INTRODUCTION TO SEISMOLOGICAL CONCEPTS RELATED TO EARTHQUAKE HAZARDS IN THE PACIFIC NORTHWEST

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The objective of this brief discussion is to acquaint you with the general aspects of the earthquake hazards in the Pacific Northwest. We will address the "why," "how big," and "how often" of earthquake occurrence. In addition, some mention will be made of the severity of effects that we may expect in this region. In order to answer the questions concerning "where" and "why," we will call on some general concepts of plate tectonics. Answering the "how big" question will require a discussion of earthquake magnitude and other means of characterizing the "size" of an earthquake. The question of "how often" will cause us to look at some elementary statistics of earthquake distributions and the importance of the historic record. Finally, our discussion of the severity of effects will necessitate the introduction of the idea of how we characterize destructive ground motion and how the severity of motion depends on the local situation.

Whether or not the scientific community is ever able to reliably predict earthquakes, engineering decisions need to be made every day based on our present state of understanding of the earthquake risk. Thus, the principal task of a seismologist interested in reducing the hazards due to earthquake is to develop an understanding of how geologic and seismologic parameters affect motion. This is necessary because we need to predict in advance the nature of ground motion for an earthquake that has not yet occurred and all we have to look at is the geology and the record of past earthquakes.

PLATE TECTONICS AND EARTHQUAKES

The plate tectonic model of planet Earth is the starting place for understanding the "why" and "where" of earthquake occurrence. In the simplest sense, earthquakes are the "noise" or creaking and grinding disturbances that accompany the motion of tectonic plates. In this view, the plates (with associated continents riding along on top of some of them) do not move smoothly at rates of a few centimeters a year; rather, they move spasmodically, with a jump during each large earthquake, such that the average motion viewed over thousands (or millions) of years is several centimeters per year. Of course, the entire plate does not have to lurch forward during a single earthquake, but significant distortion and

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movement could be expected every time a large portion of any its boundaries slips. That earthquakes are associated with the boundaries of these plates can easily be seen by looking at Figure 1, which illustrates the global pattern of earthquake activity. Narrowing our view to the Pacific Northwest, we have the plate configuration illustrated in Figure 2.

Plate Boundaries

A plate has three types of boundary--a spreading ridge boundary, a subducting zone boundary, and a transform fault (or edge) boundary. In the simplest view, the ridge has the smallest earthquake occurrences because the lithosphere is thin and hot (weak) near a ridge and, thus, a large area of potential slip (and, thus, a large volume in which to store strain energy) does not exist. In contrast, the subduction zone boundary appears to be the place where the world's largest earthquakes (great earthquakes) occur. This is because the lithosphere is cooler, thicker, and stronger and because a larger area of potential slip exists (the entire interface between the overriding and underthrusting plates). Transform faults or plate edges appear to be intermediate between these two extremes with a limit on the depth extent of faulting, but with a horizontal extent that can be quite large as in the case of Chile, Turkey, and California. It would appear that large earthquakes, but perhaps not truly great earthquakes, are possible on transform faults. The distinction between "large" and "great" for engineering purposes ultimately may be important because of the size of area affected rather than because of distinction in the severity of ground motion. This is true since in recent years it has become clear that even moderate earthquakes can produce very severe ground motion locally.

Subduction Zones

Looking in more detail at the conditions that affect the potential "size" of earthquakes on subduction zones, we find that the two most important parameters seem to be the age of lithosphere and the rate of plate motion (covergence). A simple model of the downgoing slab, which progressively grows cooler and thicker as it moves out from its source region at the spreading ridge, is that it is sinking vertically under its own weight while also being subjected to relative horizontal convergence as the overriding plate moves over it. All other things being equal, the faster it tends to sink because of negative buoyancy, the less normal stress there will be between the two plates and the more likely it will be able to move smoothly (without a stick-slip type motion) and, thus, the smaller the earthquakes are likely to be. In the limit of a plate that is sinking so fast that it is actually separating (trying to separate) from the overriding plate, it is unlikely that large earthquakes could occur at all. The single most important parameter that seems to control the density of the downgoing plate and, thus its buoyancy, is its age. The older and colder the plate, the more dense it is and the faster it will sink. The other parameter is the plate velocity (covergence rate).

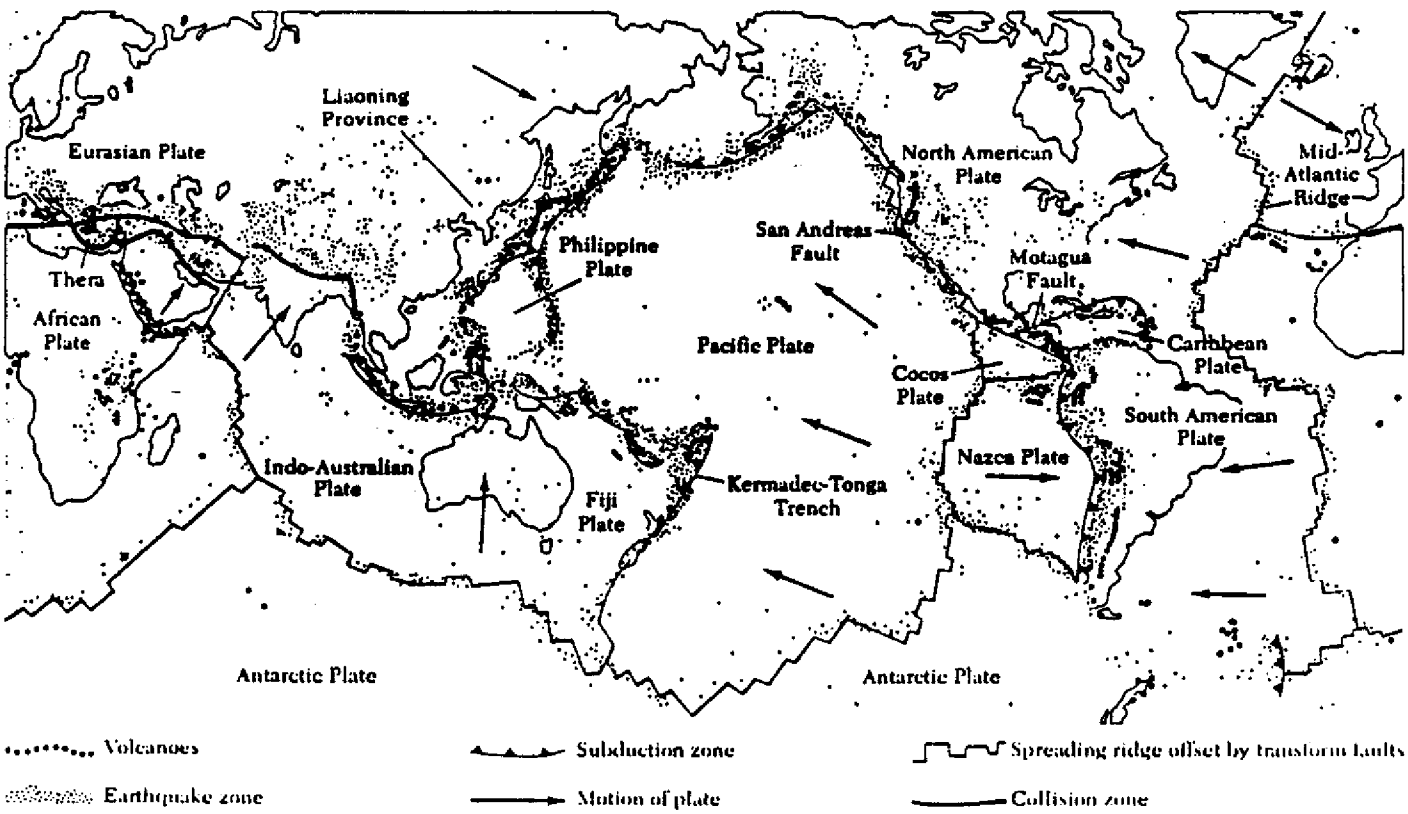

FIGURE 1 World map showing relation between the major tectonics and plates and recent earthquakes and volcanoes. Earthquake epicenters are denoted by the small dots, and the volcanoes by the large dots.

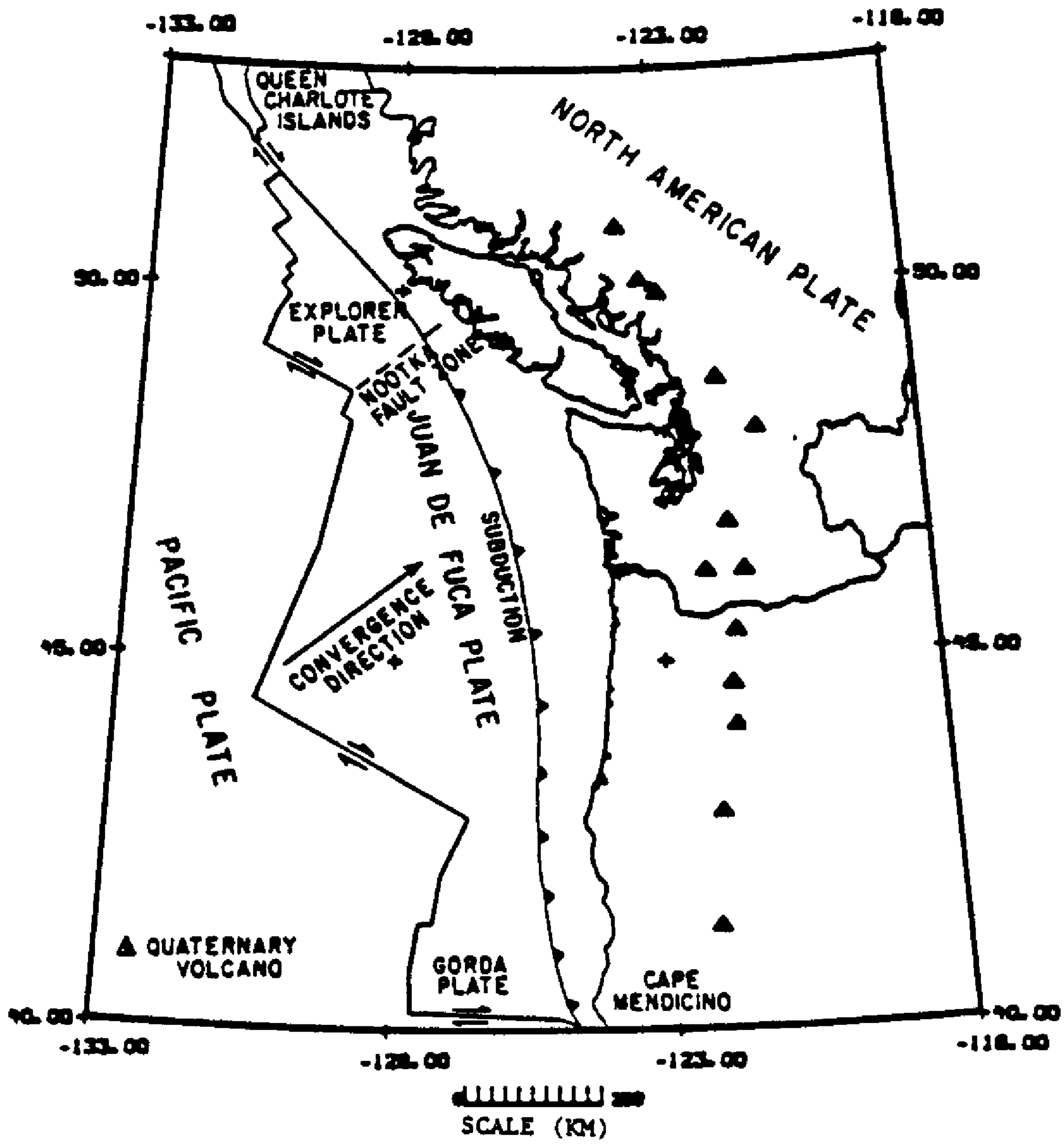

FIGURE 2 Juan de Fuca Plate map.

Here, for a constant sinking rate, the faster the two plates are converging, the more normal stress there will be locking the surface between them. This, in turn, leads to a situation of large stress accumulation and, thus, large earthquakes.

The correlation between lithospheric age and convergence rate shows, for example, that in the Pacific Northwest, where the Juan de Fuca plate has an age of less than 20 million years off the coast of Washington and a convergence rate of about 3.5 cm/yr., the expected value of moment magnitude for the largest possible earthquake is 8.25. The scatter in the data revealed in the multiple regression work by Heaton and Kanamori would cause one to put an uncertainty of about ± 0.4. The remarkable thing about this analysis is that here we have a region where the historic record is less than two centuries and there are no reports of earthquakes larger than around 7.5 and, yet, a model based strictly on geologic data and the plate tectonic hypothesis leads to a prediction of an earthquake as large as 8.5.

Transform Faults

In trying to apply similar kinds of basic physics to transform faults to see what parameters influence the maximum size of earthquakes, we have much less success. It appears to be only the top 20 or so kilometers of crust that can support brittle fracture; therefore, the size of the possible slip area is controlled primarily by the length of the fault. Complexity of the fault, lateral inhomogenieties and bends or kinks, appears to be important in determining how long a section might rupture in a single earthquake event. Thus, the detailed surface geology is critical and no generalizations can be made. Transform faults or plate boundaries are of several varieties depending on which types of plate boundaries the transforms connect. Plate edges between two offset ridges (RR transform) can be easily modeled with a piece of cardboard in which two slots are cut and through which two pieces of paper (appropriately marked with magnetic stripes) can be pulled. Two lessons are learned from this paper model. First, the relative motion on the transform fault connecting the two ridges is opposite to that which would be expected if one thought that the ridges had been offset by a fault that connected them and that they originally had been a throughgoing feature. More important from the standpoint of assessing possible earthquake size, however, is that the ends of the fault, which extend beyond the ridges and are called fracture zones (FZ), have no relative motion and, thus, can be viewed as fossil faults on which there will be no earthquakes generated. Thus, a fracture zone that is a thousand kilometers long can generate a rupture only as long as the segment joining the two actively spreading ridges. Even in the case of the transform fault, the plate tectonic hypothesis provides some important guidance as to the earthquake potential of this feature. My own view is that we have seen only the beginning of the way in which our understanding of the physics (and chemistry) of the earth will affect our assessment of future earthquake hazards.

FAULT AND EARTHQUAKES

Up to this point, we have viewed the only source of earthquakes to be plate boundaries, and our view of plates has been one of a grand scale where there are some 17 major plates comprising the entire surface of the planet. Looking closer, we find that this view is only an approximate one and that the earth is very much more complicated. In some instances the plate boundaries are razor sharp and easy to identify, whereas in others the boundary may be spread out over hundreds of kilometers or greatly obscured by the possible subdivision of the plate into many smaller platelets (the term "microplate" is starting to become popular). When we come to the hard question of estimating the future earthquake activity in a region, it sometimes seems that we have simply substituted one crystal ball for another when we try to invoke ideas of plate tectonic models and the plates themselves are not easily understood. Let us leave the simple plate viewpoint for the moment, recognizing that even if we had a simple plate model at depth, what we would see at surface is likely to be obscured by the local geology (e.g., mountains, sedimentary basins). In examining how the surface rocks may deform or fracture (fault) in response to deeper plate movement, we can use some the ideas of fracture mechanics to relate stresses to resulting fault type and pattern.

Normal Faults

A normal fault is one in which the slip direction is down-dip in such a way that you would expect to develop if the region were stretched and the blocks readjusted accordingly. Typically the dip of normal faults is quite steep, between 45 and 90 degrees. (Remember, dip is measured from the horizontal downward). In terms of earthquake potential, one would not expect a great deal of normal stress pressing the two sides of the fault together since the region is undergoing horizontal tension (being pulled apart). Thus, all other things being equal (which in geology they never are), one would not expect the largest earthquakes to occur on such faults. Substantial earthquakes, however, have been observed on normal faults (e.g., Dixie Valley, Nevada, in 1954 and Hebgen Lake, Montana, in 1959). These faults had vertical displacements of up to 4 or 5 meters over distances of nearly 100 km so they were "big" earthquakes by any measure but they were not "great" earthquakes in the sense of the Alaskan earthquake of 1964. Our 1949 earthquake near Olympia (magnitude 7.1) was apparently on such a fault although it occurred on the deep part of the subducted slab where we cannot directly observe it.

Reverse Faults (and Thrust Faults)

A reverse fault is also a fault on which the slip is in the direction of dip, but in this case it is the upper block (hanging wall) that is pushed up so the sense of motion is opposite to that discussed for the normal fault. Typical dips for reverse faults are 45 degrees or less. When the dip gets to be very shallow, almost horizontal, then the term "thrust" fault is used to describe it. There are numerous examples of nearly horizontal thrust sheets where, over geologic time, the upper block has slid many miles on top of the lower sheet. One could expect large normal stresses to develop across such faults (since the two sides of the fault are being pushed together) and, thus, very energetic earthquakes. A recent example of a thrust type earthquake was that in the San Fernando region of southern California in 1971. Since the Juan de Fuca plate is being thrust beneath North America, this is the type of faulting that could conceivably occur beneath western Washington. Should this occur, there would likely be quite severe ground motion over the entire region from the Pacific Coast inland to the Cascade Mountains.

Strike Slip Faults

Finally, we have the case of nearly vertical fault surfaces with slip in the horizontal direction. Such faults are called "strike slip" and are classified as to right or left lateral depending on the sense of motion with respect to an observer standing on one side of the fault and looking across it. The famous San Francisco earthquake of 1906 (magnitude 8.25) occurred on the San Andreas fault, which is a right lateral strike slip fault. During that earthquake the fault slipped as much as 17 feet in some places. The recently noted alignment of earthquakes through Mt. St. Helens extending to the northwest is believed to be a strike slip fault based on indirect seismological evidence although geologic data that would confirm slip on this fault has not yet been uncovered.

Earthquake Potential of Mapped Faults

Examination of virtually any geologic map will reveal that there are a multitude of faults on a variety of scales present nearly everywhere. In fact. the density of faulting on maps seems to depend largely on how carefully the area has been mapped by geologists and how good the exposures of bedrock are. Areas like the Puget Sound region may not show many faults, for example, if they are covered by a thick blanket of recent glacial material which makes them inaccessible for geologic mapping. The scale of faulting varies from tiny, millimeter-size features you can see in a rock fragment up to global-size features that are best seen in satellite imagery. Obviously not all these features have the same potential for generating earthquakes. Size or length of faulting is an obvious distinction, but perhaps the most important characteristic is the age of most recent movement.

Age of Most Recent Movement

Most observed faults are very old, representing past periods of deformation under stress conditions that are very different from what we have today. In geology we do our forecasting somewhat like the meterologist does his when he uses the "strategy of persistence"--i.e., the most likely conditions for tommorrow are more of what we have seen today. In that sense, the faults most likely to cause a problem by generating earthquakes are the ones that have the most recent history of movement. The development of radioactive age dating techniques, particularly those that involve short half life elements like Carbon 14, and can be used to date materials as young as thousands of years, provides the means to distinguish very young and, thus, potentially dangerous faults from those that that are old and no longer active. Investigations are generally made by trenching across the fault trace, or boring through it, with careful mapping of the materials on either sides. The key is to find features that are continuous across the fault and to date these features. For example, an old soil layer that lies uninterrupted across a dip slip fault and has an age of 2,000 years tells us that the fault has not moved in at least 2,000 years. Conversely, if the soil layer were disturbed, it would establish that the fault had moved sometime (exactly when could not be said) in the past 2,000 years.

In western Washington our heavy glacial cover obscures most fault features that might be useful in assessing the record of past earthquakes (and guessing the future ones). Some evidence of ancient fault motion on the Olympic Peninsula was developed a number of years ago by dating trees that were submerged as a possible effect of fault-dammed streams. Some lineaments are visible in air photographs of the Cascade Mountains and in side-looking radar imagery (SLAR), but their significance is not as clearly understood as would be the case in California or Nevada where the overall record of surface geology is much better preserved. In the Mojave Desert of California, fault scarps that moved thousands of years ago are so well preserved they look as if they might have moved yesterday. In contrast, here in the Northwest the rate of growth of vegetation (such as Douglas fir) and the erosion due to heavy rainfall are so great that faults can easily be obscured in a short period of time. In addition, the plate tectonic configuration is basically different in the Pacific Northwest from what it is in California. In California, the boundary between the Pacific and North American plates is a nearly vertical fault plane (or collection of planes) that intersects the surface of the earth producing obvious features (e.g., the San Andreas fault). In contrast, our plate boundary in the Northwest lies beneath us, the gently dipping interface between the Juan de Fuca plate and the North American plate. Its only intersection with the surface where one might look to see its expression is under water several hundred miles offshore.

Definition of Capable Fault

The technology for recovering the history of fault movement has developed remarkably during the past decade driven by society's need to assess the "capability" of faults in connection with large dams and nuclear power plants. There are no firm rules to tell us how old a fault has to be before we can classify it as inactive. It seems to be a sliding scale depending on how high the stakes are. In the case of nuclear power plant siting, a specific criteria has evolved in which a fault that has moved at least once in the past 50,000 years must be considered "capable." Generally, however, if there is no evidence of movement since the last period of glaciation, approximately 10,000 years, it appears unlikely that future movement will occur.

CRUSTAL DEFORMATION

Obviously, with all the plates stretching, squeezing and colliding with one another, there should be some possibly measurable deformation going on between earthquake occurrences. In the earliest days of seismology, an earthquake was attributed to either explosive action or magma movement deep in the earth. It wasn't until the 1891 earthquake at Mino-Owari in Japan that serious consideration was given to sudden fault slip being the cause of an earthquake. The excellent set of geologic and geodetic data that was collected before and after the 1906 San Francisco earthquake, however, really set the stage for the first rational explanation of earthquake sources, the "elastic rebound" theory.

A number of fundamental questions remain to be answered concerning the slow deformation that precedes (and follows) major earthquakes. The tools to measure these effects are available, primarily laser distance measuring devices both land-based and satellite-based, but since the motions are slow, it is going to take quite a few more years before many of the questions are satisfactorily answered. For example, how does the stress increase in the years (possibly centuries) leading up to the earthquake? Is it rather steady, simply building gradually to a point of failure and then starting over again to produce a periodic recurrence of earthquakes? Alternatively, is the stress quiescent most of the time, with rapid periods of buildup just prior to large earthquakes? These two possible scenarios lead to quite different strategies for predicting future earthquakes.

SEISMIC WAVES

We have been using sudden fault slip or rupture as a working model for an earthquake source. The phenomenon that we normally associate with an earthquake, however, is ground-shaking. What's the relation between these two observations? The ground-shaking we notice some distance away from an earthquake (and some time after the faulting occurred back at the hypocenter) is simply the effect of seismic waves that have traveled from the hypocenter to our point of observation. The principal shaking motion that is experienced in an earthquake is due to two broad categories of seismic waves, namely, "body waves" and "surface waves." The term "body wave" means a disturbance that travels directly through a solid medium, choosing a path that is the quickest possible route between source and receiver. There are two general types of body wave, compressional or P waves and shear or S waves. Surface waves travel along the surface of the earth in a manner somewhat analogous to water waves. They also come in two varieties--Love waves that produce strictly horizontal shaking and Rayleigh waves that cause vertical as well as horizontal shaking.

For a number of fundamental reasons, the frequency of both types of surface waves, Love and Rayleigh, is much lower than that for the direct body waves, P and S. As a result, surface waves are of much more concern for long period structures such as bridges and high-rise building than for more conventional structures. Simple consideration of how the wave energy spreads out in a surface wave (two-dimensional or cylindrical traveling along the surface) compared with body waves (three-dimensional, spherical waves traveling through the medium) tells us that the wave amplitude will die off faster with distance for a body wave than it will for a surface wave. As a result, if a site is near an earthquake, it will most likely be the body waves that do the damage, whereas if the epicenter is a long distance away, it is more likely that the surface waves will present the largest motion.

EARTHQUAKE SIZE

We have now established that earthquakes are the sudden slip or rupture on a fault plane and that the shaking we observe is a result of seismic waves produced by that fault slip. Intuitively, we might expect more intense shaking from a fault that had a relatively large amount of slip. We also might expect more intense shaking if the fault surface on which slip took place was a large one since that would permit constructive interference effects to occur. As a result, the measure of earthquake "size" should somehow include both the amount of slip as well as the size of the fault area.

Now, the observable quantity we have available to measure earthquake size is generally a seismogram. Only very rarely do we have the opportunity to directly measure fault slip and area. Thus, we need a measure of earthquake size that depends on something we can measure on a seismogram, such as the amplitude of some particular seismic wave. In the early development of the magnitude scale, Charles Richter at Caltech simply measured the maximum amplitude on seismograms. To avoid differences in the response of different kinds of instruments, he restricted himself to a particular type, namely, the Wood-Anderson torsion seismograph. This instrument has two attractive attributes for development of a magnitude scale. First, it is a very "broad band" instrument that responds uniformly to vibrations of both very short and very long period. Second, since it is a mechanical-optical device, there are no amplifiers, variable resistors, or, in fact, any knobs at all that can be twiddled to change its sensitivity. Thus, it is nearly "technician proof," and even years after an earthquake has been recorded, one can have confidence in the published sensitivity of the instrument.

Richter Local Magnitude, M_L

Richter noted that the maximum amplitude on seismograms behaved in a organized way. Although there were rapid variations in amplitude and a lot of scatter in data, he found that the maximum amplitude data formed a one-parameter family of curves when the logarithm of the amplitude was plotted versus the logarithm of distance. The free parameter was some kind of arbitrary number which denoted the "size" of the earthquake. He defined that number as the local magnitude and it has been denoted as M_L. There is an arbitrary "starting point" for this scale and he chose it such that a magnitude 0 shock would have an amplitude of 1 mm at a distance of 100 km.

Body Wave Magnitude, m_b

Richter didn't specify which seismic wave he was measuring, he simply chose the largest excursion on the record. Since the instrument was measuring horizontal motion and since he was generally dealing with local (nearby) earthquakes, the maximum always corresponded to the SH wave. Subsequent work using earthquakes from further distances showed that this process was inadequate. As waves travel through the earth they preferentially lose their high frequency constituents and, thus, appear longer in period (lower frequency) the further away you observe them. It was found that dividing the amplitude by the period provided a convenient and useful way to normalize out this effect. It was also necessary to have a scale based on compressional waves as recorded on vertical instruments. The resulting relationship with some empirical corrections added to make it fit reasonably well with the M_L scale looked like:

$$m_b = \log(A/T) + 0.1\Delta + 5.9,$$

where A is the amplitude of ground motion, T is the period of the wave, and Δ the distance.

Surface Wave Magnitude, M_s

It soon became clear that a single number, either M_L for nearby earthquakes or m_b for distant ones, wasn't adequate to describe the "size" of an earthquake. Two earthquakes of the same magnitude might produce remarkably different damage effects, and they certainly could write remarkably different looking seismograms. One of the big differences was in the amount of surface waves generated, and this observation soon led to the development of yet another magnitude scale. It utilized the amplitude of Rayleigh waves at a period of 20 seconds. Because of some waveguide effects in the earth, this period usually corresponds to the maximum part of the train of Rayleigh waves and is thus easy to identify. The resulting expression for surface wave magnitude, again adjusted so that it corresponds as closely as possible with the other magnitude scales, is:

$$M_S = \log A + 1.66\log\Delta + 2.0.$$

Seismic Moment

In addition to these empirical studies, which led to several magnitude scales that were very useful in classifying earthquakes, there were mathematical developments that led to a characterization of the strength of a seismic source. In the differential equations that describe the motion of an elastic medium, there is a source term expressed as a force. We have no way to describe an earthquake as some kind of force system since we are unable to observe forces directly in the earth and it seemed that there was no apparent way to use an earthquake as the source term in the equations of motion. This situation improved after the development of a mathematical representation theorem that showed how a dislocation (fault slip) model could be expressed as an equivalent force. An important parameter was identified in the resulting equations, the product of rock strength, fault area, and average slip:

$$M = \mu A\bar{u}.$$

It has the dimensions of a "moment," (i.e., force times length) so it was called "seismic moment." Here was a parameter that could be measured from a seismogram and could also be directly related to observations that a geologist could make in the field. It also formed the basis of a calculation of energy or work done during an earthquake, and this, in turn, was used to develop yet another (hopefully the last) magnitude scale, the so-called moment magnitude.

STATISTICS AND RECURRENCE CURVE

One of the first ways of utilizing the magnitude scale was in examining the size distribution of earthquakes. It is immediately clear that there are more small earthquakes than large ones so the question concerns whether the distribution behaves in some organized fashion. The answer, of course, is yes! If we choose a particular area of the earth and record earthquakes over some specific time, then plotting the log of N_M, the cumulative number of earthquakes that exceed magnitude M as a function of magnitude, yields a straight line:

$$\log N_M = a - bM.$$

The intercept "a" is a measure of how active the region is and the slope "b" tells us how many small shocks there are for each large one. We will have only a segment of a straight line because we will run out of data at both ends of the magnitude distribution. There will be some magnitude so small that it will escape detection by our seismic networks, and there will be some upper limit, namely the largest magnitude shock that has occurred during our time of observation. Within this range of magnitudes, the distribution generally does fit a straight line quite well with the slope ranging from 0.5 to 1.2.

An important question concerns how far we can extrapolate such a line to predict the rate of occurrence of earthquakes larger than those that have already been observed. It would be very convenient if one could record and count earthquake statistics in a region for a short period of time, say several months or even several years, and from this data determine both the maximum magnitude that could be expected in the region and its recurrence period. Unfortunately, this procedure doesn't work because without some additional information about the faults, their behavior, and the age of most recent movement, we do not know how to extrapolate the earthquake statistics to large magnitude. To illustrate this, Figure 3 shows the earthquake distribution for the Puget Sound region. Figure 4 shows a map distribution of the earthquakes that have occurred in Washington since 1841. Note that the largest event shown is the 1949 Olympia earthquake and that if this curve is a fair representation of the long-term seismicity, we should expect a repetition of a shock of this size every 130 years on the average. Can we extend the

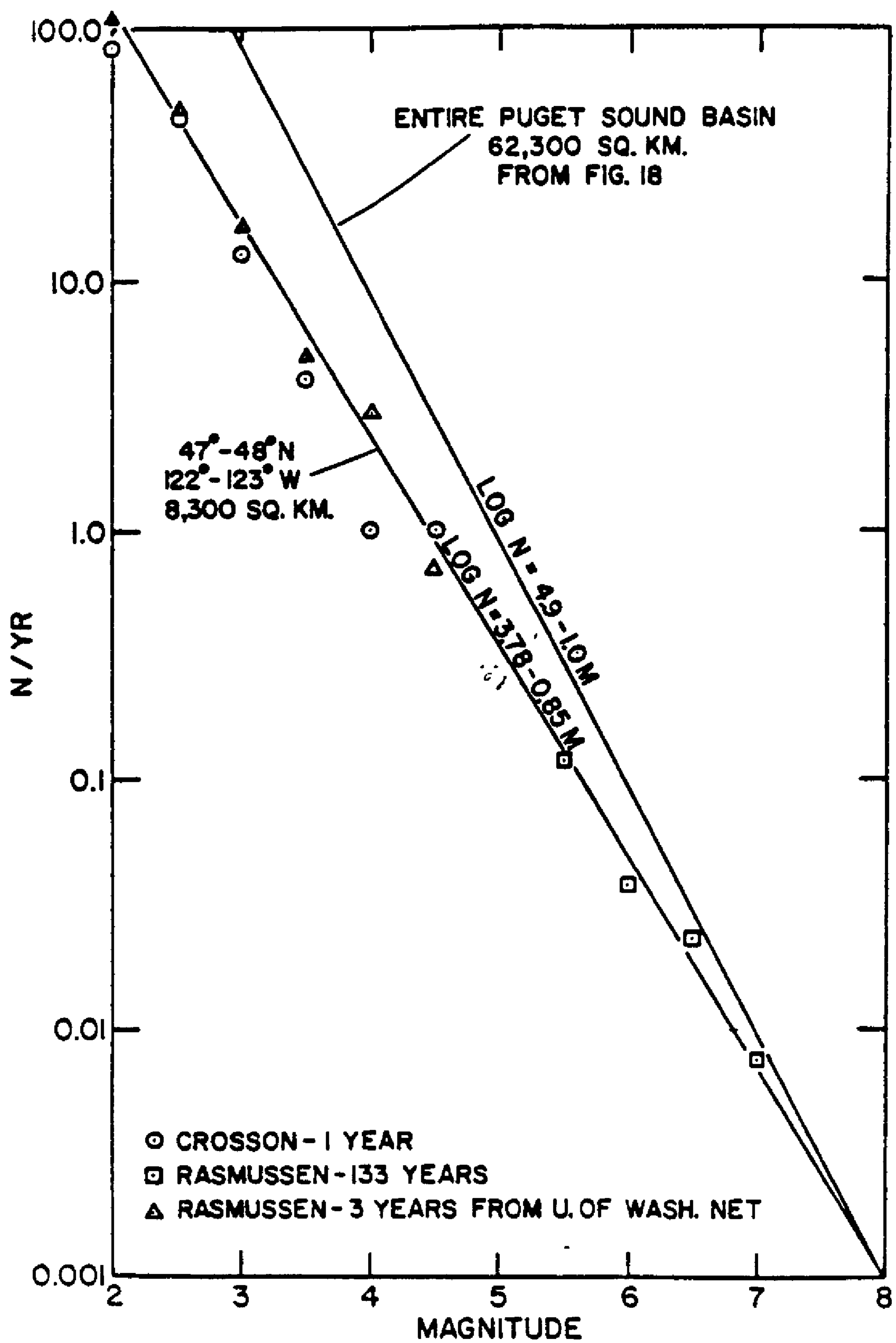

FIGURE 3 Frequency of occurrence vs magnitude for the entire Puget Sound Basis and the one degree area.

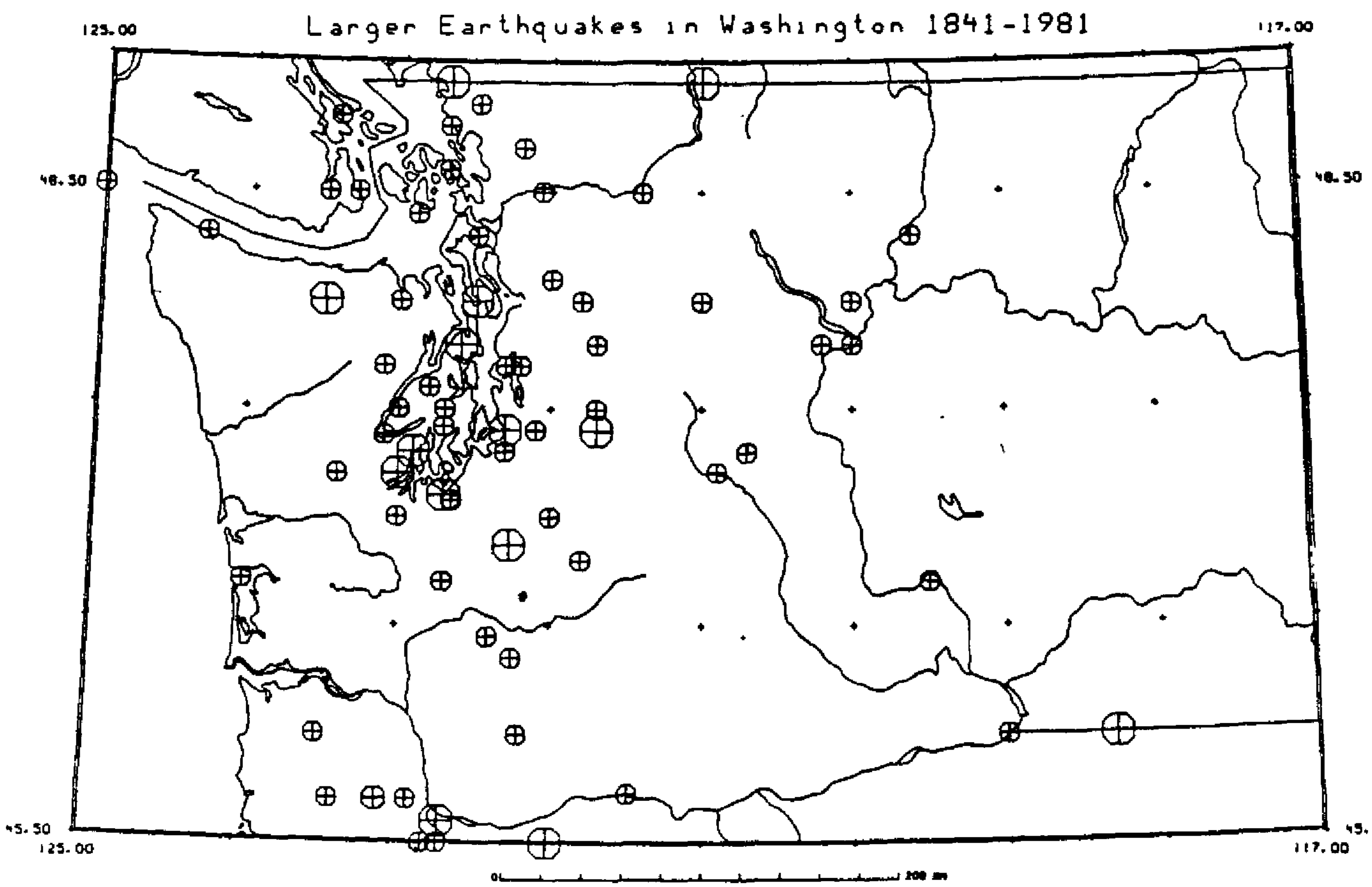

FIGURE 4 Larger earthquakes in Washington, 1841-1981.

curve to larger magnitude? If we do, how often would we "expect" a magnitude 9.0 quake and would this make any geologic sense? The Pacific Northwest is a good illustration of the pitfalls of using such curves because we suffer from a very short historic record, a poorly preserved surface geologic record, and a plate geometry not well suited for producing surface fault scarps. Thus, the critical information needed to intelligently use the meager earthquake statistics is simply not available.

GROUND MOTION

When the ground shakes during a nearby earthquake, we may (it does require some luck) obtain a record (strong motion seismogram) that displays the history of ground-shaking. A considerable amount of information is present in such records, but for our purposes we will mention only a few parameters that can be easily obtained. First, we have the maximum of acceleration, velocity, and displacement. In Figure 5 we illustrate a ground motion recording from the 1949 Olympia earthquake, magnitude 7.1, arguably the largest earthquake to have occurred in historic time. Note that acceleration is measured as a percent of the acceleration of gravity (g) or in units of cm/sec^2 reached a value of 134 cm/sec^2 or 13 percent g for this particular record. Velocity and displacement records are obtained by integrating the original acceleration record once and twice, respectively. Second, we have the duration of strong shaking, which can be defined, for example, as the length of time during which the shaking exceeded some particular value such as 5 percent g. Finally, we have some measure of the frequency content, basically a measure to describe how the energy of shaking is distributed between high and low frequencies. A variety of measures are possible ranging between simply the period of ground during which the maximum motion occurred to a response spectrum which displays the maximum motion that would be encountered by hypothetical buildings (single degree of freedom pendulums) of differing resonant frequency.

Intensity

A completely different way to characterize ground motion is through its damage effects on structures. Earthquake intensity scales are used for this purpose. For the United States, the modified Mercalli scale is the most popular. It characterizes ground motion from I to XII by a series of descriptions ranging from I as barely perceptible through VI where we see the onset of building damage to XII where one has "total destruction." The principal usefulness of such scales is to characterize the "size" of ancient earthquakes for which there are no measurements of actual ground motion. Another useful measure is the area over which the earthquake was felt since this information can often be easily determined from old newspaper reports by simply noting in what localities the shaking was felt.

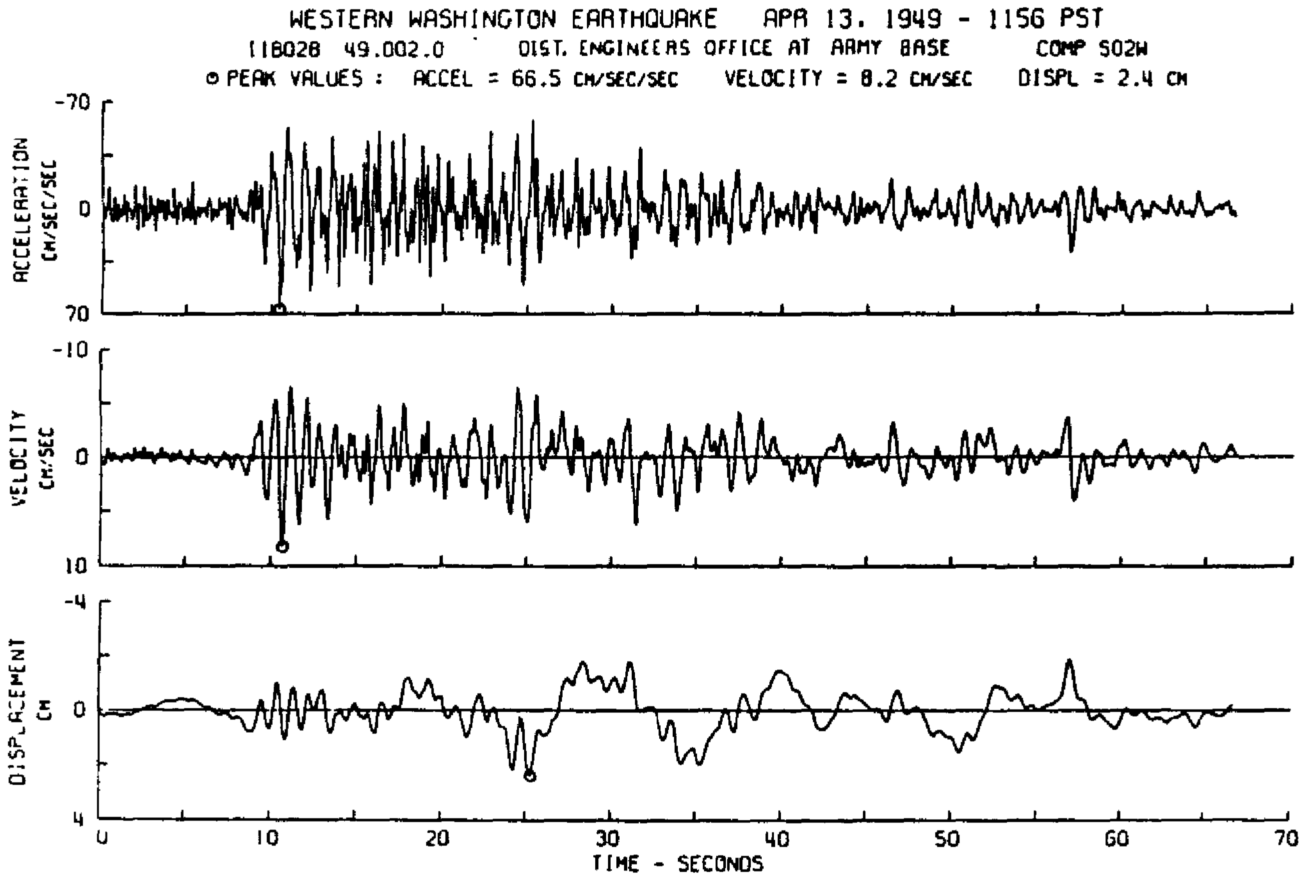

FIGURE 5 Western Washington earthquake, April 13, 1949, 1156 PST IIB028 49.002.0 Dist. Engineers Office at Army Base Comp. S02W Peak Values: Accel. = 66.5 cm/sec/sec Velocity = 8.2 cm/sec Displ. = 2.4 cm.

Attenuation Curves

Obviously, any of these "measures" of ground motion will be more severe for an observation site close to the earthquake than it will be for a more distant location. Attenuation curves are the device we use to display this relation. Any parameter can be used to construct an attenuation curve, even intensity. Typically we display the logarithm of peak horizontal acceleration as a function of distance for one particular size earthquake. The shape of this curve depends critically on a number of seismologic and geologic parameters such as fault type, depth, crustal thickness, and specific dissipation (Q^{-1}). This last parameter is a measure of how much of the elastic energy in a wave is converted to heat as the wave passes through the crust. Thus, each region will have its own distinctive curve. Such a curve, when constructed with locally derived ground motion data, together with a recurrence curve, also locally derived, and a map of the potential earthquake source regions are the basic ingredients that one needs to calculate seismic risk.

CONCLUSIONS

Western Washington lies on top of an active subduction zone. Although the characteristics of this zone are not yet well understood, comparing it with other subduction zones around the world leads us to predict that an earthquake as large as 8.25 on the moment magnitude scale could happen here. The effects of such an earthquake would not be localized to a narrow fault zone such as is the case for the San Andreas fault in California but might be spread widely from the coast inland to the Cascade Mountains and from Vancouver Island to the Columbia River. Although the scientific evidence points toward the possibility of an earthquake of this size, we have not yet been able to determine if such an event is likely to occur once per century or once per millennium. It is this rate of occurrence that will determine if the risk from such a large earthquake is greater than the risk we know for certain exists due to the repetition of smaller historical earthquakes such as those of 1949 and 1965.