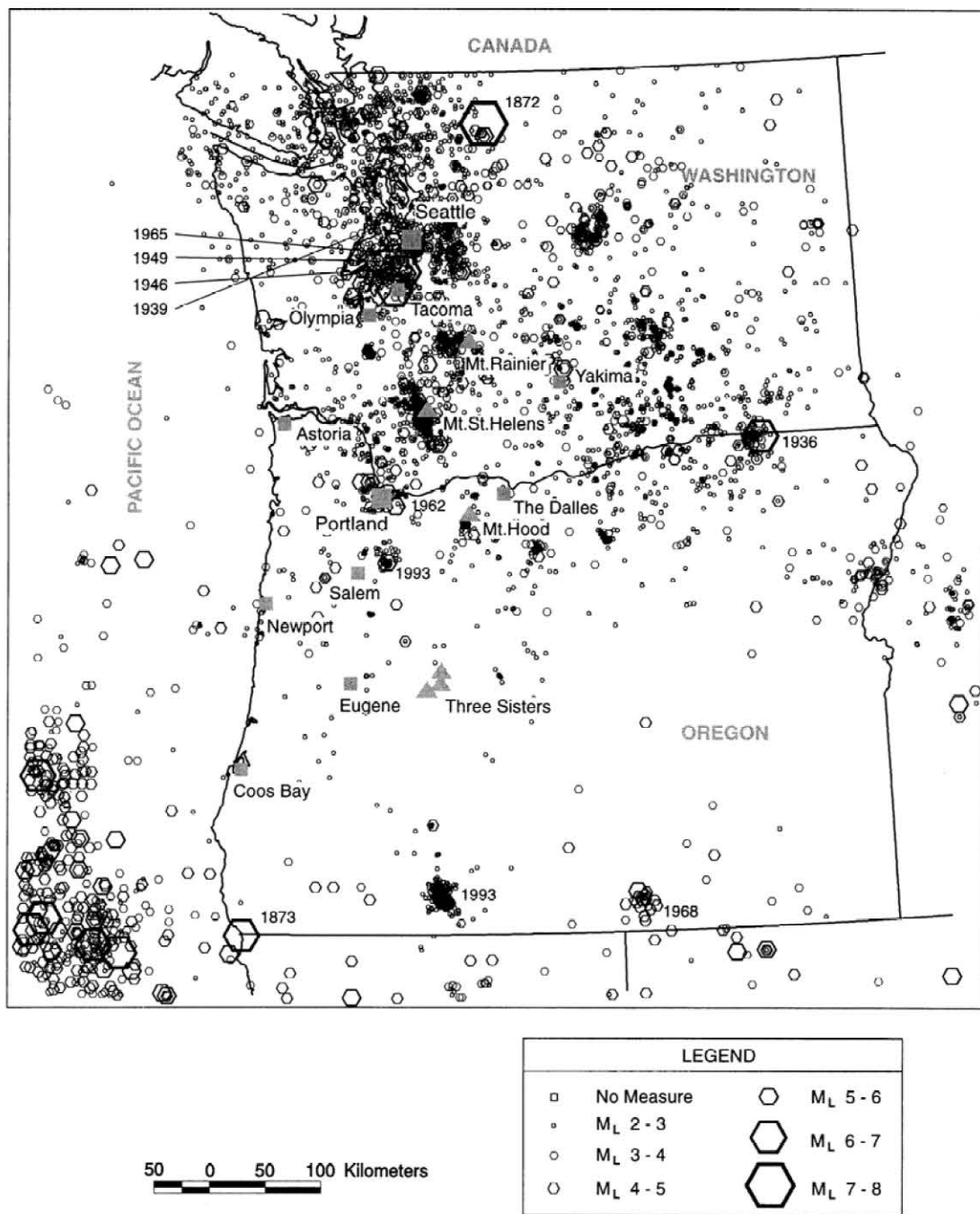


Figure 1. Historical seismicity, 1841 through 1996,  $M_L$  2.0 and greater. Data courtesy of University of Washington and U.S. Geological Survey.

analysis of its felt area and crustal attenuation. The occurrence of this large magnitude, apparently crustal, earthquake is extremely important with respect to potential seismic hazards in the Pacific Northwest.

Possibly the largest historical earthquake in Oregon occurred on 23 November 1873. This event of estimated  $M_L$  6<sup>3</sup>/<sub>4</sub> occurred near the Oregon-California border (Figure 1) with a maximum reported intensity of MM VIII in the Smith River Valley north of Crescent City, California (Topozada *et al.*, 1981). Chimneys were knocked down in Crescent City, Port Orford, Grants Pass, and Jacksonville. 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 (see Figure 4 in Topozada *et al.*, 1981), its epicentral uncertainty is large, and the event could have occurred in northernmost California or southernmost Oregon. The lack of aftershocks led Ludwin *et al.* (1991) to suggest that the earthquake may have occurred within the subducting Gorda (or Juan de Fuca) plate of the Cascadia subduction zone. Alternatively, the event may have been crustal in origin and occurred far enough offshore such that no aftershocks were felt (Ludwin *et al.*, 1991).

In the next seven decades after the 1870s, several earthquakes between M 6 and 7 occurred in Washington (none in Oregon) but none was particularly damaging. (Two M 7 earthquakes occurred in 1918 and 1946 in the northern Puget Sound of British Columbia.) However, in 1949 the southern Puget Sound was struck by a  $M_s$  7.1 earthquake just before noon on 13 April. The event had its origin at a depth of 54 km beneath an area between Olympia and Tacoma (Baker and Langston, 1987). The event caused eight deaths and numerous injuries, and \$150 million (1984 dollars) in damage (Noson *et al.*, 1988). This earthquake, which occurred in the subducting Juan de Fuca plate beneath the North American continental plate, was followed by another intraslab event of  $M_s$  6.5 on 29 April 1965. This earthquake killed seven people, injured many, and caused \$50 million (1984 dollars) in damage (Noson *et al.*, 1988). Like the 1949 event, the impact of the 60-km-deep 1965 earthquake was somewhat minimized because of its great depth.

In Washington, significant earthquakes in addition to those in 1872, 1949, and 1965 are the 12 November 1939  $M_s$  5<sup>3</sup>/<sub>4</sub> event, which was felt at a maximum intensity of MM VII, and the 14 February 1946  $M_s$  6.3 event (MM VII), both of which occurred in the southern Puget Sound. This earthquake caused damage in Seattle, Tacoma, and Olympia (Noson *et al.*, 1988) (Figure 1).

In Oregon, other significant earthquakes besides the 1873 event include the 16 July 1936  $M_L$  6.1 Milton-Freewater event; the recent 25 March 1993  $M_L$  5.6 Scotts Mills event; and the 21 September 1993  $M_L$  5.9 and 6.0 Klamath Falls events (Figure 1). Of special note was the 6 November 1962  $M_L$  5<sup>1</sup>/<sub>2</sub> ( $M_w$  5.2; Yelin and Patton, 1991) Portland earthquake, which actually had its epicenter in neighboring Vancouver, Washington (Figure 1). This event was possibly the most significant Oregon earthquake up to its time because it came as a surprise to most residents in the region particularly since it was so widely felt and because damage, though not major, was significant (Dehlinger and Berg, 1962). For more comprehensive discussions of the historical seismicity of the Pacific Northwest, I refer readers to papers by Noson *et al.* (1988), Ludwin *et al.* (1991), Bott and Wong (1993), Yelin *et al.* (1994), Wong and Bott (1995), and Weaver and Shedlock (1996).

### Limitations of the Historical Record

For most of the western U.S. outside the broad Pacific-North American plate boundary in California, paleoseismic investigations indicate that the recurrence intervals of most crustal faults range from a few thousands to more than 100,000 years. Based on limited studies, faults in the Pacific Northwest appear to have similar rates of activity. For example, studies of the Seattle fault suggest that no event has occurred for at least 5,500 years prior to the most recent earthquake, which occurred at about 900 A.D. (Bucknam *et al.*,

1996). Thus, as is often the case, the historical seismicity record only samples a fraction of typical crustal fault recurrence intervals resulting in an incomplete picture of the seismicity of the Pacific Northwest.

To what extent the historical earthquake record may be incomplete, or for that matter representative, is usually not resolvable. Even for the period of coverage, in this case 1820 to present, the completeness of the record varies significantly with detection capability. Prior to seismographic coverage of the Pacific Northwest, the historical record is probably only complete for earthquakes larger than  $M_L$  6 (Ludwin *et al.*, 1991; Wong and Bott, 1995). Detection of any earthquakes during this time was based solely on observers' reports which are a function of population distribution and density. As seismographic coverage of the region slowly improved through the first half of the 1900s, so obviously did the completeness of the historical record. The detection of small magnitude earthquakes,  $M_L$  2.5 and smaller, in the Pacific Northwest did not come about until after 1970 when the University of Washington began installation of their regional seismographic network (Ludwin *et al.*, 1991).

Besides being brief and incomplete, the pre-instrumental catalog contains many events that have no or uncertain estimated magnitudes and locations. A significant example is the size and location uncertainties of the 1872 North Cascades earthquake. Uncertainties in both magnitude and location also exist for many of the early instrumentally recorded events. These uncertainties can have a significant impact on the estimation of earthquake recurrence. The lack of focal depth estimates of the vast majority of pre-instrumental earthquakes also hinders our understanding of seismic sources in the Pacific Northwest.

## APPROACHES IN EVALUATING SEISMIC HAZARDS

In seismic hazard evaluations, whether they are deterministic or probabilistic in nature, the historical earthquake record may provide the only basis for characterizing some seismic sources. Given the catalog limitations, statistical techniques have been developed in the past few decades to try to compensate for weaknesses in catalogs such as incompleteness (e.g., Stepp, 1972) and magnitude uncertainties and variabilities (e.g., Zúñiga and Wyss, 1995). A problem of particular note has been the assignment of magnitudes to pre-instrumental earthquakes based on intensity data. Standard practice has been to use the relationship between maximum MM intensity and magnitude developed by Gutenberg and Richter (1956). In their analysis of Washington's largest earthquakes, Noson *et al.* (1988) estimated magnitudes based on felt area using a California-based relationship. Both approaches can result in significant uncertainties in magnitude and, hence, in estimating earthquake recurrence.

Numerous studies have been performed to investigate significant pre-instrumental earthquakes including estimating their size (e.g., 1872 earthquake by Malone and Bor, 1979) and to search for events not previously included in the historical record. These efforts, however, can only focus on the time period since the region has been settled. My discussion of how the historical earthquake record is used in seismic hazard evaluations follows.

In state-of-the-art seismic hazard evaluations, two categories of active seismogenic faults are considered: (1) those that are known (mapped) either because they are expressed at the earth's surface or known from subsurface imaging (e.g., seismic reflection surveys) and (2) unknown or buried (hidden) faults that have no surficial expression. With regard to the latter, the concept of the "random" earthquake has been traditionally used in seismic hazard analyses. The random earthquake approach, where events are assumed to occur randomly in both space and time (Poissonian), has been particularly useful in addressing the hazard from "background" seismicity. Background events are those that do not appear to be associated with known geologic structures. In the Pacific Northwest, much of the observed seismicity appears to be background in nature and, thus, buried faults appear to be prevalent throughout the region.

Adequately characterizing background seismicity is crucial in many regions because in probabilistic assessments, they often dominate the seismic hazard at short return periods (high annual probabilities of occurrence) (e.g., Wong and Olig, 1997). To address the seismic hazard from background earthquakes, I am aware of three basic approaches. The first approach attempts to define a probabilistic epicentral distance for the random earthquake of specified magnitude. Based on earthquake recurrence parameters for the region of interest and the assumption that events are randomly distributed, the radius of a circle about a site can be calculated for the probability of an earthquake occurring in the magnitude interval  $M \pm \Delta M/2$  during a time  $t$  (Wood and Ostenaar, 1984). This probabilistic approach to obtain a specific earthquake for seismic design or safety evaluation is often used in deterministic analyses in an attempt to address ground-shaking hazard from the random earthquake. The U.S. Bureau of Reclamation consistently uses this approach in their dam safety studies (Wood and Ostenaar, 1984) and has done so for numerous dams in the Pacific Northwest.

The second approach is the traditional use of areal source zones or seismotectonic provinces in a Cornell-McGuire-type probabilistic seismic hazard analysis. In this approach, it is assumed that the seismicity characteristics of a specified area are generally uniform such that the assumption of a random distribution of earthquakes is valid. Based on the maximum magnitude of the background earthquake and the earthquake recurrence of the specified area, the hazard from these areal sources can be incorporated into probabilistic analysis. In the Oregon Department of Transportation statewide probabilistic ground motion maps (Geomatrix Consultants, 1995), the state was subdivided into 15 seismotectonic provinces (e.g., Southern Willamette Valley, Western Blue Mountains) and the hazard was calculated from each of these provinces as well as from all identified active faults.

Unlike the other two techniques, the final approach does not require the assumption of random earthquake occurrence. In this approach, it is assumed that the past is the key to the future, i.e., the historical seismicity will reflect sites of future activity. The historical seismicity is smoothed by a set of Gaussian filters and the hazard is calculated directly from this distribution by assuming a maximum magnitude and the recurrence within each grid cell. For example, the recent national ground-shaking hazard maps produced by the U.S. Geological Survey used a Gaussian smoothing technique (Frankel *et al.*, 1996). They also employed areal source zones to account for the potential nonstationarity of seismicity. In the Pacific Northwest, the areal source zones were a Cascades zone superimposed on a broad zone that covers most of the western U.S. (Frankel *et al.*, 1996).

### **Assessment of Maximum Background Earthquake**

In any approach to address the hazard from background earthquakes, the maximum magnitude needs to be defined. In the western U.S., the conventional approach has been to assume that the minimum threshold for surface faulting represents the upper size limit for background earthquakes. In much of the western U.S., this threshold ranges from a  $M_L$  (or  $M_w$ ) 6 to  $6\frac{1}{2}$  (e.g., Arabasz *et al.*, 1992; dePolo, 1994). It is believed that earthquakes larger than  $M_L$   $6\frac{1}{2}$  will be accompanied by surface rupture, and repeated events of this size will produce recognizable fault-related geomorphic features.

In the Pacific Northwest, west of the Cascades, however, the threshold of surface faulting is not defined. Possibly the most important factor dictating this threshold is the thickness of the seismogenic crust. In most of the western U.S., the seismogenic thickness ranges from about 15 to 20 km (Smith and Bruhn, 1984; Wong and Chapman, 1990). In contrast, for the region west of the Cascades, the seismogenic crust appears to be quite thick, from 25 to 30 km based on observed seismicity (Ludwin *et al.*, 1991). This thickened seismogenic crust is often observed in regions overlying subduction zones (Tichelaar and Ruff, 1993). Thus the surface faulting threshold in this portion of Washington and Oregon could be  $M_w$  7 or higher, because faults may be deeply seated and therefore may never rupture to the surface (Wong *et al.*, 1994). A schematic



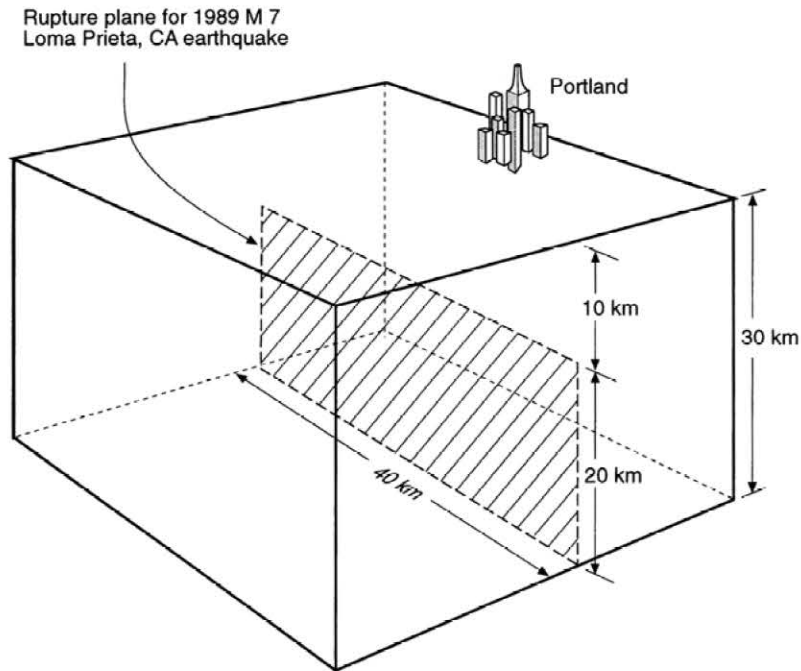


Figure 2. Schematic illustration showing how earthquakes up to  $M_w$  7 can exhibit no geologic surficial expression west of the Cascades.

illustration of this point using the rupture dimensions of the 1989  $M_w$  7.0 Loma Prieta, California, earthquake is shown on Figure 2.

This possibility, in addition to the fact that the source of the largest crustal earthquake in the historical record, the 1872  $M$  7.4 North Cascades event, is unknown (possibly indicating a buried fault) requires that a maximum background earthquake of at least  $M_w$  7 be considered in seismic hazard analyses in western Washington. Whether this is the case for western Oregon will be discussed later.

In eastern Washington and Oregon, seismicity appears to be confined to typical upper crustal depths less than 15 km (Ludwin *et al.*, 1991). Thus a surface-faulting threshold and maximum background earthquake of  $M_w$  6 $\frac{1}{2}$  appears appropriate particularly for most of eastern Oregon, which is located within the Basin and Range Province.

In recognition that a crustal background earthquake of at least  $M_w$  6.5 can occur anywhere in Oregon, state building code provisions require that such an event be considered as a minimum in seismic design studies for critical and high-occupancy facilities (M. Mabey, Brigham Young University, personal communication, 1997).

#### Identification of Active Faults and Seismic Zones

It has long been recognized that the association of seismicity with a geologic structure was sufficient evidence to demonstrate that the structure was active and seismogenic. Although cases have been relatively

few in number, recent earthquake activity has identified both mapped and unmapped active faults in the Pacific Northwest (e.g., Wong and Bott, 1995). In Oregon, mainshock-aftershock sequences such as the 1993 Scotts Mills and Klamath Falls earthquakes have delineated the mostly buried Mt. Angel fault (Thomas *et al.*, 1996) and the Lake of the Woods fault zone (Braunmiller *et al.*, 1995), respectively, although neither was preceded by any significant historical seismicity. Other faults in Oregon that appear to have manifested themselves seismically include an unmapped fault near Adel in 1968 (Wong and Bott, 1995), possibly the Mt. Hood fault (Weaver *et al.*, 1982; Geomatrix Consultants, 1995), and the Portland Hills and Frontal fault zones (Blakely *et al.*, 1995).

In Washington, seismicity has delineated two prominent seismic zones, the St. Helens (Weaver and Smith, 1983) and the Western Rainier (Stanley *et al.*, 1996) zones, which exhibit no late-Quaternary geologic expression at the surface, although the former intersects the volcano of the same name. No specific faults or fault zones have been identified within these "seismic zones," thus accounting for the use of the term although it is apparent that the causative structures are faults. Other notable seismic source zones include the Darrington zone east of Seattle (Zollweg and Johnson, 1989) and the Goat Rock zone (Stanley *et al.*, 1996). Seismicity may also be associated with the Seattle fault (Gower *et al.*, 1985) and the Southern Whidbey Island fault (Johnson *et al.*, 1996). Thus instrumentally recorded seismicity in the Pacific Northwest has been valuable in delineating active faults and seismic zones, with and without surficial expression and will undoubtedly reveal other sources in the future.

### Estimating Earthquake Recurrence

A key element in evaluating seismic hazards, particularly on a probabilistic basis, is the assessment of earthquake recurrence. Since the 1940s, it has been shown that the earthquake recurrence in a region can be characterized by the Gutenberg-Richter relationship of  $\log N = a - bM$ , where  $N$  is the annual cumulative number of earthquakes,  $M$  is magnitude, and  $a$  and  $b$  are the familiar recurrence parameters. To properly calculate the recurrence parameters of an areal source zone for use in probabilistic seismic hazard analysis, adjustments need to be made to the historical catalog.

For example, Bott and Wong (1993) calculated the recurrence for the Portland region, following the maximum-likelihood procedure developed by Weichert (1980). The historical record was adjusted for incompleteness using the procedure of Stepp (1972), and dependent events (foreshocks and aftershocks) were removed to satisfy the assumption of Poisson behavior of independent events (Also, if active faults are included in the hazard analysis, those thought to be associated with such faults need to be removed to avoid double-counting their contribution to hazard). All event magnitudes were converted to equivalent  $M_L$  values. The recurrence parameters of  $b$  and  $a$  of  $0.84 \pm 0.07$  and 2.55, respectively, were estimated for the Portland region. This recurrence results in a return period for earthquakes of  $M_L$  6.0 and greater of about 325 years, with the uncertainty in this value being at least several decades. For  $M_L$  5.5 events and greater, the return period of 100 to 150 years is consistent with the occurrence of the 1962 Portland and 1993 Scotts Mills earthquakes in the 150-year historical period. For  $M_L$  6.5 and greater earthquakes, the return period is estimated to be approximately 800 to 900 years.

Over the years, the calculation of seismicity recurrence parameters for hazard analysis has unfortunately taken on the appearance of being simple and straightforward. And yet the uncertainties in the resulting parameters have become somewhat unappreciated in standard practice. For instance, it is often the case, particularly in regions outside the Puget Sound, that the historical record, even before all the proper adjustments have been made, contains only a small number of events (e.g., eastern Oregon). Of course, what constitutes a sufficient number of events to calculate robust recurrence parameters cannot really be answered in a general way. Certainly, the range in magnitudes of such events should be as wide as possible, at

least more than 3 to 4 magnitude units, and should hopefully include larger events in the region. The uncertainties in magnitude values, particularly for pre-instrumental events and in converting from different scales, the issues regarding the definition of a seismotectonically uniform area in which the recurrence is to be calculated, the adequacy of current techniques to remove all dependent events whose behavior appears to vary from region to region, and a host of other issues raised in research the past few decades, makes the estimation of recurrence a very uncertain business. Finally, the question always remains as to whether the historical record is adequately sampling the long-term seismicity behavior of the region. Despite all these issues, recurrence estimation based on the historical record has been proven to be a very viable approach if properly done and if based on an adequate record.

Paleoseismology has been a vital tool in assessing the recurrence of surface-faulting earthquakes on active faults. However, for an estimation of the recurrence of smaller nonsurface faulting yet potentially damaging earthquakes along a fault, it is generally assumed that the slope of the recurrence curve is the same as the  $b$ -value for the region calculated from the historical catalog. This is the case regardless of the recurrence model (exponential or characteristic) assumed appropriate for the fault.

To estimate the recurrence for seismic sources beyond the visibility of geologists, the historical earthquake record is the only viable approach. For example, the recurrence of the intraslab region of the Cascadia subduction zone in Washington, the source of the destructive 1949 and 1965 Puget Sound earthquakes, can only be determined by the historical record at this time. Geomatrix Consultants (1995) calculated a return period for  $M_w$  7 and greater Benioff earthquakes of about 130 years for the Washington portion of the subduction zone.

## **IMPLICATIONS TO SEISMIC HAZARDS**

An examination of the historical earthquake record raises several other issues regarding the earthquake potential of the Pacific Northwest and its implications to seismic hazards. My discussion of these issues follows.

### **Intraslab Earthquake Potential in Oregon**

Prior to the last decade, the absence of observed seismicity along the megathrust of the Cascadia subduction zone essentially led many earth scientists to conclude that large earthquakes were not possible and that the plate interface was aseismic. However, paleoseismic studies in the past decade have indicated the contrary (e.g., Atwater et al., 1995) and now the earth science community has adopted the position that events of  $M_w$  8 to 9 have and can occur along the megathrust.

In contrast, the occurrence of the 1949 and 1965 earthquakes as well as several other historical events greater than  $M_L$  5 in the Puget Sound region (Figure 1) attest to the significant earthquake potential within the Juan de Fuca plate. Abundant small magnitude seismicity as recorded by the University of Washington's Pacific Northwest network images quite well the geometry of this subducting plate in western Washington. This same seismic monitoring, however, also indicates that very few events appear to be associated with the Juan de Fuca plate in western Oregon (Figure 3).

Unlike the Puget Sound region, no large historical Benioff earthquakes have occurred in Oregon with the possible exception of the 1873 Crescent City earthquake. Weaver and Shedlock (1996) propose that the potential for large intraslab earthquakes, which will occur at depths of 45 to 60 km, extends throughout western Washington and as far south as westcentral Oregon. Because their proposed intraslab source zone

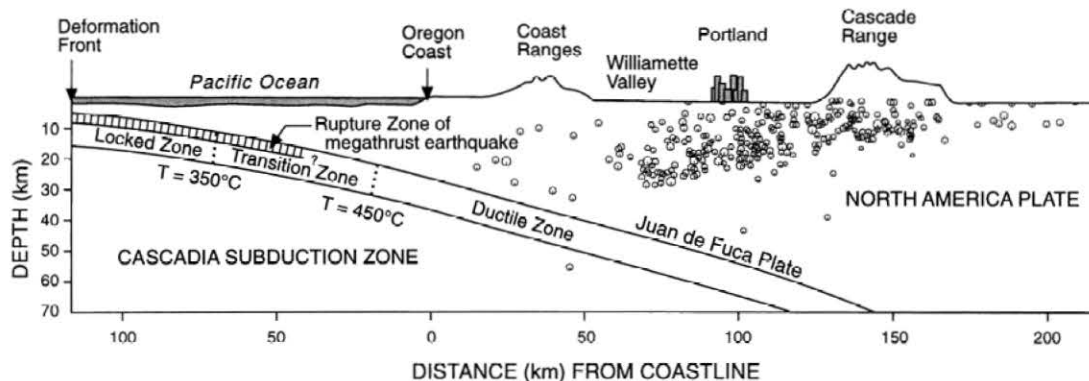


Figure 3. Cross-sectional view of well-located seismicity (1972-1994) through the Cascadia subduction zone at the latitude of Portland. Model of the subduction zone is from Hyndman and Wang (1993).

is based on both the plate geometry and historical seismicity, Weaver and Shedlock (1996) chose not to characterize southwestern Oregon as a source zone with the exception of the 1873 epicentral area.

Rogers *et al.* (1996) suggest that the low rate of seismicity in Oregon may be due to (1) the megathrust is unlocked and the continental crust and slab are in a low stress state or (2) the megathrust and overlying crustal faults are locked and in a stage of the seismic cycle where few events are produced. These opposing explanations underline the fact that the historical earthquake record cannot in itself be used to discern either of the two models. Based on a very poor historical catalog, Geomatrix Consultants (1995) estimated that the return period for  $M_w$  7 and larger Benioff earthquakes beneath western Oregon is about 800 to 900 years or about 6 to 7 times longer than the return periods for western Washington.

At this time, I believe the question of whether large Benioff earthquakes are possible beneath western Oregon remains unanswered. Further studies such as recent deep seismic imaging (e.g., Trehu *et al.*, 1996) may shed new light on the physical characteristics and geometry of the Juan de Fuca plate beneath western Oregon. It is probably these properties that control the rate of internal deformation within the subducting plate and dictate whether large Benioff earthquakes are possible.

Assuming that a potential for large intraslab earthquakes beneath western Oregon exists, what is their maximum magnitude? Based on the largest Benioff earthquakes observed worldwide, the maximum size of these events is about  $M_w$  8. The maximum size of the intraslab earthquake assumed appropriate for the Cascadia subduction zone has been based on the size of the 1949 earthquake and the relative thinness of the subducting plate and is suggested to be  $M_w$  7.5 (Rogers *et al.*, 1996; Ludwin *et al.*, 1991). This value has generally been used in seismic hazard evaluations in the Pacific Northwest. Whether this maximum magnitude is appropriate for Oregon also remains unresolved at this time.

### Crustal Seismicity in Oregon

Crustal seismicity in Oregon appears to be an order of magnitude lower in occurrence than in Washington (Figure 1), due to some degree to the sparser seismographic coverage of the state. It is obvious, however,

that the difference indicates a lower seismicity rate in Oregon as compared to Washington. There are several significant issues regarding crustal seismicity in Oregon. First, what is the largest crustal earthquake that can occur in the state? Paleoseismic studies of faults in eastern Oregon indicate that earthquakes of at least  $M_w$  7 in size have occurred in prehistoric times and are therefore likely in the future (Hemphill-Haley *et al.*, 1993; Pezzopane and Weldon, 1993).

In contrast, no such faults capable of generating  $M_w$  7 events have been identified to date in western Oregon, and as previously described, no historical earthquakes greater than  $M_w$  6 have occurred with the possible exception of the 1873 Crescent City event. As discussed earlier, faults of sufficient dimensions capable of generating  $M_w$  7 and greater earthquakes can potentially occur in western Oregon without detection (Figure 2). Blakely *et al.* (1995) have identified several mostly buried faults in the Portland Basin whose lengths suggest that they could generate approximately  $M_w$  7 earthquakes assuming that the entire fault ruptures in a single event. However, very little is known about the geometry of these faults and the question of whether they are even seismogenic remains unanswered. The 40-km-long Oatfield fault as depicted by Blakely *et al.* (1995), for example, appears to be structurally segmented and thus it seems unlikely that this fault could generate a  $M_w$  7 earthquake. More extensive studies of these faults in the Portland Basin will be required to define their earthquake potential.

Based on the recurrence calculated by Bott and Wong (1993) for the Portland region, possibly the most seismically active area in Oregon, the return period of  $M_w$  7 earthquakes is probably on the order of several thousands of years and thus it is not surprising that no such event has been observed during historical times. Other regions in western Oregon, which are less seismically active, will have even longer average return periods for  $M_w$  7 earthquakes.

In summary, although no definitive geologic evidence indicates active faults in western Oregon that have and could rupture in  $M_w$  7 or larger earthquakes, no reasonable geologic or seismologic arguments are against such a possibility. Given the lower level of seismicity and possibly lower crustal strain rates in western Oregon relative to western Washington (where  $M_w$  7+ events have occurred), the frequency of large earthquakes in western Oregon is likely to be much lower. The compressional tectonic stress regime of western Oregon (Zoback and Zoback, 1989) indicates that such events will be the result of reverse/thrust, strike-slip, or oblique faulting. The 1993 Scotts Mills mainshock was due to reverse faulting (Thomas *et al.*, 1996) and small earthquakes in the Portland Basin appear to result from both strike-slip and reverse faulting based on focal mechanism data (Yelin and Patton, 1991; Yelin, 1992) (Figure 4). The existence of blind reverse and thrust faults capable of  $M_w$  6½ and larger earthquakes in regions analogous to the Portland Basin are not uncommon. Two examples include the Puget Sound and the Los Angeles basins.

A second issue is whether crustal seismicity is stationary in western Oregon. In other words, do sites of historical and current seismicity reflect sites of future earthquake activity? Based on Oregon's historical record, the answer is equivocal. A possible example of stationarity is the Portland area, which has been one of the most, if not the most, persistently active area in the state (Bott and Wong, 1993). Earthquakes of  $M_L$  5 and greater have been observed since 1877. The Pine Valley Graben-Cuddy Mountain area east of Baker is another area of fairly continuous seismicity in Oregon (Wong and Bott, 1995).

In contrast, the 1993 Scott Mills earthquake, though possibly preceded by several small events at the northernmost end of the Mt. Angel fault (Werner *et al.*, 1992), was considered a surprise in terms of its location and particularly magnitude. The Mt. Angel fault was previously thought to be a potentially active fault based on a few geologic observations (Unruh *et al.*, 1994). Similarly, the 1993 Klamath Falls earthquakes were not preceded by any known significant historical seismicity (Wiley *et al.*, 1993).



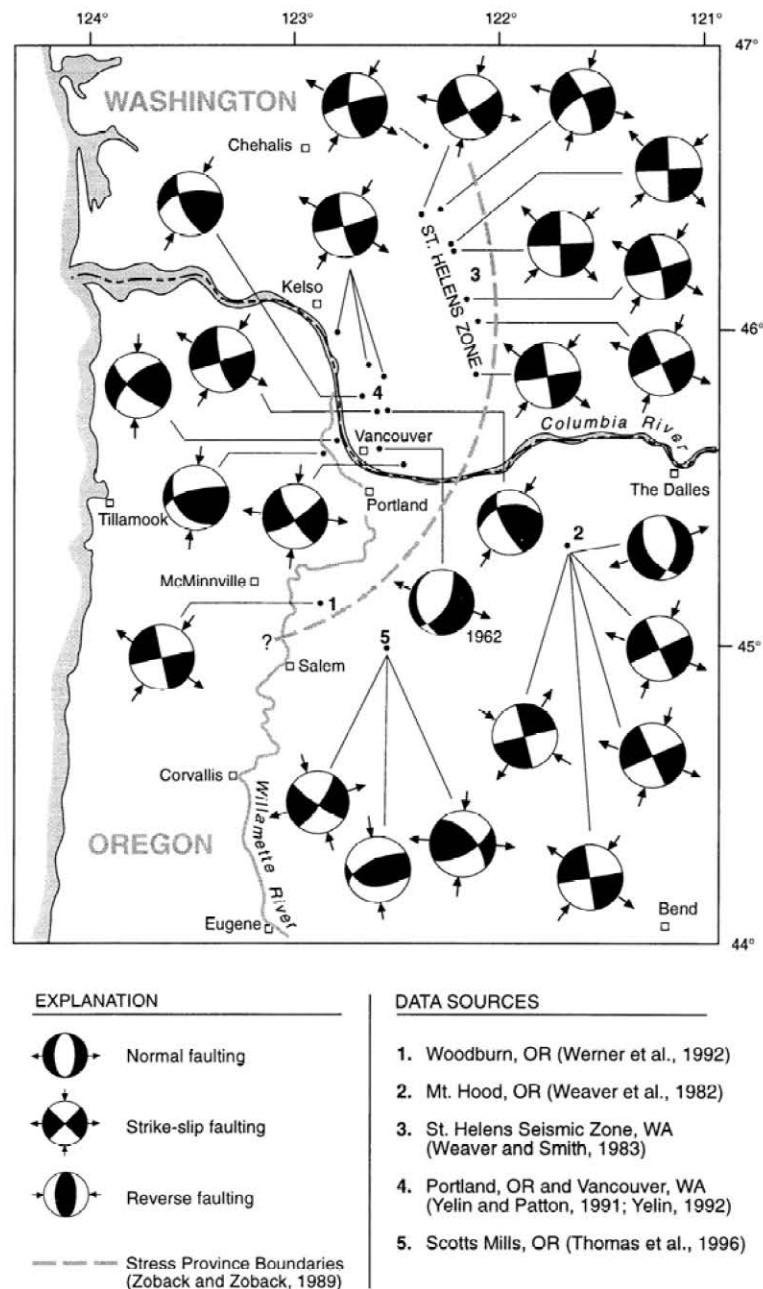


Figure 4. Crustal earthquake focal mechanisms of northwestern Oregon and southwestern Washington.

Given what we believe are long recurrence intervals for most of the crustal faults in Oregon, it is not unexpected that seismicity is to any great degree, nonstationary. Such observations are typical of most if not all of the western U.S. outside of California. Thus the historical seismicity record cannot be totally relied upon

to predict areas of future seismicity in Oregon. Only extensive fault investigations can significantly improve our knowledge of potential seismic sources in western Oregon although as discussed earlier, many important buried faults will likely go uncharacterized due to their lack of surface expression.

## STRONG EARTHQUAKE GROUND SHAKING AND HISTORICAL EARTHQUAKES

Instrumental recordings, both seismograms and accelerograms, provide much of the basis for our understanding of earthquake processes and earthquake effects. Given the relatively late seismographic coverage of Washington (since 1970) and in particular, Oregon, our knowledge of the sources and source processes of earthquakes in the Pacific Northwest has been hampered. This issue also extends to the characterization of strong earthquake shaking in the region. The Pacific Northwest has had relatively few strong motion instruments (accelerographs) and thus few records of significant historical earthquakes are available. Table 1 lists those significant free-field records ( $>0.01$  g) that are known to exist. The largest earthquake in modern times, the 1949 Olympia event, was only recorded at two strong motion sites. The second largest earthquake, the 1965 Seattle-Tacoma event (the Portland record was less than 0.01 g), has only four free-field stations recordings (two horizontal and a vertical component). The 1962 Portland earthquake was recorded

**TABLE 1**

### SIGNIFICANT FREE-FIELD STRONG MOTION RECORDS

Earthquake	Station	Epicentral Distance	Focal Depth	Peak Horizontal Accelerations	Maximum Intensities
13 April 1949 $M_s$ 7.1	Olympia Highway Test Lab	5 km	54 km	0.16 g 0.28 g	MM VIII in Tacoma, Olympia, and Seattle
	Seattle Corps of Engineers Office	60 km	54 km	0.07 g 0.07 g	
29 April 1965 $M_s$ 6.5	Seattle Federal Office Building	21 km	60 km	0.07 g 0.07 g	MM VII in Tacoma, Olympia, and Seattle; MMVIII in localized areas of Seattle and vicinity
	Olympia Highway Test Lab	61 km	60 km	0.14 g 0.20 g	
	Tacoma County-City Building	21 km	60 km	0.06 g 0.06 g	
25 March 1993 $M_L$ 5.6	Detroit Dam Downstream	43 km	15 km	0.06 g 0.06 g	MM VII in Woodburn and Molalla

(0.10 g) but only by a non-free-field instrument in the basement of the 11-story Portland State Office Building.

Though few in number, the available strong motion records have been invaluable in both seismological research and for direct use in seismic design studies. In particular, the 1949 and 1965 strong motion records have provided the bases for several studies of earthquake ground shaking in the Puget Sound region. Langston (1981) modeled ground motions from the 1965 earthquake and concluded that attenuation due to the thick sediments beneath Seattle and Tacoma is significantly higher than beneath Olympia. In a follow-up study, Langston and Lee (1983) modeled the ground motions in the Duwamish Valley and noted that the observed damage in the 1965 earthquake could be explained by significant amplification that occurred due to basin focusing and impedance contrasts within the basin sediments. Ihnen and Hadley (1986) used three-dimensional ray-tracing techniques to simulate the 1965 records and also noted significant amplification due to the basin focusing and soft soils. Silva *et al.* (1997) modeled the 1949 and 1965 events using a stochastic numerical modeling technique to assess the effects of the unconsolidated sediments. They concluded that frequency-dependent soil amplification and damping controlled the amplitudes and spectral content of the ground motions at periods less than 1.0 sec. Thus the available strong motion data have been crucial in understanding earthquake ground shaking in the Puget Sound region. However, due to the lack of records for Oregon, no similar strong motion evaluations based on empirical data have been performed. Because the large majority of the population in the Pacific Northwest resides in the Puget Sound and Willamette valleys, it is vital that site and basin effects on ground motions be fully understood and predicted (Wong *et al.*, 1993; Silva *et al.*, 1997).

One of the most important uses of strong motion is the development of region-specific attenuation relationships. The lack of such records for crustal earthquakes in the Pacific Northwest, however, has prevented the development of such relationships. Current seismic hazards practice has assumed that the crustal earthquake relationships derived from principally California strong motion data (e.g., Boore *et al.*, 1993) are appropriate for Washington and Oregon. This may or may not be the case (e.g., Atkinson, 1995). Likewise, current subduction zone relationships used in the Pacific Northwest (e.g., Crouse, 1991 and Youngs *et al.*, 1997) have been based on data from other subduction zones such as Japan, Mexico, and Chile and/or numerical modeling. Given that attenuation in the Cascadia subduction zone may be different from these other subduction zones and that these differences will be more pronounced at long distances (beyond 50 km), the evaluation of the appropriateness of these other relationships to the Pacific Northwest is critical. A recent set of preliminary relationships for Cascadia has been developed by Atkinson and Boore (1997) based on numerical modeling. Their relationships appear to underpredict ground motions for large events ( $M_w > 7$ ) at distances of 100 km or more based on comparisons with strong motion data from other subduction zones.

It has now been verified in recent earthquakes such as the 1994 Northridge, California, that ground motions generated by reverse faults can be significantly higher than for strike-slip faults at distances less than 20 km (Abrahamson and Somerville, 1996) and also that near-field source effects such as rupture directivity can be important (Somerville *et al.*, 1997). The presence of active reverse faults such as the Seattle fault in Seattle and possibly the Portland Hills fault in Portland would suggest that understanding these types of source effects on ground motions is important. The lack of strong motion data even for the moderate-sized events that have occurred such as the 1993 Scotts Mills earthquake obviously hinders this understanding.

A final significant issue is the value of dynamic stress drops for Pacific Northwest earthquakes. Increasingly, the importance of stress drops in controlling the levels of high frequency ground shaking has been realized. The average stress drop for western U.S. crustal earthquakes is 70 to 100 bars (Atkinson, 1995). Based on an analysis of an extensive data set of recordings from crustal and Benioff events of  $M$  3 to 7, Atkinson (1995) estimated an average Brune stress drop of 30 bars. Unfortunately, to my knowledge, no

dynamic stress drops have been computed for the 1949 and 1965 Puget Sound intraslab earthquakes. A lower stress drop would indicate lower high-frequency ground motions for a given magnitude. Further studies on earthquake stress drops with an emphasis on attempting to find differences between events from different source regions (interplate, intraplate, and crustal) should be performed.

## CONCLUSIONS

The historical earthquake record for the Pacific Northwest extends back to 1820. Although this record only provides a small sample of the long-term earthquake behavior in the region, it provides a valuable and for some uses, the only means by which some types of seismic sources can be characterized. Without such characterizations, reliable seismic hazard evaluations in the Pacific Northwest would not be possible. In particular, recorded seismicity has provided (1) the basis for identifying several seismic sources that possessed no surficial expression and (2) has allowed for an estimation of the earthquake recurrence of the Benioff zone of the subducting Juan de Fuca plate and crustal background seismicity. Significant issues particularly relevant to Oregon raised by this albeit incomplete historical record are (1) does a potential for large Benioff earthquakes ( $M_w > 7$ ) exist?; (2) are  $M_w$  7 and larger crustal earthquakes possible in the western half of the state?; and (3) does the historical record provide clues to sites of future significant seismicity or should one assume that significant earthquakes can occur anywhere? It is obvious that further studies particularly in paleoseismology are necessary before progress will be made in answering these questions. Finally, the relative lack of strong motion data has hindered the assessment of strong ground shaking and the factors that affect it in the Pacific Northwest. Such factors include seismic source, attenuation, basin, and geologic site effects. Hopefully, the ongoing efforts to increase the number of strong motion instruments in the region will help remedy this situation.

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# **Young faulting and the characterization of earthquake sources in Oregon**

by Silvio K. Pezzopane,  
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## **NOTE**

This paper will be published separately in *Oregon Geology*.

## **A method for predicting slope instability for earthquake hazard maps: Preliminary report**

by  
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and

*Yumei Wang, Oregon Department of Geology and Mineral Industries*

### **ABSTRACT**

The Oregon Department of Geology and Mineral Industries (DOGAMI) and the U.S. Geological Survey (USGS) are conducting a pilot study, mapping earthquake-induced slope instability (i.e., landslide potential) in the greater Eugene-Springfield area in Lane County, Oregon. In this area, development in potentially unstable hillslope areas is dramatically increasing, and the 1:24,000-scale map resulting from this study is intended for use by local governments for regional planning, design, and mitigation purposes. The method for producing the map is derived from state-of-practice dynamic slope-stability and liquefaction analyses, empirical correlations of slope instability with engineering properties of materials, and manipulation of data on local topography, engineering geology, and hydrology using geographic information system (GIS) tools.

Specific types of data used to produce the map include: (1) distribution of geologic units, determined from published geologic maps and from field mapping carried out as part of the present study; (2) engineering properties of materials in each geologic unit, determined from field mapping and from laboratory testing of selected materials; (3) slope inclinations, determined using digital elevation models (DEMs) with a 30-ft grid spacing; (4) regional hydrology, determined from well data, mapping of springs and seeps, and hydrologic modeling; (5) distribution of existing landslide deposits, as shown on unpublished DOGAMI maps; (6) distribution of artificial slope alterations, such as large road- and railroad cuts; and (7) ground motions from design scenario earthquakes, including a local floating earthquake and a far-field subduction-zone event.

Using these data, slopes steeper than 25° are analyzed using empirical criteria that relate slope instability to degree of weathering, strength of cementation, spacing and openness of rock fractures, and hydrologic conditions. To perform this analysis, statistically averaged values for each of these properties as measured in outcrop, were assigned to each geologic unit within the study area. Slopes between 5° and 25°, which in the study area are commonly mantled with aprons of heterogeneous colluvium, are evaluated with a dynamic slope-stability analysis that uses slope inclinations, engineering geologic characteristics of geologic units, and shaking parameters from design earthquakes as inputs. Existing landslides, most of which have formed on slopes within this range, are treated as a separate category. Slopes gentler than 5° are analyzed for liquefaction and resultant lateral spreading using state-of-practice engineering analyses. Results of these analyses are then combined to produce a final map with five slope instability categories (Very High, High, Medium, Low, and Nil potential for slope failure).

Techniques developed in this pilot study are intended to be applicable to relatively rapid, regional-scale mapping of seismic slope instability in other areas with a wide variety of topographic, geologic, and hydrologic characteristics.

## INTRODUCTION AND PURPOSE

Many types of earthquake hazards can be evaluated and mitigated to an acceptable level of risk before future damaging earthquakes strike. Slope instability can be a great threat, especially in urban areas with concentrated development on unstable slopes. Many recent earthquakes have caused significant loss of life and property damage from earthquake-induced landslides.

This paper focuses on dynamic (i.e., earthquake-induced) slope instability for a wide range of landslide failure types for slopes that range from steep to gentle. Steep slopes are most susceptible to rockfalls and other fast moving landslides, moderate slopes are susceptible to deep-seated rotational and translational block slides, and gentle slopes are susceptible to liquefaction-induced lateral spreading. The described method for producing a hazard map is derived from dynamic slope-stability and liquefaction analyses, empirical correlations of slope instability with engineering properties of materials, and manipulation of data on local topography, engineering geology, and hydrology using geographic information system (GIS) tools.

The final map will be produced at a scale of 1:24,000, using 30-ft grid digital elevation model (DEM) data, and will provide information for regional planning, design, and mitigation. It should serve as a useful tool to reduce hazards through effective land-use and emergency planning, regional vulnerability studies, identifying areas that would benefit from site-specific studies, and providing a cost-effective means to prioritize mitigation efforts.

The pilot study presented here is part of a larger project to map earthquake hazards and reduce risks in Eugene and Springfield. The project area includes three 7½-minute U.S. Geological Survey quadrangles (Eugene West, Eugene East, and Springfield) and the remaining areas within the urban growth boundary that lie outside the three quadrangles. The project area totals about 200 mi<sup>2</sup>. This project involves working closely with an advisory task force composed of local community members and determining ways to prudently implement mitigation. It also includes establishing a temporary local seismograph network to monitor local and distant earthquakes to gain a better understanding of local sources and ground response, and performing an evaluation of structural seismic vulnerability on a limited number of selected buildings.

## BACKGROUND

From its beginnings in the 1840s, the population of the Eugene-Springfield metropolitan area is now approaching 200,000 and continues to increase. The population within the metro plan boundary (Figure 1), which covers an area slightly larger than both the city and urban growth boundaries, is projected to increase by approximately 57 percent between 1990 and 2020 (Meacham, 1990).

As expansion spreads into the hillslope areas, which tend to be more difficult to develop, slope instability threatens both existing and new developments. Many areas are prone to landsliding, due to the nature of the geology, topography, and climate.

Numerous nonseismic landslides have occurred in the hills of the study area. For example, many landslides were activated during the heavy winter rainfall of 1996-97, including one that affected a ¼-mi section of a major arterial road, remobilizing for the fifth time in the last 20 years. This landslide was recently repaired with a rock buttress, at a cost of approximately half a million dollars. Another example from the spring of 1997 is the failure of a retaining wall at the construction site of a new single-family dwelling. Construction of this hillside home has been postponed, and the site is under investigation.

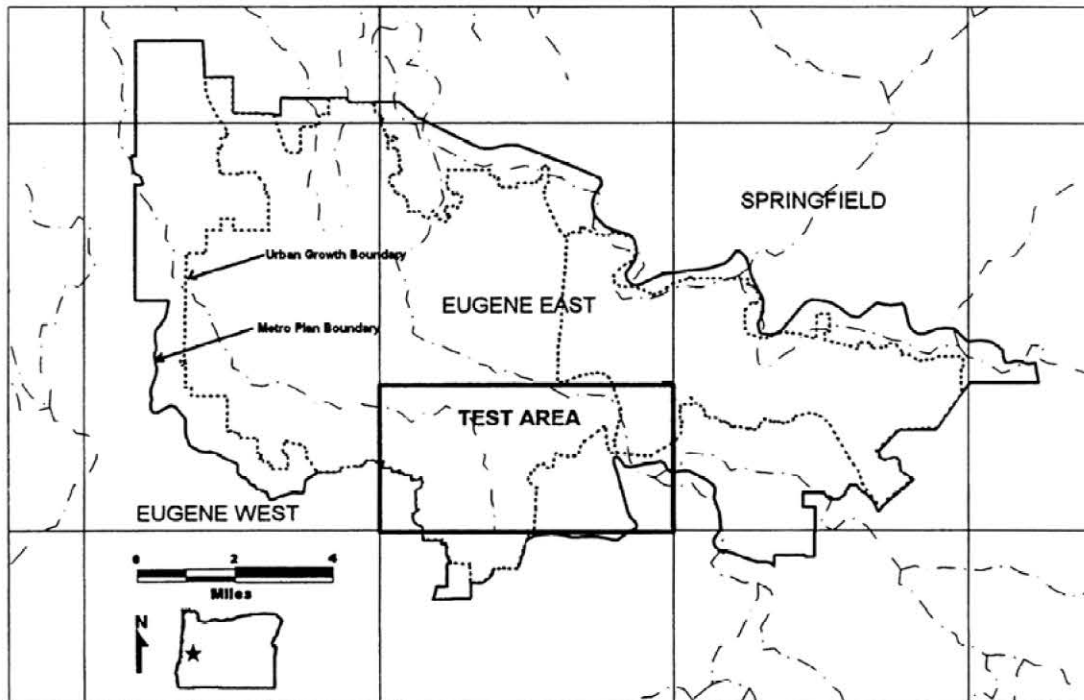


Figure 1. Location map showing test area for pilot project and its location with respect to urban growth and Metro Plan boundaries of the Eugene-Springfield area as well as main topographic quadrangles. Dash-dot lines mark major drainages.

### GEOGRAPHIC SETTING

The study area is located in the southern reach of the upper Willamette basin near the confluence of the Coast and Middle Fork Willamette Rivers and the McKenzie River. It includes hills bounding the valley, with the Cascades on the east flank and the Coast Range on the west and south. The climate is moderate in temperature and has an average annual precipitation of 40 in. Generally, the elevation of central Eugene and Springfield is about 400 ft.

### GEOLOGIC SETTING

The Willamette Valley geomorphic province is a broad lowland separating the Oregon Coast Range from the Cascade Range. This terrain is part of the forearc basin associated with the Cascadia subduction zone. The smooth alluvial plain of the Willamette Valley is interrupted by occasional flood and stream channels. The valley floor is underlain by Tertiary rocks ranging from Eocene to Miocene in age, including volcanic flows and intrusions and tuffaceous sediments and sedimentary rocks. In most places, bedrock units are overlain by Quaternary-age alluvium and thus are not well understood in detail.

The geologic units in the test area are described in Table 1 and shown on Figure 2 (Walker and Duncan, 1989).



Table 1. *Geologic units in the Eugene-Springfield metropolitan area. Modified from Walker and Duncan (1989).*

Symbol	Age	Description
Qal	Holocene	<b>Alluvium</b> —Clay, silt, sand, and gravel in river and stream channels
Qoal	Holocene/ Pleistocene	<b>Older alluvium</b> —Poorly consolidated clay, silt, sand, and gravel marginal to active stream channels and filling lowland plains of Willamette River Basin and tributary drainages
Qt	Holocene/Pleistocene	<b>Terrace and fan deposits</b> —Elevated deposits of alluvial silt, sand, and gravel along main drainages in the Coast Range and in the western Cascade Range
Tub	Miocene	<b>Basalt and basaltic andesite flows and flow breccias</b> —Grades laterally into palagonitic tuff and breccia and into clastic sedimentary rocks
Ti	Oligocene	<b>Mafic intrusions</b> —Sheets, sills, and dikes of massive granophyric ferrogabbro; some bodies strongly differentiated and include pegmatitic gabbro, ferrogranophyre, and granophyre
Tf	Oligocene/Eocene	<b>Fisher Formation, undivided</b> —Predominantly continental volcanoclastic rocks, including andesitic lapilli tuff, breccia, water-laid and air-fall silicic ash, and interbedded basaltic flows
Te	Oligocene/Eocene	<b>Eugene Formation</b> —Thin- to moderately thick-bedded, coarse- to fine-grained arkosic, micaceous, and, locally, palagonitic sandstone and siltstone, locally highly pumiceous, assigned to the upper Eocene to middle Oligocene, marine Eugene Formation
Tfb	Eocene	<b>Basaltic flows</b> —Flows, some of which may be invasive into the undivided Fisher Formation (unit Tf), and undivided and questionable sills that may intrude the undivided Fisher



Figure 2. *Simplified geologic map of test area, modified from Walker and Duncan (1989). Coordinates approximate. Geologic unit symbols keyed to Table 1 above.*

### SEISMIC SETTING

The physiographic setting of the Pacific Northwest results from its plate tectonic setting. From northern California to British Columbia, oceanic plates, including the Juan de Fuca plate, are being subducted beneath the North American plate along the Cascadia subduction zone. Earthquakes can occur within the subducting Juan de Fuca plate (intraplate earthquakes), in the overriding North American plate (crustal earthquakes), or along the interface between the two plates (subduction zone earthquakes) (Figure 3). All three possible earthquake types (subduction, intraplate, and crustal) can severely impact the study area, and each was considered as part of this study.

Although no damaging earthquakes have occurred during historic times, small local earthquakes have been recorded. A recent study that focused on evaluating ground response in Eugene and Springfield (R. Weldon and S. Perry-Huston, University of Oregon Geological Sciences Department, unpublished data) included the recording of several very small local earthquakes. In January 1996, a cluster of small earthquakes occurred about 25 km east of Eugene. Later, in May 1996, a small earthquake occurred about 5 km north-northwest of downtown Eugene. These earthquakes have not been identified with any specific fault structure (S. Perry-Huston, personal communication, 1997) but are considered to be a potential threat to local communities. The study area is located about 100 km east of the Cascadia deformation front, where several large-magnitude subduction zone earthquakes are thought to have occurred in the past few thousand years (Atwater, 1996).

This study evaluates the ground response, as influenced by local site conditions, using estimated ground motions from strong earthquakes. A strong local crustal earthquake or great subduction zone earthquake would likely produce significant ground shaking for all areas. Bedrock ground motions incorporated in the study were developed by Geomatrix Consultants (1995).

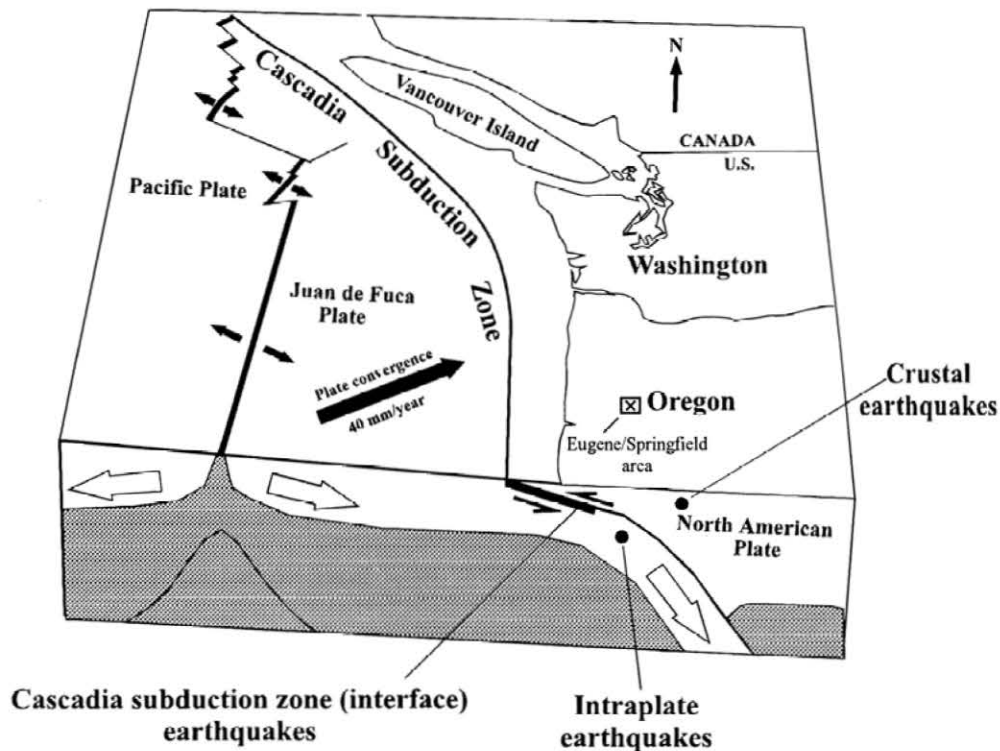


Figure 3. Pacific Northwest earthquake setting. Map and cross section showing the Cascadia subduction zone, typical locations of the three types of earthquakes, and the study area (X).

## DATA COLLECTION

The method used to evaluate earthquake-induced slope instability requires information on the geologic units (their distribution and engineering characteristics), slope angles, hydrology, and the occurrence of existing landslides and large artificial slope alterations.

The distribution of geologic units was determined from published geologic maps (Walker and Duncan, 1989; Vokes and others, 1951) and from additional mapping carried out as part of the study. The engineering properties of materials in each geologic unit were determined from field mapping, laboratory testing of selected materials, in situ tests, and engineering judgment. Field work included the mapping of over 200 outcrops considered to be reasonably representative of the geologic units. In situ tests included downhole shear wave velocity profiles, surface refraction lines, and standard penetration testing. Slope inclinations were determined using geographic information system (GIS) tools and digital elevation models (DEMs) with a grid spacing of 30 ft. Regional hydrology was determined from borehole and well data, mapping of springs and seeps, and also hydrologic modeling conducted by the local water departments. Existing landslide deposits were mapped as part of the study. Input on active landslides was provided by local consultants and public works department staff. Lastly, artificial slope alterations, such as large road- and railroad cuts were identified. The method does not specifically address slope aspect, vegetation, and human effects (such as logging and grading practices).

## ANALYSES

Slopes in the test area were divided into four groups: (A) existing landslides; (B) steep slopes, greater than  $25^\circ$ ; (C) moderate slopes, ranging from  $5^\circ$  to  $25^\circ$ ; and (D) gentle slopes, less than  $5^\circ$  (Figure 4). It was assumed that groups (B), (C), and (D) have fundamentally different modes of dynamic failure. Consequently, different analytical techniques were applied to these groups as shown on Figure 5.

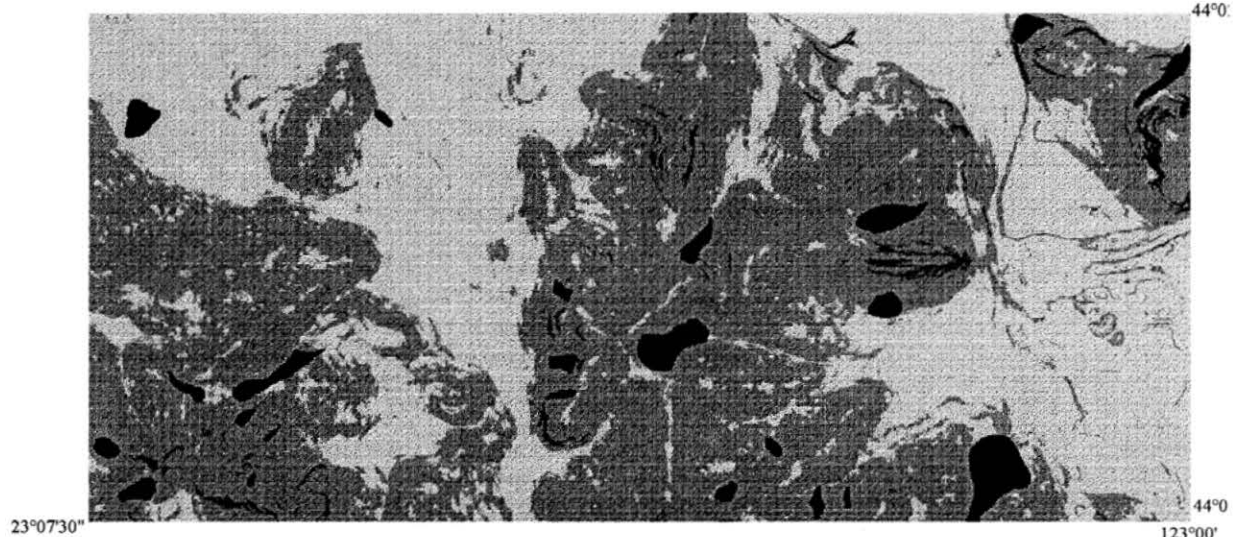


Figure 4. Distribution of slope types in test area (approximately, bottom third of Eugene East quadrangle).

- Group A, existing landslides
- ▨ Group B, steep slopes,  $>25^\circ$
- ▤ Group C, moderate slopes,  $25^\circ$  to  $5^\circ$
- ▥ Group D, gentle slopes,  $<5^\circ$

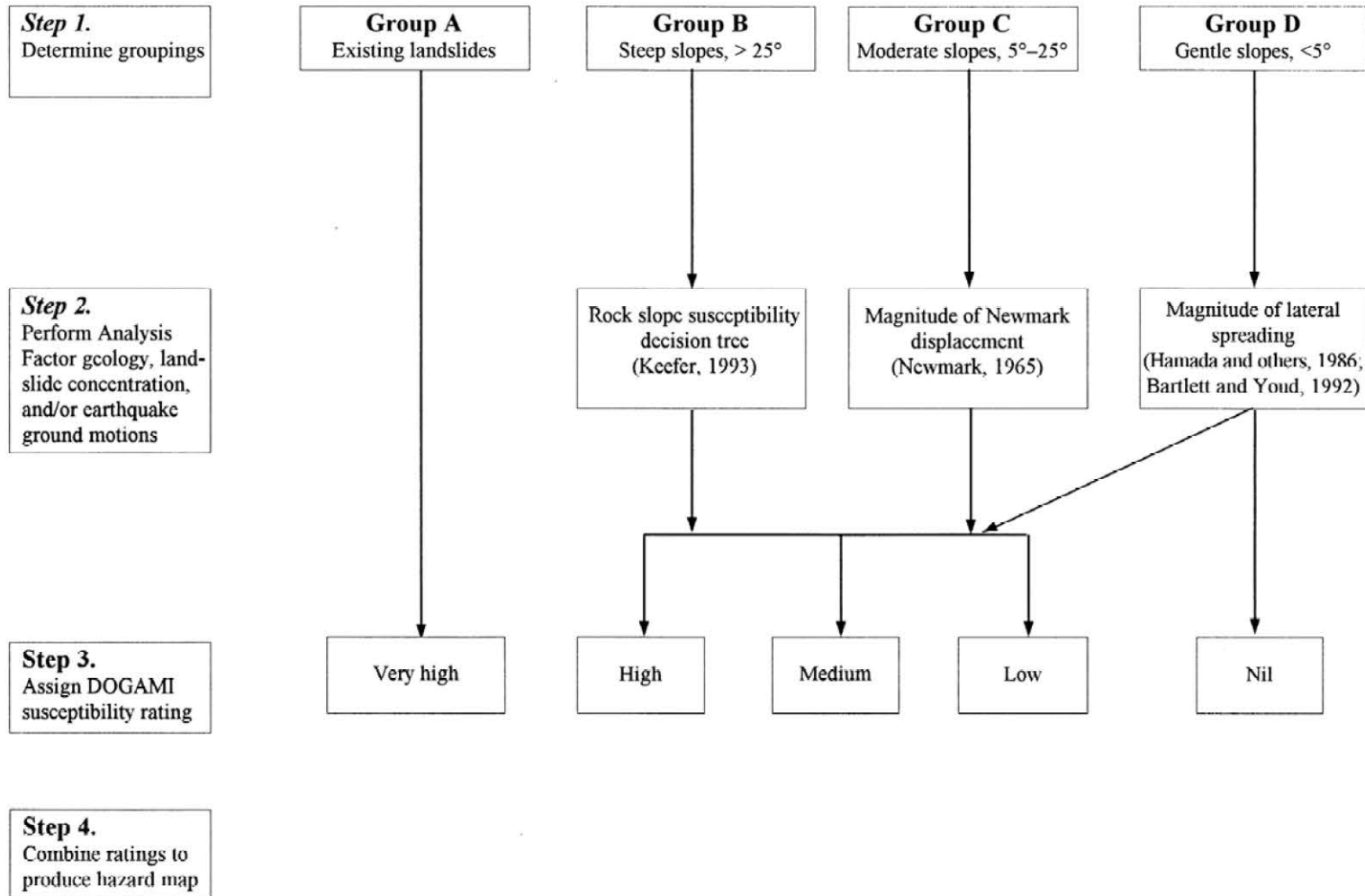


Figure 5. Method of ratings flow chart.

#### (A) Existing landslides

The movement characteristics of existing landslides are highly variable and range from actively moving to stable. To understand the nature of each existing landslide would require numerous site-specific evaluations. In the absence of this landslide information, it was assumed that the slip planes are at reduced shear strengths of unknown values, and that existing landslide masses are inherently unstable under earthquake loading. Thus, existing landslides were assigned to the very high susceptibility rating. No analytical techniques were applied.

#### (B) Steep slopes

Slopes greater than 25° were assigned to Group B, steep slopes. Engineering properties of geologic units, including degree of weathering, strength of cementation, spacing and openness of rock fractures, and hydrologic conditions, were mapped in outcrops. Each outcrop was assigned to a mapped geologic unit. Then, each geologic unit was evaluated for susceptibility to slope failure using a decision tree outlined in Keefer (1993) and shown in Figure 6.

For each geologic unit, the average value from the rating category was analyzed using empirical criteria that relate slope instability to area (Keefer, 1993). Keefer (1993) related engineering properties observable in outcrop to landslide concentration, expressed as number of landslides (LS) per square kilometer. For the geologic units within the test area, each outcrop was rated according to Figure 6, and then the results from the total number of outcrops were averaged for each geologic unit, using the following relationship:

$$LS/km^2 = (32)(\% \text{ extremely high}) + (8)(\% \text{ very high}) + (2)(\% \text{ high}) + (0.125)(\% \text{ low}).$$

Subsequently, each geology unit was assigned a new susceptibility rating compatible with the DOGAMI earthquake hazard rating system of high, medium, or low (Table 2) on the basis of the value calculated for landslides per square kilometer as shown on Table 3.

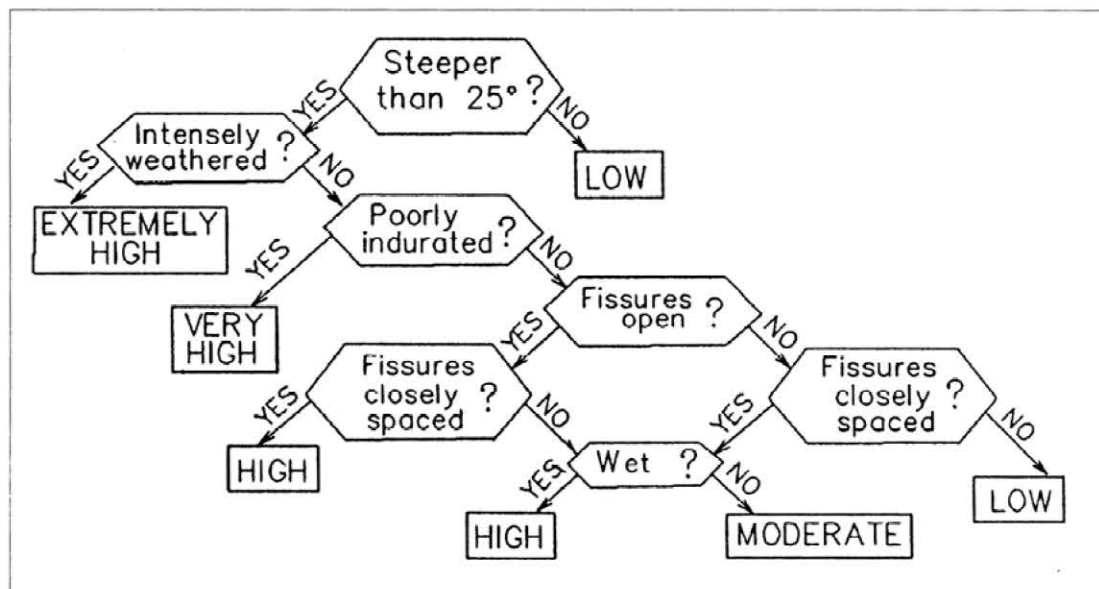


Figure 6. Decision tree for susceptibility of rock slopes to earthquake-induced landslides (from Keefer, 1993). For the test area, all slopes were assumed to be wet.

Table 2. DOGAMI rating system of landslide concentration

LS/km <sup>2</sup>	DOGAMI rating
> 2	High
1–2	Medium
< 1	Low

Table 3. Landslide concentration and ratings for geologic units

Geologic unit <sup>1</sup>	LS/km <sup>2</sup>	DOGAMI rating
Tfb	10.82	High
Te	5.34	High
Tf	2.73	High
Tub	1.83	Medium
Ti	1.30	Medium

<sup>1</sup> After Walker and Duncan (1989)

The following illustrates how the susceptibility for a specific geologic unit (Tub) was determined. A total of 34 outcrops was mapped and evaluated in accordance with Keefer's (1993) method: Thirty-one of 34 of the outcrops, or 91 percent, were assigned a susceptibility rating of High (example in Figure 7), and 3 of 34, or 9 percent, were assigned a rating of low. Landslide concentration is determined as follows ( $2 \times 0.91 + 0.125 \times 0.09 = 1.83$ ) and gives a result of 1.83 landslides per square kilometer. This value falls into the Medium susceptibility rating shown on Table 3.

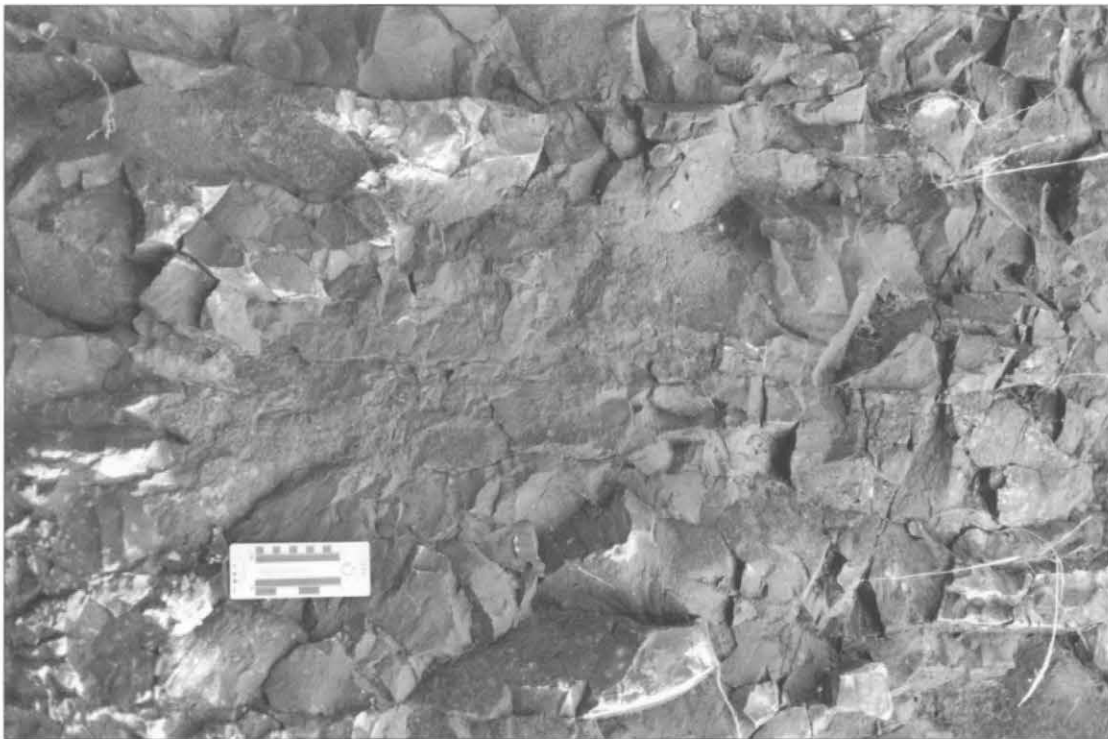


Figure 7. Sample of outcrop of geologic unit Tub, rated according to decision tree for landslide susceptibility by Keefer (1993) as shown in Figure 6. Unit Tub is not intensely weathered, not poorly indurated, fissures are not open and are closely spaced, and it is wet. This results in a rating of High.



### (C) Moderate slopes

Slopes ranging from 5 to 25° were assigned to Group C, moderate slopes. For moderate slopes, we assumed that coherent, relatively deep-seated translational and rotational slides are the most common modes of failure (Keefer, 1984). Moderate slopes in the study area are commonly mantled with aprons of heterogeneous colluvium. Our method for rating these slopes is based on the dynamic slope stability analysis of Newmark (1965), as verified and extended to regional-scale use by Wilson and Keefer (1983, 1985), Wicczorek and others (1985), Jibson (1993, 1996), and Jibson and Keefer (1993).

The selected earthquake input parameters included two controlling events: A magnitude 8.5 subduction zone earthquake at a 100-km epicentral distance and a magnitude 6.5 event at 10-km distance. Arias Intensity ( $I_a$ ) values were determined based on magnitude and epicentral distance according to the equation developed by Wilson and Keefer (1985),

$$\log I_a = M - 2 \log R - 4.1,$$

where  $I_a$  is in meters per second,  $M$  is moment magnitude, and  $R$  is earthquake source distance in kilometers. Next, assuming an infinite slope failure (where  $\alpha$  equals the slope inclination) and a selected factor of safety ( $FS$ ), an equation by Newmark (1965),

$$a_c = (FS - 1) g \sin \alpha,$$

is used to calculate the critical acceleration ( $a_c$ ). Here,  $a_c$  is the critical acceleration in terms of  $g$ , the acceleration due to Earth's gravity;  $FS$  is the static factor of safety, and  $\alpha$  is the angle from the center of the landslide mass to horizontal. The Newmark displacement ( $D_N$ ) was then determined from the relationship (Jibson, 1993; Jibson and Keefer, 1993)

$$\log D_N = 1.460 \log I_a - 6.642 a_c + 1.546.$$

Finally, each slope was assigned a DOGAMI susceptibility rating of high, medium, or low, based on the calculated  $D_N$ .

### (D) Gentle slopes

Slopes less than 5° were assigned to Group D, gentle slopes. For gentle slopes, we calculated lateral spreading (i.e., slope instability) susceptibility for Quaternary-age geologic units that are prone to liquefaction failure.

Geologic units that were pre-Quaternary in age were assumed to be stable and were automatically given a susceptibility rating of nil. Areas of Quaternary-age units were separated in order of depositional age, with artificial fill and the youngest deposits generally being the most vulnerable to slope movement. The selected earthquake input parameters, taken from an Oregon Department of Transportation study (Geomatrix Consultants, Inc., 1995), include a time history with a 2,500-yr return interval. The controlling event is an approximate magnitude 8.5 subduction zone earthquake at about a 100-km epicentral distance.

To evaluate for lateral spreading susceptibility, we first estimated the site effects of local geology on ground shaking, using SHAKE91, which is a commercially available program for analyzing one-dimensional site-response of vertically propagating (normally incident) shear waves at a level site (Idriss and Sun, 1992). Peak rock accelerations on the synthetic acceleration time history were scaled to 0.34  $g$  and used as input parameters in SHAKE91. The peak surface accelerations determined from SHAKE91 analysis were used as input accelerations in the liquefaction analyses.

Next, liquefaction was analyzed by two methods: The first by Robertson and Fear (1996), which is an improvement of a method developed by Seed and others (1984) and is based on standard penetration test (SPT) measurements, and the second by Andrus and Stokoe (1996), which is based on shear wave velocity measurements. Both methods were used to maximize the available in situ data in the study area and account for the uncertainties associated with evaluating the predominantly gravelly soils in the study area.

For soils that are prone to liquefaction, lateral spreading was estimated in two approaches: that of Hamada and others (1986) and that of Barlett and Youd (1992). Both methods are appropriate for sandy soils, whereas the study

area has predominantly gravelly soils. These methods, however, appeared to be the best available techniques. Because the maps indicate susceptibility in a relative sense, as compared to an absolute sense, this extrapolation was considered to be acceptable.

According to Hamada and others (1986), lateral spreading satisfies the following equation:

$$D_H = 0.75 T^{0.5} \theta^{0.33},$$

where  $D_H$  is lateral spreading, in meters;  $\theta$  is ground slope, in percent; and  $T$  is thickness of the liquefied layer, in meters. According to Bartlett and Youd (1992), lateral spreading due to liquefaction satisfies:

$$\log D_H = -15.787 + 1.178 M - 0.927 \log R - 0.013 R + 0.429 \log S \\ + 0.348 \log T_{15} + 4.527 \log (100 - F_{15}) - 0.922 D50_{15},$$

where  $D_H$  is lateral spreading, in meters;  $M$  is moment magnitude;  $R$  is horizontal distance to the nearest seismic energy source, in kilometers;  $S$  is ground slope, in percent;  $T_{15}$  is the cumulative thickness, in meters, of saturated cohesionless soils with  $(N_1)_{60}$  value  $\leq 15$ ;  $F_{15}$  is the average fines content, in percent;  $D50_{15}$  is mean grain size.

To illustrate this method, we used drill hole data from Test Site ES-2 (Figure 8), which is located in the Eugene West quadrangle (Figure 1). SHAKE 91 was run, and a peak surface acceleration of 0.56  $g$  was achieved. Liquefaction was analyzed using methods of Robertson and Fear (1996) and Andrus and Stokoe (1996). The results are compared in Figure 9. Next, lateral spreading was calculated by two methods: We first used Hamada and others (1986), where  $T = 9.0$  m. Then we used Bartlett and Youd (1992) and assumed  $M = 8.5$ ,  $R = 100$  km,  $D50_{15} = 1.0$  mm,  $F_{15} = 5$  percent, and  $T_{15} = 5$  m. The results are shown in Table 4. For a given liquefiable deposit, steeper slopes have a tendency toward greater lateral spreading displacements than for gentler slopes.

A susceptibility rating of high, medium, low or nil was assigned according to possible lateral spreading displacements in a relative sense.

### Susceptibility ratings

Applying susceptibility ratings within each of the four groups (A, existing landslides; B, steep slopes; C, moderate slopes; and D, gentle slopes) requires professional judgment. The last step involves combining the independent analytical results from each group to produce a coherent, uniform, relative hazard susceptibility map. Results from each group fall within one of five susceptibility ratings for dynamic slope instability: very high, high, medium, low, and nil (Figure 5).

## DISCUSSION

The method described in this paper is still under development and results are preliminary. The method is being applied to the pilot study area and will then undergo additional review. Additional research is needed in many areas of predicting dynamic slope instability for large regions. In this study, several difficulties were encountered. (1) For steeper slopes, the slope angles calculated from the DEMs with GIS tools were lower, in many areas, than those measured in the field. In addition, in places, they were lower than slope angles calculated from contours shown on USGS topographic maps. To improve modeling slopes on the basis of DEMs, calibrating and applying a correction factor may be necessary. A correction factor could be determined by calibrating slope angles calculated from DEMs to survey measurements conducted in the field. In this study, no correction factor was applied. (2) For this study, all existing landslides were, conservatively, assumed to be problematic and consequently were rated in the very high susceptibility category. It is important, however, to distinguish active landslides from inactive and stable ones to estimate hazards better. Research on developing better techniques to assess the hazards without extensive field mapping and analyses is needed. (3) Determining the hazards for moderate slopes is complex, due to the many variables involved with landslide susceptibility. To calibrate the reliability of our method, more post-earthquake field calibrations should be performed.

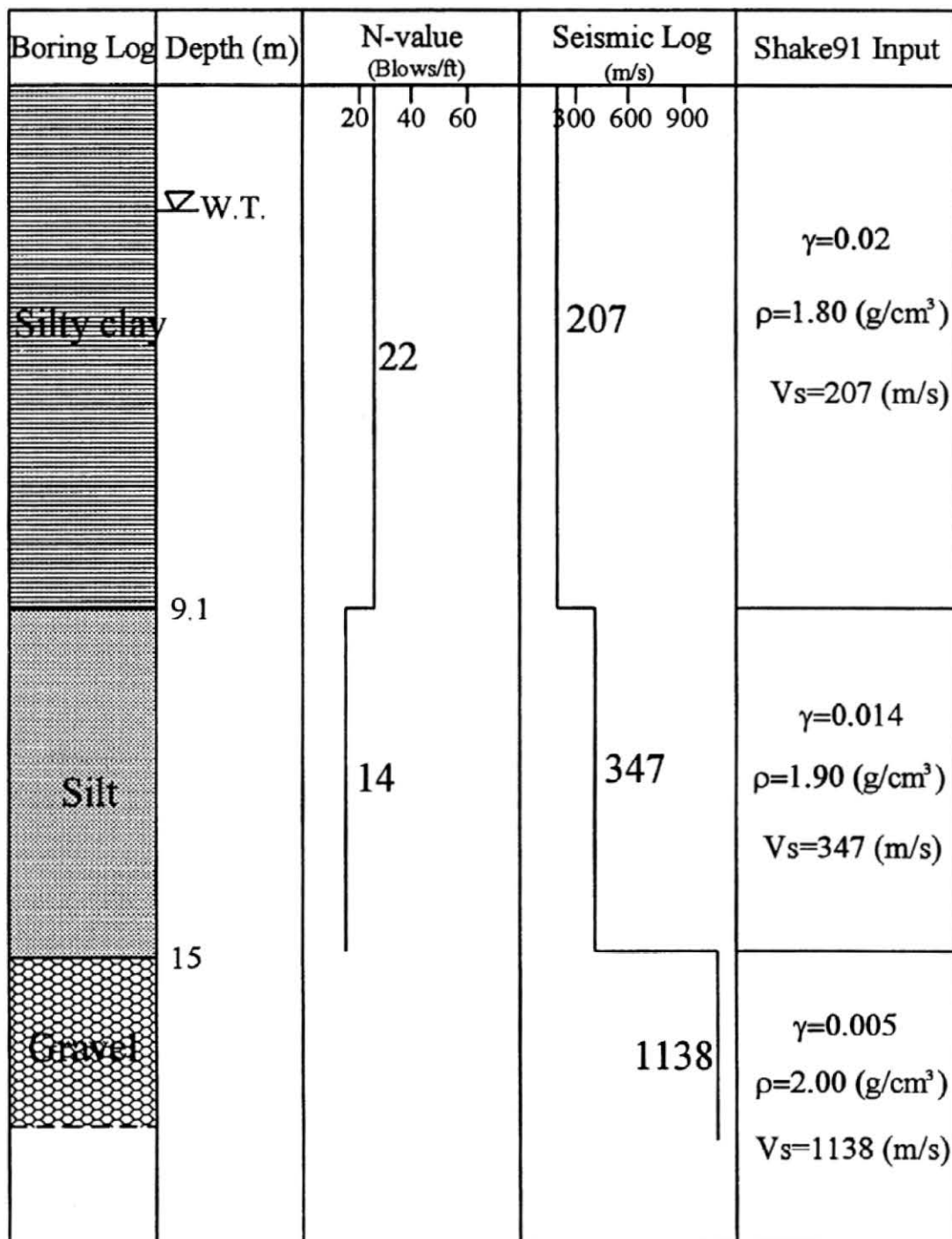


Figure 8. In situ test results and soil dynamic parameters at site ES-2. .T. = Water table;  $\gamma$  = damping ratio.

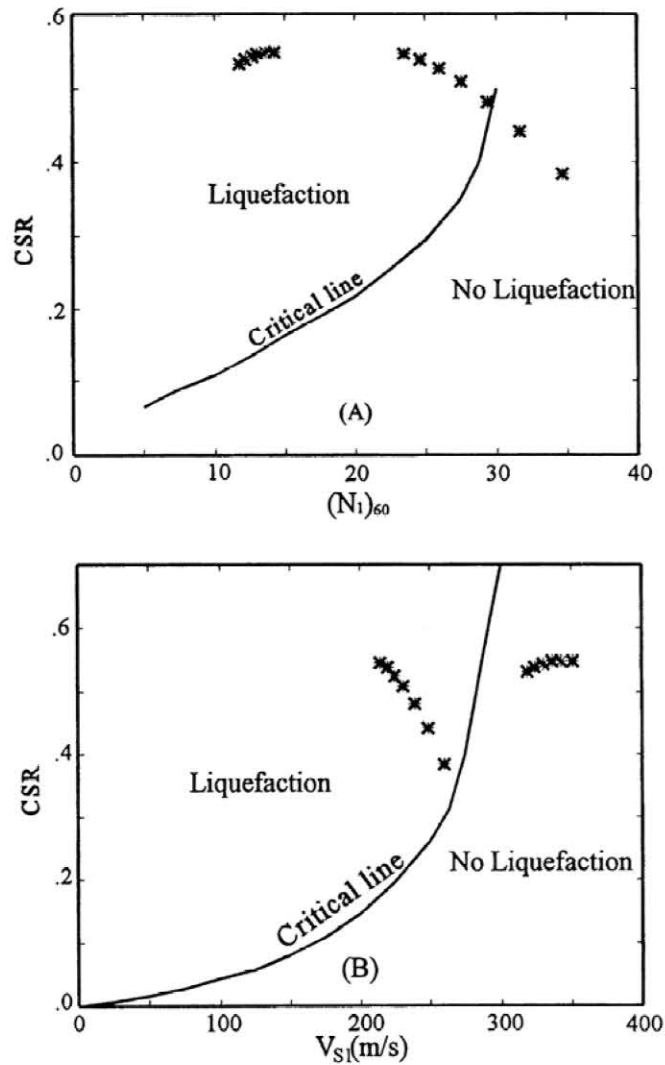


Figure 9. Comparison of results from liquefaction analyses. (A) shows results based on Robertson and Fear (1996); (B) shows results from Andrus and Stokoe (1996). CSR = cyclic stress ratio;  $(N_1)_{60}$  = corrected blow counts in blows per foot;  $V_{Si}$  is corrected shear wave velocity.

Table 4. Lateral spreading displacements ( $D_H$ ) for test site ES-2

Slope (degrees)	$D_H$ (m)	
	after Hamada and others (1986)	after Bartlett and Youd (1992)
1	2.68	0.28
2	3.40	0.38
3	3.88	0.45
4	4.28	0.51
5	4.59	0.56

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# **LIQUEFACTION EVIDENCE FOR THE STRENGTH OF GROUND MOTIONS FROM A CASCADIA SUBDUCTION EARTHQUAKE ABOUT 300 YEARS AGO**

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## **ABSTRACT**

Paleoseismic studies conducted in the coastal regions of the Pacific Northwest in the past decade have revealed evidence of crustal downdropping and subsequent tsunami inundation, attributable to a large earthquake along the Cascadia subduction zone which occurred approximately 300 years ago, and most likely in 1700 AD. In order to characterize the severity of ground motions from this earthquake, we report on results of a field search for seismically induced liquefaction features. The search was made chiefly along the coastal portions of several river valleys in Washington, rivers along the central Oregon coast, as well as on islands in the Columbia River of Oregon and Washington. In this paper we focus only on the results of the Columbia River investigation. Numerous liquefaction features were found in some regions, but not in others. The regional distribution of liquefaction features is evaluated as a function geologic and geotechnical factors at each site in order to estimate the intensity of ground shaking.

Searched islands in the Columbia River are located as close as 35 km from the coast. Liquefaction effects from the earthquake of about 300 years ago were found to be plentiful yet sporadic within about 90 km of the coast. The effects of liquefaction were found to be relatively minor in comparison to effects in the meizoseismal zones of great earthquakes elsewhere that have occurred in historical time. A preliminary geotechnical analysis similarly indicates that the peak surface accelerations required to produce the observed liquefaction features were quite low in light of current estimates for the magnitude of this earthquake which range from M 8 to M 9.

The regional inventory of liquefaction features and preliminary geotechnical analysis is interpreted as providing evidence for only moderate levels of ground shaking in coastal regions of Oregon and Washington. These ground motion intensities are much lower than the estimates derived in various investigations using theoretically- and statistically-based models to predict onshore ground motions for M 8 and 9 earthquakes. Given the influence of estimated ground motion from Cascadia subduction zone earthquakes on seismic hazard studies within 100 km to 150 km of the coastline, we strongly advocate that more paleoliquefaction and geotechnical field studies are needed to resolve this large difference.

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## INTRODUCTION

Characterizing the seismic hazard along the Cascadia Subduction Zone in the Pacific Northwest region of the United States (Fig. 1) has become increasingly important over the past decade as compelling evidence has been assembled for late-Holocene seismicity. Active convergence between the Juan de Fuca and North American lithospheric plates is resulting in a zone of subduction, in which multiple earthquake sources have the potential for causing significant ground shaking. The source zones are illustrated in Figure 2 and include earthquakes occurring along the subduction zone (interface and intraslab events) and shallower focus, crustal sources. The absence of significant historic seismicity along most of the Cascadia Subduction Zone places reliance on geological evidence to interpret the earthquake potential.

Recent paleoseismic studies in coastal Oregon and Washington have focused on seismicity caused by movement along the interface. Buried tidal-marsh soils in these regions provide strong evidence for prehistoric, late-Holocene episodes of sudden submergence accompanied by tsunamis. The submerged marshes reflect flexure of the crust by tectonic thrusting along the Cascadia Subduction Zone (Atwater, 1987, 1992; Darienzo and Peterson, 1990). Atwater (1992) has interpreted the subsidence stratigraphy to indicate that at least two great earthquakes have struck the coast of Washington during the past 2,000 years. The presence of widespread, abruptly buried soils provides evidence for an earthquake which occurred approximately 300 years ago, and a less widespread buried soil indicates that another event occurred between 1,400 and 1,900 years ago. This work is supported by the discovery of liquefaction features at sites located between Vancouver, B.C. (Clague et. al., 1997) and the lower Columbia River region (Obermeier, 1995) which appear, within the accuracy of the radioisotope dating techniques, to be correlated in time. Great (M 8 to 9) subduction earthquakes have been inferred partly on the basis of the exceptional length along the coast that appears to have been downdropped simultaneously. Radiocarbon dating allows that possibly more than 700 km of coastline were tectonically warped simultaneously in the most recent event, approximately 300 years ago (Nelson and Atwater, 1993). Most recently, documentation of a moderate-sized tsunami (2 m run-up height) in Japan, on January 26, 1700 AD, has been advocated as corroborating that the subduction earthquake of ca. 300 years ago was of M~9 (Satake et. al., 1996). For discussion purposes in this report, we initially assume that both the age and approximate magnitude interpretations based on these investigations are correct.

The recent paleoseismic studies have provided valuable data for the periodicity of large earthquakes along the Cascadia subduction zone, yet these studies have not addressed the ground motions associated with these events. Considering altogether the location of the subsided zone (Atwater, 1992), heat flow data (Hyndman and Wang, 1993; Hyndman, 1996), and strain data (Savage and Lisowski, 1991; Draggert et al., 1994), the easternmost extent of the potential rupture zone is most likely a small distance offshore (0 to 20 km) along the central portion of the subduction zone, in southwestern Washington and northwestern Oregon (Geomatrix, 1995). A sectional view showing the implicated extent of coseismic rupture is depicted in Figure 2. These models have not been verified, however, due to the absence of a historical subduction earthquake in the region. The uncertainty associated with the location of the seismogenic portion of the subduction zone (potentially tens of kilometers) has a significant impact on the strength of shaking postulated at coastal sites.

The uncertainty of the strength of onshore shaking and the implications for seismic hazard evaluations in the region prompted us to initiate a search for liquefaction evidence of strong shaking. The premise of the search was that a great subduction earthquake in proximity to the coast should have produced a multitude of large-sized liquefaction features, even in sediments of only moderate liquefaction susceptibility. The search focused on locating liquefaction features in the banks of rivers throughout the field area of southwestern Washington and northwestern Oregon. The primary objectives were as follows: (1) to search for and catalog

earthquake-induced liquefaction features along the coastal portions of the Chehalis, Columbia, Humptulips, and other rivers, (2) to verify that these features were formed during the Cascadia subduction zone earthquake of 1700, (3) to estimate the severity of shaking ground motions in the region by comparison with ground failure caused by liquefaction during historic,  $M > 8$  earthquakes, and (4) to make preliminary estimates of the ground surface accelerations using geotechnical methods. The results of the investigation along the Columbia River are presented herein.

Several hundred seismically-induced liquefaction features (clastic dikes) have been discovered in both the Columbia River islands and in banks of the smaller rivers. Dikes interpreted to be associated with the subduction event of 1700 were discovered in 1992 in islands of the lowermost Columbia River (Obermeier, 1993). In 1993, a field study by many researchers provided further geologic verification that the dikes on the Columbia River islands were induced by the earthquake of 1700 (Atwater, 1994). During the 1993 study, tube sampling of sediment was conducted in conjunction with the geotechnical investigations at scattered sites exhibiting evidence of liquefaction. However, most of the test sites were suitable only for verification of a seismic origin to the clastic dikes (i.e., limited geotechnical data was obtained). Still, a preliminary assessment of severity of shaking was attempted by Obermeier (1995). The method involved selecting a thickness of liquefied sediment that was thought to be much more than adequate to fracture the cap of cohesive non-liquefiable sediment and yield venting at the paleosurface, in accordance with criteria developed by Ishihara (1985) and later validated by Youd and Garris (1995) for earthquakes of  $M 5.3$  to  $8$ . Based on the penetration resistance of the less susceptible portion of that thickness, the peak ground surface acceleration can then be estimated. This technique has been modified by several investigators (e.g., Martin and Clough, 1994; Pond, 1996) and verified using historical earthquakes. The Ishihara method relates peak acceleration required to cause pervasive ground failure at the surface by the mechanism of hydraulic fracturing (i.e., ground failures not due to lateral spreading). The method permits estimating an upper bound in strength of shaking at sites where the observed liquefaction features do not penetrate through the non-liquefiable capping soils. The method of Ishihara seems ideal to apply to many of the Columbia River islands.

The focus of this report is to present the results of the geologic and geotechnical studies at the Columbia River islands, and then to assess the likely severity of shaking induced by the earthquake of 1700.

### **The Field Search**

The field investigation along the lowermost Columbia River focused primarily on large islands between Marsh Island, located about 35 km east of the coast, and Bonneville Dam, located more than 150 km east of the coast. Islands searched in which there was at least 0.5 km of cleanly exposed outcrop in vertical section are listed below in the order of west to east: Karlson, Marsh, Brush, Horseshoe, Woody, Tenasillahe, Price, Hunting, Wallace, Deer, Sauvie, Bachelor, Government, and Reed. Sand deposits, having grain sizes in the range most highly susceptible to liquefaction, were observed at low tide to underlie all these islands except Sauvie and Bachelor Islands. At Government and Reed Islands, the susceptible deposits that were observed lie above the level of low tide. Ages of susceptible sediments examined on all islands exceed that of the downdropping event of ca. 300 years ago. No islands with suitable outcrop exceeding a few hundred meters in length were found nearer the coast than Marsh Island.

Sand-filled dikes of seismic liquefaction origin were discovered from Marsh Island upstream as far east as Deer Island, located 90 km from the coast. Further inland, the banks along the Sandy River, near Portland, were also searched at the confluence with the Columbia River. At this site about 125 km from the coast, small sills and small dikes, some doubtlessly from seismic liquefaction, were discovered in the banks of Sandy River.

#### Sites downstream from Deer Island

The islands in this portion of the Columbia River (Fig. 3) originated mainly as immense sand bars, locally cut by channels that were later filled with fine-grained sediment. Thick deposits of sand underlie large portions of most of these islands. The islands are flat, poorly drained, and swampy. Large portions of the islands are submerged during the highest tides. Strong currents and wave action are severely eroding the banks of many islands, resulting in the exposure of clean vertical banks as much as 2 meters above water level. Significant areas (tens of thousands of square meters) are also being scoured clean in plan view.

The banks of the islands between Marsh Island (35 km inland from the coast) and Wallace Island (65 km inland) expose mainly structureless clay-rich silt deposits intercalated with a few thin (1-2 cm) silt horizons (Fig. 4). The soft clay-rich deposits are generally dark blue (unoxidized) and exhibit moderate- to high-plasticity. Very weakly developed wetland soil horizons also occur. The age of the clay-rich cap just above the level of low tide is less than 800 years and more than 600 years on most islands, on the basis of radiocarbon ages of plant material.

Regional stratigraphic control on the clay-rich cap is excellent. About 1.5 m below the top of the banks there is a tan horizon of a few centimeters in thickness, which is exceptionally rich in volcanic ash. Chemical analysis shows that the ash originated as tephra of an eruption from Mount St. Helens, in A.D. 1480-1482 (Peterson, 1992). About 10 to 15 cm beneath the tan horizon is a blue-grey horizon, generally several centimeters thick, also rich in ash. Therefore, the radiocarbon ages on fossil marsh plants (600 to 800 years) and the ash data show that exposed sediments are old enough to record liquefaction associated with the 1700 downdropping event, but are not old enough for the event 1,400 to 1,900 years ago postulated by Atwater (1992). Radiocarbon data on organic material from several meters below the ash layers show that the submerged sands were deposited between 1,500 and 2,000 years at many places (Atwater, 1994).

Sand is exposed at low tide immediately beneath the clay cap on many islands. Gradation analyses reveal poorly graded, fine- to medium-grained, clean and silty sand. The sand deposits are very extensive both vertically and laterally. Sediment and ground-water conditions on many islands are nearly ideal for formation of large liquefaction-induced features. Not only should the thin cap (commonly 1 to 3 m) enhance venting (Ishihara, 1985; Obermeier, 1989), but the ground-water table has almost certainly been at or within a meter or so of the ground surface since the islands formed. The tidal range at these islands is about 2 to 2.5 m, and high tides inundate parts of the islands and doubtlessly have done so for at least several hundred years.

Hundreds of small sand dikes were observed in about 9 km of clean vertical banks between Marsh Island and Deer Island. Figure 4 illustrates stratigraphic and venting relations commonly associated with the dikes, downstream from Cathlamet, Washington. A thin sand sheet lies on a very weakly developed soil about 1 m below the present surface. The sand sheet is 1 to 4 cm in thickness and is as wide as 10 m. Connected to the sheet is a nearly vertical, narrow planar dike that widens downward markedly and can be traced to the source stratum of sand. Where pits were dug in sand just beneath the dikes, flow structures in the sand could be observed going into the base of the dike. Within the uppermost 5 to 15 cm of most dikes, the dike width normally is several millimeters or less. For the widest dikes (those dikes wider than 15 cm), however, sidewalls seem to be parallel and nearly vertical (indicating they formed as lateral spreads). Sills up to 5 cm in thickness were observed. Sills appear abundant locally along the base of the cap, but not within the cap. At many places dikes were exposed in plan view, where it could be seen that most dikes trend parallel to and increase in abundance toward the streambanks or the shoreline. (This parallelism shows that many probably formed by the mechanism of lateral spreading.)

The widest dikes observed on the islands are listed in Table 1. Dikes widths correlate approximately with severity of shaking (Bartlett and Youd, 1995), and the regional trend of dike widths has been found to be a good parameter for locating meizoseismal regions (Munson et al., 1995; Obermeier, 1996a, 1997; Pond, 1996). Although many of the widest dikes may have been removed by erosion (Atwater, 1994), much of the search area was in regions protected from severe erosion, and the trend of wider dikes toward the coast is so clear-cut as to indicate the validity of the trend.

TABLE 1 Widths of three widest dikes on islands, as function of distance from the coast.

ISLANDS (No. of largest dikes at each island)	DISTANCE FROM THE COAST (km)	WIDTH OF THREE WIDEST DIKES (cm)
Marsh (2), Brush (1)	35	30, 30, 20
Hunting (1), Tenasillahe (2)*	50	20, 15 - 20, 15
Wallace (3)	60	15, 15, 5
Deer (3)	90	8, 5, 3

\*The widest dike at Hunting Island was probably a fracture to vent, whereas elsewhere the widest dikes are probably from opening of a lateral spread.

Not listed in Table 1 are parameters such as abundance of dikes per unit length or unit area, as was compiled by Atwater (1994). Dike density is largely controlled by the mechanisms that contribute to dike formation (Obermeier, 1996a, 1997). Principal mechanisms are lateral spreading, hydraulic fracturing, and surface oscillations. Thus, interpretations of relative severity of shaking based on dike density are equivocal without distinguishing the mechanism that created the dikes. On the Columbia River islands it is presumed that most of the dikes probably originated from lateral spreading. This assertion is based on the parallelism of dikes with the banks. Dike density associated with lateral spreading depends strongly on proximity to a stream bank (e.g., see aerial photograph of Fig. 18A, Obermeier, 1996a), therefore this parameter is a poor indicator where there has been significant erosion of stream banks, as on some of the Columbia River islands.

Dike density from other than lateral spreading on the islands clearly decreases markedly going upstream from the region of Marsh - Hunting Islands (35 to 50 km from the coast). Both at Wallace Island (60 km) and further upstream at Deer Island (90 km), where very large areas in plan view are available for inspection, dike density from hydraulic fracturing and surface oscillations ranges from very sparse to virtually nonexistent even though the cap is typically very thin (1 to 2 m). Even at islands within 35 km of the coast, dike density from these mechanisms is very low to zero over very large areas, where the cap is 3 m or less in thickness.

All dikes downstream from Cathlamet are interpreted to have been caused by the coastal subsidence event of 1700, for the following reasons: (1) the radiocarbon ages of sticks on and near the surface of venting are consistent with the event of 1700; (2) trees rooted in sediments above vented sand have the same maximum ages (about 200 years); (3) the relation of the tephra layer to the surface onto which sand was vented is regionally the same; and (4) in a regional sense maximum dike widths tend to increase toward the coast, and the density of dikes from hydraulic fracturing and surface oscillations appears to increase significantly toward the coast. The 1-m thickness of silt and clay above vented sand is interpreted to have been deposited because of submergence, following the regional downdropping from the subduction zone earthquake of 1700. The

thickness is controlled by the level of high tide, with there being no deposition of sediment above high tide. This 1-m thickness lies well within the estimates of coastal tectonic submergence (0.5 to 2 m) by Atwater (1987, 1992), and Darienzo and Peterson (1990).

Upstream from Cathlamet, the regional stratigraphic control provided by the two thin ash-rich strata is missing for a portion of Wallace Island and for all of Deer Island. However, radiocarbon data indicate that the dikes cut through sediments older than the earthquake of 1700. At Wallace Island, plant debris cut by dikes have a radiocarbon age of  $420 \pm 60$  years. At Deer Island, an archeological hearth in sediments above deposits cut by dikes has a radiocarbon age of  $690 \pm 70$  years.

In contrast to further downstream, the stratigraphic control and radiocarbon dating are insufficient to narrowly bracket the ages of the dikes at Wallace Island and at Deer Island. The range of possible ages admits that the Wallace and Deer Islands dikes could have been caused by earthquakes other than the event of 1700. For example, small liquefaction-induced ground failures were observed in the nearby town of Longview during the M 7.1 Olympia earthquake in 1949 (Chleborad and Schuster, 1990). Nonetheless, the regional pattern of maximum dike widths and the general upstream attenuation of dike abundance agrees with the overall pattern for a subduction earthquake near the coast in about 1700.

#### Sites upstream from Deer Island

Many islands were searched between Deer Island and Bonneville Dam. At least 25 km of well exposed, vertical banks of silty clay were searched upstream from Deer Island to islands about 15 km upstream from Portland. Ages of the soil deposits examined in streambanks greatly exceed the earthquake of 1700, as shown by archeological evidence and by severity of weathering (formation of B-horizons) in sediments of the banks. No liquefaction effects were observed in the portion of the search upstream from Deer Island to Portland. However, it is likely on the basis of data from water-wells and engineering borings in the region that all the outcrops searched are underlain by clays that are too thick (many meters) to have permitted ground failure from liquefaction at depth during the earthquake of 1700, at least for low to moderate levels of seismic shaking.

Evidence for liquefaction upstream from Deer Island was discovered in the vicinity of the Sandy River, about 15 km east of Portland. At this location the river terraces are elevated about 5 m above the Columbia River, which permitted observation of deformation features to much greater depth than in the marshy islands downstream. The liquefaction effects are very minor. Throughout the region, numerous small sand sill-like features occur along the base of the cap. Small detachments from the cap have also sunk into underlying sand. The dikes are very small features that pinch together upward. Their widths are less than a centimeter and heights are less than 0.5 m. Similar small features have been observed in outcrop on Reed Island (Peterson and Madin, 1997), which is an island in the Columbia River a few km upstream from Sandy River. Both on Reed Island and in the banks of Sandy River there is a thin stratum of ash that corresponds chemically to a Mt. St. Helens ash, radiocarbon dated at  $410 \pm 70$  yr BP. This ash stratum has been intruded and warped by the underlying fluidized sand (Siskowic et al., 1994). Peterson and Madin (1997) also report several more islands in the vicinity of Sandy River and Reed Island that have small dikes (generally 3 to 6 cm maximum width); stratigraphic relations and radiocarbon ages permit that the dikes were induced by the subduction earthquake of 1700. Still, there remains the question of the seismic source. Local historical seismicity near Portland records six events of magnitude 5-6 (Wong and Bott, 1995), so earthquakes strong enough to produce small dikes seem to occur relatively often in the area.



## Geologic Indications of Strength of Shaking

Multiple lines of geologic evidence indicate that the severity of shaking was not especially strong on the islands in the Columbia River. This includes the islands in the lowermost part of the river, within 35 km of the coast. The geologic evidence is provided mainly by comparison of liquefaction effects with effects elsewhere from earthquakes of  $M > 8$ . Three independent ground-failure parameters are critiqued below: (1) ground shattering and warping, (2) dike height and severity of venting, and (3) lateral spreading.

A very meaningful comparison is provided by a study recently completed, which had the objective of searching for liquefaction effects of the M 9.2 Alaskan earthquake of 1964 (Walsh et al., 1995). Indeed, the Alaskan study was initiated to provide a basis for comparison with the relatively insignificant liquefaction effects in the Columbia River islands. The method of the Alaskan search was to use a boat to examine great lengths of stream banks, which was the method used in the Columbia River study. The Alaskan search was conducted in four primary areas where regional modified Mercalli Intensities (MMI) values were IX to VII. These MMI values are considered to be representative of the intensities that would have been expected at the Columbia River sites located downstream of Deer Island during the earthquake of 1700 provided that the rupture extended near the coast.

One of the study sites in the meizoseismal region (MMI VIII to IX), located near Portage, is about 30 km perpendicular distance from the closest portion of the rupture surface. Although dikes as wide as 1.9 m from lateral spreading were found at this site, dikes wider than 25 to 30 cm were relatively uncommon. Hundreds of narrower dikes were discovered. Severe ground failure, manifest as surface warping, anastomosing dikes and formation of numerous clasts by breakup of the cap, appears to have been commonplace.

For comparison, the widest dike discovered in the Columbia River islands is 30 cm, and the phenomenon of ground shattering was not observed. The thin cap was not observed to have been warped, as would be expected if surface shaking had been very strong and if much venting had occurred. Overall, liquefaction-induced ground failure effects discovered in the meizoseismal region of the 1964 earthquake are much more pervasive than on the Columbia River islands, even if it were assumed that all the large dikes on the islands have been removed by bank erosion since the Cascadia earthquake of 1700. In light of the conclusions reached by Walsh and others, it is significant to note that the Alaskan and Columbia River field areas include similar geologic environments which were subsided following the respective earthquakes. The subsequent, rapid deposition of intertidal cohesive soils over sandy soils was hypothesized by Walsh and others to enhance the preservation of the liquefaction features.

The Alaskan search was also conducted in a region of tectonic uplift where regional MMI values were only VII to VIII (roughly 85 km to 100 km from the vertical projection of the rupture surface), and in which liquefaction effects were purportedly severe during the earthquake although little field verification was done after the earthquake. No liquefaction effects were discovered by Walsh and his coworkers (1995) in one region where they examined 20 km or so of stream banks. This was interpreted to imply that all effects of liquefaction can be removed in areas of tectonic uplift, even where liquefaction has been severe. We suggest that the findings of Walsh and others to imply that in regions of such low MMI values, i.e., VII and VIII, effects of liquefaction can be very localized.

Effects of liquefaction can also be compared with another historic great earthquake. Ground failure from liquefaction throughout the meizoseismal region of the 1811-12 New Madrid earthquakes is literally an order of magnitude more severe than on any of the Columbia River islands (e.g., Obermeier, 1989; 1996a).



Regional MMI values throughout the meizoseismal region are estimated to have been X-XI (Nuttli, 1981). Dikes as wide 0.5 to 1 m are commonplace, even many hundreds of meters away from any stream banks or moderate slopes. Hydraulic fracturing has caused multitudes of large sand blows to develop over large areas, through a cap as thick as 6 to 7 m. Surface oscillations have caused ground failure at distances as much as tens of kilometers beyond the meizoseismal region. Dikes as much as 15 cm wide are ubiquitous. Great quantities of sand were vented over large areas, even to the extent of forming a continuous veneer of sand as much as a meter in thickness over hundreds of hectares.

The liquefaction susceptibility is generally only moderate through the meizoseismal region of the 1811-12 earthquakes (Obermeier, 1989). Possibly enhanced ground motions associated with a very high earthquake stress drop occurred during the 1811-12 earthquakes (Hanks and Johnston, 1992), which may have induced especially severe liquefaction effects in the meizoseismal zone. In addition, the liquefaction susceptibility is probably a little higher in the meizoseismal zone of the 1811-12 earthquakes than at most of the Columbia River islands (Obermeier, 1995). Still, any difference in susceptibility does not seem nearly large enough to explain the huge difference in severity of liquefaction effects. Thus, the geologic evidence suggests that the strength of ground shaking on the Columbia River islands was not especially strong.

The field observation that only very minor venting occurred throughout the islands in the lowermost Columbia River indicates a situation that is ideal for application of the Ishihara method, as used by Martin and Clough (1994). The observation of minor venting, even through a thin cap, indicates that the cap thickness on the islands was adequate to suppress severe ground disruption. This interpretation is supported by the lack of other manifestations of severe liquefaction such as those we noted for the Alaskan and New Madrid earthquakes (e.g., shattering of the cap or development of anastomosing dikes).

## **Geotechnical Evaluation of Strength of Shaking**

### **Method of analysis**

A preliminary estimate of peak accelerations required to produce the observed liquefaction features is made using the method of Ishihara (1985), which relates the ground surface accelerations, thickness of liquefied zone, and surface evidence for liquefaction. Ishihara's relations are shown in Figure 5. These empirical relations were established at near-horizontal sites where the mechanism of hydraulic fracturing (cohesive cap) or upward erosion of the cap (non-liquefiable cohesionless cap) caused by the liquefaction of underlying sand resulted in the ground failures observed at the ground surface. The relationships are therefore not strictly applicable for sites where lateral spreading or surface oscillations have caused ground failure. Both lateral spreading and surface oscillations commonly cause ground failure at lower accelerations than those required for hydraulic fracturing (Seed et al., 1983, 1986; Youd and Garris, 1995; Pond, 1996; Obermeier, 1997), at least for sand deposits that are loose to moderately compact. As a result of a lower threshold for lateral spreading at sloping sites, or adjacent to free-faces, field relations depicted in Figure 6 are observed in paleoseismic studies. The height  $H_1$  is the maximum attributable to hydraulic fracturing alone. Discussion of field conditions suitable for the Ishihara method are given by Obermeier and Pond (1997).

The method of Ishihara (1985) is based on the ground motions required to produce evidence of liquefaction at the ground surface. In areas where liquefaction-induced dikes have been injected into the cap of non-liquefiable soil yet do not penetrate to the surface, the method can be used to estimate upper bound ground surface accelerations. This implies that if the ground shaking had been stronger, then the dike would have completely penetrated the cap and surface venting of liquefied soil would have occurred.

The procedure of Seed and others (1983; 1986) is used to estimate the acceleration required to liquefy the sand beneath the cap. Estimates are made according to field measurements on the sands, using the Standard Penetration Test (SPT). Figure 7 shows the simplified liquefaction boundary curve for M 8 earthquakes developed by Seed and others. The procedure of Seed and others was originally intended to provide a conservative bound for the  $(N_1)_{60}$  values in sandy soils that indicate a susceptibility to liquefaction. The catalog of liquefaction case studies has been reevaluated by several investigators in order to establish the liquefaction boundary curves for a 50% probability of liquefaction (e.g., Liao et. al., 1988; Loertscher and Youd, 1994). The various curves developed for M 8 earthquakes are provided for comparison in Figure 7. The recent magnitude scaling factors proposed by Arango (1996) were used in plotting the liquefaction boundary curves by Seed and others (1986) and Liao and others (1988). This plot has been used to estimate the peak ground surface accelerations at the liquefaction sites.

It should be noted that this method of evaluating the threshold acceleration required to induce surficial evidence of liquefaction may result in conservative estimates of acceleration (i.e., higher intensity shaking than actually occurred during the causative earthquake) because of the following factors: (1) densification of the near surface sands during and after the earthquake of 1700, and (2) the earthquake of 1700 is assumed throughout the liquefaction evaluation to be M 8, despite independent geologic evidence for a greater magnitude. The sandy deposits at the study sites have undoubtedly been subjected to numerous earthquakes since the 1700 event. Observations made after the 1949 Puget Sound earthquake confirmed that liquefaction occurred at numerous sites along the banks of the Columbia River (Chleborad and Schuster, 1990). It is likely that the density of the sandy soils has increased during the past 300 years. This would result in higher penetration resistances than existed prior to the earthquake of 1700, and would increase the computed peak surface accelerations. The influence of this effect on the estimated pga values cannot be estimated in any more than a qualitative manner.

In light of the variability of the geologic environment in the region and the uncertainty posed by the obliteration of many of the liquefaction features since the earthquake of 1700, it is not considered possible to precisely ascertain the magnitude of the earthquake from the liquefaction features alone. We therefore initially assume in the following analyses that the earthquake was a M 8 subduction event. Based on the magnitude scaling factors reported by Arango (1996), the use of liquefaction boundary curves for a M 9 earthquake would yield computed pga values that are approximately 35 to 55 percent smaller than corresponding pgas determined for an M 8 earthquake.

Preliminary geotechnical investigations were performed at the following islands, located at increasing distances from the coast: Marsh Island (35 km), Brush Island (35 km), Price Island (45 km), (Hunting Island (50 km), Wallace Island (65 km), and Deer Island (90 km). A portable dynamic penetrometer was used to obtain the penetration resistance (number of blows required to drive the probe 30.5 cm) of the sand deposits beneath the cap. The dynamic penetrometer that we used is a small device, much smaller than the split-spoon sampler normally used for the SPT procedure. This difference in sizes required converting the penetrometer blow counts to a corresponding SPT value. The correlation with the SPT was established at several Columbia River sites from side-by-side comparisons.

Both hand augering and the vibrocore technique were used to obtain samples of sand for gradation analysis. The hand auger could obtain samples only in the upper meter, or so, of sand directly beneath the cap. The vibrocore technique works by vibrating a tube into the ground. In essence, the tube advances by liquefying the sand at the tip of the tube. This method typically obtained samples to a depth of 3 to 4 m into the sand, which in most places was about the same as the maximum depth of the penetration test nearby. However, the

vibracore method was used at a relatively small number of sites in comparison to the penetration tests.

#### Estimation of Peak Surface Accelerations

The source sands are fine, uniformly sized, and usually contain some silt (approximately 10 to 20 percent). The soil profiles at liquefaction sites generally have a 2 to 4 m thickness of soft, plastic silty clay that is underlain by loose to medium-dense sand. The sand extends beyond the maximum depth of investigation, 6 m. Geotechnical investigations throughout the region reveal approximately 60 m of sandy soils underlain by interbedded sands and silty soils, with bedrock located in several borings at depths between 170 m to 210 m.

Relations at Marsh Island are relatively uniform in terms of geologic and geotechnical settings, and thus the setting is ideal for evaluating the ground motions associated with the formation of liquefaction features. Field penetration resistances (listed in Atwater, 1994) have been normalized to  $(N_1)_{60}$  values. The normalized blow counts shown here are typical of those at most of the other islands in the sense that a very loose zone of sand lies directly beneath the cap, within the uppermost meter. With increasing depth, the  $(N_1)_{60}$  values increase markedly. On many other islands the  $(N_1)_{60}$  blow counts vary much more horizontally, but even on those other islands the  $(N_1)_{60}$  values generally tend to increase significantly with depth. The thin loosened zone likely corresponds to a region loosened by liquefaction (Obermeier, 1996a). A summary of the data from the field study are in reports by Atwater (1994) and by Obermeier (1995).

The values of  $(N_1)_{60}$  on Marsh Island are nearly the same over a long horizontal distance, in a region where dikes did not penetrate all the way to the top of a 1.5 m-thick cap, and in a region far removed from possible effects of lateral spreading. Application of the Ishihara method to Marsh Island, in the manner devised by Martin and Clough (1994), is shown in Figure 8. The figure applies to a M 8 earthquake. The abscissa on the figure is related to the thickness of the sand layer that liquefied, in terms of the percent of the thickness with penetration data. The solid line shows the peak ground acceleration that would have been required to cause venting at the surface, in terms of thickness of liquefied sand. The accelerations required to liquefy various thicknesses of the sand are shown as dashed and as dotted lines, depending on fines (silt and clay) content. Limited gradation analyses demonstrate that the source sediments consist of poorly graded sand with generally less than 15% nonplastic silt. The intersection of the lines yields the back-calculated estimate of acceleration. The peak ground surface acceleration is estimated to have been approximately 0.1 to 0.15 g at this site.

The results of this analysis for the other islands are presented in Table 2. The table shows that within 35 km of the coast, the maximum psa is about 0.25 g. Yet, farther from the coast, at Price and Hunting Islands, the pga values are substantially higher. The higher values could have resulted from three sources: (1) the heights of many of the dikes could have been enhanced by lateral spreading, (2) the penetration blow counts are highly variable on these islands, which makes it virtually impossible to be confident of the properties of the source sands that actually created the dikes without extensive in situ testing, and (3) some of the "dikes" were interpreted from very thin sand intrusions within samples taken using the vibracore, and thus the intrusions could have been an artifact of a sampling procedure that inherently liquefies loose sandy soils during sampling. The variability of soil conditions at several sites resulted in widely varying trends in the penetration resistances in virtually side-by-side soundings. Our calculations in Table 2 were made by assuming the dike height was controlled by the properties of the sand at the nearest penetration test; that assumption is often invalid because of the tendency of liquefied material to flow horizontally along the base of an impermeable layer before making a dike that extends upward (e.g., Obermeier, 1996b; Tuttle and Barstow, 1996). Horizontal flow is especially a problem along the base of a channel-fill deposit such as at Price Island and part of Hunting Island. Altogether, back-calculated accelerations on Price and Hunting Islands are very suspect, and may be

considerably too high.

Back-calculated results on Wallace Island, 65 km from the coast, yield values ranging from about 0.1 to 0.25 g. Stations 4 and 5 in the table are the only ones that can be reasonably associated with hydraulic fracturing alone; values of  $p_{ga}$  here range from 0.15 to 0.25 g.

At Deer Island, 90 km from the coast, dikes were very sparse in plan and sectional views throughout a very large area (Obermeier, 1995). All dikes were small, and none was observed to have cut through more than a 1.5 m thickness of cap. It is likely that nearly all were caused by lateral spreading. Only very rarely did a dike cut by hydraulic fracturing through a cap as much as one meter in thickness. Penetration data indicate it is likely the surface accelerations were even lower than on Wallace Island.

At dike sites on the islands near Portland, at about 110 km from the coast, insufficient geotechnical data have been collected to evaluate the strength of shaking (C.D. Peterson, Portland St. U., oral comm., 1997). Whatever the source of the small dikes in the region, either a local earthquake or a great subduction earthquake near the coast, the very small sizes and the spotty occurrences of liquefaction effects argue strongly for a low level of seismic shaking, on the order of the lower limit of the threshold for liquefaction.

To summarize, between 35 and 90 km from the coast, our analysis indicates that peak surface accelerations only exceptionally might have much exceeded 0.25 g. On four islands, including two nearest the coast, the peak surface accelerations could have been as low as 0.1 to 0.2 g to produce the liquefaction features observed in the field. On two islands located 45 to 50 km from the coast, the accelerations were likely higher, but cannot be estimated reliably because of the highly variable and complex geologic setting on those islands.

#### Bedrock Accelerations

Acceleration-attenuation relationships for M 8 to M 9 Cascadia earthquakes, based on regression analyses of empirical data (e.g., Crouse, 1991; Youngs, in Geomatrix, 1995) and numerical modeling techniques (e.g., Cohee et. al., 1991; Wong and Silva, 1996), are shown in Figure 9. Although these relationships provide guidance for selecting ground motion parameters, considerable uncertainty exists in their application to Cascadia subduction zone earthquakes due to the absence of empirical data. Below we use the approximate ranges of peak surface accelerations (Table 2) to back-calculate the peak bedrock accelerations that were produced by the earthquake of 1700. Our back-calculated values are then compared with the estimates of others.

We have evaluated the dynamic response of the soils along the lower Columbia River valley using three lines of evidence: (1) regional intensities of historic earthquakes, (2) ground motion amplification factors that have recently been developed for use in seismic design provisions, and (3) results from one dimensional dynamic soil response analyses. Because of the lack of measured dynamic soil properties at the island sites, our study has primarily evaluated whether the soil deposits along the lower Columbia River valley are more likely to amplify or attenuate bedrock motions.

The Modified Mercalli Intensity data for the 1949 Puget Sound earthquake demonstrate an increase of as much as three units adjacent to the Columbia River in the vicinity of Astoria (Malone and Bor, 1979). This large increase is interpreted as being due primarily to two factors: ground motion amplification due to dynamic soil response, and the presence of poor soils prone to seismically-induced ground failures.

Table 2 Estimated Peak Ground Accelerations at Columbia River Islands

ISLAND AND BORING NO.	CAP THICKNESS (m)	APPROX. PGA (g)	PROBABLE MODE OF GROUND FAILURE	CONFIDENCE IN PGA	BEST ESTIMATE OF PGA AT ISLAND
MARSH ISLAND (35 km from coast)					
1	1.4	0.10	Lateral spreading	Poor	0.15 g
2	1.3	0.15	Lateral spreading	Poor	
3	1.8	0.1-0.15	Lateral spreading	Poor	
4	1.3	0.1-0.15	Lateral spreading	Poor	
5	1.1	< 0.15	Hydrofracturing	Good	
6	1.0	< 0.15	Hydrofracturing	Good	
BRUSH ISLAND (35 km from coast)					
1	1.5	0.18	Lateral spreading	Poor	0.2 - 0.25 g
2	1.5	0.17	Lateral spreading	Poor	
3	1.5	0.23	Lateral spreading	Poor	
4	1.5	0.24	Hydrofracturing	Good	
PRICE ISLAND (45 km from coast)					
1	3.5	< 0.35 - 0.4	Unknown	Poor	Indeterminant
2	2.7	0.27 - 0.30	Unknown	Poor	
3	2.8	0.25	Unknown	Poor	
4	5.0	< 0.35 - 0.40	Unknown	Poor	
5	4.0	< 0.4	Unknown	Poor	
HUNTING ISLAND (50 km from coast)					
1	2.7	< 0.4	Lateral spreading	Fair	0.25 - 0.30 g
2	2.7	< 0.5	No evidence of liquefaction	Fair	
3	1.9	0.25 - 0.3	Hydrofracturing (?)	Fair	
4	2.0	0.3	Hydrofracturing	Fair	
5	3.0	0.4	Hydrofracturing	Fair	
6	1.9	0.3	Hydrofracturing	Fair	
7	1.7	0.25 - 0.3	Hydrofracturing	Fair	
8	2.0	0.3	Hydrofracturing	Fair	
9	2.7	0.4	Hydrofracturing	Fair	
10	2.3	0.2	Hydrofracturing	Fair	
WALLACE ISLAND (65 km from coast)					
1	1.0	0.15	Lateral spreading	Poor	0.2 - 0.25 g
2	1.0	0.15 - 0.20	Lateral spreading	Poor	
3	1.0	0.10 - 0.15	Lateral spreading	Poor	
4	1.1	0.15 - 0.20	Hydrofracturing	Good	
5	1.1	0.20 - 0.25	Hydrofracturing	Good	



The case for the enhancement of ground motions along the Columbia River is supported by the intensity-dependent amplification factors recently developed for the FEMA/NEHRP Recommended Provisions for Seismic Regulations (1995). Based on the deep deposits of predominantly sandy soils (Soil Profile D) prevalent throughout the Columbia River study area, amplification ratios ( $PHA_{\text{surface}}/PHA_{\text{rock}}$ ) of 1.4 to 1.6 are recommended.

In order to establish credible bounds on site-specific ground motion amplification ratios for the study sites, a suite of one dimensional dynamic soil response analyses was performed using the equivalent linear model SHAKE91 (Idriss and Sun, 1991). The dynamic properties of the soils were estimated from empirical correlations with standard geotechnical parameters (e.g., depth, penetration resistance, void ratio) taken at sites in the region. Parametric studies were used to evaluate the influence of the soil and bedrock properties on the soil response. The input ground motions used in the numerical modeling included several recorded motions from M 7+ western United States crustal earthquakes, as well as a synthetic acceleration time history developed for a M 8 Cascadia subduction zone earthquake (Cohee et. al., 1991). For the broad range of ground surface shaking estimated to have occurred during the 1700 earthquake (i.e., 0.10 g to 0.35 g) the computed amplification ratios varied from 1.5 to 2.5. This range is in close agreement with the values determined using the simplified FEMA procedure (1995) and the value of 2.2 used by Cohee and others (1991) based on empirical data.

The influence of 2D or 3D "basin" effects due to bedrock topography on the characteristics of the ground motions was not evaluated. This effect has been shown in several regions to increase the amplitude and duration of earthquake ground motions, and would thereby increase the likelihood of liquefaction. Given the soil depth-to-valley width ratios of roughly 1/20 to 1/30, and the location of the islands in the central portions of the Columbia River valley, the effects of bedrock topography on the intensity and duration of strong ground motions would be very minor. Omitting this factor is judged to be only slightly conservative (i.e., leading to higher estimated peak ground accelerations).

Synthesis of the three methods of evaluation leads to the conclusion that the soil deposits at the lower Columbia River sites probably amplified bedrock motions. A lower bound amplification ratio of 1.5 seems conservative to us, and should yield bedrock accelerations at least as high as actual values. Using this ratio yields peak horizontal accelerations in bedrock of approximately 0.10 g to 0.25 g, with best estimates ranging from about 0.10 g to 0.15 g, at distances between 35 and 65 km from the coast.

## DISCUSSION

Liquefaction effects of the 1700 earthquake indicate that the ground motions were of no more than moderate intensity within a few tens of kilometers of the coast, even on the Columbia River islands. Geologic indicators for liquefaction-induced ground failures, such as formation of clastic dikes, venting of sand, and deformation and warping of the fine-grained cap above the liquefied sand, are minor in comparison with effects of liquefaction documented in the meizoseismal regions of other large, historic earthquakes. Despite the possibility that erosion of the islands has removed some of the largest dikes, all the other indicators clearly show only minor ground failure due to liquefaction.

The geologic evidence for the strength of ground surface shaking has been supplemented with a quantitative, albeit approximate, geotechnical analysis to estimate the strength of shaking. The assumptions that we used in the geotechnical analysis were made as to yield high (i.e., conservative) values of strength of shaking. While the influence of several of these assumptions on the computed accelerations can be quantified



(e.g., liquefaction susceptibility due to M 8 or M 9 earthquakes, ground motion amplification ratios), the impact of other approximations (e.g., post-earthquake densification, limited geotechnical data) on these estimates is indeterminable. Still, we believe that our back-calculated peak bedrock accelerations are accurate within roughly 0.10 g.

Our estimates of peak bedrock accelerations in the Columbia River valley range from 0.10 g to 0.20 g, within 35 to 65 km of the coast (assuming that the causative earthquake was M 8). These results corroborate the findings along several other coastal rivers in Washington and Oregon (Obermeier and Dickenson, *submitted for publication in the BSSA*). It appears that no where within the region investigated were the peak ground surface accelerations during the 1700 earthquake more than 0.25-0.3 g. Numerous sites located within 35-40 km of the coast provide evidence for even lower peak ground surface accelerations. These values apply throughout the field area, from the Columbia River extending 100 km to the north. Such low accelerations are indicated by findings in two very different geologic settings. In one setting, Columbia River islands are underlain by very thick, extensive deposits of fine sand. In the other setting, terraces along smaller rivers are underlain by sandy gravel and thin layers of medium sand.

The estimated peak horizontal accelerations for rock sites, as a function of distance to the rupture, are shown in Figure 10. For the figure we have assumed that the rupture zone is located 15 km offshore at a depth of 20 km (Hyndman and Wang, 1993; Hyndman, 1996). The attenuation curves of Youngs (Geomatrix, 1995) for M 8 and M 9 Cascadia earthquakes are provided for comparison. Our bedrock accelerations agree reasonably well with that of Youngs, for a M 8 or slightly stronger earthquake located about 15 km offshore. The peak accelerations on rock are generally overpredicted by the attenuation relationship for the M 9 scenario. This generally poor agreement could be due to (a) uncertainties in the peak surface accelerations reported herein, (b) uncertainties in the current attenuation relationships for Cascadia subduction zone earthquakes, (c) a rupture zone which did not extend to within tens of kilometers of the coast, or (d) the earthquake of 1700 was approximately M 8. Again, it should be noted that if we had performed the liquefaction evaluation for a M 9 event, our computed accelerations would have been considerably smaller.

A possible explanation for the discrepancy between the estimated accelerations is provided by recent studies of the faulting mechanisms along the Cascadia subduction zone. Recent studies by Atkinson (1995) and by Wang and others (1995) indicate that stress drops of large Cascadia subduction earthquakes in the Puget Sound-Vancouver Island region are relatively low, and range from approximately 3 to 10 MPa. Earthquake stress drop affects computed moment magnitude as well as the intensity of the ground motions (Hanks and Johnston, 1992). The relatively low stress drops that appear to be representative for Cascadia subduction zone earthquakes would almost certainly result in relatively low levels of ground shaking. We emphasize that at least two independent lines of evidence (i.e., crustal heat flow modeling and recorded ground motion data) other than our liquefaction-based interpretations indicate that accelerations from Cascadia subduction earthquakes are relatively low.

An example showing the strong influence of stress drop on computed acceleration is shown by the results of numerical ground motion modeling by Wong and Silva (1996). They assumed a stress drop of about 6 to 7 MPa, and their results at two source-to-site distances are plotted in Figure 12. Reducing the stress drop used in the numerical model by a factor of two (i.e., to 3 Mpa following the work of Atkinson) would reduce the computed peak rock accelerations by about 40-50 % (Wong, personal comm., 1997).

Our study provides initial estimates for the strength of shaking generated by the earthquake of 1700. Our interpretations are based on a relatively large geologic base but a relatively small geotechnical base. There

is a need for much more geotechnical study of sediments of the Columbia River islands. The study should be done at islands such as Marsh Island, where the geotechnical properties of the sand are not highly variable in the lateral direction; the study should not be done at locales such as Hunting Island, where the properties vary so much as to make interpretations equivocal.

There is also a need to do an extensive paleoliquefaction search in central Oregon, because the relatively low ground shaking intensities computed along the Columbia River are supported by the observations by the first author in the central coast of Oregon. Several kilometers of stream banks in the vicinity of Alsea Bay and Siletz Bay were inspected for evidence of liquefaction. Despite the searched areas being underlain by saturated, loose, clean to slightly silty sands (Wang and Priest, 1995), no liquefaction features were discovered in sediments that are probably at least slightly older than the earthquake of 1700. This lack of liquefaction features within 5 to 10 km of the coast appears inconsistent with proximity to the meizoseismal region of a M 8 or 9 subduction earthquake. Additional work in this region is warranted before conclusions are reached regarding the significance of these preliminary observations. Nonetheless, the absence of liquefaction features in highly susceptible sediments that pre-date the earthquake of 1700 provides qualitative data for the assertion that ground surface motions from that earthquake were low in the vicinity of the central Oregon coast.

### CONCLUSIONS

1. The liquefaction features discovered along the Columbia River provide conclusive evidence for seismic shaking from the subduction earthquake of 1700 AD.
2. Ground failure from liquefaction induced by the 1700 Cascadia earthquake was relatively moderate throughout the field area. The size and abundance of the liquefaction features is considerably less substantial than is found in the meizoseismal regions ( $\text{MMI} > \text{VIII}$ ) of large historic earthquakes elsewhere.
3. In the study area, from the Columbia River northward about 100 km, very strong onshore ground surface shaking ( $\geq 0.25\text{ g}$ ) from a plate-boundary subduction earthquake probably did not extend much onshore the past few thousand years. In some of the field region it does not appear that these levels of ground shaking have been exceeded throughout Holocene time.
4. The liquefaction data seem consistent with multiple scenarios for the earthquake of 1700: (a) a M~8 earthquake with the rupture zone extending as far east as the coast, (b) a M 8.5 to 9 earthquake with the rupture zone located several tens of kilometers offshore, or (c) a large earthquake with source parameters, such as stress drop, that resulted in relatively low intensity ground motions.
5. A seismic source zone capable of liquefying sandy gravel is likely located near the Chehalis River valley, about 60-80 km inland from the coast.
6. There is a great need to search for evidence of liquefaction from the earthquake of 1700 AD throughout the coast of Oregon, in order to verify the levels of shaking.

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The Oregon Department of Transportation provided the field penetrometer and the Port of Portland permitted access to several sites for the calibration study. Tim Roberts (Strata Consultants) performed most of the ODOT penetrometer testing on the Columbia River islands and Mark Barkau (Woodward Clyde Consultants) assisted with the calibration study. This support is gratefully acknowledged.

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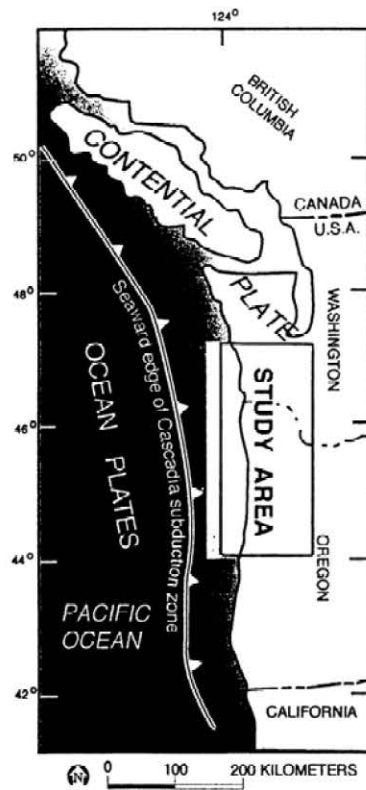


Figure 1: Cascadia subduction zone and the study area.

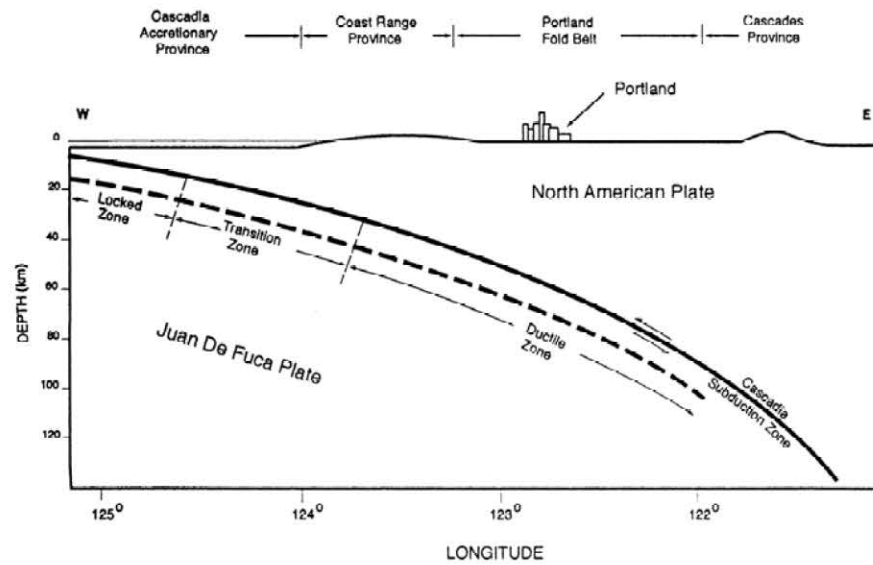


Figure 2: Schematic cross section of the Cascadia subduction zone (after Wong et. al., 1993).

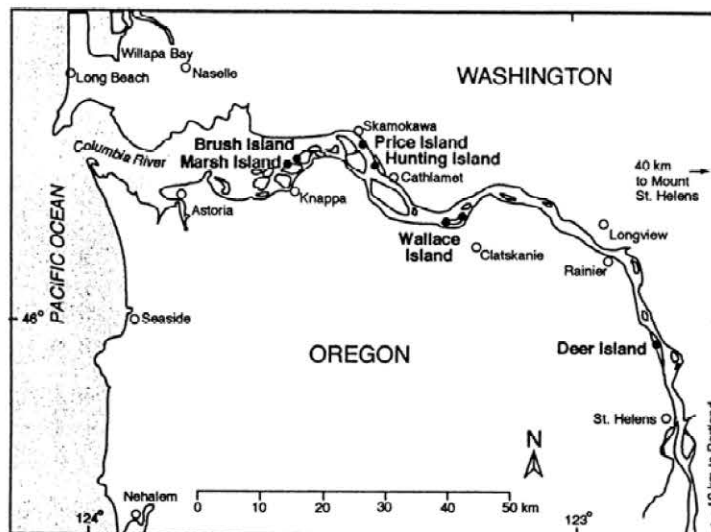


Figure 3: Map showing Columbia River islands where detailed geologic and preliminary geotechnical studies were conducted (after Atwater, 1994).

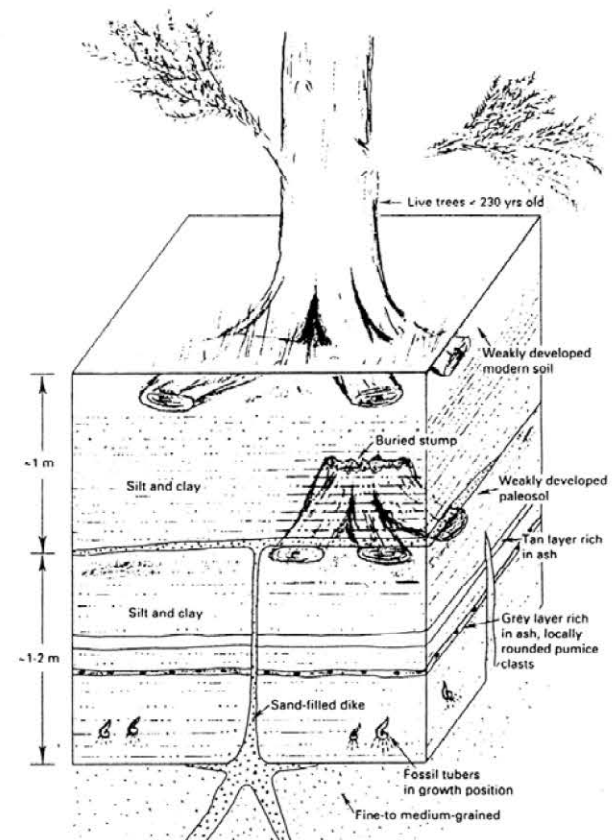


Figure 4: Block diagram showing typical field relations at liquefaction sites on the islands in the lowermost Columbia River.

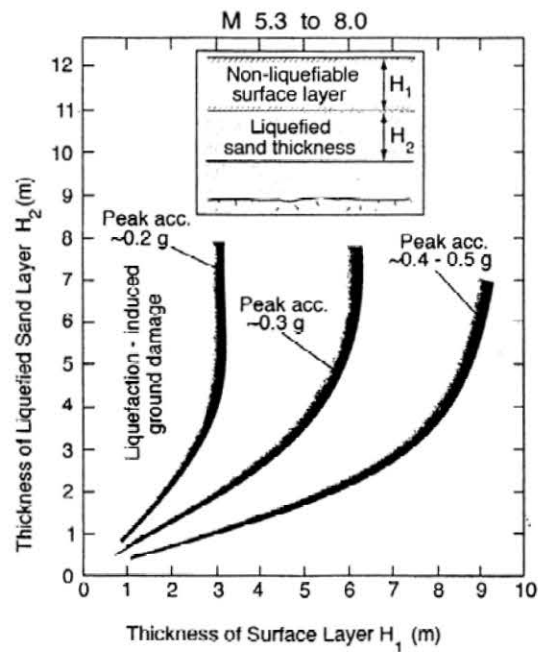


Figure 5: Boundary curves relating thickness of nonliquefiable surface layer to thickness of the liquefied zone, as a function of  $p_{gas}$  required to induce venting at the ground surface (from Ishihara, 1985).

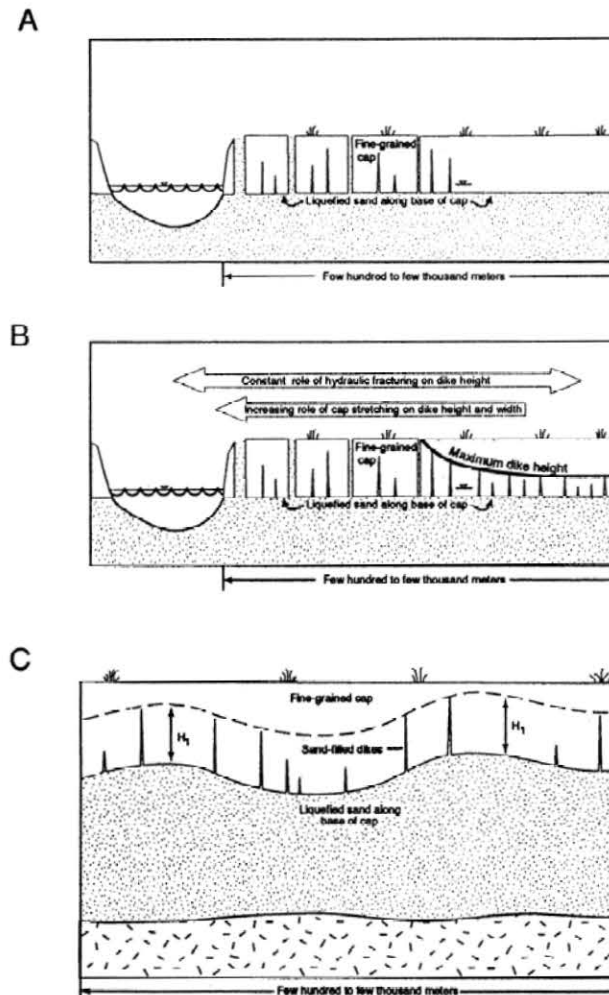


Figure 6: Vertical sections showing liquefaction-induced failure due to (a) lateral spreading, (b) lateral spreading and hydraulic fracturing, and (c) hydraulic fracturing alone.

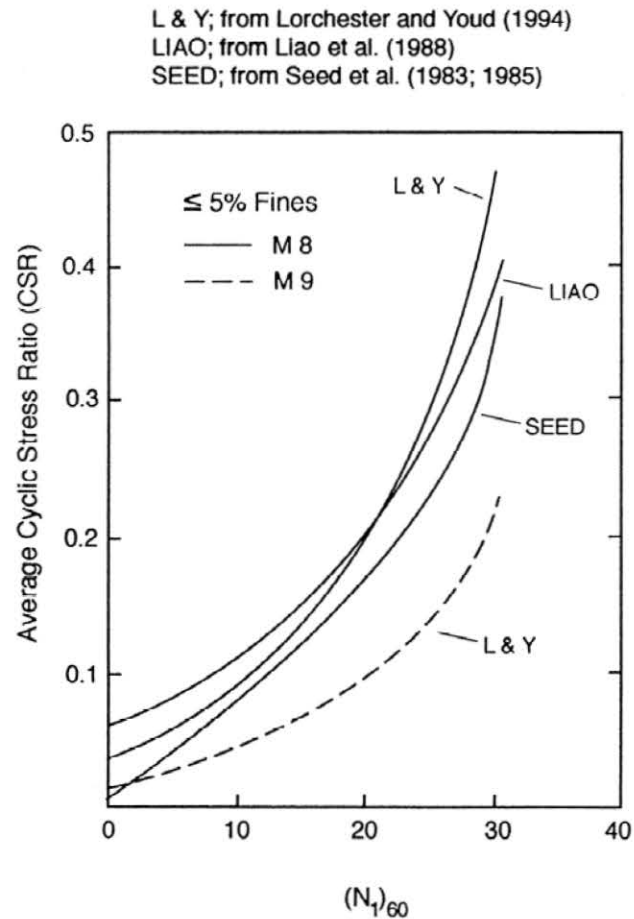


Figure 7: Relationship between stress ratio causing liquefaction and  $(N_1)_{60}$  Values for Sands (modified from Seed et. al. , 1986).

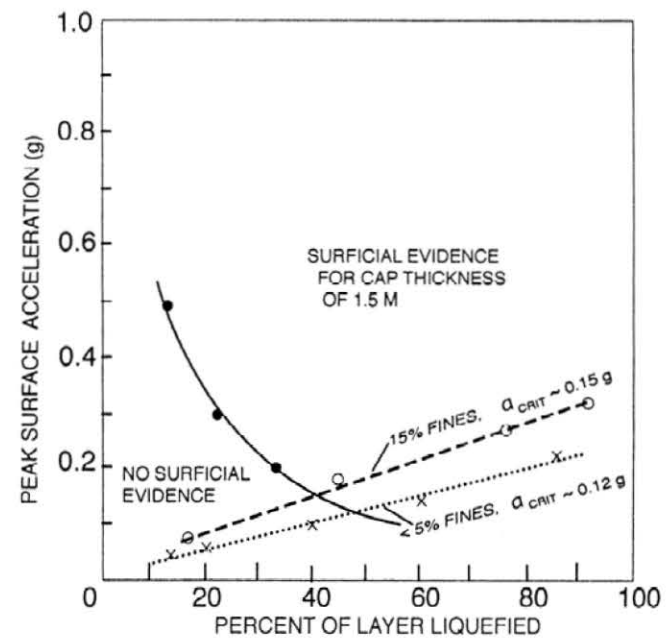


Figure 8: Estimation of the pga at a Marsh Island site using the technique of Martin and Clough (1994).

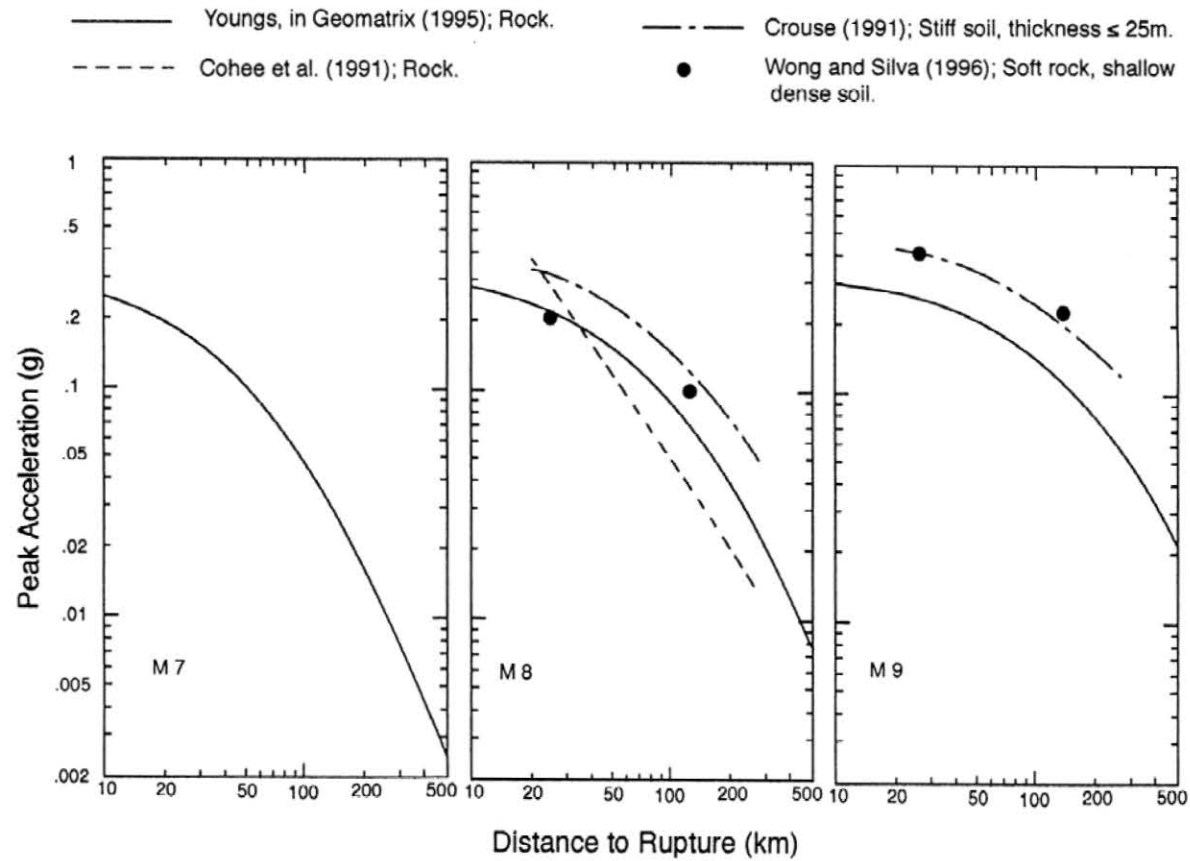


Figure 9: Attenuation relationships for peak accelerations at weak rock and stiff soil sites proposed for interface earthquakes along the Cascadia subduction zone.

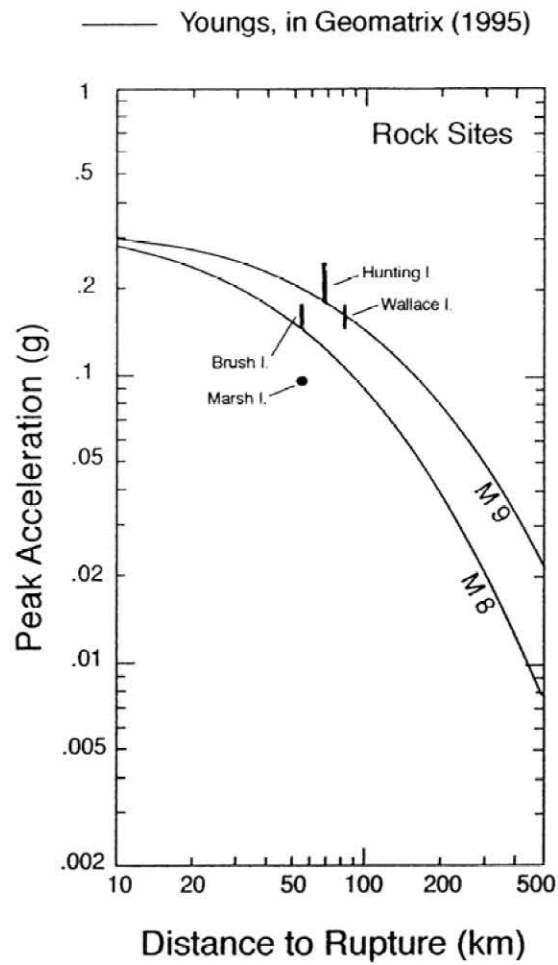


Figure 10: Comparison of the estimated peak accelerations on rock based on liquefaction evidence and selected attenuation relationships for Cascadia subduction zone earthquakes. Liquefaction evaluation based on a M 8 earthquake.



# Ground Motion Attenuation In Subduction Zones

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## Abstract

Exposure and vulnerability to the effects of earthquakes is increasing as urban centers grow, especially in tectonically active areas. The economic and social effects of earthquakes can be reduced through a comprehensive assessment of seismic hazard and risk that leads to increased public awareness, seismically sensitive land-use planning, and the implementation of seismically sound building construction codes. State-of-the-art estimates of expected ground motion at a given distance from an earthquake of a given magnitude are a fundamental inputs to earthquake hazard assessments. Seismic design criteria for any engineered structure depends on plausible, reproducible estimates of the expected ground motions from earthquakes on nearby (or reasonably distant) faults, during the expected lifetime of the structure. These estimates are usually equations, called attenuation relationships, that express ground motion as a function of magnitude and distance (and occasionally other variables, such as type of faulting). Different tectonic environments give rise to different ground motion attenuation relationships.

## Introduction

Earthquakes are among the most deadly and expensive natural disasters affecting humankind. Vulnerability to the effects of earthquakes is increasing as urban centers grow, especially in tectonically active areas. Although all states within the United States (US) experience earthquakes, most of the

seismicity occurs along the western margin plate boundary. From Cape Mendocino northward, the western margin of the US is part of the large circum-Pacific subduction system. There are multiple seismic sources in subduction zones: intraplate earthquakes in both the under- and over-riding plates, and interplate earthquakes. Furthermore, the interplate earthquakes in subduction zones are very large earthquakes. The ten largest earthquakes of the twentieth century (e.g. the largest instrumentally recorded earthquakes) are all interplate subduction (collision) zone earthquakes (Table 1). The second largest known earthquake occurred in the Pacific Northwest. The 1964 Alaska earthquake was catastrophic because large urban centers were effected. A repeat of the 1964 Alaska, or the probable 1700 Pacific Northwest earthquake, would be economically and socially devastating.

Table 1. Earthquakes of Large M

Event	Year	M
Chile	1960	9.5
Alaska	1964	9.2
Aleutian	1957	9.1
Kamchatka	1952	9.0
Ecuador	1906	8.8
Aleutian	1965	8.7
Assam	1950	8.6
Kurile Islands	1963	8.5
Chile	1922	8.5
Banda Sea	1938	8.5

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The economic and social effects of earthquakes (and other natural disasters) can be reduced through a comprehensive assessment of seismic hazard and risk that leads to increased public awareness, seismically sensitive land-use planning, and the implementation of seismically sound building construction codes. Clear, well-documented assessments of seismic hazard are the first and fundamental step in the mitigation process. With the recent and pending publications of the 1996 US National Seismic Hazard Maps (Frankel et al., 1996), the Southern California Earthquake Center Phase II (Working Group on California Earthquake Probabilities, 1995) and Phase III (hazard calculations for several earthquake scenarios effecting the Los Angeles basin) reports, *Putting Down Roots in Earthquake Country* (southern California), *living on shaky ground* (northern California), and the *The Next Big Earthquake: Are you Prepared* series (San Francisco Bay Area and southern Alaska), interest in seismic hazard assessments has steadily increased. The need to understand seismic hazard assessments has expanded from the professional seismic engineering and scientific communities to national, state, county, city, and local public officials, the private sector, educators at all levels, and the general citizenry. Demand for information has never been greater, nor has scrutiny of all steps of the processes that were used to produce the assessments.

### Estimation Of Ground Motion

State-of-the-art estimates of expected ground motion at a given distance from an earthquake of a given magnitude are a fundamental inputs to earthquake hazard assessments. The determination of seismic design criteria for any engineered structure depends on plausible, reproducible estimates of the expected ground motions from earthquakes on nearby (or reasonably distant) faults, during the expected lifetime of the structure. These estimates are usually equations, called attenuation relationships, that express ground motion as a function of magnitude and distance (and occasionally other variables, such as type of faulting).

Ground motion attenuation relationships may be determined in two different ways: empirically, using previously recorded ground motions, or theoretically, using seismological models to generate synthetic ground motions which account for the source, site, and path effects. There is overlap in these approaches, however, since empirical approaches fit the data to a functional form suggested by theory and theoretical approaches often use empirical data to determine some parameters.

The most commonly mapped ground motion parameters are horizontal and vertical peak ground acceleration (PGA), peak ground velocity (PGV), and 5%-damped spectral acceleration (SA) for a given site classification. The 1996 US National Hazard maps include PGA and 0.2, 0.3, and 1.0 s SA with a 10%, 5%, and 2% chance of exceedance in 50 years (Frankel et al., 1996), assuming a "firm-rock" site. The 1996 Canadian Seismic Hazard maps include PGA, PGV, and 0.1, 0.15, 0.2, 0.3, 0.4, 0.5, 1.0 and 2.0 s SA with a 10% chance of exceedance in 50 years (Adams et al., 1996), assuming a "firm ground" site.

The parameters that must be clearly defined in order to estimate ground motions are: earthquake magnitude, type of faulting, distance, and local (receiver) site conditions (classification). Moment magnitude ( $M$ ) is the preferred magnitude measure, because it is directly related to the seismic moment of the earthquake. Style of faulting needs to be specified because, within 100 km of a site, strike-slip earthquakes generate smaller PGA and SA than reverse and thrust earthquakes, except for  $M \geq 8.0$  (Boore et al., 1993; 1994; Campbell and Bozorgnia, 1994).

Different source-to-site distance measures are used by different researchers. A complete summary can be found in Abrahamson and Shedlock (1997). The attenuation relationships discussed in this paper use one of the following:  $r_{rup}$ , the closest distance to the rupture surface;  $r_{scis}$ , the closest distance to the seismogenic rupture surface (assumes

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that near-surface rupture in soft sediments is non-seismogenic (Marone and Scholz, 1998)); or  $r_{\text{hypo}}$ , the hypocentral distance.

Table 2. Site Classification Schemes

NEHRP 1994	BOORE et al. 1993	CAMPBELL	IDRISS SADIGH et al. YOUNGS et al.
A $V_s > 1500 \text{ m/s}$	A	Hard Rock	Rock
B $760 < V_s < 1500 \text{ m/s}$	A $750 \text{ m/s} < V_s$	Hard Rock	Rock
C $360 < V_s < 760 \text{ m/s}$	B $360 < V_s < 760 \text{ m/s}$	Soft Rock	Rock/Stiff Soil
D $180 < V_s < 360 \text{ m/s}$	C $180 < V_s < 360 \text{ m/s}$	Firm Soil	Deep Soil
E $V_s < 180 \text{ m/s}$	D $V_s < 180 \text{ m/s}$	Soft Soil	Soft Soil
		Shallow Soil	

There are also several site classification schemes, ranging from a description of the physical properties of near-surface material to very quantitative characterizations. Table 2 is summary of the most commonly used site classifications, compared to the classification given by the 1994 National Earthquake Hazard Reduction Program (NEHRP) *Recommended Provisions for the Development of Seismic Regulations for New Buildings* (BSSC, 1994). The NEHRP classifications serve as the standard, since they have been adopted by the Building Seismic Safety Council, the International Conference on Building Code Officials, and several State Structural Engineers Associations. The correspondences to the site classifications used by others are approximate, not absolute.

Different tectonic environments give rise to different ground motion attenuation relationships. Data collected within each of the different tectonic environments usually are inadequate to uniquely characterize the region, so averaged attenuation relationships are determined. Currently, three categories of regional ground motion attenuation relationships are used in seismic hazard assessments: shallow crustal earthquakes in active tectonic regions (e.g. western North America), shallow crustal earthquakes in stable continental regions (e.g. central and eastern North America), and subduction zones (e.g. northwest North America). This paper compares and contrasts regional ground motion attenuation relationships in subduction zones.

### Seismic Sources In Subduction Zones

Both inter- and intraplate earthquakes occur in subduction zone environments. Throughout the Pacific Northwest and Alaska, earthquakes occur in three distinct source regions: 1) interface (or subduction zone) earthquakes along the zone of contact between two plates (either the Pacific-North America, the Gorda-North America, or the Juan de Fuca-North America contacts); 2) shallow intraplate (crustal) earthquakes within the crust of the overriding North American plate; and 3) intraplate earthquakes within the subducting plates. Of the three sources, crustal earthquakes in the North American plate and intraplate events within the subducting plates have formed the basis of seismic hazards analyses in the Pacific Northwest while the 1964 interplate earthquake has formed the basis for analyses in Alaska. One of the enigmas of the Cascadia subduction zone is that no recorded earthquakes have occurred on the Juan de Fuca-North America or Gorda-North America interfaces. In most subduction zones, it is the plate interface that produces the great magnitude (8+; Table 1) thrust earthquakes, as in Alaska in 1964. Despite uncertainty surrounding the details of how and when great subduction zone earthquakes may occur in the Pacific Northwest, there is growing acceptance of the past occurrence of these events, most recently in 1700 (Satake, et al., 1996). Thus, Canadian and US scientists have incorporated the possibility of great subduction zone earthquakes into seismic hazard analyses (Frankel, et al., 1996; Adams, et al., 1996).

The different earthquake source zones and physical properties of the crust throughout the subduction zone require the use of several ground motion relationships in seismic hazards assessments. Ground motions from intraplate earthquakes in the overriding plate are estimated using relationships derived for shallow crustal earthquakes in active tectonic regions. Ground motions from intraplate earthquakes occurring in the subducting plate and the great interface earthquakes require the use of attenuation relationships developed for deeper events. Both the Canadian and US hazard maps incorporated multiple attenuation relationships in the hazard calculations to represent the uncertainty in modeling strong ground motions from the various sources.

### Attenuation Relationships

Shallow earthquakes in active tectonic regions have provided the largest amount of ground motion data and the largest number of ground motion attenuation relationships. With this large data set, effects of parameters other than just magnitude, distance, and site condition can be evaluated. For example, in most recent attenuation models, there is a distinction between the ground motion from reverse events and strike-slip events called the style-of-faulting factor (Table 3).

Campbell (1997) summarized the results of several years work developing empirical attenuation relationships for horizontal and vertical PGA, PGV, and SA in active tectonic regions. Idriss (1991) and Sadigh and others (1997) developed attenuation relationships for shallow crustal earthquakes using strong motion data recorded in California. All three sets of relationships illustrate that ground motions are larger for reverse/thrust earthquakes than for strike-slip earthquakes.

There are few strong motion recordings from subduction zone earthquakes in the United States, so most attenuation models for subduction zone events are primarily based on recordings from Japan and South America. Most subduction zone events are recorded at large distances because the events tend to be deep or offshore. The sparse data within 30 km leads to a large uncertainty in the extrapolation of these models to short distances. The exception is the recording of the 1985 Michoacan earthquake by the Guerrero array, which had stations as close as 13 km to the epicenter.

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Table 3. Summary Information on Attenuation Relationships

Model	Period (s)	Comp	Site Conditions	Model Parameters <sup>1</sup>
Anderson	0.0	H, V	Rock	$M_s$ or $m_b$ , $r_{rup}$
Atkinson/Boore	0.0-2.0	Ave H	Rock, Soil Amp factors	$M$ , $r_{hypo}$
Campbell	0.0-4.0	Ave H, V	Hard Rock, Soft Rock, Soil	$M$ , $r_{seis}$ , $F_1$ , $D$
Crouse	0.0-4.0	Ave H	Stiff Soil	$M$ , $r_{hypo}$ , $H$
Idriss	0.0-5.0	Ave H	Rock/Stiff Soil, Deep Soil, Soft Soil	$M$ , $r_{rup}$ , $F_2$
Sadigh <i>et al.</i>	0.0-4.0	Ave H	Rock, Soil	$M$ , $r_{rup}$ , $F_2$ , $HW$
Youngs <i>et al.</i>	0.0-4.0 <sup>2</sup>	Ave H	Rock/Stiff Soil, Deep Soil, Soft Soil	$M$ , $r_{rup}$ , $F_3$ , $H$

Anderson (1997) developed a nonparametric model for PGA in a subduction zone, using data from the Guerrero, Mexico, accelerograph network. This approach differs from all the others attenuation functions in that Anderson determines a table of PGA values, for specific values of  $M$  and  $r_{hypo}$ , and an interpolation rule for intermediate values. Anderson's model is valid for earthquakes with magnitudes ranging from less than 3 to 8.1.

Atkinson and Boore (1997) developed preliminary ground motion relationships for the Cascadia region that may be used to predict ground motions from  $M < 7$  earthquakes at all distances and to predict conservative ground motions from larger earthquakes at distances less than 100 km.

Crouse (1991) used nearly 1000 records from subduction zone earthquakes "considered representative of the Cascadia subduction zone" to develop attenuation relationships for firm-soil sites (comparable to Boore *et al.* class C, Table 2) in the Pacific Northwest. Adams *et al.* (1996) adjusted these relationships to class B for use in their hazard maps by adding a period-dependent constant determined from the Boore *et al.* (1993) relationships. We use these modified Crouse relationships in our comparisons.

Youngs *et al.* (1997) developed attenuation relationships for subduction zone interface and intraslab earthquakes. They illustrated that peak ground motions from subduction zone earthquakes attenuate more slowly than those from shallow crustal earthquakes in tectonically active regions and that intraslab earthquakes produce larger peak ground motions than interface events for the same magnitude and distance.

<sup>1</sup> Model Parameters:

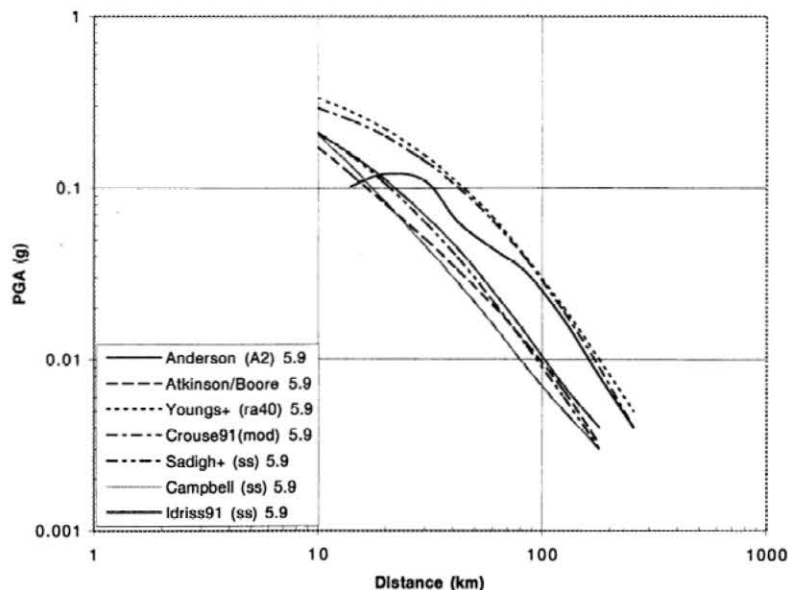
$F_1$  = 1 for reverse or reverse/oblique, 0 otherwise  
 $F_2$  = 1 for reverse, 0.5 for reverse/oblique, 0 otherwise  
 $F_3$  = 1 for intra-plate, 0 for interface  
 $D$  = depth to basement rock at receiver (km)  
 $H$  = hypocentral depth (km)  
 $HW$  = 1 for hanging wall sites, 0 otherwise

<sup>2</sup> Period range is 0.0-3.0 s for rock site conditions.

## Discussion

The median (except for Anderson (1997), which predicts the largest) PGA predicted by seven attenuation relationships for a  $M$  5.9 intraplate earthquake in Cascadia are shown in Figure 1. The modified Crouse (1991) and Youngs et al. (1997) relationships predict higher PGA (a factor about 2) at all distances than the other five, although the PGA predicted by the nonparametric model of Anderson is comparable at distances greater than 80 km. The higher predictions from the Youngs et al. (1997) relationship is not surprising. Youngs et al. (1997) developed these relationships expressly for intraslab earthquakes, which have a lower rate of attenuation of peak ground motions than do earthquakes in the shallow crust. Furthermore, Youngs et al. (1997) illustrate that peak ground motions increase with depth of intraslab earthquakes. The Youngs et al. (1997) curve shown in Figure 1 assumes a 40 km hypocenter depth, greater than the shallow crust depths assumed by the other relationships (except Anderson (1997)). Similarly, the higher values calculated using the modified Crouse (1991) relationship are due to the initial assumptions in their development. The original Crouse (1991) relationships were developed for "firm soil" sites, which consistently yield higher values of peak motions than do rock or firm rock sites. Period dependent terms were used to modify the Crouse (1991) relationships for an intraslab/subduction zone environment; thus, the modified Crouse (1991) relationships should have higher peak motions than relationships developed for the shallow crust.

Cascadia 5.9

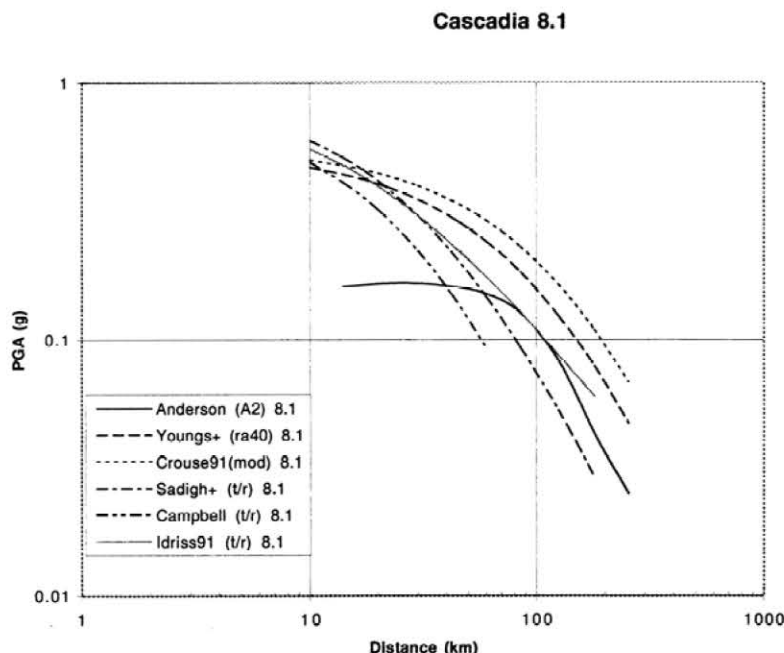


**Figure 1.** Peak ground acceleration (PGA), in units of  $g$ , predicted by seven attenuation relationships for a  $M$  5.9 earthquake in the Cascadia region. (ss) means the values are predicted for a strike-slip or normal earthquake. + signifies that there are co-authors of the functions. (A2) refers to the model designation given by Anderson (1997). (ra40) means that the values are calculated assuming an intraslab earthquake at 40 km hypocentral depth. (mod) signifies that the values are calculated from the modified version of the Crouse (1991) relationships.



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The median PGA predicted from a  $M$  8.1 earthquake in Cascadia is plotted for six attenuation relationships in Figure 2. With the exception of Anderson (1997), the relationships predict essentially the same PGA within the first few tens of km of the fault. At about 100 km from the fault the predicted PGA values differ by less than a factor of two. Just as at lower magnitudes, the higher PGA values are predicted by relationships developed for use in intraslab/subduction zone environments and the lower values are predicted by relationships developed for use with shallow earthquakes in active tectonic regions.



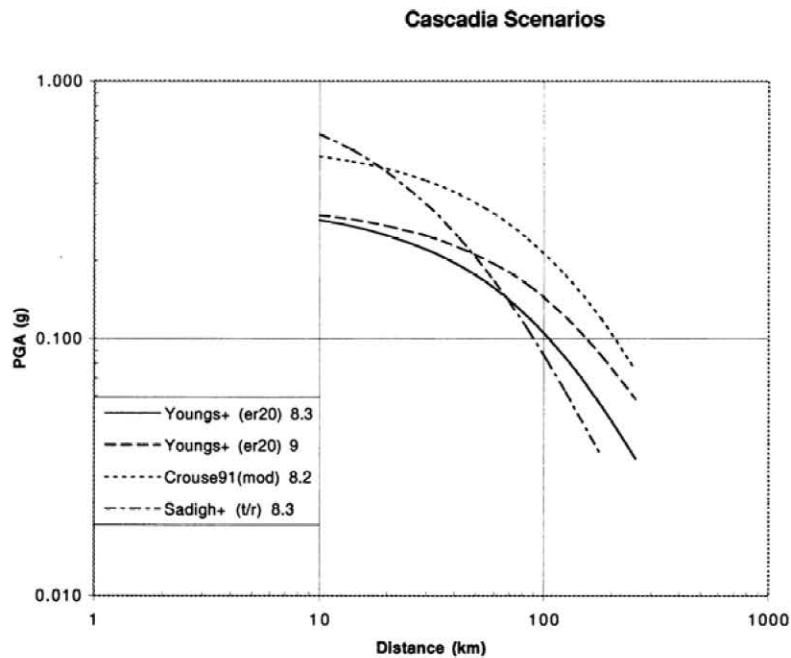
**Figure 2.** Peak ground acceleration (PGA), in units of  $g$ , predicted by six attenuation relationships for a  $M$  8.1 earthquake in the Cascadia region. (t/r) means the values are predicted for a thrust or reverse earthquake. + signifies that there are co-authors of the functions. (A2) refers to the model designation given by Anderson (1997). (ra40) means that the values are calculated assuming an intraslab earthquake at 40 km hypocentral depth. (mod) signifies that the values are calculated from the modified version of the Crouse (1991) relationships.

The nonparametric approach of Anderson (1997) yields curves that exhibit different behavior than the regression analysis approaches (Figures 1 and 2). Anderson (1997) found that PGA saturates as magnitude increases close in to the fault ( $< 25$  km distance) and PGA decreases more rapidly with distance for small earthquakes ( $M < 6$ ). Overall, however, the predicted PGA values fall between the interplate and intraslab earthquake PGA values predicted by Youngs et al. (1997) for  $M \leq 7$  earthquakes (Anderson and Lei, 1994). Although the use of the nonparametric approach is constrained by the magnitude and distance ranges of recorded data, it does model greater complexity of ground motion than regression analyses.

Based on the comparisons of PGA predicted from  $M$  5.9 (medium) and 8.1 (large) earthquakes, the conservative approach to seismic hazard mapping in Cascadia would be to use the attenuation relationships developed for the region (i.e. Youngs et al., 1997, or the modified Crouse, 1991).

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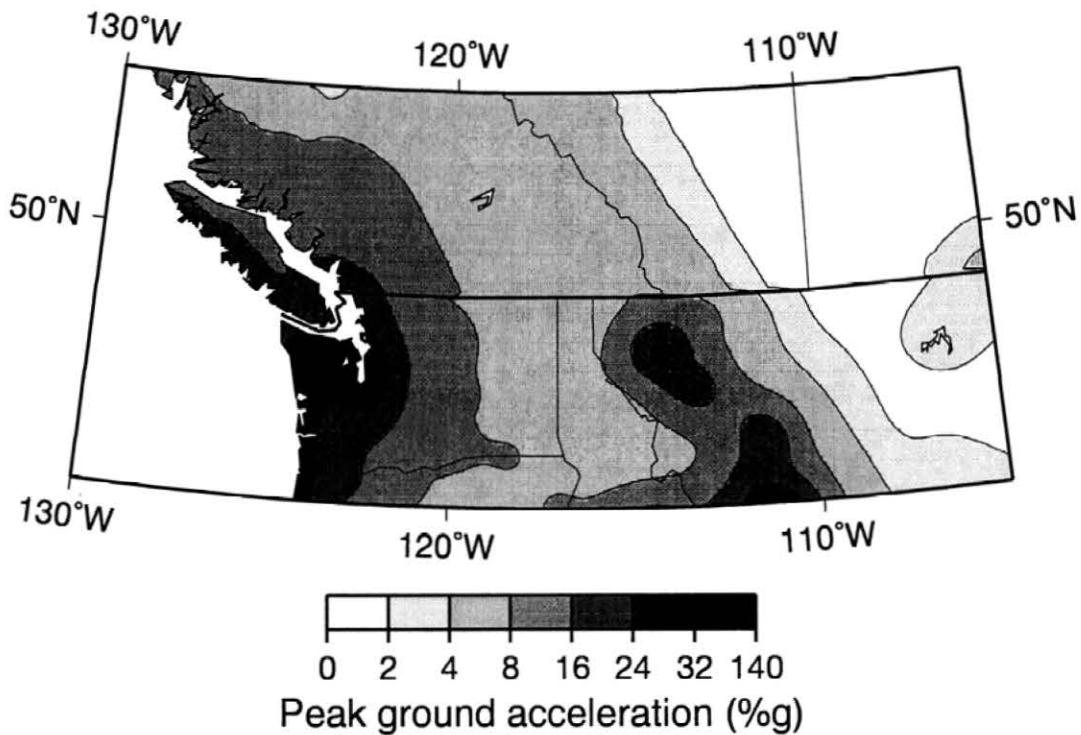
Both the 1996 Canadian and US National Seismic Hazard maps incorporate scenario Cascadia subduction earthquakes. The scenario great earthquake adopted by Canada for the maps is a **M** 8.2 with the center of energy release approximately one-third of the distance from the locked zone into transition zone. Canada assumed the modified Crouse (1991) attenuation relationship to predict the resulting ground motions. The US maps incorporate two weighted scenarios for the great Cascadia subduction earthquake (Frankel et al., 1996). The first scenario (weight = 0.67) floats a **M** 8.3 rupture zone along the subduction zone, allowing enough **M** 8.3 earthquakes to rupture the entire subduction zone every 500 years. The second scenario (weight = 0.33) allows a **M** 9.0 earthquake to rupture the entire subduction zone every 500 years. The US assumed the Youngs et al. (1997) attenuation relationships for both scenarios, equally weighted with the Sadigh et al (1997) for the first scenario and alone for the second. The median PGA predicted by each of these relationships for the different scenarios is shown in Figure 3. The predicted PGA values differ by roughly a factor of two at all distances. The values predicted by Sadigh et al. (1997) are largest within the closest few tens of km of the fault, but rapidly decrease to the smallest at distances greater than about 60 km. However, the Sadigh et al. (1997) relationships were developed using strong motion data primarily from California earthquakes and stations (shallow earthquakes in an active tectonic region). Their extrapolation to a great Cascadia subduction earthquake provides a conservative prediction for near-fault ground motions. The Youngs et al. (1997) relationships provide the conservative estimates at greater distances.



**Figure 3.** Peak ground acceleration (PGA), in units of *g*, predicted by three attenuation relationships for the US and Canadian scenario subduction earthquakes in the Cascadia region. (t/r) means the values are predicted for a thrust or reverse earthquake. + signifies that there are co-authors of the functions. (er20) means that the values are calculated assuming an interslab earthquake at 20 km hypocentral depth. (mod) signifies that the values are calculated from the modified version of the Crouse (1991) relationships.

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The US and Canada had other differences in their hazard mapping assumptions in addition to the attenuation functions and scenario Cascadia subduction earthquakes. For example, Canada used clearly defined source zones (Adams et al., 1996); the US used combinations of  $M \geq 3$  earthquake recurrence rates and background seismicity (Frankel et al., 1996). The different assumptions made by Canadian and US scientists in their hazard maps result in important differences in the maps (Figure 4). In Cascadia, the PGA contours are offset across the Canada/US border, by as much as 100 km in the lowest hazard ranges (Figure 4). PGA predictions for the highest hazard, and coincidentally the most densely populated, areas agree reasonably well, although the US map predicts higher PGA in the Seattle-Vancouver (BC) urban areas than does the Canadian map.



**Figure 4.** Peak ground acceleration (PGA), in units of %g, having a 10% probability of being exceeded in 50 years. Values shown for the US are from Frankel et al. (1996). Values shown for Canada are from Adams et al. (1996).

## Summary

The complicated earth structure and tectonics of subduction zones require that multiple attenuation functions be used in seismic hazard assessments. There are three types of earthquake source regions in Cascadia: 1) interplate earthquakes; 2) shallow intraplate earthquakes in the overriding plate; and 3) intraslab (intraplate) earthquakes in the subducting plate. Ground motions differ from each source. The conservative approach to seismic hazard assessment in Cascadia (and other subduction zones) includes using attenuation functions developed expressly for subduction regimes along with those developed for shallow earthquakes in active tectonic regions, and including scenario earthquakes in the hazard calculations.

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