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Earthquake Hazards in the Pacific Northwest of the United States

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EARTHQUAKE HAZARDS IN THE PACIFIC NORTHWEST: AN OVERVIEW

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Foreword

This paper is one of a series dealing with earthquake hazards of the Pacific Northwest, primarily in western Oregon and western Washington. This research represents the efforts of U.S. Geological Survey, university, and industry scientists in response to the Survey initiatives under the National Earthquake Hazards Reduction Program. Subject to Director's approval, these papers will appear collectively as U.S. Geological Survey Professional Paper 1560, tentatively titled "Assessing Earthquake Hazards and Reducing Risk in the Pacific Northwest." The U.S. Geological Survey Open-File series will serve as a preprint for the Professional Paper chapters that the editors and authors believe require early release. A single Open-File will also be published that includes only the abstracts of those papers not included in the pre-release. The papers to be included in the Professional Paper are:

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Nelson, Alan R. and Personius, Stephen F., "The potential for great earthquakes in Oregon and Washington: An overview of recent coastal geologic studies and their bearing on segmentation of Holocene ruptures, central Cascadia subduction zone"

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ABSTRACT

Scientific research on earthquake hazards in the Pacific Northwest through 1991 suggests that the hazard is greater than previously recognized. A great earthquake of $M \geq 8$ on the thrust fault that separates the North America and Juan de Fuca plates (commonly termed the Cascadia thrust or megathrust fault) or a large earthquake of about $M \approx 8$ or less on shallow faults in the continental crust would be of significantly larger magnitude and would produce much greater damage than previously experienced by the modern inhabitants of this region. Furthermore, damaging events similar to the 1949 and 1965 Puget Sound earthquakes, which occurred in the subducted plate, are expected to recur. In some circumstances, marginally different locations or larger magnitudes for these events could substantially increase the levels of damage compared to that observed historically.

A variety of geological and geophysical data suggest that great earthquakes (earthquakes with magnitudes between 8 and 9.5) have occurred in the last 7000 years on the Cascadia thrust fault. First, marsh deposits along portions of the Washington and Oregon coast record episodic sudden submergence. The sudden submergence events, which have been inferred from peaty soils overlain by tidal mud deposits in coastal estuaries, have been ascribed to tectonic subsidence associated with past great earthquakes along the Cascadia thrust fault. In some cases, the buried peaty soils are capped by thin sand layers that suggest sand–bearing tsunamis coincided with the drowning of tidal wetlands during some of the earthquakes. Second, geodetic measurements suggest that the continental crust in western Washington is undergoing compression, where the direction of maximum compressional increase is parallel to the direction of convergence between the Juan de Fuca and North America plates. Third, Pleistocene and Holocene folding and faulting of offshore and coastal sediments give evidence of deformation that may have accompanied subduction. Fourth, the Cascadia subduction zone has several characteristics similar to those in other subduction zones that have produced great earthquakes historically. Fifth, ground shaking during prehistoric earthquakes may have triggered thirteen sea–floor turbidite deposits at widely scattered locations along the Cascadia margin. Most scientists believe that these and other data are consistent with continued subduction, and some believe the data are consistent with seismic subduction and great prehistoric earthquakes.

High–precision radiocarbon ages suggest that the last sudden submergence and tsunami occurred about 300 years ago. The time between such earthquakes, however, may range from a few centuries to more than a millennium. At present, a reliable forecast of the time until the next event is not possible. Nonetheless, because the time until the next event is unknown and its effects are likely to be great, such an event should be considered in earthquake risk assessment and disaster planning for the region .

The consequences of a great earthquake on the Cascadia thrust fault are significant because of the regional exposure to most types of earthquake hazard. Damaging shaking may reach across much of western Oregon, western Washington, and northern California in a great earthquake. This type of event may also trigger widely scattered ground failure in susceptible areas. Tsunami and seiche that may accompany a great earthquake can further compound the levels of damage from ground failure and shaking in coastal zones.

Some scientists, however, argue for aseismic subduction of the Juan de Fuca plate. For example, no great earthquakes have occurred since the arrival of European settlers, nor have sensitive seismographs detected small earthquakes on the Cascadia thrust fault. The case for aseismic subduction is also based on inferences about fault zone properties and slab temperatures that suggest the downdip width of the locked section of the thrust fault is limited. The maximum magnitude of great Cascadia earthquakes may be limited if the model conditions prevail, or seismic subduction may even be precluded. It is also argued that some estuary submergence events may have causes other than great Cascadia earthquakes. For example, the effects of eustatic sea level rise, dating procedures, and local tectonics cloud the interpretation of some of the submergence events. Yet, the evidence for great earthquakes appears stronger than the evidence against such events.

The potential for shallow continental crust earthquakes, of less than about magnitude 8, is also increasingly apparent. New geologic evidence indicates that faults capable of generating large continental crust earthquakes may exist in western Washington, western Oregon, and northwestern California, including some urban areas. These earthquakes have the potential to produce locally higher levels of damage than great subduction earthquakes when they underlie or pass near urban areas. The long–term hazard from earthquakes on these faults is unknown because many shallow faults may be unmapped, and earthquake recurrence rates and maximum magnitudes on these faults are unknown.

Intermediate–depth earthquakes, at 40–80 km in the subducted Juan de Fuca plate, similar in type to those that occurred in 1949 and 1965, are expect to recur. These earthquakes may nucleate in a shallower part of the subducted plate, occur closer to urban areas, or have greater magnitude than those in the

past. A maximum magnitude of about 8 is possible for this type earthquake. Thus, these events have the potential to produce greater damage and life loss than observed historically. Earthquake risk assessment and disaster planning efforts should also include these hypothesized more damaging earthquakes.

Much new research, presented in this volume and other recent studies, provides the stimulus to increase awareness of the earthquake hazard in the Pacific Northwest and to encourage steps to reduce the risk. Such steps should involve hazard prediction, loss estimation, citizen education, disaster management, urban planning, and continued evaluation of earthquake building codes and code enforcement.

PREFACE

PROGRAM GOALS

To increase knowledge and understanding of geologic hazards and risk in the Pacific Northwest, the U. S. Geological Survey initiated a research program focused on this problem in the late 1970's (the volcanic hazard has traditionally been treated separately, for example, see Crandell and others, 1979). Funding for these studies was increased in 1987 to support a 5–yr accelerated research program under the aegis of the National Earthquake Hazards Reduction Program. These studies were primarily conducted under an element of the National program termed "Urban and Regional Hazards Assessment," which previously supported intensive studies in Salt Lake, Utah, Los Angeles, and San Francisco, California. Typically, scientists, engineers, and sociologists from Federal agencies, state geological organizations, universities, and private industry have participated in this research program. Some of the scientific studies from the first three years of the program are presented in chapters of this Professional Paper volume. This particular chapter summarizes the state of knowledge concerning geologic hazards and risk in this region through 1991, including the studies herein and other relevant studies.

VOLUME GOALS

The broadest goals of this volume are to promote the recognition, assessment, and reduction of the earthquake hazard in the region. Engineers, planners, decisionmakers, and land and building owners should, in the long term, use this new information to reduce the effects of expected future earthquakes in Washington, Oregon, California, and British Colombia.

Because individuals in many segments of society are responsible for understanding earthquake hazards and making decisions to reduce earthquake effects, this volume is likely to have a broad audience. To some extent, the diversity in the contents of this volume reflects the diversity of the audience. While many of the research chapters necessarily address other specialists having some familiarity in the geological and geophysical sciences and earthquake hazard mitigation, this overview chapter and the chapters on implementation of hazard reduction techniques are intended to inform both technical and non-technical readers. To further aid the non–technical reader, we have defined technical terms used throughout the volume in the Glossary (Appendix B). Because some issues are inherently complex and can not be simplified without losing significant context, we recommend other discussions to further clarify some aspects of the problem or to provide additional detail and a different perspective (i.e., Heaton and Hartzell, 1987; Noson and others, 1988; Riddihough, 1978; Rogers, 1988a; Rogers, 1988b; Shedlock and Weaver, 1991).

PACIFIC NORTHWEST: DEFINING THE GEOGRAPHICAL STUDY REGION

The Pacific Northwest, in some contexts, may refer to a very large region, including all of Oregon, Washington, northern California, and western British Columbia, Canada. In this volume, the Pacific Northwest refers primarily to the region west of the Cascade Range in Oregon and Washington, northwestern California, and southwestern British Columbia, including Vancouver Island (see fig. 1a, b for place names and major geologic features). The geographic emphasis for our study area developed for several reasons. Large damaging earthquakes are part of the regional documentary record, yet for various reasons, we do not yet have confidence in the estimates of earthquake hazard. New data have increased our concern about the potential for great subduction zone earthquakes and large shallow earthquakes. The rapid increase in population density and urban development during the last decade has also increased the risk. Finally, to some extent, this study area includes a region broad enough to span much of the subduction tectonic province, permitting us to properly address the problems across a single province.

While our discussion of the Pacific Northwest includes the large region described, much of the research contained in this volume necessarily focuses on topical studies within smaller areas, especially in the states of Washington and Oregon. Unfortunately, the topical and geographic limits of this research preclude a full description and understanding of the hazard at present. Research concerning the regional earthquake hazard has also exposed a number of important unresolved issues. These issues, which are the subject of ongoing scientific dialogue are, to some extent, a consequence of the remarkable diversity and complexity of past and present tectonic processes, a fact that permits multiple interpretations of the same data in some cases. But the uncertainty concerning some interpretations also results from scant data and the infancy of scientific research on some topics.

Future data collection, new techniques, and continued research will ultimately provide a greater level of understanding of the hazard than is currently possible. The present state of knowledge, however, is sufficient to continue planning and decision making to reduce the effects of future earthquakes. Both the steps to increase understanding of the hazard and efforts to reduce the hazard should be viewed as long-term goals spanning decades or more, steps that will ultimately increase the level of protection in the urban environment in this region against earthquake hazards.

RECOGNIZING POTENTIAL EARTHQUAKE HAZARDS AND RISK IN THE PACIFIC NORTHWEST

The earliest inhabitants of the Pacific Northwest suffered the consequences of a wide range of geological hazards (Heaton and Snavely, 1985; Grant and Minor, 1991). Even though generations of retelling have left some aspects unclear, stories still remain among native peoples concerning some of these events. Stories told by the Makah and other tribes suggest the occurrence of a large tsunami along the coast of Washington. Yurok mythology relates the existence of a powerful earthquake god (Heaton and Snavely, 1985), and the Chinook have traditions describing earthquakes that shook their houses and raised the ground (Gibbs, 1955); Gibbs (1955) noted evidence for recent elevation and depression of the land at Shoalwater (now Willapa) Bay, Washington, a locale of these Indians. Recent findings about the geology of the region suggest that at least some of these traditions have basis in fact (i.e., Atwater, this volume).

Some early events show up unequivocally in the documentary record. A large ashfall between 1770 and the early part of the nineteenth century contributed to sickness and death in the Sanpoil tribe (possibly several eruptions from Mt. Hood, Glacier Peak, or Mt. St. Helens) (Coombs and others, 1977). The earliest earthquake in the region's documentary history occurred in 1820 (Coombs and others, 1977) in association with an eruption of Mt. Rainier (then Mt. Tacoma). Settlers at Fort Nisqually, Washington felt another early earthquake in 1833 (Bradford, 1935). European settlers and Indians alike reported the effects of the magnitude 7.4, 1872 Lake Chelan earthquake, which was felt from Eugene, Oregon, to British Columbia. The direct and indirect effects of this earthquake caused many deaths among the Indians. The total number of deaths is unknown; nonetheless, multiple accounts mention the death of Indians from rock falls, emissions of poisonous gases, and famine following the earthquake (Coombs and others, 1977). Altogether there have been at least 57 earthquakes in the Pacific Northwest large enough to produce some damage (Modified Mercalli intensity of VI or greater; Coffman and von Hake, 1973). A few of these events caused or had to potential to cause substantial damage (MMI ≥ VII) (table 1; figs. 2–4).

Recently collected geologic and geodetic data (Adams, this volume; Atwater, this volume; Nelson and Personius, this volume; Peterson and Darienzo, this volume; Savage and others, 1991; Darienzo and Peterson, 1990) suggest that the region could experience major earthquakes along the Cascadia subduction zone in the future. Future rupture of a 300– to 1200–km segment of the Cascadia fault would produce a great earthquake of magnitude 8 to 9.5. (Heaton and Hartzell, 1987), which is much larger than any historical earthquake in the region. While historical earthquakes have had a significant impact, no earthquakes have occurred in modern times in the Pacific Northwest having the intensity or the geographic extent of damage that is likely in a great Cascadia thrust fault event.

Unfortunately, few data currently exist concerning earthquake characteristics such as the frequency, size, and location of future damaging earthquakes and the expected level and geographic distribution of ground shaking and ground failure. Furthermore, the frequency of large damaging earthquakes is low compared with many subduction zones around the world, and no great (M≥8) earthquakes have occurred in the Pacific Northwest in the 200–year historical record of European settlers. The apparent discrepancy between the historic and geologic records, differing interpretations of these and other relevant data, and scant data on some aspects have produced uncertainty and debate regarding the degree of earthquake hazard from great earthquakes.

Furthermore, it is difficult to establish the full consequences of a great earthquake in this region for several reasons. First, some properties of the source, including the lateral and downdip extent of the expected rupture, are not established; the dimensions of the locked portion of the fault strongly control the magnitude of the expected earthquake and the proximity of the radiating section of the fault to urban areas. Second, the attenuation of strong ground shaking in this region is not adequately established for any earthquake source type. Third, the characteristics of ground shaking effects related to geologic site conditions in urban areas are just beginning to be studied (King and others, this volume; Madin, this volume; Wong and others, 1990), yet these effects could strongly influence the levels of shaking in urban areas. Until many of these issues are better resolved, our understanding of the level of hazard from an event of this type will remain clouded.

It is likely, however, that should a great

Table 1. Large Damaging (MMI ≥ VII) Earthquakes of the Pacific Northwest since 1833.

Local Date	Location	Maximum Intensity (MMI)	Magnitude (M_s)
December 14, 1872	Near Lake Chelan, Washington	IX	7–7.4[1]
November 22, 1873	Southwestern Oregon	VIII	6.7[2]
October 12, 1877	Cascade Range, Oregon	VIII	
March 7, 1893	Umatilla, Oregon	VII	
April 14, 1898	Mendocino County	VIII–IX	
March 17, 1904	About 60 km NW of Seattle	VII	
January 11, 1909	North of Seattle, near the Washington/British Columbia border	VII	
December 6, 1918	Vancouver Island, B.C.	VIII[1]	7
January 24, 1920	Straits of Georgia	VII[1]	
January 22, 1923	Off Cape Mendocino (offshore)	VII–VIII	7.3[3]
July 15, 1936	Northern Oregon or southeastern Washington, near Freewater	VII	5.7[1]
November 13, 1939	NW of Olympia, Washington	VII	5.8[1]
April 29, 1945	About 50 km SE of Seattle, Washington	VII	
February 15, 1946	About 35 km NNE of Tacoma, Washington	VII	6.3
June 23, 1946	Vancouver Island, B.C.	VIII[1]	7.2[5]–7.3[6]
April 13, 1949	Between Olympia and Tacoma, Washington	VIII	7.1[6]
April 29, 1965	Between Tacoma and Seattle, Washington	VIII	6.5[7]
November 8, 1980	North of Cape Mendocino (offshore)	VII	7.0–7.2[4]
August 17, 1991	Near the coast of northern California	VII	6.0[8]

Values adapted from Algermissen (1983) unless footnoted. [1]Estimated maximum intensity or magnitude estimated from intensity data. Unless otherwise indicated, magnitude (M) in this volume refers to moment magnitude (M_w) or surface wave magnitude (M_s). See the glossary for a definition of these terms. The 1872 magnitude is estimated as M = 7.4 (Malone and Bor, 1979) assuming a depth of 60 km; M = 7 is estimated from the maximum intensity of IX (Itself estimated assuming a shallow crustal depth by Hopper and others (1982)). [2]Toppozada and others (1981). [3]Coffman and von Hake (1973). [4]Person (1981). [5]Rogers and Hasegawa (1978). [6]m_b, Abe (1981). [7]m_b or M_s, Algermissen and Harding (1965). [8]U.S. Geological Survey (1991). Two other earthquakes occurred within 24 hours of this more damaging event with magnitudes between 6.3 and 7.1, but these events were farther offshore and northwest of the listed event.

earthquake occur, the resulting shaking and ground failure will affect a wide region because of the great length and shallow dip of the fault; furthermore, other hazards such as tsunami and seiche are also likely. Thus, in a single great earthquake, this combination of effects could produce concurrent damage in a number of urban centers along the western regions of Oregon, Washington, northern California, and British Columbia.

Aside from the great earthquake issue, however, the documentary and geologic records tells us that large earthquakes can be expected on faults other than the Cascadia thrust. For example, large continental crustal earthquakes have occurred on Vancouver Island and possibly in the northern Cascades. Geologic evidence suggests that young shallow faulting may exist close to major population centers in both the Puget Sound and Willamette Basin regions (i.e., Yeats, this volume; Wilson and others, 1979; Harding and others, 1988; Bucknam and Barnhard, 1989). Because of the proximity of shallow faults to developed areas, moderate to large earthquakes on these faults might be as damaging in urban areas as more distant earthquakes on the Cascadia thrust fault.

Continued subduction of the section of the lithosphere below the thrust fault is likely to produce 1949– and 1965–style earthquakes with no further convergence between the North America and Juan de Fuca plates. The occurrence of earthquakes similar to those in the past, but with slightly different magnitudes or locations, could lead to greater loss of life and property than observed historically. Together, the 1949 and 1965 earthquakes in the Puget Sound region caused well over $200 million (1984 dollars) in property damage and 15 deaths (May and Noson, 1986). In 1975, the U. S. Geological Survey (1975) estimated that an earthquake similar to the 1949 event could lead to 2200 deaths, 8700 serious injuries, and as many as 23,500 homeless in the Washington counties of King, Kitsap, Mason, Pierce, Snohomish, and Thurston. The hypothetical Benioff zone earthquake (see the Glossary) produced effects that were more severe than the 1949 and 1965 events for several reasons. The assumed magnitude (M=7.5) was larger, the assumed hypocenter was at shallower depth (50 km), and the assumed epicenter was closer to urban areas than the earlier earthquakes (of the scenarios considered in this study, the greatest life loss and injury was produced by an earthquake hypocenter beneath south Seattle). Nonetheless, the assumptions are realistic for a Benioff zone event, which was a recognized earthquake source at the time of the study.

Infrastructure and population have significantly increased in Washington and Oregon since these estimates were made, adding to the number of lives and dollars at risk. May and Noson (1986) estimated that since the U. S. Geological Survey (1975) study, the population in the Puget Sound region alone has increased by 25 percent, and the total assessed value has increased by 240 percent. At present, however, we do not know the effect of these changes on anticipated losses.

Although the documented record of moderate earthquakes in the region has, in some cases, prompted application of building codes to the development of modern infrastructure, it is possible that the hazard has been underestimated and that existing actions to reduce risk are inadequate considering much new data. The most populous portions of Washington and Oregon lie within seismic zones 2 and 3 of the Uniform Building Code (International Conference of Building Officials, 1991), classifications that have special earthquake resistant design requirements for certain types of structures. However, absence of the largest hypothesized events from the documentary record of the region may weaken the resolve among the contemporary citizens to adapt and enforce hazard reduction techniques. We know, for example, that some locales have not strictly enforced earthquake elements of the building codes (May and Noson, 1986). Lack of experience, debate about the level of hazard, and our scant knowledge regarding fundamental aspects of earthquake occurrence in this region hamper efforts to seek hazard reduction measures.

THE NATURE OF EARTHQUAKE HAZARDS

A large earthquake can cause widespread regional damage, resulting from several different geologic and seismic effects. An earthquake occurs when rocks on opposite sides of a fault move or slip abruptly, in response to stress that exceeds the strength of the fault. In map view, the fault and associated fractures are commonly confined to a narrow zone a few meters to a few kilometers wide and several meters to a thousand or more kilometers long. The downdip extent of the fault, referred to as the fault width, can range from a few meters for a very small earthquake (magnitude < 2) to one hundred kilometers or more for a very large earthquake (magnitude ≥ 7.5). Not all earthquakes occur on faults that produce slip at the earth's surface, a fact that damaging earthquakes of the Pacific Northwest commonly demonstrate (i.e., Weaver and Smith, 1983). Nonetheless, faults with surface expression do exist in this region (plate 1), particularly offshore between the Cascadia thrust fault and the coast, and some of these faults (both onshore and offshore) have the potential to produce earthquakes.

From experience in other regions, faults commonly inflict damage on buildings, roadways, pipelines, or other structures lying within the fault zone with surface expression. Disruption of lifelines by faulting can have significant consequences (Earthquake Engineering Research Institute, 1986). Young mappable faults intersect the land surface in this region in several locales and geophysical data suggest that many such faults may exist that are obscured by sediments (i.e., Finn, 1990; Harding and others, 1988; Yount and Gower, 1991).

The destruction from an earthquake, however, usually extends much beyond the rupture zone. Damaging energy, in the form of seismic waves, propagates away from the fault to distances as great as several hundred kilometers in a large earthquake. The strong shaking that occurs when the energy reaches the earth's surface is responsible for the largest dollar and life loss in most earthquakes (figs. 2 and 3). Although shaking levels generally decrease with distance from the rupture because of the natural attenuation of the Earth and geometric spreading of seismic waves, levels of ground shaking can increase at sites underlain by soft soils or alluvium and/or deep sedimentary basins. This effect is commonly termed site amplification or simply the site effect. This phenomenon, which may be enhanced at distant sites by seismic wave reflections from deep layers, was clearly demonstrated in several recent earthquakes, such as the 1985 Michoacan, Mexico, 1985 Chile, and 1989 Loma Prieta, California, events (Algermissen and others, 1985; Singh and others, 1988; Theil, 1990). In the 1985 Michoacan earthquake, collapsed or damaged buildings in Mexico City were nearly 400 km from the source. Soil amplification was a factor affecting shaking in the 1949 and 1965 Washington earthquakes (Algermissen and Harding, 1965; Mullineaux and others, 1967; U.S. Geological Survey, 1975) and is expected to be a factor during future earthquakes in the Pacific Northwest (Silva, this volume; King and others, this volume; Madin, this volume).

Strong or prolonged shaking also produces other damaging geologic effects such as landslides (fig. 4), liquefaction, and lateral spreading. In some earthquakes, landslides and soil failure may produce greater damage and lifeloss than ground shaking, particularly for urban areas in mountainous terrain. Noson and others (1988) note that as many as 14 earthquakes have caused landslides in the state of Washington between 1872 and 1980. The 1949 and 1965 earthquakes triggered at least 105 separate landslides (Chleborad and Schuster, 1989). Potentially disastrous prehistoric landslides are known to have occurred, such as the Osceola mud flow that extended from the slopes of Mount Rainier down the White and Puyallup rivers as far as the present-day location of Auburn and Tacoma (Crandell, 1971). Earthquake shaking may have induced some prehistoric landslides. For example, about 1100 years ago, large-scale landsliding submerged blocks along the shore of Lake Washington (Jacoby and Williams, 1990), timing that is approximately coincident with other regional geologic events that may be earthquake related.

Liquefaction is a phenomenon wherein soils of certain types loose strength owing to shaking-induced flow of water from depth towards the surface; lateral spreading is a related behavior of soils that results in permanent horizontal displacement of the ground. Both phenomena can be destructive to facilities that lie in ground failure zones. Earthquakes have produced these effects in the Pacific Northwest and may have caused as much as 25 percent of the damage in the 1949 and 1965 earthquakes (Grant, 1989).

Ground failure will occur in the future in the Pacific Northwest, induced by both rainfall and earthquake shaking. Schuster and Chleborad (1989), Chleborad and Schuster (this volume), and Grant and others (this volume) among others have discussed the locations of past landslides and the areas susceptible to future landsliding.

Earthquakes and landslides offshore can produce large sea-wave trains or tsunamis (see McCulloch, 1985 for a review). Tsunamis travel away from the source area at speeds up to 800 km/hour (500 mi/hour) (McCulloch, 1985). Their height is low in the open ocean, but can reach tens of meters on approaching shore, causing flooding of low-lying areas as much as thousands of miles from the source. The outer coasts of Washington, Oregon, and California are not only susceptible to tsunamis generated by earthquakes rupturing the Cascadia subduction zone (Hebenstreit, 1988), but are also susceptible to smaller waves from distant earthquakes along the Aleutian and Japanese trenches (McCulloch, 1985). The tsunami from the 1964 Alaska earthquake damaged the Alaskan, British Columbian, Washington, Oregon, and northern California coasts (including 13 killed and $11 million in damage, primarily at Crescent City, California (McCulloch, 1985)). Sand deposits in coastal Washington and Oregon suggest that tsunamis accompanied several large earthquakes on the Cascadia subduction zone in the last 5,000 yrs (i.e., Atwater, 1987; Atwater, this volume; Bourgeois and Reinhart, 1988; Darienzo and Peterson, 1990; Peterson and Darienzo, this volume). Bucknam and Barnhard (1989) found evidence for recent uplift of 7 m at Restoration Point in Puget Sound. If an event of this type occurs suddenly, it is likely that a tsunami is generated. Permanent modification of the shoreline can also be expected. Some evidence does exist that a tsunami occurred in Puget Sound in association with this uplift (Atwater, personal comm. 1992).

Finally, damaging seiches can occur in the Pacific Northwest. A seiche is analogous to resonant sloshing of water in a bucket that has been disturbed. These waves, which are observed in lakes and other bodies of water (which must be mostly enclosed), can produce considerable damage and flooding along shorelines. Seiches are commonly induced by the long-period ground shaking that accompanies earthquakes, but can also be induced by submarine or shoreline ground failure, sudden uplift or subsidence, or sudden tilt (McCulloch, 1985). Past earthquakes have caused seiches in Lake Washington and Lake Union in Seattle, as well as at other lakes in Washington and Oregon (see Thorsen, 1988, for a summary). It is reasonable to question whether the seismic- or non-seismic-induced collapse of a large delta such as the Skagit Delta could produce a damaging water wave (Finn and others, 1989; McCulloch, 1985).

PACIFIC NORTHWEST EARTHQUAKES: ORIGIN, LOCATION, SIZE, AND TIMING

Most earthquakes in the Pacific Northwest are either directly or indirectly related to the interaction of the crustal plate system of the region, which includes the Pacific, Juan de Fuca, Gorda, Explorer, and North America plates (fig. 1b) (i.e., Atwater, 1970; Fox and Engebretson, 1983; Spence, 1989). Much evidence suggests that these brittle plates converge obliquely along the 1200 km-long Cascadia subduction zone, causing the oceanic plates to descend, or subduct, beneath the continent (fig. 5a,b,c). This process can lead to several types of earthquakes at disparate geographic locations not only offshore, but throughout much of the western regions of Washington, Oregon, northern California, and British Columbia (figs. 6 and 7). We classify the earthquakes most likely to affect population centers in this subduction zone into four principal types based on fault location (fig. 5c): 1.) shallow crustal earthquakes in the North America plate; 2.) earthquakes that initiate on faults within the subducted oceanic lithosphere below the thrust fault; at Cascadia these events partly define the dipping Benioff zone and commonly extend from the deepest section of the locked thrust to as much as 80 km depth; (3) thrust earthquakes on the main boundary fault between the oceanic and continental plates; these earthquakes are part of the Benioff zone in some subduction zones; and (4) shallow earthquakes within the oceanic plates or along their margins. Earthquake types that nucleate within plates are termed intraplate events, and those along plate boundaries are termed interplate events. All except earthquake type (3) have been observed historically in the Pacific Northwest. Evidence for type (3), however, may exist in the geological record (i.e., Atwater, this volume; Atwater and others, 1991; Peterson and Darienzo, this volume).

In this volume, we limit discussion of some types of earthquakes because they are of less importance in seismic hazard evaluation or are outside the study area. For example, significant earthquakes are likely east of the Cascade ranges, which could affect urban areas on both sides of the Cascades. The largest earthquake in Washington and Oregon's documentary history (M $\approx$ 7-7.4; Milne, 1956; Hopper and others, 1982; Malone and Bor, 1979; table 1) in 1872 must be noted in this discussion, although it probably occurred outside our study region near Lake Chelan. This event is likely to have been a crustal earthquake (Algermissen, 1983), but no surface expression of the fault has yet been found (Shannon and Wilson, 1977). As recent experience in California demonstrates, however, shallow earthquakes on some faults such as blind thrusts do not always rupture to the surface. Many young crustal faults exist east of the Cascades (Plate 1), and some of these faults may be capable of generating ground motions that would affect urban areas on both sides of the Cascades (i.e., Peity and others, 1990; Hawkins and others, 1989a). Although such faults are not the focus of this volume and we do not discuss them further, they do contribute to the overall hazard of a region that includes the study area.

We also largely ignore the earthquake potential along the spreading ridges because, for the most part, these events are too distant from urban areas to cause widespread damage. Earthquakes along the fracture zones, such as the Blanco and Mendocino fracture zones, or within the continental and oceanic plates offshore, have produced significant damage in the past. The 1980 earthquake (Table 1), for example, was an event of this type that injured 6 people, caused a bridge collapse, and produced maximum MMI VII near Eureka, California (U.S. Geological Survey and National Oceanic and Atmospheric Administration, 1982). Three earthquakes, ranging in magnitude from 6.0 to 7.1 occurred offshore of northern California in 1991 (U.S. Geological Survey, 1991). The smallest of these events, which was near the coast, produced maximum MMI VII in coastal northern California. If the M=7.1, 1991 event had occurred close to shore, damage along coastal California and Oregon would have been substantial. While we acknowledge the importance of these offshore structures in assessing the earthquake hazard of this region, we do not focus on them in this volume. In the following sections, we emphasize the potential for damaging earthquakes within the continental crust, within the subducted plate and lithosphere, and on the Cascadia thrust.

CONTINENTAL CRUSTAL EARTHQUAKES

Some parts of the continental crust in the Pacific Northwest are clearly seismically active, as demonstrated by the fact that many historical low-magnitude earthquakes and several large historical earthquakes occurred in the continental plate (fig. 6) (Ludwin, 1989). Low-magnitude seismicity in the continental plate illuminated the north-northwest striking right-lateral St. Helens seismic zone (figs. 6, 8) (Weaver and Smith, 1983). The 1918 (m_b=7.2) and 1946 Vancouver Island (m_b=7.3) earthquakes were also crustal events (Rogers and Hasegawa, 1978; Cassidy and others, 1988). These events may have a second-order relation to subduction. If the thrust fault is locked, stresses produced by subduction or other relative plate motions may be transferred to the continental plate (i.e., Fitch, 1972).

Young faults recently recognized by geologists must have produced shallow earthquakes in recent prehistoric time. Plate 1 presents a map showing faults with known or suspected Quaternary displacements and geophysically inferred structures in the Pacific Northwest (note that the decrease in density of youthful faulting in Oregon at the Oregon–California border apparent in Plate 1 is not a tectonic feature, but rather is an artifact of the greater emphasis on fault mapping in California). Near Puget Sound, the Saddle Mountain East, Saddle Mountain West, and Hood Canal faults are of Holocene age; Dow Mountain and Cushman Valley faults are of Late Quaternary age (Wilson, 1983). Faults within and bounding the Portland Basin in Oregon appear to be seismically active (Yelin and Patton, 1991). Evidence exists for as many as three Holocene slip events on the Little Salmon fault zone in northwestern California and for less frequent and smaller events on the nearby Mad River fault zone (Clarke and Carver, 1989; Carver, 1987; Carver and Burke, 1987b; Carver and others, 1989; Kelsey and Carver, 1988; Clarke and Carver, 1992). Bucknam and Barnhard (1989) observe Holocene uplift at Restoration Point on Bainbridge Island and at Lynch Cove near Belfair, Washington, which may be related to slip on a shallow fault that cuts through post-glacial deposits on the floor of Puget Sound (Yount and Gower, 1991). Deformation on local shallow faults may have produced subsidence at Coos Bay, Oregon (Nelson and Personius, this volume).

Mapping of offshore deposits shows many offshore faults and folds in the continental plate. Snavely (1987) and Snavely and others (1980a) discussed Holocene faults both onshore and on the continental shelf that have the potential to produce destructive earthquakes. Clarke (1990), Clarke and Carver (1992), and Clarke (1992) mapped many offshore fold and thrust structures. The age of these structures is not precisely known. Given that blind thrusts may not rupture through rocks as young as the age of faulting, some faults may be younger than the youngest rock unit penetrated. Where these faults come ashore, in some cases, they have been age-dated as Holocene and Pleistocene (i.e., Clarke and Carver, 1989). Clarke (1990) stated that these offshore "...structures originated or were reactivated as a consequence of deformation accompanying plate convergence during Pliocene to Holocene time." Offshore faults and folds cutting Holocene sediments are also abundant on the continental margin of Washington (Wagner, 1985; Wagner and others, 1986) and in the Strait of Juan de Fuca (Wagner and Tomson, 1987). Finally, Harding and Barnhard (1987) and Harding and others (1988) discussed features in seismic reflection lines that offset or disturb the sediments on the floor of Puget Sound and Lake Washington. In some cases, submarine slumps may produce these offsets, but in other cases the depth penetration and dip of the features are evidence of faulting of Holocene age.

Some of these faults may represent secondary rupture in response to slip on the Cascadia thrust (Clarke and Carver, 1992). The return time for earthquakes on such sympathetic faults could be linked with the return time for great earthquakes on the Cascadia thrust fault. In other cases, however, slip on continental faults may be independent of the Cascadia thrust.

Whether these faults represent secondary or primary rupture, many are long enough to be considered potential sources of large damaging earthquakes. On the basis of historical seismic data, a maximum magnitude of 7.4 is justified, if events such as the 1872 north Cascades earthquake are possible at other locales west of the Cascade Range. A similar value is suggested by the Vancouver earthquakes. The St. Helens zone (Weaver and Smith, 1983) produced an earthquake of M_L 5.5 in 1981. Grant and Weaver (in press) suggest that this fault is capable of an event as large as magnitude 6.8. Some mapped faults in the region may be capable of maximum magnitudes greater than 7.4. For example, based on fault offsets in Holocene sediments on the Little Salmon, McKinleyville, and Mad River fault zones, Clarke and Carver (1989; 1992) and Dengler and others (1991) infer paleoearthquakes associated with these faults having magnitudes ranging between 7.5 to 8+.

The lack of data hampers establishment of a reliable upper limit for the magnitude of shallow continental crust earthquakes, which, in any case, should be determined separately for each fault. Despite the evidence cited, little is presently known about the locations, size, or recurrence interval of earthquakes on shallow crustal faults, and it is likely many shallow faults with the potential to produce large earthquakes have not been recognized. It is not yet possible to state the location of all such events. Even in cases where the location of surface faulting is known, few data are yet available about the timing of past movement. Nevertheless, existing fault data suggest that large shallow earthquakes are possible at some locations west of the Cascades, and the size and location of these faults suggests that the hazard from continental crust earthquakes may be as large as or greater than that from other source types.

EARTHQUAKES WITHIN AND BELOW THE SUBDUCTED PLATE

The Benioff zone is the inclined zone of seismicity defining the seismic portion of the subducting oceanic lithosphere, observed in active subduction zones worldwide. Commonly, this zone of seismicity includes earthquakes on the thrust fault and earthquakes within the subducting oceanic lithosphere below the thrust fault. In the Pacific Northwest, the thrust fault has not produced detectable earthquakes (Ludwin and others, 1991). Thus, in this region, the Benioff zone is defined only by earthquakes below the thrust fault. Large damaging Benioff zone earthquakes have occurred in the past in this region, and similar events will likely occur in the future. Two of the largest historic earthquakes, which occurred in 1949 (m_b=7.1) and 1965 (m_b=6.5) at depths between 50 and 60 km, are Benioff zone events. Ludwin and others (1991) suggest that the 1873 earthquake near the Oregon–California border may have been a Benioff zone earthquake because it lacked aftershocks, a feature that is common to many such events (Astiz and others, 1988). Although the evidence for a Benioff zone in this region was lacking early on (Crosson, 1972), more recently, improved seismic network capabilities, improved location techniques, and an extended period of monitoring have shown that small–to–moderate magnitude earthquake hypocenters in Washington and western British Columbia clearly define a Benioff zone (Ludwin and others, 1991). Evidence is still lacking for a seismically defined Benioff zone beneath Oregon (discussed below). Strong evidence, however, does exist for the presence of a downgoing slab beneath both Oregon, Washington, northern California, and western British Columbia from studies of seismic-wave travel times (i.e., Crosson and Owens, 1987; Langston, 1981a; Lin, 1974; McKenzie and Julian, 1971; Michaelson and Weaver, 1986; Owens and others, 1988; Rasmussen and Humphreys, 1988; Taber and Lewis, 1986; Zervas and Crosson, 1986).

Gravitational pull of the oceanic slab into the earth's mantle is likely the fundamental process that causes Benioff zone earthquakes (Astiz and others, 1988; Spence, 1987). The gravitational force not only causes earthquakes in the subducted slab by extensional faulting, but is also the predominant driving force that loads the thrust fault. The subducted slab can also deform in modes other than by normal faulting. For example, differential extension between faulted segments of the downgoing slab may cause strike–slip movement (fig. 5b) (Baker and Langston, 1987; Taber and Smith, 1985).

Benioff zone earthquakes do not have a well-established upper-bound magnitude. Astiz and others (1988) noted that, worldwide, earthquakes of this type as large as 8 are observed, and this maximum magnitude has been suggested by some scientists (Coppersmith and Youngs, 1990). The probable thickness of the subducted slab at Cascadia may imply that the subducted slab can sustain events no larger than M=7.5 (Washington Public Power Supply System, 1988). To compute the probabilistic seismic hazard of the region, Perkins and others (1980) assumed an upper limit of M=7.9.

THE POTENTIAL FOR GREAT THRUST FAULT EARTHQUAKES IN THE CASCADIA SUBDUCTION ZONE

Both prehistoric and future great earthquakes with magnitudes between 8 and 9.5 have been hypothesized for the Cascadia thrust fault on the basis of a wide variety of evidence (Savage and others, 1981; Heaton and Kanamori, 1984; Heaton and Hartzell, 1986; Heaton and Hartzell, 1987). Scientific controversy exists concerning this hypothesis, however, because no data unequivocally indicate either prehistoric great earthquakes or the preconditions for such an event in the future. Furthermore, the Cascadia thrust fault has not produced detected earthquakes at any magnitude level during the 90–year instrumental record nor are any known to have occurred on the thrust during the 200–yr historical record (Ludwin and others, 1991). The scientific controversy centers on several related questions. First, is the Juan de Fuca plate continuing to subduct beneath North America? Second, is subduction, if ongoing, seismic or aseismic? Third, if seismic subduction is possible, how large can the earthquake be, and how likely is it?

In the following, we discuss potential for great earthquakes on the Cascadia plate boundary and some aspects of the debate on this topic. The discussion is an outline intended to review and cite the most significant arguments for and against active seismic subduction. Tables A1 and A2 (Appendix A) summarize the text discussion that follows in the next several sections. The discussion and citations are not intended to be all inclusive; readers can find a greater level of detail and other arguments peripheral to this topic in the references cited.

SUBDUCTION: ACTIVE OR INACTIVE?

Geophysical Data

Sea–floor magnetic lineaments are the principal evidence for active subduction over the last several million years. These lineaments are the remnants of the magnetic field of the earth and ancient reversals in this field recorded by ocean–bottom rocks as they cool after ejection from the oceanic ridges (Raff and Mason, 1961; Vine and Matthews, 1963). These features and other geophysical and geological data show that the Juan de Fuca ridge is spreading at a rate of about 3 cm/yr in a east–southeasterly direction with respect to the Pacific plate, whereas the North America plate is moving at about 5.8 cm/yr in a southeasterly direction with respect to the Pacific plate (Atwater, 1970; Nishimura and others, 1984; Riddihough, 1984). These motions resolve into east-northeast–west-southwest directed relative convergence (fig. 1b) between the Juan de Fuca plate and the North America plate at about 2.5 to 4.3 cm/yr.

Compelling evidence for active subduction also derives from strain–rate (see the Glossary) measurements in Washington. The direction of maximum contraction inferred from geodetic measurements of strain (N59^{0}E±6.6^0: Olympic Peninsula; N68^{0}E: Seattle (Savage and others, 1991); E3^{0}S: western Oregon (P. Vincent, written comm., 1991) is subparallel with the convergence direction (N68^{0}E) determined from plate motions and, thus, is consistent with active subduction. Variable measurements from Vancouver Island, B.C. area (N2^{0}E±13^0: Queen Charlotte Strait; N18^{0}E±11^0: Port Alberni, (Dragert, 1991)) suggest complexity in the subduction process in the Vancouver Island region, but are consistent in magnitude with a locked thrust fault model (Dragert, 1991).

Tide gauge and leveling data indicate gradual uplift of coastal areas in both Washington and Oregon. Savage and others (1991) showed that this observation is consistent with ongoing shortening of the North America plate (fig. 5b).

Several investigators (Silver, 1972; Snavely and others, 1980b; Snavely, 1987, McInelly and Kelsey, 1990, Clarke, 1992, Kelsey, 1990, among others) concluded, from a variety of offshore geophysical data, that the outer continental margin and coastal regions are undergoing active youthful (Holocene at some locales) folding and faulting consistent with subduction of the Juan de Fuca plate. The lateral extent of the deformation suggests that ongoing subduction also includes the Explorer (Riddihough, 1978; Hyndman and others, 1979; Hyndman and others, 1990) and Gorda sub–plates (Kelsey and Carver, 1988).

Seismicity

One of the observations leading to controversy is that no great earthquakes have been observed during the past 200 yrs on the Cascadia thrust fault (Rogers, 1988a), and the fault zone where such events are expected to initiate is seismically quiet, even at the lowest detectable magnitude levels (Crosson, 1972; Crosson, 1983; Ludwin and others, 1992). Earthquakes are also lacking over large regions of the offshore continental plate in spite of the youthful faulting and folding of offshore sediments noted above. Seismicity is also absent at the Juan de Fuca ridge and over large sections of the oceanic plate west of the trench (fig. 6), although this observation is common to many subduction zones worldwide (Am. Assoc. Petroleum Geol., 1981). Seismicity does occur, however, along the offshore fracture zones and within the Gorda south plate. We discuss these dilemmas in the following paragraphs.

The lack of historic seismicity in the subduction thrust zone has been interpreted in several ways: (1) the subduction process has stopped; that is, there is no longer any closing motion between the North America and the Juan de Fuca plates or slab pull is insufficient to initiate slip on the thrust fault (i.e., Crosson, 1972; Farrar and Dixon, 1980); (2) the plates are unlocked and the subduction process proceeds without generating earthquakes (i.e., Ando and Balazs, 1979); and (3) the two plates are strongly locked, but converging nonetheless; this condition would lead to compression and bending of the continental plate, and storing of elastic energy, to be released in a future great earthquake (i.e., Savage and others, 1981; Heaton and Kanamori, 1984). In the latter case, the historical period is presumably part of the seismic cycle between great earthquakes.

In the simplest terms, the seismic cycle refers to the buildup and release of stress on faults that results in alternating periods of quiescence and seismicity. The relationship between this cycle and earthquake occurrence is not yet completely clear (Kanamori, 1981). In one model of the seismic cycle, however, motion of crustal plates produces strain accumulation resulting in several seismicity phases. At low strain levels, few earthquakes may occur at a low or background rate. As strain accumulates, the rate of earthquake occurrence commonly increases throughout the strained volume of crustal rock. At larger strain levels, some faults or fault segments lock and become quiescent again, although seismicity may continue on surrounding faults. As strain increases further, a sequence of foreshocks may develop within the formerly quiet zone, temporally and spatially near the impending mainshock. When strain reaches a critical level on a given fault, the mainshock earthquake occurs followed by a series of aftershocks, after which the cycle repeats. If the rate of long–term strain accumulation is constant and the fault zone properties are uniform, the time between mainshocks is dependent on the amount of slip in the last mainshock (Bufe and others, 1977). Under these conditions, following this simplistic model, an earthquake with large slip will produce a longer wait time until the next event than an earthquake with small slip.

This model partially stems from the absence of seismicity on known active faults, a commonly observed behavior (Lay and others, 1982). Rogers (1988a) cites the subduction zone in New Zealand near Wellington as an example of a fault that produced a large subduction zone earthquake in the last century having no detectable present–day seismicity on the thrust fault. Seismic gaps along other subduction zones may forecast the potential for future large earthquakes (i.e., Kelleher, 1970; McCann and others, 1980). The San Andreas fault in California, although not a thrust fault, provides another example of this behavior. As noted by Heaton and Kanamori (1984) the section of the fault that ruptured in 1906 is now seismically quiet, whereas the creeping section that has not produced a large earthquake historically is seismically active at low magnitude levels. Thus, we can neither take the existence of small earthquakes as conclusive evidence of the potential for large events nor the absence of small earthquakes as evidence of a fault with no potential for large events (Lomnitz and Nava, 1983). Nevertheless, based on these examples, the Cascadia thrust fault may be in a quiescent period of the seismic cycle, given other data suggesting that plate motions continue.

For example, seismicity occurs on some boundaries of the regional plate system that includes the boundaries between the North America, Juan de Fuca, Gorda, and Explorer plates. Because the Mendocino, Blanco, and Nootka fracture zones (fig. 6) exhibit seismic activity, it is possible to determine if slip rates on these boundaries inferred from earthquake magnitudes are consistent with independently determined plate motion rates. Hyndman and Weichert (1983) showed that slip rates calculated from the magnitudes of earthquakes on the plate boundary faults accord with slip rates based on global plate kinematics. Their result, however, did not resolve which plates are in motion, but only showed that relative motion occurs across these faults at rates consistent with plate motion reconstructions.

Examination of slip directions and stresses derived from earthquake focal mechanisms provides another means of observing relative crustal plate motions, but these observations are equivocal regarding active subduction. Focal mechanisms and seismicity alignments indicate right–lateral strike slip on the Mendocino and Blanco fault zones (McEvilly, 1968; Spence, 1989). Right–lateral slip on these faults is consistent with either transport of the Juan de Fuca plate into the trench, a fixed Juan de Fuca plate and northwest transport of the Pacific plate, or motion of both plates. Only the first and third alternatives are wholly consistent with active subduction. As noted in the following paragraphs, the second alternative may or may not be consistent with active subduction, depending on other assumptions.

Furthermore, the greatest compressive stress direction inferred from focal mechanisms is principally north–south in the Pacific Northwest (Spence, 1989). If convergence of the two plates is ongoing and the thrust fault is locked, a more east–northeast compressive stress direction should be observed, parallel to the convergence direction. This expectation is contrary to most of the focal mechanism observations.

Spence (1989) computes stress directions and magnitudes throughout the Juan de Fuca plate and portions of Washington and Oregon based on a model that only requires motion of the Pacific and North America plates and a locked subduction zone. The model assumes no motion of the Juan de Fuca plate. The orientation of the greatest compressive stress computed using this model roughly fits the focal–mechanism inferred direction of greatest compression. The agreement between modeling and data implies that the Juan de Fuca plate west of the trench is not presently in subducting compression, although the modeling does not preclude subduction in the future (see additional discussion of this modeling below, in the section on "Strain rate measurements/crustal shortening and focal mechanisms").

Seismic evidence for motion of the subducted slab below the locked zone strengthens the argument for active subduction. Taber and Smith (1985), who studied the focal mechanisms for small earthquakes in the Benioff zone (in locations similar to earthquakes within the subducting lithosphere shown in fig. 7b), found that these events were predominantly normal–fault earthquakes with downdip extensional axes. Both the 1965 and 1949 earthquakes also exhibit downdip extension axes (Weaver and Baker, 1988). Extension in the Benioff zone is consistent with active subduction because the slab in the Benioff zone is being pulled into the mantle by gravitational forces. Spence (1989) argued that the slab–pull force extends the downgoing lithosphere producing normal–fault earthquakes with downdip extension axes because the locked thrust fault pins and resists downdip slab forces. If the thrust fault is unlocked, experience in other subduction zones suggests that an ongoing series of normal–fault earthquakes should occur out to the trench and farther west (Spence, 1989), as the slab pull force causes bending of the oceanic lithosphere. If the thrust fault undergoes a seismic cycle, normal–fault earthquakes near the trench are only expected after a subduction-zone great earthquake permits an extensional stress pulse to propagate updip past the momentarily unlocked fault. At Cascadia, normal–fault events do not extend to the trench, although they do extend beneath the Olympic peninsula (Taber and

Smith, 1985), suggesting that a portion of the thrust fault is locked.

The 1949 earthquake focal mechanism is strike slip rather than normal faulting. From ground-motion modeling studies, Baker and Langston (1987) inferred left-lateral strike slip on an east-west striking fault for this event. This slip style can be produced by differential motion between different segments of the downgoing slab (fig. 5b). Thus, an east-west tear in the slab could result in strike slip on this fault (Baker and Langston, 1987), if the southern segment of the slab moved downdip relative to the northern segment. In any case, this event suggests active subduction.

Other Views and Complexities

The interpretation of active subduction, however, is not incontrovertible. In earlier studies, Crosson (1972), Milne and others (1978), and Sbar (1983) noted the lack of a distinct Benioff zone and an earthquake-inferred greatest-horizontal-compressive stress orientation different from the convergence direction. From these data they argued that subduction may have ceased. Over time, however, new data on the seismicity of the region accumulated that improved the definition of the Benioff zone beneath Washington (Crosson, 1983; Crosson and Owens, 1987; Ludwin and others, 1992). Yet, the lack of a distinct Benioff zone beneath Oregon (Weaver and Michaelson, 1985) has not been adequately explained.

Bolt and others (1968) discussed seismic and geologic evidence that the Gorda basin and North America plates form a single plate owing to the penetration of the San Andreas fault to the northwest beyond the Mendocino fracture zone. This hypothesis implies that a southern section of the Cascadia subduction zone is permanently locked.

Other arguments have been given to suggest that subduction has ceased. The pattern of ocean-bottom magnetic reversals indicates that the ocean-ridge spreading rate decreased during the last 6 million yrs (Carlson, 1982; Riddihough, 1977; Riddihough, 1984; Duncan and McElwee, 1984). Unfortunately, the last age for which active spreading can be calculated is about 700,000 yrs before present; thus, the magnetic lineament data provide no constraint on the ridge spreading rate to the present day. Furthermore, at present, the Juan de Fuca ridge is nearly aseismic, a fact that may imply spreading has ceased. If spreading has slowed and subduction rates are low, the thrust fault could have permanently locked during the last 700,000 yrs (i.e., Atwater, 1970).

However, other evidence or arguments counter the ceased subduction hypothesis. For example, seismic rates are commonly low along many actively spreading ridges worldwide (Am. Assoc. Petroleum Geologists, 1981). And even though the Juan de Fuca ridge is aseismic, tens of kilometers of new crust have formed in some sections of the Juan de Fuca ridge (Delaney and others, 1981; Heaton and Hartzell, 1987) during the last 700,000 yrs, and heat flow values over the spreading ridge are comparable to those over other active spreading ridges (Korgan and others, 1971).

Some regard the lack of an offshore trench as evidence that subduction has ceased (i.e., Riddihough, 1978). However, high sedimentation rates from the Colombia and Fraser rivers since the last glaciation, a slow subduction rate, and a buoyant oceanic plate may obscure or inhibit development of a prominent trench. Geophysical data support this inference (i.e., Kulm and Peterson, 1984). Finally, Heaton and Hartzell (1986) suggested that the poorly-defined sediment-filled Cascadia trench is a characteristic and consequence of youthful, buoyant, strongly-locked, active subduction zones worldwide.

In our view, the evidence favors ongoing active convergence across the Cascadia thrust fault. Active slab pull could drive subduction and the eventual occurrence of large earthquakes on the thrust fault (Spence, 1987). The most direct evidence for contemporary subduction derives from the offshore deformation of young sediments in the accretionary wedge, strain data that indicate active compression of the continental plate, coastal uplift inferred from leveling and sea level data, and the existence of a Benioff zone having earthquakes with downdip extensional stress axes beneath Washington.

IS THE CASCADIA THRUST FAULT SEISMIC?

Even though active subduction is widely accepted, considerable debate exists about whether or not the thrust fault is seismic or aseismic (Acharya, 1985; Ando and Balazs, 1979; Atwater, 1988a; Byrne and others, 1988; Byrne and Sykes, 1987; Heaton and Kanamori, 1984; Heaton and Kanamori, 1985; Sammis and others, 1988; Spence, 1989; Sykes, 1989; Sykes and Byrne, 1987; Washington Public Power Supply System, 1988; West and McCrumb, 1988a; West and McCrumb, 1988b; Coppersmith and Youngs, 1990). In a recent polling of 14 experts, (Coppersmith and Youngs, 1990) probabilities were elicited about the potential for significant earthquakes ($M_W > 5.0$) on the thrust fault; the responses ranged from a probability near zero to near 1. Although the estimates averaged about 0.5, the assessments displayed bimodal peaks near 0.5 and 0.9. Significant new data suggesting seismic subduction, discussed in this summary, were not known to these experts at the time of their evaluation in 1988.

A locked thrust fault and accumulating stress are prerequisites to large subduction zone earthquakes, although not all agree on this point (Chen and others, 1982; Washington Public Power Supply System, 1988). Nonetheless, the seismic/aseismic debate about the Cascadia thrust fault centers on data that are relevant to whether the fault is locked and accumulating compressional stress. Unfortunately, these data sometimes lead to equivocal inferences. Though assessment of the potential for large earthquakes is not settled, we review the principal aspects of the discussion in the following paragraphs. Table A2 (Appendix A) also outlines the evidence on both sides of the debate.

One question to be addressed is, what constitutes unequivocal evidence that a great earthquake will occur on the Cascadia thrust fault? On the one hand, the only completely unequivocal evidence is probably a great earthquake occurrence. On the other hand, reasonably consistent evidence of several types combined with models that fit these data well using statistical measures could constitute a case providing a high level of confidence in the hypothesis. If this level of confidence is high for a group of observers, then most observers would conclude that such earthquakes will occur. For example, if all the following were true then, in our view, confidence in the hypothesis would be high: (1) consistent times of coastal submergence at widely separated locales, such that only slip on the Cascadia thrust could have produced the effect; (2) evidence of ground shaking in the form of tsunamis, landslides, liquefaction, and turbidites, each having dates consistent with the submergence events; (3) a model of the locked subduction zone that explains the geodetic observations, sea level changes, and paleosubmergence events; and (4) development of models that permit stick–slip over portions of the thrust fault given inferences about temperatures and pore pressures along the thrust fault. Conversely, one might reject the hypothesis because of the following evidence: (1) lack of synchronism among submergence events along the coast and/or between submergence events and shaking evidence; and (2) development of models that explain geodetic data, sea–level data, and rapid coastal submergence in terms of aseismic slip. As will be demonstrated in the following sections, confidence in the hypothesis is stronger than confidence in its rejection. Arguably, the most significant evidence suggests that at least some segments of the Cascadia thrust fault are locked, that subduction continues, and that seismic slip on the thrust fault can be expected in the future (Atwater, this volume; Heaton and Hartzell, 1987, and 1986; Savage and others, 1991; Spence, 1989).

Inferences from Models of Plate and Fault Zone Properties

In one model of the plate–boundary thrust fault (Byrne and others, 1988), the fault is characterized by four elements (fig. 5c), proceeding downdip from the surface trace of the fault: (1) the deformational front, or leading edge of the continental plate, where clay–laden sediments are relatively soft and affected by high–pore pressure; (2) a zone of igneous rock and dewatered, lithified, consolidated sediments, referred to as the backstop, that is strong enough to support stress accumulation; (3) continental crustal rocks; and (4) a thin zone of unconsolidated sediments along the thrust fault that accumulated on the oceanic crust, before their subduction. These unconsolidated sediments form a boundary of variable thickness between the subducted slab and the accreted continental rocks. High fluid pressure and subducted unconsolidated sediments could act to reduce the region over which a locked thrust fault is possible in zones (1) and (2) (Byrne and others, 1988; Byrne and Sykes, 1987; Crosson, 1989; Crosson, 1990; Kulm, 1989; Magee and Zoback, 1989; Sykes and Byrne, 1987). This model is a hypothesis, and many of its features may be difficult to test. For example, the pore pressure and temperatures at the depth of the thrust fault can only be inferred from shallow measurements and modeling, and the behavior of thrust faults under conditions characteristic of subduction is not yet well understood. Finally, some recent geophysical data suggest that sediments are mostly scraped off the downgoing Juan de Fuca plate at some locales and are not subducted (Hyndman and others, 1990). The long–term aseismic behavior of the thrust fault, however, could ultimately validate this working hypothesis.

In zones (2) and (3), temperatures on the fault may exceed 300–450° C (Davis and others, 1990; Hyndman and others, 1990), which would prevent stick–slip motion on this section of the fault (Scheidegger, 1984; Byrne and others, 1988; Sammis and others, 1988). Temperature measurements in shallow sea–floor boreholes and modeling of thermal behavior of the crust form the basis of this hypothesis. Under some conditions, which may apply to the Cascadia fault, these effects could reduce the potential locked fault zone to zero (Byrne and others, 1988; Byrne and Sykes, 1987; Sykes and Byrne, 1987).

The age and temperature of the downgoing slab form the basis of other hypotheses about Cascadia subduction. Subduction of a young relatively hotter oceanic plate may slow subduction because it is more buoyant, reducing the slab–pull force and increasing coupling with the continental plate (Spence, 1989). In this model, large earthquakes may be possible even with a reduced rate of subduction (i.e., Heaton and Kanamori, 1984; Spence, 1989). Contrary to this interpretation, however, others believe evidence exists that a very youthful subducting plate and relatively slow rate of subduction could create a condition of aseismic subduction (Kanamori and Astiz, 1985).

Worldwide Subduction Zone Analogs

To various degrees, each subduction zone is unique. The behavior of one subduction zone does not necessarily unambiguously signal the behavior of another, as noted by Heaton and Hartzell (1986) and Spence (1989). Nonetheless, one means of assessing the likelihood that the Cascadia subduction zone can produce great earthquakes is to compare its characteristics with other active subduction zones worldwide. Heaton and Kanamori (1984), Heaton and Hartzell (1987 and 1986), and Rogers (1988a) found that Cascadia shares many characteristics with subduction zones off southwestern Japan, southern Chile, and Colombia, all zones that have produced great earthquakes historically. Some of these characteristics are: (1) an age and convergence rate that, combined, group this zone with others that have produced great earthquakes; (2) lack of an active back-arc basin; (3) absence of seismicity in the Benioff zone below about 80 km; (4) lack of a well-defined offshore trench; and (5) a shallowly-dipping Benioff zone.

We refer the reader to these studies for the details comparing Cascadia with other subduction zones. We cite one case, the Rivera subduction zone off the coast of Mexico, as an important example having properties closely paralleling Cascadia. Both plates are similar in size and share several other significant characteristics (table 2). The most significant difference is that the Rivera plate may have produced an M_S=8.2 earthquake in 1932 (Singh and others, 1985), yet the Juan de Fuca thrust fault has produced no significant earthquakes in the past 200 years. Unfortunately, this analog is not unambiguous because some suggest that the 1932 earthquake occurred on the Cocos thrust fault to the south of the Rivera plate (Eissler and McNally, 1984; Washington Public Power Supply System, 1988).

Table 2. Comparison of subduction zone characteristics: Juan de Fuca and Rivera plates (Rogers, 1988a)

Characteristic	Juan de Fuca	Rivera
Age of subducting plate (millions of years)	7	10
Convergence rate (cm/yr)	4.5	2.5
Contemporary seismicity on the thrust fault	aseismic	nearly aseismic
Contemporary seismicity on the spreading ridge	nearly aseismic	nearly aseismic

Elevation Changes: Sudden Submergence and Tilt of the Earth's Surface

In active locked subduction zones, the compression of the continental margin leads to secular changes in elevation and shortening of the crust (fig. 5b). Some elevation changes may occur abruptly at the time of a great earthquake. Secular changes in tilt at some locations may also accompany the elevation change. Because of the complexity of elevation and tilt behavior as a function of geographic position, dip of the fault, position of the locked portion of the fault, and the rheological properties of the continental rocks and fault zone, the observations at different locales may appear contradictory, leading to differing interpretations in some cases. Progress is being made, however, in developing models that explain the diverse range of observations.

The strongest evidence for seismic subduction arises from three sources. First, multiple sequences of buried peats in coastal intertidal lowlands occur that appear to be caused by sudden submergence of coastal areas (Atwater, this volume; Atwater, 1987; Peterson and Darienzo, this volume; Grant and McLaren, 1987). Second, thin sand layers overlie some peats suggesting that rapid submergence produced a tsunami. Third, as many as 13 episodes of offshore turbidites occur in the Cascadia Channel, the Astoria Canyon, and at two sites off Cape Blanco (Adams, this volume). Turbidites are slumps of sea floor sediments that liquefy and flow over large distances. Excessive sedimentation or earthquake shaking could trigger turbidite flow (Adams, 1990). The significance of these 13 turbidites is that they all lie above the Mazama ash (a widespread volcanic deposit about 7500 years old (Adams, this volume)), and they issue from Cascadia channel tributaries with widely separated sources—evidence that turbidites occurred at different locations at about the same time. Fourth, landslides or other ground failure events occur in several locales that may temporally correlate with subsidence (Schuster and others, 1992; Williams and Jacoby, 1989; Jacoby and Williams, 1990).

From studies of estuary deposits, evidence in the form of buried peat horizons has been found at about 50 radiocarbon-dated locations showing that coastal lowlands have undergone episodic, rapid submergence (Atwater, 1987; Atwater, this volume; Peterson and Darienzo, this volume). Some of these locations also exhibit 300-yr old tree stumps and snags now surrounded by salt- and brackish-water tidal marshes. Because these trees must have grown above the reach of salt water, it is likely that rising sea level or land subsidence killed these trees. Study of tree–ring growth patterns and the presence of intact grasses in the buried peat deposits, however, suggest that sudden submergence of coastal lowlands most likely killed these organisms (Atwater and Yamaguchi, 1991).

If the dating of submergence events, turbidite deposits, liquefaction, and paleolandslides showed common ages at many locales over a several hundred kilometer north–south section of the subduction zone, the evidence for great subduction zone earthquakes of magnitude 8 or larger might be conclusive. Evidence for ground shaking in the form of liquefaction, landslides, and turbidity deposits that was synchronous with rapid coastal elevation changes would be particularly important data supporting earthquake occurrence. As shown by Nelson (in press a) and Atwater (1992), however, ^{14}C ages show centuries of scatter, making the temporal correlation of different types of events at many sites hard to test. The combined difficulty of obtaining samples representative of the age of a particular subsidence or ground failure event and scatter or ambiguity inherent in age–dating techniques produces this scatter. Of course, such scatter could, in part, also be due to the inadequacy of the earthquake model. Nonetheless, the data suggest a correlation in the time of occurrence of some of these events.

For example, Atwater and others (1991), taking advantage of more accurate age estimates afforded by high–precision ^{14}C techniques and tree–ring data, concluded that sudden submergence in areas 55 km apart was the result of coordinated slip on one fault or one system of faults. Because the subduction thrust fault is the only fault recognized to extend through the entire 55–km long area, Atwater and others (1991) associated the submergence event with a Cascadia thrust–fault earthquake. The high–precision radiocarbon dating fixed this submergence in the interval A.D. 1695 to 1710 (referred to below as time A) (95% confidence interval; 281–296 years before present) (Atwater and others, 1991). Less precise ages indicate that sudden submergence occurred about this time at other locales along the Washington and Oregon coast, suggesting that the 300–yr old event ruptured more than 55 km of the thrust fault.

The record of probable tsunami deposits in estuary sediments is important evidence of prehistoric large earthquakes. Modeling the geographic distribution of sand atop the 300 yr–old buried peat (time A) at Willapa Bay, its landward deposition pattern, and amount of sedimentation, Reinhart and Bourgeois (1989) found that a 10 m tsunami could have produced this sand deposit, and flooding by river discharge was an improbable cause. Atwater and Yamaguchi (1991) argued that the landward thinning of the sand deposits and their coarseness suggest "an extraordinary, landward–directed surge that comprised pulses and approximately coincided with rapid submergence of coast above the Cascadia subduction zone" (see also, Atwater, 1987; Darienzo and Peterson, 1990).

Approximately coincident timing of geologic events older than time A may exist at scattered locales (Williams and Jacoby, 1989). For example, approximate coincidence of several types of geologic events (i.e., sudden submergence, sudden uplift, landsliding, and sand venting) exists at two separate ages, about 600–1700 (time B) and 1400–1900 years ago (time C, possibly 2 events in this age range; Atwater, 1992). (Although Atwater quotes an age range for each event, time B has been referred to as the 1150 YPB (years before present) event, and event C has been referred to as the 1700 YPB event (Williams and Jacoby, 1989)). Lack of coseismic subsidence at Copalis River estuary at time B raises a question concerning whether this event was produced by movement on the plate boundary fault or another fault. Subsidence is observed, however, at other coastal Washington and Oregon sites at time B (Atwater, 1988b; Peterson and Darienzo, this volume; Darienzo and Peterson, 1990; Grant, 1989). The Copalis site does exhibit evidence for vented sand deposits that are approximately coincident with time B, but Atwater (1992, p.1917) attributes this deposit to folding and/or faulting of deep aquifers rather than shaking induced liquefaction. Landslides also occurred at time B and time C at Lake Washington (Williams and Jacoby, 1989; P.L. Williams, personal comm., 1992). Landslides at about time B formed within a 30–km radius of the Saddle Mountain east fault (Schuster and others, written communication, 1992). This fault and others nearby may have also slipped at time B (Wilson, 1983). Uplift at Restoration Point and Lynch Cove occurred at time B (Bucknam and Barnhard, 1989; Bucknam, 1991). These geologic events could have resulted from a subduction zone earthquake, a continental crust earthquake(s), or a temporal clustering of both types of earthquakes.

Geologic data at a few locales suggest that some vertical coastal movements may have interpretations other than great earthquake events. For example, buried estuary peats at some locales can be attributed to localized tectonic processes or sedimentation (Nelson and Personius, this volume; Nelson, in press b). Muhs and others (1990) also noted that local structures can affect the amount and rate of uplift. Although explanation of these data does not always require great subduction zone earthquakes, the data do imply that active deformation of coastal regions is ongoing. This deformation is likely to be in response to active subduction.

Much evidence exists for tilt of the earth's surface in the Pacific Northwest. Geologic evidence exists for long–term eastward tilting of coastal regions (Adams, 1984), and geodetic (i.e., Reilinger and Adams, 1982; Vincent and others, this volume) and sea level data indicate that eastward tilt of these regions continues at present (Adams, 1984). In earlier studies, Ando and Balazs (1979) interpreted the down–to–the–east tilt of the outer coastal zones in Washington and Oregon, the falling sea level at Neah Bay, Astoria, and Crescent City, and rising sea level at Seattle and Friday Harbor with a model of coastal uplift and inland subsidence that was consistent with aseismic subduction. Their model required down–to–the–west preseismic tilt and subsidence of the leading edge of the continental plate produced by aseismic slip along the thrust fault. They concluded, based on comparison with data from Japan, that landward tilt is inconsistent with seismic subduction.

In contrast to Ando and Balazs (1979), others interpreted the elevation and tilt data to indicate seismic subduction. For example, Adams (1984) interpretation of these data is for seismic accumulation of inelastic deformation near the plate margin and elastic deformation farther inland, consistent with a locked thrust fault. Savage (1983), Weaver and Smith (1983), and Darienzo and Peterson (1990) suggested that some regions of a coast may undergo preseismic subsidence while other coastal zones rise (fig. 5b). The behavior of a particular locale depends, at least in part, on its location with respect to the crest of the uplift, the geometry of the thrust fault, and the position of the locked zone (Atwater, 1988a; West and McCrumb, 1988b). Savage and others' (1991) successful modeling of strain and elevation data was based on active subduction and a locked thrust fault (further discussion below). Vincent and others (1989) and P. Vincent (written comm., 1991), using a combination of north–south and east–west leveling data, also concluded that these data support subduction with strong interplate coupling.

On the coasts adjacent to some subduction zones worldwide, ancient beach erosional features, termed marine terraces, are uplifted when large thrust earthquakes occur. The scarcity of uplifted Holocene marine terraces along the coasts of Oregon and Washington (Lajoie and others, 1982) has been used as an argument that the Cascadia subduction zone is not seismic (West and McCrumb, 1988a; Sykes, 1989). Late Quaternary terraces do exist at some Pacific Northwest locales, such as at Cape Blanco, Oregon (Kelsey, 1990) and the Cape Arago–Bandon region (McInelly and Kelsey, 1990). Although Kelsey (1990) finds evidence of a Holocene uplifted beach berm at Cape Blanco, the five terraces at Cape Blanco and Cape Arago are of Late Quaternary age and not Holocene. The terraces were deformed and tilted, and the berm was uplifted as a result of the formation of local folds during the Holocene at the two locations. These features may be coseismic, secondary deformation related to subduction, as suggested by McInelly and Kelsey (1990) and Kelsey (1990).

Terraces are not always present at locales having historic great earthquakes (for instance, southern Chile; Heaton and Hartzell, 1986). Other factors such as high coastal erosion rates in the Pacific Northwest could have obscured terraces (Heaton and Hartzell, 1987). Furthermore, the position of the crest of the uplifted zone at Cascadia may be such that rapid submergence of the coast occurs at some locales at the time of the earthquake, whereas uplift is required to produce marine terraces.

Evidence for past and present changes in elevation and land tilt from geologic and geodetic data strongly suggest, but do not prove, that multiple prehistoric earthquakes alternately gradually uplifted some coastal areas between earthquakes and abruptly submerged the same zones when each earthquake occurred. That measurable ongoing uplift and tilt can be modeled by convergence of a locked thrust fault strongly suggests, but does not prove that such events will repeat.

Strain Rate Measurements/Crustal Shortening and Focal Mechanisms

From stress orientation and strain–rate measurements, it is possible to infer the state of coupling between the Juan de Fuca and North America plates. High strain rates and a greatest compressive–stress direction parallel to the plate convergence would imply converging plates, a locked thrust fault, and the potential for large earthquakes. On the other hand, low strain rates and/or compressive stress directions not parallel to convergence may imply weak coupling and fewer earthquakes or even aseismic subduction.

Data on stress directions arise primarily from three sources: earthquake focal mechanisms, strain measurements, and borehole breakouts. Earthquake focal mechanisms provide an estimate of the principal stress orientations by fitting models of earthquake ground–motion radiation to observed first–motion directions and amplitudes. The average focal–mechanism stress directions are commonly assumed to indicate the principal stress orientations at earthquake depths. Borehole breakout is a term that refers to elliptical elongation and cracking of the walls of a borehole under stress. The short and long axes of the ellipse align parallel to the greatest– and least–horizontal compressive stresses, respectively. Like strain measurements, borehole breakouts reflect the stress orientation near the surface.

Savage and others (1991) (see also Savage and others, 1981; Savage, 1989; Savage and Lisowski, 1991) found that, although strain rates are low (lower in fact than indicated by earlier measurements), the strain–rate data from both Puget Sound and the Olympic Peninsula were predicted by a model of the thrust fault locked offshore and convergence between the Juan de Fuca and North America plate of about 4 cm/yr. This model predicted low strain rates and fitted the observed uplift rates in western Washington determined from tide gages. Savage's success in fitting this variety of data is especially significant because parameters used in the model were determined independently, rather than fitting the observations by parameter variation. Still, the observation of low strain rates, the limited geographic extent of strain measurements, and the possibility that permanent deformation of the Olympic peninsula produces the observed strain effects precludes unequivocal interpretation of a locked subduction zone.

Other interpretations of the strain data also suggest that the two plates are actively converging and locked. Melosh (1987; 1989), using finite element techniques, modeled uplift and strain data and concluded that a locked thrust fault and plate convergence fit the data best.

Other modeling studies support this conclusion. For example, analog stress modeling (Fox and Engebretson, 1983) of a locked thrust fault predicted some aspects of the young geologic deformation history for the past 7 my.

On the other hand, Crosson (1986; 1990) and Ma and others, (this volume) interpreted observations of low horizontal strain rate, episodic strain rate, and north–south compression in the Puget Sound region as arguments for aseismic subduction. Crosson (1986) also argued that strain–rate error bars are comparable to the observations themselves and, therefore, can not be used to unequivocally support a locked subduction zone.

Other apparent conflicts arise in the comparison of stress and strain–rate principal stress directions. The direction of compressive stress inferred from focal mechanisms differs substantially from that inferred from strain-rate measurements. The strain–inferred direction of greatest compression approximately parallels the direction of convergence between the Juan de Fuca and North America plates. In contrast, the greatest compressive stress direction inferred from earthquake focal mechanisms, occurring in either the continental plate or in the Juan de Fuca plate (west of the thrust fault), is commonly perpendicular to the convergence direction. Focal mechanisms for earthquakes in both plates generally support a north–south compressive stress direction that characterizes the interaction between the Pacific and North America plates, as observed along the San Andreas fault in California. This discrepancy has been used to argue against active Cascadia subduction. While the discrepancy between the orientation of focal mechanism stresses and principal strain–rate directions is uncommon among locked subduction zones around the world (Uyeda and Kanamori, 1979), we note one exception to this generality along the Nankai trough. In this case the convergence direction between the Eurasian and Philippine plates is northwest, but the compressive stress direction inferred from focal mechanisms, strain rates, and geologic deformation is east–west to east-southeast–west-northwest (Tabei, 1989). This subduction zone, however, has a documented history of great earthquakes. Furthermore, the apparent difference in stress directions may be found in the fact we are comparing orientations derived from stress (strain) rates with that related to the ambient stress.

This difference was noted by Melosh (1989) who attempted to reconcile the stress direction conflict with a model involving elastic rebound of the continental crust. In the early to middle phases of the earthquake cycle, the continental plate may be in a state of extension because of elastic rebound of the continental plate following the last earthquake (Ando and Balazs, 1979; Spence, 1987). In this phase of the subduction cycle, the stress orientations would reverse, so that the direction of extension is parallel to the direction of convergence, but because convergence continues, a compressional strain *rate* is observed along the direction of extension. As the cycle continues, the rebound is absorbed as the continental plate recompresses. Late in the cycle, earthquake– and strain–inferred compressive–stress directions would coincide, and a thrust earthquake would ultimately occur.

Spence (1989) suggested that a focal–mechanism–inferred compressive–stress direction not parallel to the convergence direction is the result of temporally varying stress–field superposition. Spence argued that stresses reflecting the Pacific and Juan de Fuca plate interaction and slab pull are transferred to the North America plate through a locked thrust fault. In stress–field modeling studies, Spence (1989) assumed that the stress produced by the Pacific–North America plate interaction presently dominates over stresses related to slab pull. Spence's model that best fits observed north–south pressure axes only required motion of the North America and Pacific plates. A passive Juan de Fuca plate was assumed. Subduction was not required, although the model implicitly permits subduction when stresses induced by slab pull, by accumulation of ridge–spreading forces over time, or by overriding of the continental plate become sufficiently great. This model qualitatively accounted for the differences between strain– and focal–mechanism–inferred pressure orientation.

Because of the complex nature of deformation in a subduction zone, extension commonly occurs in the continental crust during active subduction. Evidence exists that east–west extension and north–south compression may be a pattern that persisted during a significant part of the Cenozoic (62–12 Ma) in Oregon and southern Washington. Wells (1990) interpreted clockwise paleomagnetic rotations of forearc, arc, and backarc rocks with a complex model of clockwise block rotations in the arc and forearc regions. Wells inferred that, in Oregon, the rotation produced equal amounts of extension and right–lateral shear, whereas in Washington smaller rotations were produced primarily by right–lateral shear. The orientation of the inferred shear is about northwest. Wells (1990) cited the Mt. St. Helens shear zone as an example of contemporary deformation fitting this model. Observed temporal clockwise rotation of dike– and normal–fault strikes into the Quaternary led Priest (1990) to observe that arc extension may continue to the present and that the direction of least compression is about east–west in arc and forearc rocks. If this deformation persists to the present–day, one can infer that the North America–Pacific plate interaction dominates the contemporary stress pattern, producing oblique subduction and backarc extension (Wells, 1990).

In our view, the evidence largely favors a locked thrust fault with the potential for a great subduction zone earthquake. We acknowledge, however, many counter arguments (besides the references cited see also Washington Public Power Supply System, 1988).

CASCADIA THRUST FAULT: MAXIMUM MAGNITUDE EARTHQUAKE

The question of maximum magnitude (M_{max}, assumed to be moment magnitude, M_w) on the Cascadia fault is of vital concern. Earthquake size or magnitude is one of the most significant factors relating the level, duration, and lateral extent of strong shaking. The length and downdip width of the rupture determine the earthquake size and, thus, the level, duration, and geographic extent of shaking. In the worst–case scenario, the fault rupture zone may extend the north–south length of the Cascadia subduction zone and eastward to the Puget–Willamette trough. Alternatively, if the rupture area is limited in size, the geographic area affected is proportionately reduced. Furthermore, in some scenarios, a limited rupture size may lie farther from urban areas. Because the extent and position of the fault rupture control many aspects of earthquake effects, it is important that the most reliable estimates possible be obtained for this parameter.

The distribution and amount of slip on the fault also control the magnitude, shaking levels, and duration of shaking in an earthquake. This aspect of the earthquake source, however, is difficult to determine before the earthquake occurs. Thus, empirical relations between maximum fault rupture dimensions and magnitude or other empirical data are commonly used to establish the maximum magnitude.

Differences of opinion exist about M_{max} that are fundamentally related to the coupling of the Juan de Fuca and North America plates, the segmentation of the Cascadia subduction zone, and the length and downdip width of the fault likely to rupture in a single event. Using worldwide data, the maximum expected magnitude has been empirically related to the subduction rate and age of the subducting plate (Heaton and Kanamori, 1984). High rates and youthful subducting plates are associated with the largest magnitudes. Heaton and Kanamori (1984) used this observation to empirically predict magnitude 8.3 $\pm$ 0.4 for the Cascadia subduction zone. Noting the similarities between the Cascadia and southern Chile subduction zones, however, Heaton and Hartzell (1986) concluded that rupture of 1000 km of the Cascadia subduction zone could produce an earthquake as large as magnitude 9.5. Both Adams (1990) and Rogers (1988a) computed a maximum magnitude of 9.1 for the Juan de Fuca segment using different assumptions. Rogers (1988a) estimate is based on fault area and rupture of the Cascadia fault on the 900 km–long Juan de Fuca segment from the Nootka fault on the north to the Gorda south plate, assuming a 100 km fault width. The magnitude is related to the fault area by an empirical relation (Wyss, 1979). Inclusion of the Gorda south plate in this event raises the magnitude to 9.2. Adams (1990) assumed a convergence rate of 45 mm/yr and an average recurrence interval of 590 years to compute an average slip of 26 m per earthquake, if all slip is seismic. He further assumed that the fault would rupture 750 km from the Blanco to the Nootka fracture zone with a 100 km width. The magnitude was computed from the definition of earthquake moment and an empirical relation between magnitude and moment. In the following discussion, we review the evidence that could limit the length or downdip width of the fault, and thus, limit the maximum magnitude to values less than that just discussed.

Limits on Maximum Magnitude Related to Fault Segmentation

Segmentation refers to the possibility that the downgoing slab west of the thrust fault is composed of two or more plates separated laterally by fracture zones. Segmentation raises the possibility that the plates subduct independently of one another. Variations in coupling along the fault strike may also lead to earthquakes that rupture limited sections of the fault. The zones of strong coupling are termed asperities, and earthquakes can rupture one or more asperities. This leads to the possibility of a range of possible earthquake sizes, where very large events that rupture through several segments/asperities are the least likely. Commonly, segmented thrust faults rupture only one segment in an earthquake, although cases exist with probable multiple-segment rupture in a single large event (Ando, 1975; Kanamori and McNally, 1982), as in the 1000 km–long 1960 Chile rupture. There are no examples worldwide, however, for the rupture of an entire subduction zone in one earthquake (Spence, 1989). Cases of "fault rupture contagion" (Perkins, 1987) also exist in which rupture of a segment leads to a sequence of ruptures of adjacent segments at time intervals ranging from seconds to centuries (Kelleher, 1970; Mogi, 1974; Ando, 1975). The magnitudes of sequenced events that occur over an extended time period are likely to be smaller than that of a single synchronous rupture of several segments occurring over several minutes. Although segmentation of the Cascadia fault does not necessarily categorically limit the maximum magnitude, it suggests the possibility of a range of earthquake sizes on the fault and, perhaps, less frequent multi-segment great earthquakes. Great earthquakes with magnitudes exceeding M 8.0 are possible, however, with the rupture of individual segments.

On the one hand, some data suggest prehistoric rupture across segment boundaries on the Cascadia fault. For example, 13 post–Mazama–ash turbidite events on the Juan de Fuca and Gorda plates (Adams, 1990) may have been synchronous, which suggests widespread shaking triggered these events. Furthermore, the most recent estuary subsidence events in Washington are approximately synchronous with slip on major onshore faults in northern California (Carver and Burke, 1987a; Carver, 1989). All of these events may be a response to motion on the thrust fault (Clarke and Carver, 1989). If rupture of these and other onshore faults occurred at the same time as abrupt coastal subsidence, the large geographic spacing between the events suggests slip on more than one thrust fault segment. Of course, several alternative interpretations are also possible, including independent rupture of several segments at about the same time.

On the other hand, the evidence for a segmented Cascadia thrust fault appears strong. It is likely, for example, that the Gorda plate on the south and the Explorer plate on the north separated from the Juan de Fuca plate about 4 million yrs before present (Riddihough, 1984). In addition, however, the 800 km–long Juan de Fuca section of the Cascadia fault, which extends from the Nootka fault on the north to the Blanco fracture zone, may be further segmented. Morgan (1968) shows at least 5 faults in the Juan de Fuca plate offsetting magnetic lineaments. Like the Juan de Fuca ridge and the Cascadia thrust fault, however, these faults, if seismically active, are active at very low rates and/or low magnitude levels.

A segment boundary may exist at 46^0N (Magee and Zoback, 1989; Coppersmith and Youngs, 1990). A change in stress directions and uplift rates has been observed between 45^0N and 46^0N that could be related to segmentation (Mitchell and others, 1991; Palmer and others, 1991). From bore hole breakouts, focal mechanisms, and other data, Weaver and Smith (1983), Zoback and Zoback (1989), and Magee and Zoback (1989) inferred that the direction of greatest horizontal stress (A) and the direction of plate convergence (B) is similar in Washington north of latitude 46° N and west of the Cascade Range, indicating a locked thrust fault in that zone. Their data are consistent, however, with a difference between (A) and (B) from south of 46^0 N to the Mendocino fracture zone (see also, Werner and others 1991), which they inferred to indicate weak coupling between the subducting slab and the continental crust south of 46^0 N in Oregon. In contrast to this interpretation, however, Ma and others (this volume) concluded that the data do not require a rotation in the compressive stress orientation south of 46^0 N, although their analysis does not preclude a rotation.

In recent work, sidescan sonar, SeaBeam bathymetry, and single/multichannel seismic records show 3 to 7 left–lateral strike–slip faults off the coast of central and northern Oregon that may be active segment boundaries in the Juan de Fuca plate (Goldfinger and others, this volume; Goldfinger and others, 1990; Kulm and others, 1991; Goldfinger and others, 1992). These faults strike N62^0W to N78^0W subparallel to the Blanco fracture zone, and there is evidence that slip on some of these faults at some locales occurred during the Holocene (Kulm and others, 1991). Although Goldfinger and others (this volume) have inferred a block rotation mechanism for the offshore left–lateral faults, they also note other interpretations are possible. Much remains unclear about the kinematic significance of these faults. Wells and Coe (1985) interpreted older onshore faults in southwestern Washington of similar orientation and slip style (west–northwest and left lateral) as a secondary response (Riedel shears) to clockwise block rotation within a major system of north–northwest striking right–lateral faults.

Portions of the Juan de Fuca plate downdip from the thrust fault may also be segmented (Michaelson and Weaver, 1986), although the evidence on this point is not conclusive (Rasmussen and Humphreys, 1988). From seismicity data, the dip direction of the Juan de Fuca plate appears to vary from northeast under northwestern Washington to about east–southeast under southwestern Washington and northern Oregon (Crosson and Owens, 1987; Weaver and Baker, 1988) (fig. 5b). These data suggest that the dip angle varies as well, from about 15–20^0 north and south of an arch in the plate that extends beneath

Puget Sound, to about 10–12^0 along the arch axis. Recent reflection studies by Hyndman and others (1990) partially confirm this interpretation. In any case, Weaver and Michaelson (1985) interpreted these data as two plate segments separated by a transition zone. In further analyses, Michaelson and Weaver (1986) suggested that the Juan de Fuca plate may be segmented in three sections in Washington and northern Oregon between about 43^0 N and 49^0 N. Based on these data the maximum segment length is about 250 km, and rupture of this segment could produce an earthquake of maximum magnitude 8.25±0.25 (Washington Public Power Supply System, 1988). Rupture of the Gorda south plate segment could produce a maximum magnitude of 8.4 (Clarke and Carver, 1991; Clarke and Carver, 1992). Unfortunately, we do not know whether segmentation of the slab below the thrust implies segmentation of the slab in the locked thrust zone and/or west of the thrust fault. At present, it is also unclear whether plate geometry, such as the bend in the slab, could control the lateral extent of faulting.

Serial rupture of Cascadia fault segments would produce discordant subsidence that might be detected by high–precision dating, if the discordance exceeds a few decades (Atwater and others, 1991). In fact, considerable variability in subsidence event ages is observed, but the variability among ages at individual localities is commonly as large as the variability among localities (Atwater, 1992; Nelson, in press a).

Peterson and Darienzo (this volume) argued that the presence of probable tsunamigenic sands only on the tops of buried peats rather than within the intervening bay–mud layers is inconsistent with segmented ruptures because ruptures of adjacent or even more distant segments would likely produce tsunami deposits interspersed in the peat horizons of the non–rupturing segments. We believe, however, that segmented ruptures with close temporal spacing could also produce sands only at the top of the peats, and the rupture of one segment might not produce tsunamis large enough to breach bay–mouth bars in adjacent or distant segments, inhibiting the deposition of tsunami sands in bay–mud deposits of the adjacent segments.

The issue of segmentation and its bearing on maximum magnitude remains unresolved. While Cascadia fault segmentation is likely, at present, no argument or data preclude rupture of multiple segments. That is, for the present, a maximum rupture length of 800–1200 km for this fault should be considered plausible, but we have little idea of the likelihood of such an event. Conversion of the range of possible segment lengths into expected maximum magnitudes requires other assumptions about the downdip width of the fault and the amount of slip. The maximum expected magnitude even with segmentation, however, ranges between 8 and 9+ (Coppersmith and Youngs, 1990; U.S. Nuclear Regulatory Commission, 1991).

What is the Anticipated Downdip Extent of Rupture?

The width of the fault, namely the downdip extent of rupture and the area over which stick–slip displacement occurs is not well defined, but is an important unresolved issue bearing on the maximum magnitude of the earthquake and the level of strong ground shaking in inland urban areas. The locked section of the fault radiates seismic energy at the time of slip. Several competing views exist concerning the size of the locked section of the thrust fault. The thrust fault could rupture downdip for 100–200 km from the offshore deformation front to the intersection of the thrust fault with the continental crust–mantle boundary at a depth of about 30–40 km (Weaver and Shedlock, 1989). Others suggested that the locked zone may extend from the trench eastward to the continental slope and outer shelf (Davis and others, 1990) or farther eastward to the outer coast (Savage and others, 1981; Weaver and Shedlock, this volume). Some of these models proposed the existence of a locked zone and transition zone (Clarke and Carver, 1991; Savage and others, 1991). In the transition zone, which is assumed to exist downdip from the locked zone, both seismic and aseismic slip are expected. In the extreme view, no section of the fault locks (Byrne and Sykes, 1987). We review these hypotheses in the following paragraphs.

Strain data near Seattle are consistent with a locked thrust fault extending 120 km from the offshore continental margin to about the coast (Savage and others, 1981; Weaver and Smith, 1983; Lisowski and others, 1987). Melosh (1987; 1989) discussed finite element modeling of slip and strain accumulation along the fault that also fits tilt observations and the record of sudden coastal submergence and gradual uplift. In this model the eastern limit of the locked fault is west of the outer coast.

In a study based on additional strain–rate data from the Olympic Peninsula, Savage and Lisowski (1991) fitted both the strain–rate and the geologic data (interseismic uplift rates at the coast and coseismic subsidence) well. This model, which assumed a convergence rate of 4 cm/yr, required a locked fault extending from the offshore trace of the thrust fault about 100 km eastward. Farther eastward, the fault is assumed to be unlocked, but the continental crust in this region accumulates compressional strain. The abrupt slip accompanying the earthquake occurs in the locked zone, but may propagate into the unlocked zone. In this model, the total downdip extent over which seismic radiating slip occurs is 175 km or less.

The zone of seismic slip would be less than 175 km if slip near the updip and downdip tips of the rupture occurred as rapid post-seismic creep (Savage, personal comm.). We infer from this model that the energy radiating section of the fault may extend east of the outer coast well into the Olympic Peninsula.

Arguments have been advanced suggesting that the Cascadia fault is unable to sustain stress over sections of the accretionary wedge and at depths below about 20 km owing to properties of the Juan de Fuca plate and conditions along the fault zone. In these scenarios, plastically deforming soft sediments and high pore pressures in the leading edge of the accretionary wedge limit the western extent of faulting (Sykes and Byrne, 1987; Byrne and Sykes, 1987; Byrne and others, 1988); subduction of significant quantities of young sediments and unusually high temperatures of the subducting slab limit the eastern extent of faulting. The youthfulness of the slab and the covering of sediments may produce a slab temperature that prevents stick-slip (Scheidegger, 1984; Davis and others, 1990). The volume of sediment deposition in the Cascadia trench, produced by the Columbia and Fraser rivers, is greater than that in some active subduction zones considered an analog to Cascadia. These young sediments not only thermally insulate the downgoing oceanic plate, but also insulate the accretionary wedge, raising its temperature and, subsequently, that at the thrust fault boundary. Under such conditions, the shallow section of the thrust fault could slip aseismically. In this case the locked fault width, if a locked section existed, is limited to the zone between the backstop (relatively more competent continental rocks) and the depth at which the fault crosses the 400–450° C isotherm (i.e., the brittle–ductile boundary, fig. 5c). This boundary is estimated to be at a depth of between 20 km (Davis and others, 1990) and 35 km (Byrne and others, 1988). This limited section of the fault zone probably lies beneath the continental slope and the outer shelf. The locked section of the fault is narrower or non-existent, if soft sediments carry below the backstop (Sykes and Byrne, 1987). These factors could significantly limit the size of earthquakes on the thrust fault or, perhaps, preclude their occurrence.

This model raises several issues. First, Scheidegger (1984) shows a zone of 3.4 km/s sediments that thicken eastward and overlie the basalt at one locale 30 km west of the base of the slope. He interprets this high velocity layer as thermally dewatered sediments, which presumably could sustain shear stress. These and other data argue against subduction of significant quantities of soft sediments at some locales. Second, estimation of the depth and map position of the 450° C isotherm on the top of the subducted slab is highly dependent on a number of assumptions, such as accretion and erosion rates, initial thermal gradient, and frictional heating on the thrust fault (Barr and Dahlen, 1989). Finally, the estimated position of the 450° C isotherm computed by Davis and others (1990) applies only to the region off Vancouver Island. Because of the arch in the subducting slab, the map position of this isotherm on the top of the slab is probably well onshore in the model of Crosson and Owens (1987) or near shore in the model of Weaver and Baker (1988).

Differences exist in the position of the locked thrust fault relative to the backstop in various models. The possible existence of multiple backstops and/or differing definitions for this term may be responsible for these differences. For example, in another view, some suggest that the locked section of the fault lies updip from the backstop (i.e., Savage and others, 1991; Wang and Hyndman, 1991; Clarke and Carver, 1991). In these models, the accretionary wedge is aseismic from its leading edge to roughly 20 km downdip. In this zone, the thrust fault also slips aseismically. A locked section lies further downdip that is of variable width as a function of geographic position. Near the Olympic peninsula the locked section may extend 95 km downdip (Wang and Hyndman, 1991); in northern California the locked section may extend 70–80 km downdip (Clarke and Carver, 1991). Downdip from the locked section, a transitional zone may exist that can exhibit both aseismic and seismic slip. This zone is also variable width as a function of position, but may range between an 80 km length near the Olympic Peninsula to 65 km off Oregon (Wang and Hyndman, 1991), or it may extend to the subducted plate knee bend (Clarke and Carver, 1991) (fig. 5c), where the plate dip changes from 11° to about 25°.

Definition of the location and downdip width of the locked zone is an important parameter required to model strain and stress data and secular vertical motion of the coastal regions. Definition of the locked section of the fault, however, may not fully specify the section of the fault that radiates energy at the time of the earthquake (Wang and Hyndman, 1991). The locked section of the fault is expected to radiate seismic energy when rupture occurs; however, if the rupture propagates into unlocked sections of the fault either updip or downdip, it is possible that these sections also radiate energy. Thus, the locked section of the fault may only be a lower bound on the fault dimensions radiating seismic energy. Very little is known about this aspect of subduction zone earthquakes.

Models that assume the rupture zone extends to onshore regions implicitly suggest greater hazard to urban areas. Because these models fit several kinds of data, more weight should be accorded them compared to models that are based on indirect evidence and hypothesis. Nevertheless, specification of the fault width is a matter of current debate not yet resolved.

Continued monitoring of regional strain rates, broader strain network coverage, and refined modeling techniques are likely to produce more certain estimates of the width and length of the locked fault in the future. Research is also needed, however, to better define the relationship between the size of the locked zone and the section of the fault that radiates energy in subduction zone earthquakes.

EARTHQUAKE RECURRENCE

Earthquake recurrence estimates are expressed in several ways. Specification of the precise time, location, and magnitude of the next earthquake is termed earthquake prediction. Unfortunately, at present, methods of earthquake prediction are still being devised and have not been adequately tested. The next level of specification is the calculation of probability of specific earthquake magnitudes on specific faults over a given period. This calculation depends on the historic seismic record, the geologic paleoseismic record of inter–event times, and the time since the last event. This level of specification of earthquake occurrence has been termed an earthquake forecast. At present, official earthquake forecasts are only available for selected faults in California (U.S. Geological Survey, 1988). In the Pacific Northwest, studies are underway to determine paleoseismic inter–event intervals and the time since the last event, data that are the basis of preliminary earthquake forecasts (see below). The lowest level of specification of earthquake recurrence is a mean rate of occurrence for a given magnitude level in a given region. When the logarithm of the mean rate of occurrence is plotted as a function of magnitude, the values form a straight line that is referred to as the Gutenberg–Richter recurrence relationship. In the Pacific Northwest, at present, even this level of specification is difficult because the historic record of earthquakes in the region is too short.

It is worth noting that specification of a mean rate of earthquake occurrence, while useful for some applications, does not necessarily describe the temporal behavior of earthquake occurrence in specific cases. That is, while earthquakes may behave according to the Gutenberg–Richter statistical law for a region that includes multiple faults when averaged over sufficient time, earthquakes in sub-regions or on specific faults commonly recur with considerable irregularity.

The diverse types of earthquake sources in the Pacific Northwest complicate estimation of earthquake recurrence in comparison with most other regions of the U. S. Thrust fault earthquakes, oceanic plate earthquakes, continental crust earthquakes, and intraslab earthquakes are each likely to have different mean recurrence intervals for a given magnitude. Even for a given source type, sub-regional recurrence variations likely exist; these variations, for example, are demonstrated by lower seismic rates in western Oregon relative to western Washington and lower rates within the oceanic plates relative to that along their margins.

Although we have much uncertainty about seismic rates, it is possible to compute estimates for each source type. At present, we must rely on scanty historic seismic data to estimate recurrence rates of continental crust earthquakes. Ongoing paleoseismicity studies of individual surface faults may eventually yield recurrence interval data for crustal prehistoric earthquakes, at least for large events on selected individual faults. Sparse historic earthquakes in the subducted lithosphere form a weak data base for estimation of recurrence rates of Benioff zone earthquakes. Because no historic seismic data exist for thrust–type subduction zone events, estimation of recurrence intervals must rely entirely on plate convergence rate estimates and geologic data, such as the time intervals between rapid subsidence events (Atwater, this volume) or submarine slope–failure events (Adams, this volume). These complexities add to our uncertainty about earthquake recurrence for each source type.

Although the recurrence rates for crustal, Benioff zone, and great subduction earthquakes are not yet well established, estimates have been calculated for several source types and source areas (table 3). The event magnitudes given in table 3 are the magnitudes at which a given author specified the rate. (In some cases, the author also inferred these magnitudes to be the maximum magnitude. Notwithstanding the magnitude values given in this table, in our view, the discussion of maximum magnitude given above seems most plausible.) These return periods give the reader a rough basis for estimating the recurrence of damaging earthquakes for each source type. Large crustal earthquakes ($M \approx 6$) have a mean return period of about 200 yrs (Crosson, 1989) based on seismic data in the Puget Sound region. Note, however, that 2 to 4 crustal earthquakes with magnitudes between 7 and 8 have occurred in the Pacific Northwest since 1833 (1880 Oregon?, 1872 Washington?, 1918 and 1946 Vancouver Island). Study of paleoearthquakes on several individual faults in northern California and southern Oregon indicated the average return period (M=7.6–7.8) on these faults to be about 500 to 600 yrs (Clarke and Carver, 1989; Carver, 1989). From the small seismicity data base, magnitude 7.4 earthquakes in the Benioff zone have about a 200–yr return period (Crosson, 1989).

On the basis of age–dating of buried estuary peats, subduction zone earthquakes occurred on average every 600 yrs for the last 3500 yrs, but with an irregular interseismic interval in detail (i.e., Yeats, 1989; Atwater, this volume; Grant and others, 1989;

Table 3 Return Periods§ (range in years) for Large Earthquakes in the Pacific Northwest (rounded to the nearest hundred) (magnitude at which the rate is estimated)

Data Source	Geologic Intervals	Mean Seismic Intervals		
	A Subduction Events (Shallow)	B Benioff Events	C Crustal (except Subduction)	B+C Events
Atwater (this volume; south coastal Washington)	(300–1200)	—	—	—
Peterson and Darienzo (this volume; central coastal Oregon)	(300–700)	—	—	—
Clarke and Carver 1991, Gorda south plate	(150–550) (M=8.4)			
Heaton and Hartzell (1986)†	(100–250) (M=8.5) (250–500) (M=9.5)			
Adams (this volume); Adams (1984)†	600 (200–500)	—	—	—
Based on Juan de Fuca convergence rates (see text; Rogers, 1988a); (or based on rupture of all Cascadia segments)	(400–1300) (600–2000)	—	—	—
Based on strain–rate measurements (see text; Savage and Lisowski, 1991)	(400–1100) 550	—	—	—
Crosson (1989; Puget Sound)	—	200 (M=7.4)	200 (M=6.1)	—
Rasmussen and others (1974; Puget Sound)	—	—	—	100 (M=7.0)
Perkins and others (1980; Puget Sound)	—	—	—	100 (M=7.0)
Perkins and others (1980; Willamette Basin)	—	—	—	700 (M=7.0)

§This table shows the variability in return period values based on either historic seismicity, geologic, or strain data. Of course, earthquakes may not recur at regular intervals, in which case, return period is a less useful concept. Still the variability in these values and the range about which these measurements center provides a sense of the inter–event times expected in this region based on seismic, geologic, and strain data. †Rates inferred from strain measurements and/or comparison with other subduction zones.

Darienzo and Peterson, this volume). Turbidite occurrence averaged over an interval of about 7500 yrs yielded an average rate remarkably similar to that deduced from buried estuary peats. The uniform thickness of turbidite deposits suggests that subduction earthquakes happen at relatively uniform intervals of about 590±170 yrs (Adams, this volume). On the basis of both data sets, the most recent event occurred 300 yrs before 1990, although earlier dates do not always coincide for turbidites and subsidence events. Approximate or possible correspondence, however, exists for events at 300, 1100–1200, and about 4300 yrs ago (Adams, this volume). The lack of accord between event dates over the entire record for the two data sources may have explanation in their measurement and interpretation. For example, several mechanisms, such as erosion or oxidation, could lead to the removal or masking of some markers in the subsidence record. It is also plausible that some large events do not produce uplift at the coast, as has been observed for some events in Chile. Furthermore, estimation of the date of turbidite deposition, based on the thickness of layers that separate turbidite deposits and assumptions about erosion of this layer, is only approximate (Adams, this volume). Nonetheless, the close agreement between the long–term mean rates of event occurrence and the similarity in some of the age estimates from two independent data sets is compelling, but still equivocal, evidence that the geologic record marks the effects of paleoseismic earthquakes.

Continued refinement of paleoseismic techniques and extension of the data base to older events will help test these results. Further study will also clarify the degree to which errors in radiocarbon dating measurements or other errors bias these results and help determine whether some events are missing in the geological record.

If we know the rate at which the Juan de Fuca and North America plates converge, and if we know or assume the total slip occurring in a great earthquake, it is possible to calculate the return period between events (return period = total slip/slip rate). Rogers (1988a) estimated the mean repeat time for great earthquakes on the Cascadia thrust fault by combining convergence rate estimates with assumptions about maximum slip during great earthquakes and the ratio of seismic slip to total slip. Using this technique, return period estimates range from 400 to 1300 yrs for the Juan de Fuca segment. The recurrence intervals obtained using this method for the Explorer plate (100–300 yrs) and the Gorda plate (70–200 yrs) are small, leading Rogers (1988a) to suggest these segments have the highest probability of rupture.

Inference about the interseismic interval for great earthquakes from strain rate measurements appears to be in rough accord with the geologic data in Washington and Oregon. Forty to one hundred μstrain accumulated before fault rupture in other great earthquakes, independent of magnitude (Sbar, 1983; Adams, 1984; Heaton and Hartzell, 1986). Given the compressional strain rate of 0.092 μstrain/yr (Savage and others, 1991, and a similar strain rate for western Oregon, i.e., 0.08 μstrain/yr (P. Vincent, written comm., 1991), the period between earthquakes is about 400 to 1100 yrs. Repeating the calculation using the 50–μstrain maximum estimated for the great subduction zone earthquakes in Chile in 1960 and Alaska in 1964 (Plafker and Savage, 1970; Plafker, 1972), the calculated return period is about 550 yrs, a value approximating the long–term mean rates deduced from geologic data.

Using current estimates of the mean repeat time, the time since the last earthquake, and the standard error in these estimates, it is possible to forecast the probability of a large earthquake on the Cascadia fault over specified time periods. Adams (this volume) uses the turbidite recurrence to compute a 10% probability of a great earthquake in the next 100 yrs and a 25% probability in the next 200 yrs. He notes that these estimates could be in error by a factor of 2 and does not include the possibility of events clustering in time (Adams, this volume). We consider the values quoted from this study preliminary estimates because of present uncertainties in the underlying assumptions and data.

Note that the time since the last great earthquake, based on geologic studies, is about 300 yrs, whereas the average repeat time is about 600 yrs. If the repeat times are regular, one could infer that the next event is 300 yrs hence. If the repeat times vary between 300 and 1200 yrs, however, it is difficult to infer the time of the next event with statistical significance or practical utility. Although much is known, unfortunately many questions remain unanswered concerning the timing and recurrence of great earthquakes.

LOWER SEISMIC RATES IN OREGON

Compared to western Washington, much of western Oregon exhibits lower seismic activity (i.e., Weaver and Michaelson, 1985). Although a subducted plate is present beneath western Oregon, seismicity defining the Benioff zone is absent, and few earthquakes are detected in the continental crust. The meager number of seismograph stations in Oregon, however, probably reduces our ability to detect events with magnitudes below some magnitude threshold. Nevertheless, it is likely that a difference does exist in the underlying activity rate at magnitude levels where detection is complete. The cause of this difference and it's implication for earthquake hazard in Oregon are important problems that have not been satisfactorily addressed at present. Two simplified hypotheses, similar to those used to explain the lack of seismicity on the thrust fault, may be invoked; 1) the thrust fault is unlocked in the Oregon section and the continental crust and subducted slab are in a low–stress state producing few earthquakes (i.e., Magee and Zoback, 1989); 2) the thrust fault and continental faults are locked, but the seismic cycle is such that few events are presently produced in either the continental crust or the Benioff zone (i.e., Spence, 1989). As with many tectonic issues in this region, existing data do not provide an unequivocal interpretation. The historical aseismicity in western Oregon implies lower short–term hazard compared with western Washington, but the long–term hazard is not necessarily lower. This conclusion would require, for example, definitive evidence that the subduction process is aseismic in this section of the Cascadia fault. At present, we consider the long–term earthquake hazard in Oregon comparable to other sections of Cascadia because the geologic evidence suggests an active Cascadia thrust fault; furthermore, both offshore and onshore youthful faulting and folding suggest active continental earthquake processes.

In detail, several concepts related to these two hypotheses can or have been considered, but some of these ideas raise further dilemmas. For example, Riddihough (1984) suggested that subduction is slowed in the Oregon Cascadia section relative to the Washington section because, including the Gorda north plate, the subducting slab adjacent to southern Oregon is younger than that adjacent to Washington. In this model, a young oceanic slab has more buoyancy and locks more effectively with the continental crust, reducing subduction rates (Kanamori, 1971; Uyeda and Kanamori, 1979; Ruff and Kanamori, 1980; Ruff and Kanamori, 1983). Differences in buoyancy cause the Juan de Fuca plate to pivot about a pole located south of the plate, with slower subduction on the south and faster subduction on the north (Menard, 1978; Riddihough, 1984). In this case slowed subduction on the southern section of the Juan de Fuca plate could reduce strain and seismic rates in the continental crust of Oregon.

Several issues arise in the consideration of Riddihough's (1984) hypothesis. From earthquake locations (Crosson and Owens, 1987) and travel time studies (Green and others, 1986; Michaelson and Weaver, 1986; Owens and others, 1988), an arched subducted slab dips more shallowly beneath northern Washington and has steeper dip to the north and south of this section (fig. 5a,b). If the slab beneath Oregon is more youthful (i.e., more buoyant), one might expect the plate there to be shallower, not steeper, than that beneath Washington, although other factors, such as oblique convergence, varying strike of the Cascadia fault, and plate buoyancy, could combine to shape plate geometry (Crosson and Owens, 1987).

If the subducted plate is arched, coupling of the plates is expected to be greatest where the plate dips more shallowly and to be less in the steeply dipping sections such as that beneath Oregon (Weaver and Michaelson, 1985). Decreased coupling of the thrust fault beneath Oregon differs from Riddihough's (1984) model, but is also consistent with lower seismic rates in western Oregon. An unlocked thrust fault beneath Oregon would not only permit aseismic subduction, but would reduce seismic rates in the continental crust and subducted lithosphere (Weaver and Michaelson, 1985; Rogers, 1985). For example, an unlocked condition would reduce the slab–pull extensional stress necessary to produce Benioff zone earthquakes.

If motion of the Juan de Fuca plate is independent of the Gorda (north and south) and Explorer plates, the buoyant plate hypothesis produces a result diametrically opposed to Riddihough's (1984) conclusion. Rogers (1985) argued that buoyancy was greater on the northern section of the Juan de Fuca plate (i.e., considering the Juan de Fuca plate alone, age increases from north to south). This hypothesis differs from Riddihough's 1984 model, which treated the Juan de Fuca and the Gorda north plate as a single plate. In Rogers' (1985) model, the age difference increases the coupling of the plates on the north compared with the south. In this interpretation, the section of the Cascadia trench off the Oregon coast is moving seaward (referred to as trench roll–back) at about the same rate as the westward component of North America–Juan de Fuca relative plate motion. Rogers (1985) supported this interpretation by noting that partial compressional unloading of the continental crust in Oregon may have permitted extensional style volcanism in western Oregon in the last 2.5 ma. Priest (1990) also argued that east–west extension in Oregon and eastward migration of the volcanic arc during the last 7–4 million years supports compressional unloading. Thus, a difference in the style of tectonism between the northern and southern sections of the continental crust adjacent to Cascadia could also result in differential seismic rates.

Other hypotheses have also been suggested. For example, a complete break in the downgoing slab at depths between 40 and 100 km could reduce slab–pull stress (Michaelson and Weaver, 1986; Weaver and Michaelson, 1985; VanDecar and others 1991). In an extension of Riddihough's (1984) model, the lack of a Benioff zone beneath Oregon might be explained by reduced slab–pull forces owing to the buoyancy of young subducted crust. Spence (1989), for example, argued that a slow subduction rate at Oregon permits the subducted plate temperature to increase. Under this condition, the density of the downgoing slab beneath Oregon decreases relative to that beneath Washington, reducing the slab pull force. Spence used seismic velocity changes in the slab noted by Michaelson and Weaver (1986) to infer slab density.

Neither the lack of slab seismicity nor these models, however, unequivocally demonstrate that large Benioff zone earthquakes (similar to those in Washington in 1949 and 1965) or thrust fault earthquakes are unlikely beneath Oregon.

GROUND–SHAKING HAZARD ESTIMATION

The accurate estimation of the ground–shaking levels in an earthquake requires much information about the earthquake rupture, including its depth, rupture area, orientation, and the slip distribution on the fault. Other factors are equally important, such as the distance between the fault and the location under consideration, the properties of the rock or soil underlying the site (including basin geometry), and the attenuation of shaking that is expected between the fault and the site. Earthquakes occurring in the continental crust, Benioff zone, or on the thrust fault must each be treated separately in shaking estimates because of factors such as differences in earthquake depth, expected maximum magnitude, and various source parameters. Much of the data required to estimate earthquake ground motions may not be available until after an earthquake occurs. Uncertainty about the parameters for each of the source types, including the likely position of ruptures both within and between the North America and Juan de Fuca plates, magnifies the difficulty of computing ground-shaking estimates in the Pacific Northwest. Thus, prediction of shaking levels requires estimation of these parameters from scant geologic, geodetic, and seismic information and earthquake behavior in other locales.

Ground-shaking estimates can take several forms. For example, some maps depict the geographic variation in ground shaking with measures such as Modified Mercalli intensity, peak acceleration, spectral acceleration, spectral velocity, measures of relative motion at sites underlain by soil compared to rock (see the Glossary for a definition of these terms), or some other parameter. Ground-motion maps are useful in the design of ordinary structures, urban planning, earthquake emergency response, zoning, estimation of losses, etc. Methods are also available to estimate these ground-motion parameters for specific sites, a practice that is used in the design of important structures. Site-specific shaking estimates commonly include more detailed information than is feasible for map estimates and, thus, are probably more accurate than map values.

Ground shaking time series (i.e., ground acceleration versus time) and related spectra are commonly estimated from theoretical and hybrid (combinations of empiricism and theory) methods. Estimation of the model parameters is difficult and intensive computations are required with these methods, thus, this type of prediction is most commonly applied in site studies for critical structures. In the following, we review various ground shaking estimates for hypothetical and historical Pacific Northwest earthquakes to give the reader a sense of the expected levels of shaking for various earthquake types and to outline the uncertainties surrounding these estimates.

PROBABILISTIC GROUND-MOTION ESTIMATES

One type of empirically-based ground-motion estimate takes the form of statements about the level of ground shaking that will not be exceeded during a specified exposure period with a specified probability level. Maps of uniform probabilistic ground-shaking hazard are computed for specified site conditions, commonly firm soil or rock. For example, the mapped parameter may be the peak horizontal acceleration, peak velocity, or spectral acceleration having a 10% probability of being exceeded in an exposure period of 50 yrs. The Applied Technology Council (1978) used such maps (updated every three years) for design ground-motion specification included in their proposed seismic regulations for buildings. The National Earthquake Hazards Reduction Program "NEHRP Recommended Provisions for the Development of Seismic Regulations for New Buildings" (Building Seismic Safety Council, 1988) uses similar specifications for ground motion in the new 1988 Uniform Building Code (International Conference of Building Officials, 1991).

Computation of probabilistic hazard maps requires (1) the location, depth, and orientation of active faults, the seismic rate on each fault for the range of earthquakes capable of generating strong shaking, (2) an estimate of the maximum magnitude for each source, (3) the ground-motion attenuation curve as a function of magnitude and distance from the source, and (4) an estimate of each parameter's variability. In lieu of complete information about specific faults, seismic rates are commonly attributed to source zones that may be composed of a few or no specific faults, "background faults" (a uniformly spaced fault grid of specified orientation and maximum magnitude), and uniformly distributed point sources for events of magnitude less than about 6 to 6.5. Seismic rates may be distributed on these elements uniformly or may be apportioned based on seismic observations or geologic constraints. In principal, the maximum magnitude could be defined as the maximum magnitude observed in the source zone or on a specific fault. However, maximum magnitude for some source zones is unknown because the historic period is too short to have observed the largest event. In such cases, the maximum magnitude may be estimated from the length of young faults in the source zone, from comparison of the tectonic setting with similar circumstances in other locales, or assumed to be a value arbitrarily equal to or larger than that observed historically. Commonly, the attenuation function is based on historic strong-motion data, which is predominantly composed of records of shallow earthquakes in California.

Past probabilistic ground-shaking hazard estimates for the Pacific Northwest suggested the hazard is significant. These estimates, such as those by Algermissen and others (1982), Perkins and others (1980), or Algermissen and others (1990) were based on the assumption that damaging future events will recur as they have in the past, that is, as shallow- and intermediate-depth earthquakes, no larger than magnitude 8.2. Past probabilistic estimates did not incorporate great subduction zone earthquakes because evidence for such events was not yet recognized. These probabilistic ground-shaking hazard maps show that levels of shaking that might be exceeded in a 50-yr period with a 10 percent probability of exceedance are comparable to that in the Salt Lake City, Utah, area (table 4) or to that in the moderately seismic areas of California, such as the central valley and sections of northern California (Algermissen and others, 1990).

Algermissen (1988) also produced preliminary probabilistic hazard estimates accounting for potential subduction zone earthquakes. He assumed, based on the paleoseismic data, that a subduction zone earthquake of magnitude 8.5 could recur with an average recurrence interval of either 300, 600, or 900 yrs, a range that roughly spans the values suggested by geologic studies. These calculations did not account for the time since the last event, but assumed a Poissonian earthquake occurrence process. Besides different from the value given in table 4 for Seattle. This result held for all three assumed recurrence intervals. For the deeper rupture extending to Puget Sound, the probabilistic peak acceleration and velocity were about equal to that in table 4 when the 900-yr average recurrence interval was assumed, but the peak velocity was double the table 4 value when the 300-yr recurrence interval was assumed.

More recent unpublished modeling of the subduction zone is in general agreement with the probabilistic mapping results just discussed. The important conclusion is that, if the rupture surface in the subduction zone is more than about 125 km distant from Seattle and, if average recurrence rates of large $M_s \approx 8.3$ earthquakes equal about 500 years, large subduction zone shocks have little effect on the maximum probabilistic acceleration (10% chance of exceedance) at Seattle for periods of interest (exposure times) of up to about 100 years (Algermissen, written comm., 1991).

Recent work (Algermissen, written comm., 1991) also indicated that for an exposure time of 50 years probabilistic spectral accelerations at 0.3 and 1.0 sec (5% damping) are of the order of 85 cm/sec^2 and 40 cm/sec^2, respectively, at Seattle. Spectral accelerations for a 250-year exposure time are about double the 50-year values.

The result that the probabilistic hazard with and

Table 4. Probabilistic hazard estimates: peak horizontal acceleration and peak horizontal velocity with a 10% probability of exceedance in Y years on firm soil

	Exposure Time (Y years)			
	50	250	50	250
Location	Peak Acceleration (%g)		Peak Velocity (cm/s)	
Seattle, Washington	30[1,3]	41[3]–60[1]	17[3]–28[1]	23[3]–52[1]
Portland, Oregon	16[1,3]	30[3]–36[1]	10[3]–15[1]	15[3]–30[1]
Salt Lake City, Utah	29[1]–35[2]	60[1]–70[2]	22[1]	60[1]
San Francisco, California	80[1]	>80[1]	80[1]	>80[1]
Memphis, Tennessee	30[1]	40[1]	20[1]	40[1]

[1]Algermissen and others (1990) (includes parameter uncertainty for some parameters); [2]Youngs and others (1987) (includes parameter uncertainty); [3]Perkins and others (1980) (does not include parameter uncertainty).

the 3 separate recurrence intervals assumed, two fault models were separately applied. In one model, the eastern extent of slip was assumed to be the point at which the fault penetrated a depth of 20 km, which limits its extent to points west of the outer coast. In the second model, the eastern extent of slip was assumed to be the point at which the fault penetrated a depth of 50 km, which permits the rupture to extend as far east as Puget Sound. For the rupture model to 20-km depth, Algermissen (1988) found that the 50-yr exposure-time peak acceleration with a 10% probability of exceedance was not appreciably

without inclusion of large thrust fault events is mostly unchanged for some ground motion parameters arises in part because of the assumed distance of the rupture surface from the urban areas of Washington and Oregon relative to sources in the continental plate and Benioff zone and, in part, because of the infrequency of great subduction zone earthquakes relative to the other earthquake sources. This result does not imply that subduction zone events are an inconsequential component of the earthquake shaking hazard. First, strong shaking in a large subduction zone event will be widespread. While mapped probabilistic ground–motion estimates would demonstrate this fact clearly, point estimates, such as discussed above (Algermissen, 1988), do not confer the geographic extent of shaking. Furthermore, statements about the probabilistic peak–acceleration and to a lesser extent peak–velocity hazard do not adequately represent the hazard owing to low–frequency ground motions of long duration that would be of great importance in assessing the effect of this type of earthquake on tall buildings and bridges. The efficiency of a great earthquake in generating low–frequency motions is also of prime importance in increasing the degree and distribution of landsliding and liquefaction. Second, note that probabilistic hazard calculations presently do not fully account for the effects of site geology, which can significantly increase the level of ground shaking at sites underlain by low–velocity sediments or at sites affected by ground–motion focusing due to basin–geometry. Finally, peak acceleration or velocity computed by this method does not necessarily represent the maximum ground motion expected in each source zone. The maximum expected motion can be estimated by deterministic means (see below) or by using probabilistic ground motions for long exposure periods.

Other limitations of probabilistic ground–motion maps are widely recognized. The calculations are no better than the accuracy of the input parameters used in the models. For example, the return period and expected size for each type of earthquake in the Pacific Northwest is poorly determined, particularly for continental crust earthquakes. Probabilistic maps could understate the hazard for continental crust earthquakes. Although past probabilistic hazard calculations assumed that large continental crust earthquakes were possible, it is increasingly apparent that the number of shallow faults capable of producing large earthquakes and the rate of earthquakes on these faults is presently unknown. Stated another way, given the complexity of the deformation in the Pacific Northwest, it is likely that the historic record does not adequately sample the seismic rate or the modes of deformation, and there is presently no means to entirely account for this fact in probabilistic hazard calculations. On the other hand, McGuire and Barnhard (1981) found in a study of the Chinese earthquake record that the best hazard estimate for the next 50–year interval is the most recent 50–yr period, a result that supports the utility of probabilistic forecasts.

The utility of probabilistic hazard maps stems from the need of users to compare the relative hazard across geographic regions given our current state of knowledge in a way that integrates different historical seismic rates, ground–motion attenuation, maximum magnitude, and tectonic style (National Research Council, 1988). The absolute hazard level predicted by these methods is useful only insofar as the parameters incorporated into such maps are an accurate representation of physical earthquake processes; however, the quality of this knowledge itself varies geographically. As we continue to improve our knowledge of earthquake processes, probabilistic estimates of the ground shaking hazard can serve as a guide in decisions made by designers, code writers, regulators, owners, and public officials. Probabilistic hazard calculations serve to establish the change in risk incurred for various decisions, within the constraints of the models and our present state of knowledge.

For example, probabilistic hazard calculations are commonly used in design decisions and now form the basis for design ground motions in most building codes. Ideally, design levels should vary for buildings of different types. The use of building design motions corresponding to longer exposure periods, for example, is a conservative approach that can be used with critical facilities to mitigate the effects of imperfect return period and time of occurrence estimates for great subduction zone earthquakes. The design level may be higher for schools and hospitals, where the risk for human life loss is higher, than for certain types of low–occupancy low–rise commercial structures. Thus, for a type of structure with low economic value or life–loss risk, a ground motion associated with a 50–yr exposure period (or shorter) may be adequate, while for a structure with high value, life–loss risk, and/or a long projected lifetime, a ground motion associated with a 250–yr exposure period (or longer) may be the design motion required to achieve an acceptably low level of risk. The point is that probabilistic statements about the ground–shaking hazard permit one to assess the relative risk incurred for various design ground–motion decisions.

OTHER EMPIRICAL GROUND–MOTION ESTIMATES

As an example of estimated shaking values in map form, the U.S. Geological Survey (1975) produced maps of expected Modified Mercalli intensity for several scenario Benioff zone earthquakes in Washington. These hypothesized magnitude 7.5 events were assumed to nucleate at 50 km depth at the location of the 1965 event and at a location near Seattle. Modified Mercalli intensities were estimated for a 6–county area that included Seattle, Tacoma and Olympia. The Modified Mercalli predicted intensities in the Seattle area ranged between VI and IX, depending on near–surface geologic conditions. These maps were the basis of casualty and damage estimates in the 6–county region (we summarized the results in the Introduction section of this chapter).

An important method of predicting the ground motion for subduction zone and Benioff zone earthquakes is to fit earthquake observations recorded in other subduction zones with an equation that includes terms for magnitude, geometric and anelastic attenuation, and site conditions. This technique has been used in several studies relevant to the Pacific Northwest (Youngs and others, 1988; Youngs and Coppersmith, 1989; Crouse and others, 1988). Although the data distribution and data properties are generally inadequate to independently determine all the coefficients in an equation expressing each of these factors, commonly one or more of the parameters are fixed by theory or other independent information. Furthermore, ground motions generated by individual subduction zones may differ because of unmodeled parameters, such as dip of the fault, distribution of slip on the fault, and the type of earthquake (i.e., a thrust fault or Benioff zone event). Still, these methods are among the most reliable currently available to predict deterministic ground shaking values. The probabilistic computations also require empirical equations of this type.

Tables 5a,b summarize empirical ground–motion values predicted using the relations developed by Youngs and others (1988) and Youngs and Coppersmith (1989) for both great thrust–fault and Benioff–zone earthquakes. First we discuss the thrust fault case. In Table 5a, in the first scenario, we assume the thrust fault rupture extends eastward as far as the 40 km depth contour on the fault (scenario 1). Some disagreement exists concerning the geographic position of the 40 km contour. Crosson and Owens (1987) placed this contour 35 km west of Seattle, assuming that the thrust fault lies within the seismically quiet zone between the crustal earthquakes and the events within and below the subducted slab (seismicity cross section in fig. 7b is representative of the data). Weaver and Baker (1988) inferred a position 90 km west of Seattle based on the position of the Benioff zone, but assumed the thrust fault lies along the top of the Benioff seismicity. From refraction profiles and other geophysical data, Mooney and Weaver (1989) placed the 40–km contour about 30 km west of Seattle. These contour depths correspond to a 50–100 km range in closest distance to the fault. Closest distance to the fault (R) is the equation parameter required in Youngs' (1988; 1989) formulation. To complete our example calculations, we use the contours of Mooney and Weaver (1989). The shortest distance to the fault is about 50 km to Seattle and 40 km to Portland. In scenario 2, the thrust fault ruptures to the 20–km depth contour on the fault. The shortest distance to the rupture is about 135 km for Portland and about 160 km for Seattle. We assume that portions of the rupturing segment of the Cascadia thrust fault are adjacent to, but west of each city. The magnitude we assume is arbitrary and independent of the implicitly assumed fault length.

The computed results, presented in Table 5a, demonstrate several important points about the ground shaking hazard at Seattle and Portland for thrust fault earthquakes (see the Glossary for a definition of terms used in this table). First, under any assumption about the position of the rupture, an earthquake of this size is expected to produce damaging ground motion at least as far east as the urban areas of the Puget–Willamette trough. Second, the values expected at Portland are marginally higher than those expected at Seattle. Third, the values at Seattle or Portland under scenario 2 could be roughly twice the values under scenario 1. Fourth, based on Youngs (1988) equations, the ground–motion values on soil are as much as twice the values on rock. Other data, however, suggest that rock–soil site-spectral differences can be larger than shown in Table 5 for some site types at some periods (i.e., Algermissen and others, 1985; Rogers and others, 1991; Theil, 1990).

Table 5a. Empirically Predicted Ground Motions† for Seattle or Portland for a M=8.5 Thrust Fault Rupture Based on the Relations of Youngs and others (1988) and Youngs and Coppersmith (1989).

	M=8.5 Thrust Fault Rupture							
	Rock		Soil		Rock		Soil	
	Rupture to the 40–km contour				Rupture to the 20–km contour			
	R=50 km (Seattle)	R=40 km (Portland)	R=50 km (Seattle)	R=40 km (Portland)	R=165 km (Seattle)	R=145 km (Portland)	R=165 km (Seattle)	R=145 km (Portland)
Peak ground acceleration (g)	0.20	0.20	0.30	0.35	0.10	0.10	0.15	0.15
Maximum 5% spectral acceleration at 0.3 and 0.7 s (g)	0.40–0.30	0.45–0.35	0.70–0.50	0.75–0.55	0.15–0.10	0.20–0.15	0.30–0.20	0.35–0.25
5% spectral velocity at 1 s period (cm/s)	30	35	50	55	10	15	20	25

Table 5b. Empirically Predicted Ground Motions† for Seattle or Portland for a M=7.9 Benioff Zone Rupture Based on the Relations of Youngs and others (1988) and Youngs and Coppersmith (1989) and Observed Values in the 1949 and 1965 Earthquakes.

	M=7.9 Benioff Zone Rupture—Predicted Ground Motions		M=7.1, 1949 Earthquake—Observed Ground Motions	M=6.5, 1965 Earthquake—Observed Ground Motions
	Rock	Soil	Soil (OHT)+	Soil (OHT)+
	R=50 km		R=49 km	R=91 km
Peak ground acceleration (g)	0.25	0.4	0.28§	0.20§§
Maximum 5% spectral acceleration at 0.3–0.7 s (g)	0.5–0.35	0.9–0.6	0.98–0.38¶	0.43–0.29¶
5% Spectral velocity at 1 s period (cm/s)	35	60	65¶	26¶

†Predicted acceleration is rounded to the nearest 0.05g, and predicted velocity is rounded to the nearest 5 cm/s. Predicted spectral acceleration values are computed by the appropriate transformation of spectral velocity given by the relations of Youngs and others (1988) and Youngs and Coppersmith 1989. +Largest of two horizontal components. OHT—Olympia Highway Test Lab (§Langston, 1981b; §§Baker, 1987). ¶The observed spectral values are taken from Silva and others (this volume).

Table 5b presents calculated ground–motion values using equations developed for Benioff zone earthquakes by Youngs and others (1988) and Youngs and Coppersmith (1989). To illustrating their equations, we postulate a Benioff zone earthquake of magnitude 7.9 directly beneath either Seattle or Portland, almost a worst–case scenario for an event of this type. Ground motions for these Benioff events are substantially larger than that expected from the thrust fault events. Table 5b also shows the ground motion values observed in the 1949 and 1965 earthquakes at the Olympia Highway Test Lab. Although the observed and predicted values can not be compared directly because of differences in magnitude and distance, a casual comparison shows that the empirical predictions are reasonable.

Uncertainty about the position of intraplate continental crust faults and the potential maximum magnitude of earthquakes on these faults precludes estimation of the ground shaking hazard associated with them. If faults such as the Portland Hills, Hood Canal, Saddle Mountain, or the geophysically inferred structures beneath Puget Sound are capable of generating earthquakes, the Seattle and Portland levels of ground shaking from earthquakes on these faults could exceed that from either thrust fault events or Benioff zone events.

THEORETICAL GROUND–MOTION ESTIMATES—BENIOFF ZONE EARTHQUAKES

Theoretical estimation of ground shaking for Pacific Northwest earthquakes at specific sites has been attempted for a few scenarios. Ground–motion time histories and spectra at a specific site can be produced using theoretical models that incorporate fault rupture, propagation path, and site properties. For example, Langston (1981b) modeled the velocity– and displacement–time histories that were recorded during the 1965 Benioff zone earthquake at Seattle, Tacoma, and Olympia. Significant discrepancies between the modeled and recorded motions were attributed to insufficient knowledge about earth structure and the source. The inadequacy of this model led to more realistic attempts, such as ray–tracing calculations in a two–dimensional model of Duwamish river valley (Langston and Lee, 1983). This study showed that order of magnitude changes in ground–motion displacements were predicted over distances of 200 m produced by basin–geometry focusing, an effect which Langston and Lee (1983) suggested explains the geographic variability in ground shaking during the 1965 earthquake. A three–dimensional model was also evaluated with ray–tracing techniques (Ihnen and Hadley, 1986) and compared with the records for the 1965 earthquake. In this study maps of peak acceleration for the Puget Sound region showed variability of a factor of 10 at a given distance from the fault, due primarily to focusing by basin geometry and secondarily to amplification by near–surface low–velocity layers. Their model predicted peak accelerations observed at Seattle and Tacoma reasonably well, but underestimated the peak acceleration at Olympia by a factor greater than 4. Ihnen and Hadley (1986) indicated that their results qualitatively agreed with the observed intensity variations in the 1965 earthquake. By way of comparison with the predictions given above (table 5b), a large fraction of Ihnen and Hadley's ground motion values grouped in the range between 0.03 and 0.3 g at epicentral distances between 0 and 70 km, though some of their predicted values were as large at 0.6 g near the mouth of the Duwamish River.

Silva and others (this volume) used a stochastic ground–motion model and random–vibration theory techniques to predict the ground motions for the 1949 and 1965 earthquakes. Their computations assumed seismic energy release at a point on the fault nearest the site, attenuating propagation between source and site, and an attenuating flat–layered near–surface velocity model that approximates the site effect. In these simulations, the known hypocenter for these earthquakes is the assumed point of energy release. The peak acceleration values predicted by this technique compare favorably with the observed values (table 5b) except at station OHT (Olympia Highway Test Lab) for the 1965 earthquake. For this station, the predicted values are too low by a factor between 2 and 3 (Silva and others, this volume, table A1). They attributed this discrepancy to excessive model attenuation in the near–surface sediments. Silva and others' (this volume) predicted spectral acceleration curves also compare favorably at most periods with the observed spectral accelerations in the 1965 and 1949 earthquakes at stations OHT and FED (Federal Office Building in Seattle). These comparisons demonstrate the value of using this technique for predicting ground motions in this region at other sites or for other types of earthquakes.

THEORETICAL GROUND–MOTION ESTIMATES—THRUST FAULT EARTHQUAKES

Table 6 compares the results of three theoretical modeling studies for scenario earthquakes on the Cascadia thrust fault (Silva and others, this volume; Cohee and others, this volume; Heaton and Hartzell, 1986). Silva and others (this volume) applied the model described above to predict the ground motions for a thrust fault event. In these simulations, they assumed the point of energy release to be at 70 km hypocentral distance from the FED site in Seattle. Thus, this model simulated rupture to the 40 km depth contour on the thrust fault (see discussion above).

Cohee and others (this volume) applied a Green's function technique (see the Glossary) with empirical source functions recorded during the 1985 Chile thrust event at two sites, one underlain by rock and the other underlain by soil. This source function is modified by theoretical functions that simulate the effects of gross crustal structure and geometrical attenuation. Two magnitude 8 thrust fault earthquakes are modeled in this manner, one in Washington (A) and one in Oregon (B). Each rupture zone had a different length, downdip width, and dip, largely in accord with the plate geometry determined by Weaver and Baker (1988). Rupture extends to the 40 km depth contour on the fault in both models. The model included the effects of crustal structure, and implicitly included near-surface site effects. The calculations separately assumed three distributions of energy release on the fault: shallow (upper third of the fault), middle (middle third), and deep (lower third). The largest accelerations at Seattle and Portland occur for the deep rupture models (for both A and B), which places the greatest energy release nearest the Puget-Willamette trough. Peak accelerations at Portland (for the deep model B: 0.1 g for rock and 0.2 g for soil) were slightly lower than that at Seattle (table 6).

Table 6. Theoretical Deterministic Ground-Motion Parameters at Seattle for a Cascadia Thrust Fault Rupture

Reference	Magnitude	Closest Distance to the Fault (km)	Peak Ground Acceleration (g)		Maximum 5% Spectral Acceleration (0.3–0.7s) (g)	
			Rock	Soil	Rock	Soil
Silva and others (this volume)	8	70	0.1	0.2	0.3–0.2	0.5
Cohee and others (this volume); (model A, see text)	8	81	0.1–0.15 (deep model)	0.25 (deep model)	0.2–0.1	0.3+
Heaton and Hartzell (1987; 1989)	9.5	68	0.25 (avg. conditions†)		0.7 (avg. conditions†)	

Accelerations have been rounded to the .05 g. †Average of soil and rock site records were used. The values are quoted for Puget Sound rather than Seattle. +Cohee and others (1991).

Heaton and Hartzell (1987 and 1989) applied an empirical Green's function technique, whereby they summed the radiated ground shaking contribution of lower magnitude events distributed on the fault to simulate the radiation of a magnitude 9.5 earthquake. In their study, an assortment of records produced by thrust-fault earthquakes in Japanese and Alaskan subduction zones served as Green's functions, ranging in magnitude from 7.4 to 8.2. Heaton and Hartzell considered three fault models, but the values we quote in table 6 are for the fault model that ruptures from the trench axis east to the coast. In their study and that of Cohee and others (this volume), the ground-motion values were substantially larger than those shown, if the rupture extended eastward to Puget Sound. Extension of the rupture eastward is inferred, if the slab is arched into the thrust fault region, as suggested by some data sets.

SUMMARY OF GROUND–MOTION ISSUES

Several important factors may control the level of ground shaking in a great subduction zone earthquake. First, the eastward limit of energy-releasing rupture will have a strong influence on the level of shaking in the urban areas of the Puget-Willamette trough. If the thrust fault ruptures no farther east than the outer coast, shaking damage will be limited, compared with a rupture extending beneath the Olympic Peninsula. A rupture limited to the outer coast would reduce the shaking effects particularly for low-rise buildings compared to that shown in the preceding tables. Even in this scenario, however, the hazard for some classes of structures greater than 2–5 stories in height may remain high, particularly those underlain by geologic basins or thick sediments. The amplification of long-period ground shaking by sediment basins and the lower rate of attenuation of long-period motions with distance could lead to damage of some tall structures. The Michoacan earthquake off the coast of Mexico damaged and collapsed some types of buildings in Mexico City about 360 km from the epicenter. Buildings in the 7 to 12 story height range were particularly affected at some Mexico City locales because geologic conditions (Singh and others, 1988) increased the level and duration of shaking to damaging levels for buildings in this class (Smolka and Berz, 1989). The 1989 Loma Prieta magnitude 7.1 earthquake also produced damage as much as 80 km from the rupture zone owing to the effects of site geology (i.e., Theil, 1990). While the most damaging geologic conditions in Mexico City and San Francisco are extreme cases, similar deposits are found at some sites in the Pacific Northwest. Furthermore, a Cascadia fault rupture to the outer coast results in a hypocentral distance to the

Puget–Willamette trough of less than about 160 km, considerably closer than the Mexico City/Michoacan earthquake case.

Second, the length and downdip width of rupture not only affect the level of shaking at a given point, but they also determine the size of the geographic area affected. An earthquake of magnitude 8 to 9+ can be expected to produce damage from the outer coastal areas to the urban areas of the Puget–Willamette trough. Simultaneous damage to a region of this size poses serious problems for emergency response and recovery.

For Benioff zone earthquakes, an important issue is whether these events are possible at any location in the subducted lithosphere for depths below about 40 km and above 80 km. If so, then these events are possible below Seattle, Portland, and other urban areas of the Puget–Willamette trough (Weaver and Shedlock, this volume). Occurrence of the largest possible ($M_{max} \approx 7.5$–8; U.S. Nuclear Regulatory Commission, 1991; Weaver and Shedlock, this volume) Benioff zone events at locations near or beneath these urban zones would produce significantly larger ground motions than were observed at these locations in the 1949 and 1965 earthquakes.

Almost every aspect of continental crust earthquakes is an issue at present. The potential locations, maximum magnitudes, and shaking attenuation are poorly known for this type of earthquake in this region. Nonetheless, geologic evidence for youthful faults of this type exists, and such events have occurred in the region historically. Significant damaging ground motions could result from moderate-to-large continental crust earthquakes occurring near urban areas. Notably, this type of earthquake has the potential to produce more intense damage to structures of all classes than thrust fault and Benioff earthquakes, if the shallow event occurs within or near an urban area. An event of this type, however, would likely affect a smaller region than a great Cascadia earthquake. Few ground motion estimates have been produced for this type of event in the Pacific Northwest to date, primarily because of uncertainty about which faults should be considered sources of future events.

USERS AND USES OF EARTHQUAKE–HAZARD INFORMATION

Earthquakes are a recognized hazard in the Pacific Northwest, a fact that has resulted in scientific and engineering studies to assess the hazards. This volume contains some of the significant new information produced by these studies. This information when translated for and transferred to potential users (table 7) should result in the selection and adoption of appropriate earthquake–hazard reduction techniques, as, for example, those introduced by Kockelman (list 1, last chapter of this volume). This volume provides an opportunity to build a bridge between the producers and the users of earthquake–hazard information in the Pacific Northwest.

Table 7. Examples of potential users of earthquake—hazard information in the Pacific Northwest

City, county, and multicounty government users	Federal government users (con't)
Local building, engineering, zoning, and safety departments	Farmers Home Administration
City and county offices of emergency services or management	Federal Emergency Management Agency
County tax assessors	Federal Housing Administration
Mayors and city council members	Federal Insurance Administration
Multicounty planning, development, and preparedness agencies	Federal Power Commission
Municipal engineers, planners, and administrators	Forest Service
Planning and zoning officials, commissions, and departments	General Services Administration
Police, fire, and sheriff's departments	National Institute of Standards and Technology
School districts, administrators, and teachers	National Oceanic and Atmospheric Administration
State government users	National Security Council
Building Codes Agency, Oreg.	Nuclear Regulatory Commission
Building and Construction Safety Inspection Division, Wash.	Small Business Administration
Department of Community Development, Wash.	**Other national users**
Department of Ecology (Dam Safety Section), Wash.	American National Red Cross
Department of Energy, Oreg.	Applied Technology Council
Department of Environmental Quality, Oreg.	American Association of State Highway and Transportation
Department of Education, Wash. and Oreg.	Officials American Public Works Association
Department of Geology and Mineral Industries, Oreg.	American Red Cross
Department of Information Services, Wash.	Association of Engineering Geologists
Department of Land Conservation and Development, Oreg.	American Association of State Geologists
Department of Transportation, Wash. and Oreg.	Earthquake Engineering Research Institute
Department of Water Resources, Oreg.	International Conference of Building Officials
Division of Emergency Management, Wash. and Oreg.	National associations of cities, counties, and states
Division of Geology and Earth Resources, Wash.	National Association of Insurance Commissioners
Division of Risk Management, Wash.	National Center for Earthquake Engineering Research
Fire Marshall, Oreg.	National Institute of Building Sciences
Fire Protection Services Division, Wash.	Natural Hazards Research and Applications Center
Legislature	**Private, corporate, and quasi-public users**
Museum of Science and Industry, Oreg.	Civic, religious, and voluntary groups
National Guard	Concerned citizens
Office of the Governor	Construction companies
Utility and Transportation Commission, Wash.	Consulting planners, geologists, architects, and engineers
Federal government users	Extractive, manufacturing, and processing industries
Army Corps of Engineers	Lending and insuring institutions
Department of Energy	Landowners, developers, and real-estate salespersons
Congress and congressional staffs	News media
Department of Housing and Urban Development	Professional and scientific societies (including geologic, engineering, architecture, and planning societies)
Department of Interior (Bureau of Land Management, Bureau of Reclamation, Geological Survey)	University departments (including geology, geophysics, civil engineering, structural engineering, architecture, urban and regional planning, and environmental departments)
Department of Transportation	Utility districts
Environmental Protection Agency	

Responsibility for reducing earthquake hazards resides in all segments of society. Potential users of earthquake–hazard information include many people at community, state, and national levels, both public and private (table 7). Users may be scientists and engineers who use the information to resolve scientific or technical questions, planners and decisionmakers who must consider hazards in the context of other land use and development criteria, and interested citizens who must consider the likelihood of future earthquakes before making personal decisions.

Various levels of government, business, industry, the services sector, voluntary organizations, professional societies, and special–interest groups play important roles in reducing earthquake hazards. The last chapter of this volume presents examples of some group's utilization of earthquake–hazard information to reduce casualties, damage, and socioeconomic interruptions.

CITY, COUNTY, AND AREAWIDE GOVERNMENT USERS

City and county governments are empowered and obligated to provide for the public health, safety, and welfare of their citizens. For example, the Washington State Legislature (1990) requires certain counties and cities to adopt development regulations precluding land uses or developments that are incompatible with designated "critical areas." The minimum guidelines for classifying critical areas include seismic hazard areas defined as "areas subject to severe risk of damage as a result of earthquake induced ground shaking, slope failure, settlement, soil liquefaction, or surface faulting" (Washington State Department of Community Development, 1990). Public liability is an added incentive as discussed by Perkins and May (this volume). Legislatures commonly authorize and require the adoption and administration of local zoning, subdivision, building, grading, and safety ordinances. For example, recognizing the likelihood and severity of a future earthquake, one of the largest cities in the United States—Los Angeles—inventoried and evaluated its unsafe unreinforced masonry bearing–wall buildings. It then enacted an ordinance providing for the owners' strengthening or demolition of 8,000 such buildings scheduled over a 14–yr period.

The various plans, codes, and overlays introduced by Booth and Bethel (last section of this volume) are particularly good reduction techniques into which earthquake–hazard information can be incorporated. For example, King County's "Sensitive Areas" ordinance can be amended to include more specific earthquake–hazard information—ground shaking severity, landslide potential, and liquefaction potential—as it becomes available.

The principal governmental resource at the site of an emergency or disaster is usually local government agencies. They will, of necessity, be heavily involved in preparing for, as well as in coordinating on–the–scene response to, recovery from, and reconstruction after, damaging earthquakes. For example, school preparations have included an earthquake safety program (Martens, 1988), earthquake emergency planning booklet (Noson and Martens, 1987), and a manual for securing nonstructural building components (Washington State Superintendent of Public Instruction, 1989).

STATE GOVERNMENT USERS

The states have the ultimate non–Federal public responsibility for the health, safety, and welfare of their people. For example, the Oregon State Land Conservation and Development Commission (1990, p. 8) adopted statewide planning goals which are mandatory and guidelines which are suggested. One of the goals requires that developments "subject to damage or that could result in loss of life shall not be planned or located in known areas of natural . . . hazards without appropriate safeguards." State agencies incorporate their own seismic–safety standards into their operations and also work with local governments to encourage community seismic–safety efforts. California mapped official zones of potentially–active fault rupture areas. It then mandated that all cities and counties require a geologic report for these areas prior to development and a minimum setback from the trace of active faults for most human occupancy structures.

Officials in almost every state government agency (table 7) need to consider seismic safety in carrying out their duties. For example, the Washington State Seismic Safety Council (1986) recommended essential actions that both the legislature and eight state agencies should undertake for a long–range seismic–risk reduction program. In appendices to its recommendations, the Council identified the primary and secondary earthquake hazards, compared its earthquake activities to California's, and discussed potential losses, vulnerability of school buildings, liability, and risk of chemical accidents. The Council also inventoried the hazard–reduction activities of selected state agencies.

In 1990, the Washington legislature directed the Department of Community Development (which includes the Division of Emergency Management) to create a Seismic Safety Advisory Committee. Funding was provided

> ...solely for the department to develop a seismic safety program to assess and make recommendations regarding the state's earthquake preparedness. The department

> shall create a seismic safety advisory board to develop a comprehensive plan and make recommendations to the legislature for improving the state's earthquake preparedness. The plan shall include an assessment of and recommendations on the adequacy of communications systems, structural integrity of public buildings, including hospitals and public schools, local government emergency response systems, and prioritization of measures to improve the state's earthquake readiness...

This advisory committee presently includes participation by 15 state agencies, business and industry representatives, and 2 local engineering consulting firms. The advisory committee has completed its initial report that recommends priorities for action, including establishment of a seismic safety oversight committee, improved emergency planning, strengthening of buildings, and strengthening of lifelines (Seismic Safety Advisory Committee, 1991).

Likewise, in Oregon, a bill was passed in 1989 (Senate Bill 955) directing the Department of Geology and Mineral Industries (DOGAMI) to assess and mitigate earthquake hazards. In 1990, the Governor appointed the Seismic Safety Advisory Commission (SSPAC) to provide policy guidance on earthquake hazard reduction. Also, in 1990, DOGAMI created a new Earthquake Engineer position. In 1991, the legislature established the SSPAC in two separate Senate Bills (SB 96 and SB 309). Senate Bill 96 also takes the following actions:

1.) Establishes earthquake drills in public, private, and parochial schools with 50 or more pupils.
2.) Amends the state building code to :
    a.) Evaluate new essential facilities, hazardous facilities, major structures (greater than 6 stories and greater than 60000 sq ft, structures greater than 10 stories, and all parking structures.
    b.) Install strong–motion accelerographs in and near selected buildings.
    c.) Review geologic and engineering reports for seismic design of new large or critical buildings.
    d.) File for public use non–interpretive data from seismic site evaluations.
    e.) Establish a surcharge of 1% of building permit fees to support the above activities.

Senate Bill 309 authorizes the Building Codes Agency to adopt regulations to require correction of unsafe conditions caused by earthquakes in existing buildings. The bill also grants building codes inspectors authority to inspect for post–earthquake damage.

Blair and Spangle (1979, p. 24) noted that the "location and construction of public facilities, management of State lands, provision of services, and delegation of powers and responsibilities to local governments should all reflect an awareness that damaging earthquakes are inevitable." Public awareness of the costs of damaging earthquakes, and techniques to reduce such costs, have been addressed in the Pacific Northwest in various ways. These include the Council's recommendations (introduced above), bibliography and index to seismic hazards (Manson, 1988), serial publications such as Oregon Geology and Washington Geology (formerly Washington Geologic Newsletter), conferences and workshops (Walsh and others, 1990), outreach programs, and guidebooks (Noson and others, 1988).

FEDERAL GOVERNMENT USERS

The federal government plays a crucial role in stimulating state and local governments to improve their seismic–safety efforts. The development and distribution of loss scenarios are a part of this role. For example, the U.S. Geological Survey (1975) prepared a study of potential earthquake losses in the Puget Sound area and is funding an earthquake–hazard assessments program in the Pacific Northwest. Federal efforts to reduce risks have evolved from a national commitment to provide disaster relief to states and localities devastated by earthquakes. One of these efforts includes the preparation and wide distribution of a series of guidebooks for reducing earthquake hazards (see list 3 in the last chapter of this volume).

Federal agencies are also responsible for the seismic safety of their own facilities. After the 1971 San Fernando (southern California) earthquake, the U.S. Veteran's Administration (1976; 1986) developed earthquake–resistant design requirements for VA hospital facilities and established seismic protection provisions for furniture, equipment, and supplies for these hospitals. In addition, the U.S. Government, through the Federal Emergency Management Agency (which leads the National Earthquake Hazards Reduction Program), formulates plans for national actions in response to catastrophic earthquakes. These plans include large–scale preparedness for, response to, and recovery from major damage or loss of life. Such plans may include search and rescue, evacuation, provision for medical care, food, shelter, and police protection, and other emergency help.

The federal government, through the U. S. Nuclear Regulatory Commission, is also responsible for the safe siting, design, and operation of nuclear power plants. As such, they have commissioned many studies by the applicants and engineering firms and reviews on the geologic and seismic hazards that might affect nuclear facilities in the Pacific Northwest (i.e., U.S. Nuclear Regulatory Commission, 1991). The U. S. Nuclear Regulatory Commission recently increased the magnitude of the Safe Shutdown Earthquake for Cascadia thrust fault events to M=8.25 (U.S. Nuclear Regulatory Commission, 1991), partly in recognition of work sponsored by the U. S. Geological Survey. The U.S. Nuclear Regulatory Commission also funds an independent research program on earthquake hazards.

The federal government is also responsible for the seismic safety of many dams. As part of this responsibility, both the Army Corps of Engineers and the U.S. Bureau of Reclamation conduct studies concerning the tectonic setting of dams and earthquake and volcanic hazards to these facilities in the Pacific Northwest (i.e., Hawkins and others, 1989a). The Bureau of Reclamation also recognizes M=8.25 as the maximum magnitude subduction event on the Cascadia thrust fault (i.e., Peity and others, 1990).

OTHER NATIONAL USERS

Voluntary organizations (which provide aid to victims of disasters), national institutes, and scientific and professional organizations make major contributions to seismic safety. For example, the American National Red Cross (1986) prepared a resource guidebook that includes strategies on how to identify and use resources, conduct research and evaluations, work with Red Cross and non–Red Cross resources, and develop public relations. The earthquake preparedness program of Snohomish–Island County, Washington was included as one of 43 examples of Red Cross chapter activities.

Researchers from many disciplines, including seismology, geology, earthquake engineering, and the social sciences, play a major role in reducing earthquake hazards by helping local, state, and federal planners and decisionmakers formulate and evaluate their plans and programs. Experts from the research community provide technical advice to special advisory groups that prepare model building codes such as the Uniform Building Code by the International Conference of Building Officials (1991). Scientific and professional organizations are major contributors to the worldwide exchange of information through publications and technical meetings.

PRIVATE, CORPORATE, AND QUASI–PUBLIC USERS

The private (including corporate) sector exercises perhaps the strongest influence on engineering design and land development in the United States. In choosing where to locate and how to construct facilities, industrial and commercial developers are sensitive to tax and other incentives offered by state and local governments, to favorable labor and retail markets, to transportation networks, and to terrain and water resource conditions. Often, natural hazards are not weighted heavily in these development considerations unless they are likely to affect the facilities within their amortized lifetime or period of ownership. According to the Executive Office of the President:

> Business, industry, and the services sector play the lead roles in constructing new buildings and in developing land. Seismic design provisions in local codes . . . are minimum standards. Thoughtful businessmen interested in providing a safe environment for their consumers and employees, and in protecting their capital investment will want to give careful consideration to earthquake hazards in planning, constructing and maintaining their facilities . . . . In some instances short–term profits may be reduced to increase the long–term benefits of saving lives, reducing property damage, and maintaining the functioning of the economy in the face of a major earthquake.

The seismic retrofitting of the Heritage Building and Union Station (Perbix, 1990a; Perbix, 1990b) in Seattle are examples of good business decisions.

An important, but less discussed, aspect of engineering design includes the securing of non–structural building components—parapets, bookcases, office equipment, light fixtures, and partitions. A particularly good guidebook on reducing nonstructural earthquake damage has been prepared by Reitherman (1983). The hazard–reduction techniques recommended in this guidebook are relatively inexpensive and are fully applicable to the Pacific Northwest.

The last chapter of this volume emphasizes evaluation and revision of hazard–reduction techniques. Another good example is a report on the 1989 Loma Prieta earthquake by the California Association of Hospitals and Health Systems (1990). This report addressed structural, nonstructural, staffing, communications, cost reimbursement, media relations, and research issues. Many recommendations were made to improve building safety, preparedness, response, and recovery.

PREREQUISITES

The effective use of geological, seismological, and engineering information to avoid damage or to reduce loss requires a considerable effort by both the producers and the users of the information. For an integrated hazard assessment method, see Preuss and

Hebenstreit (last section of this volume). Unless scientific and engineering results are specifically translated, the effective user community is limited to other geologists, seismologists, and engineers. If users do not become proficient in interpreting and applying technical information, the information is likely to be misused or even neglected in the decisionmaking process.

A study by Kockelman (1990) on the use of earth-science information by planners and decisionmakers in Utah showed that the most effective use of hazard information was achieved when maps that clearly depicted the likelihood, location, and severity of the hazards were provided. Furthermore, the use of this information for hazard reduction was more likely when the information was delivered to potential users along with technical assistance and encouragement. The effectiveness of any earthquake-hazard reduction program is dependent upon the awareness, understanding, and motivation of engineers, planners, and decisionmakers—public and private. May (this volume) addresses the prospects for hazard reduction in the Pacific Northwest.

SUMMARY

In part, the stimulus for accelerating earthquake hazards research in the Pacific Northwest was the recognition that this region may have experienced magnitude 8 to 9.5 earthquakes along the Cascadia thrust fault during the last 7000 yrs. Other factors continue to increase interest in the earthquake hazard here. For example, recent research indicates that large continental crust earthquakes are possible in or near the principal urban areas of the Pacific Northwest. Given our current level of understanding, it is also likely that Benioff zone earthquakes similar to the 1965 and 1949 events will recur. Some of these earthquakes could be larger ($M \approx 8$ or less) or initiate closer to urbanized areas of the Puget–Willamette trough than in the past. Furthermore, this area has also undergone major increases in population and economic value at risk since the last large earthquake in 1965. Taken together, these considerations provide incentive for continued evaluation of the tectonics and earthquake hazards of the region. The significance of this hazard also provides impetus for continued evaluation of the suitability, level of implementation, and effectiveness of existing plans and regulations to reduce future earthquake effects.

We find that the most important issues regarding the level of earthquake hazard in the Pacific Northwest are summarized as follows.

(1) Large shallow crustal earthquakes are likely in the future, but at present little is known about the recurrence of these events or their potential location. New geological data suggest, however, that such earthquakes are possible at locations close to urban areas and that events of this type (not necessarily on the faults near urban areas) could be as large as $M \approx 8$.
(2) Great earthquakes are possible on at least some segments of the Cascadia thrust fault, and most scientists believe that these earthquakes could have magnitudes at least as large as 8, although magnitudes as large as 9–9.5 have been suggested.
(3) Unfavorable ground conditions in the Puget–Willamette trough are expected to substantially increase the shaking hazard at some sites, particularly for high-rise structures underlain by deep sedimentary basins.
(4) The extent of downdip rupture in a subduction earthquake on the Cascadia fault will strongly control shaking levels in the principal urban areas. A model fitting both strain and uplift rates suggests the fault could rupture downdip to points beneath the Olympic Peninsula, which would substantially increase shaking levels relative to models that limit rupture to the outer coast or further west.
(5) Future large Benioff zone earthquakes are likely, and some scientists believe that these events are possible within the subducted lithosphere from western British Columbia to northwestern California. The probable depth of these events ranges between 40 and 80 km. Their maximum magnitude is likely to be between 7.5 and 8.0. Thus, events of this type appear to be possible with locations and maximum magnitudes that would produce substantially greater damage than the historic Benioff zone earthquakes.

Progress in understanding the potential for great earthquakes or continental crust earthquakes will come from continued paleoseismicity studies, instrumental seismicity studies, and expanded geodetic measurements. Much additional work is also needed to produce useful maps that depict the effects of geologic conditions on ground shaking and the areas of various types of ground failure. While much remains to be done to further our understanding of the earthquake hazard in the Pacific Northwest, progress has been made in several areas, as evidenced by the results presented in this volume.

Although this collection of research papers will be of interest to a wide audience, informing a diverse audience is hampered by the complexity of the problem. The controversy surrounding some issues in the Pacific Northwest, further complicates communication. These factors prescribe substantial technical discussion to convey both the hazard and the limitations of our present data and methods of analysis. Most importantly, however, the historical and geological lessons in hand tell us that earthquake hazards in this region are substantial and that this fact should be made clear to those responsible for the safety of others and the preservation of property in future earthquakes. An important goal of this volume is to provide information to these users and other citizens concerning the scientific state of knowledge about Pacific Northwest earthquake hazards. In addition, clarification of the issues will improve the focus of scientific research on factors that strongly control hazard assessment. This process will lead to increased understanding and recognition of the problem, to more informed decisions by engineers, planners, land and building owners, and public officials, and, ultimately, to the adoption of appropriate risk reduction techniques.

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APPENDIX A

Table A1 Principal Arguments/Counter-arguments† Regarding Contemporary Active/Inactive Subduction in the Cascadia Subduction Zone.

Arguments/Evidence In Support of Active Subduction	Counter–Arguments	Arguments/Evidence In Support of Inactive Subduction	Counter–Arguments
Sea–floor magnetic lineaments indicate spreading of the Juan de Fuca ridge and global plate motion reconstruction indicates convergence between the Juan de Fuca and North America plates	The inferred rate of sea–floor spreading decreased from 6 my years to 700 ky before present, the time of the last magnetic reversal. Thus, if sea–floor spreading continued to decrease since then, subduction may have also ceased. Plate motion reconstructions are presently only loosely constrained.	Seismicity is lacking on the Cascadia thrust fault and low at the Juan de Fuca ridge.	Seismic gaps exist on many faults that have produced large earthquakes. The Juan de Fuca ridge continues to produce new crust and has heat flow values comparable to other active ridges. Subduction could continue without an active ridge spreading, owing to slab pull forces and overriding by the continental plate.
Greatest compressive stress direction inferred from strain data in Washington is subparallel to the plate convergence direction and compressive strain is increasing.	—	Greatest compressive stress direction inferred from earthquake focal mechanisms and bore hole breakouts is NNE to NNW in western Washington, Oregon, northern California, British Columbia, and interior to the Juan de Fuca plate; this direction is not parallel to the convergence direction (ENE).	Active subduction stresses may be in a low–stress phase of the earthquake cycle. Landward counterclockwise rotation of the stress orientation may exist because of permanent deformation of the continental shelf and slope. Or post paleoearthquake continental rebound may have produced present day extensional strain subparallel to the maximum strain rate direction.
The observed horizontal–strain and coastal–uplift rates are successfully modeled by a locked thrust fault converging at the independently estimated plate convergence rate.	Observed strain rates could be produced by permanent deformation of the continental plate without subduction. The strain measurements may be influenced by processes near the earth's surface, such as topography or local deformation.	Low horizontal–strain rates are observed near Puget Sound and on the Olympic Peninsula which could indicate low or ceased subduction, especially if the error in strain rate measurements is comparable to the measurements.	Strain rates may change with time. Low strain rates, however, have been predicted on the basis of a locked thrust fault and Juan de Fuca–continental plate convergence.
A Benioff zone exists beneath Washington and western British Columbia and seismicity is present along the Mendocino, Blanco, and Nootka fracture zones.	A Benioff zone is lacking beneath Oregon and northern California.	A poorly defined trench could be evidence that subduction has ceased.	Buoyant oceanic crust that is well coupled, but still undergoing active subduction may lead to a poorly defined trench based on comparisons with other subduction zones world wide. High sedimentation rates in the Pacific Northwest obscure the trench by filling it with sediments.

Table A1 Principal Arguments/Counter-arguments† Regarding Contemporary Active/Inactive Subduction in the Cascadia Subduction Zone (con't.).

Arguments/Evidence In Support of Active Subduction	Counter–Arguments	Arguments/Evidence In Support of Inactive Subduction	Counter–Arguments
Presence of normal fault earthquakes in the Benioff zone suggests active slab pull against a locked thrust fault. Slab pull is the predominant mechanism driving subduction.	Normal faulting in the Benioff zone could continue in the subducted slab after the thrust fault had permanently locked.	—	—
A subducted slab exists downdip from the thrust fault which supports subduction.	This evidence supports ancient subduction, but does not provide evidence of current subduction.	—	—
Offshore and onshore Late Pleistocene and probable Holocene folding and faulting are consistent with subduction .	Tectonic processes other than subduction could produce some of these structures.	—	—

†See the text for the argument citations and additional discussion.

Table A2 Principal Arguments/Counter–arguments Regarding Contemporary Seismic/Aseismic Subduction in the Cascadia Subduction Zone†.

Arguments/Evidence In Support of Seismic Subduction	Counter–Arguments	Arguments/Evidence In Support of Aseismic Subduction	Counter–Arguments
Geologic evidence for multiple episodes of sudden prehistoric coastal subsidence exists.	Rapid eustatic changes in sea level could produce similar effects in some cases. Coastal subsidence at some locales may be produced by deformation on shallow or local faults. Coastal submergence may also have occurred during interseismic periods at some locales. The ages of peat burial are not demonstrably equal at all estuaries. Aseismic slip on the plate boundary could produce subsidence.	Lack of earthquakes along the thrust fault suggests aseismic subduction.	Other active subduction zones are commonly aseismic for some periods of the seismic cycle. If only the offshore section of fault is locked, it may be difficult to detect small events in that region using existing onshore networks.
Sand deposits atop peats, buried by rapid coastal subsidence, suggesting the occurrence of tsunamis.	Over decades following rapid or gradual coastal subsidence, estuary flooding by high tides, unusual river discharge or tsunamis from distant earthquakes could, in some cases, also produce these layers.	Holocene coastal terraces are mostly absent in the region.	Terraces may be absent because of repeated coseismic coastal subsidence and the position of the coast relative to the locked thrust fault. Continuing eustatic sea level rise or high rates of coastal erosion may have obliterated terraces.
Thirteen episodes of submarine turbidites are observed in offshore sea–fan deposits less than 7500 years old.	Could be triggered naturally by excess sedimentation or by earthquakes other than subduction events.	—	—
Observed horizontal–strain and coastal–uplift rates are successfully modeled by a locked thrust fault converging at the independently estimated plate convergence rate.	Observed strain rates could be produced by permanent deformation of the continental plate without subduction. The strain measurements may be influenced by processes near the earth's surface, such as topography or local deformation.	Low horizontal strain rates are observed near Puget Sound and on the Olympic Peninsula which could indicate low or ceased subduction, especially if the error in strain rate measurements is comparable to the measurements.	Strain rates may change with time. Successful modeling of strain rates is a counter.
Leveling data that show tilt to the north in Oregon and eastward tilt of the Coast Ranges in Washington are consistent with a locked thrust fault.	—	In one model, eastward tilt is expected for a zone undergoing aseismic subduction.	In other models, either eastward or westward tilt may be observed depending on position relative to the locked thrust fault.
The Cascadia subduction zone is analogous to subduction zones off S. Chile, Mexico, southwestern Japan, and Colombia.	Cascadia subduction zone has unique features. No other active subduction zone duplicates Cascadia.	—	—

Table A2 Principal Arguments/Counter–arguments Regarding Contemporary Seismic/Aseismic Subduction in the Cascadia Subduction Zone (con't.)†.

Arguments/Evidence In Support of Seismic Subduction	Counter–Arguments	Arguments/Evidence In Support of Aseismic Subduction	Counter–Arguments
Exceptionally young, buoyant oceanic slab should couple and store stress more effectively with the continental plate. A locked thrust fault is supported by the downdip tension direction in the Benioff zone that results from slab pull against a locked thrust fault and by the presence of seismicity in the continental plate.	If the fault locks too strongly or the subducted crust is too buoyant, the slab pull force may be too small to overcome the frictional stress on the fault. Normal faulting in the Benioff zone could continue in the subducted slab after the thrust fault had permanently locked.	High heat flow and high fluid pressure in the accretionary wedge and subducted soft sediments along the thrust fault may prevent accumulation of the required stresses for a great subduction zone earthquake.	Inferences about the temperatures, fluid pressures, rock properties, and presence of subducted soft sediments along the thrust fault are difficult to test and are considered by some to be circumstantial evidence.
Offshore and onshore Late Pleistocene and probable Holocene folding and faulting is consistent with subduction.	Could also be formed during aseismic subduction processes, particularly in the soft sediments of the accretionary wedge.	—	—
On the one hand, the mostly consistent compressive stress direction (NNE to NNW) inferred from focal mechanisms and well–breakouts in both the Juan de Fuca and North America plates suggests a locked thrust fault. See Table A1 for other stress related arguments that apply.	On the other hand, the lack of agreement between this stress direction and the convergence direction suggests that subduction produced compressive stress is low compared to the compressive stress being produced by the Pacific plate/North America plate interaction. See Table A1 for other stress related arguments that apply.	—	—
Holocene paleolandslides are known in the region that yield ^{14}C ages similar to those for some inferred subduction earthquakes.	Rainfall or local earthquakes could also have produced paleolandslides.	—	—

†See the text for the argument citations and additional discussion. Because a locked thrust fault is a prerequisite for seismic subduction, some arguments center on whether the effects of a locked fault are present in the various observations. Neither arguments nor counter–arguments are conclusive at present, and in some cases further response to counter–arguments is possible (see the original studies referenced in the body text for this level of detail). In our view, however, the weight of the evidence favors seismic subduction. In particular, the evidence for sudden subsidence, tsunami deposits atop buried peats, and the modeling of strain data by a locked thrust fault are highly significant. Additional work to improve the accuracy of radiocarbon dating and connection of these events with shaking evidence, such as liquefaction and landsliding, could make the case for great subduction earthquakes unequivocal.

APPENDIX B: GLOSSARY OF TECHNICAL TERMS

(This glossary has been adapted from Ziony and Kockelman (1985).

Acceleration. The time rate of change of **velocity** of a reference point during an earthquake. Commonly expressed in percentage of gravity (g), equal to 980 cm/s^2.

Accelerogram. The record from an **accelerograph** showing ground **acceleration** as a function of time.

Accelerograph. A compact, rugged, and relatively inexpensive instrument that records the signal from an **accelerometer**. Film is the most common recording medium.

Accelerometer. A sensor whose output is almost directly proportional to ground acceleration. The conventional strong motion accelerometer is a simple, nearly critically damped **oscillator** having a **natural frequency** near 20 Hz.

Accretionary wedge. Sediments that accumulate and deform where the oceanic and continental plates collide. These sediments are "scraped" off of the top of the downgoing oceanic crustal plate and are added to the leading edge of the continental plate.

Active fault. A **fault** that is considered likely to undergo renewed movement within a period of concern to humans. Faults are commonly considered to be active if the fault has moved one or more times in the last 10,000 yrs, but faults may also be considered active in some cases if movement has occurred in the last 500,000 yrs.

Aftershocks. Secondary tremors that may follow the largest or main shock of an earthquake sequence. Such tremors can extend over a period of weeks, months, or years.

Alluvium. Loosely compacted gravel, sand, silt, or clay deposited by streams.

Amplification. An increase in seismic signal **amplitude** within some range of **frequency** as waves propagate through different earth materials. The signal is both amplified and deamplified at the same site in a manner that is dependent on the frequency band. The degree of amplification is also a complex function of the level of shaking, such that as the level of shaking increases, the amount of amplification decreases. Shaking levels at a site may also be increased by the **focusing** of seismic energy by the geometry of the sediment velocity structure, such as basin subsurface topography.

Amplitude. Zero–to–peak value of any wave-like disturbance.

Arias intensity. A ground–motion parameter derived from an **accelerogram** and proportional to the integral over time of the **acceleration** squared. Expressed in units of velocity (meters per second or centimeters per second).

Aseismic. A fault on which no earthquakes have been observed is termed aseismic.

Attenuation. A decrease in seismic signal **amplitude** as waves propagate from the seismic **source.** Attenuation is caused by geometrical spreading of seismic wave energy and by the absorption and scattering of seismic energy in different (termed anelastic attenuation) Earth materials. Q and kappa are values used in modeling the attenuation of ground motions.

Basement. Igneous and metamorphic rocks that underlie the main sedimentary rock sequences of a region and extend downward to the base of the **crust**.

Bedrock. Relatively hard, solid rock that commonly underlies softer rock, sediment, or soil.

Benioff zone. A dipping planar zone of earthquakes that is produced by the interaction of a downgoing oceanic crustal plate with the continental plate. These earthquakes can be produced by slip along the **thrust interface fault** or by slip on faults within the downgoing plate as a result of bending and extension, as the plate is pulled into the **mantle**. Slip may also initiate between adjacent segments of down–going plates below the thrust fault. The Benioff zone in the Pacific Northwest is not as well developed as it is in other subduction zones. The events that are observed do not appear to be produced by slip along the **thrust interface zone**. Also known as the Wadati–Benioff zone.

Body wave. Seismic wave propagated in the interior of the Earth. **P** and **S waves** are examples.

Brittle–ductile boundary. A depth in the crust across which the thermo–mechanical properties of the crust change from brittle above to ductile below the boundary. A large percentage of the earthquakes in the crust initiate at or above this depth on high–angle faults; below this depth fault slip may be aseismic and fault slip may grade from high angle to low angle.

Bulk density. The weight of a material divided by its volume, including the volume of its pore spaces.

^{14}C age date. A relative age obtained for geologic materials containing bits or pieces of carbon using measurements of the proportion of radioactive carbon (^{14}C) to daughter carbon (^{12}C). These dates are independently calibrated to calendar dates.

Coherent slides. Landslides that consist of a few relatively intact blocks of rock or soil that move together. The basal failure surface of most such slides is several meters or tens of meters below the land surface.

Cohesionless. Refers to a sediment whose shear strength depends only on friction, because there is no bonding between the grains. Typical of clay–free sandy deposits.

Colluvium. Loose soil or rock fragments on or at the base of gentle slopes or hillsides. Deposited by or moving under the influence of rainwash or downhill creep.

Compressional wave. See **P wave**.

Convolution. A mathematical operation that describes the action of a linear system on a signal, such as that of a filter on a seismic signal.

Corner frequency. That **frequency** at which the curve representing the **Fourier amplitude spectrum** of a recorded earthquake seismic signal abruptly changes slope.

Creep. Slow, more or less continuous movement occurring on faults, owing to ongoing **tectonic** deformation, or on slopes owing to gravitational forces.

Critical facilities. Structures whose ongoing performance during an emergency is required or whose failure could threaten many lives. May include (1) structures such as nuclear power reactors or large dams whose failure might be catastrophic; (2) major communication, utility, and transportation systems; (3) involuntary– or high–occupancy buildings such as prisons or schools; and (4) emergency facilities such as hospitals, police and fire stations, and disaster–response centers.

Crust. The outermost major layer of the Earth, ranging from about 10 to 65 km in thickness worldwide. The continental crust is about 40 km thick in the Pacific Northwest. The thickness of the oceanic crust in this region varies between about 10 and 15 km. The crust is characterized by **P–wave** velocities less than about 8 km/s. The uppermost 15 to 35 km of the crust is brittle enough to produce earthquakes. The seismic crust is separated from the lower crust by the **brittle–ductile boundary.**

Damping. The reduction in **amplitude** of an **oscillation** owing to matter's internal absorption of energy. Also see **geometrical attenuation.**

Design earthquake. The postulated earthquake (commonly including a specification of the **ground motion** at a site) that is used for evaluating the earthquake resistance of a particular structure.

Dip. Inclination of a planar geologic surface (for example, a fault or a bed) from the horizontal.

Dip slip. See **fault**

Directivity. An effect of a propagating fault rupture whereby earthquake **ground motion** in the direction of propagation is more severe than that in other directions from the earthquake source.

Displacement. The difference between the initial position of a reference point and any later position. (1) In seismology, displacement is the ground motion commonly calculated by integrating an **accelerogram** twice with respect to time and is expressed in units of length, such as centimeters. (2) In geology, displacement is the permanent offset of a geologic or manmade reference point along a **fault** or a landslide.

Disrupted slides and falls. Landslides that are broken during movement into chaotic masses of small blocks, rock fragments, or individual grains. The basal failure surface of most such slides is within a few meters of the land surface.

Earthquake hazard. Any physical phenomenon associated with an earthquake that may produce adverse effects on human activities. This includes **surface faulting**, ground shaking, landslides, **liquefaction, tectonic deformation, tsunami**, and **seiche** and their effects on land use, man–made structures, and social–economic systems.

Elastic dislocation theory. In seismology, a theoretical description of how an elastic Earth responds to fault slip, as represented by a distribution of **displacement** discontinuities.

Epicenter. The point on the Earth's surface vertically above the point (focus or hypocenter) in the crust where a seismic rupture nucleates.

Fault. A fracture along which there has been significant **displacement** of the two sides relative to each other parallel to the fracture. Strike–slip faults are chiefly vertical fractures along which rock masses have shifted horizontally. Dip–slip faults are chiefly inclined fractures along which rock masses have shifted vertically. If the rock mass above an inclined fault is depressed, the fault is termed normal slip, whereas the term reverse slip (or thrust) indicates that the side above the fault is elevated. Oblique–slip faults have significant components of both strike and dip slip along them.

Fault–plane solution. An analysis using stereographic projection or its mathematical equivalent to determine the attitude of the causative fault and its direction of **slip** from the radiation pattern of seismic waves using earthquake records at many stations. The analysis most commonly uses the direction of **first motion** of **P waves** and yields two possible orientations for the fault rupture and the direction of **seismogenic** slip. From these data, inferences can be made concerning the principal axes of **stress** in the region of the earthquake. The principal stress axes determined in this method are the compressional axis (also the P–axis, the axis of greatest compression, or σ_1), the tension axis (also the T–axis or axis of least compression or σ_3), and the intermediate stress axis (σ_2).

Fault scarp. Step-like linear landform coincident with a **fault trace** and caused by geologically recent slip on the fault.

Fault trace. Intersection of a fault with the ground surface; also, the line commonly plotted on geologic maps to represent a fault.

Filter. In seismology, a physical system or a mathematical operation that changes the waveform or **amplitude** of a signal.

Filtering. Attenuation of certain frequency components of a seismic signal and the **amplification** of others. For a recorded signal, the process can be accomplished electronically or numerically in a digital computer. Filtering also occurs naturally as seismic energy passes through the Earth.

First motion. On a seismogram, the direction of **ground motion** as the **P wave** arrives at the **seismometer.** Upward ground motion indicates an expansion in the **source** region; downward motion indicates a contraction.

Focal–mechanism solution. See **fault–plane solution.**

Focus. See **hypocenter.**

Focusing. See **amplification.**

Fourier amplitude spectrum. The relative **amplitude** at different component frequencies which are derived from a **time series** by Fourier analysis.

Fourier transform. The mathematical operation that resolves a time series (for example, a recording of ground motion) into a series of numbers that characterize the relative amplitude and phase components of the signal as a function of frequency.

Frequency. Number of cycles occurring in unit time. **Hertz (Hz)**, the unit of frequency, is equal to the number of cycles per second.

Fundamental period. The longest period for which a structure shows a maximum response. The reciprocal of natural frequency.

G or g. See **acceleration.**

Gaussian noise spectrum. The spectrum of a time history whose sample values are generated by random selection from a population that has a specified mean and standard deviation. The values (ordinates) have a "bell–shaped" distribution about the mean. This type of spectrum is multiplied by a source spectrum to obtain predicted ground motion spectra for scenario earthquakes.

Geodetic. Refers to the determination of the size and shape of the Earth and the precise location of points on its surface.

Geometrical attenuation. That component of **attenuation** of seismic wave amplitudes owing to the radial spreading of seismic energy with distance from a given source.

Geomorphology. The study of the character and origin of landforms.

Geotechnical. Refers to the use of scientific methods and engineering principles to acquire, interpret, and apply knowledge of Earth materials for solving engineering problems.

Green's function. A mathematical representation that, in reference to earthquake shaking, is used to represent the **ground motion** caused by instantaneous **slip** on an small part of a fault. Green's functions can be summed over a large fault surface to compute the ground shaking for a large earthquake rupturing a fault of finite size. The fractional fault slip events that are summed can be records from small earthquakes on the fault or they can be theoretically computed small earthquake records, summed for discrete positions on the fault.

Gravity. The attraction between two masses such as the earth and objects on its surface. Commonly referred to as the acceleration of gravity. Changes in the gravity field can be used to infer information about the structure of the earth's lithosphere and upper mantle. Interpretations of changes in the gravity field are generally applied to gravity values corrected for extraneous effects. The corrected values are referred to by various terms, such as free air gravity, Bouguer gravity, and isostatic gravity depending on the number of corrections.

Ground motion. General term referring to the qualitative or quantitative aspects of shaking of the Earth's surface from earthquakes or explosions.

Halfspace. A mathematical model bounded by a planar surface but otherwise infinite. Properties within the model are commonly assumed to be homogeneous and isotropic, unlike the Earth itself, which is heterogeneous and anisotropic.

Hertz (Hz). A unit of frequency. Expressed in cycles per second.

Holocene. Refers to an age between the present and 10,000 years before present. Faults of this age can be considered active based on the observation of historic activity on faults of this age in other locales.

Hypocenter. The point within the Earth where an earthquake rupture nucleates. Also commonly termed the **focus.**

Intensity. A subjective numerical index describing the severity of an earthquake in terms of its effects on the Earth's surface, on humans and their structures. Several scales exist, but the scales most commonly used in the U.S. are the Modified Mercalli scale and the Rossi–Forel scale.

Intraplate/Interplate. Intraplate pertains to processes within the earth's crustal plates. Interplate pertains to processes between the plates.

Isoseismal. A line on a map bounding points of equal **intensity** for a particular earthquake.

Kinematic. Refers to the movement patterns that result in a particular deformation.

Late Quaternary. Refers to an age between the present and 500,000 years before present. Faults of this age can be considered active based on the observation of historic activity on faults of this age in other locales.

Lateral spreads and flows. Landslides that commonly form on gentle slopes and that display rapid fluid–like flow movement.

Least–squares fit. An approximation of a set of data with a curve such that the sum of the squares of the differences between the observed points and the curve is a minimum.

Liquefaction. Process by which water–saturated sediment temporarily loses strength, usually because of strong shaking, and behaves as a fluid.

Lithology. The description of rock composition and texture.

Lithosphere. The outer solid portion of the Earth, including the **crust** and uppermost **mantle.** The lithosphere is about 100 km thick, although its thickness is age dependent. The lithosphere below the crust is brittle enough in some locales to produce earthquakes by faulting.

Love wave. A type of seismic surface wave having a horizontal motion that is transverse to the direction of propagation.

Magnetic polarity reversal. A change of the Earth's magnetic field to the opposite polarity that has occurred at irregular intervals over geologic time. Polarity reversals can be preserved in sequences of magnetized rocks and compared with standard polarity change time scales to estimate geologic ages of the rocks. **Spreading oceanic ridges** commonly preserve this pattern of polarity reversals, and this pattern can be used to determine the rate of ocean–ridge spreading. The pattern of reversals recorded in the rocks is termed sea–floor magnetic lineaments.

Magnitude. A number that characterizes the relative size of an earthquake. Magnitude is based on measurement of the maximum motion recorded by a **seismograph** (sometimes for earthquake waves of a particular **frequency**), corrected for attenuation to a standardized distance. Several scales have been defined, but the most commonly used are (1) local magnitude (M_L) (commonly referred to as "Richter magnitude"), (2) surface–wave magnitude (M_S), (3) body–wave magnitude (m_b), and (4) moment magnitude (M_w). Scales (1–3) have limited range and applicability and do not satisfactorily measure the size of the largest earthquakes. The moment magnitude (M_W) scale, based on the concept of **seismic moment,** is uniformly applicable to all sizes of earthquakes, but is more difficult to compute than the other types. In principal, all magnitude scales could be cross calibrated to yield the same value for any given earthquake, but this expectation has proven to be only approximately true, thus, the need to specify the magnitude type as well as its value.

Mantle. That part of the Earth's interior between the metallic core and the **crust.**

Microzonation. Geographic delineation at local or site scales of areas having different potentials for hazardous effects of earthquakes (can include any of the **earthquake hazards** and their effects).

Moho. A discontinuity in seismic velocity that marks the boundary between the earth's **crust** and **mantle.** Also termed the Mohorovičić discontinuity, after the Croatian seismologist Andrija Mohorovičić (1857–1936) who discovered it. The boundary is between 25 and 60 km deep beneath the continents and between 5 and 8 km deep beneath the ocean floor.

Moment magnitude. See magnitude.

Newmark analysis. A numerical technique that models a potential landslide as a rigid friction block resting on a slope and uses a strong–motion record to calculate the expected displacement of the block under earthquake shaking.

Natural frequency(ies). The discrete frequency(ies) at which a particular elastic system vibrates when it is set in motion by a single impulse and not influenced by other external forces or by damping. The reciprocal of fundamental period.

Normal stress. That **stress** component perpendicular to a given plane.

Oscillator. A mass that moves with oscillating motion under the influence of external forces and one or more forces that restore the mass to its stable at–rest position. In earthquake engineering, an idealized damped mass–spring system that is used as a model of the **response** of a structure to earthquake **ground motion.** A **seismograph** is also an oscillator of this type.

P wave. A seismic **body wave** that involves particle motion (alternating compression and extension) in the direction of propagation.

Paleoseismic. Referring to the prehistoric seismic record as inferred from young geologic sediments.

Peak value. Largest value of acceleration, velocity, or displacement recorded for a particular earthquake time–history.

Peak acceleration. The largest value of ground acceleration recorded during an earthquake at a specific site. Peak acceleration most commonly refers to the maximum horizontal component of ground acceleration, although maximum vertical accelerations are also measured. Peak accelerations decrease on average with increasing distance from the fault and with decreasing magnitude of the event. Peak acceleration is an important ground–motion parameter for engineering applications because it is related to the maximum force applied to the building.

Pedogenic. Pertaining to processes that add, transfer, transform, or remove soil constituents.

Period. The time interval required for one full cycle of a wave.

Period band. That range of periods being considered in an analysis of **ground motion.**

Phase. (1) A stage in periodic motion, such as **oscillation,** measured with respect to a given initial point and expressed in angular measure. (2) A pulse of seismic energy arriving at a definite time.

Physiographic. Refers to the character and distribution of landforms.

Plate tectonics. A well–proven theory that considers the Earth's **crust** and upper **mantle** to be composed of a number of large, thin, relatively rigid plates that move relative to one another. Interaction along plate boundaries commonly results in earthquakes. Several styles of faults bound the plates, including thrust faults along which plate material is subducted or consumed in the mantle and spreading ridges along which new crustal material is produced.

Poisson distribution. A probability distribution that characterizes discrete events occurring independently of one another in time.

PSRV. Pseudo–relative velocity response spectrum. See **response spectrum**.

Q. See **attenuation**

Quaternary. The geologic time period comprising the last 2 million years.

Radiometric. Pertaining to the measurement of geologic time by the analysis of certain radioisotopes in rocks and their known rates of decay.

Random vibration theory. A theoretical formulation that links band–limited **Gaussian noise spectra,** representing the spectra of earthquake ground motions, for example, with its corresponding **time history peak values.**

Rayleigh wave. A seismic **surface wave** causing an elliptical motion of a particle at the free surface, with no transverse motion.

Ray tracing method. A computational method of computing ground shaking estimates that assumes that the ground motion is composed of multiple arrivals that leave the source and are reflected or refracted at velocity boundaries according to Snell's Law. The amplitudes of reflected and refracted waves at each boundary are recalculated according to the law of conservation of energy.

Recurrence interval. The average time span between events (such as large earthquakes, ground shaking exceeding a particular value, or liquefaction) at a particular site. Also termed return period.

Reflection. The energy or wave from a seismic source that has been returned (reflected) from an interface between materials of different elastic properties within the Earth.

Reflector. An interface between materials of different elastic properties that reflects seismic waves.

Refraction. (1) The deflection of the ray path of a seismic wave caused by its passage from one material to another having different elastic properties. (2) Bending of a **tsunami** wave front owing to variations in the water depth along a coast line.

Regression analysis. A statistical technique applied to data to determine, for predictive purposes, the degree of correlation of a dependent variable with one or more independent variables. See **least–squares fit.**

Relaxation theory. Concept wherein radiated seismic waves of an earthquake result when stored strain within the Earth is released at the time of **slip** along a fault; adjacent fault blocks reach new states of equilibrium.

Rigidity. See **shear modulus**.

Residual. The difference between the measured and the predicted values of some quantity.

Resonance. An increase in the **amplitude** of vibration in a elastic body or system when the **frequency** of the shaking force is close to one or more of the **natural frequencies** of a shaking body.

Response. The motion in a system resulting from shaking under specified conditions.

Response spectrum. A curve showing the mathematically computed maximum response of a series of simple damped harmonic **oscillators** of different **natural frequency(ies)** to a particular record of earthquake ground **acceleration.** Response spectra, commonly plotted on tripartite logarithmic graph paper, show the oscillator's maximum **acceleration, velocity,** and **displacement** as a function of oscillator frequency for various levels of oscillator **damping.** A computational approximation to the response spectrum is referred to as the pseudo–relative velocity response spectrum **(PSRV).** These curves are used by engineers to estimate the maximum response of simple structures to complex ground motions. For example, the 5% spectral acceleration at one second is the maximum acceleration of the top of a structure with 5% damping whose natural period of vibration is one second when subjected to a specified input ground acceleration record.

Return period. See **recurrence interval.**

Root mean square. Square root of the mean square value of a random variable.

Rupture front. The instantaneous boundary between the slipping and locked parts of a fault during an earthquake. Rupture in one direction on the fault is referred to as unilateral. Rupture may radiate outward in a circular manner or it may radiate toward the two ends of the fault from an interior point, which is referred to as bilateral.

Rupture velocity. The speed at which a **rupture front** propagates during an earthquake.

S wave. A seismic **body wave** that involves a shearing motion in a direction perpendicular to the direction of propagation. When it is resolved into two orthogonal components in the plane perpendicular to the direction of propagation, SH denotes the horizontal component and SV denotes the orthogonal component.

Sand boil. Sand and water ejected to the ground surface as the result of liquefaction at shallow depth; the conical sediment deposit that remains as evidence of liquefaction.

Sea–floor magnetic lineaments. See **magnetic polarity reversals.**

Segmentation. The property of faults to be broken along their length by other faults that cross them or to be limited in length by other factors such as topography or bends in the **strike** of the fault. Segmentation can limit the length of faulting in a single earthquake to some fraction of the total fault length, thus, also limiting the size of the earthquake.

Seiche. Oscillation of the surface of an enclosed body of water owing to earthquake shaking.

Seismic hazard. See **earthquake hazard.**

Seismic impedance. Seismic **P wave** velocity multiplied by the **bulk density** of a medium.

Seismic wave. An elastic wave generated by an impulse such as an earthquake or an explosion. Seismic waves may propagate either along or near the Earth's surface (for example, **Rayleigh** and **Love waves**) or through the Earth's interior **(P and S waves).**

Seismicity. The geographic and historical distribution of earthquakes.

Seismic moment. A measure of the size of an earthquake based on the area of fault rupture, the average amount of slip, and the shear modulus of the rocks offset by faulting. Seismic moment can also be calculated from the amplitude spectra of seismic waves.

Seismic risk. The probability of social or economic consequences of an earthquake.

Seismic zonation. Geographic delineation of areas having different potentials for hazardous effects from future earthquakes. Seismic zonation can be done at any scale, national, regional, local or site. Also see **microzonation.**

Seismogenic. Capable of generating earthquakes.

Seismogram. A record written by a seismograph in response to ground motions produced by an earthquake, explosion or other ground–motion sources.

Seismometer/Seismograph. A damped oscillating mass, such as a damped mass–spring system, used to detect seismic wave energy. The motion of the mass is commonly transformed into an electrical voltage. The electrical voltage is recorded on paper, magnetic tape, or another recording medium. This record is proportional to the motion of the seismometer mass relative to the earth, but it can be mathematically converted to a record of the absolute motion of the ground. Seismograph is a term that refers to the seismometer and its recording device as a unit.

Separation. The distance between any two parts of a reference plane (for example, a sedimentary bed or a geomorphic surface) offset by a fault, measured in any plane. Separation is the apparent amount of fault **displacement** and is nearly always less than the actual **slip.**

Shear modulus. The ratio of shear stress to shear strain of a material during simple shear.

Shear stress. That stress component on a plane, such as a fault plane, that results from forces tangential to the plane.

Shear wave. See **S wave.**

Slip. The relative **displacement** of formerly adjacent points on opposite sides of a fault, measured in the fault surface.

Slip model. A kinematic model that describes the amount, distribution, and timing of slip associated with a real or postulated earthquake.

Slip rate. The average rate of **displacement** at a point along a fault as determined from geodetic measurements, from offset manmade structures, or from offset geologic features whose age can be estimated. It is measured parallel to the dominant **slip** direction or estimated from the vertical or horizontal **separation** of geologic markers.

Soil. (1) In engineering, all unconsolidated material above **bedrock.** (2) In soil science, naturally occurring layers of mineral and (or) organic constituents that differ from the underlying parent material in their physical, chemical, mineralogic, and morphologic character because of **pedogenic** processes.

Soil profile. The vertical arrangement of soil horizons down to the parent material or to bedrock. Commonly subdivided into A, B, and C horizons.

Source. (1) The geologic structure that generates a particular earthquake. (2) The explosion used to generate acoustic or seismic waves.

Spectral acceleration. Commonly refers to either the **Fourier spectrum** of ground acceleration or the **pseudo–relative acceleration spectrum**

Spectral amplification. A measure of the relative shaking **response** of different geologic materials. The ratio of the **Fourier amplitude spectrum** of a **seismogram** recorded on one material to that computed from a seismogram recorded on another material for the same earthquake of explosion.

Spectral ratio. See **spectral amplification.**

Spectrum. The distribution or range of values of a characteristic of a physical system. In seismology, commonly a curve showing **amplitude** and **phase** as a function of **frequency** or **period.**

Spreading oceanic ridge. A fracture zone along the ocean bottom that accommodates upwelling of mantle material to the surface, thus creating new crust. This fracture is topographically marked by a line of ridges that form as molten rock reaches the ocean bottom and solidifies.

Standard deviation. The square root of the average of the squares of deviations about the mean of a set of data. Standard deviation is a statistical measure of spread or variability.

Standard penetration resistance. A measure of relative density expressed by the number of blows (blow count) needed to push a probe a standard distance into sediment. The "standard penetration test" determines the number of blows required to drive a standard sampling spoon 1 ft into the sediment by repeatedly dropping a 140–lb weight from a height of 30 in.

Station. A ground position at which a geophysical instrument is located for an observation.

Stick slip. The rapid displacement that occurs between two sides of a fault when the shear stress on the fault exceeds the frictional stress. Stick slip displacement on a fault radiates energy in the form of seismic waves creating an earthquake.

Strain rate. Strain measurements are computed from observed changes in length on the earth's surface, usually along multiple paths. Because the changes in length are observed over varying time periods and path lengths, they are expressed as change in length divided by the measurement distance divided by the measurement time period. This number, which is expressed as the change in length per unit length per unit time, is termed the strain rate. These measurements are used to infer the directions of principal strain/stress rates near the earth's surface.

Stratigraphy. The study of the character, form, and sequence of layered rocks.

Stress. Force per unit area acting on a plane within a body. Six values are required to characterize completely the stress at a point: three normal components and three shear components.

Stress drop. The difference between the **stress** across a fault before and after an earthquake. A parameter in many models of the earthquake source that has a bearing on the level of high–frequency shaking that the fault radiates. Commonly stated in units termed bars (1 bar = 1 kg/cm^2 or megapascal = 10 bars).

Strike. Trend or bearing, in relation to north, of the line defined by the intersection of a planar geologic surface (for example, a fault or a bed) and a horizontal surface.

Strike slip. See **fault**

Strong motion. Ground motion of sufficient **amplitude** and duration to be potentially damaging to engineered structures.

Subduction. A **plate tectonics** term for the process whereby the oceanic **lithosphere** descends beneath the continental **lithosphere.**

Subduction thrust fault. The fault that accommodates the differential motion between the downgoing oceanic crustal plate and the continental plate as **subduction** occurs. This fault is the contact between the top of the oceanic plate and the bottom of the newly formed continental **accretionary wedge.** Also commonly referred to as the plate–boundary thrust fault, the thrust interface fault, and the megathrust fault.

Surface faulting. Displacement that reaches the earth's (or sea floor) surface during **slip** along a fault. Commonly accompanies moderate and large earthquakes having focal depths less than 20 km. Surface faulting also may accompany aseismic **tectonic** creep or natural or man–induced subsidence.

Surface wave. Seismic wave that propagates along the Earth's surface. **Love** and **Rayleigh waves** are the most common.

Surface–wave magnitude. See **magnitude.**

Tectonic. Refers to rock–deforming processes and resulting structures that occur over regional sections of the lithosphere.

Teleseismic. Pertains to earthquakes at distances greater than 1000 km

Time history. The sequence of values of any quantity (such as a ground–motion measure) during the time span of an event. Also termed time series.

Transform fault. A special variety of strike–slip fault that accommodates relative slip between other tectonic elements, such as oceanic plates.

Traveltime curve. A graph of arrival times, commonly **P or S waves,** recorded at different points as a function of distance from the seismic **source.** Seismic velocities within the earth can be computed from the slopes of the resulting curves.

Tsunami. An impulsively generated sea wave of local or distant origin that results from large–scale sea– floor displacements associated with large earthquakes, major submarine slides, or exploding volcanic islands.

Tsunamigenic. Refers to those earthquake **sources,** usually along major **subduction** zone plate boundaries, such as those bordering the Pacific Ocean, that can generate tsunamis.

Tsunami magnitude (M_t). A number used to compare sizes of tsunamis generated by different earthquakes and calculated from the logarithm of the maximum amplitude of the **tsunami** wave measured by a tide gauge distant from the **tsunami** source.

Turbidites. Sea bottom deposits formed by massive slope failures where rivers have deposited large deltas. These slopes fail in response to earthquake shaking or in response to excessive sedimentation load. The temporal correlation of turbidite occurrence for some deltas of the Pacific northwest suggests that these deposits have been formed by earthquakes.

Velocity. The time rate of change of **displacement** of a reference point during an earthquake. Can be calculated by integrating an **acceleration** record once with respect to time. Expressed in centimeters per second.

Velocity structure. A generalized regional model of the Earth's **crust** that represents crustal structure by layers having different assumed seismic velocities.

Water table. The upper surface of a body of unconfined ground water at which the water pressure is equal to the atmospheric pressure.

Wavelength. The distance between successive points of equal **amplitude** and **phase** on a wave (for example, crest to crest or trough to trough).

FIGURE CAPTIONS

Figure 1. a. This map shows place names mentioned in the text. The shaded outline depicts counties noted in the text. The solid triangles indicate major Quaternary volcanoes. Maps that follow show other place names and the names of geologic features. These abbreviations are used: LC—Lynch Cove; RP—Restoration Point; FH—Friday Harbor; PR—Puyallup River; WR—White River; SD—Skagit Delta; DRV—Duwamish River Valley. b. Map showing major geologic features noted in the text. Arrows along the fracture zones show the slip directions across fault zones. The dashed arrow indicates the direction of relative convergence between the North America and Juan de Fuca plates. Ridges are major fractures in the lithosphere where plates separate and new crustal material is emitted. The fracture zones are strike–slip transform faults that accommodate differential motion between plates. The dashed line marking the continental margin is the western boundary of the main thrust fault where the oceanic plate begins to underthrust or subduct beneath the continental plate.

Figure 2. This photo shows ground shaking damage from the 1949 earthquake to unreinforced masonry in Seattle. (Photo credit: George Cankonen/Seattle Times).

Figure 3. This photo shows ground shaking damage from the 1949 earthquake to the brick veneer on a Centralia building. The parapet collapse resulted in one death. (Photo credit: A.L. Miller Collection, University of Washington Archives)

Figure 4. This photo shows landslide damage to the railbed between Olympia and Tumwater in the 1965 earthquake (Photo credit: Greg Gilbert Daily Olympian).

Figure 5. These drawings depict the Cascadia subduction zone from several different perspectives, emphasizing one or more aspects discussed in the text. Drawings a.) and b.) have been exaggerated to emphasize certain points, but c.) is drawn to scale. a.) This view from space shows the relationship between the principal geographic and geologic features in the Pacific Northwest. The shaded region is the contact zone between the North America and Juan de Fuca plates, termed the Cascadia thrust fault. The lines on this fault are drawn primarily to aid in visualization of a three–dimensional effect, but the fault may be broken in one or more sections. MFZ is the Mendocino fracture zone. b.) A roughly east–west cross section through the Earth's crust and upper mantle near the latitude of Puget Sound is depicted in this simplified drawing. The drawing shows the compression and warping of the continental plate that occurs between great earthquakes and the resulting variability in uplift and tilt that is expected as a function of geographic position. The warping of the upper plate is greatly exaggerated. The exaggerated arch in the lower plate is shown in both a.) and b.). The subducted plate has lower dip along the top of the arch, extending the thrust fault farther inland near Puget Sound than north or south of Puget Sound. The drawing also shows the potential for segmentation within the subducted plate. The segmentation may extend into the thrust fault or even into the oceanic lithosphere west of the subduction zone, but at present no evidence exists to connect potential segmentation offshore with that below the thrust fault. c.) This cross section is for a region near the latitude of Puget Sound. The drawing depicts the positions of the most significant types of earthquakes in the Pacific Northwest. The largest historic earthquakes in this region were in the Benioff zone and in the continental crust. Earthquakes in the Benioff zone occur both within the subducted oceanic crust and underlying lithosphere. In most locales along the Cascadia subduction zone, the accretionary wedge is nearly aseismic, however, a few historic earthquakes occasionally occurred in the younger sections of the accretionary wedge (age increases from west to east), and considerable seismicity occurs in the accretionary wedge west of Vancouver Island and west of the northern California coast (Figure 6). Many earthquakes also initiate in the oceanic plate in these two locales (west of the region depicted in this drawing). Some suggest that a thin section of soft sediments are subducted along the thrust fault, which are not shown in this figure. The brittle–ductile boundary, here arbitrarily drawn at a depth of 25 km, is the probable depth

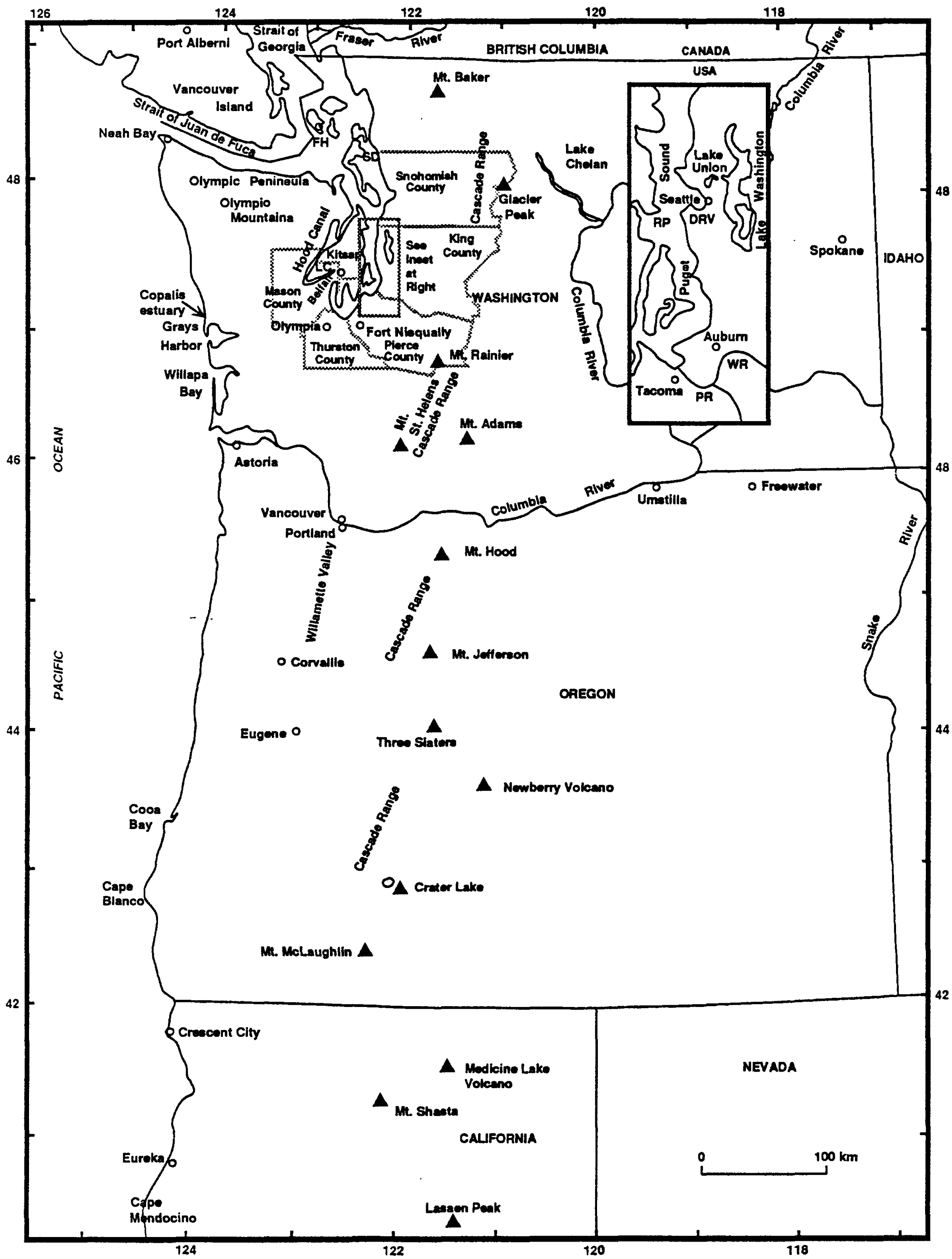
limit for seismic slip on the thrust fault and the depth limit for earthquake occurrence within the continental rocks. In this drawing, the backstop is approximately collocated with the outer arc ridge (Snavely, personal comm., 1991; Snavely in press), a zone of uplift that has elevated the sea floor. The position and dip direction of the backstop are not well determined and more than one backstop may exist.

Figure 6. A map of epicenters for the largest earthquakes in the Pacific Northwest for the period 1833–10/1991 (Ludwin and others, 1992; U.S. Geological Survey, 1991). Double lines indicate oceanic spreading ridges; solid lines and straight dashed lines indicate the approximate position of major transform faults (adopted from Riddihough, 1983); the curved dashed line indicates the inferred position of the continental margin (and inferred westernmost extent of the Cascadia thrust fault; adopted from Connard and others, 1984).

Figure 7. a.) This map shows epicenters with well–constrained depths (Ludwin and others, 1991) and the location of cross section AA'. b.) This cross section AA' shows earthquake hypocenters relative to the position of the top of the subducting oceanic plate inferred by Mooney and Weaver (1989).

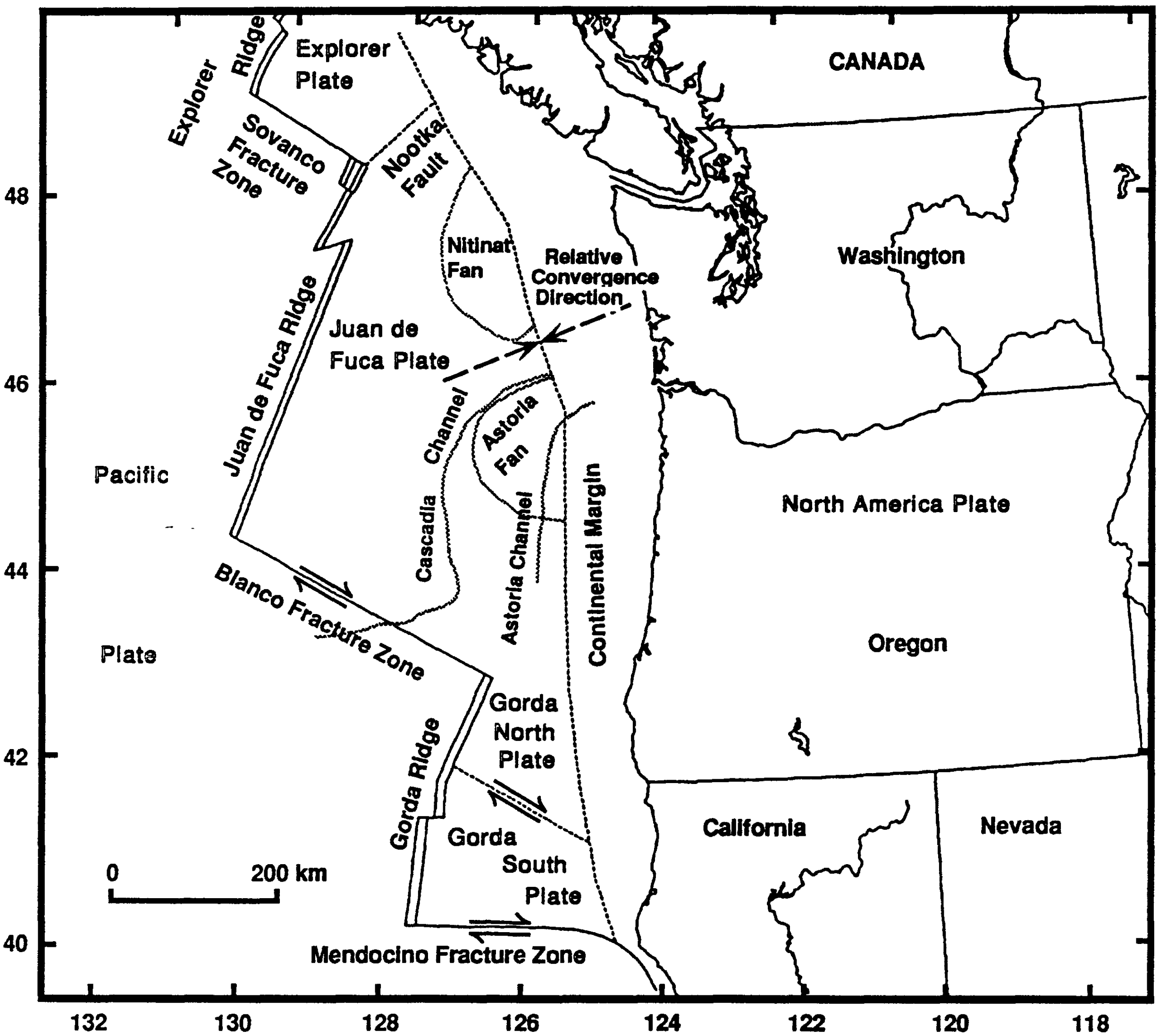
PLATE EXPLANATION

Plate 1. This map shows known or suspected faults with Quaternary displacements in the Pacific Northwest. The numbers accompanying structures are indexed to the principal references. Offshore faults shown on this map do not necessarily cut reflectors interpreted to be Quaternary in age because ruptures on blind thrusts do not reach the surface (S.H. Clarke, personal communication, 1991). Inclusion of a structure on this map does not necessarily imply that it is seismogenic. Many of these faults, particularly offshore, probably represent secondary rupture, possibly during subduction zone events. Map reference key: 1.) Wagner and Tomson (1987); 2.) Wagner and others (1986); 3.) Qamar and Zollweg (1990); 4.) Gower and others (1985); 6.) Walsh and Logan (1985); 7.) Howard and others (1978); 8.) Wilson and others (1979); 9.) Crosson and Frank (1975); 10.) Grant and Weaver (1986); Weaver and Smith (1983); 11.) Yelin and Patton (1991); 12.) Malone and others (1978); 13.) Shaffer and West (1989); 14.) West and Shaffer (1989); 15.) Tolan and Reidel (1989); U.S. Department of Energy (1988); 16.) Campbell and Bentley (1981); 17.) Goldfinger and others (1992); 18.) Madin and Pezzopane (1990); 19.) Geomatrix Consultants (1990); 20.) Hawkins and others (1989c); 21.) Pezzopane and Weldon (1990); 22.) McInelly and Kelsey (1990); 23.) Hawkins and others (1989a); 24.) Sherrod (1989); 25.) Geomatrix Consultants (1989); 26.) Hawkins and others (1989b); 27.) Hemphill–Haley and others (1989); 28.) Clarke (1990); 29.) Clarke (1992); 30.) Clarke and Carver (1992); 31.) Jennings (1985).



126
124
122
120
118
Port Alberni
Strait of Georgia
Fraser River
BRITISH COLUMBIA
CANADA
USA
Vancouver Island
Strait of Juan de Fuca
Neah Bay
FH
Mt. Baker
Olympic Peninsula
Olympic Mountains
Snohomish County
Cascade Range
Glacier Peak
Lake Chelan
Hood Canal
Kitsap
Mason County
See inset at Right
King County
WASHINGTON
Copalis estuary
Grays Harbor
Olympia
Fort Nisqually
Thurston County
Pierce County
Mt. Rainier
Willapa Bay
Mt. St. Helens
Cascade Range
Mt. Adams
Columbia River
Sound
Lake Union
Seattle
RP
DRV
Lake Washington
Puget
Auburn
WR
Tacoma
PR
Spokane
IDAHO
OCEAN
PACIFIC
Astoria
Umatilla
Freewater
Vancouver
Portland
Columbia River
Mt. Hood
Willamette Valley
Cascade Range
Mt. Jefferson
Corvallis
OREGON
Snake River
Eugene
Three Sisters
Newberry Volcano
Coos Bay
Cascade Range
Cape Blanco
Crater Lake
Mt. McLaughlin
Crescent City
Medicine Lake Volcano
NEVADA
Mt. Shasta
CALIFORNIA
0
100 km
Eureka
Cape Mendocino
Lassen Peak
48
46
44
42

Figure 1a



Explorer Ridge
Explorer Plate
CANADA
Sovanco Fracture Zone
Nootka Fault
Nitinat Fan
Relative Convergence Direction
Washington
Juan de Fuca Ridge
Juan de Fuca Plate
Cascadia Channel
Astoria Fan
Pacific
Plate
North America Plate
Astoria Channel
Continental Margin
Blanco Fracture Zone
Oregon
Gorda North Plate
Gorda Ridge
Gorda South Plate
California
Nevada
0
200 km
Mendocino Fracture Zone
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Figure 1b

Figure 2

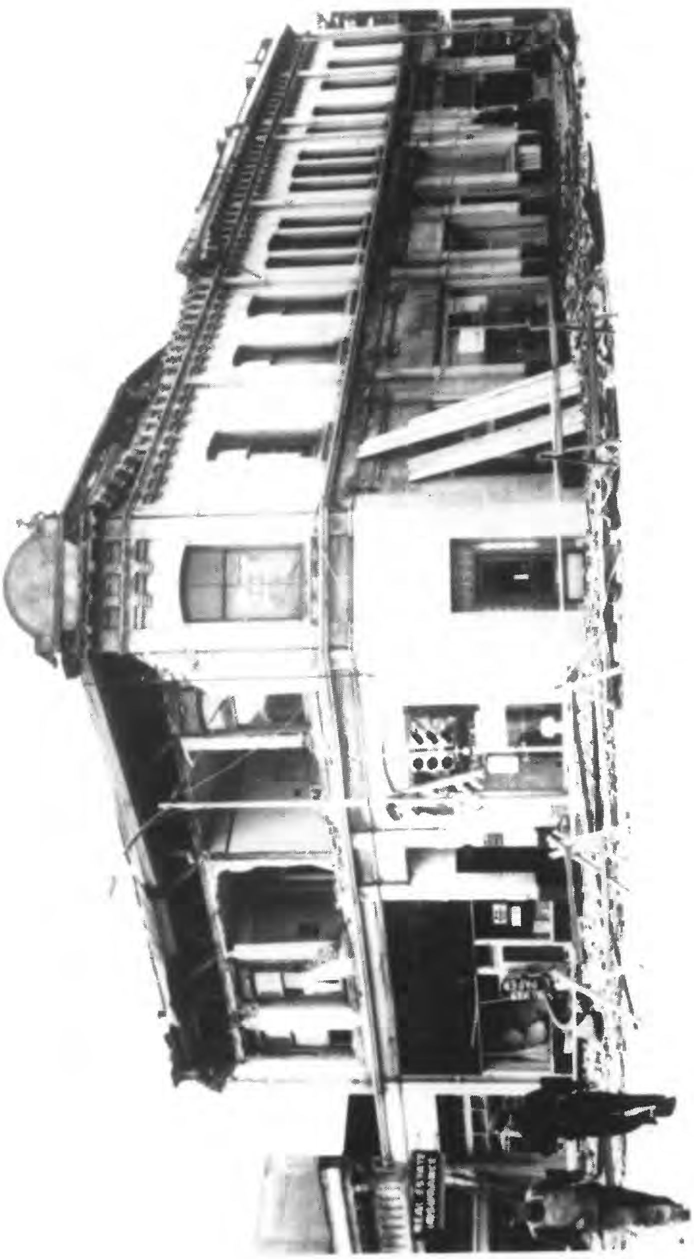

Figure 3

Figure 4

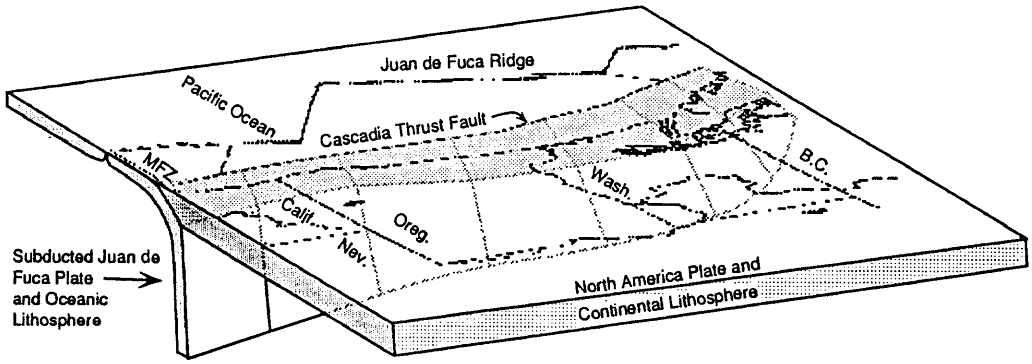

Figure 5a

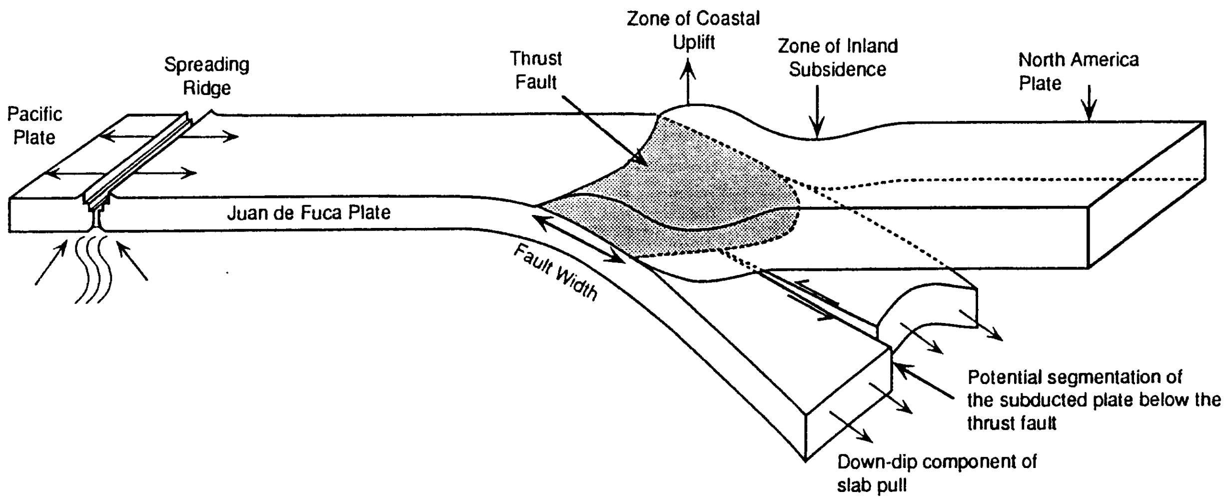

Figure 5b

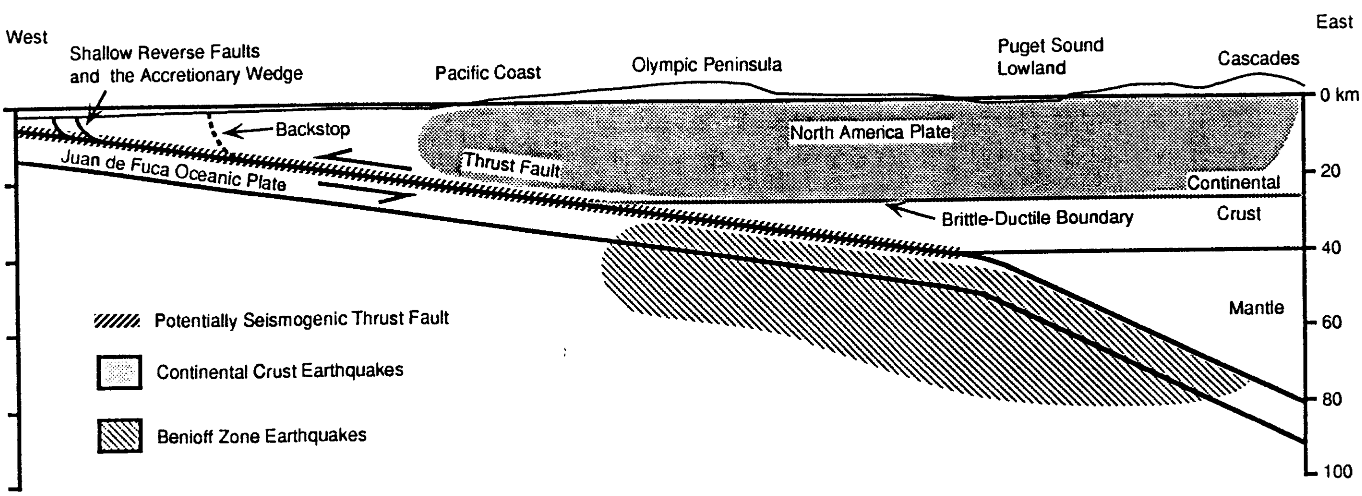

Figure 5c

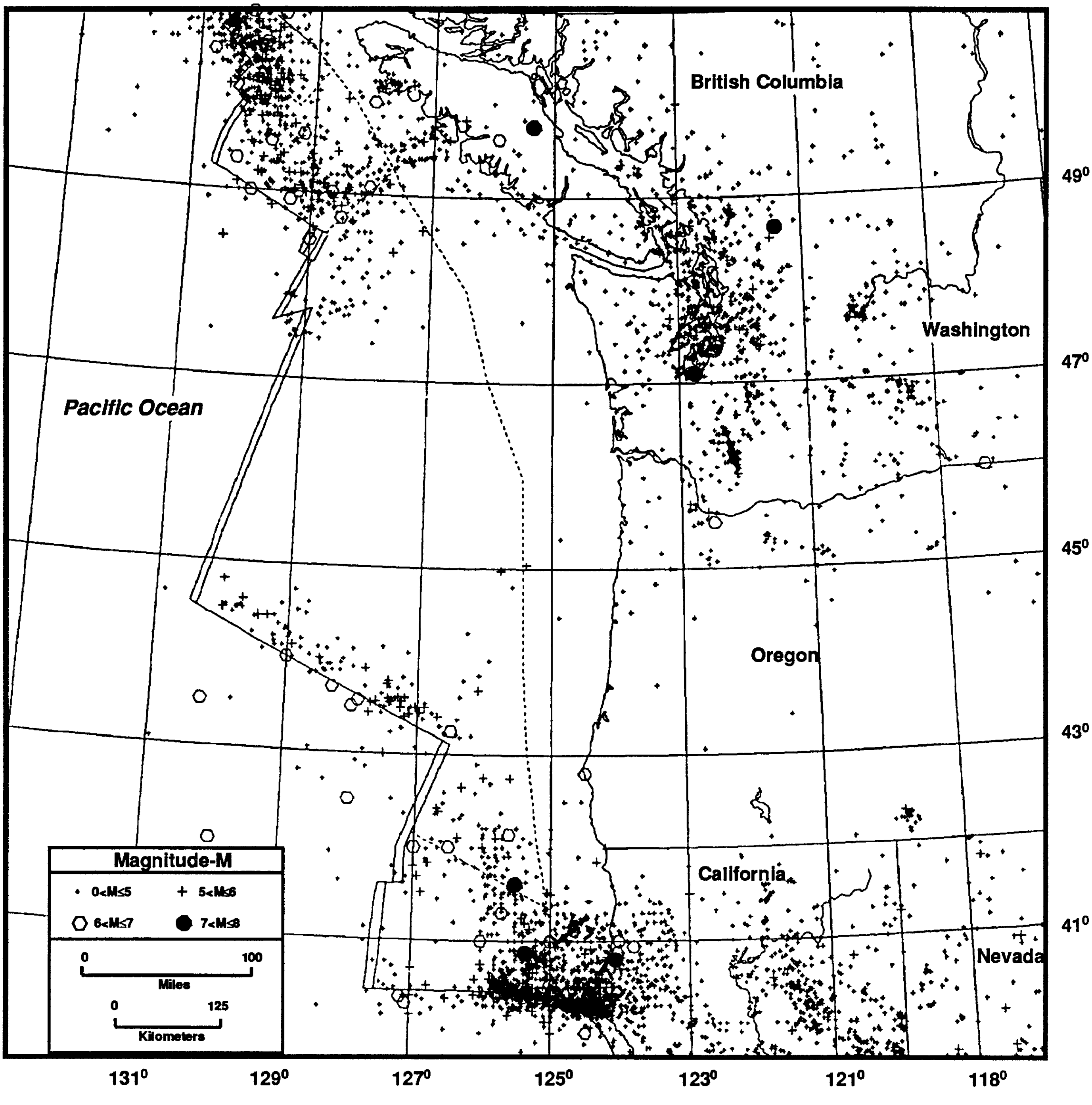
British Columbia
Washington
Pacific Ocean
Oregon
California
Nevada
49°
47°
45°
43°
41°
131°
129°
127°
125°
123°
121°
118°
Magnitude-M
0<M≤5
5<M≤6
6<M≤7
7<M≤8
0
100
Miles
0
125
Kilometers

Figure 6

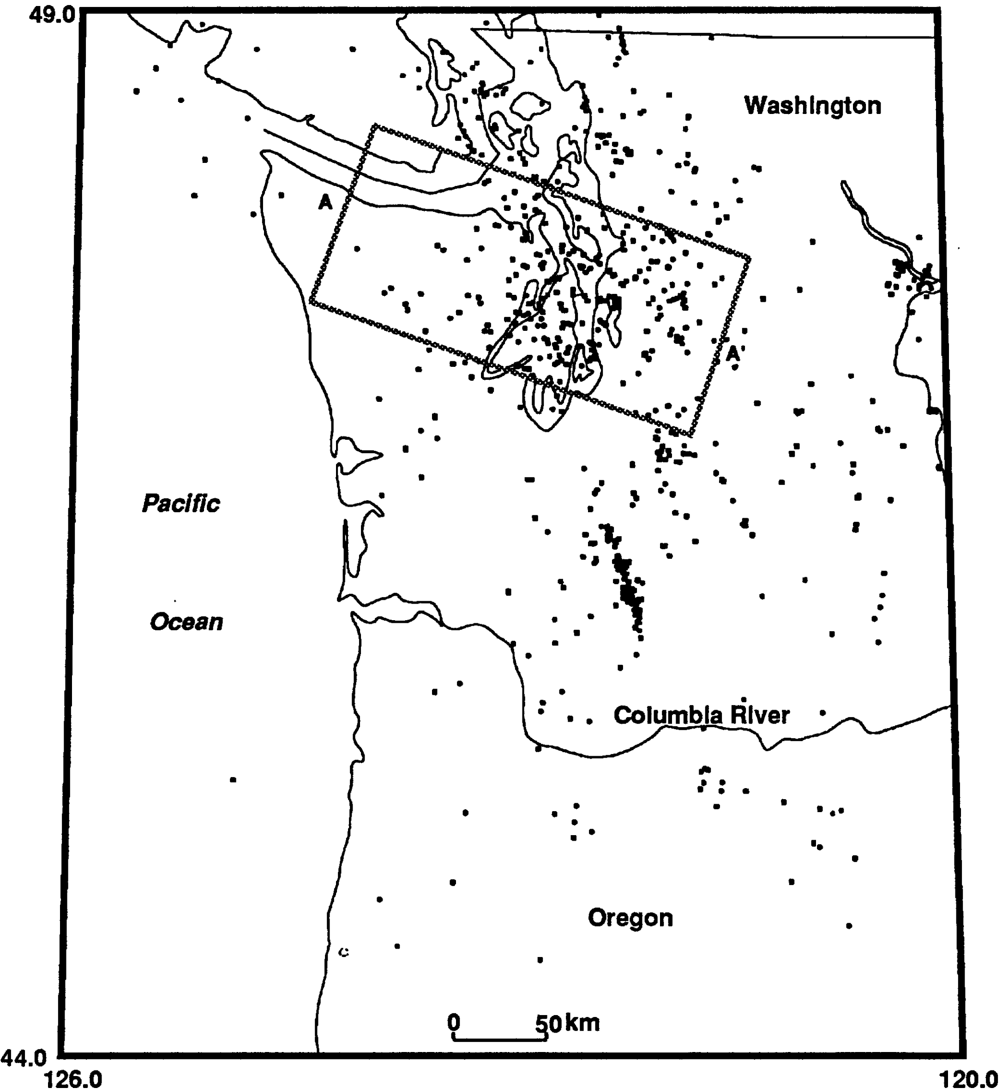
49.0
Washington
A
A'
Pacific
Ocean
Columbia River
Oregon
0
50km
44.0
126.0
120.0

Figure 7a

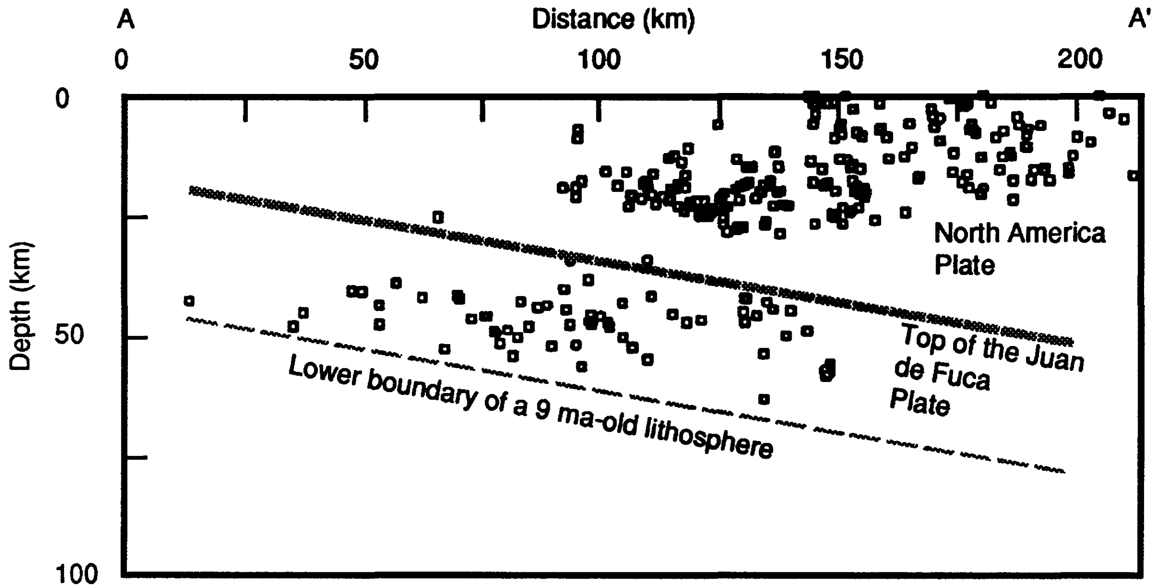
A
Distance (km)
A'
0
50
100
150
200
0
50
100
Depth (km)
North America Plate
Top of the Juan de Fuca Plate
Lower boundary of a 9 ma-old lithosphere

Figure 7b