PRELIMINARY ESTIMATES OF THE STRENGTH OF PREHISTORIC SHAKING IN THE COLUMBIA RIVER VALLEY AND THE SOUTHERN HALF OF COASTAL WASHINGTON, WITH EMPHASIS FOR A CASCADIA SUBDUCTION ZONE EARTHQUAKE ABOUT 300 YEARS AGO

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U.S. Geological Survey Open-File Report 94-589

1995

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ABSTRACT

Liquefaction-induced features recently discovered in islands of the Columbia river show that seismic shaking accompanied coastal downdropping about 300 years ago. These findings agree with previous studies that attributed the downdropping to tectonism along the plate-boundary of the Cascadia subduction zone. A regional inventory of liquefaction-induced features has been used as a means to bracket the strength of prehistoric subduction earthquakes. The searched areas focused on islands in the Columbia River, and the banks of the Chehalis and Humptulips Rivers in western Washington.

A preliminary geotechnical engineering analysis, made to assess the strength of shaking required to form the dikes in the islands, yields results consistent with theoretically and statistically derived accelerations for a M~8 earthquake whose energy source is located about 35 kilometers offshore. The liquefaction effects are evaluated by using the procedure of Seed et al., in terms of 0.5 probability estimated in a logistic regression. The method is not dependent on locating the most distal effects of liquefaction in the field, but rather requires testing at sites of marginal liquefaction.

In the Chehalis River valley scattered dikes of earthquake-induced liquefaction origin were also discovered, but only at distances further than 60 km from the coast. Source beds commonly were sandy gravel, indicating strong shaking was required for their formation. Ages of the dikes are not well constrained, but most formed within the past two to three thousand years. Some of these dikes greatly predate the subduction event of about 300 years ago. Most or all of these dikes in the Chehalis River valley likely had a local seismic source not directly related to rupture of the plate boundary.

Nearer the coast, in the Humptulips River valley, there is a virtual absence of dikes in sandy gravel that probably should liquefy during very strong shaking. The paucity of dikes very near the coast indicates that onshore seismic shaking could not have been very strong within the past few thousand years. Therefore, for any very large plate-boundary subduction earthquake of the past few thousand years, an offshore energy source of shaking is probable.

If the magnitudes of subduction earthquakes within the past two thousand years much exceeded M 8, the accelerations seem to have been extraordinarily low or alternatively the locked zone was much further than 35 km offshore. The dominant vibration frequency may also have been very low.

INTRODUCTION

Buried tidal-marsh soils in coastal Washington and Oregon provide excellent evidence for prehistoric, late Holocene episodes of sudden submergence accompanied by tsunamis. The submerged marshes most likely reflect flexure of the crust by tectonic thrusting along the interface where the Juan de Fuca plate is subducting beneath the north American plate (Atwater, 1987, 1992; Darienzo and Peterson, 1990). Inferred earthquakes originated by subduction of the oceanic (Juan de Fuca) plate beneath the continental (North America) plate -- the Cascadia subduction zone (Fig. 1). Great (M 8 to 9) subduction earthquakes have been inferred partly on the basis of the exceptional length along the coast that appears to have downdropped at about the same time, and perhaps simultaneously (Atwater, 1987, 1992; Darienzo and Peterson, 1990; Nelson and Atwater, 1993). Radiocarbon dating of coastal downdropping allow that possibly more than 700 km of coastline was tectonically warped simultaneously in the latest downdropping event (Nelson and Atwater, 1993). Key evidence is provided by the regional extent of downdropping and presence on the buried tidal marsh soils of a sheet of sand interpreted as having been laid down by a tsunami. Geologic evidence for widespread multiple turbidites in the offshore region likewise suggests the possibility of great Holocene earthquakes (Adams, 1990).

The possibility of recurrent great subduction earthquakes in coastal Washington and Oregon was first suggested from indirect geophysical evidence and by comparison of the Cascadia subduction zone with tectonic settings worldwide (Heaton and Hartzell, 1987). However, no strong Cascadia earthquake has occurred during the time of written history in the Pacific Northwest, some 200 years. Modern seismicity on the subduction zone is limited to scattered, small earthquakes -- none with thrust mechanisms except along the southern boundary of the subduction zone in northern California.

Atwater (1992) inferred from subsidence stratigraphy that at least two great earthquakes have struck the coast of Washington, including the region around the Columbia River valley, during the past 2,000 years. The presence of widespread, abruptly buried soils provides strong evidence that one event was about 300 years ago (ca. 300 ybp), and a less widespread buried soil indicates that another event occurred between 1,400 and 1,900 years ago. The portion of the plate boundary that ruptured and provided energy for seismic shaking was most likely a small distance offshore (few tens of kilometers), considering altogether the location of the subsided zone (Atwater, 1992), heat flow data (Hyndman and Wang, 1993), and strain data (Savage and Lisowski, 1991; Draggert et al., 1994). A sectional view showing the implicated extent of coseismic rupture is depicted in Figure 2.

The strong shaking from great subduction zone earthquakes in the vicinity of the coast would be expected to have produced abundant liquefaction features, even in sediments having moderate to low liquefaction susceptibility. Therefore, I undertook a search for liquefaction features in the banks of rivers near the coast. The rivers examined are shown in Figure 3. For purposes of discussion, the search areas are designated in two categories, the Columbia River valley and smaller river valleys.

This text is intended to be readily understandable to both geologists and engineers. Some few terms, though, are used in their geotechnical engineering sense because of the lack of geologic equivalents for a more quantitative description of certain sediment properties. For example, the state of compactness of sand (i.e., relative density) is constrained by semiquantitative descriptors such as "very loose" to "very dense". The term "liquefaction susceptibility" refers to the ease at which a saturated sediment liquefies, and may be qualified with terms such as "very low" to "very high". The term "fluidization" is a geologic descriptor and refers to the process whereby flowing water exerts sufficient drag to suspend grains of sediment, thereby destroying original bedding and transporting grains; for fluidization to occur does not require liquefaction, but only flowing water.

This text has been written so as to permit the reader to go directly to the DISCUSSION section, on page 22.

COLUMBIA RIVER VALLEY

Many large islands in the lowermost Columbia River were searched between Marsh Island, located about 35 km east of the coast, and Bonneville Dam, located more than 150 km east of the coast (Fig. 3). (See Appendix, Note 1, for listing of islands searched.) No islands with suitable outcrop exceeding a few hundred meters in length were found nearer the coast than the vicinity of Marsh Island. The banks along the Sandy River, near Portland, were also searched at the confluence with the Columbia River.

Sand-filled dikes of seismically induced liquefaction origin were discovered from Marsh Island upstream to as far west as Deer Island, located 90 km from the coast. Further inland, about 125 km from the coast, possible small sills and small dikes that may have formed by seismic liquefaction were discovered in the banks of Sandy River, as well as in the banks of the Columbia River near the confluence with Sandy River.

Sites Downstream from Deer Island

The islands in this portion of the Columbia River originated as braid bars on a grand scale. Thick deposits of sand underlie large portions of most of these islands, which are flat, poorly drained, and swampy. Large portions of the islands are submerged during the highest tides. Strong currents and wave pounding are severely eroding many islands and as a result have sculpted clean vertical banks as much as 2 meters above water level. Significant areas (tens of thousands of square meters) are also being cleaned in plan view by tides.

The banks of the islands between Marsh and Wallace Island (Wallace Island is about 65 km inland) expose mainly structureless clay-rich silt deposits intercalated with a few thin (1-2 cm) silt horizons (Fig. 4). The clay-rich deposits are generally dark blue (unoxidized), and soft (engineering sense). Very weakly developed wetland soil horizons also occur. Age of the clay-rich cap just above the level of low tide is less than 800 years and more than 600 years on most islands, on the basis of conventional radiocarbon ages on fossil marsh plants (genus Scirpus) found in growth position.

Regional stratigraphic control on the clay-rich cap is excellent. About 1.5 m below the top of the banks there is a tan horizon of a few centimeters in thickness. This horizon is exceptionally rich in volcanic ash. Very locally, rounded pumice clasts as large as 5 cm in diameter occur in this ash horizon. Chemical analysis shows that the ash and pumice of the tan stratum has minerals and elements identical with those of tephra (set W) of an eruption from Mount St. Helens, in A.D. 1480-1482 (C.D. Peterson, 1992, Portland State Univ., written comm.). About 10 to 15 cm lower is a blue-grey horizon, generally several centimeters thick, also rich in ash. Therefore, the radiocarbon ages on fossil marsh plants

(600 to 800 years) and the ash data show that exposed sediments are old enough to record liquefaction associated with the ca. 300 ybp downdropping event, but probably are not old enough for the event 1,400 to 1,900 years ago (e.g., Atwater, 1992). Several meters below the ash layers, below water level, radiocarbon age data show that the sands probably have ages between 1,500 and 2,000 years at many places (Atwater, 1994).

Sand is exposed at low tide immediately beneath the clay cap on many islands. This sand ranges from being clean, fine- to medium-grained to clayey, silty fine sand. Extensive deposits of clean sand are widespread, commonplace, and they are tens of meters thick at many places. Sediment and ground-water conditions on many islands are nearly ideal for formation of large liquefaction-induced features. Not only should the thin cap (commonly 1 to 3 m) enhance venting (Ishihara, 1985; Obermeier, 1989), but the ground-water table has almost certainly been at or within a meter or so of the ground surface since the islands formed. The tidal range at these islands is about 2 to 2.5 m, and high tides inundate parts of the islands and doubtlessly have done so for at least several hundred years.

More than three hundred dikes were found along more than 9 km of clean vertical banks between Marsh Island and Deer Island. Maximum dike widths tend to decrease upstream (Fig. 3, Table 1), although erosion almost certainly has removed some of the largest (see Atwater, 1994). At some places dikes were exposed in plan view. The density of dikes in plan view also generally decreases going upstream, although large variations occur from site to site. Only 10 small dikes were found on a large outcrop area on Deer Island, which is far inland from the coast (90 km), whereas the highest density (with many tens of dikes) was found on Brush Island ((35 km), the site located nearest the coast. No dikes were observed on large parts (perhaps one-half) of all the larger islands, even where the cap was thin.

Figure 4 illustrates stratigraphic and venting relations of liquefied sand commonly observed downstream from Cathlamet. A thin sand sheet lies on a very weakly developed soil (organically enriched) that is about 1 m below the present surface. The sand sheet is 1 to 4 cm in thickness and is as wide as 10 m. Connected to the sheet is a nearly vertical, narrow planar dike that widens downward markedly and can be traced to the source stratum of sand. Where pits were dug in sand just beneath the dikes, flow structures in the sand could be observed going into the base of the dike. Within the uppermost 5 to 15 cm of most dikes, the dike width normally is several millimeters or less. For the widest dikes (those dikes wider than 15 cm), however, sidewalls seem to be parallel and nearly vertical. Sills as thick as 5 cm were observed. Sills appear abundant very locally, but numerous vertical exposures to the base of the cap show that sills commonly do not occur within the cap.

Samples of small sticks were collected at many sites to bracket the time when venting occurred. Typical of others, at one site a small limb with a stem lying on the vented sand had a radiocarbon age of 230 $\pm$ 70 years (calendar age between A.D. 1520 and 1810). A similar limb with a stem about 15 cm beneath the vented sand had a similar radiocarbon age of 250 $\pm$ 70 years, which for practical purposes is the same age as the stick lying on the vented sand. The oldest trees rooted in sediments above vented sand are on the order of 200 years, as determined from tree ring counts. The oldest tree has an age of 230 years (Atwater, 1994). Sand at one island (Price) was observed to have vented at the level of a buried swamp surface delineated by dead roots connected to living trees more than 350 years old; the live roots are at the modern surface, at the level of highest tides (Atwater, 1994).

All dikes located downstream from Cathlamet (Fig. 3) are interpreted to have been caused by the earthquake associated with the coastal subsidence of ca. 300 ybp, for the following reasons: (1) the radiocarbon ages of sticks on and near the surface of venting are consistent with the event of ca. 300 ybp; (2) trees rooted in sediments above vented sand have the same maximum ages (about 200 years); (3) dikes tend to increase in abundance toward the coast; (4) maximum dike widths tend to increase toward the coast, and (5) the relation of the tephra layer to the surface onto which sand was vented is regionally the same. The 1-m thickness of silt and clay above vented sand is interpreted to have been deposited because of submergence, following the regional downdropping from the subduction zone earthquake of ca. 300 years ago. The thickness is controlled by the level of high tide, with there being no deposition of sediment above high tide. This 1-m thickness lies well within the estimates of coastal tectonic submergence (0.5 to 2 m) by Atwater (1987; 1992), and Darienzo and Peterson (1990). Interpretations on the basis of the five factors above, with the exception of the ages of the living trees, were made in 1992 by Curt Peterson (Portland State Univ.), by David Tabaczynski (USGS), and by me. The following year copious details supporting this interpretation were provided by Atwater (1994).

The 1- to 4-cm thick sheet of vented sand may be exceptionally thin because of erosion caused by tidewater and wave action. The surface of venting is submerged at moderate to high tide. Tidewater flows relatively fast and waves are large in this area, so any large cones of sand initially vented to the surface could have been beveled off, and the sand scattered over a large area. A contributing factor could have been subaqueous venting.

The regional stratigraphic control provided by the two thin ash-rich strata (Fig. 4) is missing only for a portion of Wallace Island (upstream part) and for Deer Island. However, where the ash layers are absent, radiocarbon data indicate that the sediments cut by the dikes or that lie above the dikes have ages that are slightly older to much older than the earthquake of ca. 300 years ago. At Wallace Island, plant debris cut by dikes yielded a radiocarbon age of 420 $\pm$ 60 years (Atwater, 1994). At Deer Island, an archeological hearth that I excavated in sediments above deposits cut by dikes yielded a radiocarbon age of 690 $\pm$ 70 years.

The stratigraphic and radiocarbon control is insufficient to narrowly bracket the ages of the dikes at Wallace Island and at Deer Island, in contrast to further downstream. The range of possible dates of origin admits that the Wallace and Deer Islands dikes could have been caused by earthquakes other than the event of ca. 300 ybp. For example, small sand blows were observed to have formed (Chleoborad and Schuster, 1990) in the nearby town of Longview (Fig. 3) during the 1949 Olympia earthquake (M 7.1). Still, the regional pattern of maximum dike widths and the general upstream attenuation of dike abundance agrees with the expected overall pattern for the subduction earthquake event of ca. 300 ybp (Fig. 3). To provide a conservative (high) estimate of the strength of shaking for the earthquake of ca. 300 ybp, it will be assumed that the dikes at Wallace and Deer Islands were caused by that earthquake.

Sites Upstream from Deer Island

Many islands were searched between Deer Island and Bonneville Dam. At least 25 km of well exposed, vertical banks of silty clay were searched upstream from Deer Island to Government Island (about 10 km upstream from Portland, Fig. 3). Ages of the banks examined greatly exceed the earthquake of ca. 300 ybp, as shown by archeological evidence and by severity of weathering (formation of B-horizons) in sediments of the banks. No liquefaction effects were observed in this portion of the search region. However, it is likely that all the outcrops searched are underlain by clays that are too thick (greater than several meters) to have been hydraulically fractured by any liquefaction at depth during the event of ca. 300 ybp, on the basis of data from water-wells and engineering borings in the region.

The only suspected evidence for liquefaction upstream from Deer Island was discovered in the vicinity of the Sandy River, about 15 km east of Portland (Fig. 3). Here there are terraces elevated about 5 m above the Columbia River, which permitted observation of deformation features to much greater depth than in the marshy islands downstream. The possible liquefaction effects are very minor. Evidence is present in terraces that form both banks along Sandy River from the confluence with the Columbia River to more than 1 km upstream. Evidence is also in a 1-km long exposure of the same terrace that forms the bank of the Columbia River, just upstream from Sandy River. Throughout the region, the base of a very friable cap appears to have been corraded, and numerous small sand sill-like features occur along the base of the cap. Rare features thought to be planar sand dikes intrude up into the cap in the vicinity of Sandy River. The dikes are very small features that pinch together upward. Their widths are less than a centimeter and heights are less than 0.5 m. Within the source sand, small planar sand dikes cut irregularly throughout at least the upper 2 m. Similar small features were observed in outcrop on Reed Island (Curt Peterson, 1994, Portland State U., oral comm.), which is an Island in the Columbia River a few km upstream from Sandy River (Fig. 3). Both on Reed Island and in the banks of Sandy River, there is a thin stratum of ash that corresponds chemically to a Mt. St. Helens ash, radiocarbon dated at 410 $\pm$ 70 ybp. This ash stratum has been intruded and warped by the underlying fluidized sand (Siskowic et al., 1993; Anderson et al., 1994).

The small sill-like features and corrasion along the base are pervasive laterally along the contact of the source sand with the capping deposit, where the cap is underlain by clean sand. The friable silty sand cap is about 2 m in thickness. The source sand is fine-to medium-grained, with no clay and almost no silt. Thickness of the source sand in outcrop typically exceeds 3 m in the vicinity of Sandy River, and possibly is much greater. At Reed Island, though, the source sand can be seen at low tide to be bounded beneath by a thick clay (Curt Peterson, 1993, Portland State Univ., oral comm.).

The nearly ubiquitous lateral development of the minor corrasion effects along the base of the cap, combined with the unusually large number of sill-like features relative to dikes, raises the question of whether these effects were caused by earthquake-induced liquefaction. Plausible alternate origins include effects of artesian conditions, and wave-induced liquefaction during flooding. Artesian springs probably can be eliminated because of the widespread development of fluidization effects on opposite banks of the Sandy River, as well as development on Reed Island in the Columbia River. Hydraulic connection of all these widespread sites (several km apart) with an artesian source is extremely implausible; in addition, the source sand on Reed Island is underlain by a thick clay. Wave-induced liquefaction likewise would seem to be eliminated because the features occur in different river valleys. Furthermore, small tabular dikes along Sandy River extend at least 2 m below the base of the cap, where they can be seen to cut across thin silty sand beds. I have observed similar small dikes cutting across sand beds at depth, caused by seismic liquefaction, to be commonplace in the meisoseismal region of the 1811-12 New Madrid earthquakes (Obermeier et al., 1990, Fig. 22C).

Even though the dikes and possible sills may have a seismic origin, there remains the question of the seismic source. Local historic seismicity near Portland records six events of magnitude 5-6 (Siskowic et al., 1994). These local events have left no apparent evidence of liquefaction in the youngest flood deposits (0-100 radiocarbon ybp) along Sandy River. The observation that the dikes and sills cut and warped the Mt. St. Helen's ash, radiocarbon dated at 410 $\pm$ 70 ybp, allows the possibility that these intrusions were induced by the subduction earthquake of ca. 300 ybp. This possibility is enhanced by the observation that historic sediments (0-100 ybp) in the region appear to be undeformed (Siskowic et al., 1994).

Preliminary Assessment of Strength of Shaking

A preliminary quantitative assessment of the strength of shaking for the earthquake of ca. 300 ybp is made by using a conventional geotechnical engineering method, the procedure of Seed et al. (1983, 1985), to estimate threshold accelerations in the source sands that liquefied; these accelerations are then adjusted to 0.5 probability in a logistic regression. The adjusted accelerations are then compared with predictions from seismological models and statistical analysis.

The Seed et al. procedure (Fig. 5) is used by engineers to predict sites where liquefaction could occur in the event of a future earthquake. The procedure is based mainly on data from sites where liquefaction has been severe enough to cut through the cap and vent sand and water to the ground surface. In order to use the curves of Figure 5, the relative density of the source sand must be evaluated. This evaluation is generally done by the Standard Penetration Test (SPT) blow count method. This blow count is then adjusted to account for factors such as depth to water table, depth of source stratum, and fines content, which yields a N_1 value (Fig. 5).

Sites where only marginal liquefaction occurred are most suitable for paleoseismic analysis. In this study at the Columbia River islands, sites most suitable for using the Seed procedure were at Wallace and Deer Island, because of the presence of especially large areas with only minor evidence of liquefaction effects, although large areas with no surface evidence of liquefaction are also plentiful at islands downstream.

Venting to the surface during the earthquake of ca. 300 ybp probably occurred upstream at least as far as Wallace Island; at one locale on Wallace Island there are dikes that pinch together very near the ground surface of 300 years ago (within ~1 m), and at another locale there are dikes at depth connected to an overlying buried sand horizon that originated either as a sill or, more likely, as vented sediment. At both these locales, engineering tests were performed to assess the in-situ properties of the source sand beds.

The dikes at Deer Island were observed only in plan view. No dikes were seen extending into a nearby 2 to 3 m high vertical outcrop that was stratigraphically above the plan view exposure. Length of the vertical outcrop was about 700 m. Because of this absence of evidence for venting, paucity of dikes, and the small size of the dikes at Deer Island, the assumption that venting only marginally developed at these sites on Deer Island would seem to be reasonable. Engineering tests were performed on the source sands at these sites on Deer Island.

Engineering data were also collected nearer the coast, at Hunting Island and at Marsh Island. Almost all coring and penetration testing sites, other than at Deer Island, were selected by Brian Atwater (USGS) or by Steve Palmer (Washington DNR). The test sites they selected generally were not chosen where only scattered small dikes had developed, but at sites of lateral spreading.

Engineering data were collected in the general vicinity of sites of liquefaction along Sandy River, where there were no observations of venting, so backcalculations of accelerations there should yield upper bound values.

The field engineering data for all test sites, except sites at Deer Island and along Sandy River, are reported in Atwater (1994). Data at Deer Island were collected by the writer. Sandy River results are from Anderson et al., 1994.

The Wallace Island and Deer Island sites are particularly important because ground motions in alluvium using SHAKE can be modeled with more confidence here than further downstream, where depth and properties of alluvium can only be estimated more approximately. Crystalline bedrock lies directly beneath alluvium at Wallace and Deer Islands. Results of a seismic refraction survey by W. D. Mooney (U S Geological Survey, written comm., 1993) shows that crystalline rocks with a P-wave velocity of 4 km/s extend to a depth of 2 km, and at greater depth the velocity increases to 6 km/s.

Methodology

The curves for the Seed et al. procedure in Figure 5 are based on worldwide observations during many earthquakes (Seed et al., 1983, 1985). The curve locations were chosen primarily by judgement. The curves were developed from plan view sites where liquefaction effects were observed, mainly in the form of sand blows, indicating that liquefaction at depth was extensive. An absence of sand blows was taken to mean that any liquefaction was minor. The curves were not intended to indicate whether or not minor liquefaction occurred, but rather whether or not a threshold severity of liquefaction occurred. This threshold thickness of a source sand can exceed 1 to 2 m in many field settings where the cap thickness is 1 to 2 m (Ishihara, 1985; Obermeier, 1994). Some data points used to develop the Seed et al. curves were probably taken at sites where liquefaction was rather severe, in that large quantities of sand were vented. A secondary source of data used to develop the curves was from sites where there was no more than very minor lateral spreading (on the order of 8-10 cm in width) (Liao and Whitman, 1986; C.S.C. Liao, oral comm., 1994). The method is intended for level ground sites; the method is not intended for sites of significant lateral spreading, especially near free faces, because of the ease with which lateral spreading can develop there. For paleoseismic analysis, therefore, the sites that are most applicable are where small sand blows developed on level ground, far from any free faces. Sites of significant lateral spreading can yield backcalculated accelerations that are too high (Castro, 1987; E.C. Pond, Virginia Tech., written comm. 1994).

Liao et al. (1988) have used statistical regression methods to assess the probability of liquefaction based on the Seed et al. model above. Their results are shown in Figure 6, normalized in terms of magnitude. The curves apply for a M 7.5 earthquake.

Although the Liao et al. (1988) results basically agree with the trend of the Seed et al.

curve, it can be seen from Figure 6 that there are some noteworthy differences. Figure 6A shows that for clean sand deposits, the shape of the curves differs slightly, and the Liao et al. curve at 50 percent probability almost everywhere lies above the Seed et al. curve. This difference is understandable in terms of the original intent of the Seed et al. curve for engineering design, where there is a need to obtain a conservative (low) estimate of the strength of shaking required to cause liquefaction. If, however, the intent is to reverse the process to backcalculate the maximum ground acceleration that could have occurred at a site where only marginal liquefaction with limited venting occurred, then the engineering conservatism in the Seed et al. curve could lead to a value that may be too low. Whether or not the Seed et al. value is too low depends on the constraints on liquefaction with venting at site; these constraints can depend very much on factors such as cap thickness, strength of the cap, and depth to the water table. If all factors are favorable for venting (discussed below), the Seed et al. curves probably provide a good assessment for backcalculations. Unless conditions are for venting are restrained considerably, the Liao et al. 50 percent probability curve should almost certainly lead to an assessment that is not too low.

It was noted above that the curves of Figures 5 and 6 make no attempt to directly account for field conditions not conducive to venting. Elimination of such sites would likely increase the probability associated with the curves of Figure 6. In general, an increasing cap thickness increasingly restrains venting. A cap thickness exceeding about 5 m is sufficient to prevent venting in many field situations that I have observed in the central U. S. A cap thickness of only 1 to 2 m, though, seems to be about optimal for extensive venting (Obermeier, 1994). A 1.5 to 4 to meter thickness range is the norm at the lower Columbia River islands. Another factor of importance is the whether or not there are sufficient roots in a cap to greatly increase the tensile strength of the cap, thereby inhibiting hydraulic fracturing to form dikes in the cap. For this condition, sills can form extensively. Such extensive root development was not observed in the lower Columbia River islands. It has also been suggested to me that the cap in the lower Columbia River islands may be too soft to form dikes at many places; my field experience throughout the New Madrid seismic zone and in the Wabash Valley of Indiana-Illinois shows though that soft clays and silts, at least as soft as those of the islands, readily fracture in a brittle manner and form dikes extensively (Obermeier, 1994). A deep water table also cannot be invoked as a mechanism that inhibited venting on the lower Columbia River islands. The water table doubtlessly was very shallow on the islands at the time of the earthquake of ca. 300 ybp. The only factor that may have had an influence on venting on the islands was the presence of well-defined bedding planes near the base of the cap, which could have enhanced sill development; however, this same condition in the New Madrid seismic zone and the Wabash Valley did not prevent extensive formation of dikes. On the lower Columbia river islands, therefore, all properties of the cap and the field setting appear to be very favorable to extensive formation of dikes with venting, if liquefaction at depth had been anything but minor.

In summation, I believe that the curves of Seed et al. would probably yield assessments of peak accelerations that range from being most reasonable to only slightly too low, at sites of marginal liquefaction where the cap is thin on the lower Columbia River islands. To be conservative, though, the 0.5 probability curve of Liao et al. will be used for the backcalculations. (For the Columbia River islands, increasing the Seed et al. accelerations by about 20 percent raises them to the 0.5 probability curve of Liao et al.)

Some uncertainty in using the procedure of Seed et al. to backcalculate accelerations is whether or not there have been large changes in the blow count values of the source stratum. Where liquefaction has been severe, field data indicate there may be large changes, mainly increases, in the N_1 values of the source sands. Such changes are shown in reports by Chameau et al. (1991), Koizumi (1966, p. 40), Kishida (1966, p. 75), and Kawakami and Asada (1966, p. 18). Loose, thick sands appear to be especially susceptible to an increase in blow counts. No guidelines are available to estimate changes in density caused by an occurrence of liquefaction. Even though the reports cited above indicate an overall tendency for densification of the source sand, a review of the literature (Obermeier, 1994) shows that there also appears to be a strong tendency for formation of an especially loose sand in a thin zone just below the cap. This sand directly below the cap is likely looser than before the occurrence of liquefaction. The loosening probably occurs in response to seismic liquefaction because of (1) water flowing up from deeper source zones, (2) water collecting beneath the cap (Elgamal et al., 1989; Dobry and Liu, in press), and (3) sills that obviously were in a very fluidized condition at the time of their emplacement forming along the base of the cap (Tuttle and Barstow, in press). I believe that the thickness of the loosened zone is probably less than 1 meter except in regions where liquefaction of thick source beds has been extreme; this judgment is based on the field observations by me and by Tuttle and Barstow (in press) that sills only exceptionally exceed a meter in thickness even in the meisoseismal region of the 1811-12 New Madrid earthquakes, and that SPT data in that region (Obermeier, 1989) generally indicate the possibility of only a thin ($\leq$ 1 m) zone of loosening.

I suspect that below the thin zone of loosening beneath the cap, there may have been some densification of the source beds at liquefaction sites on the Columbia River islands. The field measurements of blow counts in the source beds seem to be higher than values on sands in the region measured by highway departments. Liquefaction should have occurred preferentially at sites of loosest sands, yet the blow counts at the liquefaction sites generally seem higher than blow counts in sands at bridge crossings. (However, the high blow counts at the liquefaction sites may also be a reflection of the technique used to collect the blow counts, discussed below.) Figure 6 shows that a small increase in a relatively high N_1 value, such as that of the Columbia River islands, can cause a significant increase in the backcalculated acceleration.

Some densification of sands at the Columbia River islands may also have occurred because of seismic shaking since the earthquake of ca. 300 ybp. The 1949 earthquake of Olympia (M 7.1) caused liquefaction in the Columbia River valley that was reported at Longview and at Cathlamet (Fig. 3) (Chleborad and Schuster, 1990). Seismic shaking even without liquefaction also commonly causes densification (Castro, 1987).

Despite the possibility of an increase in the value of N_1 caused by the earthquake of ca. 300 ybp, the acceleration corresponding to the 0.5 probability curve in the likely densified zone will be used to estimate the peak accelerations. In addition, the blow counts will be conducted in a manner as to eliminate any influence of a loosened zone just below the cap. The backcalculated accelerations should thereby not be too low, and are probably somewhat high.

Another parameter that has a bearing on the backcalculated acceleration is the duration of strong shaking. An increase in shaking duration reduces the peak acceleration required for liquefaction. For an earthquake of M~8, doubling the duration of strong shaking can decrease the threshold acceleration by about 10 to 15 percent (National Research Council, 1985, Table 4-2). If the duration of strong shaking of the subduction earthquakes was extraordinarily long, use of the Seed et al. procedure (0.5 probability value) should again yield estimates of acceleration that are too high.

Once near-surface acceleration estimates are made, the ground motions at the alluvium-bedrock contact at depth can be estimated. This is done by using one-dimensional program SHAKE (Schnabel et al., 1972) to deconvolve the surface accelerations to bedrock motions. These bedrock motions are then compared with predictions of accelerations using seismological models and statistical relations for a M 8 Cascadia plate-boundary subduction earthquake.

Field Geotechnical Investigation

Preliminary geotechnical investigations were performed at the following islands, located at increasing distances from the coast: Marsh Island (35 km), Hunting Island (50 km), Wallace Island (65 km), and Deer Island (90 km). Samples for estimating grain sizes of the source beds were collected by driving an aluminum tube (~10 cm diameter) into the ground using the vibracore method. The source sands were found to be fine, uniformly sized, and with some silt at most places (Fig. 7). In sectional view, the liquefaction sites where blow count data were collected generally have a 2 to 4 m thick cap of soft, plastic silty clay (see Atwater, 1994,p. 54) underlain by moderately dense sand in the engineering sense of Terzaghi and Peck (1967). The liquefiable sand extends at least to the maximum depth of investigation, 6 m, at all sites tested.

The raw blow count values (see Atwater, 1994) of the source sands were found to tend to increase with depth. Directly beneath the cap there is generally a thin zone (less than 0.7 m) of sand much looser than at depth. This thin zone possibly corresponds to a region loosened by liquefaction (discussed previously).

The field procedure for obtaining blow count values made use of a hand auger to first advance a hole through the surface clay layer. A dynamic penetrometer (Note 2) then was used to obtain a continuous record of the penetration resistance (recorded as the number of blows required to drive the probe 1-foot increments) of the underlying sand. The dynamic penetrometer is a small portable device, much smaller than the tool normally used to measure the penetration resistance. This required converting the raw blow counts to the penetration resistances corresponding to the equivalent Standard Penetration Test (SPT) N-value. The suggested conversion factor of the manufacturer was used. These N values were then normalized to account for in-situ overburden stress conditions of 1 kg/sq cm, and an adjustment was made to account for the fines content according to the method described in Seed et al. (1983, 1985). (An increasing silt content decreases the liquefaction susceptibility.) This procedure yielded the value (N_1) in Figure 6 that is used for the analysis of strength of shaking.

Insufficient data have been collected to carefully calibrate the dynamic penetrometer to SPT blow count values. The calibrated values are best viewed as tentative. The dynamic penetrometer N-values seem to be somewhat higher than those using standardized tools by the highway departments, as noted previously. Thus results using the dynamic penetrometer are suspected to yield backcalculated accelerations that are too high.

Analysis of Data

Sands beneath the cap were tested in their uppermost 3 to 5 m. For the liquefaction analysis, use was made of two N_1 numbers, the median and the lowermost 25 percentile values. In this way the N_1 values in most of the upper 1 m of sand (the possible loosened zone) are eliminated from the analysis. (Dickenson et al. (1994) used the mean and the mean minus one standard deviation values of N_1, which is a slightly less conservative method of analysis).

Penetration data were collected at five test holes near dikes on Wallace Island. The median N_1 value was 18 and the 25 percentile N_1 value was 14. Estimated peak accelerations for a M 7 earthquake are 0.13 and 0.17 g, respectively, for the 25 percentile and median values, using the Seed et al. method. For a M 8 earthquake, peak accelerations are 0.11 and 0.15 g, respectively, for the 25 percentile and median values. Thus, the range is estimated to be 0.11 to 0.17 g for M 7-8 earthquakes. Three of the penetration test sites were near a 15-cm wide lateral spread, however, which casts doubt on the backcalculated value.

A more reasonable assessment of actual peak acceleration is probably provided at a site on Wallace Island far from any possible effects of lateral spreading, where dikes pinched together near the surface but did not vent. One hole extended to a depth of 6 m, where a relatively loose sand was encountered. The median value of N_1 was 26, and 25 percentile value was 24. Refinement between such close numbers is not warranted. Peak accelerations using the Seed et al. procedure for M 7 and M 8 earthquakes are estimated as 0.22 g and 0.19 g, respectively. Table 2 shows results using the Seed et al. procedure and the 0.5 probability value of Liao et al. (The foregoing shows that the calculated accelerations using the cyclic stress ratios are not very sensitive to this range of assumed earthquake magnitudes; the N_1 value of the source sand is the much more important parameter for the analysis.)

This method of evaluation was repeated at the other islands. Venting at two other islands, nearer the coast than Wallace Island, still was sporadic over large areas, although numerous closely spaced, pinching upward dikes also formed over large areas. At Hunting Island, the borehole with highest blow counts at a site of venting had a median N_1 value of 28 and lower quartile N_1 value of 26; these values are in the uppermost 3 m of possible source sands. For M 7 and 8 earthquakes, respectively, these yield accelerations of about 0.24 g and 0.19 g. At a site of probable strong venting, above a 20-cm wide dike, N_1 values are no higher than at the site of venting described above, but the lack of grain size data makes interpretations equivocal.

A 10-cm diameter tube sample driven into the sand beneath the cap at Hunting Island contained a portion of a dike that was intersected at a depth of about 5 m below the reference ash layers. Estimated N_1 values of sand beneath the dike were quite high (about 35) but whether or not this sand with the high N_1 values is the source material is unknown. Radiocarbon data on a stick just beneath the dike yielded an age of 1670 years; such a large age admits the possibility that the deep dike developed during the two to four major earthquakes that probably shook this region between 1670 and 300 ybp (Atwater, 1994). Overall, the best estimate of peak ground acceleration for the earthquake of ca. 300 ybp is a value exceeding 0.2 to 0.25 g at Hunting Island, but not greatly so.

At Marsh Island, nearer the coast (35 km), the widest (30 cm) dike (a lateral spread) occurs in the area of lowest blow counts: median N_1 values here are about 15 to 20, which is generally lower than elsewhere in the region and makes the presence of the sizeable lateral spread understandable. The N_1 values in a large area of no dikes have a median value of about 24 to 28, and a 25 percentile value of about 20 to 24, although there is some question of interpretation of values because grain size data are sparse in this area. (Possibly more silty or muddy sands underlie this site, which would cause the adjusted N_1 values to be higher.) An estimate of peak accelerations a little in excess of 0.2 to 0.25 g again seems to be a reasonable assessment for a M 7 or 8 earthquake.

On Deer Island, 90 km from the coast, no venting was observed and the scattered dikes are small. Thus the Seed procedure values corrected to the 0.5 probability may overestimate actual accelerations. Because dikes formed, though, the estimates are likely to be only a little too high. No tube samples were collected to verify the amount of fines (silt and clay) in the source strata, but bucket auger samples (8 cm in diameter) indicate the fines content was 10 percent or less . At 5 scattered penetration test sites near the dikes, the median N_1 value was 20 and the 25 percentile value was 16. For a M 7 earthquake, these N_1 values yield peak accelerations ranging from 0.14 to 0.17 g. For a M 8 earthquake, these values yield peak accelerations ranging from 0.11 to 0.15 g.

Estimated accelerations at all the island sites for a M 8 earthquake are in Table 2. The table also shows the accelerations reported for the Sandy River sites (Anderson et al., 1994). Also shown are estimated acceleration in bedrock beneath the liquefaction sites, and the estimated distance from the energy source (locked portion of the subduction zone) proposed by Hyndman and Wang (1993). Their best estimate is that the locked zone lies within 35 km of the coast in the vicinity of the Washington-Oregon boundary. Hyndman and Wang note that there is an uncertainty of about $\pm$ 30 km in the landward boundary of the locked zone.

The strength of shaking in bedrock beneath the surface of alluvium is estimated by using the calculated peak surface accelerations. At the relatively low levels of surface shaking associated with the ca. 300 ybp event, worldwide observations show that moderately thick deposits of loose and soft alluvium (such as alluvium beneath the Columbia River islands) likely amplify the peak bedrock accelerations and enhance the duration of the ground motions (Idriss, 1990). Amplification doubtlessly occurred in the 1949 Olympia, Washington earthquake (M 7.3), when it was observed that Modified Mercalli intensities in the Columbia River valley were higher in areas underlain by moderately thick alluvium than where bedrock was near the surface (Malone and Bor, 1979).

A dynamic soil mechanics analysis using SHAKE was modeled for Wallace Island by using subsurface information from a boring log nearby. From the surface downward there is a surface layer of loose silty sand extending to a depth of 12 m, underlain in turn by 50 m of medium dense to dense silty sand, 29 m of very dense silty sand, and 100 m of very dense clayey sand and gravel. Basalt bedrock is at a depth of 190 m. Recorded bedrock motions from western U.S. crustal earthquakes with M > 7.0, as well as simulated ground motions developed for a M 8 Cascadia earthquake (Cohee et al., 1991) were used as the input motions to estimate amplification of motions in alluvium above bedrock.

Results of the one-dimensional dynamic response study are reported by Dickenson et al. (1994). Their best estimate of amplification of the peak acceleration in bedrock at the surface of alluvium at Wallace Island is a factor of 2.5. Due to the uncertainties in the shear wave velocities (V_s) of the sediments, Dickenson et al. varied these values by $\pm$ 10%. Parametric studies were also performed to evaluate the sensitivity of amplifications to variations in V_s. The layer velocities were modified in an alternating sequence, resulting in different impedance contrasts for each calculation. The amplification of bedrock accelerations consistently fell within the range of 2 to 3. Whereas the amplification factor of 2.5 is somewhat higher than will be readily accepted by much of the community of geotechnical-seismological practitioners. Dickenson (oral comm., 1994) is of the opinion that the factor of 2 is a reasonable lower bound for the estimated levels of shaking in the region of Wallace Island. The possibility remains, though, that the amplification was extraordinarily high because of resonance within the narrow Columbia River valley (a two-dimensional effect).

At Deer Island, the depth of sediments almost certainly exceeds 100 m. For such a thickness the amplification factors for Wallace Island probably are representative for Deer Island. Thick sediments also probably underlie Hunting and Marsh Islands. Insufficient data were available to make estimates for the Sandy River sites.

Recent ground motion studies of the Cascadia subduction zone (Cohee et al., 1991; Geomatrix, 1993; Youngs et al., 1988; Youngs et al., 1993) provide estimates for the peak ground (bedrock) accelerations generated during M 7 to 9 subduction zone earthquakes. Figure 8 presents the most recent interpretations for M 8 and M 8.5 earthquakes. These interpretations have evolved from a composite of theoretical studies and statistical analysis of subduction earthquakes. The theoretical solution is a stochastic-based approach, using band limited, random vibration theory. The theoretically derived data points (dots) in Figure 8 show scatter because of variations used in locations of asperities and the mode of rupturing. At distances further than 100 km from the locked zone, the best fit of data for both theoretical and statistical solutions is essentially the same. Figure 8 shows that peak accelerations at rock sites during a M 8 event 35 km offshore would be estimated to be less than 0.13 g in the vicinity of Marsh Island (70 km from the locked zone), to less than approximately 0.10 g at Wallace Island (100 km from the locked zone), and less than about 0.08 g at Deer Island (125 km from the locked zone). The duration of strong shaking associated with these models is on the order of 50 to 70 seconds.

Table 2 shows that the analysis based on the liquefaction data yields accelerations that basically are the same as the best estimate provided by model values for a M 8 earthquake. However, the standard deviation for the empirical (statistical) analysis portion is quite large, so the agreement of accelerations does not constitute proof. According to R.R. Youngs (Geomatrix Consultants., oral comm., 1994), a factor of 1.9 applies to the first standard deviation of empirical acceleration values. This exceeds the factor of about 1.7 required to increase accelerations from a M 8 to a M 9 earthquake (Youngs et al., 1988, Fig. 12). The credibility of the accelerations in Figure 8 is enhanced in my opinion, though, because both statistical and theoretical solutions yield the same acceleration values at distances corresponding to the island sites.

Geologic Indicators for Strength of Shaking

Previously it was noted that the lack of evidence for abundant venting at island sites nearest the coast (within~ 35-45 km) suggests that surface shaking in that vicinity was not greatly stronger than at Wallace Island. Other qualitative geologic indicators similarly suggest that the ground shaking was not particularly strong for a long duration, even in the vicinity of the largest dikes (30-cm wide) at and near Marsh Island (35 km from the coast). The basis for this interpretation is that field observations worldwide show a reasonably well defined statistical relation between the region of strongest shaking and largest lateral spread movements and, in turn, widest dikes (Youd and Perkins, 1987). Field observations discussed below in one of the most intensely studied areas of severe historic liquefaction, the meisoseismal region of the 1811-12 New Madrid earthquakes (where three M 8-8.4 earthquakes struck), also provide guidance to observations made on the Columbia River islands.

All the liquefaction features discovered in the Columbia River islands are along eroding island banks. Some have been eroded more than 100 m during the past 100 years (Atwater, 1994) Observations of liquefaction effects at many places worldwide show that the largest lateral spread movements and largest dikes usually develop very near a stream bank or even a small scarp, within 100 m or so (see for example articles in O'Rourke and Hamada, 1989; see also aerial photographs in Obermeier, 1989). However, field data in the meisoseismal zone of the 1811-12 New Madrid earthquakes show that, where there has been very strong shaking, there are many major exceptions to having very large lateral spread movements and dikes form within a few hundred meters of a stream bank of scarp, even on level or nearly level ground; Fuller (1912, p. 49) for example, states "The spacing of the (lateral spread) fissures is very variable. They are closest together when near the banks of rivers, parallel cracks only a foot or two apart sometimes occurring, while a spacing of from 10 to 15 feet is not uncommon. In the sand-blow districts the spacing varies from several hundred feet down to less than 10 feet, the resulting extrusions in the latter case forming more or less confluent sheets of sand. ..... In the case of the large fault-block fissures the spacing is greater, several hundred feet often intervening between the cracks, <u>while the space between them may be half a mile or more</u>. Isolated cracks of this type are not uncommon." Fuller also notes that fissure (lateral spread) openings of 10 to 15 feet were not unusual.

By implication, the largest dikes generally should have developed near the margins of the Columbia River islands and toward small creeks at the time of the earthquake of ca. 300 ybp. Data collected by Atwater (1994) and by me show that four of the six widest dikes (15 to 30 cm in width) at scattered islands are quite near small streams, so it would seem the worldwide observations apply to the Columbia River liquefaction sites. The maximum dike width of 30 cm discovered in the Columbia River islands is relatively small in comparison to dikes commonly induced elsewhere by seismicity. Dike widths of a meter or more are not exceptional in the meisoseismal region of the 1811-12 earthquakes (Wesnousky and Leffler, 1992, see Figs. 12 and 17). In the Wabash Valley of Indiana and Illinois, widths in excess of 30 cm are not rare even for source beds of gravelly sand that liquefied in an estimated M~7.5 prehistoric earthquake (Obermeier et al., 1993, Fig. 1). The relatively small maximum size of 30 cm in the Columbia River islands also cannot be explained by large differences in liquefaction susceptibility of the source sediments between the various geographic regions. (SPT and grain size data for the meisoseismal region of the 1811-12 earthquakes are reported by Obermeier (1989).)

I believe that it is also not reasonable to argue that erosion of banks of the Columbia River islands removed all the large dikes, because much of the 9 km of vertical outcrop searched by the writer has probably had very little erosion since the earthquake of ca. 300 ybp. Strong currents and strong wave action probably do not impact at least half these exposures, many located in coves and other protected areas where many scattered small dikes were also observed. In summary, I believe that the field evidence indicates that all very large dikes were probably not eroded away, but rather were never abundant.

A relatively low level of ground shaking in the vicinity of Marsh Island and upstream is also suggested by another type of comparison with liquefaction effects in the meisoseismal region of the 1811-12 earthquakes -- ground warping and disruption. In the meisoseismal region of the 1811-12 earthquakes, areas with a thin (1 to 3 m in thickness) cap have been strongly warped (> 1 m vertically) over large regions. Effects of liquefaction have also torn apart the ground over large areas (Wesnousky and Leffler, 1992, Fig. 8). Part of this warping doubtlessly was caused by the expulsion of a large volume of sand to the surface. Evidence for such strong warping has not been found in the lower Columbia River islands. The marker beds of ash (Fig. 4) are very nearly horizontal at almost all places, and reflect the possibility of only minor warping (see many figures in Atwater, 1994). In addition, even at Marsh Island (Fig. 3), no dikes were observed in section or plan view over large regions.

Another indicator that shaking was not particularly strong even at Marsh Island, 35 km from the coast, is that venting to the surface only occurred erratically. Venting appears to have been much restricted where the cap was no thicker than ~~as~~ 3 m.

Others have suggested to me that the properties of the soft silt cap may have prevented dikes from forming at many island sites; they suggest, instead, that only sills formed even though severe liquefaction occurred over large regions. The basis for this suggestion is the presence of numerous, small sill-like features along the base of the cap, which have been observed in samples collected in 10-cm tubes. I believe that such an interpretation is unlikely for two reasons: (1) I have never observed in other geographic-tectonic regions where extensive formation of sills beneath a soft to stiff cap was not accompanied by dikes, and (2) the mechanics of forming sills by severe liquefaction over a large region, without also forming dikes, does not seem plausible. In order to form sills over a large area, the cap must be lifted to provide space for the intrusion. The force required to lift the cap must at least equal the weight of the cap. Simple calculations shows that the uplift force to raise a 1-4-m thick cap is much higher than the tensile strength of a soft cap. Therefore, extensive development of dikes should occur. I can envision other mechanisms whereby small sills could form without severe liquefaction, however, as a result of seismic shaking.

To summarize, the geologic indicators of strength of shaking, in terms of dike width and ground warping, do not suggest disruptions consistent with what I would expect in the meisoseismal region of a great earthquake. An energy source from a great earthquake considerably west of the searched islands would be consistent, though.

SMALLER RIVERS VALLEYS

More than 100 km of streams in western Washington were searched for liquefaction features, excluding the Columbia River region. A canoe was used to make a continuous search of stream banks. These Washington streams, collectively designated as "smaller rivers", include portions of the Quinault, Humptulips, Chehalis, Newaukum, Naselle, and Grays River (Fig. 3). The banks expose fluvial terraces of mainly Holocene age. At least 20 percent of the length of rivers traversed had banks so freshly and cleanly eroded that dikes exceeding one cm in width would have been noticed. At many outcrops there is a 2 to 4 m thickness of fine grained sediment above thick sand and gravel. Within the capping sediment are soil horizons commonly about 1 m apart that are separated by much less weathered silt and clay.

A large portion of the search was along the Chehalis River. Glaciofluvial braid-bar terraces crop out at a few places along this river. The lowest glaciofluvial terraces, of latest Pleistocene age, generally are about 5 to 7 m above the level of low flow of the modern river. Inset into the glaciofluvial deposits are sediments laid down by a younger, meandering Chehalis River. The highest meander deposits, with terraces that are about 1 to 2 m lower than the lowest glaciofluvial terrace, have well-developed soil horizons (0.5 to 0.75 m zones of strong ped development) stacked on top of one another. Terraces of meander deposits with only incipiently developed soil horizons typically are yet another 1 to 2 m lower than those with well-developed soils. Similar stratigraphic and soil relations occur along the Humptulips River. Also exposed at many places along the Humptulips River are sequences of Pleistocene glaciofluvial sandy gravel that are capped over large areas by clay-rich deposits. Observations of pedological relations in conjunction with terrace levels and radiocarbon data make it clear that outcrops expose many kilometers of sediments whose ages are as much as one to two thousands of years or older, in addition to the Pleistocene deposits. There is no doubt that sediments in the great majority of outcrops (> 20 km) along all the smaller rivers exceed 300 years in age (the time of latest coastal downdropping). In addition, it is highly probable that sediments in many outcrops are older than the tectonic downdropping event of 1,400 to 1,900 years ago reported by Atwater (1992).

The Holocene terraces along the smaller rivers almost everywhere have a clay-bearing cap with a firm to stiff consistency, 2 to 5 m in thickness. At most places the cap is made up of fine-grained overbank or channel-fill deposits. The cap sharply overlies clean granular deposits, almost invariably thick sandy gravel, although locally there is as much as a 0.3 m thickness of clean sand that lies between the cap and the sandy gravel. The total thickness of granular deposits much exceeds 3 to 4 m at almost all places. Only very locally is there bedrock within a few meters beneath the cap. Depths of gravel pits and water wells in the area suggest that the terraces are commonly underlain by 20 to 30 m of sandy gravel of glaciofluvial origin. Alluvium is generally underlain by thick Tertiary deposits of unlithified sand and clay.

The depth to the water table has almost certainly been shallow (within a few meters of the ground surface) at many places during the past few thousand years, because much of the year this region's climate approaches that of a rain forest. Even if the region had been experiencing unusually prolonged dry weather at the time of an earthquake, the water table would almost certainly have been high in the many recently abandoned channels and low-lying point-bar deposits that border the rivers.

Within 60 km of the coast, the only deformation-flowage features discovered were a small number of planar dikes containing sandy gravel that had intruded into and plastically deformed the cap. These intrusions were observed at only four sites, two along the Chehalis River, one along the Satsop River, and one along Humptulips River (Fig. 9). Dikes at one site (CH-L) were as much as 0.3 m in width. Extensive slumping appeared to be involved at two sites (CH-L and CH-M), including the site with widest dikes.

Only farther inland than 70 km are there dikes that have the classic morphology of a level-ground, seismic liquefaction origin. Here the dikes typically are thin (< few cm), planar, nearly vertical intrusions that have fractured the ground in a brittle manner. The intruded fillings of sand or sandy gravel tend to fine upward. Ten liquefaction sites were found. Three of these sites (CH-I, CH-J, and CH-K) are near Oakville, about 70 km inland (Fig. 9). The other seven sites are in the vicinity of Centralia. All but one of the dikes near Centralia appear unweathered and filled with loose sand, and on that basis most were likely induced by the nearby 1949 Olympia earthquake (M 7.1). (Liquefaction was relatively widespread in the vicinity of Centralia during the 1949 earthquake (Chleoborad and Schuster, 1990). But, because one strongly cemented dike near Centralia is cut by an unweathered dike, one liquefaction event much predated the 1949 earthquake (Moses et al., 1993). Virtually all dikes in the vicinity of Centralia pinch together upward in a thick clay-rich cap (> 5 m) whose strong weathering at the surface indicates that the host sediments have an age of at least several thousand years. No geologic relations (such as cross-cutting of sediments, truncation of dikes, etc.) were observed that could help to closely bracket the time when the dikes near Centralia formed.

Some bracketing of dike ages can be done from Oakville (Fig. 9) westward, however. Radiocarbon data at four sites (CH-I, CH-J, CH-L and HU-A), in conjunction with well developed soil profiles on sediments above plastically deformed beds at two other sites (CH-M and SA-A), admit the possibility that the dikes in the general vicinity of Oakville (say, from near Oakville to Satsop River) could have originated from a single large earthquake, sometime between 1,800 and 2,500 years ago. Table 3 shows the radiocarbon data from sites CH-I, CH-J, CH-L, and HU-A. Recent erosion has truncated the dike at another site near Oakville (CH-K), and no datable material could be found to bracket the age.

I initially questioned whether the dikes near the coast (site CH-L seaward), all of which had plastically deformed the caps and two of which probably were associated with slumping, had a seismic liquefaction origin. Only with the subsequent discovery of dikes that had brittely fractured the ground, near Oakville, in combination with radiocarbon data through the region, does the evidence seem to strongly suggest to me that all the dikes from Oakville to the coast have a seismic origin. The plastically deformed caps probably indicate that the sandy gravel was only marginally able to liquefy where relatively compact (dense) (the probable state of compactness is discussed below); because of only marginal liquefaction, there was no opportunity for forceful injection to fracture the cap in a brittle manner.

The seismic source responsible for the dikes from Oakville to the coast is uncertain. The discussion following is largely speculative, but represents an attempt to sort through the factors relevant to interpretation of the source region. My preferred interpretation is that all dikes (except the dike at HU-A) were formed by an earthquake whose meizoseismal zone was in the general vicinity of Oakville, because dikes are both most plentiful and largest (widest) in this region. The near-absence of dikes along the Humptulips River and in the lowermost Chehalis River supports this source location. An epicentral region much further east is also possible, but not very likely because of the lack of dikes between Oakville and Centralia.

Plate-boundary subduction zone earthquakes, widely spaced in time, may also be the sources of the dikes from Oakville to the coast, but are very unlikely in my opinion. It is conceivable that a very fortuitous combination of factors could have caused this regional pattern of dikes in which a great subduction earthquake produced the only dike that was found very near the coast. These factors include the properties of materials beneath alluvium, the shaking characteristics of subduction earthquakes, and the liquefaction susceptibility of alluvium. The possible nature of their roles is considered below.

Shaking at the ground surface very near the coast (< 50 km) could have been exceptionally low because an extremely thick zone of weakly lithified sediment beneath alluvium dampened shaking from the subduction rupture zone. Tertiary sediments of weakly lithified clay, silt, and sand underlie the Quinault and Humptulips River valleys, and also Chehalis River valley downstream from Porter (Rau and McFarland, 1982; S. Y. Johnson, U.S. Geological Survey, written comm., 1993). These sediments are so weakly lithified that they can be excavated with hand tools at many places. Total thickness of these weakly lithified sediments exceeds 2,000 m beneath the Humptulips valley (Fig. 10). In the Chehalis valley, between Montesano and Porter, the very weak deposits are much thinner and lithified rocks come to within 1,000 m of the surface. Upstream from Porter, the thin alluvium lies on lithified bedrock.

Data from a historic earthquake elsewhere show evidence for such strong dampening of acceleration. Chapman et al. (1990) have analyzed Modified Mercalli intensity data from the 1886 earthquake (M 7.5) of Charleston, South Carolina, and they concluded that a 1,000 m thickness of very weakly lithified Tertiary sediments that veneers hard rock around Charleston highly dampened peak accelerations there. Even as far as 50 km from the epicenter of the 1886 earthquake, they estimate that the peak accelerations could have been reduced by as much as 50 percent. They emphasize the clear-cut evidence that thinner Tertiary sediments are associated with higher Modified Mercalli intensities. Their modeling analysis also shows (Chapman et al., 1990, Fig. 11), however, that the thick sediments may substantially amplify lower frequency ground motions (< few Hz), while attenuating high frequency motions (> 5 Hz). By analogy of the South Carolina geologic setting with that of the Chehalis-Humptulips valleys (where there are large thicknesses of weak sediments nearer the coast), the factors that contribute to dampening of bedrock motions increase toward the coast. Therefore, dampening so severe as to prevent liquefaction of the gravel-rich deposits near the coast during a plate-boundary subduction earthquake nearby seems plausible for a $M \geq 8$ earthquake with an energy source 35 km offshore (about 60 km from Humptulips River), because onshore accelerations could have been relatively low.

High dampening also seems plausible along those portions of the Grays and Naselle Rivers that were searched (Fig. 3). Along Grays River, mudstones and other weakly lithified sediments probably extend to a depth exceeding 2,000 m on the basis of well data (Wolfe and McKee, 1968). Probably the Naselle valley is underlain by similar deposits. Here in the general vicinity of the Columbia River, it is only upstream from Wallace Island that crystalline bedrock is known to lie directly beneath alluvium.

Strong amplification is also improbable at most places along the smaller river valleys because (1) the alluvium is too thin to promote high amplification, and because (2) of likely similarities of shear modulus values of alluvium with underlying weakly lithified bedrock. However, further than 60 km from the coast (near site CH-L and further inland), bedrock motions locally may be amplified slightly to moderately because more lithified bedrock with a higher shear modulus comes much near the surface (Fig. 10). Of particular interest is that this region of shallower bedrock has dikes of liquefaction origin. Considering all factors, amplification of bedrock motions throughout the region of the smaller rivers was almost certainly less than beneath the Columbia River islands.

Of possible relevance to liquefaction of the alluvial sandy gravel is that subduction zone earthquakes can be especially rich in long period ground motions (Youngs and Coppersmith, 1989). The long period shaking may permit pore pressures to dissipate between cycles of shearing, particularly if the source sediments have a high permeability and are not very loose in the engineering sense. (In very loose sediments, the pore pressure increases too fast to dissipate during cycles of shearing.) Most of the sandy gravel of the smaller rivers probably is very permeable on the basis of their relatively low sand content (Fig. 7, smaller rivers, common sizes); further, it could be seen in the field that there are open voids between gravels within thin layers in most deposits. Thus, some dissipation of pore pressures during cyclic shaking in such permeable gravel seems plausible.

Still, there are many places in the smaller river valleys where there are liquefiable sandy gravels containing as much as 30 to 40 percent sand and a thickness of as much as 2 m (Fig.7). Gravel with even less sand, about 2 m in thickness, flowed and vented extensively during the (M 7.3) 1983 Borah Peak, Idaho, earthquake (Andrus et al., 1991). The Borah Peak earthquake possibly was much richer in the higher frequency shaking than would be expected from a great subduction zone earthquake. However, strong subduction zone earthquake shaking probably should not be devoid of strong high frequency shaking, especially near the source of energy, and it seems probable that some liquefaction effects should occur if accelerations are sufficiently high (Note 3).

The gravels that liquefied extensively during the Borah Peak earthquake were loose to very loose in the geotechnical engineering sense (Andrus et al., 1991 ; 1992; 1994); peak accelerations at a site of extensive liquefaction (Pence Ranch) appear to have been ~0.3 g (Note 4). Even moderately dense sandy gravel can liquefy and induce venting to the surface, though; this has been observed at several paleoliquefaction sites in the Wabash River valley of the central U. S. (Eric Pond, Virginia Tech., written comm., 1994). Engineering borings for bridge foundations in the study area of Washington indicate that 2 to 3 m thicknesses of moderately dense (blow count N-values of 15-20) sandy gravels are not common but also are not rare, although looser deposits appear to be rare. Therefore, it is possible that the sediments of the smaller rivers are too dense and too permeable to have formed liquefaction features except during exceptionally strong shaking.

Plate-boundary subduction earthquakes of magnitude greater than M 8 may not have generated particularly strong shaking on the region of Humptulips River, which is the principal river that was searched extensively near the coast. The estimated peak acceleration in bedrock beneath the Humptulips River valley for a M 8 earthquake with an energy source zone located 35 km offshore is about 0.15 g, and less than 0.2 g for a M 8.5 earthquake (Fig. 8). The alluvium in Humptulips valley is probably too thin to significantly amplify bedrock motions. In view of the underlying thick, weakly lithified Tertiary sediments, accelerations at the surface plausibly are lower than the values in hard bedrock. Therefore, a level of shaking of significantly less than 0.15-0.2 g would be consistent with the lack of liquefaction features in the moderately dense sandy gravels near the coast.

It is obvious from the preceding discussion that the use of engineering and seismological techniques to determine the source of the dikes centered around Oakville-Porter is fraught with uncertainties. The most fruitful approach to interpreting the source of these dikes is to continue to search for more liquefaction features. The regional pattern of dike ages and sizes should yield the answer. The ages of dike formation must match the ages of downdropping along the coast for a plate-boundary subduction earthquake origin.

In summation, it seems very unlikely that any of the dikes along the smaller rivers were caused by the earthquake of ca. 300 ybp, with the possible exception of the dike at site HU-A. It seems only remotely possible that many of the dikes in the vicinity of Oakville-Porter were caused by previous plate- boundary subduction earthquakes.

COASTAL MARINE TERRACES

Fluidization features discovered elsewhere along the coast also probably have a seismically induced liquefaction origin. Many ancient fluidization features have been identified recently in late Pleistocene marine terrace deposits (Peterson et al., 1991; Peterson, 1992). Ages of the deposits are thought to range from 80,000 to 120,000 years. The terraces are widespread and extend along the coast from central Washington to near the California-Oregon boundary, which is a span of about 500 km. These terraces have been variably uplifted as much as a few tens of meters, which permits viewing of source beds as well as connected dikes and sills. The fluidization features are exposed in cliffs that are being cleanly eroded by ocean waves.

Source beds for the fluidization features include beach sands and sandy gravels, and lagoonal sands. The features reported by Peterson include clastic dikes as much as 5 m in height. Dikes are filled with clean sand or gravelly sand at almost all places. Dikes widths are as much as a meter in some scattered locales. At a few places dikes have penetrated upward into dune sands or have cut through lagoonal muds and peat. Sills are also abundant. The largest sills are as much as a meter in thickness. Even thin sills can extend laterally for several meters. Sills commonly extend beneath lagoonal muds and peats. Small, steeply dipping dikes branch off from these sills at many places and cut up into thin (less than 0.5 m), low-permeability strata at the surface.

The possibility that the fluidization features were caused by wave-action that induced liquefaction has been considered because the deposits were laid down under marine or shoreline conditions. Storm waves can impose significant shear stresses in sediments on the ocean bottom, even where the water depth exceeds 60 to 70 m (E. C. Clukey, Exxon Corp., written comm., 1992). Wave-induced cyclic shear stresses are thought to cause liquefaction in sands and granular deposits in a manner analogous to seismically induced liquefaction (Nataraja and Gill, 1983). The action of waves pounding on beaches also seems plausible as a mechanism for forming fluidization features. For the fluidization features in the late

Pleistocene coastal marine terraces, though, the mechanism of wave-related liquefaction seems to be eliminated at some sites because dikes extend up into dune sands, where wave action seems extremely unlikely. Additionally, some dikes and sills cut lagoonal deposits at places that probably would have been protected from wave action. Significant artesian pressures at these lagoonal sites also are implausible. Thus a seismic origin seem probable for at least some of the features along the coast. Only in northern and central Oregon have fluidization sites been critiqued for a seismic liquefaction origin.

It is not possible to determine if the source of shaking was from rupture along the plate-boundary subduction zone or from other earthquake zones. Still, the fluidization features in the marine terraces provide independent support for strong seismic shaking near the coast during late Pleistocene time.

DISCUSSION

Liquefaction features in this study have been examined in five geologic settings: Columbia River islands; valleys of smaller rivers within 60 km of the coast; valleys of smaller rivers further than 60 km from the coast; coastal marine terraces; and the valley of Sandy River. These settings are very diverse with respect to liquefaction susceptibility, amplification of shaking from the energy source zone of subduction earthquakes, and the strength of the accelerations from the energy source zone. Altogether, though, they present pieces of evidence that to some extent complement one another concerning the nature of the shaking. Below I summarize observations for each of the five field settings, and deductions about shaking for each.

1. Columbia River islands

Physical setting, liquefaction susceptibility, and liquefaction observations.

Liquefaction-induced dikes were found to extend inland at least 90 km from the coast (Fig. 3). Virtually all dikes nearest the coast probably formed during the subduction earthquake of ca. 300 ybp. Even 90 km inland, the dikes are probably from that earthquake. The island sites far inland, Wallace Island and Deer Island, typically have fewer and smaller dikes than nearer the coast. Both islands are underlain by thick alluvium that lies directly on crystalline bedrock. This simple field setting facilitates first-order estimates of amplification of ground motions in the alluvium. Field conditions at Wallace Island and Deer Island probably are conducive to moderate to high amplification of bedrock accelerations in alluvium, yet basin-edge effects in this narrow valley may further amplify the amplitude and duration of strong motions. Alluvium at islands downstream may possibly be underlain by weak Tertiary deposits. Therefore, downstream from Wallace Island, there is diminished confidence in calculations of motions in alluvium.

The Columbia River islands are typically underlain by thick, cohesionless, fine sand and silty fine sand. A thin cap of silt and clay lies over the sand, and the water table extends nearly to the ground surface. The sands generally are moderately dense in the engineering sense. Therefore, these sand bodies should be prone to formation of many large liquefaction-induced effects during strong shaking from a nearby subduction zone earthquake.

Deductions about shaking

Generally only very minor effects of liquefaction were observed at Wallace Island, and especially so at Deer Island. The back-calculated accelerations in alluvium at these two islands were quite low, which increases confidence in the linear model (SHAKE) that was used to estimate the amplification of bedrock shaking in the alluvium. At Wallace and Deer Islands (65 and 90 km from the coast, respectively), the best estimate of the backcalculated peak bedrock accelerations is 0.05 to 0.07 g (Table 2). These values are very near the acceleration for the best-fitting estimates of a M~8 earthquake, based on theoretical and empirical statistical model analyses of plate-boundary subduction earthquakes. The model analyses shows only a modestly higher acceleration for a M 8.5 earthquake. Therefore, the liquefaction data in conjunction with the model data strongly suggest that very strong bedrock shaking (> 1/4 g) did not extend far inland from the coast, irrespective of magnitude.

2. <u>Smaller river valleys--within 60 km of the coast</u>

Physical setting, liquefaction susceptibility, and liquefaction observations.

Valleys of the Quinault, Humptulips, Chehalis, Naselle, and Grays Rivers were searched within 60 km of the coast (Fig. 3). Alluvium generally has a thickness of at least some tens of meters in these valleys, though locally the thickness ranges from a few meters to more than 30 m. Within 60 km of the coast, alluvium in virtually all the valleys is probably underlain by great thicknesses, (> 1,000 to 2,000 m) of very weakly lithified sediments. Such sediments plausibly dampen accelerations beneath the alluvium. Reducing peak accelerations by a factor of 1/2 seems possible, though improbable. Very probably, bedrock accelerations from great depth were not greatly enhanced at many places in the alluvium near the surface.

The valleys are filled mainly with permeable sandy gravel, probably only moderately susceptibility to liquefaction at scattered sites. Highly susceptible sediment appears rare. Sediments much exceeding the age of the earthquakes of ca. 300 ybp and ca. 1,400 to 1,900 ybp crop out at many places. Only one small dike of probable liquefaction origin was discovered within 50 km of the coast.

Deduction about shaking

Shaking during the subduction earthquakes of ca. 300 ybp and ca. 1,400 to 1,900 ybp rarely exceeded the threshold for creating liquefaction features within 60 km of the coast. The high permeability of the alluvium and the possibility of an exceptionally long vibration period (> 1 to 2 seconds) probably played only secondary roles in the lack of liquefaction effects. Sandy gravels having the same grain size distributions as those commonly found in the smaller river valleys vented and flowed extensively at an acceleration of about 0.3 g during the 1983 earthquake (M 7.3) of Borah Peak, Idaho, and in a prehistoric M~7.5 earthquake in the Wabash River valley of Indiana-Illinois.

The accelerations required to cause liquefaction in the sandy gravels of the smaller rivers cannot be estimated with confidence because of the lack of sufficient engineering data on states of compactness. (Such engineering measurements could be made.) Still, the lack of liquefaction effects in sandy gravels, some only moderately dense, strongly suggests that the accelerations were not extremely high near the coast.

This lack of liquefaction effects indicates strongly that during the past few thousand years, the seismic source zone of any great (M≥8) plate boundary subduction earthquake must have been located tens of kilometers offshore.

3. <u>Smaller river valleys--further than 60 km from the coast</u>

Physical setting, liquefaction susceptibility, and liquefaction observations.

The region of main interest is in the Chehalis River valley near Oakville-Porter (Fig. 3), where dikes and sills almost certainly originated by liquefaction of sandy gravel. Crystalline bedrock here comes much nearer the surface than closer to the coast, which makes dampening of shaking from depth improbable. Accelerations from depth may be amplified slightly in alluvium. The alluvial sediments are almost everywhere permeable sandy gravel, and the liquefaction susceptibility appears to be essentially the same as elsewhere in smaller river valleys, including nearer the coast, on the basis of grain sizes, mode of deposition, and capping material.

Several sites with liquefaction features were found in this region. Their ages are not constrained, but many or all could have formed about 1,500 - 2,000 years ago.

Deduction about shaking.

There are no engineering data on the probable sandy gravel source beds at the liquefaction sites. Except for large dikes at one site of slumping, all dikes are small. However, very strong shaking may have been required to have liquefied the sandy gravels.

It is not known if the dikes formed during plate-boundary subduction zone earthquakes, but a local source region seems very probable.

4. <u>Coastal marine terraces</u>

Physical setting, liquefaction susceptibility, and liquefaction observations

Marine terraces of late Pleistocene age border the ocean throughout Oregon and Washington. Wave-cut exposures along the ocean show the terraces generally are made of clean sand that is many meters in thickness, and locally capped by a very thin peaty zone. Large dikes and sills, many of probable earthquake liquefaction origin, can be seen in these outcrops.

Deductions about shaking

Very strong surface shaking (>> 1/4 g) from earthquakes does not seem inconsistent with the observed effects of liquefaction. Whether plate-boundary subduction earthquakes were involved cannot be inferred.

5. <u>Valley of Sandy River</u>

Physical setting, liquefaction susceptibility, and liquefaction observations

The lowermost part of the valley of Sandy River, located near Portland about 130 km from the coast, is underlain by clean, fine- to medium-grained sand that is at least several meters in thickness and possibly much thicker. The sands typically are capped by a friable silty sand. Data for deposits beneath alluvium are insufficient to make a good estimate of the likely amplification of bedrock shaking in alluvium, but the regional setting does not suggest more than moderate amplification.

Effects of minor fluidization are nearly ubiquitous along the contact of the friable silty sand with the underlying clean sand. Sill-like features abound in the region. Only scattered small dikes occur. Radiocarbon data and chemical analysis of tephra show that the deformed beds could be as old as the subduction earthquake of ca. 300 ybp.

Engineering data on the source sand beds show that a peak acceleration of about 0.1 g would have sufficed to have caused venting, during a great subduction earthquake near the coast. No venting was observed, so the estimate of 0.1 g is probably slightly high.

Deduction about shaking

The pervasive lateral effects of fluidization along the contact of the cap with the source sand, in combination with the abundance of very small sill-like features and paucity of dikes, seems consistent with a prolonged duration of weak seismic shaking. If the shaking had been strong, causing strongly developed liquefaction, dikes should have formed much more abundantly. Instead, a prolonged duration of shaking with minor liquefaction could have caused only extensive lateral flowage. Thus the liquefaction effects seem consistent with weak shaking for a particularly long time, such as could occur from an offshore subduction earthquake of $M \geq 8$.

CONCLUSIONS

Consideration of the liquefaction-induced features in the five widely scattered geologic settings discussed above leads to the following conclusions.

1. Occurrence of a M~8 plate-boundary subduction earthquake ca. 300 ybp seems consistent with the liquefaction data, providing the energy source (locked) zone was located offshore some tens of kilometers.
2. Very strong onshore bedrock shaking (> 1/4 g) from plate-boundary subduction earthquakes probably did not extend onshore more than a few tens of kilometers in coastal Washington during the past few thousand years.
3. The model of a plate-boundary subduction earthquake with a prolonged duration of shaking, and particularly rich in long period shaking, appears to be consistent with the field observations for the earthquake of ca. 300 ybp.
4. Plate-boundary subduction earthquakes stronger than the event of ca. 300 ybp possibly have struck offshore during the past few thousand years and formed the dikes far inland (> 60 km) along the Chehalis valley. However, this scenario seems unlikely, because of the lack of liquefaction-induced features nearer the coast.
5. A seismic source zone capable of liquefying sandy gravel is likely located near the Chehalis River valley, about 60-80 km inland from the coast.

These conclusions are based largely on preliminary geotechnical engineering analysis of liquefaction effects that were observed in the study area, in conjunction with consideration of the regional pattern of liquefaction effects. The engineering analysis has been done so as to yield the most reasonable to conservative (high) estimates of accelerations. These estimates have been compared with the best estimates of accelerations for a M 8 earthquake, from both theoretical and statistical empirical models. Both models predict nearly the same peak accelerations at distances greater than 30 km from the coast. Considering that the engineering analysis of liquefaction effects yields results comparable with the best-fit solutions for these models, a value of M~8 seems a reasonable estimate for the earthquake of ca. 300 ybp. If the magnitude much exceeded a value of 8, and the locked zone came within 35 km of the coast, the accelerations were probably extraordinarily low for such an earthquake.

SUGGESTIONS FOR FUTURE RESEARCH

All the regions that I searched for liquefaction-induced features present uncertainties in interpretations either because of the presence of sediments of uncertain liquefaction susceptibility (especially the sandy gravels), or because of unknown amplification of bedrock motions (especially in the Columbia River valley). These uncertainties could probably be largely resolved with further studies. In addition, further research could probably shed light on the liklihood of whether the previous subduction earthquakes had extraordinarily low peak accelerations, accompanied by an extraordinary amount of the energy in long period vibrations.

Amplification of motions in the narrow Columbia River valley may be caused by motions reverberating from wall to wall, thereby enhancing both amplitude and duration of strong shaking. Positioning seismographs in the valley to monitor effects of small earthquakes could provide data to help resolve the possibility of such an effect. In addition, two-dimensional modeling of motions might be very useful.

There are regions along the coast where problems posed by unknown bedrock amplification and by low liquefaction susceptibility are minimized. One such region is the long spit (Long Beach) between Willapa Bay and the Pacific Ocean. The spit is probably an accumulation of sand that lies on relatively shallow bedrock over a large region. I examined several hundred meters of a freshly excavated ditch in the central part of the spit, in 1987. Ground conditions (clean sand, shallow ground water table) looked very favorable for formation of liquefaction-induced features, yet I found none. Humate development in the sand indicated an age much exceeding the earthquake of 300 ybp. I found this absence of liquefaction effects to be very surprising, because in a similar physical setting in the meisoseismal region of the 1886 earthquake (M 7.5) of Charleston, South Carolina, almost certainly I would have found liquefaction effects of the 1886 earthquake. I suggest that an extensive trenching operation to search for liquefaction-induced features on the spit of Long Beach might be very informative. A similar region along the coast appears to be located on the coast of Oregon, just south of the Columbia River.

One of the better ways to bracket the magnitude and strength of shaking of a prehistoric earthquake is to determine the regional extent of liquefaction effects. During the 1964 Alaskan earthquake (M 9.2), liquefaction effects extended 420 km from the energy source, and lateral spread movements of as much as 36 cm extended to 210 km (Youd and Perkins, 1987). For the Cascadia region, the Willamette River valley in Oregon is well positioned to have recorded the effects of great subduction earthquakes, because the valley is located only about 75 km inland from the coast. I have begun a search for evidence of liquefaction in the valley, and have discovered dikes of almost certain seismic liquefaction origin in banks of the Calapooia River, near Corvallis, Oregon. More than 25 dikes have been found at six widespread sites in the 20 km of river searched from a canoe. I suspect that all the dikes originated from a single, ancient earthquake that occurred at least several thousand years ago, as evidenced by severity of weathering of some of the dikes and the same apparent sequence of ancient (early Holocene?) sediments cut by the dikes. More field work is needed to determine the regional extent of the dikes in the Willamette River valley, and to determine their age(s) of formation.

More work is needed to bracket the strength of shaking on the Columbia River islands. It is important to place upper limits on the strength of shaking because of the possible danger to structures such as the tall buildings in Portland and Seattle.

APPENDIX

Note 1. Islands searched in which there was at least 0.5 km of cleanly exposed outcrop in vertical section are listed below in the order of going upstream: Karlson, Marsh, Brush, Horseshoe, Woody, Tenasillahe, Price, Hunting, Wallace, Deer, Sauvie, Bachelor, Government, and Reed. Sand deposits, having grain sizes in the range most highly susceptible to liquefaction, were observed at low tide to underlie all these islands except Sauvie and Bachelor Islands. At Government and Reed Islands, the susceptible deposits lie above the level of low tide. Ages of susceptible sediments examined on all islands exceed that of the downdropping event of ca. 300 ybp.

Note 2. The field penetrometer consisted of a 20 pound weight that was dropped 18 inches onto a 5/8 inch hexagonal steel rod. which had a 1 inch diameter tip. Apex angle of the tip was 60 degrees. Rods were 4 feet in length. The penetrometer was advanced in a continuous manner, and the measured penetration resistances were the number of blows required to advance the tip 6 inches. These blow counts were later converted to equivalent SPT blow count values.

Rods were rotated during testing to minimize skin friction. In sand beds, the rods were easily rotated by hand, suggesting that skin friction was negligible.

Note 3. Only limited research has been done to quantify the susceptibility of gravelly sand and sandy gravel to liquefy and then flow to form fluidization features (such as dikes). Laboratory shake-table data suggest that for gravel-sand mixtures, with the same initial state of packing (a relative density equivalent to moderate packing, in the engineering sense), increasing the gravel content from 0 to 70 percent increases the shaking (shear stress) threshold for (initial) liquefaction by a factor about 50 percent (see Ishihara, 1985, p. 334-5). In another study, Evans and Seed (1987) compared the liquefaction susceptibility (in terms of development of large pore-pressure) of well-graded (poorly-sorted) gravel, uniformly-graded (well-sorted) gravel, and sand samples in dynamic triaxial tests. The samples were compacted to the same initial relative density and then subjected to the same loading conditions. They found that the liquefaction susceptibilities were approximately equal, in contrast to results of Ishihara. Using the approach of measuring the shear strain at which pore pressures increase, Hynes (1988) found that a significantly larger threshold was required for gravels than for sands. This implies that significantly higher accelerations would be required to liquefy gravels.

Considering all sources of data, it seems highly probable that gravels with less than 30 to 40 percent sand are less likely than sand deposits, of the same relative density, to experience the large buildup of pore pressure required for liquefaction. Still, a gravel-rich deposit can be quite susceptible to development of large pore pressures, providing the deposit is confined sufficiently to prevent rapid dissipation of pore pressures during cycles of shearing.

Unquestionably a very important factor in the field controlling development of dikes is the presence of a cap of low permeability above the liquefied sandy gravel. Such a cap helps prevent extremely rapid dissipation of excess pore pressure between cycles of shearing, and permits formation of large fluidization features.

Note 4. No seismographs were near the epicentral region of the Borah Peak earthquake, but peak accelerations at the liquefaction site with coarsest gravel and least sand are estimated to have been 0.3 to 0.35 g on the basis of seismological model data (Andrus et al., 1991, p. 257; Andrus, 1994). Using Modified Mercalli intensity data in the epicentral region and aftershock data, Harder (1988) estimated the peak acceleration to be 0.29 g.

ACKNOWLEDGMENT

Appreciation is given to Curt Peterson (Portland State Univ.), who assisted greatly with field interpretations during the summer of 1992. Important contributions were provided by the following persons: David Tabaczynski, Boyd Benson, Colby Menton, John Schulene and Brian Atwater (USGS); Steve Palmer (Washington State DNR); Lynn Moses (Washington State DOT); Steve Dickenson and Tim Roberts (Oregon State Univ.), and Jim Phipps (Grays Harbor College). Thoughtful reviews of this manuscript were provided by Tom Holzer and Al Nelson (USGS), Steve Dickenson (Oregon State Univ.), and R.R. Youngs (Geomatrix Consultants). An exceptionally helpful and thorough technical review was provided by Eric Pond (Virginia Tech); Mr. Pond is currently completing his Ph. D. dissertation, which is an engineering evaluation of prehistoric strength of shaking in the Wabash Valley of Indiana-Illinois by use of liquefaction-induced effects.

Funding was provided mainly by the U S Nuclear Regulatory Commission, with support from the NEHRP Program. Oregon State University provided support for the field penetration testing conducted on the Columbia River islands. The Oregon Department of Transportation permitted our use of their field penetrometer.

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Figure 1. Map showing geographic region of Cascadia subduction zone. Box shows area searched for relict liquefaction features.

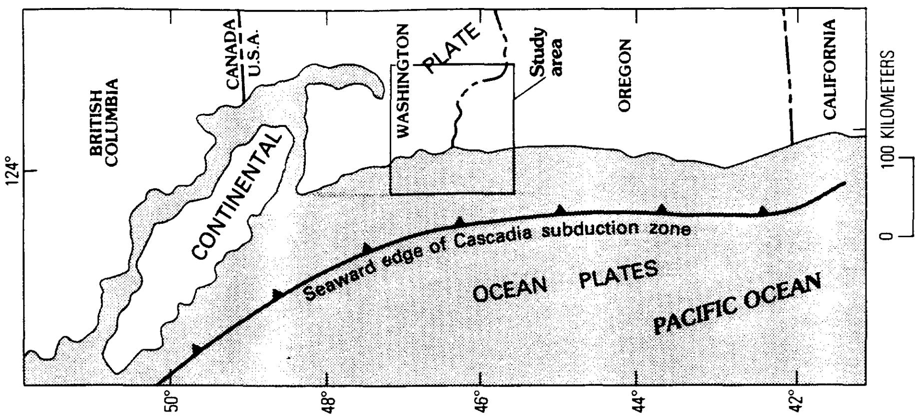


Figure 2. Schematic sectional view throught the Cascadia subduction zone at the latitude of Portland, Oregon. (From Wong et al., 1993.)

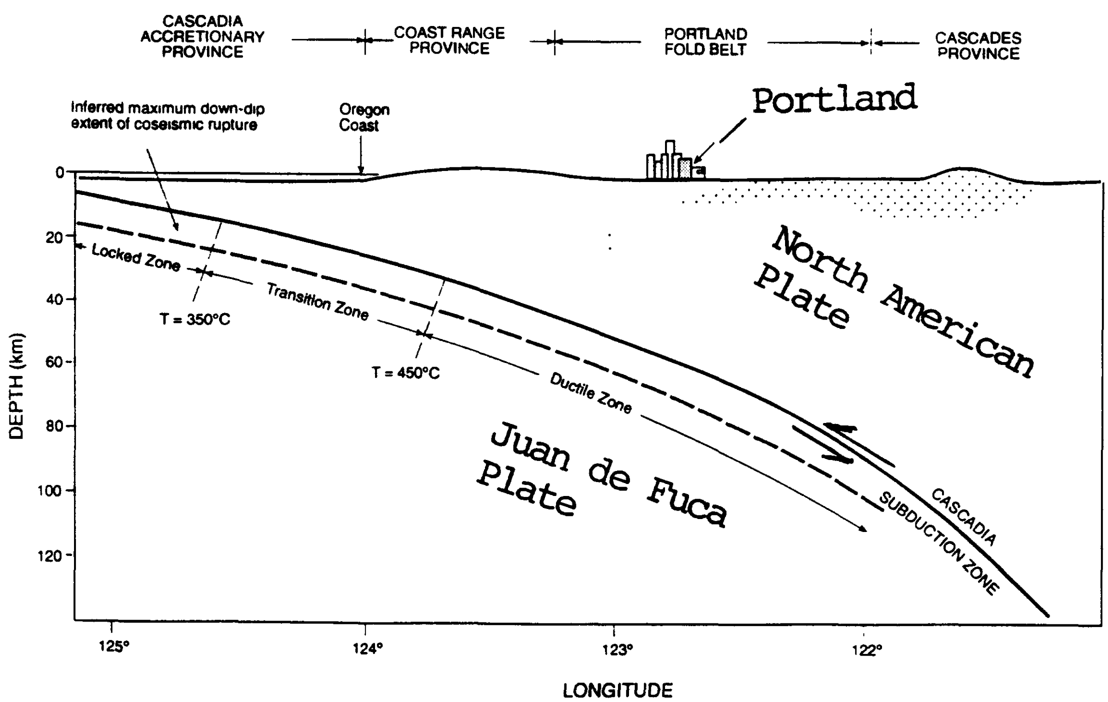


Figure 3. Map showing that part of the Columbia River where banks of islands were searched for paleoliquefaction features. These searched islands have ages between 600 and 1,000 years at most places. Sands beneath islands are fine to medium grained and generally are at least moderately susceptible to liquefaction. Maximum dike width is measured at least 1 m above the base of the dike.

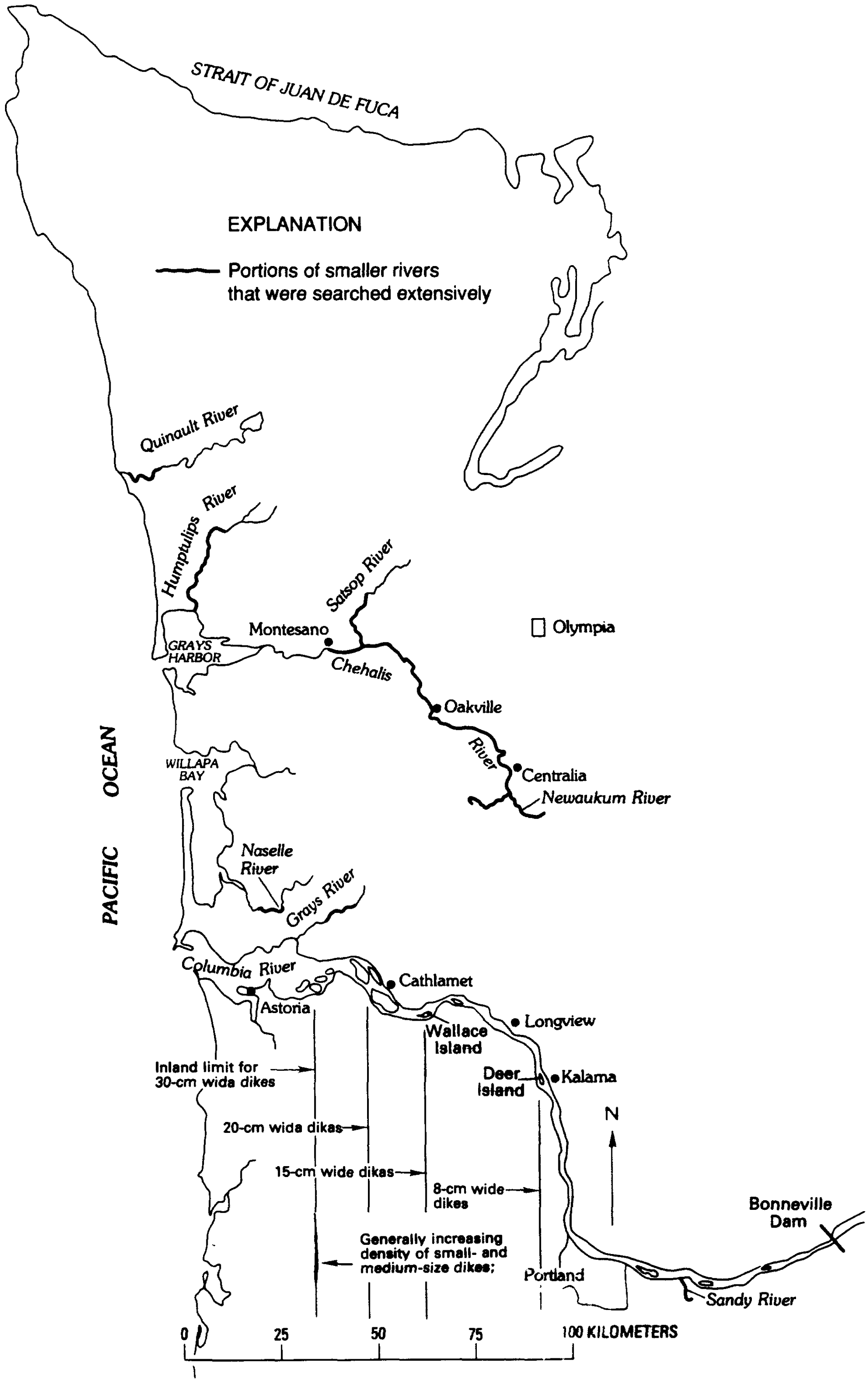


Figure 4. Block diagram showing typical field relations at liquefaction sites in Columbia River islands. A sand-filled dike cuts through a 1- to 2-m thickness of soft silt and clay with a weakly developed soil at the top. The dike connects to a thin sand sheet on the soil horizon that is buried by a 1-m thickness of silt and clay. Tubers at widespread sites collected in their growth position near the base of the stratum cut by dikes have radiocarbon ages ranging between 600 and 1,000 years. Widths of top of dikes are typically only several millimeters.

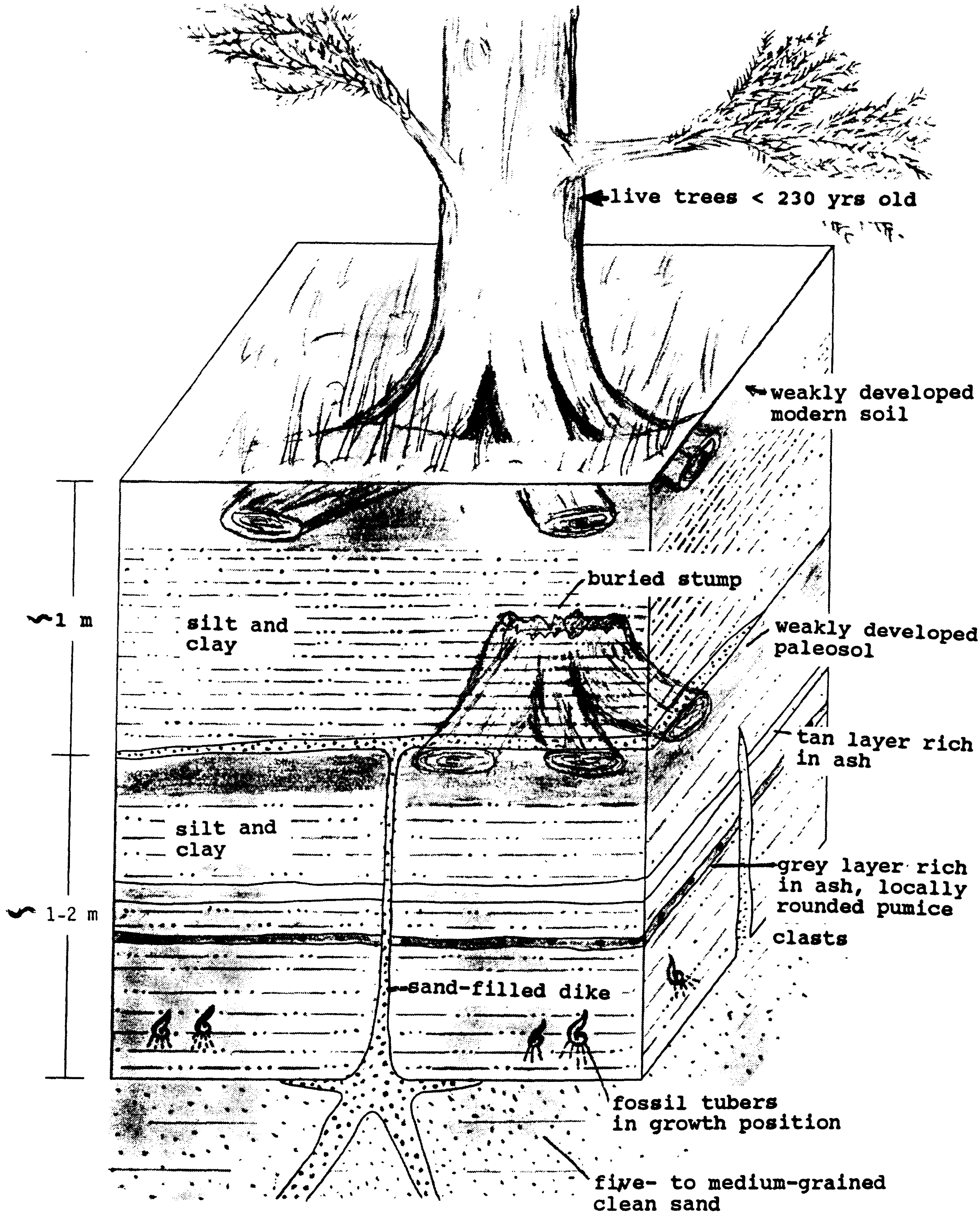
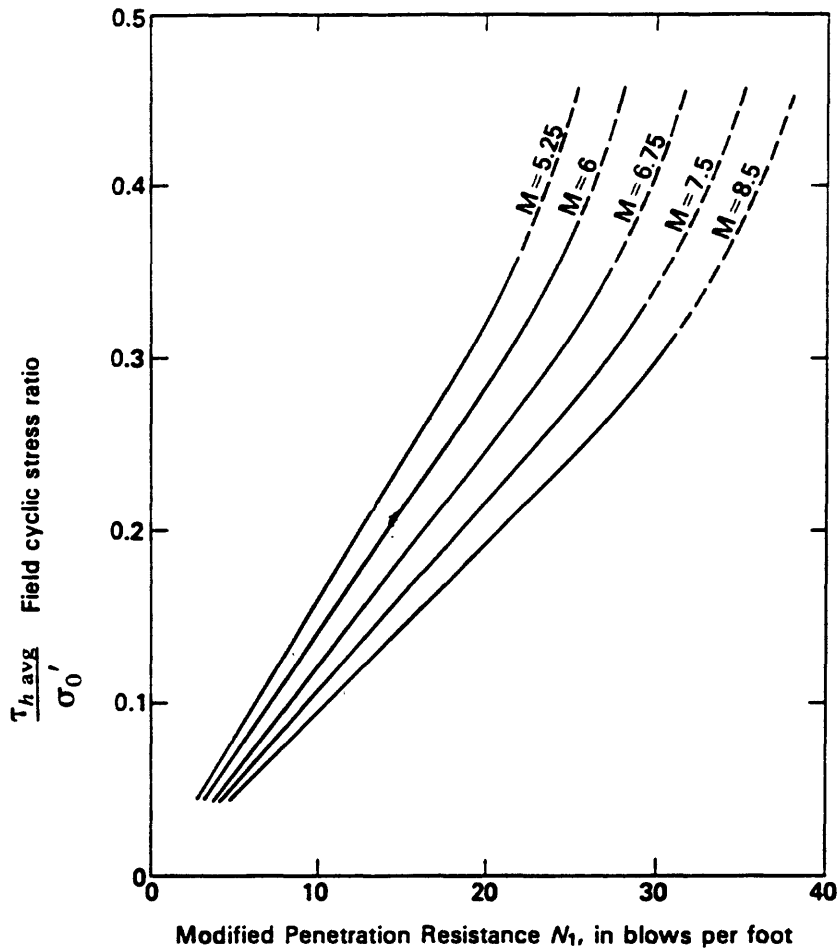


Figure 5. Curves for the method of Seed et al. (1983) used to evaluate the potential occurrence of liquefaction with accompanying venting of sand on appreciable ground cracks at a site on level ground. Curves are for clean sand deposits (average diameter >0.25 mm) and for different earthquake magnitudes (5.25 to 8.5). Points above and to the left of curves show conditions having high potential for liquefaction.



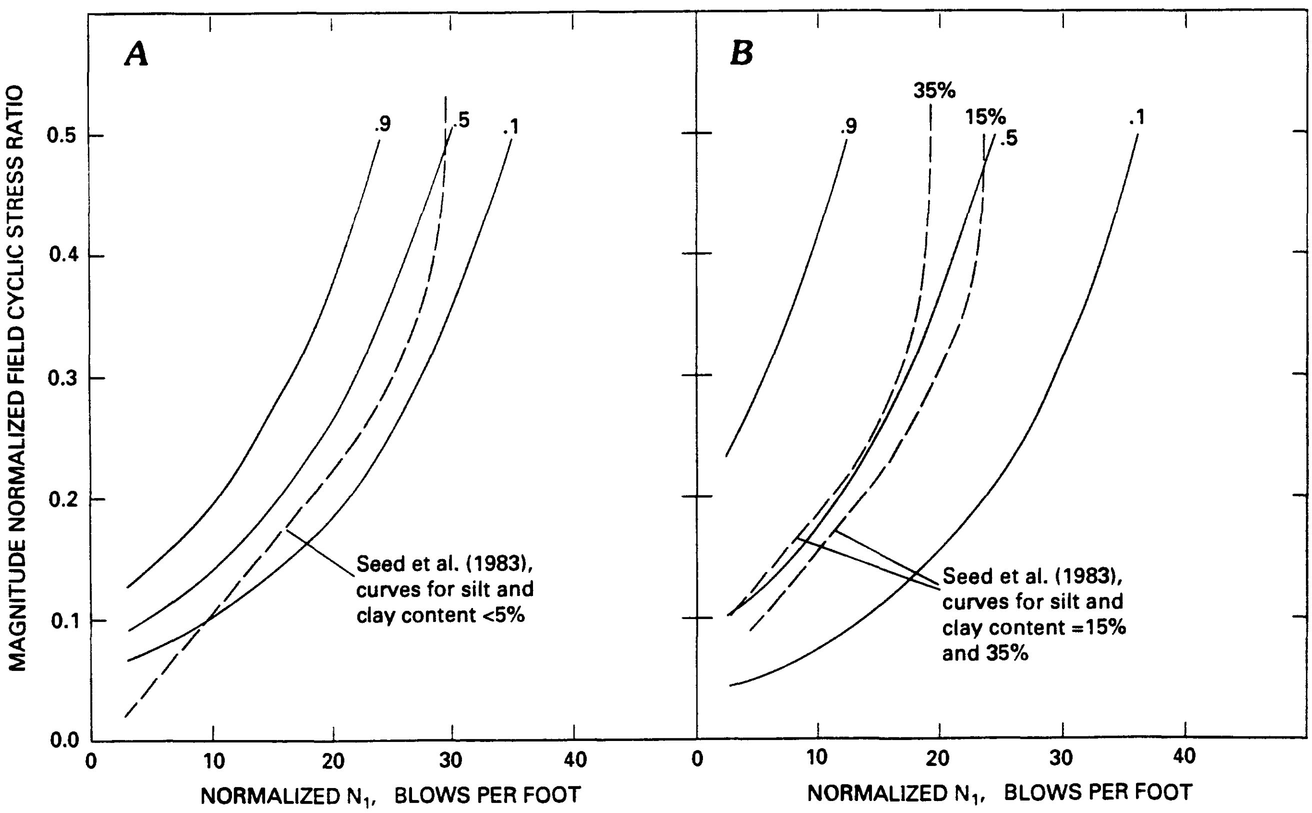
EXPLANATION

$\tau_{h\ avg}$ Average earthquake-induced horizontal cyclic shear stress

σ_0' Vertical effective stress

N_1 Standard Penetration Test blow count measured in field, modified to blow count resistance at vertical effective stress of 1 ton/ft^2

Figure 6. Probability of liquefaction with accompanying venting or appreciable ground cracking, normalized to blow count value, magnitude, and field cyclic stress ratio. A, curves for silt and clay content ≤5 percent. B, curves for silt and clay contents of 15 and 35 percent. (From Liao et al., 1988.)



Correction Factors for Magnitude Normalized Cyclic Stress Ratio (From Seed et al., 1985)

Earthquake magnitude	Correction factor
8-1/2	0.89
7-1/2	1.0
6-3/4	1.13
6	1.32
5-1/4	1.5

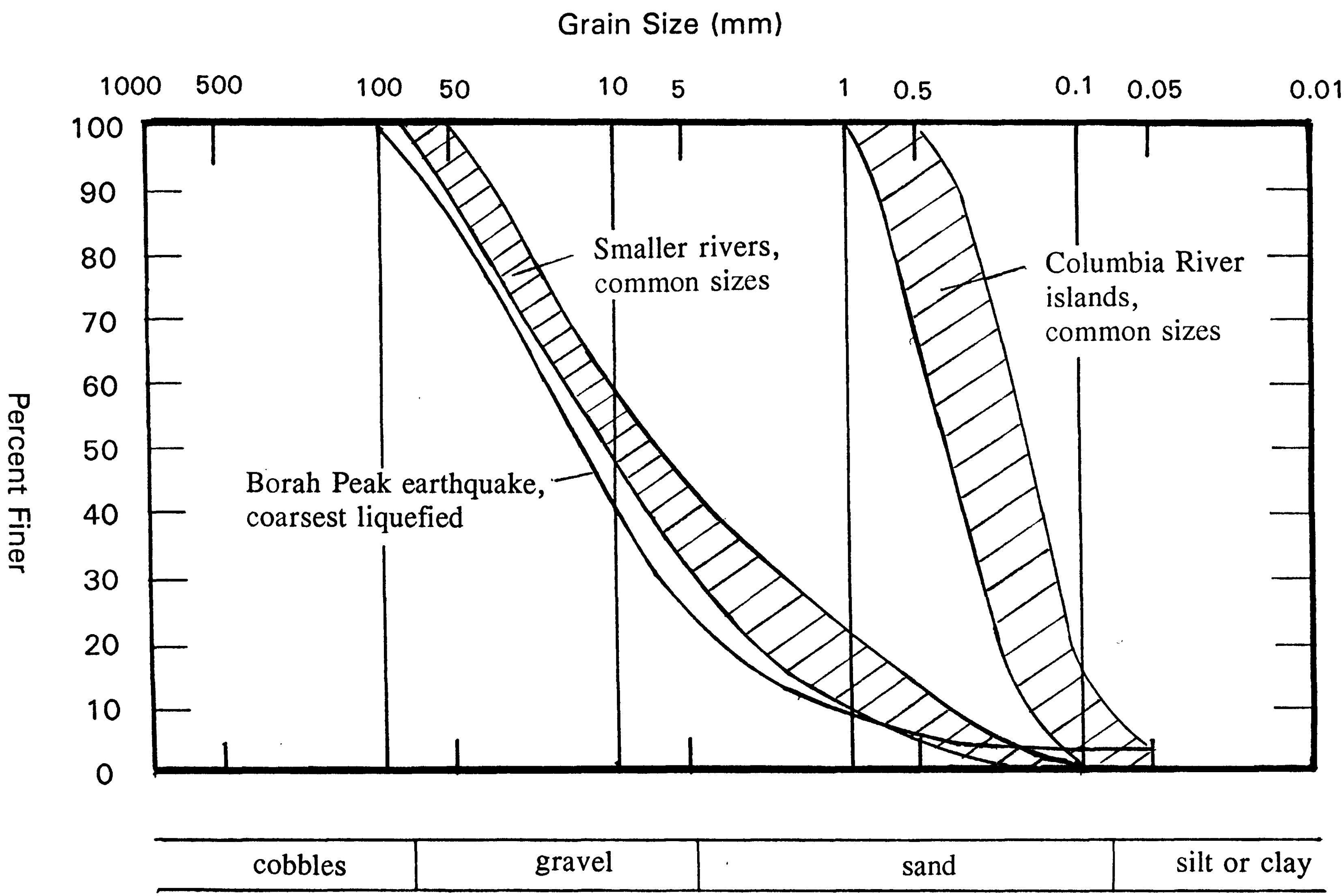


Figure 7. Grain-size curves for the following: (1) representative granular samples collected from sites that liquefied on Columbia River Islands; (2) representative granular samples collected along smaller rivers in Washington; and (3) samples from the coarsest stratum that liquefied extensively during the 1983 Borah Peak, Idaho, (M 7.3) earthquake. Borah Peak earthquake data from Andrus et al. (1991).

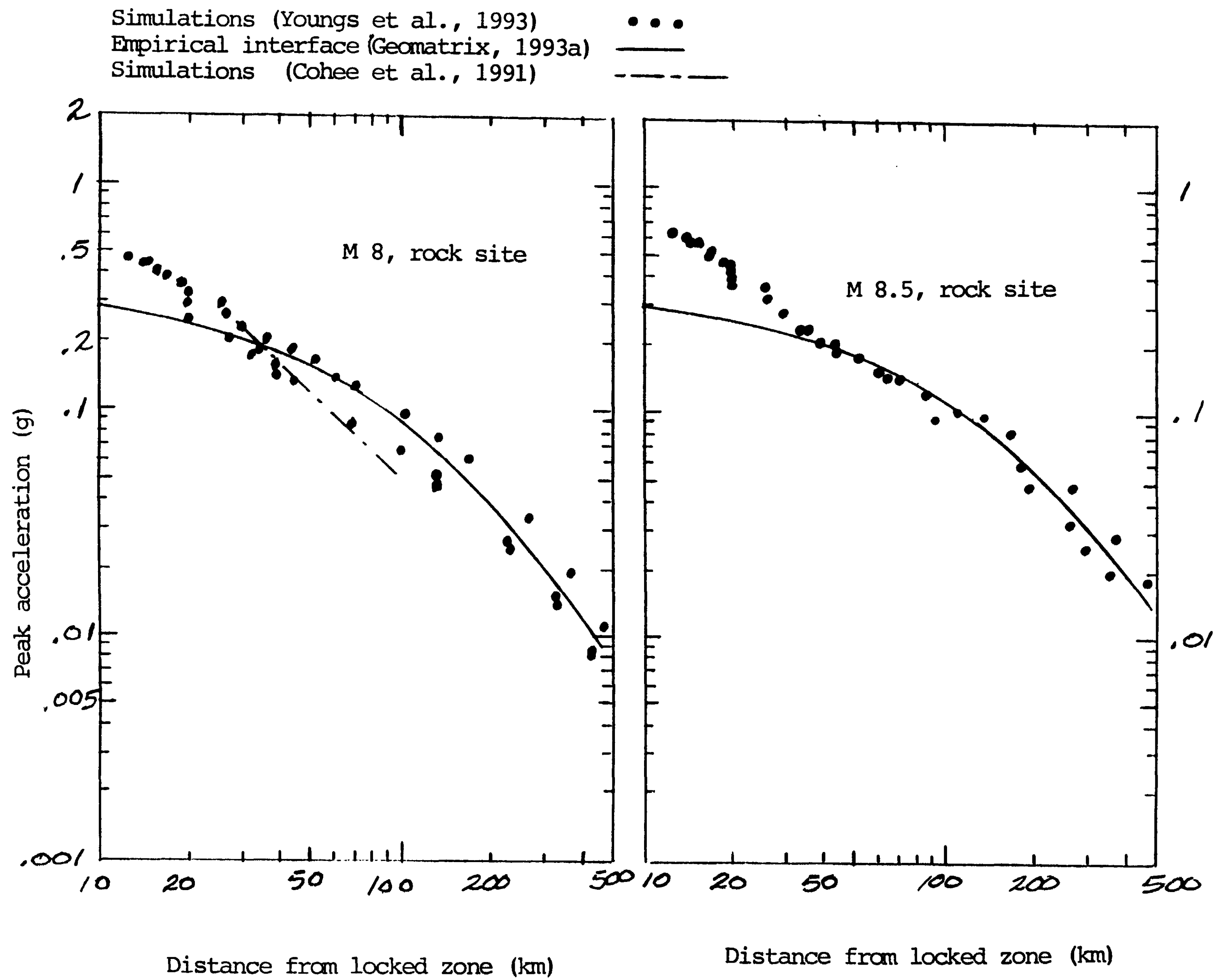


Figure 8. Model accelerations at rock sites for M 8 and 8.5 plate-boundary Cascadia subduction earthquakes. Simulations by Silva and Cohee et al. are theoretically based solutions. Empirical interface solution is a statistical analysis. (From Geomatrix, 1993a,b and R.R. Youngs, written com., 1994.)

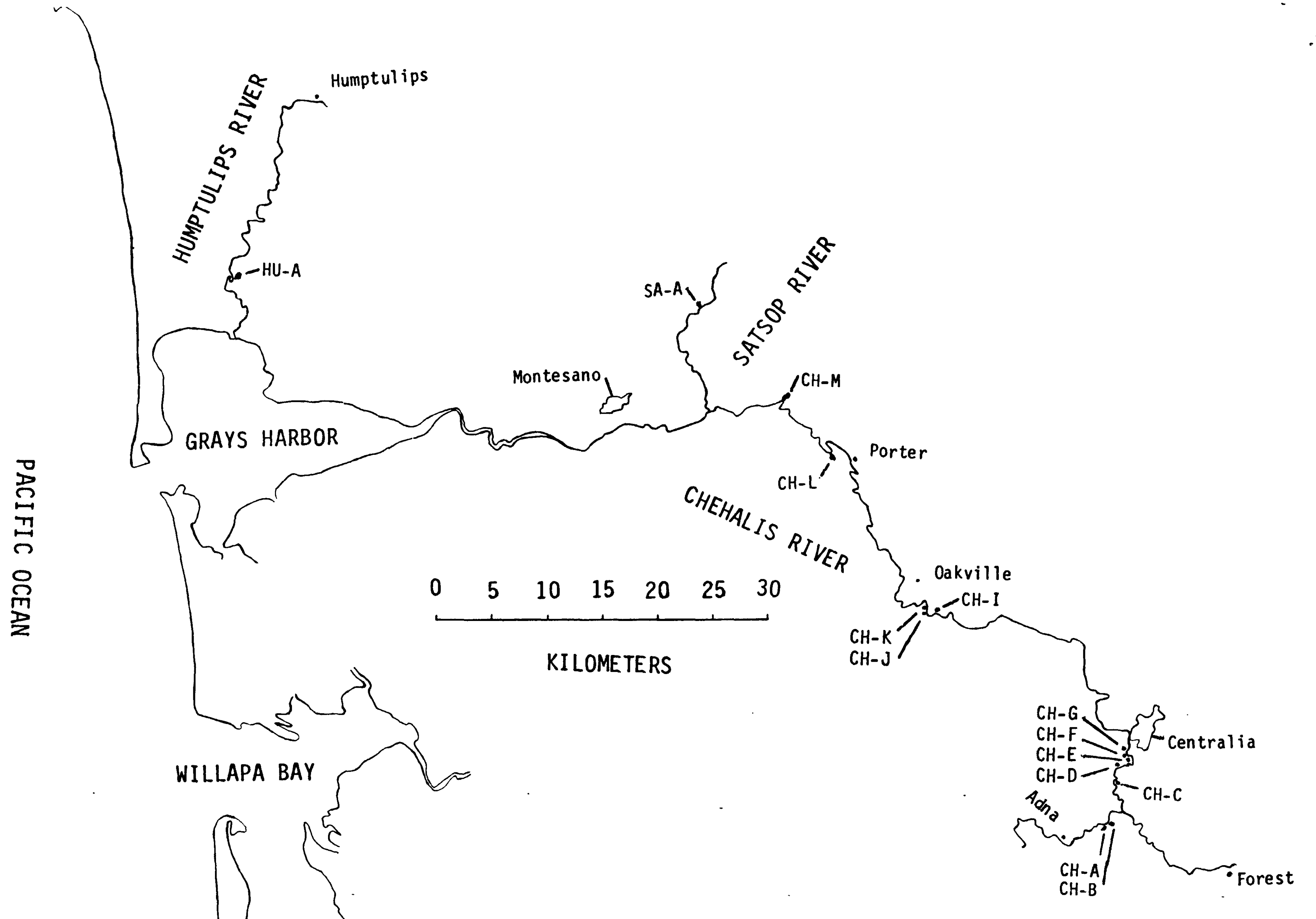


Figure 9. Locations of liquefaction sites discovered in banks of rivers in Washington. It is doubtful that liquefaction at any of the sites was caused by the subduction earthquake of ca. 300 ybp, with the exception of site HU-A.

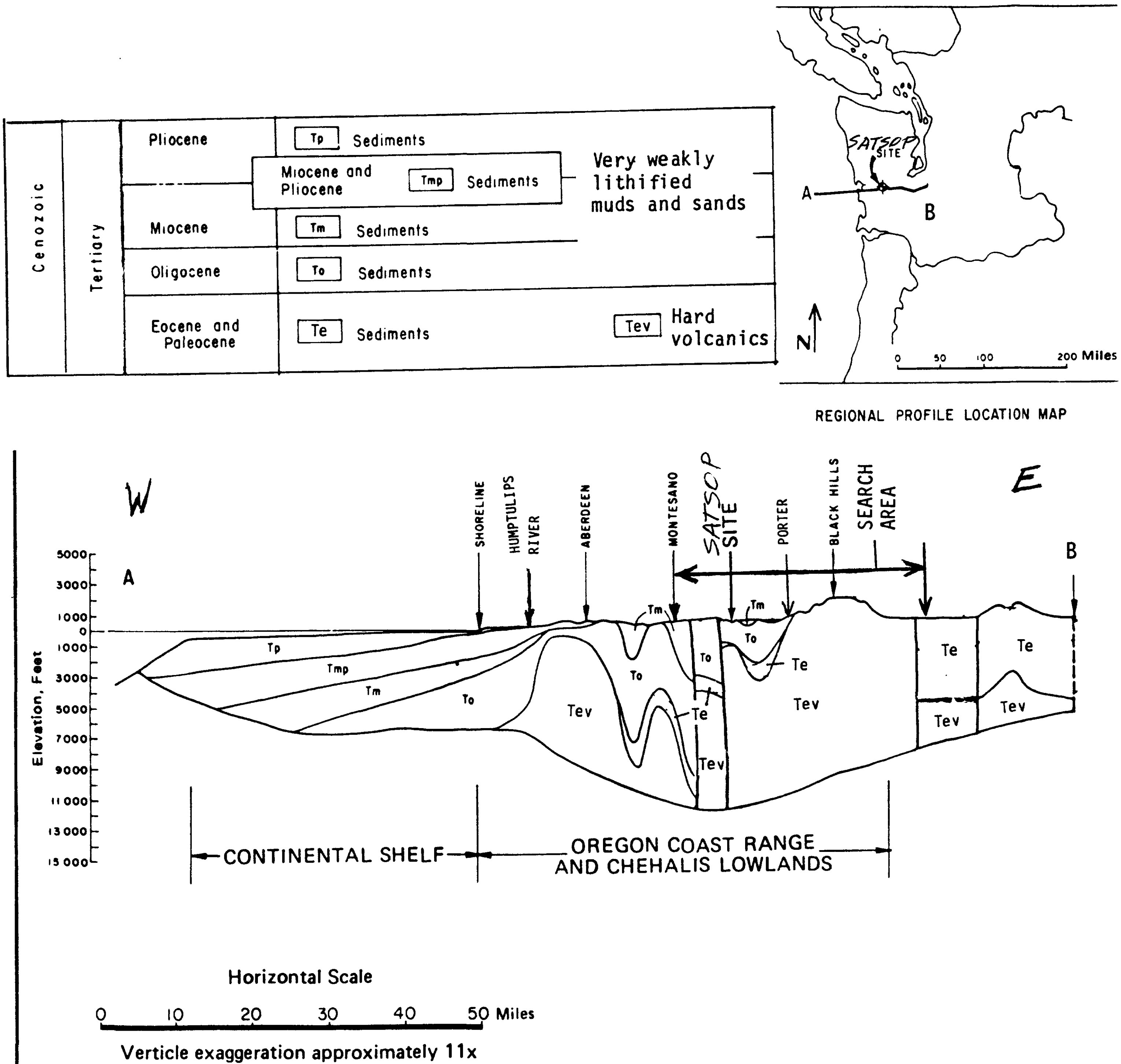


Figure 10. Schematic cross-section of geologic materials beneath alluvium, near that portion of the Chehalis River vally that was searched for liquefaction features. Hard rock lies near the surface upstream from the town of Porter. The region of the Humptulips River is underlain by very thick, weakly lithified deposits. (Modified from Fig. 2.5-10 in the report by Washington Public Power Supply Systems, 1983.)

Table 1. Widths of three largest (widest) dikes on groups of islands as a function of distance from the coast. Smallest outcrop area examined in plan view was at islands nearest the coast. Area of outcrop examined in sectional view was about the same at all groups.

Islands (number of largest dikes at each island)	Distance east from the coast (km)	Widths of three largest dikes (cm) (measurements within 1 m of the surface of venting
Marsh (2), Brush (1)	35	30,30,20
Hunting (1)*, Tenasillahe (2)	50	20,15-20,15
Wallace (3)	60	15,15,5
Deer (3)	90	8,5,3

*The widest dike at Hunting Island was probably a fracture to a vent, whereas everywhere else the widest dikes are probably openings from lateral spreading.

Table 2. Comparison of backcalculated acceleration with estimates of models, for a M 8 Cascadia subduction earthquake

Site	Distance from coast	Distance from locked zone	Backcalculated peak surface acceleration, from Seed procedure curves of Fig. 5.	Seed et al. (1983) value calibrated to 0.5 probability of Liao et al. (1988), in Fig. 6.	Backcalculated peak bedrock acceleration, best estimate (x 2.5 amplification factor)	Backcalculated peak bedrock acceleration, range (x 2 to 3 amplification factor)	Model value from Fig. 8, best estimate of peak bedrock acceleration for M 8 earthquake
Marsh Island	35 km	70 km	>0.2 to 0.25 g ?	>0.25 to .3 g	>0.10 to .12 g	>0.08 to 0.15 g	0.10 to 0.12 g
Hunting Island	50 km	85 km	>0.2 to 0.25 g ?	>0.25 to .3 g	>0.10 to .12 g	>0.08 to 0.15 g	0.09 to 0.11 g
Wallace Island	65 km	100 km	0.11 to 0.19 g	0.13 to 0.23 g	0.05 to 0.09 g	0.04 to 0.11 g	0.08 to 0.09 g
Deer Island	90 km	125 km	0.11 to 0.15 g	0.13 to 0.18 g	0.05 to 0.07 g	0.04 to 0.09 g	0.07 g
Sandy River	125 km	160 km	≤0.1 g (from Anderson et al., 1944)[1]	~.1 g ?	Likely >~0.05 g (< x 2 amplification factor)[2]	- -	0.06 g

1. The value of 0.1 g is likely a high estimate of acceleration. This estimate is based on Dutch cone data at two sites, which then were used with the Seed et al. procedure.

2. Alluvium probably has only moderate thickness in this region, which would not be conducive to very high amplification of bedrock shaking.

Table 3. Age interpretations for liquefaction at sites within 70 km of the coast

Site	Radiocarbon age and relation to field situation	Preferred age interpretation for dikes, on basis of radiocarbon data and weathering profiles	Possible range of ages for dikes
CH-I	charcoal fragment dated at 1820±60 yrs; fragment from stratum above top of dike. Top of dike is suspected to be eroded off below charcoal fragment	slightly >1,820 yrs.	could be <1,820 yrs, if dike not truncated
CH-J	charcoal fragments dated at 3,100±180 yrs; fragments from contact of cap with source stratum of sand and gravel	significantly <3,100 yrs.	could be <<3,100 yrs.
CH-L	many hemlock leaves in pristine condition dated at 3,350±60 yrs; leaves from contact of cap with source stratum of sand and gravel; charcoal dated at 4,050±90 was taken from soil horizon onto which sand probably vented	significantly <3,350 yrs.	could be slightly to significantly <3,350 yrs; strongly developed weathering profile on overbank silt and clay above vented sediment indicates an age greater than downdropping event of ca. 300 ybp.
HU-A	charcoal dated at 2,530±60 yrs; charcoal from layer cut by dike, located slightly above contact of cap and source bed of sand and gravel	moderately <2,530 yrs.	could be much <<2,530 yrs.