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THESIS APPROVAL

The abstract and thesis of Robert B. Schlichting for the Master of Science in Geology presented May 18, 2000, and accepted by the thesis committee and the department.

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ABSTRACT

An abstract of the thesis of Robert B. Schlichting for the Master of Science in Geology presented May 18, 2000.

Title: Establishing the Inundation Distance and Overtopping Height of Paleotsunami from the Late-Holocene Geologic Record at Open-Coastal Wetland Sites, Central Cascadia Margin.

Mapping and stratigraphic investigations of back barrier, open-coastal plain sites have been used to establish minimum inundation distances and wave heights of tsunami produced by great subduction zone earthquakes in the central Cascadia margin. Cascadia tsunami deposits have been reported for many coseismic subsidence events in bay marsh settings where tidal-channel features focus tsunami energy. Variable magnitude (8.5 ± 0.5 Mw), frequency (500 ± 300 yr recurrence), and rupture geometry produce widely varying computer model outcomes for Cascadia tsunami inundation. The results presented in this thesis provide specific quantitative data regarding tsunami inundation at the open coast.

Anomalous sand sheets that have been characterized consist of well-sorted beach sand that fine up-section. The thickness of the deposits vary from 45 cm to 0.2 cm, and thin in the landward direction. Many of the sand layers include detrital caps. One to three detritus and mud lamina are intra-layered in the deposits. Marine diatoms and bromine, a marine tracer, increase in concentration at each of the sand layers. C14

ESTABLISHING THE INUNDATION DISTANCE AND OVERTOPPING HEIGHT
OF PALEOTSUNAMI FROM THE LATE-HOLOCENE GEOLOGIC RECORD AT
OPEN-COASTAL WETLAND SITES, CENTRAL CASCADIA MARGIN.

by

ROBERT B. SCHLICHTING

A thesis submitted in partial fulfillment
of the requirements for the degree of

MASTER OF SCIENCE
in
GEOLOGY

Portland State University

2000

dating of select layers suggest correlation to great Cascadia, earthquake-induced, coseismic subsidence events. The sand layers are inferred to be tsunami deposited, synchronous with coseismic subsidence events. Core traverses record four, and in some cases up to seven Cascadia tsunami events. One sand layer found at many sites is dated at 600-950 cal YBP, an age that lacks a recognized coseismic horizon in the northern Cascadia margin.

Soil chronosequencing and dated subsurface erosional scarps have been used to constrain paleo-dune elevation and shoreline position. Minimum paleo-dune height ranges from 5 to 8 m. Minimum overland inundation distance at the various sites ranges from 0.3 to 1.3 km $\pm$.2 km. The tsunami origins of landward-most sand laminae are tested by marine diatom and bromine anomalies. These tracers are being used to extend paleotsunami flooding records to include low-velocity inundation distances.

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Introduction

Shorelines bordering the Pacific Ocean basin are subject to tsunami (Figure 1). Convergent tectonic boundaries, which characterize the land-water margin throughout the Pacific Basin, have historically induced tsunami in Papua, New Guinea (1996), Mindoro, Philippines (1994), the Japan Sea (1983), Alaska (1964) and Chile (1960). Whereas major damage from tsunami flooding is usually localized to shorelines in the vicinity of earthquake foci, tsunami can also have far-field effects. The 1964 Alaska earthquake, which had a magnitude moment of 9.2 (M_w 9.2), resulted in both locally devastating tsunami inundation in Alaska (Plafker *et al.*, 1969),

Figure 1. Pacific Ocean Basin showing the locations of recent destructive tsunami.

and caused significant far-field damage in Seaside, Oregon (Fiedorowicz, 1997), and Crescent City, California (Lander *et al.*, 1993).

The turbulent nature of tsunami inundation (Minoura *et al.*, 1994) induces sediment transport and can leave geologic evidence of its occurrence. Tsunami strata can be discerned where an abundant supply of sediment and a contrasting host environment exists. The sedimentary record left by tsunami inundation typically consists of well-sorted, fining-up sequences of sand that is interstratified with the mud and peat that are typical of marsh and lake settings (Foster, 1991). Intercalated and capping layers of mud and/or detritus are common in tsunami deposited sand.

It is important to note that tsunami transport is not the only way that sand can be deposited into coastal wetland and lacustrine settings. Landslides and terrestrial streams can be a source of sand from landward sources. These sources are easily distinguishable in a deposit from marine – e.g. tsunami, due to distinct differences in sand grain size, mineralogy, and texture. Liquefaction and loading processes can induce sand fountains from underlying strata, leaving anomalous sand layers in overlying wetland deposits. These types of deposits are assumed to be patchy. Hence, a sand deposit that is not broadly correlated locally (within a 25 m radius) is inferred to be from a sand fountain process, and not from a tsunami. Aeolian transport from beaches inland also occurs. However, wind transport under all but the most extreme conditions does not provide the energy necessary to transport sand by suspension over broad distances. Instead, most sand transport is by saltation over advancing sand beds or hard plain surfaces. Another possible source of sand into coastal freshwater

settings are storm surges. From a hydrodynamic standpoint, tsunami and storm surges can be similar in sediment transport dynamics.

Discriminating between storm surge and tsunami depositional processes in coastal environments is difficult (Foster *et al.*, 1991). However, the premise of this thesis is ultimately founded on the requirement to distinguish between deposits from tsunami and deposits from storm surges. Though subtle, the distinction is based on sedimentological properties. Tsunami inundation is characterized by relatively few (10 or less), long period wave trains. Hurricanes and storm surges typically consist of thousands of short period waves. As a result, tsunami-deposited sand sheets consist of fewer, but thicker inter-deposited sand layers (Jaffe *et al.*, 1998a). These interstratified sand layers are often intercalated with mud and detritus layers (Darienzo and Peterson, 1991, Atwater, 1992, Atwater and Hemphill-Haley, 1996, Nelson *et al.*, 1996b). The time necessary for settling of fine-grained and organic material between surges is plausible for a longer period wave, such as a tsunami, rather than short period waves such as those generated by a hurricane (Jaffe *et al.*, 1998a). As a result, tsunami sand sheets thin landward and fine upward over a broad area. Further, the inference that a deposit was the result of a tsunami, rather than a storm surge, can be strengthened when considering the larger geologic framework for the region. Tsunami have been inferred to accompany great Cascadia earthquakes (Satake *et al.*, 1996, Nelson *et al.*, 1996, Atwater and Hemphill-Haley, 1996, Darienzo and Peterson, 1996, Clague and Bobrowsky, 1994), and based on field observations have been inferred to flood coastlines concomitantly at sites throughout the entire Cascadia Margin, as well as the Pacific Basin (Satake *et al.*, 1996). The broad regional occurrence of the sand

layers and their apparent age correlation suggests large, region-wide marine surge events typical of a Cascadia earthquake generated tsunami.

Figure 2. Map of the Pacific Northwest Coastline showing relationship to Cascadia Subduction Zone and the locations of observed tsunami sand layers. The study sites for this paper are indicated by stars. Other sites where tsunami layers have been characterized in bays or tidal streams are shown with circles.

Tsunami sand layers exist in coastal bays and wetlands at a number of sites along the Cascadia Margin in the Pacific Northwest (PNW) (Figure 2). These have included tidal marshes at Humboldt Bay, California (Clarke and Carver, 1992), Coos Bay (Nelson *et al.*, 1996a), Alsea Bay (Peterson and Darienzo, 1996), Yaquina Bay (Peterson and Priest, 1995), Siletz Bay (Priest, 1995), Netarts Bay (Darienzo and Peterson, 1990), and the Salmon River Estuary (Grant, 1989), in Oregon and Willapa Bay (Atwater, 1987), and Grays Harbor (Atwater, 1992) in Washington, and at Port Alberni and Tofino on Vancouver Island (Clague and Bobrowsky, 1994) (Figure 2). Tsunami deposits were observed near the mouths of rivers at Lagoon Creek, California (Abramson, 1998), Alsea River, Oregon (Peterson and Darienzo, 1996), Seaside, Oregon (Fiedorowicz, 1997), the Niawiakum and Willapa Rivers, Washington (Atwater and Hemphill-Haley, 1997), the Copalis River, Washington (Atwater, 1992), and various inlets in Puget Sound, Washington (Bucknam *et al.*, 1992). In addition, tsunami deposits have been observed in fresh water lake strata from Bradley Lake in southern Oregon (Nelson *et al.*, 1996b), and Kanim and Catala Lakes exposed along the outer coast of Vancouver Island (Clague *et al.*, 1999; Hutchinson *et al.*, 1997).

Most of these sites are situated within or adjacent to tidal basins and tidal channels where tsunami wave amplification or attenuation occurs (Myers *et al.*, 1999). The objective of this thesis is to present geologic evidence of paleotsunami at open coastal sites along the Cascadia Margin where incoming tsunami are unaffected by tidal channel features. Until now, the record of marine inundation at coastal plain sites along the Cascadia Margin has not been adequately examined.

Prehistoric tsunami sand layers, verified by similar deposits in PNW estuaries have been observed to be overlying coseismically subsided soil and marsh horizons (Figure 3). These coseismically subsided horizons have been interpreted to be the result of Cascadia Subduction Zone earthquakes (Atwater, 1987; Darienzo and

Figure 3. Core logs showing buried soils from bays throughout the Cascadia Margin that have been used by various researchers to support coseismic subsidence. Each subsided peaty soil horizon has been inferred to be caused by a great (Mw 8-9) Cascadia Subduction Zone earthquake. This figure is an adaptation of core logs from published data of Clarke and Carver (1992), Peterson and Darienzo (1996), Darienzo and Peterson (1990) and Atwater and Hemphill-Haley (1997). See Figure 2 for study locations.

Peterson, 1990; Atwater and Hemphill-Haley, 1997). These great (Mw 8-9) earthquakes have a recurrence interval of 500 ± 300 years (Atwater and Hemphill-Haley, 1997), with the most recent earthquake occurring 300 years ago (Yamaguchi *et*

al., 1997). The coseismic subsidence record for the past two thousand years is well-supported by buried wetland soils in northern California at Humboldt Bay (Clarke and Carver, 1992), in Oregon at Netarts and Alsea Bay (Darienzo and Peterson, 1990; Peterson and Darienzo, 1996), and along the coast of Washington in the vicinity of Willapa Bay and Grays Harbor (Atwater, 1987, 1992) (Figure 3). The radiocarbon ages for the subsided soils at these sites are shown in Figure 4.

Figure 4. Ranges of C14 ages for samples extracted from coseismic subsidence horizons from various locations throughout the Cascadia Margin. The age ranges come from published reports. If a tsunami sand layer was observed directly above the subsided soil, it is indicated by a "ts". The average age for each of the subsided horizons is shown below the arrows. Data from Clarke and Carver (1992), Atwater (1990), Darienzo and Peterson (1990), Peterson and Darienzo (1996), and Atwater and Hemphill-Haley (1997).

Paleotsunami deposits characterized in some PNW localities are not correlated to local paleosubsidence horizons. Clague and Bobrowsky (1994) observed sand laminae in estuaries of Vancouver Island that were inferred to be deposited by the

tsunami triggered by the 1964 Mw 9.2 Alaska earthquake. Fiedorowicz (1997) described overbank sand deposits in Seaside, Oregon that were also thought to be deposited by 1964 tsunami. Clague and Bobrowsky (1994) further characterized a 500-800 year-old sand layer identified as tsunamigenic that was not associated with a corresponding local subsidence event. They suggested possible tsunami wave sources as originating from either a penultimate earthquake which occurred ~700-950 cal yr B.P. in southern Alaska (Plafker *et al.*, 1992), some other far-field seismic source in the North Pacific, or a near-field underwater landslide.

In early studies of the Cascadia Margin paleoseismicity, Atwater, (1987) and Peterson *et al.* (1988) detailed paleotsunami sand layers and interpreted their preservation to indicate regional coastal subsidence. The subsidence was then ascribed to coseismic subsidence during Cascadia Subduction Zone earthquakes. In these instances, the correlation between inferred tsunami deposition and subsidence was used to argue that large seismic events caused the subsided soil horizons (Atwater, 1987, Darienzo and Peterson, 1990). However, recent detailed analyses of tsunami deposits are now being used to further understand tsunami inundation characteristics. Observations of coastal paleoseismic stratigraphy have included variations in intracore sand layer thicknesses as a qualitative description of local paleotsunami wave magnitude (Clague and Bobrowsky, 1994; Nelson *et al.*, 1996). Detailed sedimentological analyses of landward thinning sand sheets have resulted in simple modeling of tsunami flooding at Lagoon Creek, northern California where paleotsunami were estimated to have inundation depths of 3.5-14 m and velocities of 2-5 m/s for a distance of 1 km inland from the the shoreline (Abramson, 1998).

Additional observations of field sedimentological data from the geologic record will assist in the development of tsunami-induced sediment transport models, which can be used to more accurately estimate tsunami run-up.

Figure 5. Computer modeling results of tsunami inundation at Neskowin, showing wide range of outcomes. The most catastrophic model predicts inundation elevations of 10m . The lower diagram presents a much less catastrophic model, which predicts an inundation elevation of only 3m. Representative core logs from the field site at Neskowin that were logged for this study are also shown.

The possibility of a tsunami generated from a Cascadia Subduction Zone earthquake (near-field tsunami) poses a significant hazard to coastal communities of the Pacific Northwest (Priest, 1995). As a result, computer models have elucidated

potential velocities and inundation elevations for the tsunami (Whitmore, 1993, 1994; Ng *et al.*, 1990; Priest, 1995). Recent calculations of tsunami runup from modeled Cascadia Subduction Zone rupture mechanics have varied widely in predicted runup. Priest and Baptista (1998) recently estimated, from several models, inundation elevations that ranges from 3 to 10 m for Seaside, Oregon and the Long Beach Peninsula, Washington. The coastal plain is typically 5-6 m above mean sea level. These model values provide ranges in hazard predictions from “no significant runup” to “catastrophic flooding” (Figure 5).

Numerical models of tsunami depend upon knowledge of seismically induced vertical seafloor displacement, which for the Cascadia Margin, is not known (Myers *et al.*, 1999). Attempts to model projected seafloor deformation, based on observations and inferences from fault characteristics (Goldfinger *et al.*, 1992, 1996, Roberts, 1988), generate widely varying predictions of the extent of seafloor deformation and the resulting tsunami. The simplest approach infers vertical uplift of 5 m along a deformation front, decreasing linearly to zero at the coast (Ng *et al.*, 1990). Whitmore (1993) used dislocation formulae dependent upon fault dip, strike, slip, length, width, depth and the thermal characteristics of surrounding rocks, and assumed a rectangular fault locked over its entire area prior to rupture. More recently, fault dislocation models have incorporated a wide range of three-dimensional heterogeneity over the entire locked zone of the fault (Fluck *et al.*, 1997). Assumptions about the rupture mechanics of the Cascadia Margin are speculative.

The extent of tsunami inundation is not wholly dependant upon seismic characteristics. Proximity to foci, extent and nature of seafloor displacement, shoreline

bathymetry, surface geomorphology and vegetation, tide level, meteorologic conditions, and other unknown variables play a significant role in determining the extent of the potential hazards presented by tsunami.

While vertical seafloor displacement is an important forcing mechanism, it does not completely control nearshore tsunami amplitude. Local shoreline features can significantly modify tsunami wave height. For example, studies of the 1960 Chilean earthquake showed that some near- field coastlines were not affected as greatly as far-field locations in Hawaii and California (Lander *et al.*, 1993). Larger than expected near-field tsunami were generated from the Java, Indonesia earthquake of 1994 (Mw 7.2) and the Mindoro, Phillipines earthquake of 1994 (Mw 7.2) where significant horizontal displacement of the seafloor is thought to have played a major role in tsunami wave train development (Tanioka and Satake, 1996). Earthquake-induced subaqueous landslides near the Alaska coast set up local waves that caused damage during the 1964 Alaska earthquake (Plafker *et al.*, 1969). A sub-aqueous slump is thought to have increased the impact of the 1998 Papua New Guinea tsunami. In addition, Hutchinson *et al.* (1997) suggested that the presence of dense forest stands reduced paleotsunami inundation distance at sites along the coast of Vancouver Island. In light of the comparatively great effort in numerical modeling of Cascadia tsunami runup on the open coast, little geologic field evidence exists to test the results. Published evidence of tsunami deposition is largely limited to bay deposits and freshwater pond deposits in the northernmost and southernmost parts of the Cascadia margin. One of the objectives of this thesis is to present the geologic field evidence

for periodic, large scale marine surges into back-barrier settings at various open coastal plains along the central Cascadia Margin.

This thesis will describe the litho-, chemo-, and biostratigraphy of buried sand and detritus layers from freshwater lake and marsh sites at open- coast, back-barrier localities. I will argue that the presence of the sand and detritus layers, which include up to five distinct layers at some sites, are from marine surge events with velocities great enough to entrain and transport sand up to at least 1 km inland from the shoreline. Specifically, the sand and detritus layers mapped at these locations will be presented as evidence of catastrophic tsunami inundation due to their apparent correlation to regional coseismic subsidence horizons. As such, the source of the tsunami are from great Cascadia Subduction Zone earthquakes. This work will provide relevant and useful data to estimate tsunami hazard at the coast. Tsunami wave elevation at the shoreline will be estimated, based on inferred overtopping of paleodune crests, and minimum inundation distances defined. Chemostratigraphic and biostratigraphic techniques were developed to map the maximum landward extent of tsunami inundation.

Setting

Four localities were studied for prehistoric tsunami records within the central Cascadia margin. These localities include Neskowin and Rockaway, Oregon and the Long Beach Peninsula and Grayland Plains, Washington. The study sites, as described below, were chosen because of their relatively broad geographic positions within the central Cascadia Margin (Figure 2). The broad geographical locations of

the study sites are intended to highlight the regional scope and synchronicity of tsunami inundation. In addition, study sites were selected which displayed a high potential for preserving evidence of tsunami inundation.

All of the wetland sites reported here are located in broad, open coastal plains along the central Cascadia margin. Each of them are situated landward of low beach-dune ridges. The study sites are characterized as back-barrier freshwater lake, bog, swamp and marsh settings where anomalous sand layer preservation is enhanced. Host strata at these sites predominantly consist of freshwater peat or mud. Barrier overtopping or breaching by marine water incursions are readily recorded at these sites by the presence of anomalous strata bearing lithological, biological or chemical characteristics typical of a marine source.

Long Beach Peninsula and Grayland Plains, Washington

The geomorphology of the study sites on the Long Beach Peninsula and Grayland Plains, Washington are characterized as a series of longitudinal beach-dune ridges which are oriented parallel with the Pacific Ocean shoreline (Figure 6). Seaward progradation of these dune complexes began about 4-5 ka at Long Beach, and 1-2 ka in the Grayland Plains (Peterson *et al.*, 1999, Woxell, 1998). Linear subsurface erosional scarps that parallel the longitudinal dune ridges indicate catastrophic beach retreat. The beach retreat episodes are inferred to be triggered by coseismic subsidence. Interseismic progradation results in the record of net accretion at the Long Beach Peninsula (Peterson *et al.*, 1999, Meyers *et al.*, 1996). Soil chronosequencing and radiocarbon dating of subsurface erosional scarps indicates that

the sand dunes located closest to the Pacific shoreline have been stable for at least the past 300 years (Woxell, 1998). Where not disturbed by humans, the sand ridges are vegetated with a *Picea sitchensis* (Sitka spruce) - *Gaultheria shallon* (salal) forest community. Freshwater lakes, marshes and bogs have evolved within the depressions situated between the dune ridges (Peterson and Phipps, 1992). The elevation of freshwater in these depressions is controlled predominantly by groundwater. There appear to be no direct runoff sources. As a result, the sediments which accumulate in

Figure 6. Aerial photograph of the Long Beach Peninsula, southwest Washington. The approximate crest of the dune ridges are shown by a white line (Woxell, 1998). The presence of trees on the ridges indicate the accelerated rate of soil development on these upland areas. Interdune lows are occupied by lakes and bogs. Deposits located in these interdune areas were the focus of this study.

the inter-dune marshes are biogenic, and in most cases consist of peat. Where water is too deep to support rooted plant growth, the sediment consists of fine lake muck (gyttia) overlying dune sand. Briscoe Lake, which is located on the Long Beach

Peninsula, offers an example of the varied nature in which sediments accumulate in the inter-dune depressions (Figure 6). The middle of Briscoe Lake has a water depth of approximately 1.5 m, which does not support marsh plant growth. The bottom sediments consist of clean, well-sorted dune sand. However, peat is accumulating along the margins of Briscoe Lake where marsh plant growth is vigorous and wind-wave resuspension is dampened. The south end of Loomis Lake has a peat thickness of greater than 2 m. Tsunami-induced sand sheet deposition would not be easily discerned in the middle of Loomis Lake, but would be easily detected in the peat strata along the margins of the lake.

Figure 7 Photograph of western shoreline of Loomis Lake, which is located approximately 1.2 km from the Pacific Ocean. The lake occupies an interdune low. The margins of Loomis Lake consist of peat while the central portion of the lake has a sandy bottom. Note the well-developed forest that is situated on what was a former dune ridge. The margins of Loomis Lake consist of peat while the central portion of the lake has a sandy bottom.

The stratigraphic suitability of a study site is further complicated by the timing of peat development. The youngest dunes are those located closest to the ocean shoreline. Therefore, the youngest lake/marsh settings are also located closest to the shoreline. These ocean-proximal sites are most likely to experience marine inundation, but may not be old enough to have accumulated a peat sequence with suitable depth to record a great-earthquake tsunami event. Woxell (1998) reported a delay of 500-1000 years between dune ridge soil development and basal peat development in adjacent, landward wetlands.

Neskowin, Oregon

Dune ridges, which are oriented parallel with the Pacific Ocean shoreline, separate the sea from the landward situated freshwater marshes and lakes at Neskowin. The approximate width of the sandy barrier at Neskowin is 200 m, and the elevation ranges from 3 m to 12 m (Oregon Dept. of Transportation, 1967). Major streams do not flow into the marsh in the vicinity of the study site. The peat sequences, ranging from 1-3 m in thickness, overlie fine-upper to medium-lower, well-sorted sand which is assumed to be the upper surface of a paleodune or beach backshore complex associated with a progradation of the prehistoric shoreline. The history of accretion of the sand at Neskowin has not yet been studied. The time of initiation of sand accumulation at Neskowin is not known. Basaltic bedrock and steep hillslopes of colluvium outcrop at distances ranging from 0.5 to 1.5-km inland from the shoreline. Abandoned sea stacks occur within the progradative beach plain.

Rockaway, Oregon

Rockaway is characterized by a single barrier dune-ridge complex which separates the freshwater sites from the sea. However, the stratigraphic sites at Rockaway consist of lakes, with most cores taken from Crescent Lake and Lake Lytle (Figure 8). The sediments consist of fine, organic-rich muck and mud (gyttja) and extend to thicknesses of greater than 3 m. Barrier dune elevations range from 5 m to 8 m (Oregon Dept. of Transportation, 1967), and barrier width is approximately 400 m. The dunes separating Lake Lytle and Crescent Lake from the Pacific Ocean have been

Figure 8. Aerial Photograph of Crescent Lake and Lake Lytle, Rockaway, Oregon. The solid white line outlines the crest of the dune ridge (elevation ~6m) which separates the Pacific Ocean from the back barrier wetlands.

locally downcut for residential development. Colluvium from debris flow spoils originating on steep slopes to the east of Rockaway, forms the eastern shoreline of the

lakes at Rockaway. Previous work by Eilers *et al.* (1995) at Lake Lytle suggested the presence of anomalous sand layers, possibly of tsunami origin. Carbon material sampled by Eilers from one of these layers returned an age of 1340 ± 40 radiocarbon years before present.

Methods

Stratigraphic Core Logs

Buried sand sheet deposits were analyzed from about 200 sites at the four study areas during the summers of 1997 and 1998. Shallow wetland stratigraphy was sampled using a 3-cm diameter gouge core. Depths ranged from 0.5 to 4.5 m below wetland bottom. All cores were logged in the field at the time of the sampling, with observations including depth to sand-peat/mud contact, sediment composition and structure, grain size, and significant color characteristics. Samples were collected at select strata for further radiometric dating, sand grain size analysis, and microfossil assemblage analyses. In addition, 7-cm diameter aluminum pipe cores were used for sampling, and later split for more detailed logging and laboratory analyses, including diatom and geochemical scans for marine markers. The pipe cores were driven into the ground using vibracore and pound core techniques. Vibracore extraction was performed using a hand winch mounted on a tripod for marsh sites. For pond sites, the winch system was attached to the front of an inflatable boat.

The locations of core sites in Long Beach were recorded with a Garmin model 45 XL global positioning system (GPS). Real time GPS data (± 10 m) was acquired

Figure 9. Pictures of various vibracore apparatus. Photograph in upper left shows the vibracore assembly attached to the bow of a zodiac inflatable boat. The vibracore head is being loosened from its grip on a 7-cm diameter aluminum pipe core in the lower left photograph. The pipe core is about to be winched out from its extension into the bottom of a lake. The photograph at the right shows the tripod, which is situated in a cranberry bog, that is used for support when winching out cores.

from an attached differential receiver which utilized base station data from a beacon located at Fort Stevens, Oregon. The locations of core sites in Grayland, Washington and Neskowin, Oregon were recorded using a Trimble Geoexplorer II GPS unit.

Satellite coordinate data collected in the field was post-processed using Trimble

differential correction software. The accuracy of the Trimble Geoexplorer II GPS sites is on the order of ± 1.5 m. In addition, estimated locations were approximately located on a USGS 7.5' quadrangle at the time of the coring. The locations of the gouge core sites and logs of the cores for Long Beach, Grayland, Rockaway and Neskowin are included in Appendices B, C, D, and E respectively.

Material selected for radiocarbon dating included twigs, conifer needles, and bulk peat samples (Figure 10). Twigs, conifer needles and other terrestrial detritus materials were extracted from select detrital caps. When possible, twigs were picked from sand layers. Carbon samples extracted from these strata were dated using AMS

Figure 10. Examples of horizons from which carbon samples were extracted for ^{14}C dating.

techniques, and are used to bracket the age of the marine inundation. Bulk peat samples were dated from either directly below tsunami sand layers, or directly above basal sand units. In light of the potential contamination of carbon reservoirs by younger carbon, bulk peat samples were picked free of descending root material, and were rinsed with de-ionized water. Organic materials were submitted to Beta

Analytical, Inc. of Miami, Florida for age dating. Samples were analyzed by conventional and standard atomic mass spectrometry (AMS) techniques.

Radiocarbon dates will be reported in the text of this thesis as “calibrated years before present” (calYBP). The calibration of the conventional ^{14}C age, which is reported as “radiocarbon years before present” (RCYBP) in ^{14}C laboratory summaries, is based on the calibration curves developed by Stuiver and Reimer (1993) for tree rings of known age. It is also important to note that in the reporting of these ^{14}C ages, “present” refers to the calendar year 1950. The ^{14}C ages were not converted to calendar years because of the convention established in the literature for reporting the age of Cascadia coastal subsidence events as years before present.

Geochemical Screening

Two stratigraphic cores from Briscoe Lake and one core from Neskowin were analyzed for the presence of marine geochemical markers using neutron activation analysis (INAA) techniques. The elements targeted for analysis included sodium, potassium, and bromine. In addition, a number of rare earth elements and other trace elements were included in the analysis. The sediment cores were subsampled at 2-cm intervals. Each subsample was placed into an evaporating dish which had been washed with detergent and triple rinsed with deionized water. The sample and dish were then placed in a drying oven for 36 hours. The dried samples were transferred to a stainless steel mortar and pestle, pulverized to a fine powder, and transferred to labeled plastic film containers. To minimize loss of potential tracers, the samples were not rinsed and were not screened. Each dried and pulverized INAA sample was

split into four equal parts, the final fraction of which was transferred to a massed 1-gram poly-ethylene vial (poly-vial). In addition to field samples, a lab-standard sample of 80 ppm potassium bromide solution and a lab-standard sample of coal fly ash were analyzed to provide analytical calibrations for bromine as well as a broad suite of elements.

The INAA samples were examined at the nuclear reactor at Reed College, Portland, Oregon, and were irradiated for one hour with a neutron flux of 2×10^{12} neutrons/cm²/sec. Irradiated samples were counted at Portland State University using an EG&G Ortec Germanium Crystal 92X Spectrum Master radiation counter. First counts were completed four days after irradiation, and included counting times of 3000 seconds for each sample. Standards of coal fly ash and 80 ppm potassium bromide solution were counted for 9000 seconds. The Spectrum Master counting system was calibrated using the TN4000 calibration sample, which was counted for 10,000 seconds. Standard reference tables were built from the coal fly ash and the potassium bromide standards.

Diatom Screening

Two stratigraphic cores from Briscoe Lake, Long Beach Peninsula, Washington were subsampled at 1.0-cm intervals for analysis of marine diatoms (Hemphill-Haley, 1996, Darienzo, 1987). Each subsample was treated for 72 hours with 20 mL of 30% hydrogen peroxide which had been diluted with 25 mL of deionized water. The slurry was rinsed through a 220 mesh screen with deionized

water. After settling for 30 minutes, 10 mL of the rinsate was extracted and stored in a vial for later dry-mounting onto a microscope slide.

Dry mounts of diatom samples were prepared by adding a drop of the isolated rinsate onto a microscope slide, drying on a hot plate, and preserving with a drop of preservaslide brand resin. Diatom assemblages were quantified by counting the total number of diatoms from ten randomly positioned microscope fields of view for each sample (Barnett, 1997, Hemphill-Haley, 1996, Hutchinson *et al.*, 1997). In addition, the number marine species from each of these microscope fields were also recorded. From these data a marine diatom fraction was determined for each centimeter down the core.

Sand Grain Size Analysis

Sand layers from two core transects at Neskowin were analyzed for grain size distributions using sieve fractionation techniques (Folk, 1980). These methods were employed to test for the presence of fining up sequences and fining inland trends. These trends are needed to validate turbulent suspension transport processes. Intrasand layer variations of mean grain size were analyzed by sampling an entire sand layer thickness from a given core and comparing it to an analysis of the same layer split into lower and upper fractions. Lateral variations in sand sheet grain size were analyzed by sampling correlated sand layers from distinct horizons. Sand samples were introduced to a stack of sieves ranging from mesh numbers 35 to 140, and were mechanically shaken for 20 minutes. Each size fraction was massed so that

cumulative percentages of the sizes could be plotted and moment statistics calculated (Appendix H).

Correlating Sand Layers

A number of methods were integrated to establish sand layer correlations. At a given study site, sand layers were correlated based on relative thickness, a proxy for time, between deposits, as indicated by intervals of peat or mud deposition. Some sand layers were associated with changes in the texture and color of the peat and mud. These changes reflect water level changes probably related to regional subsidence and uplift or local drainage conditions. When possible, stratigraphic markers in addition to the sand layers were utilized to make correlations. Radiometric dating (^{14}C) data was used to support the correlations.

Distinguishing a Sand Layer as Tsunamigenic

Sand-laden marine surge deposits in open coastal freshwater settings could be from two possible sources – either a tsunami, or a storm surge. In light of this potential problem, Table 1 was developed, which shows a list of characteristics that were used to distinguish between tsunami and storm surge deposits. A deposit was considered to be of marine origin if it displayed any of characteristics 1 or 2. The deposit was considered to be the result of a high velocity marine surge if it displayed any of characteristics 3 – 6. Those characteristics which are marked by an asterisk are unique for a tsunami deposit. To be considered tsunamigenic, the deposit had to display at least two of the characteristics marked by an asterisk.

Table 1. Characteristics that were used to distinguish sand and detritus layers as 1) marine derived, 2) high or low velocity surge deposited, and 3) tsunami or storm surge deposited.

Deposit Characteristic	Yes	No
1. Clasts are well-sorted, quartz-rich (derived from beach sand)		
2. Deposit shows a marine marker		
3. Deposit thickness decreases landward		
4. Clast size decreases upsection within deposit		
5. Clast size decreases in the landward direction		
6. Deposit is capped by terrestrial derived organic material (termed detritus in the text)		
*7. Deposit extends landward over a great distance (>0.5km)		
*8. Deposit contains intrasand mud or detritus layers		
*9. Deposit is broadly correlated locally (within a few km)		
*10. Deposit is broadly correlated regionally (hundreds of km)		
*11. Radiocarbon data suggests correlation to a prehistoric Cascadia earthquake event as described in the literature		

Dune Stabilization Studies

Soil profiles were measured and logged in the field at each of the study sites, except Neskowin, to verify stability of the primary barrier dune for at least the past 300 years. Analysis of B soil horizons was done using Rubification (Harden and Taylor, 1983) and Buntley-Westin (Birkeland, 1984) techniques to establish relative age chronosequence values. The duration of soil weathering is approximated by the accumulation of iron hydroxides/oxides in the soil B horizon yielding red-brown coloration. The rate of soil development is dependant on precipitation, vegetation and

parent materials. Preliminary chronosequences for Oregon coastal dunes demonstrate multi-century periods for soil B development (Burns and Peterson,). In addition, the results of sand dune radiocarbon dating, paleosol radiocarbon dating, and mapping of erosional scarps by Peterson *et al.* (1999) were used in the southwest Washington localities to establish the stability of shoreline dune ridges. The soil profile locations, stratigraphic logs, Rubification values and Buntley-Westin indices are located in Appendix F.

Statistical Analyses

Statistical methods of raw diatom data included chi-squared (χ^2) test for homogeneity and Z-test of proportions (Davis, 1986). These statistical tests were selected to show the distinction between the freshwater horizons and the marine horizons in a given core, as signified by characteristic diatom assemblages for each of these strata. It was assumed that the marine-influenced horizons were due to the anomalous incursion of ocean water into the freshwater setting. The freshwater horizons – e.g. peat and mud, occupy significant lengths of the core. The marine-influenced horizon – e.g. sand and detritus, was inferred to occupy an area adjacent to the sand/detritus-peat/mud contact. These target horizons were analyzed in greatest detail for evidence of anomalous marine diatom abundance.

As discussed above, the χ^2 test for homogeneity was performed on two intracore horizons. Each horizon in a pair was inferred to represent diatom assemblages indicative of contrasting salinity conditions. One horizon was representative of freshwater conditions, while the second horizon was inferred to

represent marine conditions. Therefore, for the χ^2 test, the following null hypothesis can be stated:

Ho: The proportions of marine diatoms in the freshwater strata (peat and/or mud) and at the upper contact of with the tsunami horizon are the same.

And the accepted hypothesis would be:

H1: The proportions of marine diatoms in the freshwater strata (peat and/or mud) and at the upper contact with the tsunami horizon are different.

The equation for the χ^2 – test for homogeneity is:

$$\chi^2 = \sum \frac{(o_i - e_i)^2}{e_i}$$

where o_i is the observed number of diatoms and e_i is the expected number. The expected number e_i is calculated from

$$e_i = \frac{(c_i)(r_j)}{\sum c_i + \sum r_j}$$

where c_i and r_j are the rows and columns, respectively (see Table 2). Degrees of freedom are calculated from:

$$v = (r-1) + (c-1)$$

Columns (c_i) corresponded to the total diatom counts (both marine and freshwater) for a given horizon, and rows (r_j) represented the total number of diatoms in a salinity condition from both horizons. A sample table is shown below:

Table 2. Row and column arrangement of diatom counts used to calculate χ^2 statistic.

	Stratigraphic Horizon		Row Total
	Inferred Freshwater Horizon	Inferred Marine Horizon	
Salinity Division	Diatom Counts	Diatom Counts	
Marine			r_1
Fresh and Brackish			r_2
Column Total	c_1	c_2	

The z-test of proportions was used to compare the proportions of marine diatoms in the freshwater strata (peat and/or mud) with proportions of marine diatoms at the contact between the freshwater lithology and the upper contact with a given inferred tsunami horizon. The null hypothesis for the z-test would be the same as that stated above for the χ^2 test. The z value is determined by the formula:

$$z = \frac{p_1 - p_2}{\sqrt{\hat{p}(1 - \hat{p}) \left(\frac{1}{n_1} + \frac{1}{n_2} \right)}}$$

where p_1 and p_2 are the proportions of marine diatoms for each of the respective stratigraphic horizons. In the equation for z, $\hat{p}$ is the pooled estimate of the proportion p given by:

$$\hat{p} = \frac{x_1 + x_2}{n_1 + n_2}$$

where x_1 and x_2 are the number of diatoms from a salinity division (marine and non-marine) in each of two samples, and n_1 and n_2 represent the total number of diatoms counted. The value q is equal to $1 - p$. The z-test uses a two-tailed level of significance (α) of 0.05 for the critical region of $|z| > 1.96$.

The correlation coefficient, r_{jk} , was calculated for degree of correlation between 1) raw diatom data and raw geochemistry data, 2) raw diatom data and lithology, and 3) raw geochemistry data and lithology. The correlation coefficient is calculated by using the relationship:

$$r_{jk} = \frac{COV_{jk}}{s_j s_k}$$

where s_j and s_k are the standard deviation of the respective data sets, and COV is the covariance between two populations. Sample calculations of each of these methods is shown in the results section of this paper.

Results

Inferred tsunamigenic sand layers were observed in 30% of the cores from the Long Beach Peninsula, 100% of cores from the Grayland Plains, 100% of cores from Lake Lytle and Crescent Lake, and 100% of cores from Neskowin. More than one tsunami layer was observed in 75% of the cores. Core site 123 from northerly Crescent Lake hosted seven anomalous sand horizons, and included a basal tsunami sand layer with an age of 2000-2325 calYBP.

Generalized Sand Layer Characteristics for the Study Area

Representative core logs of characteristic strata are shown in Figure 11. Target sand layers ranged in thickness from 0.2 cm to 45 cm. Generally, they were comprised of fine-upper to medium-upper (0.17-0.50 mm), rounded to well-rounded,

Figure 11. Detailed stratigraphic logs showing the characteristics of tsunamigenic sand layers from cores at Rockaway, Oregon (a) and Long Beach, Washington (b).

quartz-rich sand (Figure 8, Appendices A-D). In addition, the presence of terrestrial derived, organic-rich detritus layers were observed to overlie many of the sand layers. Detritus layers were also observed in cores as distinct units without sand. The sand layers in the core for Rockaway in Figure 11 displayed characteristics 1, 3, 4, 6, 7, 8, and 9 from Table 1. The sand layers for the Long Beach core in Figure 11 displayed characteristics 1, 3, 6, 7, 8, and 9 from Table 1. Hence, they are inferred to have been deposited by tsunami.

Figure 12a. Photograph of gouge core showing lamina of sand situated in a strata of peat.

Most sand layers of greater than 3 cm thickness displayed fining-up sequences of sand grain size. Many of these layers rest conformably upon abrupt contacts with underlying peat and mud. Layers of mud and terrestrial, organic-rich detritus overlie the sand layers. The upper contact of the sand/detritus layers typically grade into overlying sediments, which again, consist of peat and mud. Observed within the sand layers were horizons and laminae of terrestrial detritus, mud, and mud rip-up clasts.

The intrasand layers of detritus are particularly well-developed in proximal cores at Rockaway (Figure 11 and Appendix D), and the southerly cores at Neskowin (Appendix E). In freshwater lake sites where the hosting sediments consisted of loosely consolidated, poorly compacted mud, the older sand layers often displayed ascending and descending sand dikes (Figure 11). Thin sand sheet layers (<3 cm) and

Figure 12b. Photograph of pipe cores showing tsunami layers. These cores are from a transect at Neskowin, Oregon. The black lines show the correlation between the sand layers in the cores. Photograph above shows a sand laminae from a gouge core.

laminae (0.2-0.4 cm) of sand deposits included sharp and gradual contacts with underlying peat and mud hosting sediments. Detritus layers consisted of a mixture of wood bits, twigs, conifer needles, blades of grass, bits of bark, conifer cones, and fine-grained inorganic material such as mud and silt.

Radiocarbon Dating

The results of radiocarbon dating are shown in Table 3. The results shown on Table 3 have been listed as both Radiocarbon Years Before Present (RCYBP), and as Calibrated Radiocarbon Years Before Present (calYBP), where “present” corresponds to 1950. RCYBP corresponds to conventional, uncalibrated radiocarbon years.

Figure 13. Results of radiocarbon dating of select horizons from all of the study sites presented in this thesis. Similar dates are enclosed in a box, and have been averaged, as shown by the date to the right. Questions marks indicate dates that do not fall within the various groupings. The 265 calYBP, 1145 calYBP, 1350 calYBP, and 2170 calYBP correlate to Cascadia Subduction Zone earthquakes as outlined in the literature (Atwater, 1987; Clarke and Carver, 1992; Clague and Bobrowsky, 1994; Darienzo and Peterson, 1990).

The calibrated results are based on the calibration curves derived from tree-ring dating, which have been developed by Stuiver and Reimer (1993). Based on the calibrated results, the ages appear to bracket the following ranges for material extracted from within the tsunami layers, or directly above, or directly below the

tsunamigenic layers (Figure 13): 0-530 calYBP (core sites 189, 123, 44 – see Table 3 below), 735-975 calYBP (core sites 91, 58, 55, 172, 112, 162 – see Table 3), 920-1290 calYBP (core sites 153, 11 – Table 3), 1280-1410 calYBP (core site 61), and 2000-2325 calYBP (core site 123). Potential errors associated with dating of carbon material include the possibility that erodable marine surges could have picked up pre-event detrital material from foredune soils, leading to older dates. Bulk peat and/or detrital caps above the sand layers are potentially contaminated by descending roots, supplying younger carbon from above. In addition, the high radiocarbon production of about 300 years ago typically yields younger radiocarbon dates by about 100 years for that time period (Stuiver and Reimer, 1993).

Dune Paleosol Analysis

Dune soils and/or buried paleosols have been mapped at three of the study sites – Grayland and Long Beach (Woxell, 1998), and Rockaway. These soils contain 5-15 cm thick Bw horizons (Woxell, 1998; Appendix F), indicating relatively stable dune-ridge elevations for the past several hundred years. Soil rubification values are shown below in Table 4 for two soil profiles on the Rockaway dune ridge. The presence of paleosols in Table 4, which is indicated by a significant increase in the degree of iron oxide accumulation of the Bw horizon (brownish-orange coloration), is inferred to represent relative stability of dune elevation for the past several hundred years of soil formation. The burial of the paleosol indicates at least one event of dune ridge reactivation and sand deposition. Similar reactivations of beach dune ridges in

Table 3. ^{14}C results from core sites at Long Beach, Grayland, Rockaway and Neskowin.

*Age	**Cal YBP	Analysis Method	Material	Stratigraphic Setting	Locality	Lab Number	Core Site
50 $\pm$ 50	210-265	AMS	Stick	Cranberry bog - 37 cm below sediment surface. Resting on sand lamina	Summer Lane Transect - Grayland	Beta-122812	189
110 $\pm$ 50	0-280	AMS	Twig	Freshwater Lake - 37 cm below sediment surface. Resting on 25 cm thick sand layer.	Crescent Lake	Beta-120443	123
132 $\pm$ 5% (mod.)		AMS	Peat	Freshwater marsh - 15 cm below sediment surface. Situated above 3 cm sand layer. Sand lamina is above.	North Marsh Transect - Neskowin	Beta-120441	140
400 $\pm$ 60	300-530	Standard Radiometric	Peat	Freshwater marsh - 30 cm below sediment surface. Situated above basal sand layer.	North Briscoe Lake	Beta-111641	44
480 $\pm$ 50	465-550	AMS	Stick	Cranberry bog - 82 cm below sediment surface. Resting above sand layer.	Summer Lane Transect - Grayland	Beta-122813	188
680 $\pm$ 50	550-685	AMS	Peat, Wood bits	Freshwater Lake - 25 cm below sediment surface. Resting above diffuse sand layer.	Lake Lytle	Beta-120442	125
810 $\pm$ 40	735-795	AMS	Peat	Freshwater marsh - 62 cm below sediment surface. Overlain by a 3 cm thick sand layer	Crescent Lake	Beta-132248	91
820 $\pm$ 50	665-795	AMS	Stick	Freshwater marsh - 57 cm below sediment surface, resting above 1 cm sand layer within otherwise continuous peat section	South Briscoe Lake	Beta - 122808	58
860 $\pm$ 50	675-915	AMS	Peat	Peat bog - 72 cm below sediment surface, resting with detritus above 1 cm thick sand layer within continuous peat section	South Tinker Lake, Pacific County Offices	Beta - 122811	55
910 $\pm$ 50	655-880	AMS	Stick	Cranberry bog - 70 cm below sediment surface. Resting above sand layer.	Gould Road Transect - Grayland	Beta-122810	172
940 $\pm$ 40	840-975	AMS	Peat	Freshwater marsh - 58 cm below sediment surface. Overlain by an 18 cm thick sand layer.	Crescent Lake	Beta-132247	112
980 $\pm$ 50	770-965	AMS	Peat/ Detritus	Freshwater marsh - 70 cm below sediment surface. Overlain by 13 cm sand.	Central Transect - Neskowin	Beta-120438	162
1180 $\pm$ 100	920-1290	Standard Radiometric	Peat	Freshwater marsh - 75 cm below sediment surface. Situated below 10 cm sand layer - third layer downcore.	North Marsh Transect - Neskowin	Beta-120440	153
1240 $\pm$ 60	990-1285	Standard Radiometric	Peat	Freshwater marsh - 95 cm below sediment surface. Situated above basal sand.	South Briscoe Lake	Beta-11640	11
1310 $\pm$ 70	1065-1320	Standard Radiometric	Peat	Freshwater Lake - 225 cm below sediment surface. Situated above basal sand.	South Loomis Lake	Beta-11642	45
1450 $\pm$ 50	1280-1410	AMS	Wood bits	Peat bog - 58 cm below sediment surface, resting with detritus above 3 cm thick sand layer	South of Cranberry Road	Beta - 122809	61
2170 $\pm$ 50	2000-2325	Standard Radiometric	Wood bits, twigs	Freshwater Lake - 196 cm below sediment surface. Resting above sand layer.	Crescent Lake	Beta-120439	123

*Dates are reported as RCYBP - radiocarbon years before present, where present = AD 1950.

**Calibrated ages are based on two-sigma deviations about ^{14}C calibration curves established by Stuiver and Reimer (1993)
Radiocarbon analyses were performed at Beta Analytical, Inc., Miami, Florida

southwest Washington have been related to multiple earthquake events (Woxell, 1999). These paleosols in the Rockaway dunes have not been dated, but likely represent reactivation by one or more Cascadia earthquake, pushing the age of the initial dune ridge back to 1,100 or 1,300 cal YBP. The Neskowin barrier-spit has not been evaluated for relative dune-ridge age or elevation stability based on soil or paleosol development. Although the position of the barrier is constrained by the seaward extent of the ^{14}C -dated wetlands at Neskowin, the overtopping elevations of paleotsunami are not as well constrained for this site. Whereas the dune-ridge

Table 4. Rubification values for two soil profiles from the sand dune at Rockaway, Oregon.

Core Location - Rock-5-070898

Sample Depth (cm)	dry color		wet color		Comments
	Hue	Value/Chroma	Hue	Value/Chroma	
5	10 YR	3/2	10 YR	2/2	A Horizon
20	10 YR	5/3	10 YR	3/2	
75	10 YR	5/4	10 YR	3/4	
100	10 YR	5/6	10 YR	3/4	
125	10 YR	4/2	10 YR	3/2	Paleosol
200	10 YR	5/4	10 YR	3/4	
250	10 YR	5/4	10 YR	3/4	
300	10 YR	5/3	10 YR	3/3	
320	10 YR	4/3	10 YR	2/2	Paleosol
325	10 YR	5/6	10 YR	3/4	

Core Location - Rock-8-070898

Sample Depth (cm)	dry color		wet color		Comments
	Hue	Value/Chroma	Hue	Value/Chroma	
5	10 YR	3/1	10 YR	2/2	A Horizon
10	10 YR	5/4	10 YR	4/2	
25	10 YR	6/3	10 YR	4/2	
50	10 YR	6/3	10 YR	5/2	
100	10 YR	6/2	10 YR	5/3	
165	10 YR	6/6	10 YR	4/4	Paleosol
175	10 YR	5/6	10 YR	4/4	
200	10 YR	6/3	10 YR	4/3	
225	10 YR	6/4	10 YR	4/2	

elevations are established at Grayland, Long Beach and Rockaway for the last Cascadia tsunami (1700 AD), the elevations for the earlier tsunami are not well

constrained. Additional radiocarbon or optical dating is needed to establish the earliest ages and elevational stabilities of these dune ridges.

Long Beach Peninsula, Washington

The locations of the core sites on the Long Beach Peninsula are shown in Figure 14. Of the 65 core sites on the Long Beach Peninsula, 20 (30%) contained possible tsunamigenic horizons. These included sites 2, 3, 4, 8, 11, 35, 43, 44, 54, 55, 56, 57, 58, 59, 60, 61, 62, 63, 64, and 65. The apparent lack of sand layers observed in many of the cores from the Long Beach Peninsula will be discussed later in this thesis.

Biostratigraphy

The scan of microfossils in the upper 41 cm of core 11 located at Briscoe Lake indicated two areas of elevated levels of marine diatom species (Table 5 and Figure 15). Statistical tests for χ^2 and the z-statistic suggest that the proportion of marine diatoms is different in the peat and detritus directly above the inferred tsunami horizon than in the remainder of the strata, which is characterized by peat. Statistically significant increases in such marine diatom species as *Amphora proteus*, *Paralia sulcata*, and *Thalassiosira paficia* at 9-10 cm and 35-40 cm below the surface of the sediment suggest inundation of marine waters at these horizons (Figure 16). The correlation between the presence of detritus in the lithology and the percentage of marine diatoms for the strata from 0 cm to 41 cm yield a coefficient of 0.7. Table 5 shows the χ^2 test and z-parameter for the upper 10 cm of core 11. A tsunami horizon is interpreted to occur at the interval of 8-10 cm. As a result, the two sample populations for which χ^2 and z were determined included the inferred freshwater-

Figure 14a. Map of the southern Long Beach Peninsula showing core sites.

Figure 14b. Map of the northern Long Beach Peninsula showing core sites

Table 5. Diatom counts, core 11.

Table 3: Diatom counts, core 11.

Marine Species	Depth in Section-cm																
	1	2	3	4	5	6	7	8	9	10	11	20	21	22	23	24	25
<i>Achnanthes brevipes</i>																	
<i>Actinopteryx senarius</i>																	
<i>Amphora proteus</i>	1				1		1			9	2						
<i>Biddulphia dubia</i>																	
<i>Cocconeis scutellum</i>																	
<i>Coccinodiscus radiatus</i>										2							
<i>Dephneis surinella</i>																	
<i>Grammatophora oceanica</i>																	
<i>Hyalodiscus scoticus</i>					2												
<i>Paralia sulcata</i>			1							2			1				
<i>Thalassiosira eccentrica</i>																	
<i>Thalassiosira pacifica</i>				3	2		2	3	1	12	8	1					
<i>Trachysphenia australis</i>								1									
Total Marine:	1	0	1	3	5	0	3	4	1	25	10	1	1	0	0	0	0
Total Observed from 10 fields:	62	58	67	190	488	315	304	278	510	324	171	33	69	39	63	39	53
Percent Marine:	0.016	0	0.015	0.016	0.01	0	0.01	0.014	0.002	0.077	0.0585	0.03	0.014	0	0	0	0

Marine Species	Depth in Section-cm																				
	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40	41					
<i>Achnanthes brevipes</i>																					
<i>Actinopteryx senarius</i>																					
<i>Amphora proteus</i>								1		2		1	4	6	5	5					
<i>Biddulphia dubia</i>																					
<i>Cocconeis scutellum</i>								3	14	1		13	21	21	8	17					
<i>Coccinodiscus radiatus</i>																					
<i>Dephneis surinella</i>																					
<i>Grammatophora oceanica</i>																					
<i>Hyalodiscus scoticus</i>									1								3				
<i>Paralia sulcata</i>			1	1	1		2	7	15	1			1				12				
<i>Thalassiosira eccentrica</i>																					
<i>Thalassiosira pacifica</i>			1		1	2	1	2	7	1		18	22	28	14	12					
<i>Trachysphenia australis</i>									1			2									
Total Marine:	0	0	2	1	2	2	3	13	38	5	0	34	48	55	27	53					
Total Observed from 10 fields:	60	34	85	67	147	75	47	133	482	157		632	607	788	699	325					
Percent Marine:	0	0	0.024	0.015	0.014	0.027	0.064	0.098	0.079	0.032		0.054	0.079	0.07	0.0386	0.163					

dominated horizon from 0-8 cm, and the inferred marine horizon at 8-10 cm. The z-parameter and χ^2 test were also determined for the horizon in core 11 from 11 cm to 39 cm (Table 6). The freshwater horizon was inferred to extend from 11 cm to 34 cm, and the tsunami horizon was inferred to occur at 34 cm to 39 cm.

Figure 15. Core log and associated percentage of marine diatoms for a core from core site 11, Briscoe Lake, Long Beach Peninsula.

Figure 16. Photographs of diatom examples. *Diploneis* sp. is shown on the left, *Paralia sulcata* is in the right photo.

Table 6. Chi-squared calculation for upper 10 cm of core 11.

Null Hypothesis (Ho): For strata that do not host detritus or sand (peat), and upper contacts between peat and sand/detritus, the proportion of marine diatoms is the same.

The upper peat-sand contact occurs at 8-10 cm (see figure 15).

Observed and Expected diatom values

	Stratigraphic Setting				
	Peat (0-8 cm)		Detritus Unit (8-10 cm)		
Salinity Division	Observed	Expected	Observed	Expected	Row Total
Marine	17	40	35	12	52
Fresh and Brackish	1762	1739	495	518	2257
Column Total	1779		530		2309

Chi-Squared Test

	Stratigraphic Setting		
	Peat (0-8 cm)	Detritus Unit (8-10 cm)	
Salinity Division	(O-E) ² /E	(O-E) ² /E	χ^2
Marine	13.28	45	
Fresh and Brackish	0.3	1	59

Reject Ho if $\chi^2 > 3.82$, based on 5% confidence limit and 1 degree of freedom

Z-Test for Difference of Two Proportions

Core Site 11, 0-8 cm and 8-10 cm

Reject Ho if $Z > 1.96$

	Statistical Parameters					
Section Compared	x_1	x_2	p_1	p_2	n_1	n_2
Peat (0-8 cm) to Peat-Detritus Contact (8-10cm)	17	36	0.32	0.68	1762	1005

Test Results

$\hat{p}$	$\hat{q}$	Z	Result
0.02	0.98	-66.2	Reject Ho

The scan of the upper 25 cm of core 44 indicated one area of elevated marine species of diatoms at 25 cm below the surface of the sediment (Table 8 and Figure 17). Statistical tests for χ^2 and the z-parameter suggest that the proportion of marine diatoms is different in the peat directly above the sandy detritus than in the rest of the strata, also composed of peat. Statistically significant increases in such species as *Amphora proteus*, *Paralia sulcata*, and *Thalassiosira paficia* at 25 cm below the surface of the sediment suggest intrusion of marine water at this horizon (Figure 16).

The correlation between the location of the contact at 25 cm and the percentage of marine diatoms for the strata from 0 cm to 24 cm indicated a coefficient of 0.8. Table

Table 7. Chi-squared calculation for 11-39 cm of core 11.

Null Hypothesis (Ho): For strata that do not host detritus or sand (peat), and upper contacts between peat and sand/detritus, the proportion of marine diatoms is the same.

The detrital unit occurs at 34-39 cm (see figure 15).

CHI-Squared Test of Homogeneity - Core Site 11, 11-33 cm and 34-39 cm

Observed and Expected Diatom Values

	Stratigraphic Setting				Row Total
	Peat (11-33 cm)		Detritus Unit (34-39 cm)		
Salinity Division	Observed	Expected	Observed	Expected	
Marine	12	42	138	108	150
Fresh and Brackish	811	781	2011	2041	2822
Column Total	823		2149		2972

Chi-Squared Test

	Stratigraphic Setting		χ^2
	Peat (11-33 cm)	Detritus Unit (34-39 cm)	
Salinity Division	$(O-E)^2/E$	$(O-E)^2/E$	
Marine	21	8.0	
Fresh and Brackish	1.1	0.4	31

Reject Ho if $\chi^2 > 3.84$ - based on 5% confidence limit and 1 degree of freedom

Z-Test for Difference of Two Proportions, Core Site 11, 11-33 cm and 34-39 cm

Reject Ho if $Z > 1.96$

Section Compared	Statistical Parameters					
	x_1	x_2	p_1	p_2	n_1	n_2
Peat (11-33 cm) to Peat-Detritus Contact (34-39cm)	12	193	0.08	1.29	811	2011

Test Results

$\hat{p}$	$\hat{q}$	Z	Result
0.07	0.93	-111.8	Reject Ho

9 shows the χ^2 test and z-parameter for the upper 24 cm of the core. A tsunami horizon was inferred to occur at the interval of 24-33 cm. As a result, the two sample populations for which χ^2 and z were determined include the freshwater-dominated horizon from 0-24 cm, and the inferred marine contact at 24-25 cm. Freshwater assemblages of diatoms observed in both cores were dominated by *Eunotia pectinal*, *Pinnularia viridis*, *Gomphonema parvulum*, *Gomphonema augustatum*, and *Navicula bacillum*.

Figure 17. Core log and associated percentage of marine diatoms for a core from core site 44, Briscoe Lake, Long Beach Peninsula.

Table 8. Diatom counts for 1-cm thick samples from core 44, Briscoe Lake, Long Beach Peninsula.

Marine Species	BRI-0	BRI-1	BRI-2	BRI-3	BRI-4	BRI-5	BRI-6	BRI-7	BRI-8	BRI-9	BRI-10	BRI-11
<i>Achnanthes brevipes</i>					1	1						
<i>Actinopteryx senarius</i>												
<i>Amphora proteus</i>	2			2	2	2			2		3	
<i>Biddulphia dubia</i>												
<i>Cocconeis scutellum</i>		1	1	2	2	1	1	1		4	2	1
<i>Coscinodiscus radiatus</i>								1				
<i>Dephneis surirella</i>					1							2
<i>Grammatophora oceanica</i>					1				1	1		
<i>Hyalodiscus scoticus</i>	3	3	4	2	2		3				2	
<i>Paralia sulcata</i>	7	5	3	3	4	3	4	6	10	8	2	2
<i>Thalassiosira eccentrica</i>									1	1		
<i>Thalassiosira pacifica</i>	1	1		7	2	1	1	1		1		
<i>Trachyspenia australis</i>		1										
TOTAL MARINE	13	11	8	16	15	8	9	9	14	15	9	5
COUNTS												
F-1	10	8	23	13	16	6	15	16	20	21	23	14
F-2	8	12	22	14	13	10	21	19	14	34	15	15
F-3	11	11	19	16	20	7	19	26	19	27	17	13
F-4	11	12	14	17	12	13	22	22	18	28	18	11
F-5	14	18	20	16	14	9	18	27	16	23	22	23
F-6	12	14	17	12	18	8	17	15	20	24	20	18
F-7	11	16	11	22	8	11	14	30	15	23	19	14
F-8	19	19	14	17	12	10	17	23	22	21	11	15
F-9	14	16	11	20	10	9	16	19	18	19	17	19
F-10	22	13	12	16	9	5	19	21	13	18	19	13
TOTAL COUNTED	132	139	163	163	132	88	178	218	175	238	181	155
Percent Marine	0.0985	0.0791	0.0491	0.0982	0.1136	0.0909	0.0506	0.0413	0.08	0.063	0.0497	0.0323

Marine Species	BRI-12	BRI-13	BRI-14	BRI-15	BRI-16	BRI-17	BRI-18	BRI-19	BRI-20	BRI-21	BRI-22	BRI-23
<i>Achnanthes brevipes</i>												
<i>Actinopteryx senarius</i>												
<i>Amphora proteus</i>												
<i>Biddulphia dubia</i>												
<i>Cocconeis scutellum</i>	1							1	1			1
<i>Coscinodiscus radiatus</i>												1
<i>Dephneis surirella</i>			1									
<i>Grammatophora oceanica</i>				1								
<i>Hyalodiscus scoticus</i>	1											1
<i>Paralia sulcata</i>	5	5	7	4				1	2			5
<i>Thalassiosira eccentrica</i>			1									
<i>Thalassiosira pacifica</i>				1				1				
<i>Trachyspenia australis</i>												1
TOTAL MARINE	7	5	9	6	0	0	0	3	3	0	0	9
COUNTS												
F-1	7	14	8	10				8	7			5
F-2	9	13	9	5				5	6			3
F-3	6	14	16	15				9	8			5
F-4	9	12	15	8				4	7			4
F-5	6	10	7	12				4	6			2
F-6	4	6	13	8				7	4			2
F-7	9	18	12	10				9	7			8
F-8	11	15	11	15				3	6			3
F-9	5	16	14	6				4	3			4
F-10	7	9	12	9				3	6			4
TOTAL COUNTED	73	127	117	98	0	0	0	56	60	0	0	40
Percent Marine	0.0959	0.0394	0.0769	0.0612				0.0536	0.05			0.225

Table 9. χ^2 test for homogeneity and z-parameter for 0-25 cm of core 44.

Statistical Tests of Diatom Populations from Core Site 44

Null Hypothesis (Ho): For strata that do not host detritus or sand (peat), and upper contacts between peat and sand/detritus, the proportion of marine diatoms is the same.

The upper peat-sand contact occurs at 22-23 cm (see figure 17).

CHI-Squared Test of Homogeneity of Diatoms from North Briscoe Core

Observed and diatom values

	Stratigraphic				
	Peat (0-22 cm)		Upper Peat-Sand Contact		
Salinity Division	Observed	Expected	Observed	Expected	Row Total
Marine	165	171	9	3	174
Fresh and Brackish	2493	2487	40	3	2533
Column Total	2658		49		2707

Chi-Squared Test

	Stratigraphic Setting		
	Peat (0-22 cm)	Upper Peat-Sand Contact	
Salinity Division	$\chi^2(I)$	$\chi^2(I)$	χ^2
Marine	0.2	11	
Fresh and Brackish	0.0	431	442

Reject Ho if $\chi^2 > 3.84$ (based on 5% confidence limit and 1 degree of freedom)

Z-Test for Difference of Two Proportions

North Briscoe Core

Reject Ho if Z>1.96

Stratigraphic Section Compared	Statistical Parameters					
	x1	x2	p1	p2	n1	n2
Peat (0-22 cm) to Peat-Sand Contact (22-23cm)	165	9	0.95	0.05	2493	40

Test Results

$\hat{p}$	$\hat{q}$	Z	Result
0.07	0.93	22.2	Reject Ho

Chemostratigraphy

Instrumental Neutron Activation Analysis (INAA) of sediments from core 44 showed increases in abundance of marine geochemical markers which are correlative to the increases in marine diatom abundance. A bromine peak occurred in core 44 at 25 cm below sediment surface (Figure 18). Analyses of numerous trace elements were completed, but the element of primary interest to this study is bromine. Bromine

is a trace element common in marine water and is not supplied from terrigenous sources (Duce *et al.*, 1965, Fuge, 1988). Core 11 showed increases in bromine at 8-10 cm, 35-37 cm, and 68-70 cm below the sediment surface, indicating three marine water incursions (Figure 19). In addition, the bromine is well-correlated to the presence of woody detritus in the horizon from 0-10 cm ($r = 0.9$), and in the horizon from 36 cm to 70 cm ($r = 0.7$). The bromine concentration is strongly reverse correlated to the presence of detritus in the horizon from 12 cm to 36 cm ($r = -0.9$). Results of the initial INAA experiments are included in Appendix G.

Figure 18. Bromine concentration as it relates to lithology for core 44.

Figure 19. Bromine concentration as it relates to lithology for core 11.

Figure 20. Combined core log, marine diatom percentage and bromine concentration for cores 11 (above) and 44. For the core from site 44, the three data sets show a correlation factor of 0.65.

Stratigraphic Correlations and Resulting Minimum Inundation Distances

Two core transects from the central portion of the Long Beach Peninsula will be presented to highlight the stratigraphic correlation that exists between the inferred tsunami sand and detritus layers that were characterized. The results of a core transect orthogonal to the ocean shore across the central portion of the Long Beach Peninsula in the vicinity of Briscoe Lake are shown in a stratigraphic section (Figure 21). Three marine incursions are recorded in the cores from this transect. The upper-most tsunami horizon is observed in cores 44 and 11. A peat/detritus mixture sampled from core 44 at this horizon returned a radiocarbon age of 300-530 calYBP. The lithology at these strata consist of terrestrial, organic-rich detritus within peat for core 11, and a terrestrial detritus/sand mixture on top of a sand layer for core 44. Correlative marine diatoms and marine geochemical markers confirm a marine-forcing transport process for the anomalous deposits. The maximum observed inland extent of this strata (core 44) is situated approximately 300 m inland from the 300 B.P. subsurface erosion scarp (Woxell, 1998), and 800 m from the modern shoreline. The dune that separates cores 44 and 11 from the ocean has a current elevation of 8 m above mean sea level.

A stick sampled from above the anomalous sand layer at 64 cm below the sediment surface in core 58 returned a radiocarbon age of 665-795 cal YBP. The paleoshoreline can be constrained for this time frame, and therefore, this sand layer can be used to determine the minimum inland inundation distance of the marine surge that transported the sand. This sand layer is stratigraphically correlative in cores 11, 58, 59, 60, 61, 62 and 54 over a landward oriented, shoreline orthogonal distance of 700 m. The distance from the 1100 B.P. erosional scarp (Woxell, 1998) to the

Figure 21. Select core logs from a transect across the central Long Beach Peninsula. The map provides location of cores, subsurface erosional scarps (Meyers *et al.*, 1996), and current dune elevations.

maximum inland observed strata (core 54) for this horizon is approximately 1 km, and the distance from the modern shoreline to core 54 is 1.5 km. The layer thins in the

landward direction from a maximum thickness of 3 cm in core 58 to 0.3 cm in core 54.

A third sand horizon was also observed in cores 11, 58, 59, 60, 61, 62 and 54. A peat sample from above this layer in core 11 had a radiocarbon age of 990-1285 calYBP,

and a stick sampled from above this layer in core 61 had an age of 1065-1320 calYBP.

This layer is correlative over the same distances as the second anomalous horizon.

Figure 22 shows the stratigraphic results of four cores located in the vicinity of Tinker Lake which is situated landward of the municipality of Long Beach, Washington.

Cores 8, 55, 56 and 57 host a sand horizon consisting of fine-upper sand that is well-sorted and quartz-rich. The sand layer is capped with detrital wood, needle, and leaf fragments. A peat sample taken from above the sand layer in core 55 at a depth of 65 to 66 cm below sediment surface returned a radiocarbon age of $860 \pm \text{RCYBP}$. The most distal core (site 57) contained the thickest detrital cap (4 cm), suggesting that the bulk of the organic component of the inferred tsunami-entrained debris likely traveled further inland than the lithic component.

Using the paleoshoreline from previous work by Woxell (1998), a minimum inundation distance for the marine surge can be calculated for this vicinity. Core site 57 is located approximately 1-km inland from the current shoreline and approximately 0.5- km from the 300 and 1,100 year old erosional scarps. Current maximum dune elevation between the core sites and the shoreline is 8 m.

Grayland Plains, Washington

The location of the sites that were cored on the Grayland Plains is shown in Figure 23. Of the 27 sites that were cored, 100% contained inferred tsunami horizons. These included sites from core transects along Gould Road (Figure 13), Summer/Heather Lane, and in the vicinity of Twin Harbors Beach State Park. At least two, and possibly three sand depositional events are recorded.

Figure 23. Core sites for the Grayland Plains, central Washington coast.

Stratigraphic Correlations

The core location and stratigraphic results of a core transect along Gould Road, Grayland Plains, Washington are presented in Figure 24. Cores 169 and 170 were taken from the surface of a thickly wooded swamp. The remainder of the transect (cores 171-175) was sampled from within cranberry bogs. As a result, the upper 10-20 cm consisted of sand fill material. Cores 169 and 170, the two most proximal cores from this location, host a sand layer within 25 cm of the sediment surface. In both of these cores the layer consisted of a mixture of detritus, i.e. sticks, twigs, leaves, conifer cones, and well-sorted, quartz-rich sand. The layer in core 169 also contained two chunks of wood. Core 172 revealed a laminae of quartz-rich, well-sorted sand at a depth of 33 cm. Based on relative depth in the section, this layer is stratigraphically correlative to the sand layers in cores 169 and 170. An additional 2-cm thick sand layer was observed at 56 cm below sediment surface in core 172. This sand layer consisted of clean (e.g. little mud and detritus), well-sorted, quartz-rich sand. A stick sampled from the top of this sand layer in core 172 indicated a radiocarbon age of 655-880 calYBP. Again, based on relative depths of this sand layer, it is stratigraphically correlative with similar sand layers observed in cores 171 and 173. A thin layer of detritus was observed directly above this layer in cores 172 and 173. The upper sand layers from cores 169, 170 and 172 are inferred to have been deposited by the tsunami that occurred during the last great Cascadia earthquake at 1700 AD, based on the radiocarbon date of 250 calYBP (Yamaguchi *et al.*, 1997). The lower sand layers observed in cores 171, 172 and 173 are also inferred to be tsunamigenic and

Figure 24. Core logs from a shoreline orthogonal transect for the Grayland Plains, central Washington coast.

have a date of 850 cal YBP. Anomalous terrestrial organic-rich detritus and sand layers were observed at much deeper in cores 174 and 175. Given the relatively deep depth of these layers, it is presently unclear how these layers correlate to the dated horizons to the west. However, basal peat ages from this area are about 1000 cal YBP (Woxell, 1999), so these lower horizons might correlate to the 800 year old horizons to the west.

Inundation distance can be roughly constrained by the observation that core site 172 is located approximately 1.5 km inland from the modern shoreline. Minimum tsunami amplitude can be constrained based on the fact that modern dunes that separate the ocean from the core sites have a maximum elevation of 6 m. Although, paleodune analysis and soil chronosequencing has yet to be completed for the Grayland Plains, it is known that the largest and highest dune-ridge was in place during the 1700 AD earthquake (Woxell, 1998). The 1700 AD erosion scarp is mapped at the western flank or directly under the big dune ridge, about 0.5 km from the modern shoreline. A second scarp thought to correlate to either the 1,100 or 1,300 ya earthquake is located on the eastern flank of the big dune ridge. Therefore, the paleotsunami inundation is at least 1 km from the pre-existing shoreline.

Neskowin, Oregon

The locations of the core sites near Neskowin, Oregon are shown in Figure 25. Of the 43 sites that were cored, 100% contained inferred tsunamigenic horizons. Four shore-normal (west to east) transects and one north to south core transect were drilled. One west-east core transect and the north-south core transect were located in a marsh

in the northern region of the Neskowin. Two core transects were directed across the central region of the Neskowin dune complex. The third west to east transect occurred

Figure 25. Core sites for Neskowin, Oregon.

in the southern portion of the Neskowin dune complex, where it is situated in the vicinity of the commercial and residential sector of Neskowin. Sand grain size

analyses were completed for sand layers from the north marsh and central marsh transects to test for fining-up vertical and fining-inland lateral facies changes.

Neskowin Sand Grain Analysis

Grain size analyses for sand samples from Neskowin are presented in Figure 26 and 27. The sand layer at depth 63-74 cm from core 153 (Figure 26) was sampled for sand grain size analysis. The sand from this horizon was split into an upper and lower fraction. Each fraction was sieved, and the results are plotted as a cumulative percentage vs. grain size in mm (Figure 26). The upper fraction (63-71 cm) had a mean grain size of 0.221 mm, a median of 0.22, and a standard deviation of 0.0385 mm. The lower fraction (71-74 cm) had a mean grain size of .239 mm, a median of .239 mm and a standard deviation of .044 mm. The standard deviations of both samples are representative of a high degree of sorting (Folk, 1980). A t-test was performed on the two populations to test for the equivalency of the mean grain sizes of the sample (Table 9). The null hypothesis (H_0) and alternative are shown below for this test:

H_0 : The mean grain size from the 63-71 cm fraction is representative of the same grain size population as the mean grain size from the 71-74 cm fraction.

H_1 : The mean grain size from the 63-71 cm fraction is representative of the same mean grain size population as the mean grain size from the 71-74 cm fraction.

The t-test results gave a t-statistic of 3.26. For a two-tailed test with 7 degrees of freedom t-critical is 1.9. Given that the t-statistics is outside the value for t-critical, the null hypothesis can be rejected. Based on these results, it is concluded that the sand layer from 63-74 cm in core 153 shows a fining up sequence.

Nesk-10-980707

Core Site - 153

Sampled Horizon = 71-74 cm

Total Mass = 178.16 g

Median = .239 mm (2.08 ϕ)

D90 = 1.76 (.29 mm)

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.080	3.64	5.7	7.11	1.41	0.79	0.79
140	0.106	3.24	4.88	8.35	3.47	1.95	2.74
100	0.150	2.74	5.55	9.83	4.28	2.40	5.14
80	0.180	2.47	5.71	14.64	8.93	5.01	10.15
65	0.208	2.27	5.07	28.49	23.42	13.15	23.30
60	0.250	2.00	5.68	70.85	65.17	36.58	59.88
50	0.300	1.74	5.55	63.88	58.33	32.74	92.62
45	0.355	1.49	5.64	17.89	12.25	6.88	99.49
		1.00					100

D16 in mm = 0.195 (2.36 ϕ)Mean: 0.239 mm (2.08 ϕ)D50 in mm = 0.239 (2.08 ϕ)Standard Dev (s_1): 0.044 mm (0.27 ϕ)D84 in mm = 0.283 (1.82 ϕ)

Core Log

Nesk-10-980707

Core Site 153

Sampled Horizon = 63-71 cm

Total Mass = 71.15 g

Median = .22 mm (2.18 ϕ)

D90 = 1.86

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.080	3.64	5.7	6.22	0.52	0.73	0.73
140	0.106	3.24	4.88	6.46	1.58	2.22	2.95
100	0.150	2.74	5.55	7.74	2.19	3.08	6.03
80	0.180	2.47	5.71	11.20	5.49	7.72	13.75
65	0.208	2.27	5.07	19.92	14.85	20.87	34.62
60	0.250	2.00	5.68	37.14	31.46	44.22	78.83
50	0.300	1.74	5.55	17.05	11.50	16.16	95.00
45	0.355	1.49	5.64	8.30	2.66	3.74	98.74
		1.00					100

D16 in mm = 0.183 (2.45 ϕ)Mean: 0.221 mm (2.17 ϕ)D50 in mm = 0.22 (2.18 ϕ)Standard Dev (s_1): 0.0385 mm (.255 ϕ)D84 in mm = 0.26 (1.94 ϕ)

Figure 26. Sieving data and cumulative frequency plots for upper and lower fractions from a tsunamigenic sand layer, core site 153, showing fining up of grain size.

An F-test was performed of the two populations to test for the equivalency of the variance of the samples (Table 10). For this test the null hypothesis and alternative hypotheses are:

Ho: The variance in grain size for the 71-74 cm sand fraction is the same as the variance in grain size for the 63-71 cm sand fraction.

H1: The variance in grain size for the 71-74 cm sand fraction is not the same as the variance in grain size for the 63-71 cm sand fraction.

The F-test results gave an F-statistic of 1.31. Each sampled population represents 7 degrees of freedom, which results in an F-critical of 3.79 with a confidence of 5%. Given that the F-statistic falls within the range of F-critical, the null hypothesis cannot be rejected. Therefore, based on the F-test, the source and mechanisms of transport are similar for the upper and lower fractions.

Similar results were shown from a sand layer at depth 57-71 from core site 159 (Appendix H). Whereas fining up sequences were noted qualitatively while coring in the field, these sieving data provide quantitative verification of these observations. These data also suggest sand deposition at these sites occur by fluid transport of sufficiently high velocity to entrain and maintain sand in suspension over the distance of observed landward transport.

Sand grain size analysis was also performed on correlative sand layers to quantify landward-fining of the sand sheets. The sand layer at depth 63-74 cm below the surface from core 153 returned a mean grain size of 0.233 mm. The correlative sand layer from the landward core located at core site 141 (56-60 cm below surface) had a mean grain size of 0.226. The cumulative percentage plot of the sieving data

from each of the sand layers is shown in Figure 27. The median grain size for the layer from core 141 is 0.226 mm, and for the correlative layer from core 153 the median grain size is 0.233. Given the near identical sand grain distributions from each sand layer, a definitive conclusion about landward fining of the sand sheets cannot be discerned over these inundation distances of 300 meters. Similar results were obtained when comparing correlative sand layers from sites 156 and 159 (Appendix H).

Table 10. F- and t-tests for upper and lower sand fractions from 63-74 cm sand horizon, core 153.

t-test		Statistical Equations (D values were obtained graphically from Figure 26)		
$\bar{X}_1 = 0.239$ mm		Mean ($\bar{X}$) = (D16+D50+D84)/3		
$\bar{X}_2 = 0.221$ mm		Standard Dev. (s) = (D84-D16)/2		
$s_1 = 0.044$ mm		Median = D50		
$s_2 = 0.0385$ mm				
$s_p = 0.011048998$				
$s_e = 0.005524499$				
$t = 3.26$				
t-critical= 1.9 ($\alpha = .05$, 7 degrees of freedom)				
F-test				
F= 1.31				
F-critical= 3.79 ($\alpha = .05$, 7 degrees of freedom)				

Neskowin Stratigraphic Correlations

Figure 28 presents the core location and stratigraphic results of a representative transect across the north marsh, Neskowin, Oregon. Four additional core transects were completed for the Neskowin area and are detailed in Appendix E. Cores 153 and 141, the two most proximal cores from this location, contain an anomalous sand layer within 30 cm of the sediment surface. A sample of peat extracted from above this layer in core 153 returned a modern ^{14}C date. Two additional sand layers were observed in most of the cores that were sampled from the north marsh area at Neskowin. The second layer in core 153 was easily correlative across the marsh by

the presence of a distinctive underlying dark brown mud layer. The third sand layer in core 153 is also easily correlated across the marsh. A sample of peat from below this layer in core 153 returned a ^{14}C age of 920-1290 calYBP. The uppermost sand layers in cores 153 and 141 are inferred to have been deposited by the tsunami that occurred

Nesk-10-980707 Map ID - 153

Analyzed Horizon = 63-74 cm

Total Mass = 248.41 g

Median = 2.1

D90 = 1.78

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
Fines	0.080	3.64	5.7	13.33	1.93	0.78	0.78
140	0.106	3.24	4.88	14.81	5.05	2.03	2.81
100	0.150	2.74	5.55	17.57	6.47	2.60	5.41
80	0.180	2.47	5.71	25.84	14.42	5.80	11.22
65	0.208	2.27	5.07	48.41	38.27	15.41	26.63
60	0.250	2.00	5.68	107.99	96.63	38.90	65.52
50	0.300	1.74	5.55	80.93	69.83	28.11	93.64
45	0.355	1.49	5.64	26.19	14.91	6.00	99.64
		1.00					100

Nesk-9-980707 Map ID - 141

56-60 cm

Total Mass = 134.25 g

Median = 2.14

D90 = 1.77

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
Fines	0.080	3.64	5.7	7.82	2.12	1.58	1.58
140	0.106	3.24	4.88	7.79	2.91	2.17	3.75
100	0.150	2.74	5.55	8.7	3.15	2.35	6.09
80	0.180	2.47	5.71	14.56	8.85	6.59	12.69
65	0.208	2.27	5.07	24.69	19.62	14.61	27.30
60	0.250	2.00	5.68	62.1	56.42	42.03	69.33
50	0.300	1.74	5.55	36.58	31.03	23.11	92.44
45	0.355	1.49	5.64	14.89	9.25	6.89	99.33
		1.00					100.00

Figure 27. Sand grain size distribution for correlative sand layers from two cores at Neskowin. The near identical distributions indicate that landward fining of grain size did not occur between cores 153 (more proximal to shoreline) and 141. Core 141 was located approximately 50 meters landward of core 153.

Core Logs

Figure 28. Inland-oriented core transect from the marsh located north of Neskowin, Oregon.

during the 1700 AD earthquake. The lower sand layers observed in many of the cores are inferred to have been deposited by pre-1700 AD tsunami events.

Inundation distance for the 1700 AD tsunami can be approximated by the observation that core 141 is located approximately 300 m inland from the modern shoreline. The minimum distance from the modern shoreline of the additional two sand layers observed in all of the cores from the north marsh transect is at least 600 m (Figure 24). The aerial extent of these sand layers shows an inland-oriented distance of 250 m, which establishes a minimum inundation distance for the tsunami that deposited these layers. Sand layers corresponding to marine surges prior to 1700 AD have been mapped up to 1.0-km inland from the modern shoreline in the central marsh region of Neskowin (Appendix E). Future work on establishing the sand spit evolution at Neskowin will more tightly constrain these inferred inundation distances. Of particular interest is the maximum inundation of tsunami as recorded by deposits at Neskowin.

Rockaway, Oregon

The locations of the sites that were cored in the vicinity of Rockaway are shown in Figure 29. Of the 30 sites which were cored, 100% hosted inferred tsunamigenic horizons. These included sites from transects across Crescent Lake, Lake Lytle, and along the canal connecting the two lakes that back the sand dune barrier at Rockaway. Vibracores collected from Lake Lytle and Crescent Lake record up to six anomalous sand layers hosted in soft, organic-rich mud. The most proximal core from Crescent Lake records at least five, and possibly seven inferred tsunami

Figure 29. Map and core locations for Rockaway, Oregon.

inundation events. Terrestrial, organic-rich detritus overlying the lowest sand layer in this core (core 123) has a ^{14}C date of 2000-2325 calYBP. A detailed log of a proximal core from Rockaway is shown in Figure 11.

Stratigraphic Correlations

A robust sand layer was traced along the canal that connects Lake Lytle and Crescent Lake (core sites 91-94, Figure 30). Peat sampled from beneath this layer in core 91 returned a ^{14}C age of 735-795 calYBP, suggesting that the layer was deposited by an event prior to the 1700 AD earthquake. An additional robust sand layer was traced along a transect across the peninsula which extends into the southern end of Crescent Lake (core sites 108-114, Figure 29). Peat sampled from beneath this layer in core 112 showed a ^{14}C age of 840-975 calYBP, again suggesting that the layer was deposited by an event prior to the 1700 AD earthquake. Thin sand layers were observed in the upper 30 cm of cores 108 and 109 that probably correlate to the AD 1700 event.

Liquefaction features, i.e. intruded contacts and sand dikes (see figure 11), were prevalent in the cores extracted from the bottom of Lake Lytle which consisted of soft, poorly compacted, organic rich mud with a gelatinous texture. Some of the sand layers showed detrital caps and intra-layer detrital laminae. It is unclear whether some of the sand laminae are inundation deposits or represent liquefaction sills (Figure 31). Based on the ^{14}C date of 550-685 calYBP, and the associated liquefaction features for the uppermost sand layer for core 77, it is inferred that this layer was deposited prior to the 1700 AD earthquake. Its possible correlation to uppermost sand

Figure 30. Stratigraphic results of a core transect along the canal adjoining Lake Lytle and Crescent Lake in the vicinity of Rockaway, Oregon.

layers in other cores from Lake Lytle will require closer spaced vibracoring for confirmation.

Figure 31. Representative cores from Lake Lytle, Rockaway, Oregon. The prevalence of liquefaction features makes correlation of sand layers difficult.

Summary of Results

Table 11 represents a summary of the sand layers which were mapped at the various sites and their inferred correlation to Great Cascadia Earthquakes as outlined in the literature (Atwater and Hemphill-Haley, 1996, Darienzo and Peterson, 1995).

Table 11. Summary of results for sand layers that were characterized at various locations.

*Earthquake Date	Long Beach			Grayland		
	Number of cores where it was observed	Thickness Range	** Observed Inland Inundation Distance	Number of cores where it was observed	Thickness Range	** Observed Inland Inundation Distance
1700 AD	5, mostly as detritus layers	1-4 cm	700 m	14	.3 - 10 cm	1.0 km
800 YBP	11	1-3 cm	1.3 km	6	.2-5 cm	1.5 km
1100 YBP	5	1-3 cm	1.3 km	No dates to verify		
1300 YBP	1	.5 cm	1.3 km	No dates to verify		
1700 YBP						
2500 YBP						

*Earthquake Date	Neskowin			Rockaway		
	Number of cores where it was observed	Thickness Range	** Observed Inland Inundation Distance	Number of cores where it was observed	Thickness Range	** Observed Inland Inundation Distance
1700 AD	15	.2-15 cm	300 m	11	.2-20 cm	200 m
800 YBP	23	.2-20 cm	1.0 km	28	.2-25 cm	500 m
1100 YBP	13	.2-8 cm	1.0 km	12	2-25 cm	400 m
1300 YBP	?			2	5-10 cm	200 m
1700 YBP	?			2	2-4 cm	200 m
2500 YBP				1	2 cm	200 m

*Earthquake date is taken from figure 4, which was adapted from the literature.

** This distance is measured from the modern shoreline. Paleoshoreline position is unknown.

Discussion

Calling a Sand Layer a Tsunami Layer

The study sites are all in pond or wetland depositional settings. Biogenic (peat), organic (gyttja), and fine-grained sediments are the primary facies in these environments. Therefore, the presence of well-sorted, medium-grained, quartz-rich sand in these strata is considered to be anomalous. Based on Hjulstrom's erosion and sedimentation velocity curves (Pye, 1994), flow velocities of 20 cm/s are needed to entrain sand particles. Flow velocities of at least 2 cm/s are necessary to continue to transport the particles. In light of the Hjulstrom curve, these anomalous layers are inferred to have been transported into these quiet water settings by high-velocity hydrodynamic processes. Clearly, a marine surge mechanism picked up well-sorted, quartz-rich sand off the nearshore, beach, or dune ridges and carried it inland. Admittedly, when analyzing a core from a location within a few hundred meters of the beach, it is difficult to definitively conclude that the anomalous sand layer is tsunamigenic. On some coastlines, a storm surge could have also deposited such a sand layer. Anomalous sand layers are observed in southeastern U.S. coastal strata that have been deposited by hurricane surges. However, when analyzing a suite of cores within the regional context of the PNW, the possibility of a storm surge depositional process can begin to be ruled out.

In tidal wetlands along the entire length of the Cascadia Margin, subsided wetland horizons have been documented, radiocarbon dated, and correlated. These regional subsidence events are concluded to be coseismically induced (Atwater, 1987,

1992; Darienzo and Peterson, 1990). Subsided horizons have been documented throughout the Cascadia Margin by a number of researchers (Clarke and Carver, 1992, Darienzo *et al.*, 1994, Peterson and Darienzo, 1996, Atwater, 1987, 1992).

Anomalous sand layers were observed to rest immediately above a number of these subsided soils, and are concluded to have been deposited by tsunami that were generated by large Cascadia subduction-zone earthquakes. It is important to note that the overlying sand layers are closely associated with the coseismically subsided horizons in the tidal basins, and are rare elsewhere in the strata (Atwater and Hemphill-Haley, 1997, Darienzo and Peterson, 1990). It is possible that storm surges occurred concomitant with Cascadia Earthquake subsidence and deposited sand onto recently submerged soils. However, given the synchronicity of tsunami with subduction zone earthquakes throughout the Pacific Ocean basin, and that sand layers are directly observed to be transported inland by these tsunami (Minoura *et al.*, 1994, Jaffe *et al.*, 1998b, 1999) a strong argument can be made that the sand layers observed above the Cascadia subsided soils are tsunami deposited. The rarity of sand layers elsewhere in these cores, combined with the fact that storm surge overwashing has not been noted by historic observation for the PNW coast, supports the argument that storm-surges are not a feasible mechanism for inland sand transport of several hundred meters on this coast. Tsunami generated by Cascadia Subduction Zone earthquakes probably play the major role in the inland transport and deposition of anomalous sand layers observed in this study.

Researchers at various field sites along the Cascadia Margin where tidal basin subsidence evidence is prevalent typically characterize at least three of these cycles,

and in some cases have documented up to six during the last three thousand years (Figure 4). The sand and detritus layers observed and mapped for this study follow the same pattern (Figure 32). It is within this regional framework, as outlined in the literature regarding PNW paleoseismicity, that many of the sand and detritus layers observed at the study sites are inferred to be tsunamigenic.

Validating a Marine Source for the Targeted Incursions

Anomalous sand sheets are known to have been derived from the beach-barrier or nearshore based on sedimentological characteristics observed in the field (see core logs in Appendices B-E). These characteristics include a quartz-rich, well-sorted, subrounded sand component with a mean grain size of .177-.250 mm. Two cores from Briscoe Lake on the Long Beach Peninsula were scanned for microfossil assemblages to validate a marine source of the sand sheets. In each core, prevalence of marine diatoms *Paralia sulcata*, *Amphora proteus*, *Thalassiosira pacifica*, and *Cocconeis scutellum* increased at each of the anomalous inundation horizons (Figures 15 and 17). The results presented in Figure 11 indicated that increased levels of marine diatoms do not occur within a narrow stratigraphic horizon, as is typically reported in previous work done by Hemphill-Haley (Atwater and Hemphill-Haley, 1996), and Darienzo and Peterson (1990). Figure 15 indicates that marine diatoms persist in the strata, suggesting that 1) saline conditions may prevail for decades following a marine surge, 2) subsidence induced shoreline retreat permitted salt spray to reach the interior bog sites, and or 3) subsidence of 1-2 m could cause subsurface marine water intrusion into the ponds prior to interseismic rebound.

Figure 32. Representative core logs from each of the study sites. The regional characteristic of long time gaps (a few centuries) between anomalous sand layers, the probable regional correlation of the sand layers, and the evidence of multiple surges for some of the sand layers (core 159 and 123) suggests a tsunami surge depositional mechanism.

For this study, diatoms from the strata of coastal, land-locked lakes - e.g. interdune basin features were investigated. Low rates of water replacement for these features are demonstrated by 1) small size and absence of surface tributaries, 2) the absence of drainage outlets, 3) water level response to tidal cycles that is muted or nonexistent. Hence, the feasibility of scanning cores for the presence of marine geochemical markers (bromine) was also tested in two cores from Long Beach. Bromine was selected as the primary marine tracer because it is enriched in ocean water (Fuge, 1988). Bromine is accurately analyzed for bulk concentration using INAA techniques (Muecke, 1980). As a biophilic element (Fuge, 1988) it readily leaves a record of marine incursion into organic-rich terrestrial sites. The abrupt increase in bromine concentration correlative with sand and detritus layers is due to direct flooding of marine water into freshwater wetland sites, similar to that observed during the 1983 tsunami of the Japan Sea Earthquake (Minoura and Nakaya, 1991). However, bromine is also transferred from marine to terrestrial environments via the atmosphere, with higher concentrations observed closer to the coast (Duce *et al.*, 1965). The increase in bromine content in the strata of Briscoe Lake could record the changing proximity of the shore due to subsidence-induced erosion (Doyle, 1996). Subsequent accretion over the next century, a dynamic that has been well established for the Long Beach Peninsula (Woxell, 1998), could then increase the shoreline distance from the freshwater bog, thereby decreasing salt spray. Specifically, the initial peak of bromine observed in Figure 14 represents a marine surge origin, followed by a prolonged but decreasing bromine level. The gradual decrease in bromine content might reflect the prograding shoreline during interseismic periods.

However, the gradual decrease in bromine might also indicate a residence time within the lake for the bromine, similar to what has been suggested for marine diatoms above. As a highly soluble ion, bromine probably does not readily precipitate out of solution, but is incorporated into biomass first, and then deposited biogenically (Fuge, 1988). It is possible that the time lag in bromine peaks, especially prevalent in Figure 19, represents the slower rate of deposition typical for biogenic sources.

Additional work is needed to establish the mechanisms of sustained marine marker presence in the post-event core records. For example, similar diatom and bromine analyses need to be conducted in sites at the zero isobase where no shoreline retreat occurs following tsunami inundation. In addition, marine inundation markers should be searched-out in distal settings i.e., greater than 1 km from the shoreline to establish the maximum low velocity inundation distances of the paleotsunami.

It is worth noting that in each core from Briscoe Lake that was analyzed for diatoms, distinct sand layers were not visible. Marine incursions were inferred from increased marine geochemical markers and marine diatom abundance. Subsequent visual inspection of the cores revealed the presence of terrestrial detritus at each of these horizons - small sticks, twigs, and needles, which are prevalent in the duff of forested areas. The detritus was entrained and redeposited as the marine surge flowed through the forest which abuts the shoreline on the Long Beach Peninsula. These observations confirm the utility of diatom assemblage studies to map the inland extent of tsunami inundation, as proposed by Hemphill-Haley (1996). Analyzing for marine geochemical markers might also provide a less time intensive alternative to performing the diatom assemblage analyses.

Implications for Understanding the Past 3 (or 4?) Cascadia Tsunami

The Elusive 1700 AD Event

The great ($M_w > 8.5$) Cascadia Subduction Zone earthquake that is reported to have occurred in 1700 AD (Yamaguchi, *et al.*, 1997, Satake, *et al.*, 1996) resulted in 1-2 m of coseismic subsidence (Atwater, 1987, Darienzo and Peterson, 1990) in northern Oregon and southern Washington. The magnitude for this earthquake has been estimated to be at least $M_w \sim 9$ (Satake, *et al.*, 1996). Recent computer models of potential tsunami have resulted in inundation elevations ranging from 3-10 m and velocities of 2-4 m/s for much of the central Cascadia Margin (Myers, *et al.*, 1999). Based on the extreme catastrophic nature of these predicted inundation scenarios, as presented in the literature, one could easily conclude that the tsunami that occurred with the 1700 AD earthquake must have also been extremely large. However, the results presented in this paper indicate that the 1700 AD tsunami is smaller than the previous 3-5 tsunami events. Of the 110 cores that showed tsunami sand deposits, only 8 are inferred to include a record of the 1700 AD event. This observation is most illustrative at Long Beach, where the 1700 AD event is marked by anomalous detritus layers in only two cores (Figure 29). Further, there were no anomalous sand layers observed on the Long Beach Peninsula which could be attributed to the AD 1700 event. The weak nature of the 1700 AD tsunami record has been observed at Seaside, Oregon (Fiedorowicz, 1997) and at Cannon Beach, Oregon (Peterson, pers com., 1999).

The elusive nature of the 1700 AD tsunami deposits are highlighted to emphasize an important point as it pertains to tsunami hazard mitigation (Figure 33).

At present, the Cascadia tsunami excitation parameters are not sufficiently constrained to accurately predict nearfield tsunami runup from computer modeling. More specifically, the assumed nature of the fault dislocation and subsequent seismic release influences the respective magnitude of tsunami excitation and predicted runup (Kanamori and Kikuchi, 1993, Tanioka and Satake, 1996). Further, the disconnection between the magnitude of a predicted Cascadia Subduction Zone earthquake and the impact of the resulting tsunami can be exemplified by the fact that larger than expected tsunami can be generated from earthquake-induced submarine landslides (Adams, 1990). Smaller than expected inundation can occur if tide levels are at an extreme low (Hutchinson et. al, 1997). The presence of a dense forest near the shoreline, such as the Long Beach Peninsula, can reduce the inundation potential of the tsunami (Hutchinson *et al.*, 1997). Local features such as a broad, shallow continental shelf or large and adjacent river and bay mouths can attenuate tsunami energy (Myers *et al.*, 1999). By comparison, the thickness and sediment characteristics of a tsunamigenic sand sheet do provide a constraint on tsunami runup (velocity and magnitude) and, therefore, hazard potential (Abramson, 1998).

Due to variability of rupture mechanics, tides, and paleoshoreline conditions, a number of paleotsunami inundation records should be analyzed for representative

Figure 33. Representative core logs for proximal sites at the Long Beach Peninsula showing the ephemeral nature of the tsunami deposits, particularly for the 1700 AD event, which is apparent by the presence of terrestrial detrital layers and confirmed via diatom assemblage and geochemical marker studies of the cores (see Figure 16).

localities. A deterministic hazard analysis could be based on the maximum inundation event in the geologic record. Alternatively, a frequency weighted distribution of inundation magnitudes can be analyzed for a probabilistic hazard analysis. Additional work is warranted to extend these paleotsunami inundation records – i.e. longer time intervals at more localities, to address these statistical hazard analyses.

Pre 1700 AD Tsunami – Part 1: Evidence in Support of an 800 ya Cascadia Subduction Zone Earthquake

At least two sand layers that pre-date the 1700 AD event were observed and described at all of the study sites. Cores recording up to six anomalous sand sheets

were extracted from the bottoms of Crescent Lake and Lake Lytle at Rockaway, Oregon. The second sand layer in the strata, which will be referred to as sand layer B, was observed and ^{14}C dated in numerous cores from all of the study sites (Figure 34). Two peat samples directly beneath a broadly correlated sand layer from Crescent Lake had an age of 735-795 cal YBP and 840-975 cal YBP respectively. A terrestrial detritus layer sampled above a sand layer from Tinker Lake (Figure 22) had an age of 675-915 cal YBP. A bark-free stick which was overlying a sand layer from the southern portion of Briscoe Lake (Figure 17) on the Long Beach Peninsula corresponded to a date of 665-795 cal YBP. A bark-free twig overlying a sand layer at Grayland, which extended inland for approximately 0.8 km, had an age of 655-880 cal YBP (Figure 24). A peat sample from beneath a sand layer at Neskowin had an age of 770-965 cal YBP. These various date ranges bracket an age for this sand layer of 740-860 cal YBP (Figure 34). This age range represents one standard deviation about the mean of all the samples. A well-established coseismic subsidence record for this time frame is not reported for northernmost Oregon or Washington.

A tsunami deposit with a similar age (500-800 cal YBP) has been reported in the literature from the outer coast of Vancouver Island by Clague and Bobrowsky (1994). These authors speculated that the paleotsunami was from a far-field

Figure 34. Radiocarbon age ranges for various samples from the four study sites (bold lines) which bracket an age of approximately 800 cal YBP. The hatched stripe represents 1 standard deviation about the mean for all of the samples shown.

earthquake in Alaska, a submarine landslide, or a previously unrecognized Cascadia earthquake. Subsidence horizons corresponding to this age have been characterized by Clarke and Carver (1992) along the southern Cascadia margin, and Peterson and Darienzo (1996) along the central Oregon coast. Further, Atwater and Hemphill-Haley (1997) reported two ages in this range for sticks sampled from what were inferred to be tsunami deposits in channel fill from the Niawiakum River in southern coastal Washington. Descriptions of this event from northern California and southern Washington suggest that the subsidence record is weak and discontinuous, possibly indicating an origin from local faulting. However, the ^{14}C results presented in this thesis for sand layer B argue for a nearfield tsunami that occurred 740-860 years ago. This event is recorded by tsunami deposits along the full length of the margin. Based

on the thickness (5 cm) and extent of the deposits (200 km), the tsunami is inferred to have been produced by a Cascadia rupture. Additional tsunami and subsidence mapping is needed to verify this hypothesis.

Pre 1700 AD Tsunami – Part 2: The Large Tsunami of 1100-1350 ya

A third sand layer was observed at many of the study sites, which will be referred to as sand layer C. The ^{14}C ages that correspond to this layer come from a sample of peat underlying a sand layer at Neskowin with an age of 920-1290 cal YBP, a sample of peat and detritus overlying a sand layer at Long Beach with an age of 990-1285 cal YBP, and some bits of wood overlying a sand layer at Long Beach with an age of 1280-1410 cal YBP. It is unclear from the ^{14}C data whether this sand layer at the various study sites is correlative, and if so, whether it records one event (~1150 years ago) or two events (~1150 and ~1300 years ago) during this time frame. Both ~1150 and ~1300 year ago subsidence events have been documented from studies of buried wetlands in Washington (Atwater, 1987), and coseismic subsidence horizons in California (Clarke and Carver, 1992), and Oregon (Darienzo and Peterson, 1990). The C sand layers are characterized by thick deposits at most locations, implying severe tsunami impact to the shoreline. Proximal cores at Rockaway and Neskowin record thicknesses for sand layer C of 30-40 cm. In addition, the layers at this horizon are widespread at each of the four study localities. For example, it was mapped inland for a distance of 1.5 km at Neskowin, where it thinned from a maximum shoreline proximal thickness of 45 cm to a thickness 3 cm, and at its landward edge it interlayered with colluvium along the eastern margin of the marsh

Figure 35. Stratigraphic correlations for sand sheets at Neskowin, Oregon (see Figure 25 for locations). Sand layers B and C are noteworthy in that they are robust, and thin in the landward direction. Sand layer C is inferred to have been deposited 1100-1300 years ago.

(Figure 23). When contrasted with the weak record from the 1700 AD earthquake, the robust nature of sand layer C clearly raises numerous questions about the variability of rupture mechanics and associated tsunami excitation in the Cascadia margin. For example, was layer C produced by a larger magnitude earthquake than the AD 1700 event? Or were local upper-plate fault ruptures triggered by the megathrust fault responsible for the anomalously large tsunami inundation? Perhaps a large slump block on the continental slope was activated by the great earthquake, thereby adding to the megathrust excitation of the paleotsunami. Additional work is needed to characterize this large tsunami event in central and southern Oregon. This will help to constrain possible rupture scenarios. To conclude, the results of this study highlight some important considerations as they pertain to hazard mitigation. For example: 1) paleotsunami vary widely in inundation magnitude in the central Cascadia margin, 2) existing rupture scenarios and associated computer models of runup do not match the geologic record of paleotsunami magnitude, and 3) local conditions of paleodune ridge topography, forest cover, and pond extent affect local inundation distances. These influences on tsunami inundation are not addressed in overland runup modeling.

Mapping Tsunami Deposits in Time Transgressive Settings

Identifying tsunami deposits is dependant upon their preservation within a contrasting sediment. For this study these included peat, inorganic mud, and organic-rich mud. One of the conclusions of this study is that the most useful hosting sediment was peat. The tightly interwoven nature of the fibers minimized alteration of the sand sheets via liquefaction processes – the peat held the sand in place from

subsequent earthquake-induced ground shaking, which provided an undisturbed record of sand deposition. However, peat deposition is dependant upon water levels. The oxidizing conditions of upland soils prevents peat accumulation. Conversely, lakes which continuously exceed water depths of 0.5 meter do not allow for rooted plant growth which is necessary for peat development. As a result, understanding the timing and control of peat development for a given location is important to correlating sand layers.

The time transgressive nature of Cascadia Margin coastal sites involves lateral and vertical relationships - shoreline accretion/erosion and interseismic uplift/coseismic subsidence. As expected from Walther's Law of correlation of facies, the two are interrelated. The mapping of placer-rich, subsurface erosional scarps correlative with Cascadia paleoseismicity on the Long Beach Peninsula (Meyers *et al.*, 1996) exemplifies the punctuating, time transgressive effect that great earthquakes can have on shoreline development. The rate of the ensuing shoreline accretion will be related to sediment supply. In the case of the Long Beach Peninsula, a large sediment supply from the nearby Columbia River meant rapid interseismic shoreline progradations during the Holocene (Woxell, 1998). The interplay of vertical and lateral land level changes clearly affect freshwater levels, and therefore, peat development. At the Long Beach Peninsula, many of the shoreline proximal wetland sites most appropriate for recording tsunami inundation do not have a long peat record, thereby precluding the most recent pre-historic tsunami evidence. The ages of the tsunami layers on the Long Beach Peninsula show a strong lateral relationship. Older tsunami events are detected in the landward direction. The preservation of

tsunami events in the vertical strata on the Long Beach Peninsula are limited by land level changes and a dynamic accretionary history.

Figure 36. Conceptual diagram showing the cycle of sand progradation, punctuated by coseismic induced land level changes. The resulting groundwater level response impacts the timing of peat development, which is important as a contrasting sediment to observe tsunamigenic layers.

Rockaway, Oregon provides an interesting contrast to the Long Beach Peninsula. One main dune ridge separates the shoreline from the back-barrier wetland sites at Rockaway (versus 7 dune ridges at the Long Beach Peninsula). Well-developed peat horizons were cored at sites within 200 m of the shoreline (Figure 30).

Thick lake sediments were cored within 200 m of the shoreline (Figure 31).

Numerous sand layers were contained within these vertical facies. A clear record of marine incursions emerges from the vertical strata at Rockaway. The short width of the barrier, and back-barrier lakes (total distance 0.5 km), minimizes the usefulness of Rockaway for directly measuring inundation distances. However, knowing the thickness of the sand layers at Rockaway might provide some information about the energy of the tsunami flooding, and from that, inundation distances might be modeled.

Inferring Paleotsunami Magnitude from the Geologic Record

Of greatest importance to tsunami hazard mitigation is knowing the potential inland distances of high velocity and low velocity inundation. The inland extent of tsunami flooding hazards are dependant upon surge velocity and the vertical amplitude of the wave. Mapping prehistoric tsunami deposits can facilitate quantifying these important tsunami-shoreline dynamics. Though beyond the scope of this thesis, the use of sand grain distributions and isopachs of various sand layers might be used for calculations of water dynamics via sediment transport equations. For example, a range of assumed hydraulic heads and Manning coefficients yield predicted velocities from the equation

$$v = \frac{1}{n} R^{2/3} S^{1/2}$$

where v is mean velocity of flow, n is Mannings roughness coefficient, R is the hydraulic radius, and S is the average gradient of the channel (Pye, 1994). Mean velocities of at least 1 m/s are needed to keep medium- to fine-grained sand in turbulent suspension. Such mean current velocities can out-pace pedestrians crossing

rough ground. Drift logs and other large debris in such currents can serve as battering rams, or form debris rams that can topple small structures and roll automobiles.

Additional evidence is needed to better constrain the higher velocities of the paleotsunami inundations.

By comparison, low velocity inundation provides different types of hazards. Depths of ponded water can become significantly increased over multiple surges. The extensive flooding could cut off evacuation routes, as well as rescue routes. The exposure to cold water of escaping swimmers and waders also presents a hazard. Minimum inundation distance for a given prehistoric tsunami can be determined by mapping the inland extent of the sand sheet. The low velocity inundation records developed in this study, i.e. detrital caps, diatoms and bromine tracers, should be sufficient to extend the inundation records to maximum inundation distances in appropriate settings with further study.

Constraining Inundation Distance

Constraining inundation distance requires that paleoshoreline be estimated. Again, this was effectively carried out at the Long Beach Peninsula using ground-penetrating radar to locate subsidence-induced beach face erosional scarps (Meyers *et al.*, 1996), which were also ^{14}C dated. A placer-laden erosional scarp represents maximum landward extent of the shoreline for a given transgressive cycle. The erosional scarp is due to coseismic subsidence, and its formation can take years to decades after the tsunami inundation has occurred. If the amount of subsidence is known, which has been estimated to be 1-2 m for the Long Beach Peninsula, then Bruun's rule (Bruun, 1962) can be used to estimate pre-earthquake (and tsunami)

shoreline location for a given cycle. Bruun predicted that for a 1-2 m relative vertical water level increase, 200-300 m of horizontal shoreline erosion would occur. For example, Figure 17 shows the location of the erosional scarps for both the 300 ya and 1100 ya earthquakes. Sand layer B has been inferred to extend a distance of 1.0 km inland from this paleoshoreline feature, which constrains the maximum landward position of the shoreline for that period. Bruun's rule predicts that the actual shoreline prior to the 1100 ya earthquake was located 200-300 m to the west of the erosion scarp. Hence, the actual distance from the inferred paleoshoreline and the furthest inland extent of sand layer B is 1.2-1.3 km. This distance represents a constraint on the minimum inundation distance for the tsunami that transported the sand in layer B. At those locations where erosional scarps have not been mapped, paleoshoreline was constrained by the location of the shoreward extent of a given tsunamigenic sand sheet. For example, at Neskowin, the distance between the proximal and distal extent of sand layer B was 0.6 km. In light of the lack of paleoshoreline information for Neskowin, this represents the minimum constraint on inundation distance for sand layer B.

Constraining the Vertical Nature of Tsunami

The magnitude of the wave height can be discerned from the height of the barrier over which the tsunami crested. Constraining the magnitude of paleotsunami wave height requires that paleodune stability, and from that, paleo-elevation of the barrier, be established. This can be done by the use of radiocarbon and optical dating techniques to establish the age of the sand dunes with depth. This was effectively carried out at the Long Beach Peninsula. The highest dune (~8 m) separating Briscoe

Lake from the current shoreline was stable 300 years ago (Woxell, 1998). Qualitative observations of dune stability can also be made by the use of soil chronosequencing techniques. The highest dunes at Rockaway and Neskowin (~6 m) were stable at least 300 years ago based on the degree of development of weathered horizons on these

Table 11. Summary tabulation of minimum inundation distances at select locations for the central Cascadia Margin using various mapping concepts.

Sand Sheet Location	Event	Distance from Current Shoreline	Minimum Inundation Distance	Method for determining paleoshoreline	Current Dune Height	Paleodune Height	Paleodune Dating Method
Briscoe Lake, Long Beach	1700 AD	.8 km	.8 km +/- .2 km	Brunn's Rule, Erosional Scarp	8 meters	8 meters	Radiocarbon
Briscoe Lake, Long Beach	800 yr ago	.8 km	1.3 km +/- .2 km	Brunn's Rule, Erosional Scarp	8 meters	?	
Briscoe Lake, Long Beach	1300 yr ago	1.4 km	.7 km	Current Extent	8 meters	?	
Tinker Lake, Long Beach	800 yr ago	1 km	.8 km +/- .2 km	Brunn's Rule, Erosional Scarp	8 meters	?	
Gould Road, Grayland	1700 AD	1.5 km	.5 km	Current Extent	6 meters	?	
Gould Road, Grayland	800 yr ago	1.7 km	.6 km	Current Extent	6 meters	?	
Crescent Lake, Rockaway	1700 AD	.5 km	.4 km +/- .1 km	Paleodune Stability	6 meters	6 meters	Soil Chronosequencing
Crescent Lake, Rockaway	800 yr ago	.6 km	.3 km	Current Extent	6 meters	?	
North Marsh, Neskowin	1700 AD	.4 km	.2 km	Current Extent	6 meters	?	
North Marsh, Neskowin	800 yr ago	1 km	.5 km	Current Extent	6 meters	?	
North Marsh, Neskowin	1300 yr ago	1 km	.5 km	Current Extent	6 meters	?	
Central Marsh, Neskowin	1700 AD	.8 km	.7 km	Paleodune Stability	6 meters	6 meters	Soil Chronosequencing
Central Marsh, Neskowin	800 yr ago	1 km	.6 km	Current Extent	6 meters	?	

Explanation

Minimum Inundation Distance - this is the distance of tsunami inundation which is based on the distance from the constrained paleoshoreline to the maximum landward situated core where the sand sheet was observed.

Method for Determining Paleoshoreline - if an erosion scarp had been mapped along the shoreline, then the paleoshoreline was estimated by using the Brunn's rule relationship that correlates a 1-2 m relative sea level rise to 200-300 meters of landward erosion of the beach. This method represents the most accurate constraint of paleoshoreline position.

If an erosion scarp had not been mapped for the shoreline, then dune soils were analyzed to verify that the dune had been stable for at least the past few hundred years. If so, then the position of the dune represented the constraint on paleoshoreline. Where paleodune stability was not established, then the maximum extent of the sandsheet represented the minimum constraint on inundation distance.

Paleodune Height - this is the inferred elevation of the dune for paleotsunami events. Note that this constraint could only be effective employed for the most recent Cascadia event, which was 300 years ago, and only if dune stability was established for that time frame.

dunes. Hence, any 1700 AD tsunami deposits at these sites can be inferred to have topped these elevations. C14 dating and soil chronosequencing has been

completed at Grayland, so the nature of the barrier height 300 years ago is known for that location as well.

Table 11 represents a summary of tsunami flooding magnitude, based on the mapping principles described above. It is important to note that these represent minimum vertical (based on inferred paleo-elevations) and horizontal (based on paleoshoreline and inland extent of correlative sand deposits) inundation constraints. The actual inundation was larger than that represented in the table. In addition, a series of figures follows, which visually represents minimum inundation distances for all of the study sites.

Figure 37. Estimated extent of paleotsunami inundation for the 1700 AD event, 800 ya event, and 1300 ya event for the Long Beach Peninsula. The maximum dune elevation, as represented by the white line, is 8 ± 1 m. The solid lines represent known distances based on correlation of sand layers in cores. Dashed lines are inferred.

Figure 38. Inferred extent of paleotsunami inundation for the 1700 AD event and the 800 ya event for the central Grayland Plains. The maximum dune elevation, as represented by the white lines, is 6 ± 1 m. The solid lines represent known inundation distances based on the correlation of sand sheets. Dashed lines are inferred.

Figure 39. Aerial photograph of Rockaway showing the inundation lines for the 1700 AD and 800 ya paleotsunami. The black dots represent core locations.

Figure 40. Plan view map of Neskowin showing the inferred paleotsunami inundation distances for the 1700 AD event and the 800 ya event. The mapped lines published by Oregon Department of Geology and Industries (DOGAMI) is shown for comparison (Priest, 1995). Figure is modified from Priest, 1995.

Conclusions

1. Anomalous sand layers have been described at Grayland Plains and the Long Beach Peninsula, Washington and Rockaway and Neskowin, Oregon. These sites are characterized as broad open coastal plains with back barrier wetlands. Based on paleofossil assemblages, the presence of marine geochemical tracers, sand grain size and the general landward thinning of the sand layers, they are interpreted to be from marine surges.
2. The broad lateral continuity and regional correlation to known great earthquake events suggest earthquake-induced tsunami as the origin of the marine surges.
3. A weak record of sand sheet deposition is observed at all of the study sites for the event corresponding to 50-330 cal YBP. More robust tsunami sand sheets occurred 715-1015 cal YBP, 970-1340 cal YBP, 1330-1460 cal YBP and 2050-2375 cal YBP.
4. Based on inferred paleodune elevations separating Briscoe Lake from the Pacific Ocean at the Long Beach Peninsula, the tsunami which occurred 50-330 cal YBP crested at least $8\text{m} \pm 1\text{m}$ and inundated at high velocity for a distance of at least 0.7 km landward. Based on the inferred paleoshoreline of the Pacific Ocean, the tsunami which occurred 715-1015 cal YBP inundated at high velocity for at least 1 km landward in the vicinity of Tinker Lake, and 1.3 km in the vicinity of Briscoe Lake.
5. Based on inferred paleodune elevations at Rockaway, Oregon, the tsunami which occurred 50-330 cal YBP crested at least $6\text{m} \pm 1\text{m}$ in the vicinity of Crescent Lake and inundated inland for at least 0.5 km landward.

6. Based on inferred paleodune elevations at Neskowin, Oregon, the tsunami which occurred 50-330 cal YBP crested at least $6\text{m} \pm 1\text{m}$ and inundated inland for at least 1 km.
7. Indicators of low velocity inundation, i.e. bromine, diatom, and detrital cap tracers, indicate further, but as yet unconstrained, landward inundation distances.

Bibliography

- Abramson, H., 1998. Evidence for tsunami and earthquakes during the last 3500 years from Lagoon Creek, a coastal freshwater marsh, northern California. Unpublished master's thesis, Humboldt State University, 75 p.
- Adams, J., 1990. Paleoseismicity of the Cascadia subduction zone - Evidence from turbidites off the Oregon-Washington margin. *Tectonics*, v. 9, p. 569-583.
- Atwater, B.F., 1987. Evidence for great Holocene earthquakes along the outer coast of Washington State. *Science*, v. 236, p. 242-244.
- Atwater, B.F., 1992. Geologic evidence for earthquakes during the past 2000 years along the Copalis River, southern coastal Washington. *Journal of Geophysical Research*, v. 97, pp. 1901-1919.
- Atwater, B.F. and Hemphill-Haley, E., 1997. Recurrence intervals for great earthquakes of the past 3500 years at northeastern Willapa Bay, Washington. U.S. Geological Survey Professional Paper 1576, 108 p.
- Barnett, E., 1997. The potential for coastal flooding due to coseismic subsidence in the central Cascadia margin. Unpublished master's thesis, Portland State University, 144 p.
- Birkeland, P.W., 1984. Soils and geomorphology. Oxford University Press, New York, 372 p.
- Bruun, P., 1962. Sea level rise as a cause of shore erosion. *Proceedings of the American Civil Engineering Society*, v. 88, p. 117-130.

- Bucknam, R.C., Hemphill-Haley, E. and Leopold, E.B., 1992. Abrupt uplift within the past 1700 years at southern Puget Sound, Washington. *Science*, v. 258, p. 1611-1614.
- Clague, J.J. and Bobrowsky, P.Y., 1994. Tsunami deposits beneath tidal marshes on Vancouver Island, British Columbia. *Geological Society of America Bulletin*, v. 106, p. 1293-1303.
- Clague, J.J., Hutchinson, I., Mathewes, R.W. and Patterson, R.T., 1999. Evidence for late Holocene tsunamis at Catala Lake, British Columbia. *Journal of Coastal Research*, v. 15, p. 45-60.
- Clarke, S.H. and Carver, G.A., 1992. Late Holocene tectonics and paleoseismicity, southern Cascadia subduction zone. *Science*, v. 155, p. 188-192.
- Dariento, M.E., 1987. Late Holocene geologic history of a Netarts Bay salt marsh, northwest Oregon coast and its relationship to relative sea level changes. Unpublished master's thesis, University of Oregon, 94 p.
- Dariento, M.E. and Peterson, C.D., 1990. Episodic tectonic subsidence of late Holocene salt marshes, northern Oregon central Cascadia Margin. *Tectonics*, v. 9, pp. 1-22.
- Dariento, M.E., Peterson, C.D. and Clough, C., 1994. Stratigraphic evidence for great subduction-zone earthquakes at four estuaries in northern Oregon. *Journal of Coastal Research*, v. 10, p. 850-876.
- Dariento, M.E. and Peterson, C.D., 1995. Magnitude and frequency of subduction-zone earthquakes along the Oregon coast in the past 3000 years. *Oregon Geology*, v. 57, p. 3-12.
- Davis, J.C., 1986. *Statistics and Data Analysis in Geology*. John Wiley and Sons, New York, 646 p.

- Duce, R. A., Winchester, J.W. and Van Nahl, T.W., 1965. Iodine, Bromine, and Chlorine in the Hawaiian marine atmosphere. *Journal of Geophysical Research*, v. 70, p. 1775-1799.
- Eilers, J.M., Bernert, J.A. and Nelson, D.O., 1995. Lake Lytle, Oregon - Diagnostics and Feasability Study: A Clear Lakes Study. E&S Environmental Chemistry Consulting, Corvallis, Oregon, 45 p.
- Fiedorowicz, B., 1997. Geologic evidence of historic and prehistoric tsunami inundation at Seaside, Oregon. Unpublished master's thesis, Portland State University, Portland, Oregon, 197 p.
- Fluck, P., Hyndman, R.D. and Wang, K., 1997. Three-dimensional dislocation model for great Earthquakes of the Cascadia Subduction Zone. *Journal of Geophysical Research*, v. 102, p. 20539-20550.
- Folk, R.L., 1980. *Petrology of Sedimentary Rocks*. Hemphill Publishing, Austin, 185 p.
- Foster, I.D.L., 1991. High energy coastal sedimentary deposits; an evaluation of depositional processes in southwest England. *Earth Surface Processes and Landforms*, v. 16, p. 341-356.
- Fuge, R., 1988. Sources of halogens in the environment, influences on human and animal health. *Environmental Geochemistry and Health*, v. 10, p. 51-61.
- Goldfinger, C., Kulm, L.D., Yeats, R.S., Applegate, T.B., MacKay, M.E. and Moore, G.F., 1992. Transverse structural trends along the Oregon convergent margin – Implications for Cascadia earthquake potential and crustal rotations. *Geology*, v. 20, p. 141-144.
- Goldfinger, C., Kulm, L.D., Yeats, R.S., Applegate, B., MacKay, M.E. and Cochrane, G.R., 1996. Active strike-slip faulting and folding of the Cascadia subduction zone plate boundary and forearc in central and northern Oregon, in Rogers,

A.M., Walsh, T.J., Kockelman, W.J., Priest, G.R., eds., Assessing earthquake hazards and reducing risk in the Pacific Northwest. U.S. Geological Survey Professional Paper 1560.

Grant, W.E., 1989. More evidence from tidal-marsh stratigraphy for multiple late Holocene subduction earthquakes along the northern Oregon coast. Geological Society of America Abstracts with Programs, v. 24, p. 9-10.

Harden, J.W. and Taylor, E.M., 1983. A quantitative comparison of soil development in four climatic regimes. Quaternary Research, v. 10, p. 342-359.

Heaton, T.H. and Hartzell, S.H., 1986. Source characteristics of hypothetical subduction zone earthquakes in the northwestern United States. Seismological Society of America Bulletin, v. 76, p. 675-708.

Hemphill-Haley, E., 1996. Diatoms as an aid in identifying late-Holocene tsunami deposits. The Holocene, v. 6, p. 439-448.

Hutchinson, I. Clague, J.J. and Mathewes, R.W., 1997. Reconstructing the tsunami record on an emerging coast: a case study of Kanim Lake, Vancouver Island, British Columbia, Canada. Journal of Coastal Research, v. 13, p. 545-554.

Jaffe, B., Richmond, B. and Gelfenbaum, G., 1998a. Predictive models for tsunami and hurricane deposits based on simple hydrodynamic and sediment transport considerations. Abstract from American Association of Petroleum Geologists-Society of Economic Paleontologists and Mineralogists annual meeting, 1998.

Jaffe, B., Gelfenbaum, G., Benson, B., Davies, H. and Nongkas, M., 1998b. Sedimentation, erosion and flow in the July 17, Papua New Guinea tsunami. AGU meeting abstract, 1998.

Jaffe, B., Gelfenbaum, G., Nongkas, M. and Davies, H., 1999. Tsunami deposits from the July 17, 1998 Papua New Guinea Event. Abstract from SOPAC STAR meeting, Suva, Fiji.

- Kanamori, H. and Kikuchi, M., 1993. The 1992 Nicaraguan earthquake: A slow tsunami earthquake associated with subducted sediments. *Nature*, v. 361, p. 714.
- Lander, J.F., Lockridge, P.A., and Kozuch, M.J., 1993. Tsunami affecting the west coast of the United States, 1806-1992. US Department of Commerce, National Oceanic and Atmospheric Administration, NGDC Key to Geophysical Records, Documentation No. 29, 242 p.
- Meyers, R.A., Smith, D.G., Jol, H.M. and Peterson, C.D., 1996. Evidence for eight great earthquake-subsidence events detected with ground-penetrating radar, Willapa Barrier, Washington. *Geology*, v. 24, p. 99-102.
- Myers, E.P., Baptista, A.M. and Priest, G.R., 1999. Finite element modeling of potential Cascadia subduction zone tsunamis. *Science of Tsunami Hazards*, v. 17, p. 3-18.
- Minoura, K., Shu, N. and Uchida, M., 1994. Tsunami deposits in a lacustrine sequence of the Sanriku coast, northeast Japan. *Sedimentary Geology*, v. 89, p. 25-31.
- Minoura, K. and Nakaya, S., 1991. Traces of tsunami preserved in inter-tidal lacustrine and marsh deposits: some example from northeast Japan. *Journal of Geology*, v. 99, p. 265-287.
- Muecke, G.K., ed., 1980. Neutron Activation Analysis in the Geosciences Short Course Handbook. Mineralogical Association of Canada, v. 5, 275 p.
- Nelson, A.R., Jennings, A.E., and Kashima, K., 1996a. Holocene intertidal stratigraphy, microfossils, rapid submergence, and earthquake recurrence at Coos Bay, southern coastal Oregon, USA. *Geological Society of America Bulletin*, v. 108, p. 141-154.

- Nelson, A.R., Kelsey, H.M., Hemphill-Haley, E. and Witter, R.C., 1996b. A 7500-year lake record of Cascadia tsunamis in southern coastal Oregon. The Geological Society of America Abstracts with Programs. 134 p.
- Ng, M.K., Leblond, P.H. and Murty, T.S., 1990. Simulation of tsunami from great earthquakes on the Cascadia Subduction Zone. *Science*, v. 250, p. 1248-1251.
- ODOT (Oregon Department of Transportation), 1967. Ocean Shores - orthophoto topographic map of Oregon Beaches, Nehalem Bay-Tillamook Bay Section (21 sheets), Nestucca Bay to Lincoln City Section (29 sheets), contour interval = 2 feet.
- Peterson, C.D. and Darienzo, M.E., 1996. Discrimination of climatic, oceanic, and tectonic mechanisms of cyclic marsh burial, Alsea Bay, Oregon. In Rogers, A.M., Walsh, T.J., Kockelman, W.J., Priest, G.R., ed., *Assessing earthquake hazards and reducing risk in the Pacific Northwest, U.S. Geological Survey Professional Paper 1560*, 306 p.
- Peterson, C.D., Darienzo, M.E. and Parker, M., 1988. Coastal neotectonic field trip guide for Netarts Bay, Oregon. *Oregon Geology*, v. 50, p. 99-106.
- Peterson, C.D., Gelfenbaum, G.R., Jol, H.M., Phipps, J.B., Reckendorf, F., Twichell, D.C., Vanderburg, S. and Woxell, L., 1999. Great Earthquakes, abundant sand, and high wave energy in the Columbia cell, USA. *Proceedings from Coastal Sediments '99: American Society of Civil Engineers*, p. 1676-1690.
- Peterson, C.D. and Madin, I.P., 1997. Coseismic paleoliquefaction evidence in the central Cascadia margin, USA. *Oregon Geology*, v. 59, p. 51-74.
- Peterson, C.D. and Priest, G.R., 1995. Preliminary reconnaissance survey of Cascadia paleotsunami deposits in Yaquina Bay, Oregon. *Oregon Geology*, v. 57, p. 33-40.

- Peterson, C.D. and Phipps, J.B., 1992. Holocene sedimentary framework of Grays Harbor Basin, Washington, USA. *Quaternary Coastal Systems of the United States, Marine and Lacustrine Systems*, SEPM Special Publication 48, p. 273-285.
- Plafker, G., Kachadoorian, R., Eckel, E.B. and Mayo, L.R., 1969. Effects of the Earthquake of March 27, 1964 on various communities. *Geological Survey Professional Paper 542-G*, 50 p.
- Plafker, G., Lajoie, K.R. and Rubin, M., 1992. Determining recurrence intervals of great subduction zone earthquakes in souther Alaska by radiocarbon dating, in Taylor, R.E., Long, A., and Kra, R.S., eds., *Radiocarbon dating after four decades: An interdisciplinary perspective*: New York, Springer-Verlag, 453 p.
- Priest, G.R., 1995. Explanation of mapping methods and use of the tsunami hazard map of the Siletz Bay area, Lincoln County, Oregon. *State of Oregon Department of Geology and Mineral Industries Open File Report O-95-05*, 69 p.
- Pye, K., ed., 1994. *Sediment Transport and Depositional Processes*. Blackwell Scientific Publications, Oxford, 397 p.
- Rogers, G.C., 1988. An assessment of the megathrust earthquake potential of the Cascadia subduction zone. *Canadian Journal of Earth Sciences*, v. 25, p. 844-852.
- Satake, K., Shimazaki, K., Tsuji, Y. and Ueda, K., 1996. Time and size of a giant earthquake in Cascadia inferred from Japanese tsunami record of January 1700. *Nature*, v. 379, p. 246-249.
- Stuiver, M. and Reimer, P.J., 1993. Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program. *Radiocarbon*, v. 35, p. 215-230.

- Tanioka, Y. and Satake, K., 1996. Tsunami generation by horizontal displacement of ocean bottom. *Geophysical Research Letters*, v. 23, p. 861-864.
- Whitmore, P.M., 1993. Expected tsunami amplitudes and currents along the North American coast for Cascadia subduction zone earthquakes. *Natural Hazards*, v. 8, p. 59-73.
- Whitmore, P.M., 1994. Expected tsunami amplitudes off the Tillamook County, Oregon coast following a major Cascadia subduction zone earthquake. *Oregon Geology*, v. 56, p. 62-64.
- Woxell, L.K., 1998. Prehistoric beach accretion rates and long-term response to sediment depletion in the Columbia River Littoral System, USA. Unpublished master's thesis, Portland State University, 143 p.
- Yamaguchi, D.K., Atwater, B.F., Bunker, D.E., Benson, B.E. and Reid, M.S., 1997. Tree-ring dating the 1700 Cascadia earthquake. *Nature*, v. 389, p. 922-923.

Appendix A

Coordinates of Core Locations

Date of Core	Number	ID Number	Location	Method	Northing	Easting
6/24/97	1	Brea-1-970624	Long Beach	Gouge Core		
6/24/97	2	Brea-2-970624	Long Beach	Gouge Core		
6/24/97	3	Bris-4-970624	Long Beach	Gouge Core	5137374	419094
6/24/97	4	Bris-2-970624	Long Beach	Gouge Core		
6/24/97	5	Bris-1-970624	Long Beach	Gouge Core		
6/24/97	6	Sayce-1-970624	Long Beach	Gouge Core	5151253	420470
6/24/97	7	Sayce-2-970624	Long Beach	ouge Core	5151260	420479
6/25/97	8	Tink-1-970625	Long Beach	Gouge Core	5134473	419303
6/25/97	9	Tink-3-970625	Long Beach	Gouge Core	5134518	419333
6/25/97	10	Tink-4-970625	Long Beach	Gouge Core	5134512	419370
6/25/97	11	Brea-2-970625	Long Beach	Gouge Core		
6/25/97	12	Deer-1-970625	Long Beach	Gouge Core	5136660	420099
6/26/97	12	Blac-4-970626	Long Beach	Gouge Core	5130180	419696
6/25/97	13	Gile-1-970625	Long Beach	Gouge Core	5137516	420591
6/26/97	14	Blac-12-970626	Long Beach	Gouge Core		
6/26/97	16	Blac-1-970626	Long Beach	Gouge Core	5130199	419751
6/26/97	17	Blac-2-970626	Long Beach	Gouge Core	5130201	419777
6/26/97	18	Blac-5-970626	Long Beach	Gouge Core	5130180	419696
6/26/97	19	Blac-13-970626	Long Beach	Gouge Core	5130281	419628
6/26/97	20	Blac-14-970626	Long Beach	Gouge Core	5130264	419497
6/26/97	21	Tink-1-970626	Long Beach	Gouge Core	5133674	419433
6/26/97	22	Bris-7-970626	Long Beach	Gouge Core	5137191	419097
6/26/97	23	Bris-4-970626	Long Beach	Gouge Core		
6/26/97	24	Bris-5-970626	Long Beach	Gouge Core		
6/26/97	25	Bris-6-970626	Long Beach	Gouge Core		
7/15/97	26	Sayce-1-970715	Long Beach	Vibracore		
7/16/97	27	Blac-1-970716	Long Beach	Vibracore		
8/3/97	28	Skat-1-970803	Long Beach	Gouge Core	5153002	420236
8/3/97	29	Bris-1-970803	Long Beach	Gouge Core		
8/3/97	30	Bris-2-970803	Long Beach	Pound Core		
8/3/97	31	Brea-1-970803	Long Beach	Pound Core		
8/4/97	32	Cran-6-970804	Long Beach	Gouge Core	5136965	419500
8/4/97	33	Cran-5-970804	Long Beach	Gouge Core	5137380	419645
8/4/97	34	Cran-4-970804	Long Beach	Gouge Core	5138165	419633
8/4/97	35	Cran-1-970804	Long Beach	Gouge Core	5138617	419622
8/4/97	36	Cran-2a-970804	Long Beach	Gouge Core	5138660	419649
8/4/97	37	Cran-2b-970804	Long Beach	Gouge Core	5138660	419649
8/4/97	38	Cran-Zc-970804	Long Beach	Gouge Core	5138660	419649
8/5/97	39	Birc-1-970805	Long Beach	Gouge Core	5138973	429216
8/5/97	40	Tink-1-970805	Long Beach	Gouge Core		
8/5/97	41	Loom-1-970805	Long Beach	Gouge Core		
8/15/97	42	Blac-1-970815	Long Beach	Vibracore		
8/15/97	43	Bris-2-970815	Long Beach	Gouge Core	5137722	419105
8/15/97	44	Bris-1-970815	Long Beach	Gouge Core	5136820	419185
8/15/97	45	Loom-5-970815	Long Beach	Gouge Core	5141766	419935
8/15/97	46	Loom-4-970815	Long Beach	Gouge Core	5142863	419760
8/15/97	47	Loom-1-970815	Long Beach	Gouge Core	5143553	419754
8/15/97	48	Loom-2-970815	Long Beach	Gouge Core	5144003	419947
8/15/97	49	Loom-1-970815	Long Beach	Gouge Core		
8/15/97	50	Loom-3-970815	Long Beach	Vibracore		
8/21/97	51	Onell-1-970821	Long Beach	Gouge Core	5126010	419310
8/21/97	52	Onell-2-970821	Long Beach	Gouge Core	5126100	418810
8/21/97	53	Onell-3-970821	Long Beach	Gouge Core	5126110	418820
8/27/97	54	Cran-3-980827	Long Beach	Gouge Core		
8/27/98	55	PCOU-1-980827	Long Beach	Gouge Core		
8/27/98	56	PCOU-2-980827	Long Beach	Gouge Core		
8/27/98	57	PCOU-3-980827	Long Beach	Gouge Core		
8/27/98	58	Bris-3-980827	Long Beach	Gouge Core		
8/27/98	59	Bris-2-980827	Long Beach	Gouge Core		
8/27/98	60	Bris-1-980827	Long Beach	Gouge Core		
8/27/98	61	Cran-1-980827	Long Beach	Gouge Core		
8/27/98	62	Cran-2-980827	Long Beach	Gouge Core		
9/1/98	63	Spen-1-980901	Long Beach	Gouge Core		
9/1/98	64	Spen-2-980901	Long Beach	Gouge Core		

Date of Core	Number	ID Number	Location	Method	Northing	Easting
9/1/98	65	Spen-3-980901	Long Beach	Gouge Core		
8/16/97	66	Lyti-1-970816	Rockaway	Gouge Core	5053036	426659
8/16/97	67	Lyti-2-970816	Rockaway	Gouge Core	5052929	426724
8/16/97	68	Lyti-3-970816	Rockaway	Gouge Core	5052622	426869
8/16/97	69	Lyti-4-970816	Rockaway	Gouge Core	5052600	426570
8/16/97	70	Lyti-5-970816	Rockaway	Gouge Core	5052600	426600
8/16/97	71	Lyti-6-970816	Rockaway	Gouge Core	5052588	426613
8/16/97	72	Lyti-9-970816	Rockaway	Vibracore	5052600	426570
8/17/97	73	Lyti-9-970817	Rockaway	Gouge Core	5052400	426825
8/17/97	74	Lyti-8-970817	Rockaway	Gouge Core	5052394	426723
8/17/97	75	Lyti-7-970817	Rockaway	Gouge Core	5052384	426667
8/17/97	76	Lyti-10-970817	Rockaway	Gouge Core	5052854	426657
8/17/97	77	Lyti-6-970817	Rockaway	Gouge Core	5052365	426585
9/4/97	78	Frie-1-970904	Rockaway	Gouge Core	5049302	426160
9/4/97	79	Frie-2-970904	Rockaway	Gouge Core	5049277	426160
9/4/97	80	Frie-3-970904	Rockaway	Gouge Core	5049285	426268
9/6/97	81	Smit-1-970906	Rockaway	Gouge Core		
9/6/97	82	Smit-2-970906	Rockaway	Gouge Core	5048134	426070
9/6/97	83	Smit-3-970906	Rockaway	Gouge Core	5048085	426063
9/6/97	84	Smit-5-970906	Rockaway	Gouge Core		
9/6/97	85	Smit-6-970906	Rockaway	Gouge Core		
9/6/97	86	Smit-7-970906	Rockaway	Gouge Core		
9/7/97	87	Neat-2-970907	Rockaway	Gouge Core		
9/7/97	88	Neat-1-970907	Rockaway	Gouge Core		
6/16/98	89	Cresc-1-980616	Rockaway	Gouge Core	5053555	426987
6/16/98	90	Cresc-2-980616	Rockaway	Gouge Core	5053763	426925
6/17/98	91	Crik-10-980617	Rockaway	Gouge Core		
6/17/98	92	Crik-12-980617	Rockaway	Gouge Core		
6/17/98	93	Crik-15-980617	Rockaway	Gouge Core		
6/17/98	94	Crik-17-980617	Rockaway	Gouge Core		
6/17/98	95	Cresc-4-980617	Rockaway	Gouge Core	5053749	426827
6/17/98	96	Crik-9-980617	Rockaway	Gouge Core	5053792	420892
6/17/98	97	Cresc-9-980617	Rockaway	Gouge Core	5052361	426111
6/17/98	98	Crik-11-980617	Rockaway	Gouge Core		
6/17/98	99	Crik-13-980617	Rockaway	Gouge Core		
6/17/98	100	Crik-14-980617	Rockaway	Gouge Core		
6/17/98	101	Crik-16-980617	Rockaway	Gouge Core		
6/17/98	102	Crik-18-980617	Rockaway	Gouge Core		
6/17/98	103	Crik-19-980617	Rockaway	Gouge Core		
6/17/98	104	Crik-20-980617	Rockaway	Gouge Core		
6/17/98	105	Cresc-3-980617	Rockaway	Gouge Core	5053732	426961
6/17/98	106	Cresc-5-980617	Rockaway	Gouge Core	5053749	426757
6/17/98	107	Cresc-6-980617	Rockaway	Gouge Core	5053703	426896
6/17/98	108	Cresc-20-980617	Rockaway	Gouge Core		
6/17/98	109	Cresc-21-980617	Rockaway	Gouge Core		
6/17/98	110	Cresc-22-980617	Rockaway	Gouge Core		
6/17/98	111	Cresc-23-980617	Rockaway	Gouge Core		
6/17/98	112	Cresc-24-980617	Rockaway	Gouge Core		
6/17/98	113	Cresc-25-980617	Rockaway	Gouge Core		
6/17/98	114	Cresc-26-980617	Rockaway	Gouge Core		
6/17/98	115	Cresc-27-980617	Rockaway	Gouge Core		
6/17/98	116	Crik-7-980617	Rockaway	Gouge Core		
6/17/98	117	Crik-8-980617	Rockaway	Gouge Core		
6/17/98	118	Crik-9-980617	Rockaway	Gouge Core	5053555	426987
6/18/98	119	Cresc-1-980618	Rockaway	Gouge Core	5053783	426820
6/18/98	120	Cresc-2-980618	Rockaway	Gouge Core		
6/18/98	121	Cresc-3-980618	Rockaway	Gouge Core		
6/19/98	122	Lyti-2-980619	Rockaway	Gouge Core	5052365	426585
6/19/98	123	Cres-1-980619	Rockaway	Vibracore	5053757	426923
6/19/98	124	Lyti-1-980619	Rockaway	Vibracore	5052422	426607
6/19/98	125	Lyti-2-980619	Rockaway	Vibracore	5052365	426585
9/5/97	126	Daly-1a-970905	Neskowin	Gouge Core		
9/7/97	127	Daly-1-970907	Neskowin	Vibracore		
7/6/98	128	Nesk-1-980706	Neskowin	Gouge Core		
7/6/98	129	Nesk-2-980706	Neskowin	Gouge Core		

Date of Core	Number	ID Number	Location	Method	Northing	Easting
7/6/98	130	Nesk-3-980706	Neskowin	Gouge Core		
7/6/98	131	Nesk-4-980706	Neskowin	Gouge Core		
7/6/98	132	Nesk-5-980706	Neskowin	Gouge Core		
7/6/98	133	Nesk-6-980706	Neskowin	Gouge Core		
7/6/98	134	Nesk-7-980706	Neskowin	Gouge Core		
7/6/98	135	Nesk-8-980706	Neskowin	Gouge Core		
7/6/98	136	Nesk-10-980706	Neskowin	Gouge Core		
7/6/98	137	Nesk-11-980706	Neskowin	Gouge Core		
7/6/98	138	Nesk-12-980706	Neskowin	Gouge Core		
7/7/98	140	Nesk-10-980707	Neskowin	Pound Core	4997582	423220
7/7/98	141	Nesk-2-980707	Neskowin	Gouge Core	4997495	423273
7/7/98	142	Nesk-4-980707	Neskowin	Gouge Core	4997497	423302
7/7/98	143	Nesk-5-980707	Neskowin	Gouge Core	4997487	423346
7/7/98	144	Nesk-6-980707	Neskowin	Gouge Core	4997476	423383
7/7/98	145	Nesk-7-980707	Neskowin	Gouge Core	4997461	423420
7/7/98	146	Nesk-8-980707	Neskowin	Gouge Core	4997478	423450
7/7/98	147	Nesk-9-980707	Neskowin	Pound Core	4997495	423273
7/7/98	148	Nesk-11-980707	Neskowin	Gouge Core	4997493	423227
7/7/98	149	Nesk-1a-980707	Neskowin	Gouge Core		
7/7/98	150	Nesk-2a-980707	Neskowin	Gouge Core		
7/7/98	151	Nesk-3a-980707	Neskowin	Gouge Core		
7/7/98	152	Nesk-4a-980707	Neskowin	Gouge Core		
7/7/98	153	Nesk-10-980707	Neskowin	Gouge Core	4997582	423220
7/8/98	154	Nesk-11-980708	Neskowin	Gouge Core		
7/8/98	155	Nesk-10-980708	Neskowin	Gouge Core		
7/8/98	156	Nesk-1-980708	Neskowin	Gouge Core	4996836	423149
7/8/98	157	Nesk-2-980708	Neskowin	Gouge Core	4996842	423263
7/8/98	158	Nesk-3-980708	Neskowin	Gouge Core	4996804	423317
7/8/98	159	Nesk-4-980708	Neskowin	Gouge Core	4996760	423382
7/8/98	160	Nesk-9-980708	Neskowin	Gouge Core		
7/8/98	161	Nesk-5-980708	Neskowin	Gouge Core		
7/8/98	162	Nesk-6-980708	Neskowin	Pound Core	4996760	423382
7/8/98	163	Nesk-7-980708	Neskowin	Pound Core		
7/8/98	164	Nesk-8-980708	Neskowin	Gouge Core		
7/9/98	165	Nesk-1-980709	Neskowin	Gouge Core		
7/9/98	166	Nesk-2-980709	Neskowin	Gouge Core		
7/9/98	167	Nesk-3-980709	Neskowin	Gouge Core		
7/9/98	168	Nesk-4-980709	Neskowin	Pound Core		
6/23/98	169	Goul-2-980623	Grayland	Gouge Core		
6/23/98	170	Goul-3-980623	Grayland	Gouge Core		
6/23/98	171	Goul-4-980623	Grayland	Gouge Core		
6/23/98	172	Goul-5-980623	Grayland	Gouge Core		
6/24/98	173	Goul-3-980624	Grayland	Gouge Core		
6/24/98	174	Goul-5-980624	Grayland	Gouge Core		
6/24/98	175	Goul-4-980624	Grayland	Gouge Core		
6/24/98	176	Twin-2-980624	Grayland	Gouge Core	5189880	415637
6/24/98	177	Twin-3-980624	Grayland	Gouge Core	5189880	415647
6/24/98	178	Twin-4-980624	Grayland	Gouge Core	5189852	415895
6/24/98	179	Goul-2-980624	Grayland	Pound Core		
6/25/98	180	Twin-1-980625	Grayland	Gouge Core		
6/25/98	181	Twin-2-980625	Grayland	Gouge Core		
6/25/98	182	Twin-3-980625	Grayland	Gouge Core		
6/25/98	183	Twin-4-980625	Grayland	Gouge Core		
6/25/98	184	Twin-5-980625	Grayland	Gouge Core		
6/25/98	185	Twin-6-980625	Grayland	Gouge Core		
6/25/98	186	Goul-1-980625	Grayland	Pound Core		
8/26/98	187	Summ-3-980826	Grayland	Gouge Core		
8/26/98	188	Summ-2-980826	Grayland	Gouge Core		
8/26/98	189	Summ-1-980826	Grayland	Gouge Core		
8/26/98	190	Heat-1-980826	Grayland	Gouge Core		
8/26/98	191	Heat-2-980826	Grayland	Gouge Core		
8/26/98	192	Heat-3-980826	Grayland	Gouge Core		
8/26/98	193	Heat-4-980826	Grayland	Gouge Core		
8/26/98	194	Goul-1-980826	Grayland	Pound Core		
8/26/98	195	Summ-4-980826	Grayland	Pound Core		

Appendix B

Long Beach Cores

Core Log Key

Peat
Sand
Mud (gyttja)
Peaty Mud
Fill - disturbed
Soil - oxidized
Sandy Peat
Detritus
Wood
Water
Missing Section

Lithological Characteristics

C - clean, well-sorted sand
 mU - medium upper grain size (350-500 μ)
 mL - medium lower grain size (250-350 μ)
 fU - fine upper grain size (177-250 μ)

ls - sand lamina
 ld - detrital lamina
 lm - mud lamina

dcap - detrital cap
 ds - disseminated sand
 dd - disseminated detritus
 dm - disseminated mud, often as rip-ups
 wc - wood chunks
 l - liquefaction features

Contact Relations

dc - diffuse contact
 gc - gradual contact
 sc - sharp contact
 ac - abrupt contact
 ic - intruded contact (liquefaction dikes)

Appendix C

Grayland Core Logs

Core Log Key

Peat
Sand
Mud (gyttja)
Peaty Mud
Fill - disturbed
Soil - oxidized
Sandy Peat
Detritus
Wood
Water
Missing Section

Lithological Characteristics

C - clean, well-sorted sand
 mU - medium upper grain size (350-500 μ)
 mL - medium lower grain size (250-350 μ)
 fU - fine upper grain size (177-250 μ)

ls - sand lamina
 ld - detrital lamina
 lm - mud lamina

dcap - detrital cap
 ds - disseminated sand
 dd - disseminated detritus
 dm - disseminated mud, often as rip-ups
 wc - wood chunks
 l - liquefaction features

Contact Relations

dc - diffuse contact
 gc - gradual contact
 sc - sharp contact
 ac - abrupt contact
 ic - intruded contact (liquefaction dikes)

Appendix D

Rockaway Core Logs

Core Log Key

Peat
Sand
Mud (gyttja)
Peaty Mud
Fill - disturbed
Soil - oxidized
Sandy Peat
Detritus
Wood
Water
Missing Section

Lithological Characteristics

C - clean, well-sorted sand
 mU - medium upper grain size (350-500 μ)
 mL - medium lower grain size (250-350 μ)
 fU - fine upper grain size (177-250 μ)

ls - sand lamina
 ld - detrital lamina
 lm - mud lamina

dcap - detrital cap
 ds - disseminated sand
 dd - disseminated detritus
 dm - disseminated mud, often as rip-ups
 wc - wood chunks
 l - liquefaction features

Contact Relations

dc - diffuse contact
 gc - gradual contact
 sc - sharp contact
 ac - abrupt contact
 ic - intruded contact (liquefaction dikes)

Appendix E

Neskowin Core Logs

Core Log Key

Peat
Sand
Mud (gyttja)
Peaty Mud
Fill - disturbed
Soil - oxidized
Sandy Peat
Detritus
Wood
Water
Missing Section

Lithological Characteristics

C - clean, well-sorted sand
 mU - medium upper grain size (350-500 μ)
 mL - medium lower grain size (250-350 μ)
 fU - fine upper grain size (177-250 μ)

ls - sand lamina
 ld - detrital lamina
 lm - mud lamina

dcap - detrital cap
 ds - disseminated sand
 dd - disseminated detritus
 dm - disseminated mud, often as rip-ups
 wc - wood chunks
 l - liquefaction features

Contact Relations

dc - diffuse contact
 gc - gradual contact
 sc - sharp contact
 ac - abrupt contact
 ic - intruded contact (liquefaction dikes)

Appendix F

Dune Soil Rubification Values – Rockaway

Field Sand Auger Logs, with hue and chroma measurements shown as measured in the field.

Core Location - Rock-2-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
2			A Horizon
5	2.5 YR	6/4	
10	2.5 YR	6/4	
25	2.5 YR	6/4	
40	2.5 YR	5/3	paleosol
50	2.5 YR	5/4	
75	2.5 YR	6/4	
100	2.5 YR	5/4	
150	2.5 YR	7/4	
200	2.5 YR	6/4	
235	2.5 YR	5/4	weak Bw
250	2.5 YR	6/4	

Core Location - Rock-3-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
5			A Horizon
10	2.5 YR	6/4	
15	2.5 YR	6/4	
25	2.5 YR	6/4	
50	2.5 YR	5/2	paleosol
75	2.5 YR	5/3	
100	2.5 YR	6/4	
125	2.5 YR	6/4	
135	2.5 YR	5/2	slight humic enrichment
145	2.5 YR	5/4	
155	2.5 YR	7/4	
250	10 YR	6/6	strong Bw

Core Location - Rock-5-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
5			A Horizon
20	2.5 YR	5/2	
75	2.5 YR	6/4	
100	2.5 YR	5/4	very weak Bw
125	2.5 YR	5/4	very weak Bw
200	2.5 YR	5/4	very weak Bw
250	2.5 YR	6/4	
300	2.5 YR	5/4	
320	2.5 YR	4/3	A horizon paleosol
325	2.5 YR	5/6	Bw

Core Location - Rock-6-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
5			A Horizon
20	2.5 YR	7/4	
50	2.5 YR	6/3	
75	2.5 YR	6/3	
100	2.5 YR	5/3	weak A horizon
150	2.5 YR	6/3	
200	2.5 YR	6/4	
220	2.5 YR	4/2	strong A horizon
225	2.5 YR	5/4	weak Bw
240	2.5 YR	6/4	
250	2.5 YR	6/3	
300	2.5 YR	6/3	

Core Location - Rock-7-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
5			A Horizon
20	2.5 YR	5/4	
50	2.5 YR	5/4	
100	2.5 YR	6/4	
150	2.5 YR	6/4	
200	2.5 YR	6/4	
300	2.5 YR	6/4	
350	2.5 YR	7/4	
365	2.5 YR	7/4	

Core Location - Rock-8-070898

Sample Depth (cm)	Hue	Value/Chroma	Comments
5			A Horizon
10	2.5 YR	5/4	
25	2.5 YR	5/4	
50	2.5 YR	6/4	
100	2.5 YR	6/4	
165	2.5 YR	6/6	weak Bw
175	2.5 YR	7/6	strong Bw
200	2.5 YR	5/4	
225	2.5 YR	6/6	

Wet and dry rubification values determined in the lab for samples from two cores.

Core Location - Rock-5-070898

Sample Depth (cm)	dry color		wet color		Comments
	Hue	Value/Chroma	Hue	Value/Chroma	
5	10 YR	3/2	10 YR	2/2	A Horizon
20	10 YR	5/3	10 YR	3/2	
75	10 YR	5/4	10 YR	3/4	
100	10 YR	5/6	10 YR	3/4	
125	10 YR	4/2	10 YR	3/2	Paleosol
200	10 YR	5/4	10 YR	3/4	
250	10 YR	5/4	10 YR	3/4	
300	10 YR	5/3	10 YR	3/3	
320	10 YR	4/3	10 YR	2/2	Paleosol
325	10 YR	5/6	10 YR	3/4	

Core Location - Rock-8-070898

Sample Depth (cm)	dry color		wet color		Comments
	Hue	Value/Chroma	Hue	Value/Chroma	
5	10 YR	3/1	10 YR	2/2	A Horizon
10	10 YR	5/4	10 YR	4/2	
25	10 YR	6/3	10 YR	4/2	
50	10 YR	6/3	10 YR	5/2	
100	10 YR	6/2	10 YR	5/3	
165	10 YR	6/6	10 YR	4/4	Paleosol
175	10 YR	5/6	10 YR	4/4	
200	10 YR	6/3	10 YR	4/3	
225	10 YR	6/4	10 YR	4/2	

Appendix G

Geochemistry Results

INAA data for trace elements from a pound core from Briscoe Lake, Washington.
Core Location 44.

Sample No.	Depth (cm)	Na	As	Br	Co	Cr	Fe (%)
RBS-10	10	5327 ± 27	14.1 ± 0.8	51.2 ± 1.4			2.56 ± 0.4
RBS-12	12	6723 ± 34	10 ± 0.9	58.1 ± 1.0			
RBS-14	14	6481 ± 33	8.94 ± 0.5	73.4 ± 2.1		102 ± 40	2.36 ± 0.3
RBS-18	18	5932 ± 30	10.7 ± 0.6	77.1 ± 2.2		82 ± 37	
RBS-20	20	7300 ± 37	14.9 ± 1.3	79.8 ± 1.6			
RBS-22	22	8980 ± 46	8.48 ± 0.5	83.2 ± 2.3			3.92 ± 0.5
RBS-24	24	8840 ± 45	11.6 ± 0.7	96.4 ± 2.7			4.4 ± 0.6
RBS-26	26	14473 ± 74	14.5 ± 0.9	70.9 ± 2.0			5.77 ± 0.8
RBS-28	28	19873 ± 101	8.8 ± 0.8	29.6 ± 0.5			3.16 ± 0.5
RBS-30	30	20528 ± 105	9.95 ± 0.6	20.0 ± 0.6	27.70 ± 6.9		2.02 ± 0.3
RBS-32	32	22247 ± 113	9.27 ± 0.6	16.4 ± 0.5		51 ± 17	2.63 ± 0.4
RBS-34	34	22195 ± 113	8.87 ± 0.5	18.0 ± 0.5	33.40 ± 8.4	118 ± 53	2.97 ± 0.4
RBS-54	54	26020 ± 133	5.8 ± 0.3	4.4 ± 0.1		126 ± 34	2.22 ± 0.3

Sample No.	Depth (cm)	K	La	Lu	Sc	Sm	Yt
RBS-10	10	4400 ± 836	14.1 ± 0.8		8.91 ± 0.4	3.66 ± 0.07	
RBS-12	12	4479 ± 851	15.45 ± 0.9	0.70 ± 0.1	11.50 ± 0.6	4.26 ± 0.08	2.15 ± 0.8
RBS-14	14	4593 ± 873	17.2 ± 1.0	0.38 ± 0.1	12.77 ± 0.6	4.48 ± 0.09	2.37 ± 0.7
RBS-18	18	4964 ± 943	15.73 ± 0.9		10.80 ± 0.5	4.20 ± 0.08	2.95 ± 0.9
RBS-20	20	4778 ± 908	16.56 ± 1.0	0.48 ± 0.1	12.16 ± 0.6	4.50 ± 0.09	2.57 ± 0.8
RBS-22	22	5557 ± 1056	17.1 ± 1.0		12.50 ± 0.6	4.65 ± 0.09	
RBS-24	24	6639 ± 1261	19.0 ± 1.1		13.00 ± 0.7	4.90 ± 0.09	2.93 ± 0.9
RBS-26	26	8493 ± 1614	16.15 ± 1.0		11.40 ± 0.6	4.05 ± 0.08	3.34 ± 1.0
RBS-28	28	11420 ± 2170	15.75 ± 0.5		10.97 ± 0.4	3.17 ± 0.04	2.12 ± 0.7
RBS-30	30	12146 ± 2308	14.85 ± 0.9		9.43 ± 0.5	2.66 ± 0.05	
RBS-32	32	16085 ± 3056	14.7 ± 0.9		9.33 ± 0.5	2.87 ± 0.05	2.51 ± 0.9
RBS-34	34	14549 ± 2764	17.5 ± 1.1		10.05 ± 0.5	3.10 ± 0.06	
RBS-54	54	16562 ± 3147	13.2 ± 0.8		9.44 ± 0.5	2.36 ± 0.04	

Instrumental Neutron Activation Analysis Data

Bromine, Sodium and Potassium

Core Site 11, South Briscoe Lake

Long Beach Peninsula, Washington

Depth (cm)	Br(ppm)	Na (ppm)	K (ppm)
2	90.0 ± 1.3	838.0 ± 5.9	181.8 ± 27.8
6	77.7 ± 0.9	841.5 ± 5.0	915.5 ± 177.0
8	59.8 ± 1.2	931.4 ± 3.8	1284.7 ± 186.8
10	68.8 ± 0.5	836.2 ± 3.3	925.1 ± 130.5
12	123.4 ± 0.8	558.3 ± 2.6	642.0 ± 119.5
14	123.8 ± 0.8	888.6 ± 3.7	141.8 ±
16	118.1 ± 1.0	775.8 ± 4.5	550.6 ± 153.0
18	120.6 ± 1.1	576.9 ± 4.8	401.1 ± 160.7
20	120.8 ± 1.0	674.8 ± 4.0	845.1 ± 147.6
22	126.3 ± 0.8	582.9 ± 4.4	563.0 ± 127.5
24	125.3 ± 1.0	626.4 ± 4.9	427.9 ± 162.0
26	121.3 ± 0.9	577.1 ± 4.4	480.4 ± 110.3
28	136.8 ± 1.4	496.6 ± 4.1	689.4 ± 172.3
30	145.3 ± 1.2	524.9 ± 3.7	
32	137.7 ± 1.1	579.0 ± 5.4	
34	115.5 ± 1.2	729.9 ± 4.7	
36	88.3 ± 0.9	665.9 ± 5.1	808.6 ± 208.7
38	89.1 ± 1.0	703.2 ± 5.2	1238.8 ± 240.6
40	86.2 ± 0.8	685.0 ± 6.9	564.5 ± 171.7
42	90.5 ± 1.2	885.7 ± 8.9	712.9 ± 223.7
44	71.4 ± 1.1	967.9 ± 6.6	750.3 ± 276.7
46	42.9 ± 0.7	890.8 ± 5.4	1203.2 ± 311.6
48	49.0 ± 0.7	898.4 ± 6.4	
50	44.0 ± 0.7	814.3 ± 6.2	1033.6 ± 256.4
52	44.0 ± 0.9	744.2 ± 6.0	
54	42.4 ± 1.0	1224.5 ± 7.7	1045.2 ± 304.6
56	37.3 ± 0.8	1168.5 ± 7.5	
58	44.6 ± 1.0	1800.1 ± 9.4	
60	47.9 ± 1.0	2116.2 ± 11.6	1338.7 ± 440.6
62	56.1 ± 1.1	2450.1 ± 11.5	1810.8 ± 236.1
64	55.0 ± 1.1	3047.7 ± 11.0	2599.4 ± 463.5
66	71.5 ± 1.1	3854.5 ± 10.4	2219.5 ± 238.6
68	61.0 ± 1.1	3050.1 ± 11.3	
70	24.1 ± 0.8	7184.8 ± 15.1	7544.0 ± 657.8

Appendix H

Sand Grain Size Analysis Data

Map showing the locations of the cores from which grain size analyses were performed

Figure ID:
UTM:

153
423220 E
4997582 N

141
423273 E
4997495 N

139

Core Logs from which sand horizons were analyzed for grain size

* = analyzed strata

Core ID: Nesk-7-980708

Nesk-6-980708

Figure ID: 156

159

UTM:

Nesk-7-980708 Core Site 163

Analyzed Horizon = 31-35 cm

Total Mass = 144.55 g

D90= 1.58

Sieve #	Sieve Size (mm)	Phi 4	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
							0
Fines	0.08	3.64	5.7	5.99	0.29	0.20	0.20
140	0.106	3.24	4.88	6.03	1.15	0.79	0.99
100	0.15	2.74	5.55	7.19	1.64	1.13	2.12
80	0.18	2.47	5.71	9.38	3.67	2.54	4.66
65	0.208	2.27	5.07	15.59	10.52	7.28	11.94
60	0.25	2.00	5.68	34.25	28.57	19.76	31.70
50	0.3	1.74	5.55	50.55	45.00	31.13	62.83
45	0.355	1.49	5.64	58.47	52.83	36.55	99.38
		1.00					100

Nesk-6-980708 Core Site 162

Analyzed Horizon = 57-66 cm

Total Mass = 288.20 g

D90= 1.58

Sieve #	Sieve Size (mm)	Phi 4	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
							0
Fines	0.08	3.64	5.7	6.27	0.57	0.20	0.20
140	0.106	3.24	4.88	7.17	2.29	0.79	0.99
100	0.15	2.74	5.55	8.82	3.27	1.13	2.13
80	0.18	2.47	5.71	13.05	7.34	2.55	4.67
65	0.208	2.27	5.07	26.11	21.04	7.30	11.97
60	0.25	2.00	5.68	62.81	57.13	19.82	31.80
50	0.3	1.74	5.55	95.55	90.00	31.23	63.03
45	0.355	1.49	5.64	111.30	105.66	36.66	98.00
		1.00					100

Nesk-6-980708 Core Site 162

Analyzed Horizon = 66-71 cm

Total Mass = 162.27 g

D90= 1.75

Sieve #	Sieve Size (mm)	Phi 4	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
							0
Fines	0.08	3.64	5.7	6.92	1.22	0.75	0.75
140	0.106	3.24	4.88	7.28	2.40	1.48	2.23
100	0.15	2.74	5.55	8.11	2.56	1.58	3.81
80	0.18	2.47	5.71	11.91	6.20	3.82	7.63
65	0.208	2.27	5.07	24.23	19.16	11.81	19.44
60	0.25	2.00	5.68	64.59	58.91	36.30	55.74
50	0.3	1.74	5.55	61.61	56.06	34.55	90.29
45	0.355	1.49	5.64	20.50	14.86	9.16	99.45
		1.00					100

Nesk-6-980708 Core Site 162

Analyzed Horizon = 57-71 cm

Total Mass = 449.57 g

D90= 1.58

Sieve #	Sieve Size (mm)	Phi 4	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
							0
Fines	0.08	3.64	5.7	13.19	1.79	0.40	0.40
140	0.106	3.24	4.88	14.45	4.69	1.04	1.44
100	0.15	2.74	5.55	16.93	5.83	1.30	2.74
80	0.18	2.47	5.71	24.96	13.54	3.01	5.75
65	0.208	2.27	5.07	50.34	40.20	8.94	14.69
60	0.25	2.00	5.68	127.4	116.04	25.81	40.50
50	0.3	1.74	5.55	157.16	146.06	32.49	72.99
45	0.355	1.49	5.64	131.8	120.52	26.81	99.80
		1.00					100

Nesk-10-980707 Core Site 153

Analyzed Horizon = 35-42 cm

Total Mass = 117.34 g

D90 = 1.73

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	6.88	1.18	1.01	1.01
140	0.106	3.24	4.88	6.85	1.97	1.68	2.68
100	0.15	2.74	5.55	8.80	3.25	2.77	5.45
80	0.18	2.47	5.71	13.93	8.22	7.01	12.46
65	0.208	2.27	5.07	25.83	20.76	17.69	30.15
60	0.25	2.00	5.68	47.95	42.27	36.02	66.18
50	0.3	1.74	5.55	34.23	28.68	24.44	90.62
45	0.355	1.49	5.64	15.75	10.11	8.62	99.23
		1					100

Nesk-10-980707 Core Site 153

Analyzed Horizon = 63-71 cm

Total Mass = 178.16 g

D90 = 1.75

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	7.11	1.41	0.79	0.79
140	0.106	3.24	4.88	8.35	3.47	1.95	2.74
100	0.15	2.74	5.55	9.83	4.28	2.40	5.14
80	0.18	2.47	5.71	14.64	8.93	5.01	10.15
65	0.208	2.27	5.07	28.49	23.42	13.15	23.30
60	0.25	2.00	5.68	70.85	65.17	36.58	59.88
50	0.3	1.74	5.55	63.88	58.33	32.74	92.62
45	0.355	1.49	5.64	17.89	12.25	6.88	99.49
		1.00					100

Nesk-10-980707 Core Site 153

Analyzed Horizon = 71-74 cm

Total Mass = 71.15 g

D90 = 1.8

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	6.22	0.52	0.73	0.73
140	0.106	3.24	4.88	6.46	1.58	2.22	2.95
100	0.15	2.74	5.55	7.74	2.19	3.08	6.03
80	0.18	2.47	5.71	11.20	5.49	7.72	13.75
65	0.208	2.27	5.07	19.92	14.85	20.87	34.62
60	0.25	2.00	5.68	37.14	31.46	44.22	78.83
50	0.3	1.74	5.55	17.05	11.50	16.16	95.00
45	0.355	1.49	5.64	8.30	2.66	3.74	98.74
		1.00					100

Nesk-10-980707 Core Site 153

Analyzed Horizon = 63-74 cm

Total Mass = 248.41 g

D90= 1.78

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	13.33	1.93	0.78	0.78
140	0.106	3.24	4.88	14.81	5.05	2.03	2.81
100	0.15	2.74	5.55	17.57	6.47	2.60	5.41
80	0.18	2.47	5.71	25.84	14.42	5.80	11.22
65	0.208	2.27	5.07	48.41	38.27	15.41	26.63
60	0.25	2.00	5.68	107.99	96.63	38.90	65.52
50	0.3	1.74	5.55	80.93	69.83	28.11	93.64
45	0.355	1.49	5.64	26.19	14.91	6.00	99.64
		1.00					100

Nesk-1-980707 Core Site 139

Analyzed Horizon = 65-71 cm

Total Mass = 150.78 g

D90= 1.8

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	7.07	1.37	0.91	0.91
140	0.106	3.24	4.88	8.57	3.69	2.45	3.36
100	0.15	2.74	5.55	9.71	4.16	2.76	6.11
80	0.18	2.47	5.71	16.19	10.48	6.95	13.07
65	0.208	2.27	5.07	26.90	21.83	14.48	27.54
60	0.25	2.00	5.68	62.20	56.52	37.49	65.03
50	0.3	1.74	5.55	49.61	44.06	29.22	94.25
45	0.355	1.49	5.64	13.41	7.77	5.15	99.40
		1.00					100

Nesk-9-980707 Core Site 147

Analyzed Horizon = 56-60 cm

Total Mass = 134.25 g

D90= 1.77

Sieve #	Sieve Size (mm)	Phi	Tare (g)	Total Mass (g)	Mass (g)	% Mass	Cum %
		4					0
Fines	0.08	3.64	5.7	7.82	2.12	1.58	1.58
140	0.106	3.24	4.88	7.79	2.91	2.17	3.75
100	0.15	2.74	5.55	8.70	3.15	2.35	6.09
80	0.18	2.47	5.71	14.56	8.85	6.59	12.69
65	0.208	2.27	5.07	24.69	19.62	14.61	27.30
60	0.25	2.00	5.68	62.10	56.42	42.03	69.33
50	0.3	1.74	5.55	36.58	31.03	23.11	92.44
45	0.355	1.49	5.64	14.89	9.25	6.89	99.33
		1.00					100