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Earthquake Hazards in the Pacific Northwest of the United States

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ESTIMATES OF SEISMIC SOURCE REGIONS FROM CONSIDERATIONS OF THE EARTHQUAKE DISTRIBUTION AND REGIONAL TECTONICS IN THE PACIFIC NORTHWEST

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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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ABSTRACT

The tectonic and geologic setting of the Pacific Northwest is that of an active subduction zone, and west of the Cascade volcanic arc, there are three distinct earthquake source regions: 1) at the interface between the Juan de Fuca and the North American plate; 2) within the subducting Juan de Fuca plate, and 3) within the crust of the overlying North American plate. The record of historical seismicity in the region has few events of magnitude 7 or greater; all of these events are thought to be within the subducting portion of the Juan de Fuca or Gorda plates, within the crust of the North American plate, or offshore of northern California in the flat-lying section of the Gorda plate. This limited historical record is used for nearly all estimates of the earthquake hazards in the region. Unfortunately, the distribution of seismicity determined using the modern seismic network is not in accord with where many of the largest historical earthquakes occurred.

The best understood source zone is for those earthquakes that occur intraplate, within the subducting Juan de Fuca and Gorda plates. Most of the damaging earthquakes in the historical record of the Pacific Northwest have been in this zone and knowledge of these events has served as the cornerstone for earthquake hazards assessments in the Puget Sound region. Generally, these events are thought to be caused by gravitational forces within the plate, and that the events typically occur where the dip of the subducting plate increases from about 10° to more than 25°. Interpretation of the current available data, including earthquake hypocenters and geophysical imaging of the subducting plates, indicates that these events should be expected along the strike of the entire subduction zone. Because of the known plate geometry most of these events are expected to be at depths of 45 to 60 km and have magnitudes as least as large as that already experienced in the region (7+). The rate of occurrence of these events is unknown, but in Oregon and Washington there are 6 events known since 1870 that have estimated magnitudes of 6 or greater.

There are no known earthquakes recorded by modern seismic networks on the interface between the two plates and in the history of the European population in the area no known large event can be associated with the interface. However, the late Holocene geologic record of subsidence in coastal intertidal marshes provides evidence consistent with the occurrence of great subduction-style earthquakes. The available geologic record suggests that the return period for these events is irregular, varying from as frequent as 100 years to longer than 1200 years; the length of the coast interpreted to have subsided during a particular event is sufficient to have generated earthquakes of magnitude 8 or greater. Some studies have suggested that the entire coast from northern California to central British Columbia could fail in a single event of magnitude 9 or greater. Despite the fact that there are many unknown details concerning intraplate events, the general understanding of large-scale plate processes allows a reliable estimate of the potential source zone for subduction earthquakes. The source zone for these events is very great, including the entire coast from the deformation front offshore to about the coastline.

Finally, the source zone for crustal events is very poorly known. The historical record includes events greater than magnitude 7 in central Vancouver Island, the North Cascades, and offshore Northern California. Until the tectonic causes of these events are better understood, it is not possible to reliably determine the source zones for these events. Considerable effort is needed to resolve the geological and tectonic details related to these events.

INTRODUCTION

The first step in assessing earthquake hazards and risks within a region requires that the nature and distribution of earthquake sources be understood. During the past ten years there has been considerable effort in the Pacific Northwest focused on determining the distribution of earthquake sources and on placing these sources within the regional tectonic setting. Two examples illustrate how successful these studies have been in changing the assessment of earthquake hazards in the Pacific Northwest. First, in 1975, Hopper and others (1975) suggested that the Juan de Fuca plate might be attached to the North American plate, thereby making it unlikely that great thrust earthquakes (of magnitude 8 or larger) would occur on the interface between the two plates. Fifteen years later, most earth scientists routinely accept convergence between the two plates, and as a consequence, most further accept the interpretation that subduction zone earthquakes on this interface will be at least as large as magnitude 8.0 (e.g., Heaton, 1990); some scientists argue for earthquakes greater than magnitude 9.0 (Heaton and Hartzell, 1987). The most compelling evidence that has caused this change in earthquake hazards assessments is found in the intertidal marshes along the coast. At some sites in Willapa Bay on the Washington coast, at least 8 discrete episodes of subsidence of the intertidal marsh surface are recorded in the geologic section of the last 4000 years (Grant and others, 1989). Each episode of subsidence is thought to have been the result of a subduction zone earthquake (Atwater, 1987), thus subduction events appear frequently in the geological record (see Atwater, this volume; Rogers and others, this volume, for more discussion).

As a second example, Perkins and others (1980) issued a revised estimate of the maximum horizontal ground acceleration for Oregon and Washington in which the maximum expected earthquake magnitude was estimated for subregions of the two states. In the southern Washington Cascade Range near Mount St. Helens, the estimate used for the maximum expected magnitude was 5.1. By 1985 the St. Helens zone had been identified, and several studies had concluded that a maximum magnitude event of 6.8 should be adopted for engineering design purposes in the region of the St. Helens zone (Grant and Weaver, in press). Both examples of changes in the assessment of earthquake hazards relied heavily on understanding the regional tectonic framework, and in the second case, the change was nearly completely dependent on seismicity studies.

Because of the subduction zone regime, there are three distinct sources of earthquakes in the Pacific Northwest: 1) crustal earthquakes that occur within the overriding North American plate, 2) intraplate earthquakes that occur within the subducting Juan de Fuca and Gorda plates, and 3) interplate earthquakes that are expected to occur at the interface between the Juan de Fuca (and Gorda) plate and the North American plate (subduction or thrust events). West of the Cascade Range the distribution of earthquake source types reflects a combination of the geometry of the subducting Juan de Fuca and Gorda plates, crustal structure within the overriding North American plate, and tectonic interactions among the North American, Pacific, and Juan de Fuca plates. East of the Cascades it is likely that the earthquake distribution reflects the tectonics and structure of only the North American plate.

Of the three source types, crustal earthquakes in the North American plate and events within the subducting plate (we will refer to these as intraplate events or Benioff zone earthquakes) have been the basis of earthquake hazard assessments for the Pacific Northwest (e.g., Algermissen, 1988). In Washington and Oregon, for example, where the historical record is thought to be complete since the 1870's at the magnitude 6 and greater level (Ludwin and others, 1991), there were two events that certainly occurred in the crust and six earthquakes that are either considered or known to have occurred within the subducting plate. One of the anomalies of the Cascadia subduction zone is that there are no large historical earthquakes that are thought to have their source on the interface. In most subduction zones it is this interface that produces the great (magnitude 8+) thrust events like the earthquake that struck the Prince William Sound area of southern Alaska in 1964. Recently, efforts have been made to incorporate at least the possibility of great thrust zone earthquakes into the regional hazard analysis (Algermissen, 1988).

This paper focuses on the extent of the three source regions for the Cascadia subduction zone. In defining the source regions, we have relied on recent compilations of earthquake catalogs for Oregon and Washington, studies of regional seismotectonics, investigations of coastal marsh stratigraphy and determinations of the plate geometry. It is clear that intraplate earthquakes, the most frequently observed of the large magnitude events (6+) in the historical record of Oregon and Washington, are understood well enough that the source region expected to produce events in the future can be specified with great confidence. Despite uncertainty surrounding the details of how and when great subduction zone thrust events may occur on the interface, there is clearly a growing acceptance of the past occurrence of these events. As the general forces that produce these events are understood from comparative studies with other subduction zones, it is possible to define the potential source regions fairly accurately. Finally, because the causes of the large magnitude crustal events in the historical record remain obscure, the portion of western Washington and Oregon which may be subject to large magnitude crustal earthquakes remains uncertain. As noted elsewhere (Rogers and others, this volume; Shedlock and Weaver, 1989), narrowing the uncertainty surrounding the occurrence of great subduction zone events and determining whether the urban centers in the Puget Sound basin and the Willamette Valley as well as those in the Columbia Plateau are subject to magnitude 7, shallow (<20 km) crustal events will require significant investments in time and effort for new experiments, research and modeling.

Our paper differs from a number of recent review articles on the earthquake distribution in this region in that we provide an overview of the earthquake distribution of the entire Juan de Fuca--North American plate system. Our plots of seismicity combine the reviews of Ludwin and others (1991) who summarized seismicity in Oregon and Washington, Uhrhammer (1991) and Hill and others (1991) who discussed seismicity of northern and central California, respectively. To complete our historical perspective, we combined the catalog of earthquakes greater than magnitude 6 provided by Ellsworth (1990) for California with similar sized events listed by Ludwin and others (1991) for Washington and Oregon and events listed by Rogers (1983a) for southern British Columbia. The scope of our review broadens that of Weaver and others (1990) who discussed the crustal earthquake distribution associated with the Cascade Range from Lassen Peak to Mount Baker in Washington.

Finally, a guide concerning magnitudes in this paper. Unfortunately for most readers, seismologists use several magnitude scales to measure the strength of an earthquake at its source. (Magnitudes are distinct from the earthquake shaking effects at discrete sites that are measured by intensities). The different magnitude scales reflect the fact that most scales are appropriate for only a small portion of the wide range of possible magnitudes. Although we have given the appropriate magnitude scale for particular earthquakes (e.g., local magnitude, M_L; body-wave magnitude, m_b; surface-wave magnitude, M_S; moment magnitude, M_w; or coda magnitude, M_c) in an effort to avoid confusion we have tried to de-emphasize the differences between the various scales and often use only the term magnitude. Generally, magnitudes greater than 8 are determined using seismic moment, magnitudes greater than about 5 but less than 8 are calculated using either body or surface waves, magnitudes less than about 5.5 usually are determined using a Wood-Anderson type of "local" magnitude, and magnitudes less than about 4.5 tend to be calculated from the coda length of the event. More discussion of the magnitude scales and definitions of seismological terms can be found in the appendix to this volume.

EARTHQUAKE DISTRIBUTION

One of the significant problems in assessing the earthquake hazards of the Pacific Northwest is the discordance between the historical earthquake record and that available from modern seismic instrumentation (from about 1950). The historical record is essential because in Washington and Oregon only the 1965 event near Seattle has been greater than magnitude 6 since the World Wide Standard Seismic Network began recording in 1962; since then in California north of Cape Mendocino all events greater than magnitude 6 have been offshore in the Gorda plate. Thus, one goal of any earthquake hazards assessment program must be to place the historical, larger magnitude earthquakes in the tectonic framework. This task is made more difficult in western Oregon and Washington by the lack of known surface faults with Pleistocene to Holocene offsets (Rogers and others, this volume).

EARTHQUAKES GREATER THAN MAGNITUDE 6

Earthquakes estimated to be larger than magnitude 6 since 1870 are restricted to a relatively small portion of the Pacific Northwest (fig. 1). In plotting the earthquakes in figure 1 we have not attempted to include earthquakes along the offshore ridge system other than in the vicinity of the South Gorda Plate. As the distance between the continental area and the offshore ridges and fracture zones increases it becomes increasingly likely that not all events greater than magnitude 6 will be felt or noted. For the continental areas of the northwestern United States the seismic catalogs are thought to be complete at the magnitude 6 level since 1870 (Ludwin and others, 1991; Ellsworth, 1990); in British Columbia the catalog of magnitude 6 events is complete since 1899 (G. Rogers, written communication, 1990).

At the southern end of the Juan de Fuca-North American plate system, these larger events have occurred both onshore and offshore of northern California and to the east of Lassen Peak in the Basin and Range province. Large events are found in the historical record from northwestern Washington and central Vancouver Island; they are notably absent over nearly all of Oregon (fig.1). Within this distribution of earthquakes, the most important events that must be incorporated into hazard assessments are those earthquakes greater than magnitude 7 (eight events total excluding the 2 events in Nevada) and a few events of magnitude less than 7.

The 1872 North Cascades earthquake is generally considered the largest earthquake known in the Pacific Northwest, (Milne, 1956), with an estimated felt area magnitude of 7.4 (Malone and Bor, 1979). It was felt over an area of over 1,010,000 square km, including Washington, northern and central Oregon, northern Idaho, western Montana, and southern British Columbia. The earthquake was followed by an extensive aftershock sequence (Milne, 1956). Study of damage reports suggests that the maximum intensity exceeded VII and may have been as high as IX on the Modified Mercalli (MM) scale (Milne, 1956). Because the available historical data do not unequivocally allow the location and depth of the 1872 earthquake to be determined, both remain subjects of controversy. The location shown in figure 1 was determined by Malone and Bor (1979) based on the intensity pattern; these authors also summarized other suggested locations of the event. Hopper and others (1982) have suggested a shallow crustal depth based upon intensity contours and the extensive aftershock sequence. All instrumentally located earthquakes (since 1970) near the epicenter proposed by Malone and Bor (1979) are shallower than 25 km.

The 1949 south Puget Sound earthquake (M_S=7.1) is one five earthquakes (in 1909, 1939, 1946, 1949, and 1965) of magnitude 6 or greater known to have occurred in the Puget Sound basin (fig. 1). The 1949 and the 1965 (m_b=6.5) earthquakes caused significant damage in the Puget Sound region (Murphy and Ulrich, 1951; Nuttli, 1952; Algermissen and others, 1965); in fact these two earthquakes have caused most of the major earthquake damage in the Pacific Northwest. These two events have instrumentally determined hypocentral depths of 54 km and 60 km, respectively (Baker and Langston, 1987; Algermissen and others, 1965). No aftershocks were felt or recorded after the 1949 earthquake; instrumentation available at the time would have detected events larger than magnitude 4.5 (M_L). Similarly, following the 1965 earthquake, no aftershocks were felt, and an examination of seismograms recorded on stations operating within the region failed to identify any aftershocks greater than magnitude 2.5 (M_c). Because these large earthquakes have not produced felt aftershocks, Rogers (1983a) noted that this provided one possible way of estimating whether historical earthquakes were within the crust (where felt aftershocks would be expected) or within the subducting plate.

Large earthquakes occurred in central Vancouver Island in 1918 and 1946. The 1918 event was about magnitude 7 and had an extensive aftershock sequence (Cassidy and others, 1988). The 1946 event of magnitude 7.3 (M_S) occurred in the mid to lower crust (about 15 to 25 km) and had almost no aftershocks (Rogers and Hasagawa, 1978). Because of the sparse population of central Vancouver Island at the times of the earthquakes the value of the damage was not large. As these events are within the crust of the North American plate (Cassidy and others, 1988), they raise serious questions as to the possibility of similar events in the heavily populated urban areas in the Pacific Northwest.

Large earthquakes off the northern California coast have been frequent during the last 120 years (fig. 1) and have done structural damage to onshore facilities. Four events are estimated to have been of magnitude 7 or greater; the largest offshore event of felt-area magnitude 7.3 occurred in 1923 and shaking from this shock damaged a number of chimneys in the small towns north of Cape Mendocino (Toppozada and others, 1981; Ellsworth, 1990). In 1980 a magnitude 7.0 (M_S) event occurred within the Gorda plate, just west of the deformation front (fig. 1). Although there was relatively little damage to structures, a highway overpass collapsed with five people on the bridge; fortunately no one was killed (Simon, 1981). The most damaging earthquake in northwestern California to date is the 1954 Eureka event (fig. 1). That earthquake was estimated to be of magnitude 6.5 (Ellsworth, 1990); one person was killed and there was considerable non-structural damage.

There are two other notable earthquakes in the historical record. An earthquake estimated to have a felt-area magnitude of at least 6.75 (Toppozada and others 1981) occurred in 1873 near the Oregon-California border (fig. 1). The earthquake was felt from near Portland, Oregon to Sacramento, California, but was felt with MM intensity VIII only in a small area near the estimated epicenter (Toppozada and others, 1981). There is no mention of felt aftershocks accompanying the mainshock in newspaper accounts of the time. This observation suggests that either the event was sufficiently far offshore that any aftershocks were not felt or, that it was deep, possibly within the subducting Gorda plate where aftershocks are not necessarily expected. The second event is the 1936 Milton-Freewater earthquake (fig. 1), the largest known event in eastern Washington. Maximum intensity was reported as MM VII (Coffman and others, 1982), and the magnitude was calculated by Gutenberg and Richter (1954) to be 5.75 (M_S). Noson and others (1988) estimated a magnitude of 6.4 for this earthquake based on the felt area. Because all earthquakes in eastern Washington located since 1970 occurred in the crust, the 1936 event is also assumed to be crustal. Numerous aftershocks were felt.

Losses from earthquakes occurring in the Pacific Northwest are dominated by those suffered during the 1949 and 1965 events. Eight people were killed in the 1949 event (Ulrich, 1949) and six died in the 1965 earthquake (Algermissen and others, 1965). In terms of 1984 dollars, $150 million of damage occurred in 1949; the 1965 event did an estimated $50 million of damage (Noson and others, 1988). The effects of the 1949 and 1965 earthquakes are currently used as the basis for engineering design, emergency response planning, and damage and loss estimates in the Puget Sound region (Hopper and others, 1975). Since the 1975 report was issued, the population of the Puget Sound region has increased by more than 600,000 people and many new buildings have been constructed. More importantly, the 1975 study considered only the effects of deep earthquakes, and did not consider the effects of either a shallow crustal earthquake (like the 1872 event) or the possible effects of a great megathrust earthquake. Thus, there is wide agreement in the scientific community that an updated discussion of the sources, the hazards posed by these sources, and revised estimates of earthquake losses are urgently needed in the Pacific Northwest.

SUMMARY OF INSTRUMENTAL SEISMICITY

Although a few seismic stations have operated in the Pacific Northwest since 1900 (see Rogers, 1983a; Ludwin and others, 1991), 1960 is usually assumed as the beginning of the instrumental period for the region. By 1960 the combination of the distribution of seismic stations and population density allowed the reporting and location of earthquakes greater than magnitude 4 (Ludwin and others, 1991). Ludwin and others concluded that since 1960, for Washington and Oregon, the earthquake catalog is complete above magnitude 4.0 (M_L); in California the catalog is complete at this magnitude level since at least 1960 (Uhrhammer, 1991). Since 1960 the number of seismic stations operating in the Pacific Northwest has greatly increased so that by 1990 much of the region was routinely monitored for earthquake activity less than magnitude 2.5 (M_c). Details of the evolution of the network can be found elsewhere (e.g., Ludwin and others, 1991; Weaver and others, 1982, 1990; Rogers, 1983b). The distribution of short-period (1 Hz) seismic stations operating in late 1990 is shown in figure 2. Most of the stations in British Columbia are operated by the Geological Survey of Canada, stations in Oregon and Washington are nearly all operated by the University of Washington, and most of the stations in California are part of the U.S. Geological Survey network. A few stations were installed in southwestern Oregon during the late summer and fall of 1990 by the University of Washington (figure 2).

For a plate-wide perspective of instrumental seismicity, we selected well-located earthquakes that have occurred since 1980 greater than magnitude 2.0 (M_c) and that met the following statistical criteria: calculated hypocentral standard errors of less than ±3 km, at least 6 P-waves used in the solutions, and a RMS of the traveltime residuals less than 0.35 seconds. Most hypocenters have statistics considerably better than these criteria. For Washington and Oregon our hypocentral data are taken from the University of Washington catalog and for California the data are from catalogs maintained by the U.S. Geological Survey. Hypocentral catalogs compiled by the Geological Survey of Canada for British Columbia are incorporated into the University of Washington catalog.

In plotting the combined catalog (fig. 3 and 4), we have subdivided our data into two depth ranges, 0-30 km and deeper than 30 km. Earthquakes greater than 30 km depth are nearly all within the subducting Juan de Fuca plate system (fig. 3). Earthquakes less than 30 km depth are within the North American plate in Washington or Oregon; however, along the northern California coast most of the shallow seismicity is probably within the subducting Gorda plate (Hill and others, 1991; Eaton, 1989). Inland from the California coast the shallow earthquakes are within the North American plate.

It is clear that the distribution of both the shallow and deep earthquakes is not uniform across the Pacific Northwest. The deep events are concentrated beneath northwestern Washington and northwestern California, with a sparser distribution of events beneath southwestern Washington and northern Oregon (fig.3). The lack of deep events beneath the southwestern Oregon coast may result in part from a lack of seismic stations (compare fig. 3 with fig. 2); however, as noted earlier, earthquakes greater than magnitude 4 would have been instrumentally recorded since about 1960. Events of this magnitude would have been felt along the Oregon coast throughout the twentieth century, so there is little doubt that at the magnitude 4 level this portion of the subducting Juan de Fuca and Gorda plates is seismically quiet.

The distribution of crustal earthquakes is likewise not uniform across the Pacific Northwest (fig. 4). Weaver and others (1990) used the changes in the crustal seismicity pattern to divide the Cascade Range and adjacent areas into four regions (termed segments); here we expand those segments westward to the coast. In the northernmost segment, from Mount Baker to Mount Rainier, nearly all of the well-located crustal earthquakes are confined to the region between the eastern Olympic Mountains and the Quaternary stratovolcanoes at the western edge of the North Cascades (fig. 4); there are few events within the North Cascades. The second segment defined by Weaver and others (1990) from Mount Rainier to Mount Hood, is the most seismically active portion of the crust. In southern Washington the St. Helens zone (SHZ) is a particularly prominent north-northwesterly striking alignment of earthquakes. Although there is an hiatus of activity east of the SHZ, in general seismicity continues in a broad zone from the area immediately west of the SHZ into southeastern Washington (fig. 4). The third segment from south of Mount Hood to just north of Mount Shasta is seismically very quiet (fig. 4). This segment has not been monitored continuously, but very few earthquakes were observed during the two years (1980-1982) of continuous operation of a 32-station network or since stations were re-installed in the central Oregon Cascade Range in 1987 (Ludwin and others, 1991). There is a marked increase in earthquake activity in the fourth segment between Mount Shasta and Lassen Peak and also along the northern California coast (fig. 4).

CRUSTAL THICKNESS AND PLATE GEOMETRY

The composition of the crust of the Pacific Northwest has been investigated using geologic and geophysical methods, but there are few reversed, high-resolution refraction or wide-angle reflection profiles in the region. Mooney and Weaver (1989) summarized the existing studies with a contour map of estimated crustal thickness beneath Washington, Oregon, and northern California (fig. 5). The sparse number of seismic lines emphasizes the need for deep seismic control in many areas. Using this map as a working hypothesis for the configuration of the Moho, there are two noteworthy features.

The first is the pronounced eastward increase in crustal thickness from 16 km at the continental margin to about 40 km beneath the western flank of the Cascade Range. Gravity modeling along two profiles in Oregon and preliminary interpretations of electrical and magnetic data collected by the EMSLAB experiment (EMSLAB Group, 1988) along a profile perpendicular to the northern Oregon coast are consistent with crustal thickening. The contours beneath the Klamath Mountains in southwestern Oregon and northwestern California lack seismic control. Nevertheless, the known thin oceanic crust and thick Cascade Range crust support the general trends represented by the contours.

The second major feature of the crustal thickness map is the presence of thick crust beneath the Cascade Range, the Puget Sound basin, and the Columbia Plateau (fig. 5). Crustal thickness is estimated to be at least 38 km over this entire region, and locally reaches 46 km in the southern Oregon Cascade Range. A gradual eastward thinning of the crust occurs beneath the Basin and Range of southeastern Oregon and northeastern California (fig. 5). The Moho shallows beneath the Okanogan Highlands in northeastern Washington, where a reflection profile (line 5 in fig. 5) has been interpreted as indicating a flat Moho at about 36 km depth beneath most of this province (Potter and others, 1986).

The geometry of the subducting Juan de Fuca and Gorda plates has been partly inferred from the location of earthquakes occurring within the plates; often these events are referred to as Benioff zone earthquakes. In the historical record of large magnitude events, it is these events that are most frequently observed. Disregarding the concentration of earthquakes offshore of Cape Mendocino, since 1870 there are at least six large earthquakes that are either known or thought to be within the subducting plate. These events occurred in 1873, along the coast near the Oregon-California border, and in 1909, 1939, 1946, 1949, and 1965, within the Puget Sound basin (fig. 6). The distribution of intraplate earthquake hypocenters (largely those shown in fig.3), indicates that the subducting Juan de Fuca plate arches upward beneath southern and central Puget Sound; beneath southwestern Washington, the plate dips to the east-southeast, changing to a northeast dip beneath northwest Washington (Weaver and Baker, 1988). The structure of the arch can be seen in figure 3, where we have indicated the eastern extent of the location of hypocenters at 60 km depth (from Weaver and Baker, 1988). Although much of the shallow plate geometry beneath Oregon is either poorly resolved or unknown, it appears that the average dip of the plate (between the deformation front and 60 km) must be greater in northern Oregon than in Washington.

In Washington and northern California there is general agreement between areas of the thickest crust (>38 km) and areas of crustal seismicity (compare fig. 4 & 5). Clearly, however, there are areas where this association does not hold, particularly from the Oregon Cascades eastward. Thus, the segmentation of the crustal seismicity pattern is probably not directly related to crustal thickness. Unfortunately, the structure of the crust is too poorly known in this region to determine if the composition of the mid-to-upper crust may be fundamental in determining areas of crustal seismicity.

Nor does the segmentation of the crustal seismicity pattern match the variation of the geometry of the Juan de Fuca plate. The crustal seismicity in the Puget Sound area lies above the arch in the Juan de Fuca plate, yet relatively sharp variations in the distribution of the earthquakes argue for the importance of unknown crustal structure rather than variations of the subducting plate geometry. Guffanti and Weaver (1988) noted that the high seismicity segment, between Mount Rainier and Mount Hood, was on the southeastern edge of the plate arch, whereas the seismically quiet segment in Oregon was inland from the section of the Juan de Fuca plate that has an increase in plate dip. However, in suggesting these associations, Guffanti and Weaver (1988) did not incorporate the historical seismicity. In particular, when the 1872 earthquake is considered, it is clear that seismic moment release in the North Cascades dominates all of the crustal seismicity south of the Canadian border (Shedlock and others, 1989).

This last observation points to an important difference between seismicity in Oregon and Washington and that throughout California (including that in the subduction regime). In Oregon and Washington seismicity shows a temporal variation on the scale of at least a few decades (Ludwin and others, 1991), whereas in California Hill and others (1991) have noted that a few years of microearthquake monitoring coupled with known mapped faults provides a good representation of the long-term (decades to a few centuries) seismicity pattern.

ESTIMATES OF THE SOURCE REGIONS

PROBABLE ZONE FOR INTRAPLATE EARTHQUAKES

The geometry of the subducting plate (summarized in fig. 3) allows the occurrence of the large earthquakes in the historical record (e.g., 1949, 1965) to be related directly to the plate configuration (Weaver and Baker, 1988). The T-axis from the focal mechanism calculated by Baker and Langston (1987) for the 1949 south Puget Sound earthquake of magnitude 7.1 (M_S) is oriented to the east-southeast, and the 20_ plunge of the T-axis was shown by Weaver and Baker (1988) to be in good agreement with the plate dip angle determined from the earthquake hypocenters. Therefore, Weaver and Baker (1988) concluded that the 1949 earthquake resulted from down-dip tensional forces within the subducting Juan de Fuca plate, an interpretation consistent with observations for many earthquakes in this depth range in other subduction zones (Isacks and Molnar, 1971). Rogers (1983a) reached a similar conclusion concerning the forces responsible for the 1965 south Seattle earthquake and a smaller event of magnitude 5.1 (m_b) in 1976 off the southeastern end of the Vancouver Island coast. Both events were at a depth of about 60 km and focal mechanisms calculated for both earthquakes were normal faulting with the T axes striking northeast and plunging down-dip (Rogers, 1983a).

Based on the agreement between the dip of the Juan de Fuca plate as inferred from earthquake hypocenters determined from the modern seismographic network and the dip of the T-axes calculated for the larger magnitude historical earthquakes, we believe that we can confidently predict the intraplate earthquake source region for the entire plate (fig. 7). We expect that future large magnitude (~7) intraplate events will occur within the Juan de Fuca plate system in the depth range of the 1949 and 1965 events. Although the depths of these events are considered to be well-known, we have chosen to bracket our source region at a shallower depth. An examination of the University of Washington seismic catalog for the years 1970 through 1990 shows that all of the intraplate earthquakes greater than magnitude 4 are below 45 km and that none have been located deeper than the 1976 event. Therefore, we have used the depth range of 45 to 60 km for our estimate of the probable source region for intraplate events (fig. 7).

We emphasize that this probable source region represents the likely areal extent within which an event may occur; the actual dimensions of the fault area associated with an earthquake of approximate magnitude 7 would be expected to be similar to the 40 km long-fault estimated for the 1949 earthquake (Baker and Langston, 1987). The queried area in southern Oregon represents the region of unknown plate geometry where no intraplate earthquakes have been located either because any events that did occur were not large enough to be detected by the existing seismic network or no events have occurred. We note that the expansion of the existing seismic network (fig. 2) will greatly help to resolve this long-standing question concerning whether this portion of the Juan de Fuca plate is currently truly aseismic. In northern California Benioff zone earthquakes again allow the plate depth to be estimated from the trench eastward to the western edge of the Cascade Range (see Cockerham, 1984; Walter, 1986), so we have shown the probable source region here between the same depth limits as in Washington and northern Oregon.

POSSIBLE SOURCE REGIONS FOR SUBDUCTION ZONE EVENTS

At nearly all convergent margins around the world, large magnitude (8+) earthquakes are known to occur; Cascadia is unusual in that there is no known large earthquake in the historical record. However, recent studies of subduction zone characteristics (Heaton and Kanamori, 1984; Heaton and Hartzell, 1986), crustal strain accumulation in Washington (Savage and others, 1981), crustal earthquakes in southwestern Washington (Weaver and Smith, 1983), and the stratigraphy of coastal marshes along the Washington and Oregon coasts (Atwater, 1987; Atwater, this volume; Grant and others, 1989) have all either concluded directly or inferred that the Cascadia subduction zone should be regarded as capable of generating great interface events.

The geological evidence from the coastal marshes is particularly compelling in terms of arguing unequivocally for the occurrence of past great subduction zone earthquakes. This evidence is multiple buried peat horizons in numerous bays along the coast (see Atwater, this volume). The preferred interpretation is that each buried peat layer represents a previous surface soil horizon that was suddenly submerged during a great earthquake. Subsequently intertidal deposits accumulated on the submerged soils, until the deposits reached a level allowing the development of a new marsh surface. At some sites in Willapa Bay submerged soils have fine-grained sands deposited directly on them. Atwater (1987; this volume) and Grant and others (1989) interpret these sands as being deposited by a tsunami, and because they lie directly on submerged peat layers, they are further interpreted as being locally generated by the same earthquake that produced the subsidence.

To date, the marsh data indicate that the return period of great earthquakes in the late Holocene has been irregular, varying from a return period of less than 100 years to more than 1200 years (Atwater, this volume; Grant and others, 1989). At numerous sites along the côast from Humboldt Bay in California to the Copalis River in Washington, there is considerable field evidence that the last event occurred about 300 years ago; both carbon-14 dating and tree-ring dating techniques are in reasonable agreement on this date (Atwater, this volume). There remains much uncertainty over both the repeat time for these great earthquakes and the probable magnitude. At present, both issues require showing synchronous behavior of specific marsh horizons at multiple sites on the coast. However, geologic techniques may not allow synchronous motion for a given marsh horizon ever to be proved, but these techniques may be able to rule out the possibility that the entire length of the subduction zone from Cape Mendocino to central Vancouver Island failed in a single earthquake.

The area where great subduction zone earthquakes occur is not in doubt, in that the shallow-dipping interface between the subducting plate and the overlying plate is the fault plane. The issues for the source area for subduction events relate to whether the entire length of the Cascadia subduction zone generates these earthquakes and the width of the fault that slips. A minimum for the length is provided by Heaton and Kanamori (1984). On the basis of an analysis of plate age and convergence rate used in a regression against the observed magnitude of interface events in other subduction zones, they suggested that in Cascadia an event of about magnitude 8.3 (M_W) would be expected given the plate age and convergence rate measured there. Such an earthquake might be expected to rupture a length on the order of 150-200 km along the subduction zone. After comparing a number of additional plate parameters such as offshore bathymetry and gravity and the historical rate of moderate (magnitude 5.7+) earthquakes, Heaton and Hartzell (1986) suggested that the entire length of the Cascadia zone (1100 km), from Cape Mendocino to central Vancouver Island might rupture in one great event of magnitude 9 or greater (M_W).

Although, as noted above, all of the marsh data are not completely analyzed, the great extent of the coast where evidence of rapid subsidence has now been documented (fig. 8) makes it likely that very large portions of the coast (at least several 100 to many 100's of km) were involved in some of the earthquakes. The impressive number of sites where multiple buried peat sequences have now been found (fig. 8) with similar dates suggests that the most likely cause of the marsh subsidence is coseismic surface deformation during a great earthquake. Because of the extent of the marsh data along the coast and the arguments of tectonic setting raised by Heaton and Hartzell (1986), we believe that the entire length of the Cascadia subduction zone is capable of generating great earthquakes, even though it is not yet clear that the entire length fails during a single event.

The second point in estimating source areas concerns the width of the rupture perpendicular to the coast. Two models, one in which the rupture is largely offshore and one in which the rupture is largely beneath the continent, bracket the range of probable source widths. In the first model, the rupture extends from the trench down-dip along the interface to a depth of 30-40 km. Because of the plate geometry, this width varies along the subduction zone with a maximum beneath northwestern Washington (≈200 km) and narrows to less than 100 km beneath central Oregon and areas further south (fig. 3). If the end of the fault rupture were as deep as 40 km, then beneath Washington a significant portion of the rupture area would be beneath land (fig. 9), whereas if the rupture ends near 30 km depth then the eastern extent of the fault area would be near the coast nearly everywhere. In the second model, in areas like Cascadia that have a very high rate of sedimentation offshore, Byrne and others (1988) have argued that as these sediments are subducted, they allow very poor coupling between the two plates from the trench landward possibly as far as the coast. With this model, the potential source area capable of generating subduction zone interface earthquakes in Cascadia is greatly reduced, consisting approximately of the area from about the coast inland to where the subducting plate begins to dip steeply eastward, perhaps at an approximate depth of 50-60 km (Byrne and others, 1988). Because of the plate geometry, this area is particularly small south of the arch beneath Puget Sound (fig. 3).

In our map of the source region for great subduction zone events, we have illustrated the case where the fault zone width extends from the deformation front offshore to just beneath the coastline (fig. 9). We have chosen this fault width because of deformation modeling results that indicate that it is difficult to explain the pattern of sudden, subsidence recorded in the coastal marsh stratigraphy without the rupture area extending offshore (Savage and others, 1981). The shaded pattern in figure 9 illustrates the great source area for these events. Based on the record from other subduction zones, interface events can be expected to occur within the entire area. We have illustrated the case where two earthquakes would fill the entire zone; other possibilities have been discussed by Heaton and Hartzell (1986). To date the geological record from coastal intertidal marshes does not rule out the possibility that the entire length of the subduction zone fails in one very great earthquake. However, the same data do not yet provide unequivocal interpretations of whether the zone might break in a series of smaller earthquakes, of approximate magnitude 8 to 8.5 (M_W).

The very great area involved in any subduction zone event does indicate the need for large-scale studies of the properties of the Cascadia subduction zone. For example, greatly expanded regional strain studies would help address the issue of whether the plate boundary breaks in a great single earthquake or in a series of smaller events. A second issue is that of the eastern extent of the fault. Clearly, if the limit of the fault zone for a great subduction event were further westward than the limit shown in figure 9, then it is possible that this would favorably influence shaking effects in the major urban areas.

KNOWN SOURCE REGIONS OF LARGE CRUSTAL EARTHQUAKES

The known areas of large magnitude (7+) crustal earthquakes in the North American plate in the Pacific Northwest are limited to central Vancouver Island and the North Cascades (fig. 1). The Vancouver Island events were probably related to the stress regime generated by the interaction of the Explorer plate (at the northern end of the Juan de Fuca plate) with the North American plate (Rogers, 1983b). The cause of the 1872 event remains uncertain in that it occurred in an area with very little contemporary seismicity and no obvious surface fault expression (Malone and Bor, 1979).

The existence of these large crustal events raises the question of whether they might occur within the urban areas of western Washington and Oregon. Unfortunately, the sparsity of known Quaternary faulting (Gower and others, 1985) and the current seismicity distribution does little to answer this question. In the Puget Sound basin the crustal earthquakes do not fall along simple, linear fault zones, but appear to be distributed throughout the crust (fig. 4). Zollweg and Johnson (1989) have recently interpreted a sequence of earthquakes on the western margin of the North Cascades as evidence of a southerly dipping fault zone, the first such zone identified in northwestern Washington. Nevertheless, it remains impossible to infer either the possibility of or argue conclusively against a future magnitude 7+ shallow crustal earthquake in Puget Sound.

Mapped Quaternary faulting in the Puget Sound basin is sparse (Gower and others, 1985) although newly collected shallow reflection data may indicate Quaternary faulting within Puget Sound (Harding and others, 1988). Late Holocene marine terraces provide evidence of abrupt uplift at two locations within the basin, Restoration Point (5 km west of Seattle) and Belfair (45 km southwest of Seattle). Such uplifts may be due to an earthquake or earthquakes (Bucknam and Barnhard, 1989). The available seismicity record in the Puget Sound basin is usually interpreted to suggest that the expected maximum magnitude crustal earthquake is less than that expected in either southwestern Washington or southeastern Washington (Ludwin and others, 1991). The observations of uplift on recent Holocene terraces by Bucknam and Barnhard (1989) raises doubts about these earlier interpretations and indicates the need for continued paleoseismic studies in the urban areas.

In contrast to the earthquake distribution in the Puget Sound basin, in southwestern Washington, much of the earthquake activity occurs along the St. Helens zone (SHZ), a right-lateral strike-slip zone that is defined for over 130 km (Ludwin and others, 1991; Weaver and Smith, 1983). Two earthquakes greater than magnitude 5 have occurred on the SHZ since 1960; in 1961 a magnitude 5.1 (M_L) event occurred south of Mount St. Helens and in

1981 a magnitude 5.5 (M_L) event occurred on the SHZ to the north of the volcano. This last event is the largest magnitude crustal earthquake recorded in the Pacific Northwest since the modern seismic network was installed. Mount St. Helens directly overlies the zone where a small (few kilometers) right-stepping offset occurs (Weaver and others, 1987). Several studies have assumed that the magmatic system beneath Mount St. Helens prevents the entire 130 km length from rupturing in a single earthquake (Weaver and Smith, 1983; Grant and Weaver, in press). Grant and Weaver (in press) compared possible source areas along the SHZ north of Mount St. Helens with observations of both fault area and magnitudes calculated from earthquakes on other strike-slip fault zones. As a result of this comparison, Grant and Weaver (in press) concluded that an earthquake in the magnitude range of 6.2-6.8 was the expected maximum magnitude event for the SHZ north of Mount St. Helens.

Our plot of crustal earthquake source areas (fig. 10) shows the regions where these events have occurred in the historical record plus the SHZ and the northern end of the San Andreas system in California. The large area shaded in the North Cascades illustrates the uncertainty in the epicentral position of the 1872 earthquake. Although Malone and Bor (1979) concluded that the event probably was on the northeastern flank of the Cascade Range, other studies (Milne, 1956), have suggested that the event was located near the United States-Canadian border. One problem highlighted by this map is that modern seismicity has no indication of the past large crustal events in central Vancouver Island and the North Cascades.

The map does emphasize the advantage of both accurate locations and an understanding of the seismotectonics responsible for crustal earthquakes, in that along the SHZ it is possible to place a large event on a specific structure, as opposed to having to consider it equally likely that the event may occur throughout a given area. We emphasize that fig. 10 represents a very incomplete assessment of the source regions of large crustal events. Considerable regional geology, local Quaternary studies, and regional-scale strain networks, as discussed by Shedlock and Weaver (1989), will be required to reduce the uncertainty in defining source regions for large crustal earthquakes in the Pacific Northwest.

DISCUSSION

Our definition of the source areas for the three types of earthquakes expected in the Pacific Northwest has relied heavily on the historical and instrumental seismicity with reference to the regional tectonic setting. This is most clearly the case with respect to the area of possible intraplate earthquakes. The existing understanding of the geometry of the Juan de Fuca plate system is sufficient to allow us to argue for events like the 1949 south Puget Sound earthquake occurring anywhere along the length of the subducting plate system. Current interpretations of marine seismic reflection data (Hyndman and others, 1990; Couch and Riddihough, 1989; Snavely, 1988) show that everywhere offshore the Juan de Fuca plate is dipping eastward at shallow angles (<10°). Onshore, even in areas where the Juan de Fuca plate is currently seismically quiet, inversion of teleseismic arrivals has been interpreted to show that below depths of 40-60 km the subducting plate is nearly vertical beneath the Cascade Range (Michaelson and Weaver, 1986; Rasmussen and Humphreys, 1988). Furthermore, Crosson and Owens (1987) have suggested that teleseismic waveforms recorded near the central Washington coast are compatible with the hypothesis that the Juan de Fuca plate is continuous (that is not faulted or offset) south of the arch beneath northwestern Oregon. Thus, despite variations in the dip of the upper portion of the plate, the known or interpreted characteristics of the Juan de Fuca plate argues strongly for the existence of a subducting plate capable of producing intraplate tensional stresses along the down-dip axis of the plate (see Spence, 1987 for a discussion of plate stresses). Indeed, if the 1873 earthquake at the Oregon-California border (Figure 1) was an intraplate earthquake as suggested by Ludwin and others (1991), then the currently seismically quiet portion of the subducting plate system experienced the second largest Benioff zone earthquake in historic times.

Whether there is any link between the segmented nature of the crustal earthquake distribution and sources for either subduction zone events or crustal earthquakes is unknown. With respect to crustal earthquakes, the segmentation outlined by Weaver and others (1990) is similar to the division of the Cascade Range suggested by Guffanti and Weaver (1988). These authors used differences in the distribution and composition of late Cenozoic volcanic vents (<5 Ma) to divide the Cascade Range into five segments. These volcanic segments were the same as those suggested here except Guffanti and Weaver (1988) had two segments in northern California. In fact, isostatic residual gravity anomalies (Blakely and Jachens, 1990) and the variations in the volume of Quaternary volcanism (Sherrod and Smith, 1990) in the Cascade Range can be used to divide the Cascade Range into segments identical to those of Guffanti and Weaver (1988). Apparently, the segmentation suggested by the seismicity pattern in the Pacific Northwest must be reflecting the same regional tectonic framework that fundamentally shaped the distribution and volume of late Cenozoic volcanism. The components of this tectonic framework (variations in crustal stress, changes in crustal structure) have left their mark in the middle and upper crust, expressed today by the gravity field and to a lesser extent, regional heat flow (Blackwell and others, 1990), but not in the overall crustal thickness (fig. 5).

The link between the regional seismotectonic fabric and late Cenozoic volcanism is likely to be particularly important in assessing the earthquake hazards in the Portland, Oregon, area. A recent study of seismicity in the region around Portland has concluded that there may be a series of en-echelon, strike-slip fault zones southwest of the SHZ (Yelin and Patton, 1991). The Portland Hills fault, a few kilometers west of downtown Portland, shows at least Pliocene offsets (Sherrod and Pickthorn, 1989), and is nearly parallel to the proposed en-echelon seismic zones. The Portland area has a history of moderate magnitude earthquakes (≈5+), but the most recent, in 1962, was probably a normal faulting event (Yelin and Patton, 1991). This sense of motion is compatible with the hypothesis that the Portland area is a basin formed by crustal extension, and localized crustal extension may explain the presence of upper Cenozoic basaltic volcanism in the Portland urban area, far west of the Cascade axis. Although currently the relation between contemporary seismicity and the Portland Hills fault zone or the basaltic volcanism is not understood, it is clear that the area warrants continued seismic monitoring to determine more completely the seismotectonic relations and the associated earthquake hazards. Unfortunately, as noted by Ludwin and others (1991), since 1970 crustal seismicity in the Portland area has been sparse compared to higher rates of activity in the 1960's.

The segmentation interpreted from the seismicity pattern may provide additional justification for assumptions that are made for modeling ground shaking from crustal earthquakes. In areas where faults are not well exposed at the surface it is necessary to divide a region into subregions with different source characteristics (e.g., Algermissen and others, 1982). Often these divisions reflect geological boundaries and the maximum magnitude earthquake expected within each subregion. If the seismic segmentation is related to broad regional processes or structures, then this will be a further constraint that can be used in determining subregions for ground motion modeling. In addition, the segmentation may determine boundaries for the use of different modeling techniques. For example, across southwestern Washington and northwestern Oregon, the definition of discrete seismic zones (as earthquakes are located) will likely provide a framework within which maximum magnitude events can be estimated from an interpretation of seismicity patterns, geologic mapping, and crustal structure studies. In northwestern Washington, however, the lack of recognized seismic zones, makes it likely that here it will continue to be necessary to rely largely on an areal approach to hazard estimation. Clearly, a more complete characterization of the crustal structure and regional tectonics is key to understanding the generation of the two widely disparate seismicity patterns found in northwestern Washington and southwestern Washington. A second point concerns the possible segmentation of the Cascadia thrust interface. If any of the processes responsible for the segmentation observed across the Cascade Range owe their genesis ultimately to the direct interaction of the Juan de Fuca plate system with the North American plate, then perhaps some fundamental rupture length along the coast might reflect this segmentation.

We emphasize that the source regions we have shown for the three earthquake types are part of a major revision of earthquake hazards facing the Pacific Northwest. As noted by Heaton and Hartzell (1987), great subduction zone earthquakes will have shaking effects that are felt over much of western Washington, western Oregon, northwestern California and southwestern British Columbia. The possibility of these earthquakes alone is enough to warrant a thorough re-examination of building codes and earthquake preparedness programs throughout the region. We believe it is critical that this re-examination begin immediately in cities in the Willamette Valley such as Eugene, Salem and Portland because these areas have been outside of earlier efforts undertaken in the Puget Sound region (Hopper and others, 1975) that at least have allowed local governments to be aware of some of the potential problems from earthquakes. The fact that Eugene and Portland have not experienced the larger magnitude earthquakes found in other parts of the Pacific Northwest (Puget Sound, North Cascades, northern California) should not be used as justification for delaying reassessments of hazards. We reiterate the fact that the short-term seismic record in the Pacific Northwest can not be viewed as representative of the long-term risk of large earthquakes. In addition, the fact that we judge the entire portion of the subducting Juan de Fuca plate capable of producing intraplate earthquakes like those of 1949 and 1965 in the Puget Sound basin, coupled with the uncertainty surrounding the extent of the areas where large crustal earthquakes should be expected, provides added urgency for a region-wide re-examination of the hazards posed by earthquakes.

Finally, we note that there may be other potential seismic source zones in the Pacific Northwest that may ultimately be important in earthquake hazard assessments. One zone that might be relevant to earthquake hazards is west of the deformation front, in the nearly flat-lying portion of the Juan de Fuca and Gorda plates. Earthquakes in this portion of plates were referred to as oceanic intraplate by Astiz and others (1988). Off the coast of northern California, this zone is particularly active. But as noted above, oceanic intraplate earthquakes in the historical record have caused relatively little damage to onshore facilities. Nevertheless, as the collapse of the highway bridge during the 1980 Eureka earthquake indicated, these events must be acknowledged in hazards assessments along the northern coast of California. Off the Oregon and Washington coasts only an occasional event is located within the Juan de Fuca plate west of the deformation front; Spence (1989) discusses a few of the known major events. Because there are so few events known off the Oregon and Washington coasts and the tectonic setting of these intraplate events is very uncertain, it is not yet possible to assess how important events west of the deformation front may be in the overall earthquake hazards assessment of the Pacific Northwest. However, if the historical record of damage from this zone from northern California is appropriate for Oregon and Washington, combined with the fact that the distance between the deformation front and the coastline steadily increases northward from Cape Mendocino, then it is likely that oceanic intraplate events will be of considerably reduced importance in most of the Northwest compared with the three zones discussed here.

SUMMARY

In the convergent margin setting of the Cascadia subduction zone, three distinct earthquake sources are possible: 1) earthquakes at the interface between the Juan de Fuca and North American plate; 2) earthquakes within the crust of the overlying North American plate; and 3) earthquakes within the subducting Juan de Fuca plate. For each source type we have defined the approximate region where we expect an earthquake of that type can be expected to occur. The probable source region for intraplate earthquakes within the Juan de Fuca plate is the best known, as we are able to combine the historical data from the 1949 and 1965 earthquakes with the modern instrumental record. The latter data have been used to infer the geometry of the Juan de Fuca plate whereas the former have been used to deduce that the large magnitude earthquakes occur at least in part in response to down-dip tensional forces within the subducting plate. We suggest that the entire subduction zone, at depths between 45 and 60 km, is capable of producing these events.

Despite many unresolved issues surrounding great subduction zone interface earthquakes, the available geological record, combined with the plate tectonic setting, is interpreted as evidence that these events have occurred in the late Holocene along the coast. Because these events occur on the shallow dipping portion of the plate interface, the general location of the events is well known. Subduction zone events represent a major threat to the population of the Pacific Northwest that has not been integrated into current hazard assessments. A program to accomplish this integration will necessarily have to consider the large scale of the source region for these earthquakes. Finally, the possibility of large crustal earthquakes in the urban areas remains very poorly studied in the Pacific Northwest. Major new initiatives will be required to determine whether the urban centers in western Washington and Oregon must contend with the problems posed by this source type.

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FIGURE CAPTIONS

Figure 1. Location of earthquakes in the Pacific Northwest with magnitudes estimated to be greater than 6 for the years 1870 to 1990; except for the Cape Mendocino area offshore earthquakes clearly related to the ridge system have not been shown. Earthquake magnitudes are scaled by two symbol sizes, with the small symbols representing magnitudes of 6.0 to 6.9, and the large symbols representing magnitudes greater than 7.0. All earthquakes greater than magnitude 7 and three earthquakes greater than magnitude 6.0 discussed in text have year of occurrence noted. Triangles are Quaternary stratovolcanoes in the Cascade Range.

Figure 2. Regional short-period seismic stations operating in the Pacific Northwest and southwestern British Columbia. Triangles are seismic stations; open circles in southern Oregon represent stations installed during the late summer and early fall of 1990 by the University of Washington.

Figure 3. Distribution of earthquakes greater than 30 km depth, 1980 to April 1990. All earthquake hypocenters have at least 6 P waves used in the solutions, with epicentral errors less than ±3km, depth errors less than ±5 km, and RMS values of less than 0.35 seconds. The 60 km line shows the approximate easternmost extent of the earthquake distribution at these depths, and is from Weaver and Baker (1988).

Figure 4. Distribution of crustal earthquake activity, 1980 to April 1990. Earthquakes are less than 30 km in depth with location statistics the same for the hypocenters shown in Figure 3. SHZ is the St. Helens zone. Volcanoes are indicated by triangles abbreviated as follows: B is Mount Baker, R is Mount Rainier, S is Mount St. Helens, H is Mount Hood, N is Newberry Volcano, SH is Mount Shasta, L is Lassen Peak. [Figure from Weaver and others, 1990].

Figure 5. Contour map of crustal thickness in the Pacific Northwest. The solid lines are reversed refraction profiles (sometimes with other supporting geophysical data); dashed lines are gravity or electromagnetic profiles; dotted lines are crustal reflection profiles; areas enclosed by dashed lines have had estimates of crustal thickness made from seismic array studies. Sources are given by Mooney and Weaver, 1989. [Figure modified from Mooney and Weaver, 1989].

Figure 6. Major earthquakes in Oregon and Washington. Small symbols are events with magnitudes estimated from felt areas to be between 6.0 to 6.9, large symbols are events with magnitudes estimated to be greater than magnitude 7. The 1872 and 1936 events are thought to be crustal, whereas all other events are either known or thought to be within the subducting plate system. [Figure from Ludwin and others, 1991].

Figure 7. Schematic map of the probable source region for intraplate, down-dip tensional earthquakes. Large magnitude earthquakes (~7-7.5) are expected anywhere within the shaded region. Question marks indicate areas where there are no earthquakes located within the Juan de Fuca plate and the plate geometry is uncertain. Base map is modified from Ludwin and others (1991).

Figure 8. Location of sites in Washington, Oregon, and northern California where coastal marsh stratigraphy studies have found evidence of sudden subsidence. Compiled from Grant and others, 1989.

Figure 9. Example of source areas for two interplate earthquakes on the shallow, dipping interface. Approximate moment magnitude for each event is given. Other combinations are possible--see text for discussion. Map base is modified from Ludwin and others (1991).

Figure 10. Known source areas for historical crustal earthquakes greater than magnitude ≈6.5; dates give the year of events greater than magnitude 7. The hatched area north of Mount St. Helens represents the segment of the SHZ where Grant and Weaver (in press) suggest a maximum magnitude earthquake in the range of 6.2 to 6.8. Map base is modified from Ludwin and others (1991).

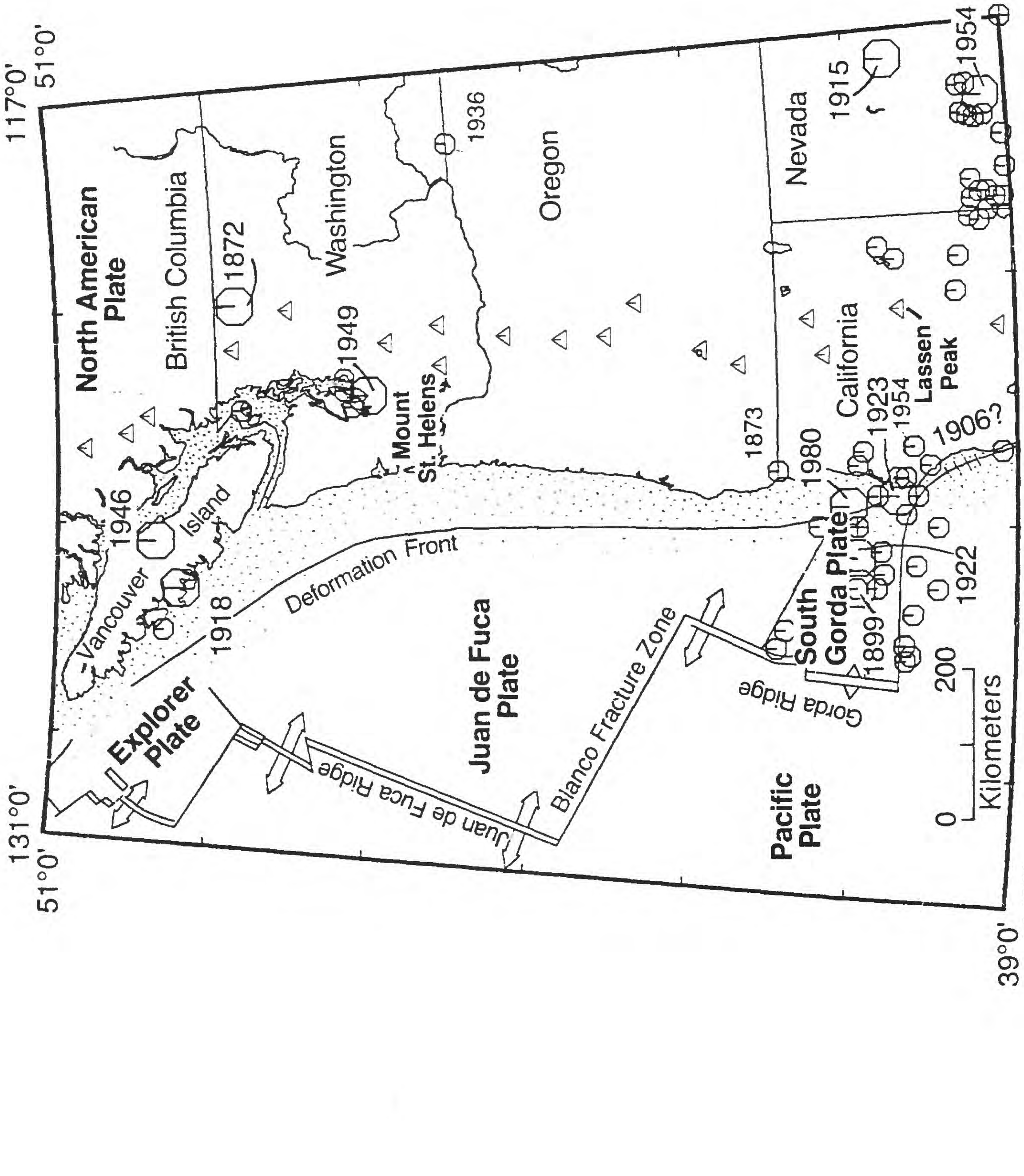

North American Plate
British Columbia
Washington
Oregon
Nevada
California
Vancouver Island
Explorer Plate
Juan de Fuca Plate
Pacific Plate
South Gorda Plate
Gorda Plate
Juan de Fuca Ridge
Blanco Fracture Zone
Gorda Ridge
Deformation Front
Mount St. Helens
Lassen Peak
1946
1918
1872
1949
1936
1873
1980
1923
1954
1906?
1915
1899
1922
0
200
Kilometers
131°0'
51°0'
117°0'
39°0'

Fig. 1

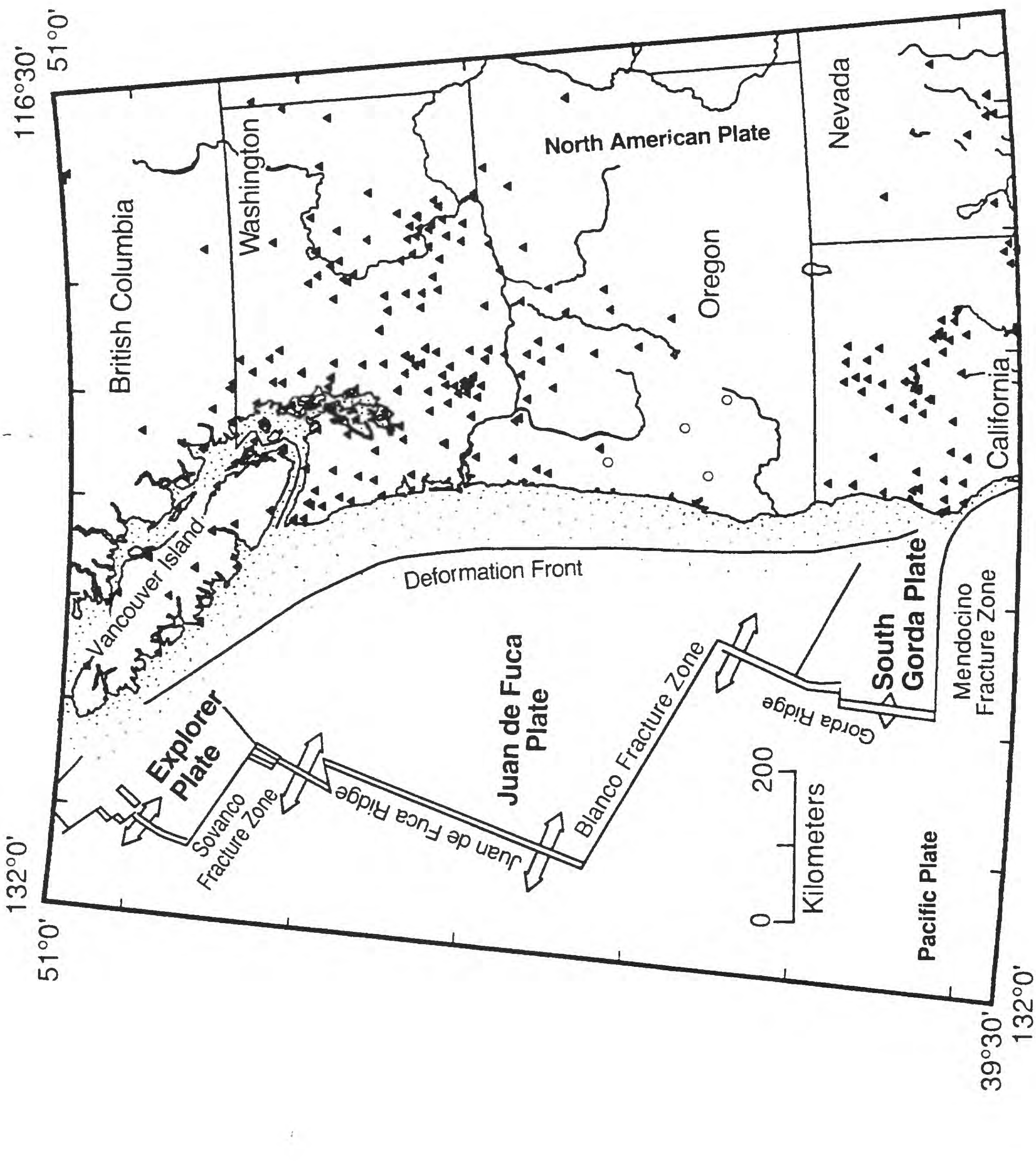

116°30'
51°0'
132°0'
51°0'
39°30'
132°0'
British Columbia
Washington
North American Plate
Nevada
Oregon
California
Vancouver Island
Deformation Front
Explorer Plate
Juan de Fuca Plate
Sovanco Fracture Zone
Juan de Fuca Ridge
Blanco Fracture Zone
South Gorda Plate
Gorda Ridge
Mendocino Fracture Zone
0
200
Kilometers
Pacific Plate

Fig. 2

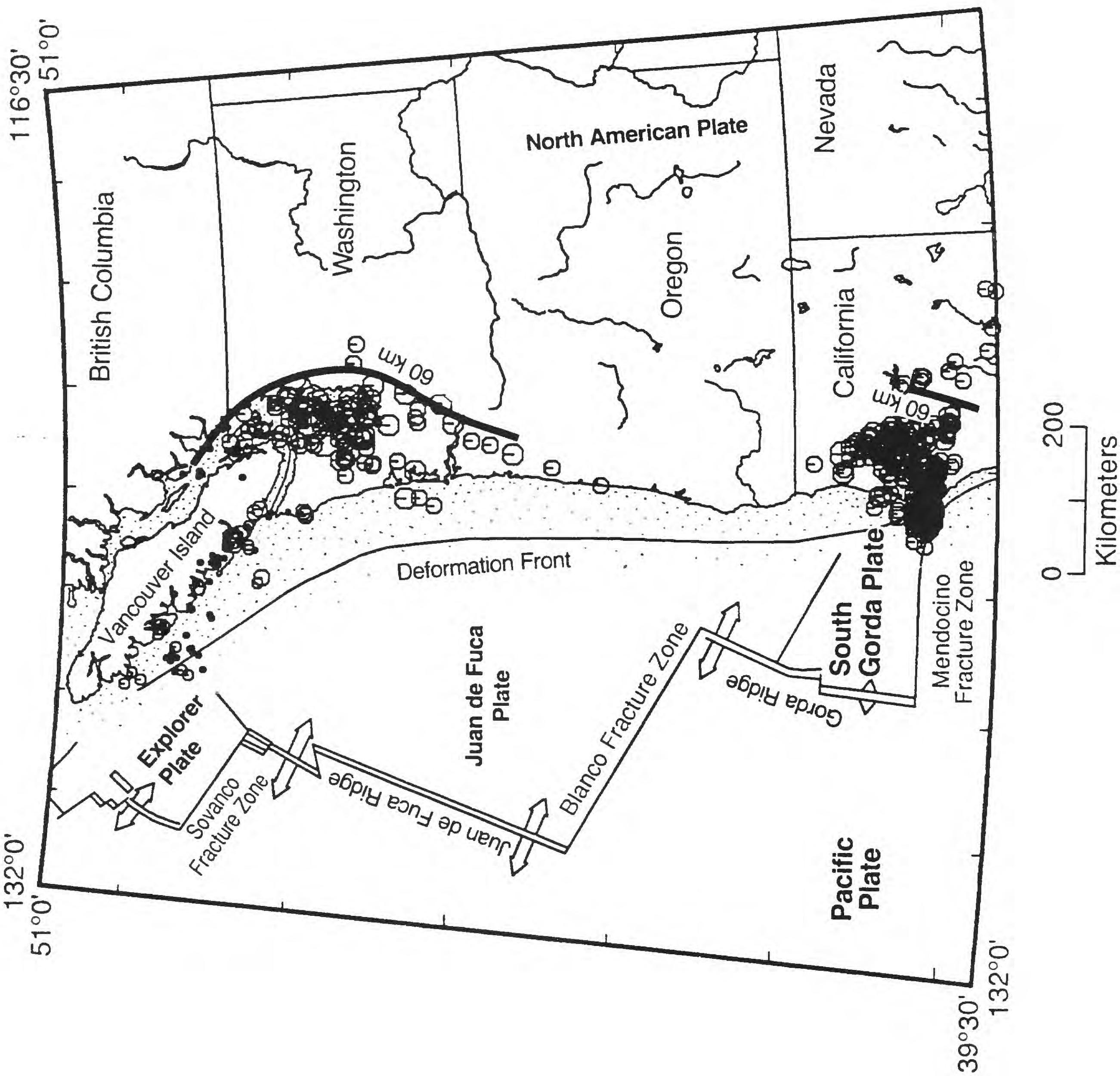
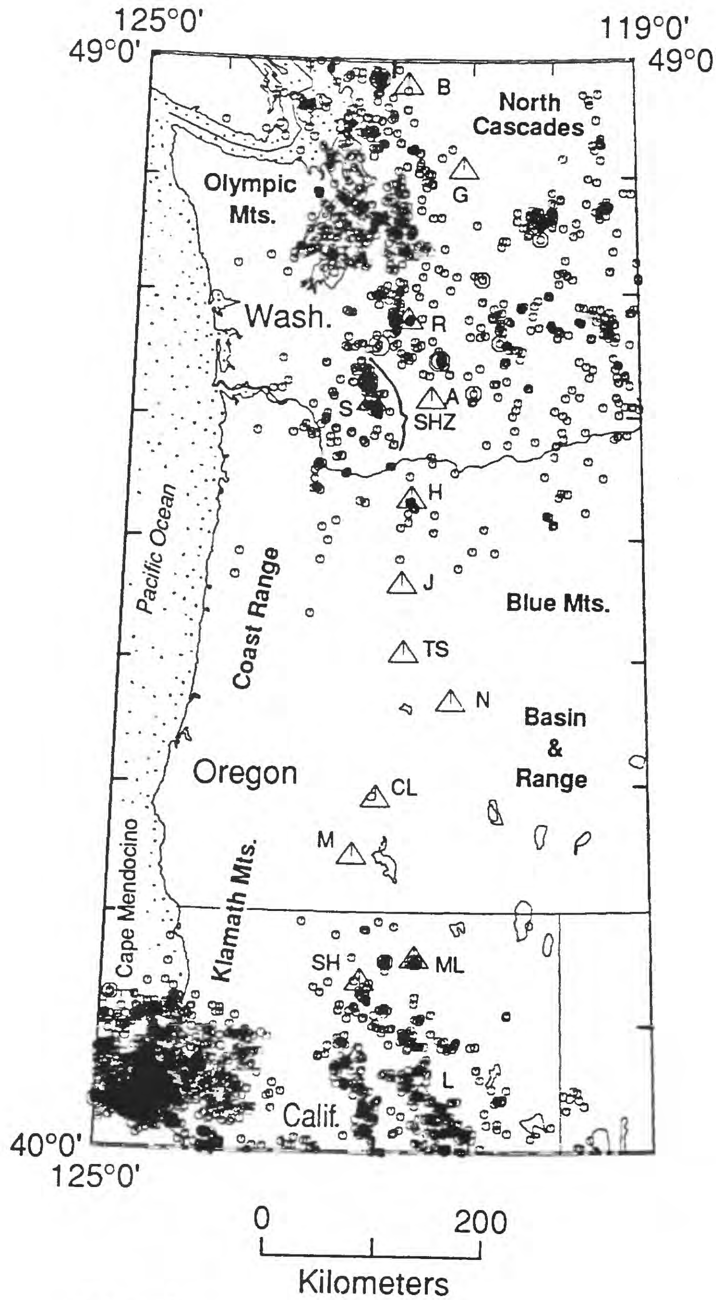
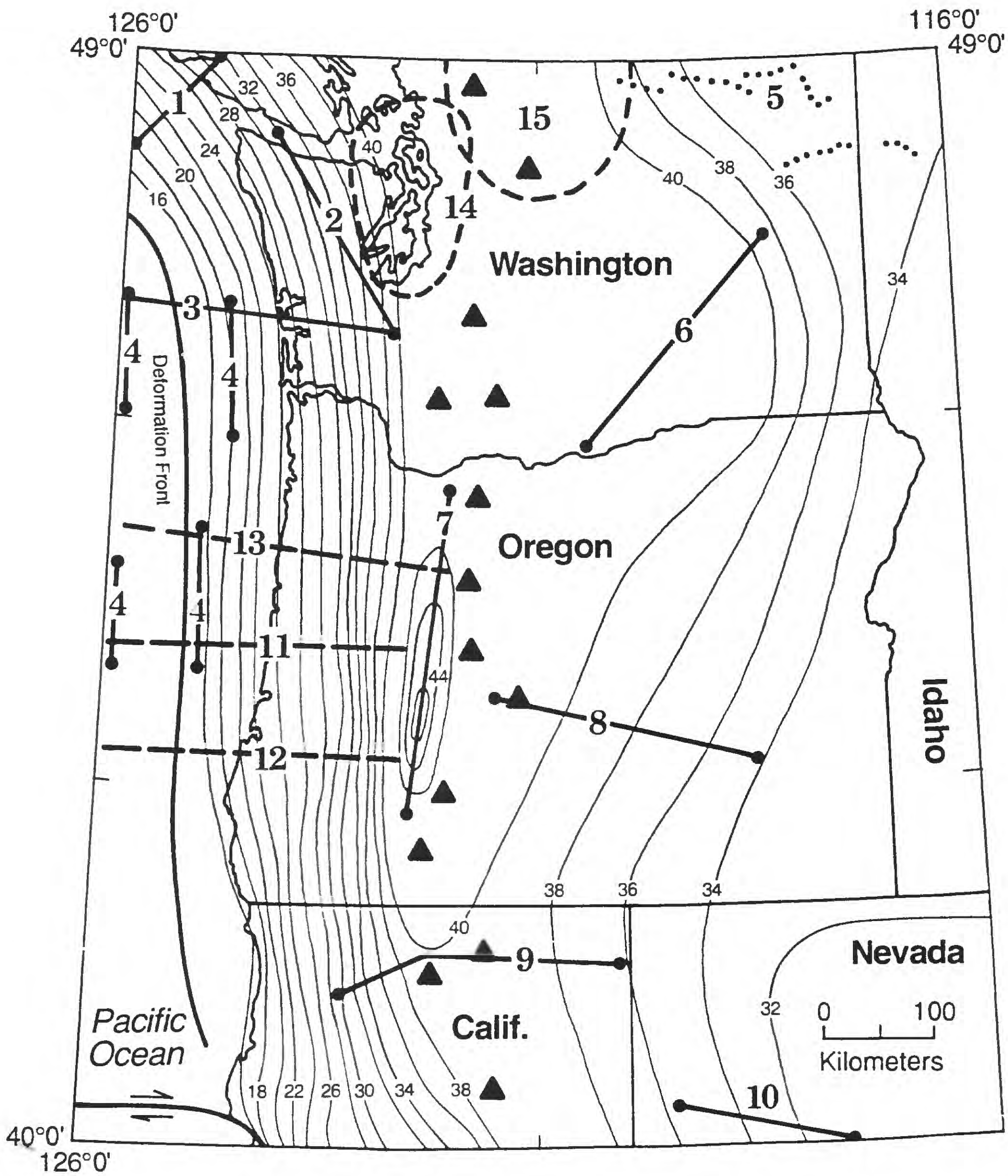

116°30'
51°0'
132°0'
51°0'
39°30'
132°0'
British Columbia
Washington
North American Plate
Oregon
Nevada
California
60 km
60 km
Vancouver Island
Deformation Front
Juan de Fuca Plate
Explorer Plate
Sovanco Fracture Zone
Juan de Fuca Ridge
Blanco Fracture Zone
Gorda Ridge
South Gorda Plate
Mendocino Fracture Zone
Pacific Plate
0
200
Kilometers

Fig. 3



125°0'
119°0'
49°0'
49°0'
B
North Cascades
Olympic Mts.
G
Wash.
R
S
A
SHZ
H
Pacific Ocean
Coast Range
J
Blue Mts.
TS
N
Basin & Range
Oregon
CL
M
Cape Mendocino
Klamath Mts.
SH
ML
L
Calif.
40°0'
125°0'
0
200
Kilometers

Fig. 4



126°0'
49°0'
116°0'
49°0'
Washington
Oregon
Idaho
Nevada
Calif.
Pacific
Ocean
Deformation Front
0 100
Kilometers
40°0'
126°0'

Fig. 5

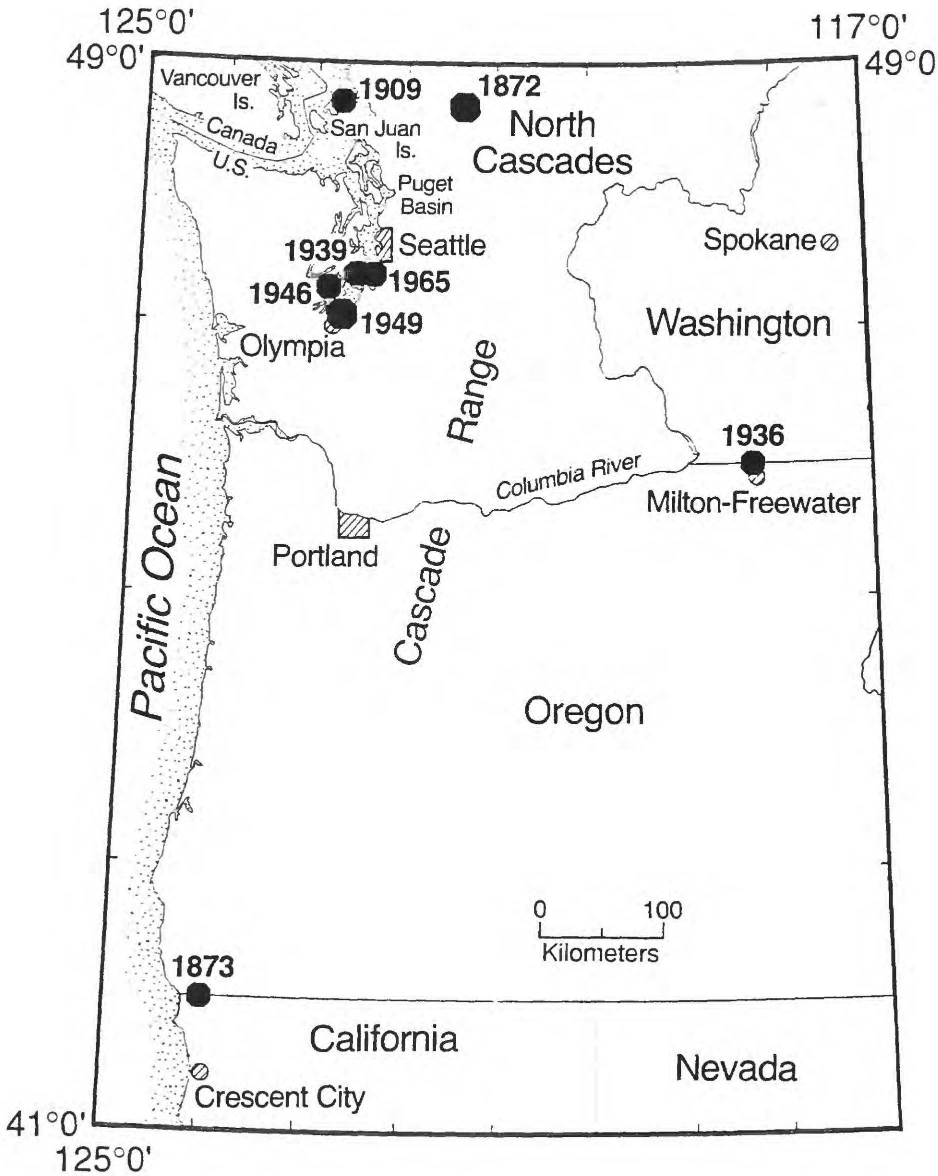

125°0'
117°0'
49°0'
49°0'
Vancouver Is.
Canada
U.S.
1909
1872
San Juan Is.
North Cascades
Puget Basin
Seattle
Spokane
1939
1965
1946
1949
Olympia
Washington
Range
1936
Columbia River
Milton-Freewater
Pacific Ocean
Portland
Cascade
Oregon
0
100
Kilometers
1873
California
Nevada
Crescent City
41°0'
125°0'

Fig. 6

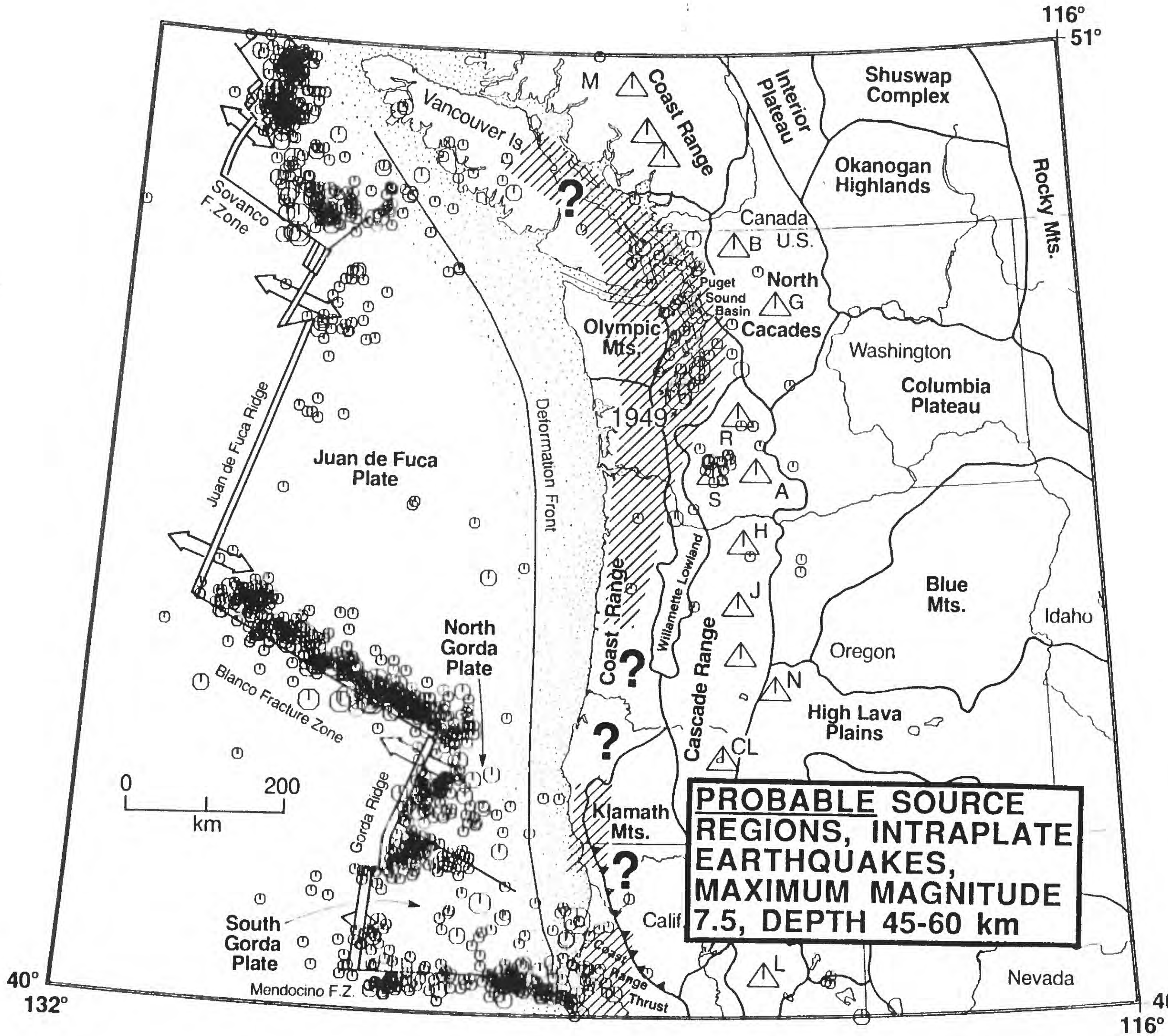
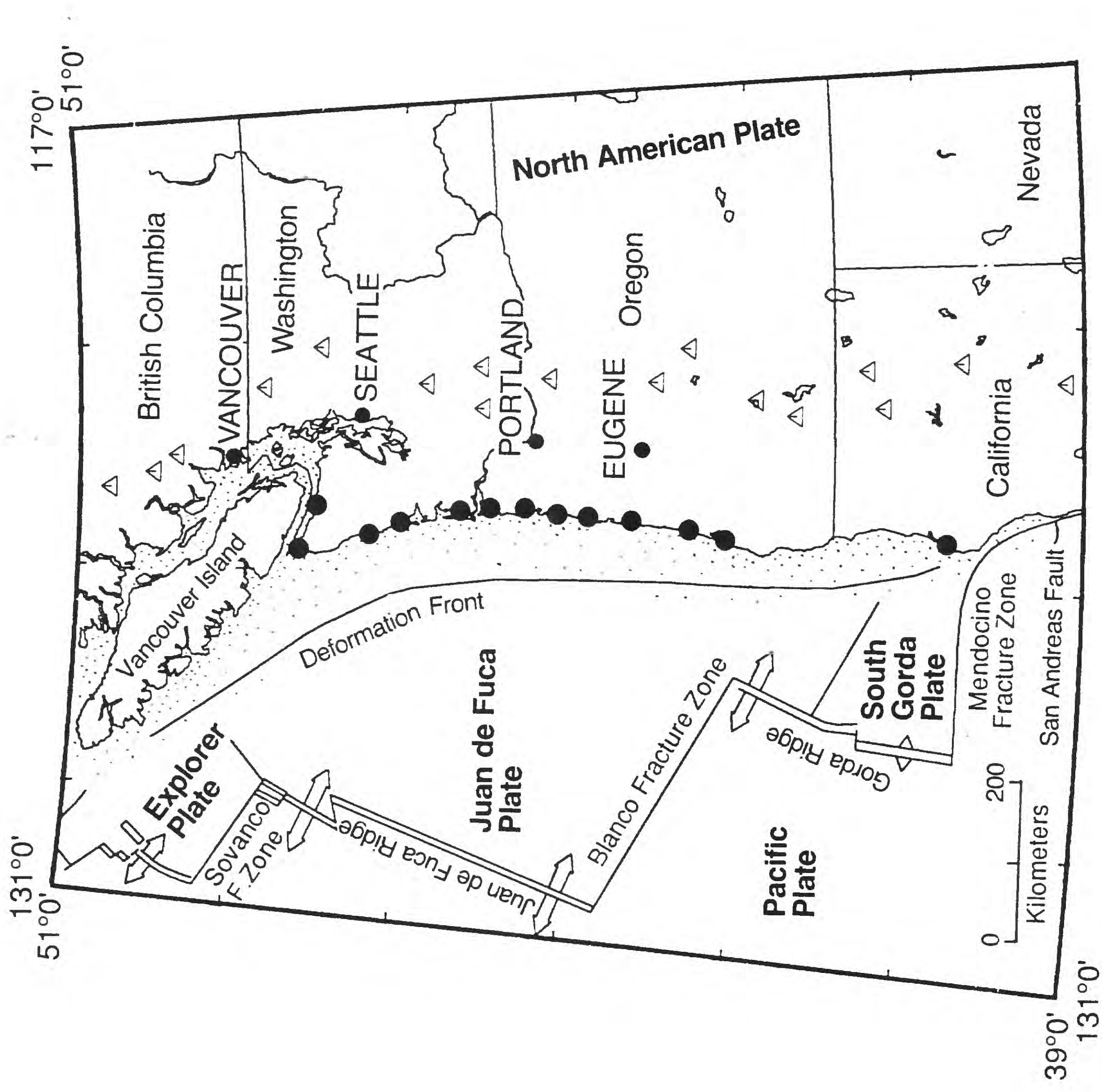
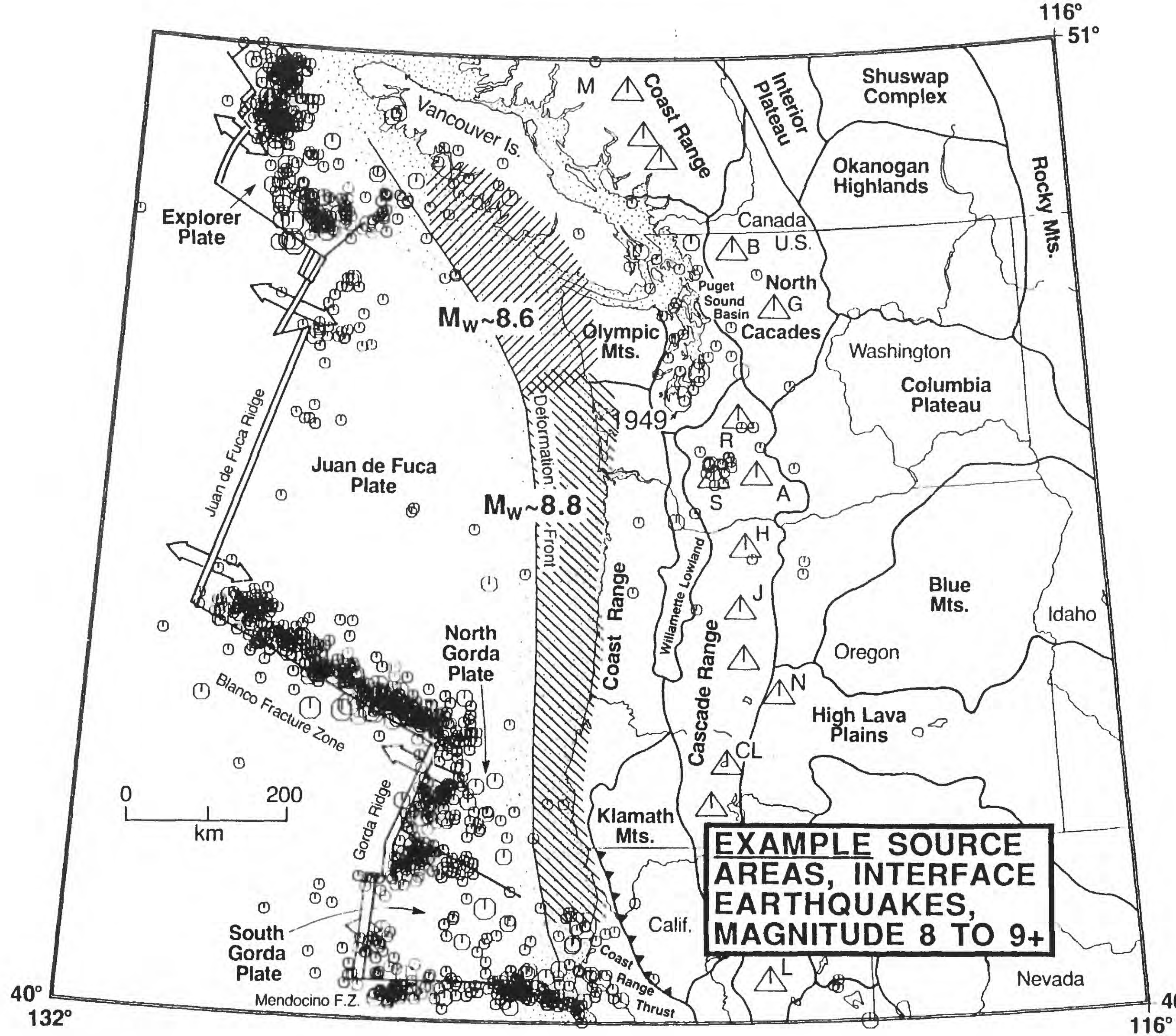

116°
51°
40°
132°
40°
116°
Shuswap
Complex
Interior
Plateau
Okanogan
Highlands
Rocky Mts.
Coast Range
M
Vancouver Is.
Canada
U.S.
B
North
G
Cacades
Puget
Sound
Basin
Olympic
Mts.
Washington
Columbia
Plateau
1949
R
S
A
H
J
N
CL
L
Blue
Mts.
Idaho
Oregon
High Lava
Plains
Willamette Lowland
Cascade Range
Coast Range
Klamath
Mts.
Calif.
Nevada
Coast Range Thrust
Deformation Front
Juan de Fuca
Plate
Juan de Fuca Ridge
Sovanco
F. Zone
North
Gorda
Plate
Blanco Fracture Zone
Gorda Ridge
South
Gorda
Plate
Mendocino F.Z.
0
200
km
PROBABLE SOURCE
REGIONS, INTRAPLATE
EARTHQUAKES,
MAXIMUM MAGNITUDE
7.5, DEPTH 45-60 km

Fig. 7



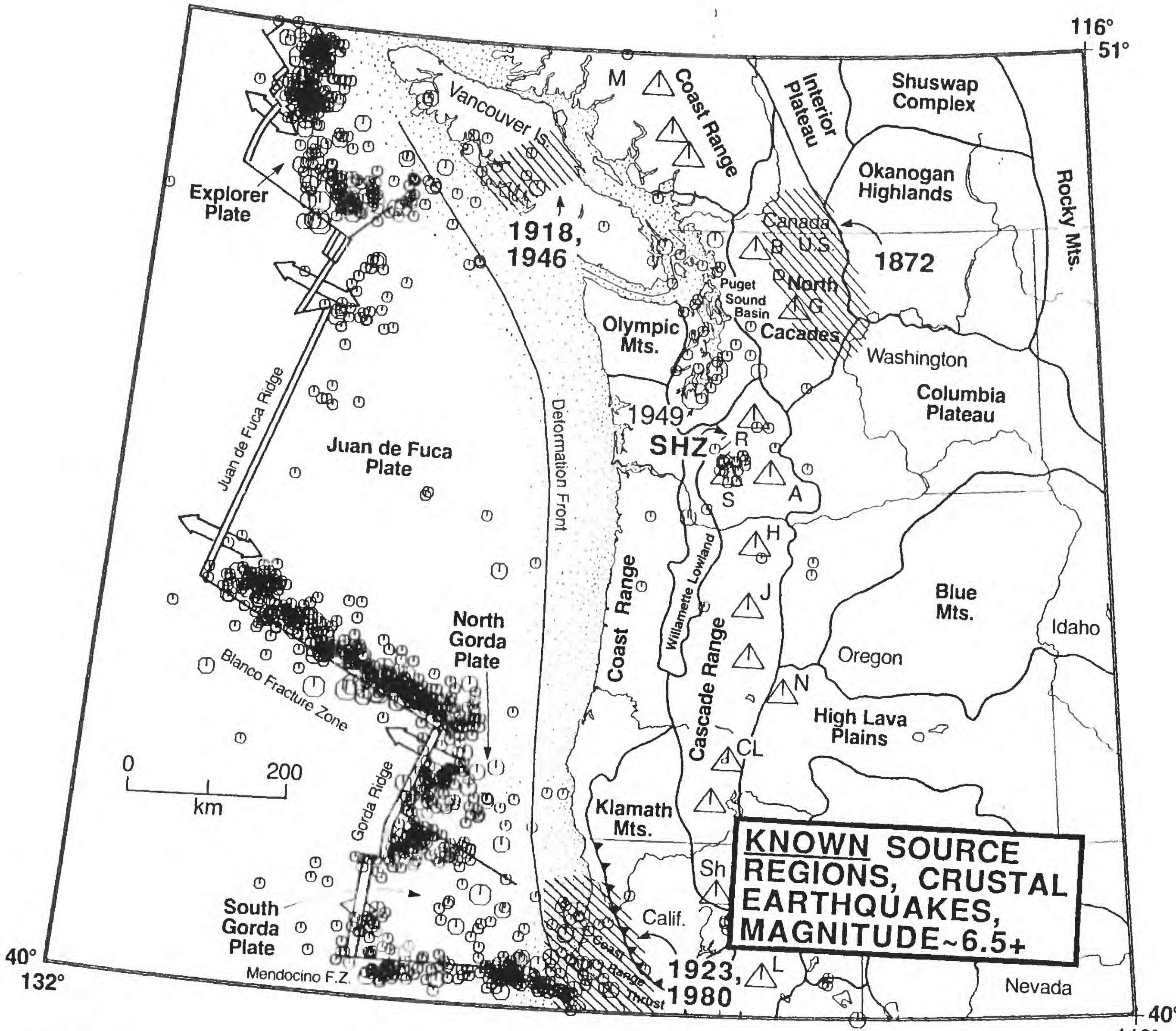
North American Plate
British Columbia
VANCOUVER
Washington
SEATTLE
PORTLAND
EUGENE
Oregon
California
Nevada
Vancouver Island
Deformation Front
Juan de Fuca Plate
Explorer Plate
Sovanco F. Zone
Juan de Fuca Ridge
Blanco Fracture Zone
South Gorda Plate
Gorda Ridge
Mendocino Fracture Zone
San Andreas Fault
Pacific Plate
0
200
Kilometers
131°0'
51°0'
117°0'
51°0'
39°0'
131°0'

Fig. 8



116°
51°
40°
132°
40°
116°
Shuswap Complex
Okanogan Highlands
Rocky Mts.
Interior Plateau
Coast Range
M
Vancouver Is.
Canada
U.S.
B
North
Cacades
G
Puget Sound Basin
Explorer Plate
Olympic Mts.
M_W~8.6
Washington
Columbia Plateau
1949
R
S
A
Juan de Fuca Ridge
Juan de Fuca Plate
Deformation Front
M_W~8.8
H
J
Coast Range
Willamette Lowland
Cascade Range
Blue Mts.
Idaho
Oregon
N
High Lava Plains
CL
North Gorda Plate
Blanco Fracture Zone
0
200
km
Gorda Ridge
Klamath Mts.
Calif.
EXAMPLE SOURCE AREAS, INTERFACE EARTHQUAKES, MAGNITUDE 8 TO 9+
South Gorda Plate
Mendocino F.Z.
Coast Range Thrust
L
Nevada

Fig. 9



KNOWN SOURCE REGIONS, CRUSTAL EARTHQUAKES, MAGNITUDE~6.5+
116°
51°
40°
132°
Shuswap Complex
Interior Plateau
Okanogan Highlands
Rocky Mts.
Coast Range
M
Vancouver Is.
1918, 1946
1872
Canada
U.S.
B
North Cacades
G
Puget Sound Basin
Olympic Mts.
Washington
Columbia Plateau
1949
SHZ
R
S
A
H
J
N
CL
Sh
L
Explorer Plate
Juan de Fuca Ridge
Juan de Fuca Plate
Deformation Front
Coast Range
Willamette Lowland
Cascade Range
Blue Mts.
Idaho
Oregon
High Lava Plains
North Gorda Plate
Blanco Fracture Zone
Gorda Ridge
0
200
km
Klamath Mts.
Calif.
Coast Range Thrust
1923, 1980
South Gorda Plate
Mendocino F.Z.
Nevada

Fig. 10