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Distribution and sedimentary characteristics of tsunami deposits along the Cascadia margin of western North America

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Abstract

Tsunami deposits have been found at more than 60 sites along the Cascadia margin of Western North America, and here we review and synthesize their distribution and sedimentary characteristics based on the published record. Cascadia tsunami deposits are best preserved, and most easily identified, in low-energy coastal environments such as tidal marshes, back-barrier marshes and coastal lakes where they occur as anomalous layers of sand within peat and mud. They extend up to a kilometer inland in open coastal settings and several kilometers up river valleys. They are distinguished from other sediments by a combination of sedimentary character and stratigraphic context. Recurrence intervals range from 300–1000 years with an average of 500–600 years. The tsunami deposits have been used to help evaluate and mitigate tsunami hazards in Cascadia. They show that the Cascadia subduction zone is prone to great earthquakes that generate large tsunamis. The inclusion of tsunami deposits on inundation maps, used in conjunction with results from inundation models, allows a more accurate assessment of areas subject to tsunami inundation. The application of sediment transport models can help estimate tsunami flow velocity and wave height, parameters which are necessary to help establish evacuation routes and plan development in tsunami prone areas.

Keywords: Tsunami deposits; Cascadia tsunami; Cascadia earthquake; Sedimentary structures; Grain size

1. Introduction

Sedimentary deposits from tsunamis are a valuable source of information for geologists and emergency planners. They provide evidence of past tsunami inundation and provide estimates of tsunami recurrence and magnitude which can be used to improve our understanding of tsunamis and help mitigate the hazards from future tsunamis. They also help evaluate earthquake hazard by providing evidence for past subduction zone earthquakes and may help estimate the magnitude of these earthquakes.

Tsunami deposits are particularly important for reconstructing the history of tsunamis in the Cascadia region of the Pacific Northwest. Cascadia is the region of the Pacific Northwest from Northern California, U.S.A. to northern Vancouver Island, B.C., Canada and is located onshore of the Cascadia Subduction Zone (CSZ)(Fig. 1). There have been no records of a great earthquake (moment magnitude>8) and associated tsunami occurring during the historical period in Cascadia. However, the absence of recorded great earthquakes on the CSZ in the recent past does not necessarily imply that the CSZ is not capable of

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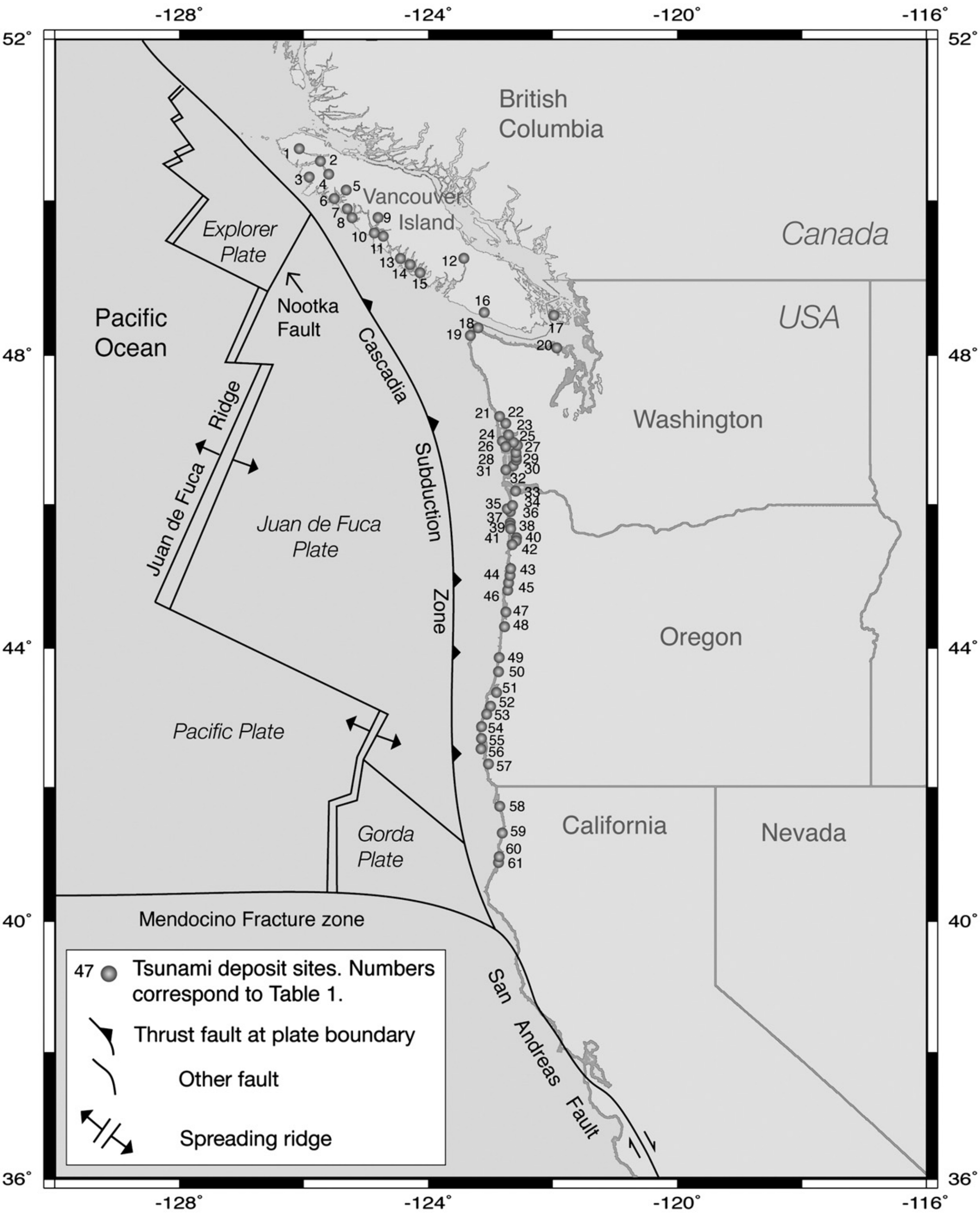

Fig. 1. Map of the Cascadia region of North America, showing the tectonic setting and location of sites where tsunami deposits have been identified.

producing them. The CSZ shares many features with other subduction zones around the world that have produced great earthquakes and tsunamis (Heaton and Kanimori, 1984; McAdoo et al., 2004). Native American oral tradition from the Pacific Northwest describes catastrophic ground shaking and associated coastal flooding occurring in the past (Heaton and Snavely, 1985; Woodward, 1990; Clague, 1995; Hutchinson and McMillan, 1997; Ludwin, 2002). Heaton and Snavely (1985) recognized there was a possible connection between the Native American oral tradition and CSZ earthquakes and tsunamis.

Over the past two decades, abundant geologic evidence has emerged to support the past occurrence of tsunamis and great earthquakes on the CSZ. Evidence that the CSZ produces large earthquakes can be seen in the turbidite record off of the Cascadia Margin (Adams, 1990; Goldfinger et al., 2003), but the evidence that these earthquakes can produce large tsunamis is most easily seen in the stratigraphy of coastal marshes and lakes. Atwater (1987) first published evidence for tsunami deposits in Cascadia in coastal marshes in Washington. Sand layers overlying buried peat were interpreted to be sediments deposited by a tsunami accompanying a great earthquake (Atwater, 1987). Since Atwater's paper was published in 1987, there has been a period of growth in the documentation and interpretation of tsunami deposits in the Pacific Northwest. Numerous studies document tsunami deposits in California, Oregon, Washington, and British Columbia (Peters et al., 2003a,b) (Fig. 2, Table 1). These studies help to increase the awareness of the tsunami hazard along the Cascadia margin.

In this paper, we summarize previously published literature on Cascadia tsunami deposits to characterize their distribution patterns and sedimentary characteristic. We examine how tsunami deposits have been used to assess the recurrence and magnitude of tsunamis in Cascadia. Then we explore the role of tsunami deposits in further mitigating the tsunami hazard for the Cascadia region.

2. Tectonic setting

The Cascadia Subduction Zone extends for about 1000 km from northern California to northern Vancouver Island, British Columbia (Fig. 1). The Juan de Fuca plate, and the smaller Gorda Plate to the south, subduct beneath the North American Plate. To the north, the Nootka fault separates the Explorer plate from the Juan de Fuca plate. The Explorer plate constitutes a broad deformational zone where convergence changes into strike–slip motion between the Pacific plate and the North American plate (Rohr and Furlong, 1995). The Juan de Fuca plate is converging with respect to the North American plate at velocities of 35–45 mm/yr (Wang et al., 1995). Subduction along the CSZ is currently locked (Hyndman and Wang, 1995) and the coast is experiencing interseismic uplift at a rate of 1–4 mm/yr (Wilson, 1993; Hyndman and Wang, 1995).

3. Cascadia tsunami deposit distribution

For this study, we consider Cascadia tsunami deposits to be those formed by a tsunami generated by a CSZ earthquake. Tsunami deposits from other sources are found in the Cascadia region, particularly in the Puget Sound area. These deposits were formed by

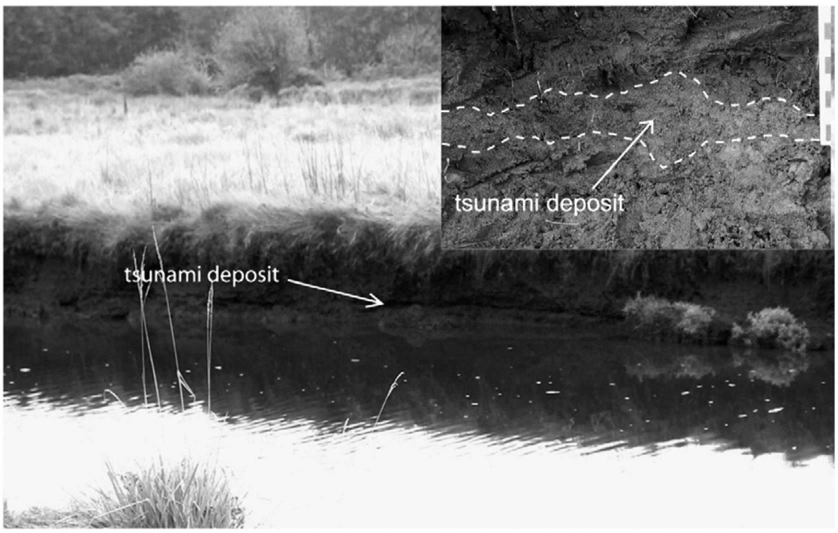

Fig. 2. Tidal marsh at low tide, Neawanna Creek, Oregon. A deposit from the 1700 AD tsunami is exposed in the cutbank as a sand layer that is more deeply eroded than the surrounding mud and peat. Inset shows a detail of the deposit. Scale bar at right of inset is in 1 cm intervals.

Table 1
Tsunami deposit locations and references

#	Site	Reference
1	Koprino Harbour, Vancouver Island, B.C.	Benson et al., 1997
2	Neroutsos Inlet, Vancouver Island, B.C.	Benson et al., 1997
3	Power Lake, Vancouver Island, B.C.	Clague et al., 2000
4	Fair Harbour, Vancouver Island, B.C.	Benson et al., 1997
5	Zeballos, Vancouver Island, B.C.	Bobrowsky and Clague, 1995; Clague et al., 2000
6	Catala Lake, B.C.	Clague et al., 1999
7	Deserted Lake, B.C.	Clague et al., 2000; Hutchinson et al., 2000
8	Louie Bay, Vancouver Island, B.C	Clague et al., 2000; Lopez and Bobrowsky, 2001
9	Channel Lagoon, Vancouver Island, B.C	Clague et al., 2000; Lopez and Bobrowsky, 2001
10	Kanim Lake, Vancouver Island, B.C.	Clague et al., 2000; Hutchinson et al., 1995, 1997
11	Hot Springs Cove, Vancouver Island, B.C.	Lopez and Bobrowsky, 2001
12	Port Alberni, Vancouver Island, B.C.	Clague and Bobrowsky, 1994a; Clague et al., 1994; Huntley and Clague, 1996
13	Kakawis Lake, B.C.	Clague et al., 2000; Lopez and Bobrowsky, 2001
14	Tofino, Vancouver Island, B.C.	Clague and Bobrowsky, 1994a,b; Huntley and Clague, 1996
15	Ucluelet, Vancouver Island, B.C.	Clague and Bobrowsky, 1994a,b
16	Port Renfrew, Vancouver Island, B.C.	Clague et al., 2000
17	Neah Bay, WA	Hutchinson and McMillan, 1997
18	La Push, WA	Hutchinson and McMillan, 1997
19	Discovery Bay	Williams et al., 2005
20	Swantown Marsh, Whidbey Island, WA	Williams and Hutchinson, 2000
21	Copalis River, WA	Atwater, 1992; Reinhart, 1991
22	North Bay, Grays Harbor, WA	Reinhart, 1991
23	Johns River, Grays Harbor, WA	Reinhart, 1991; Shennan et al., 1996
24	Grayland Plains, WA	Schlichting, 2000; Schlichting et al., 1999
25	North River, Willipa Bay, WA	Reinhart, 1991; Reinhart and Bourgeois, 1987
26	Smith River, Willipa Bay, WA	Reinhart, 1991
27	Cedar River, Willipa Bay, WA	Reinhart and Bourgeois, 1987
28	Bone River, Willipa Bay, WA	Reinhart, 1991; Reinhart and Bourgeois, 1987
29	Niawiakum River, Willipa Bay, WA	Atwater, 1987; Reinhart and Bourgeois, 1987; Reinhart, 1991; Hemphill-Haley, 1995; Atwater, 1996, Atwater and Hemphill-Haley, 1997
30	Palix River, Willipa Bay, WA	Reinhart and Bourgeois, 1987; Reinhart, 1991
31	Nemah River, Willipa Bay, WA	Reinhart and Bourgeois, 1987
32	Long Beach Peninsula, WA	Schlichting et al., 1999; Schlichting, 2000
33	Youngs Bay, Columbia River, OR	Peterson et al., 1993
34	Stanley Lake, OR	Peterson et al., 1993; Darienzo et al., 1994; Fiedorowicz, 1997; Fiedorowicz and Peterson, 2002
35	Necanicum River, OR	Darienzo et al., 1994; Fiedorowicz, 1997; Fiedorowicz and Peterson, 2002
36	Neawanna Creek, OR	Darienzo, 1991; Peterson et al., 1993; Darienzo et al., 1994; Darienzo and Peterson, 1995; Fiedorowicz, 1997; Fiedorowicz and Peterson, 2002
37	Ecola Creek, OR	Peterson et al., 1993; Darienzo and Peterson, 1995
38	Cannon Beach, OR	Peterson et al., 1993; Schlichting et al., 1999
39	Rockaway, OR	Schlichting et al., 1999; Schlichting, 2000
40	Oyster Farm, Netarts Bay, OR	Darienzo, 1991
41	Wee Willies, Netarts Bay, OR	Darienzo, 1991; Peterson et al., 1993
42	Netarts Marsh, OR	Peterson and Darienzo, 1988; Darienzo and Peterson, 1990; Darienzo, 1991; Peterson et al., 1993; Darienzo and Peterson, 1995; Shennan et al., 1998
43	Nestucca Bay, OR	Darienzo, 1991; Darienzo and Peterson, 1994; Peterson et al., 1993; Darienzo and Peterson, 1995
44	Neskowin, OR	Schlichting et al., 1999; Schlichting, 2000
45	Salmon River, OR	Grant and McLaren, 1987; Minor and Grant, 1996; Atwater et al., 1999; Nelson et al., 2004
46	Silitz Bay, OR	Darienzo, 1991; Peterson et al., 1993; Darienzo et al., 1994; Darienzo and Peterson, 1995
47	Yaquina, OR	Darienzo, 1991; Darienzo et al., 1994; Darienzo and Peterson, 1995; Peterson and Priest, 1995
48	Alsea Bay, OR	Darienzo, 1991; Darienzo and Peterson, 1995; Peterson and Darienzo, 1996
49	Lily lake, OR	Briggs and Peterson, 1992
50	Suislaw Bay, OR	Briggs, 1994
51	Umpqua River, OR	Briggs, 1994
52	Coos Bay, OR	Briggs, 1994; Nelson et al., 1998b
53	Coquille River, OR	Witter, 1999; Witter et al., 2003

(continued on next page)

Table 1 (*continued*)

#	Site	Reference
54	Bradley Lake, OR	Kelsey et al., 1994, 1995, 1998a; Nelson et al., 1998a; Ollerhead et al., 2001; Hemphill-Haley and Lewis, 2003; Kelsey et al., 2005
55	Sixes River Estuary, OR	Kelsey et al., 1993, 1994; Witter and Kelsey, 1994; Kelsey et al., 1995; Kelsey et al., 1998b
56	Elk River, OR	Witter and Kelsey, 1994
57	Eucre Creek, OR	Witter and Kelsey, 1994, 1996; Witter et al., 2001
58	Crescent City, CA	Carver et al., 1996; Garrison et al., 1997
59	Lagoon Creek, CA	Garrison et al., 1997; Abramson, 1998; Garrison-Laney, 1998
60	South Humboldt Spit, CA	Leroy, 1999
61	Eel River, CA	Li, 1992

tsunamis generated by earthquakes occurring on other faults in the region and are not included in this study. Tsunami deposits from tsunamis generated by earthquakes on distant faults also occur in the region, most notably deposits from the 1964 Alaska tsunami that was generated by a great earthquake in the Prince William Sound region of Alaska (Clague et al., 1994; Fiedorowicz and Peterson, 2002). These deposits sometimes are found at the same sites as Cascadia tsunami deposits and are included in this study for comparison with Cascadia tsunami deposits.

More than 60 sites containing potential or confirmed tsunami deposits have been identified and published along the Cascadia margin (Fig. 1, Table 1). Sedimentary deposits from tsunamis are found at numerous sites along the coastlines of Oregon and Washington. The southernmost sites occur in northern California in the region onshore of the Mendocino Triple Junction. The northernmost sites are found along the west coast of Vancouver Island, British Columbia (Peters et al., 2003a,b). Gaps are apparent in the distribution of tsunami deposits along the Cascadia coastline, particularly along the northern Washington coast. While it is difficult to assess the record of tsunami inundation for these areas, it is important to note that the absence of tsunami deposits along any particular stretch of coastline does not necessarily imply that tsunamis have not occurred at these locations in the past. The gaps may be due to conditions unfavorable to the deposition, preservation, or identification of tsunami deposits. For example, a coastline characterized by high steep cliffs may contain few areas where tsunami sediments may be deposited and any sediments deposited may have little chance of being preserved.

Tsunami deposits are preserved in three main depositional settings in Cascadia: estuaries, coastal lakes, and back-barrier wetlands. Many rivers in Cascadia have long estuaries with tidal marshes along their banks (Fig. 2). These low-lying tidal mashes are subject to tsunami inundation and have a good potential to preserve tsunami sediments. The potential for preservation of the deposit is enhanced if the marsh undergoes coseismic subsidence at the time of the tsunami. Subaerially exposed portions of the marsh that received only limited or episodic sedimentation prior to subsidence would be tidally submerged for greater periods of time, allowing deposition of fine sediment to rapidly bury and preserve the tsunami sediments. Recognition of tsunami deposits in a tidal marsh is facilitated by the contrast between the sand typical of tsunami deposits and the peat or mud characteristic of tidal marsh sedimentation. Low-lying coastal lakes in Cascadia also have a high potential to preserve tsunami deposits. High sedimentation rates and minimal erosion in coastal lakes increase the preservation potential of tsunami deposits in these settings. Coastal lakes are characterized by the accumulation of gyttja, an organic-rich mud that contrasts with the sandy tsunami deposits. Back-barrier wetlands are often located behind dune ridges along open coastal areas of Cascadia. Back-barrier marshes can preserve deposits from tsunamis that overtop or breach dune ridges. As in tidal marshes and coastal lakes, the sand of the tsunami deposit contrasts with the peat and mud typical of these wetlands.

Each depositional setting has a characteristic stratigraphy. In tidal marshes, peat develops from subaerially exposed, well-vegetated marsh soils and mud is deposited along tidal channels in the intertidal zone. Tsunami deposits in tidal marsh settings are usually seen as layers of sand overlying peat. If the tsunami is associated with coseismic subsidence, mud may overlie the sand layer (Atwater, 1987; Nelson et al., 1996) (Fig. 3A). Preservation of rooted plant material beneath the sand deposit indicates sand deposition occurred soon after subsidence, consistent with tsunami deposited sand (Atwater and Yamagucci, 1991). Lesser degrees of subsidence or tsunami deposition over marsh substrate that is not fully developed into peat may result in less pronounced differences between overlying and underlying material. A tsunami deposit overlying a buried marsh deposit that has not coseismically subsided may show no change between the overlying and underlying material and may indicate a tsunami originating from a

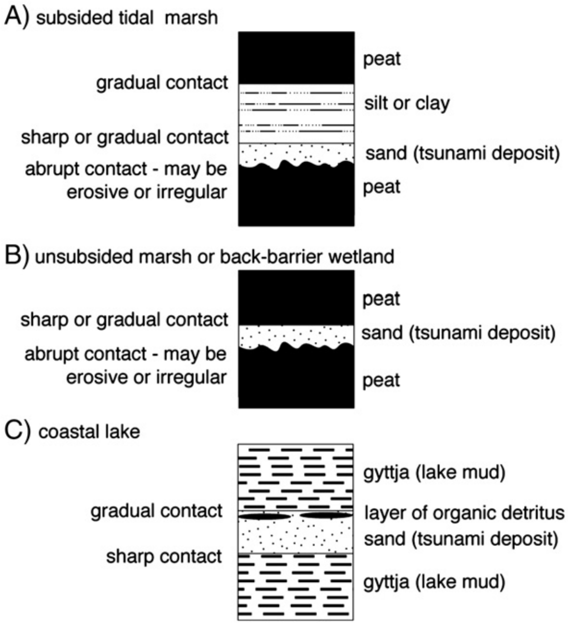

Fig. 3. Idealized stratigraphic relationships of Cascadia tsunami deposits in (A) a tidal marsh that has undergone coseismic subsidence, (B) a back-barrier wetland or tidal marsh that has not undergone coseismic subsidence, and (C) a coastal lake.

distant earthquake source or a tsunami of non-seismic origin (Darienzo et al., 1994) (Fig. 3B).

Tsunami deposits in lakes usually consist of a bed of sand layered above and below by gyttja (Hutchinson et al., 1997; Clague et al., 1999; Hutchinson et al., 2000) (Fig. 3C). A layer of organic debris and/or a massive mud may overlie the sand layer. Tsunami deposits in back barrier wetlands are characterized by tsunami-deposited sandwiched between layers of peat or mud, resembling the stratigraphy of a tidal marsh that has not undergone coseismic subsidence (Fig. 3B). Coastal lakes and back-barrier marshes are protected from tidal influences, resulting in little or no evidence for coseismic subsidence. In these settings, determining the source of the tsunami may be difficult (Clague, 1997).

4. Characteristics of Cascadia tsunami deposits

Sedimentary characteristics of tsunami deposits are important, both to determine a sedimentary origin for the deposits and to provide information about the tsunami that deposited them. Tsunami deposits reflect the high-energy conditions of the tsunami and often provide a sharp contrast to the low-energy deposits that characterize tidal marsh deposition. Considerable variation may be seen laterally within a single deposit, between correlated deposits at different sites and between deposits from separate events at a single site.

4.1. Grain size

The range of grain sizes found in Cascadia tsunami deposits is dependent on the available source material. Cascadia tsunami deposits are typically composed of medium to fine sand, silt, and clay. The source material for the sand is usually the beach and near shore. In deposits along stream banks, a portion of the sand fraction may be made up of sand entrained from stream banks and channels. Sediment eroded from hill slopes may also make up some of the coarse fraction of the deposit. The primary source for silt and clay in the deposits is mud from the tidal channels and tidal flats. Mud is often present between layers in the deposits (e.g. Benson et al., 1997; Atwater and Hemphill-Haley, 1997). Mud may also be present at the base of the deposit. Some Cascadia tsunami deposits fine landward, indicating decreasing flow velocity as the tsunami approaches the limit of inundation (Abramson, 1998; Williams and Hutchinson, 2000).

Cascadia tsunamis are also capable of transporting very large clast sizes, up to cobble size (e.g. Nelson et al., 1998b; Hutchinson et al., 2000). Coarse clasts have been found at the base of the deposit (e.g. Clague and Bobrowsky, 1994a) or at the base of layers within the deposit (e.g. Witter and Kelsey, 1996). Gravel is found adjacent to tidal channels in deposits at Fair Harbour, B.C. (Hutchinson et al., 1997).

4.2. Grading

Deposits from Cascadia tsunamis tend to be normally graded when grading is apparent. Normal grading is consistent with deposition by sediment settling out of suspension. One model for tsunami sediment transport involves a tsunami entraining sediment near the coast and carrying it inland. When the tsunami nears the limit of inundation, flow velocities become low enough that the entrained sand grains fall out of suspension, creating a tsunami deposit (Jaffe and Gelfenbaum, 2002). Normal grading is sometimes observed within the individual layers of the deposit (Witter and Kelsey, 1996; Hutchinson et al., 1997; Witter, 1999; Williams and Hutchinson, 2000; Kelsey et al., 2002). Often, however, no grading is apparent in Cascadia tsunami deposits. This may reflect uniformity in the sediment source, large flow velocities relative to the grain size of the transported sediment, or indicate other sediment transport processes are responsible for the deposit.

4.3. Thickness and geometry

Cascadia tsunami deposits usually form sand sheets that thin landward and away from channels (e.g. Atwater, 1987). The thickness of Cascadia tsunami sediments reported in the literature range from less than a centimeter to 68 cm. The sand sheets are often continuous, but near the landward edge, deposits may be discontinuous and form sand lenses. Tsunami deposits often pinch out near the forest edge at the back of marshes. Thinning of deposits up river is usually gradual and may indicate a gradual decrease in water depth inland. Along the Niawiakum River in Washington, deposits thin from 5 cm 1 km upriver from the river mouth to 0.5 cm over 3 km from the river mouth. At a site 4.5 km from the river mouth there is no tsunami-deposited sand present, though diatom evidence suggests tsunami inundation extended this far (Hemphill-Haley, 1995). In coastal lakes, deposits may be thickest near a seaward outlet, such as at Kanim Lake on Vancouver Island, British Columbia (Hutchinson et al., 1997).

Tsunami deposit thickness is not always a direct analog for the magnitude of the tsunami that created the deposit. The thickness of a tsunami deposit may be determined by several factors, including water depth, flow velocity, flow deceleration, clast size, sediment supply, and topography. The thickest deposits usually occur locally in depressions or around obstructions. At Port Alberni, British Columbia, the 1964 Alaska tsunami left a 15 cm thick deposit behind a pipeline (Clague and Bobrowsky, 1994b) while away from the pipeline, the tsunami deposited a relatively uniform sand sheet 1–2 cm thick (Clague and Bobrowsky, 1994b).

4.4. Layering

Many Cascadia tsunami deposits contain several distinct layers. Layers may extend throughout the deposit or be evident only locally. Where layers are present, 2–5 layers are typically reported, though up to 10 layers are observed in deposits from Crescent City, California (Carver et al., 1996). On deposits along the Niawiakum, Bone, and Palix Rivers in Washington, the number of layers decrease upstream (Reinhart, 1991). Layers range in thickness from a few mm to 11 cm. Pebbles are sometimes present at the base of layers (Witter and Kelsey, 1994) and a thin layer of mud may be present at the base of layers and between layers (Reinhart and Bourgeois, 1987; Reinhart, 1991; Hutchinson et al., 1997, 2000; Williams and Hutchinson, 2000). In some Cascadia tsunami deposits, layers of sand and mud alternate, forming sand–mud couplets. The deposition of sand–mud couplets may be attributed to deposition by separate waves (Reinhart, 1991). The argument for multiple waves is strengthened when the sand layers are normally graded, suggesting that sand settles out of suspension with each wave. Mud may be deposited from standing water between waves.

4.5. Sedimentary contacts

Sedimentary contacts are important because they indicate changes in the pattern of sedimentation. The basal contacts of most Cascadia tsunami deposits are abrupt, indicating a rapid change in the depositional pattern consistent with the sudden influx of sediment from a tsunami. Upper contacts range from gradual to abrupt. The upper contact reflects patterns in post-tsunami deposition.

4.6. Fossils

Both macrofossils and microfossils have been reported from Cascadia tsunami deposits. Macrofossils consist of marine mollusks and various forms of plant fossils. Marine mollusks are rare in Cascadia tsunami deposits but have been reported from deposits on Vancouver Island, British Columbia (Benson et al., 1997; Clague et al., 1999; Hutchinson et al., 2000). Plant fossils usually consist of plant material transported from either the nearby forest or marshes, or underlying marsh material. Plant detritus includes branches, twigs, bark, conifer needles, seeds, and cones. Logs are sometimes present in the deposits (Clague and Bobrowsky, 1994b). Leaf bases of *Triglochin maritimum* and tufts of *Deschampsia caespitosa*, vegetation common in Cascadia tidal marshes, are found at the base of some deposits (Darienzo et al., 1994; Atwater and Hemphill-Haley, 1997). Plant material may form a layer of detritus that caps the deposit or layers within the deposit (e.g. Witter and Kelsey, 1996; Hutchinson et al., 2000). Plant detritus may also be disseminated throughout the deposit (Schlichting, 2000). At Kanim Lake, plant material coarsens and increases in abundance towards the base of the deposit (Hutchinson et al., 1997).

Tsunami deposits from Cascadia typically contain a suite of marine, brackish, and fresh water microfossils. The most studied microfossils found in Cascadia tsunami deposits are diatoms and foraminifers. The greater than normal percentage of marine and brackish water microfossils in an otherwise freshwater environment provides evidence for a marine incursion. Specific diatom species have a narrow range of salinity, substrate, and intertidal exposure conditions in which they live. Changes in microfossil assemblages help estimate coseismic subsidence and link a tsunami deposit to a Cascadia earthquake source (Hemphill-Haley, 1995). Diatoms have been used along the Niawiakum River, Washington, to extend the limit of inundation beyond what can be determined from sedimentary deposits alone (Hemphill-Haley, 1995).

Flop-overs (vegetation bent in the direction of flow) are sometimes found in Cascadia tsunami deposits. Although preservation of flop-overs is not common, they provide one of the few clear indicators of tsunami flow direction that is preserved in tsunami deposits. Tufts of the marsh grasses *D. caespitosa* and *Potentilla pacifica* bent over in the direction of flow are present at the base of deposits along the Niawiakum River in Washington (Atwater and Hemphill-Haley, 1997). Flop-overs from these deposits indicate a landward-directed flow (Atwater and Hemphill-Haley, 1997), and suggest that at this site, deposition occurred on the uprush of the tsunami.

4.7. Rip-up clasts

Rip-up clasts are fragments of a cohesive substrate contained within a sedimentary deposit. They indicate high-energy flows and also suggest that the material was not worked for periods of time long enough to break apart the material into individual grains. In Cascadia tsunami deposits, rip-up clasts are usually composed of peat or mud. The authors observed rip-up clasts in tsunami deposits from cutbank exposures along the Niawaikum River, Washington and Neawanna Creek, Oregon, and in cores from back-barrier marshes at Neskowin, Oregon. Rip-up clasts were also reported from tsunami deposits at Lagoon Creek, California (Abramson, 1998), Eucre Creek, Oregon (Witter et al., 2001), Sixes River, Oregon (Kelsey et al., 2002), Bradley Lake, Oregon (Kelsey et al., 1995), and Copalis River, Washington (Reinhart, 1991).

5. Recurrence

Understanding tsunami recurrence intervals is critical for evaluating the risk of a tsunami impacting the Cascadia coastline. Cascadia tsunami deposits often contain organic material that can be used for radiocarbon dating. Precise and accurate dates are often difficult to determine using radiocarbon dating. The typical error reported for Cascadia tsunami deposit ages using radiocarbon techniques is on the order of ±100–200 years, which does not allow discrimination of tsunamis closely spaced in time. Dendrochronology can be used to date trees killed by coseismic subsidence into the tidal zone and can constrain radiocarbon dates to within tens of years (Atwater and Yamagucci, 1991). However, sites where this technique can be used are rare in Cascadia. It is important to note that radiocarbon dates represent a maximum possible age for Cascadia tsunami deposits because older, reworked, organic material may be present within the deposit. A maximum age may also be obtained by dating organic material from peat directly underlying the tsunami deposit. The peat represents the marsh surface prior to the tsunami. Other dating techniques used in Cascadia include ^{137}Cesium and optical dating methods. Deposits from recent historical tsunamis, such as the 1964 Alaska earthquake tsunami, may be indistinguishable from those of the more recent prehistoric tsunamis, such as the 1700 event, based on radiocarbon dates alone. ^{137}Cs dating allows discrimination of these historical tsunamis from older prehistoric events by indicating sediments deposited after the onset of nuclear testing in the 1950s (Clague et al., 1994; Benson et al., 1997). Optical dating techniques allow dating samples when organic material is not present in sufficient quantity for radiocarbon methods (Ollerhead et al., 2001; Huntley and Clague, 1996).

Using a combination of high-precision radiocarbon techniques and tree-ring growth patterns, Atwater et al. (1991) constrained the ages of a subsidence horizon associated with tsunami deposits to within a few decades of 1700 AD. Japanese historical records report that a tsunami inundated portions of Japan in January of 1700 with no associated earthquake or other local cause (Satake et al., 1996, 2003). The most likely source for this tsunami was a great earthquake on the CSZ. Based on the time required for tsunami propagation from Cascadia to Japan, the time and date of the earthquake was estimated to be 21:00 on January 26, 1700, local time in Cascadia (Satake et al., 1996). Many coastal sites throughout Cascadia contain a sand layer approximately 300 years old that is consistent with deposition by a 1700 AD tsunami (Fig. 4).

Many sites also contain tsunami deposits older than the 1700 AD event (Fig. 4). The time period represented by the marsh and lake stratigraphy along the Cascadia margin varies with location. The length of the record and time period recorded in the stratigraphy is determined by the age of the marsh or lake. Some sites have records in excess of 6000 years. Kelsey et al. (2002) record 11 tsunamis over the last 5500 years at Sixes River estuary, Oregon. At Coquille River, Oregon, deposits record 11 tsunamis over 6700 years (Witter et al., 2003). Deposits at Bradley Lake, Oregon record 13 tsunamis over

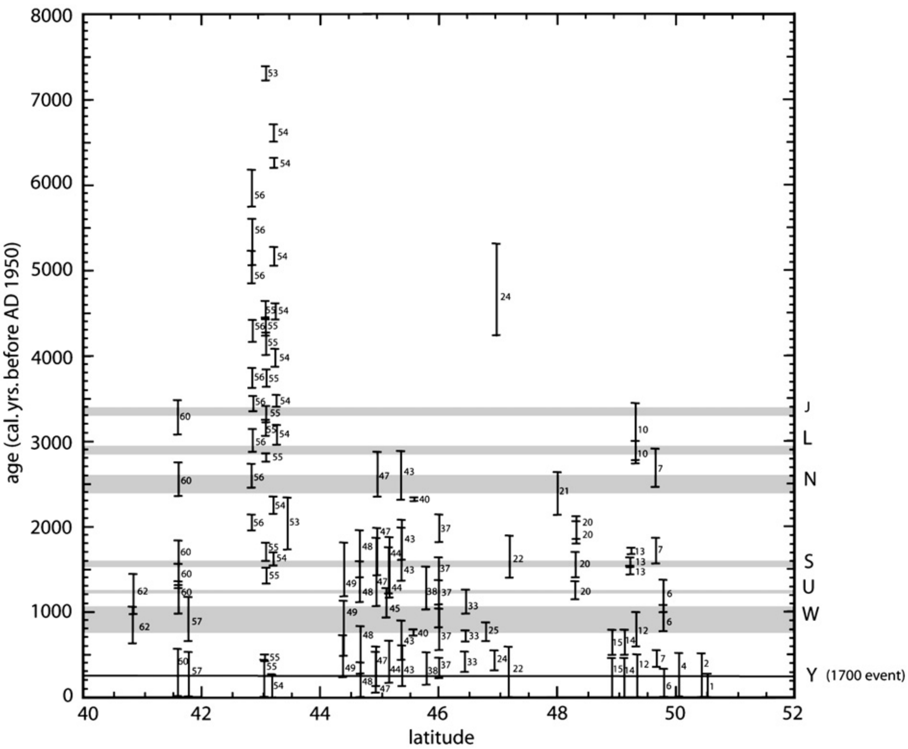

Fig. 4. Calibrated age ranges, in calendar years before 1950, reported for Cascadia tsunami deposits. Letters on the right correspond to age ranges for Cascadia Subduction Zone earthquakes based on the paleoseismic and tsunami records of SW Washington (Atwater and Hemphill-Haley, 1997). The age ranges (gray bars) correspond to the revised chronology for these events reported in Atwater et al. (2003).

7300 years (Kelsey et al., 1998a). Kakawis Lake, British Columbia has a 14,000 year-long record, yet deposits from only 3–4 tsunamis were identified there, suggesting tsunamis from other events may not have penetrated inland far enough to reach the lake (Lopez and Bobrowsky, 2001). While it is often possible to correlate deposits between cores and outcrops at a single site using radiocarbon ages and deposit depth, ambiguities in stratigraphic relationships and error in radiocarbon age make it difficult to correlate tsunami deposits with any certainty across the entire Cascadia margin.

The tsunami record preserved in coastal sediments varies throughout Cascadia (Fig. 4). At some sites there are tsunami deposits that don't correlate in time with other Cascadia tsunami deposits or with Cascadia Subduction Zone earthquake chronologies, while at other sites, more tsunami deposits are found in a given time interval (Atwater et al., 2003; Goldfinger et al., 2003). McAdoo and Watts (2004) suggest that landslide tsunamis may account for anomalous tsunami deposits found locally that are not present throughout Cascadia. Cascadia earthquakes may differ in the length and location of the rupture or in the amount of slip and this could account for differences in tsunamis and, therefore, in the deposit record. Segmentation of the CSZ may produce differing records on either side of the segment boundary (Darienzo and Peterson, 1995; Kelsey et al., 2005). Differences in the record may also be a function of tide level, sediment supply or coastal configuration at the time of the tsunami. The tsunami deposit record suggests that each tsunami affects the region uniquely and that it may be difficult to characterize a typical Cascadia tsunami.

Several attempts have been made to establish tsunami and CSZ earthquake recurrence intervals based on a combination of tsunami and subsidence events. Darienzo and Peterson (1990) calculated an average recurrence interval of 600 years based on tsunami deposits and subsidence events at Netarts Marsh, Oregon. Darienzo and Peterson (1995) calculated an average recurrence interval for CSZ earthquakes and tsunamis of 200–600 years, based on seven estuaries along 175 km of the northern Oregon coast. Atwater and Hemphill-Haley (1997) calculated an average recurrence interval for CSZ earthquakes of 500–540 years based on subsidence events along the Niawiakum River in Washington. An average recurrence interval of about 500–600 years is typical, but the records from each site do not agree precisely. The minimum and maximum time intervals expected between Cascadia tsunamis can also be estimated from the intervals between individual tsunamis. The interval between individual Cascadia tsunamis varies widely. At Netarts Marsh, Oregon, recurrence intervals ranged from less than 300 years to over 1000 years for 6 tsunamis in 3600 years of depositional record (Darienzo and Peterson, 1990).

6. Tsunami magnitude

Tsunami deposits provide important information on the magnitude of a Cascadia tsunami. Current tsunami hazard assessment practices typically rely on numerical models to assess the magnitude of a Cascadia tsunami. However, the accuracy of these models to predict tsunami magnitude may be limited by uncertainties in the inputs to these models, such as deformation at the source, size of the rupture zone, bathymetry, and topography. Information on the magnitude of past tsunamis preserved in tsunami deposits can augment tsunami modeling efforts and help evaluate model results.

6.1. Run-up and inundation distance

Tsunami run-up and inundation distance, two measures for assessing tsunami hazard, may be estimated from Cascadia tsunami deposits. Run-up is defined as the elevation of the tsunami relative to sea level at maximum inundation and inundation distance is the inland extent of tsunami penetration. However, studies of modern tsunamis document that the maximum elevation reached by a tsunami and the maximum extent of inundation is often beyond the maximum extent of sedimentation and sedimentation may be discontinuous near the limit of inundation (Jaffe et al., 2003; Gelfenbaum and Jaffe, 2003). Erosion of parts of the deposit may also result in underestimation of run-up and inundation distance. Therefore, the highest elevation and farthest distance inland a tsunami deposit is found represents a minimum estimate of the run-up and inundation distance of the tsunami. For paleotsunami studies, run-up elevations and inundation distances must be corrected using the history of uplift and subsidence subsequent to a tsunami.

Cascadia tsunami deposits have been reported at elevations up to 6 m above present mean sea level at Kanim Lake, British Columbia (Hutchinson et al., 1997). Run-up may also be estimated by the height of the barrier a wave must overtop to leave a deposit beyond it. This is not true run-up because it does not occur at maximum inundation but does give an indication of elevations reached by the tsunami. The maximum barrier elevation overtopped by a Cascadia tsunami is 8 m at Long beach Peninsula, Washington (Schlichting, 2000). Coastal uplift or subsidence that occurred during the time between deposition and elevation measurements of the deposit or barrier may have a significant effect on run-up determinations from paleotsunami deposits. When current estimates of uplift or subsidence rates are known, correcting for coastal uplift or subsidence improves run-up estimates. Changes in the height of barrier dunes over time may also influence elevation estimates.

Tsunami deposits are found up to 1 km inland on the open coast and may be found several km inland up rivers and estuaries. Tsunami deposits at Sevenmile Creek along the Coquille River estuary are 8 km from the shoreline (Witter, 1999) and at Young's Bay along the Columbia River are 10 km upriver (Peterson et al., 1993). Tsunami deposits along the Bone, Niawiakum, and Palix rivers near Willapa Bay, Washington reach up to 3 km inland (Reinhart and Bourgeois, 1987; Reinhart, 1991; Atwater and Hemphill-Haley, 1997), and along the Niawiakum River, marine diatoms have been found overlying buried soils up to 4 km inland, 1 km beyond the limit of tsunami sedimentation (Hemphill-Haley, 1995).

6.2. Shoreline change

Interpretation of tsunami deposits may be complicated by post-depositional changes in shoreline position. The Washington and Oregon shoreline is known to have been prograding at about 0.5–2.0 m/yr for the last 1000 years and up to 20–30 m/yr since the construction of jetties around 1900 (Buijsman et al., 2003). On a stable coastline, the inundation distance is the distance between the shoreline and the limit of inundation. However, where a coastline has prograded, the distance between the present shoreline and the limit of inundation is greater than the inundation distance at the time of the tsunami. Where there has been net erosion, the distance between the present shoreline and the limit of inundation is less than the inundation distance at the time the tsunami occurred.

On Long Beach Peninsula, Washington, erosional scarps in barrier dunes, delineated using ground-penetrating radar, were used in conjunction with estimates of coseismic subsidence to estimate paleoshorelines (Schlichting, 2000). The maximum inland extent of deposits from the 1700 tsunami was 800 m from the present shoreline but 300 m from a subsurface erosional scarp interpreted as the paleoshoreline. (Schlichting, 2000; Peterson et al., 2002). The distance from the modern shoreline to the maximum inland extent of deposits from an event that occurred 1100 years ago was 1500 m from the present shoreline and 1000 m from the erosional scarp correlated with the paleoshoreline at the time of this event (Schlichting, 2000).

6.3. Effect of tides on interpreting tsunami deposits

The height of the tide at the time of a tsunami can have a significant effect on the run-up and inundation. For many areas of Cascadia, a tidal range of 3 m or greater is not unusual. In addition, atmospherically-induced subtidal fluctuations in sea level may be on the order of 0.5 m (Mofjeld et al., 1997). For most Cascadia tsunami deposits, there is no dating method accurate enough to determine the time of inundation, making it impossible to determine the tide level when the tsunami arrived. However, for the 1700 AD tsunami, tides can be determined because the exact time and date have been determined for this event. Mofjeld et al. (1997) hindcast tides to determine that the 1700 Cascadia tsunami occurred during a low neap tide. Water depth, run-up and inundation distance would have been greater if the 1700 tsunami had occurred during a perigean high tide.

7. Hazard mitigation

Tsunami deposits have been successfully used to identify the tsunami hazard along the Cascadia margin. Tsunami deposits can be used to further mitigate this hazard. Including tsunami deposits on inundation maps provides an independent measure of potential inundation by a tsunami, supplementing predictions from a tsunami inundation model. While modeling has its limitations as a method of predicting tsunami inundation zones, so do tsunami deposits. Tsunami deposits are only found in locations inundated by the tsunami that were favorable to their deposition and preservation, and coastal conditions at the time of deposition my be different than current conditions. Tsunamis likely inundated low-lying areas not represented by the deposits and current coastal configuration may offer some areas protection while other areas may be more exposed. When incorporating tsunami deposits on inundation maps, it should be emphasized that the lack of a tsunami deposit does not mean that an area is safe from tsunami inundation, nor does the presence of a tsunami deposit mean that an area will be inundated in the next tsunami. However, taken together, modeling and deposits provide a better assessment of tsunami hazard than either method alone. Tsunami deposits have been included on inundation maps for the Washington and Oregon coasts (e.g. Priest et al., 1997; Walsh et al., 2000). In most cases, tsunami deposits were found in areas where tsunami inundation was predicted by the model. In some instances, such as at Hoquiam on Grays Harbor, WA, cores that did not contain tsunami deposits were taken in areas where the model predicted tsunami inundation (Walsh et al., 2000). In other places, tsunami deposits were found beyond the inundation zone predicted by the model (Walsh et al., 2000).

Another technique that shows great potential is to use data from tsunami deposits in sediment transport models to derive estimates of flow velocity and water depth. Estimates of flow velocity and water depth are important to engineers to evaluate the susceptibility of structures to damage from a tsunami and to planners to evaluate evacuation routes. Inundation models can estimate these quantities, but are limited by uncertainties in the fault parameters of the earthquake source, as well as model errors, and data errors (Myers et al., 1999). Sediment transport models offer an alternative way of estimating these quantities. Sediment transport models calculate the flow velocity or the depth of the flow using the thickness and grain size distribution of a tsunami deposit (Jaffe and Gelfenbaum, 2002). This technique has shown promising results for recent tsunamis (e.g. Papua New Guinea) and has potential for use with paleotsunamis. Sediment transport models can also predict morphological changes to estuaries as a result of a tsunami or coseismic subsidence (Gelfenbaum and Lessar, 2003).

8. Conclusions

Tsunami deposits provide evidence of past tsunami along the Cascadia margin and reinforces arguments that the CSZ is capable of producing great earthquakes. They have been identified at more than 60 locations, most commonly in low-energy coastal environments such as tidal marshes, back-barrier marshes, and coastal lakes, where identification is easiest and preservation potential is greatest. The tsunami deposits are usually composed of fine to medium sand and may contain larger clasts up to cobble size. These coarser sediments provide a sharp contrast to the peat and mud within which they lie. Tsunamis are capable of transporting large clasts and grain size is largely determined by what sediment is available for transport. Normal grading found in some of the deposits and is consistent with deposition by sediment settling out of suspension. Layers consisting of sand–mud couplets may represent deposition by separate waves of the tsunami.

Multiple tsunami deposits found at some sites indicate that as many as 13 tsunamis have affected the coast along the Cascadia margin over the last 7300 years. Differences in the tsunami records preserved at each site, both temporally and spatially, reflect the unique character of each tsunami, and may be attributed to tsunami source differences, coastal configuration, tide level, and sediment supply. The most recent tsunami associated with a great earthquake on the CSZ occurred just over 300 years ago, on January 26, 1700 AD. Average recurrence intervals of 500–600 are typical for the Cascadia Margin but the length of time between events ranges from about 300 years to over 1000 years. Cascadia tsunami deposits may extend up to a kilometer inland in open coastal settings and up to several kilometers inland in river valleys and have been found up to 6 m above present mean sea level. These figures represent minimum estimates of the distance and elevation Cascadia tsunamis have inundated.

The information gained from the interpretation of tsunami deposits has been used to help mitigate future tsunami hazard. The inclusion of Cascadia tsunami deposits on inundation maps can aid in the interpretation of these maps and, used in conjunction with results from inundation models, allow a more accurate assessment of tsunami hazard. Sediment transport modeling has great potential for providing estimates of tsunami flow velocity and wave height necessary to develop structural guidelines in tsunami-prone areas.

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