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Natural Hazards
Journal of the International Society for the Prevention and Mitigation of Natural Hazards

ISSN 0921-030X

Nat Hazards
DOI 10.1007/s11069-013-0622-1

Springer

Nat Hazards
DOI 10.1007/s11069-013-0622-1

ORIGINAL PAPER

Recurrence intervals of major paleotsunamis as calibrated by historic tsunami deposits in three localities: Port Alberni, Cannon Beach, and Crescent City, along the Cascadia margin, Canada and USA

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Received: 3 December 2012 / Accepted: 26 February 2013

Abstract We review geologic records of both historic and prehistoric tsunami inundations at three widely separated localities that experienced significant damage from the 1964 Alaskan tsunami along the Cascadia margin. The three localities are Port Alberni, Cannon Beach, and Crescent City, representing, respectively, the north, central, and south portions of the study area (1,000 km in length). The geologic records include anomalous sand sheets from marine surges that are hosted in supratidal peaty mud deposits. Paleotsunami sand sheets that exceed the thickness, continuity and/or extent of the 1964 historic tsunami are counted as major paleotsunami inundations. Major paleotsunamis (6–7 in number) at each locality during the last 3,000 years demonstrate mean recurrence intervals of 450–540 years, and within-cluster intervals (three events each) of 270–460 years. It has been 313 years since the last major paleotsunami from a great Cascadia earthquake in AD 1700. We compare the dated sequences of major paleotsunami inundations to the nearest regional records of coastal coseismic subsidence in Willapa Bay in the central margin, Waatch/Neah Bay in the northern margin, and Coquille in the southern margin. Similar numbers of events from both types of records suggest that the major paleotsunamis are locally derived (near-field) from ruptures of the Cascadia margin megathrust fault zone, rather than from transoceanic tsunamis (far-field) originating at other subduction zones around the Pacific Rim. Given the catastrophic hazard of the near-field Cascadia margin tsunamis, we propose a basic rule for reminding the general public of the need for self-initiated evacuation following a great earthquake at the Cascadia margin.

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Published online: 14 March 2013

Keywords Major tsunamis · Recurrence interval · Cascadia margin · Warning · Evacuation

1 Introduction

In this paper, we report on the recurrence intervals of major paleotsunamis along the Cascadia margin extending from southwest Canada to northwest USA. The oceanic Juan de Fuca plate is subducting under the North American plate along a megathrust fault that has not ruptured in historic time (Fig. 1). Prehistoric ruptures of the megathrust fault have produced geologic records of abrupt coastal subsidence (Darienzo et al. 1994; Atwater and Hemphill-Haley 1997), paleotsunami run-up (Clague et al. 2000; Schlichting and Peterson 2006), and offshore coseismic turbidite deposition (Adams 1996; Goldfinger et al. 2008). The different geologic records yield somewhat different numbers of rupture events during the last several thousand years (Atwater and Griggs 2012). It is not known whether every rupture of the Cascadia megathrust produces a major tsunami or whether ruptures at other Pacific Rim subduction zones have produced major tsunamis at the Cascadia margin.

Canadian and US public perceptions of Cascadia tsunami hazards (Fig. 1) are becoming confused with reported hazards from far-field tsunamis generated at other

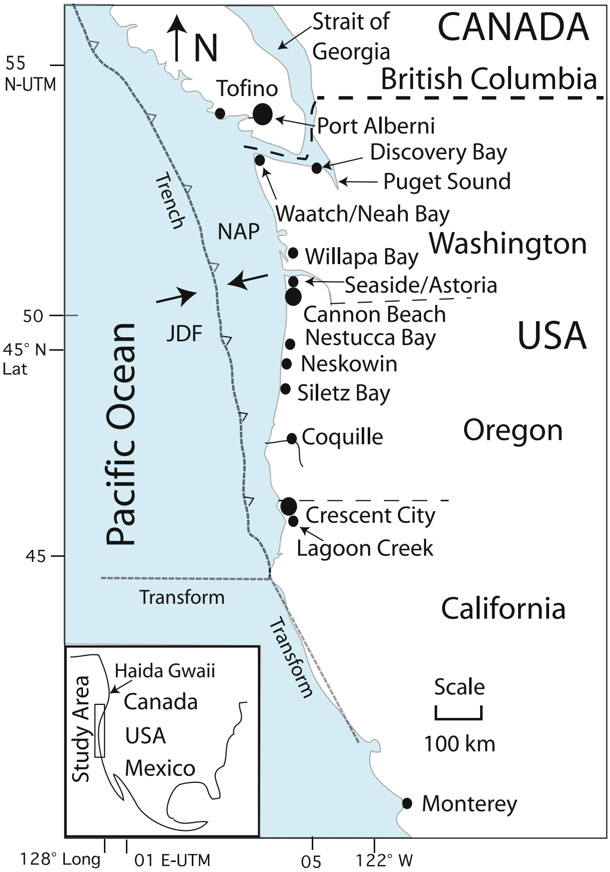

Fig. 1 Location of the three study localities—Port Alberni, Cannon Beach, and Crescent City (*large solid circles*), which are used to establish the recurrence intervals of major paleotsunamis at the Cascadia margin. The other localities that are shown (*small solid circles*) have provided geologic records of megathrust rupture and/or historic tsunami run-up values. The Juan de Fuca Plate (JDF) is subducting at the buried trench under the North American Plate (NAP) at an estimated convergence rate (*opposing arrows*) of about 4 cm year^{-1}

subduction zones around the Pacific Rim. Far-field tsunami hazards are emphasized by tsunami warning agencies that have developed transoceanic tsunami buoy systems (NOAA 2012a). The most recent far-field tsunami threat from the Tohoku Japan rupture (magnitude M_w 9.0) in March 2011 was widely broadcast by the local news media, but it did not materialize in most localities along the Cascadia margin. Coastal residents and tourists are now unsure about the need to evacuate prior to officially-authorized warnings (Seaside Tsunami Advisory Group 2011). This uncertainty was further highlighted by the confusion at the community level in British Columbia that immediately followed the M_w 7.5 earthquake on the Queen Charlotte Fault off Haida Gwaii, British Columbia, on October 27, 2012. The Pacific and West Coast Tsunami Warning Center issued a tsunami warning immediately after the earthquake. However, the British Columbia Government did not pass the warning on to coastal communities that might be affected until nearly an hour later. By that time, the tsunami, although small, had already reached many communities. In the case of a great Cascadia earthquake (M_w 8.5–9.0) and near-field major tsunami with a 15–30 min transit time to the Pacific coastline (Meyers 1994), there would not be sufficient time to wait for authorized warnings before initiating life-saving evacuation.

In this paper, we evaluate the threat posed by tsunamis to the southwest Canadian and northwest US coastlines by estimating major tsunami recurrence. We have selected three localities for the analysis of major tsunami recurrence intervals. The analysis is based on (1) the occurrence of paleotsunami sand deposits at the three localities and (2) the use of sand deposits from the 1964 Gulf of Alaska far-field tsunami at the same locations for purposes of prehistoric tsunami calibration.

The three locations are Port Alberni, British Columbia (Clague and Bobrowsky 1994a, b), Cannon Beach, Oregon (Peterson et al. 2008), and Crescent City, California (Carver 2000; Peterson et al. 2011), representing, respectively, the north, central, and south Cascadia margin regions (Fig. 1). All three localities experienced significant damage from the historic 1964 far-field tsunami. They each contain late Holocene sand sheet records of overland paleotsunami inundation of at least 200 m in distance. The inland extent, continuity, and/or thickness of paleotsunami sand sheets must exceed those of the 1964 far-field tsunami in corresponding core sites in order to qualify as what we call major tsunamis. Minor storm surges (about 1–2 m run-up) in the cold coastal waters of the study area (Peterson et al. 2008) do not reach the necessary heights or durations to produce sand sheets equivalent to 1964 historic tsunami at the three localities.

Recurrence intervals of major tsunami inundations at Port Alberni, Cannon Beach, and Crescent City (Fig. 1) are based on (1) mean recurrence intervals of events during the past 3,000 years, (2) recurrence intervals within apparent event clusters, and (3) the shortest recurrence intervals between the geologically recorded events. We compare the dated sequences of major tsunamis to dated sequences of coseismic subsidence events (Atwater et al. 2004), which are proxies for megathrust ruptures of the central Cascadia margin. We also compare records of major tsunami recurrence at the Cascadia margin to potential far-field tsunami excitation from great subduction zone earthquakes throughout the Pacific Rim (Landers et al. 1993).

The established recurrence intervals of major paleotsunami inundations at the three study localities should help refocus public attention on catastrophic inundations from near-field tsunamis produced at the Cascadia margin. We also provide an example of a basic rule for self-initiated evacuation on the open coast following a coseismic rupture of the Cascadia megathrust fault.

2 Results

2.1 Geologic records of paleoseismic recurrence intervals along the Cascadia margin

The longest continuous geologic record of episodic coseismic subsidence along the central Cascadia margin is from Nestucca Bay, Oregon (Fig. 1) (Darienzo et al. 1994). Abrupt upcore decreases in peat content suggest at least 12 events of abrupt marsh burial by intertidal mud during the past 6.5 ± 0.2 ka (Fig. 2). The cycles of gradual coastal uplift and abrupt subsidence represent, respectively, interseismic strain accumulation and coseismic strain release, as recorded landward of the strongly coupled portion of the subduction zone. Well-documented records of episodic coseismic subsidence are reported from Willapa Bay, Washington, where nine subsidence events are recorded in the past 5.0 ± 0.4 ka (Atwater et al. 2004). The longer records of episodic wetland subsidence along the Cascadia margin estuaries, such as those at Nestucca and Willapa Bays, generally occur beyond the landward extents of most paleotsunami sand sheets.

More seaward core sites in smaller tidal bays, such as in Siletz Bay, Oregon (Fig. 1) record both coseismic subsidence and associated paleotsunami inundation (Fig. 3), although record lengths are relatively short (1–2 ka). It is difficult to evaluate the magnitude or run-up height of paleotsunami inundation in the tidal bays due to uncertainties about tidal inlet geometry, tidal channel morphology, and/or the tide level during the paleotsunamis.

Paleotsunami records along the Cascadia margin are also available from supratidal settings including back-barrier wetlands, such as Neskowin, Oregon (Peterson et al. 2009), bay-head deltas, such as Discovery Bay, Washington (Williams et al. 2005), and

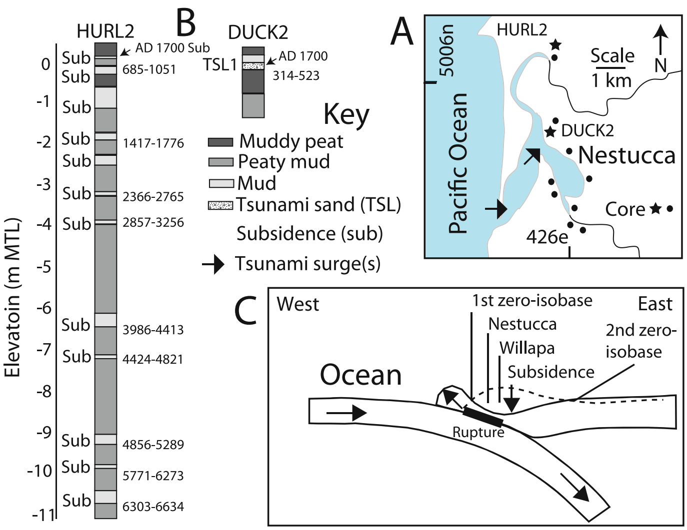

Fig. 2 **a** and **b** Nestucca Bay core sites including (*1*) distal tidal marsh site (HURL2) containing a long continuous record of episodic abrupt coastal subsidence (Sub) but no tsunami sand layers and (*2*) a proximal marsh site (DUCK2) with a record of only one subsidence event (AD 1700), but is directly associated with a tsunami sand layer (TSL). Elevation is relative to mean tidal level (MTL) and bulk radiocarbon (cal years BP 2-sigma) predate the subsidence events. **c** The positions of Nestucca and Willapa bays relative to the trough of coseismic subsidence (after Peterson et al. 2012)

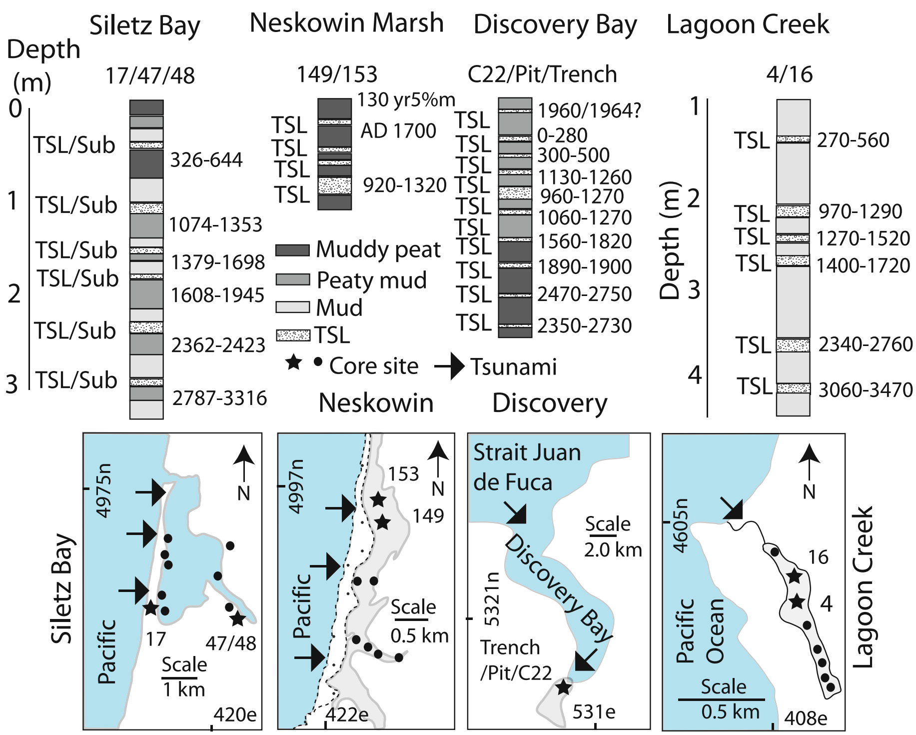

Fig. 3 Core sites (*circles*) indicating coseismic subsidence (Sub) at Siletz Bay and tsunami sand layers (TSL) in Siletz Bay, Neskowin, Discovery Bay, and Lagoon Creek along the Cascadia margin. Core logs are from numbered sites (*stars*) and are previously published. See Fig. 1 for locations along the Cascadia margin. Discovery Bay includes both coastal and inshore (Strait of Georgia and Puget Sound) paleotsunami events (Williams et al. 2005). Radiocarbon age ranges are in calibrated years before present with 2-sigma uncertainty (Stuiver et al. 2012)

beach-ridge barrage lagoons, such as Lagoon Creek, California (Abramson 1998; Garrison-Laney 1998; Carver et al. 1998) or low elevation coastal lakes (Hutchinson et al. 2000; Fig. 3). The supratidal settings generally lack high-resolution records of coseismic subsidence, but they do establish minimum run-up heights for paleotsunamis. The minimum run-up heights are based on overtopping elevations of beach barriers, delta plains, and/or alluvial plains (Peterson et al. 2008). Minimum run-up thresholds of 200 m overland inundation across subratidal settings (at least 3 m elevation) are necessary to discriminate between sustained paleotsunami inundations and other flooding events from storm surf, wind or coesimic seiching, and/or other disturbances in coastal water bodies and wetlands. The discrimination between major tsunami and minor tsunami inundation is based on comparisons of paleotsunami sand sheets against historic tsunami sand sheets in the same localities, as described later in this article.

2.2 Historic records of far-field tsunami run-ups along the Cascadia margin

The Cascadia margin is directly exposed to far-field tsunamis that cross the northeast Pacific Ocean from active subduction zones located at the north and west margins of the Pacific Ocean, as well as from Central and South America (Fig. 4). At least 18 far-field

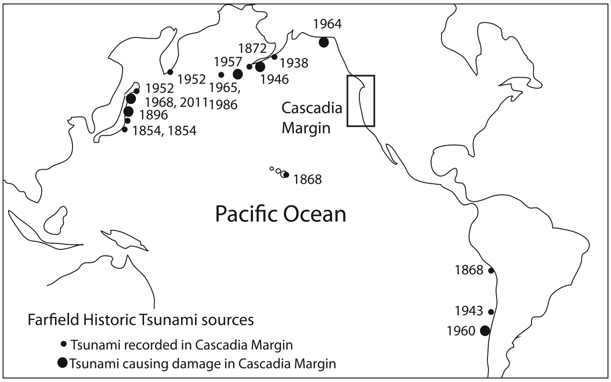

Fig. 4 Sources and dates of historic far-field tsunami (*small black circles*), as recorded on the US West Coast portion of the Cascadia margin. Data are updated to 2012 from Landers et al. (1993) and NOAA (2012b). Six events that caused damage at the Cascadia margin (*large black circles*) include Aleutian (1946, 1957), Chile (1960), Alaska (1964), and Japan (1896, 2011) tsunamis

tsunamis reached the US portion of the Cascadia margin from AD 1854–2012 (Landers et al. 1993; NOAA 2012b). Only six of the historic far-field tsunamis caused reported damage in this area. The largest and most destructive of these historic far-field tsunamis was from the 1964 Gulf of Alaska giant earthquake (M_w 9.2) (Table 1). Run-ups measured for the 1964 far-field tsunami in Port Alberni, Cannon Beach, and Crescent City reached 6.0, 3.5, and 4.9 m, respectively. The more distant far-field tsunami from the 1960 Chile M_w 9.5 earthquake had run-up heights of less the half of those from the Alaskan 1964 tsunami along the Cascadia margin. The recent 2011 Japan M_w 9.0 rupture produced run-ups on the order of 0.8, 0.2 and 2.5 m at Port Alberni, Cannon Beach, and Crescent City.

2.3 Historic Alaskan 1964 Far-field tsunami at three Cascadia localities

The exceptional run-up (6 m in height) of the 1964 tsunami at Port Alberni, British Columbia (Table 1), results from wave amplification at the terminal end of the Alberni Inlet, a narrowing fiord on the west coast of Vancouver Island (Fig. 5; Clague and Bobrowsky 1994a). The maximum water height during the 1964 tsunami in Port Alberni was 7 m NAVD88, where NAVD88 is an elevation datum at about one meter below mean sea level in the region. Tsunami surge(s) of 2–3 m above the tidal marsh at the mouth of the Somass River deposited 1–3 cm thickness of fine sand at core sites in the marsh and damaged an elevated pipeline that crossed the marsh area. Run-up of three surges along the east side of the Somass River caused extensive damage to about 240 residences and dock facilities, with total direct losses of 5–10 million in 1964 Canadian dollars (ca. 50–100 million in 2012 dollars) (Clague and Bobrowsky 1994a). The maximum run-up extent of the 1964 tsunami at Port Alberni is not addressed in this paper.

The 1964 tsunami at Cannon Beach, Oregon, comprised two surges that overtopped the south barrier beach ridge (Fig. 5), producing a modest run-up of 3.5 m above the predicted tide level (Table 1). The maximum water height was between 4.5 and 5.0 m elevation NAVD88. A sand layer, 1–3 cm thick, was deposited by the tsunami up to 200 m landward of the sand spit ridge crest. Tsunami sand was also deposited 100–150 m northeast of the lower Ecola Creek bank in supratidal spruce bogs at elevations of 3.5–4.0 m. The 1964 tsunami destroyed the Ecola Creek bridge and transported its support beams and roadway materials

Table 1 Historic tsunami run-up records from selected Pacific west coast sites

Tsunami sources and N. Pacific west cost sites	Run-up (m)	Measurement type
1960 Chile M_w 9.5		
Tofino (BC)	1.2	Tide gauge
Port Alberni (BC)	–	
Neah Bay (WA)	0.4	Tide gauge
Seaside (OR)	1.5	Observed
Cannon Beach (OR)	1.5	Observed
Crescent City (CA)	1.7	Tide gauge
Monterey (CA)	1.3	Tide gauge
1964 Alaska M_w 9.2		
Tofino (BC)	2.4	Observed
Port Alberni (BC)	6.0	Observed
Neah Bay (WA)	0.7	Tide gauge
Seaside (OR)	3.0	Observed
Cannon Beach (OR)	3.5	Observed
Crescent City (CA)	4.9	Observed
Monterey (CA)	1.4	Observed
2011 Japan M_w 9.0		
Tofino (BC)	–	
Port Alberni (BC)	0.76	Tide gauge
Neah Bay (WA)	0.43	Tide gauge
Astoria (OR)	0.18	Tide gauge
Cannon Beach (OR)	–	
Crescent City (CA)	2.47	Tide gauge
Monterey (CA)	0.70	Tide gauge

Run-up data for historic tsunamis in 1960 and 1964 for the US west coast and the Canadian coast are from Landers et al. (1993), Wigen (1964), and Public Safety Canada (2008). Run-up data for the 2011 tsunami are from NOAA (2012c)

Localities are identified by province or state: *BC* British Columbia, *WA* Washington, *OR* Oregon, and *CA* California (Fig. 1)

Historic tsunami run-up (m) is measured relative to predicted tide level, that is, the observed height of the maximum water level above the predicted tide level for the time of observed maximum surge (Landers et al. 1993)

Water heights are established from continuously operating tide gauges or from water line marks that were surveyed into benchmark elevations or a local tidal datum, shortly after the tsunami

about 600 m upstream from its position near the mouth of the tidal creek. Several motel units were moved off their foundations and carried 0.5 km into the back-barrier marsh and/or were rammed by drift log debris. Parts of the ocean beach seawall were washed away. Primary damage from the 1964 tsunami in the small Cannon Beach community was estimated to be 250,000 in 1964 US dollars (ca. 3 million in 2012 dollars) (Landers et al. 1993).

The 1964 tsunami at Crescent City overtopped a low beach ridge at 4-m elevation and produced a run-up of greater than 3.5 m, above the predicted tide at the shoreline (Fig. 5; Magoon 1966; Landers et al. 1993). A nearby tide gauge in the harbor recorded a maximum run-up of 4.9 m above the predicted tide in the harbor there (Table 1). The maximum flood height in Crescent City was about 6.5 m elevation. A sand sheet, 1–3 cm thick, was

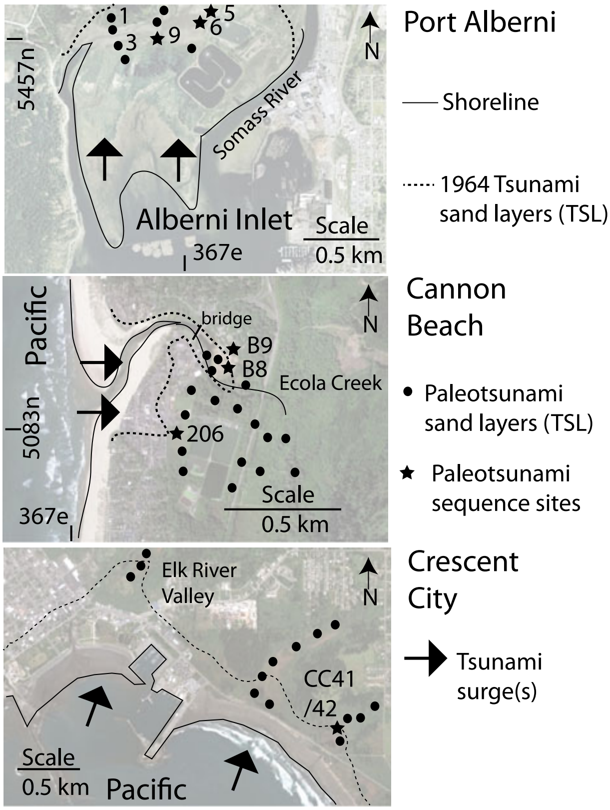

Fig. 5 Map of core sites where evidence of paleotsunami deposition has been found (*solid circles*). *Solid stars* are sites with paleotsunami radiocarbon ages. Approximate extents of the historic 1964 tsunami sand sheets (*dotted lines*) are shown for Cannon Beach and Crescent City, relative to adjacent core sites with paleotsunami sand layers

deposited by the 1964 tsunami up to 200 m inland from the beach ridge, south of Crescent City (Peterson et al. 2011). Three or four tsunami surges damaged 172 commercial buildings and 91 homes in 29 city blocks. Damage was concentrated in the downtown area, southeast of Elk River Valley and along the harbor waterfront. Twenty-one boats sank in the small harbor. Fires spread to large petroleum tanks, which exploded in succession. About 1,000 cars, shattered buildings, and other debris were swept inland by the tsunami surges, which peaked just after midnight. Total direct damage was estimated to be $11–16 million (ca. 120–250 million in 2012 dollars) (Landers et al. 1993).

We use the documented 1964 Alaskan far-field tsunami surges and associated sand sheet deposits at Port Alberni, Cannon Beach, and Crescent City to calibrate the scale of paleotsunami inundations, as recorded by the paleotsunami sand sheets, at the same localities. The paleotsunami sand sheets that exceed the 1964 tsunami deposits, in thickness, continuity, and/or landward extent, represent major tsunami inundations, as defined for this study. Several core sites each in Port Alberni, Cannon Beach, and Crescent City are used to establish minimum thresholds for major paleotsunami inundations in each of the three communities as described below.

2.4 Major paleotsunami sand layers in cores

Tidal marshes lie on a small delta plain at Port Alberni (Fig. 5) at the end of Alberni Inlet (Clague and Bobrowsky 1994a). The tidal marshes record anomalous sand and gravelly sand layers that have been traced landward in intertidal mud and peaty mud deposits (Fig. 6). Continuous drill core from one site contains nine paleotsunami sand layers below the 1964 sand layer that range from 5 to 50 cm thick and date back to ~3.5 ka ago. The last Cascadia earthquake (M_w ~ 9) in AD 1700 (Satake et al. 1996) generated a tsunami that deposited the uppermost sand layer below the 1964 tsunami layer, and it is bracketed by radiocarbon ages of 148 and 479 cal years BP. Detailed descriptions and extents of the tsunami sand layers at Port Alberni are presented elsewhere (Clague and Bobrowsky 1994a).

The wetlands at Cannon Beach, Oregon, formed behind a small barrier spit by about 3.0 ka (Peterson et al. 2008) (Fig. 5). Supratidal peaty mud deposits host seven pre-1964 paleotsunami sand sheets, 5–40 cm thick, at proximal sites within 500 m distance landward of the beach ridge (Fig. 6). The youngest paleotsunami sand layer immediately postdates a peat layer that is 310–510 years old, thereby establishing it as the deposit of the AD 1700 Cascadia tsunami. Two paleotsunami sand layers form a couplet that is between 740–910 and 1060–1290 years old. A particularly thick paleotsunami sand layer lies above a thin peat that is 1276–1411 years old. The landward extents of the four youngest paleotsunami sand sheets in the back-barrier wetlands range from 500 to 1,500 m distance from the ocean shoreline. The specific inundation distances and run-ups (10–15 m in height) of the major paleotsunami inundations in Cannon Beach are discussed elsewhere (Peterson et al. 2008; Peterson and Cruikshank 2011).

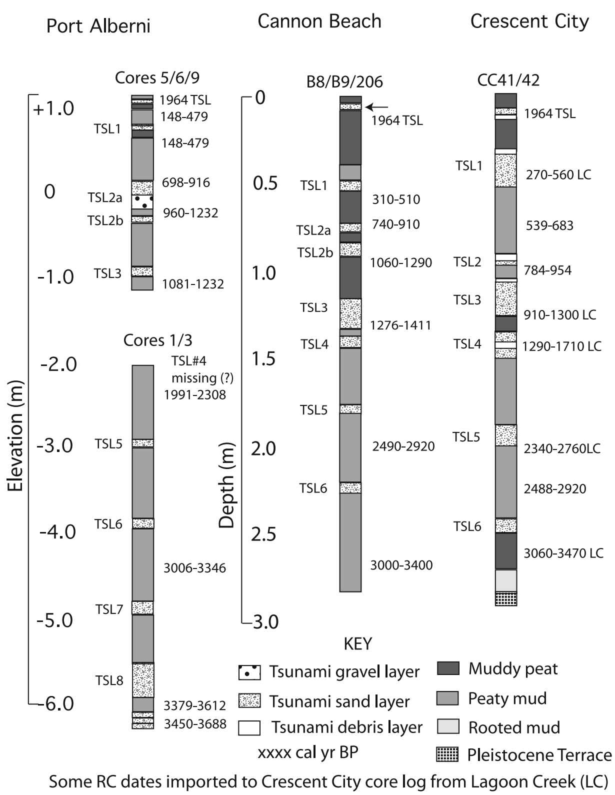

Fig. 6 Paleotsunami sand sheet records in representative cores from Port Alberni (Clague and Bobrowsky 1994a), Cannon Beach (Peterson et al. 2008), and Crescent City (Peterson et al. 2011) in the north, central, and south Cascadia margin, respectively. Radiocarbon ages have been calibrated with the CALIB_v5 or v6 curve with 2-sigma analytical error (Stuiver et al. 2012). Core locations are shown in Fig. 5

Paleotsunami sand layers in the wetlands at Crescent City (Fig. 5) are behind a stable beach ridge that has been in place since about 3 ka (Peterson et al. 2011). The small beach ridge within the Crescent City Harbor embayment lies between two rocky headlands. The wetlands, including shallow barrage ponds, contain peaty mud deposits that host six paleotsunami sand sheets, 5–50 cm thick, below the 1964 tsunami sand layer (Fig. 6). The age of the youngest paleotsunami layer is constrained by an age of 270–560 years from an underlying peat. The youngest paleotsunami layer is correlated to the last Cascadia tsunami event of AD 1700. Radiocarbon dating of several of the thicker sand sheets at Crescent City proved to be problematic. The large, high-velocity tsunami surges, estimated to have 6–8 m water column depths (Peterson et al. 2011), eroded underlying peats, and remobilized older woody debris into the sand sheets. For some of these older events, we have used the nearby Lagoon Creek paleotsunami sand layer dates of Abramson (1998) and Garrison-Laney (1998). Details on sand sheet deposits and the maximum recorded run-ups ($\sim$10 m in height) in Crescent City are presented elsewhere (Peterson et al. 2011).

3 Discussion

3.1 Major paleotsunami recurrence intervals at the Cascadia margin

The numbers of prehistoric sand sheets deposited within the past 3,000 years at Port Alberni, Cannon Beach and Crescent City are six, seven, and six, respectively (Fig. 6). For this initial analysis, we rigorously count the number of evident sand sheets at each site, but later, we argue that one additional event was missed at Port Alberni. We also count the doublet tsunami layers, TSL2a and TSL2b, as two separate paleotsunami events, although in some localities they have been previously identified as one event.

The mean recurrence intervals for the six or seven paleotsunamis at the three localities are 540 and 450 years, respectively. Each of the sand sheets exceeds the thickness, continuity, and/or landward extent of the corresponding sand sheets produced by the 1964 Alaskan far-field tsunami, so we define them here as major tsunamis. We expect the run-ups of major paleotsunamis at the three localities reached more than 6–7 m elevation, based on comparison with the 1964 tsunami run-ups and existing stable shorelines. Issues of beach barrier stability, paleosea-level elevations, tidal range, and available sand supply, which could influence tsunami sand sheet deposition, are addressed in previous papers that are specifically concerned with estimated paleotsunami runup heights on the north margin (Clague et al. 2000; Hutchinson et al. 2000; Peterson et al. 2013), central margin (Peterson et al. 2008; Peterson and Cruikshank 2011), and south margin (Carver et al. 1998; Peterson et al. 2011).

Based on the extent of paleotsunami sand sheets in Port Alberni, Cannon Beach, and Crescent City (Fig. 5), we assume that high-velocity inundations probably exceeded several hundred meters from the shorelines. Substantial infrastructure damage and associated drowning deaths would be expected in the high-velocity inundation zones from future major tsunamis, as represented by the paleotsunami sand sheets (Peterson et al. 2011).

It appears from the records that clusters of 3–4 major events occurred prior to the AD 1700 paleotsunami at Port Alberni, Cannon Beach, and Crescent City (Fig. 6). The latest three-event clusters are bracketed by radiocarbon ages for each locality as follows: 698–916 and 1981–1232 years BP at Port Alberni; 740–910 years BP and 1060–1290 years BP at Cannon Beach; and 784–954 years BP and 1290–1710 years BP at Crescent City/Lagoon Creek. If, for the sake of argument, we take the largest spread (2-sigma errors) of the bracketing ages and divide it by three events each (three events, two recurrence intervals), we obtain cluster recurrence intervals of 267, 275, and 463 years for Port Alberni, Cannon Beach, and Crescent City, respectively. These are the most conservative or longest recurrence intervals for the event clusters based on the 2-sigma error calibrated radiocarbon ages. If we use midpoints of the bracketing ages, rather than the largest spreads, the recurrence intervals for the clustered events are even less.

The shortest recurrence intervals between two dated events at each locality, again using the largest spread brackets (2-sigma error), at Port Alberni, Cannon Beach, and Crescent City are 272, 351, and 516 years, respectively. The last Cascadia rupture at AD 1700 occurred 313 years ago, which is close to the within-cluster recurrence intervals directly dated at Port Alberni and Cannon Beach.

3.2 Comparisons of regional coseismic subsidence and major paleotsunami inundations

We now compare dated sequences of paleotsunami inundations at Cannon Beach, Port Alberni, and Crescent City to the nearest corresponding records of regional coseismic subsidence at Willapa Bay, Waatch–Neah Bay, and Coquille, respectively (Fig. 1). Willapa Bay, which is 75 km north of Cannon Beach, contains evidence of six subsidence events within the past 2.8 ka (Fig. 7; Atwater et al. 2004). Willapa Bay is located well landward of the 1st zero-isobase (Fig. 2) so it should reflect elastic response of the upper plate to regional ruptures of the megathrust along the central Cascadia margin (Peterson et al. 2012). With the exception of one paleotsunami sand sheet (TSL2a), the dated sequence of major paleotsunami inundations at Cannon Beach is similar to the dated sequence of regional coseismic subsidence events over the same time period at Willapa Bay.

Post-glacial isostatic rebound has greatly limited the age of tidal marsh records on the west coast of Vancouver Island, located west of Port Alberni (Fig. 1) (Clague and Bobrowsky 1994b; James et al. 2000). For example, Tofino, British Columbia, records only the last Cascadia subsidence event and associated paleotsunami at AD 1700 (Clague and Bobrowsky 1994b; Clague et al. 2000). Cores from Port Alberni do not record any subsidence events (Fig. 6), suggesting that Port Alberni is located near the 2nd zero-isobase in the North Cascadia margin (Figs. 1, 2). The Waatch/Neah Bay wetlands, located 100 km south of Port Alberni, do provide a short record of three or four coseismic subsidence events and associated paleotsunami inundations extending back to ~1.3 ka (Fig. 7) (Peterson et al. 2013). That sequence is similar to the last four major paleotsunami sand sheets dated in the Port Alberni cores 5/6/9.

A subtle subsidence event Sub2a tentatively identified in the Waatch wetlands (Fig. 7) needs to be confirmed elsewhere, but a northern Cascadia rupture, bracketed by dates of 680 and 1060 years BP in the Waatch wetlands, could account for the second oldest paleotsunami sand layer (TSL2a) observed at Port Alberni and Cannon Beach. Clague and Bobrowsky (1994a) have suggested an alternative far-field source, the Alaskan subduction zone (Combellick 1991), for the tsunami inundation event dated 698–916 years BP at Port Alberni. Underwater seismites or debris flows in deep basin cores taken offshore of Vancouver Island do record a local shaking event at 0.7–0.9 ka (Blais-Stevens et al. 2011), but the specific rupture source is unknown.

A substantial time gap between the core transects 5/6/9 and 1/3 at Port Alberni (Fig. 6) may miss a potential major tsunami inundation (TSL4) that is recorded at Cannon Beach between dates of 1411 and 2490 years BP. A regional rupture of the central Cascadia

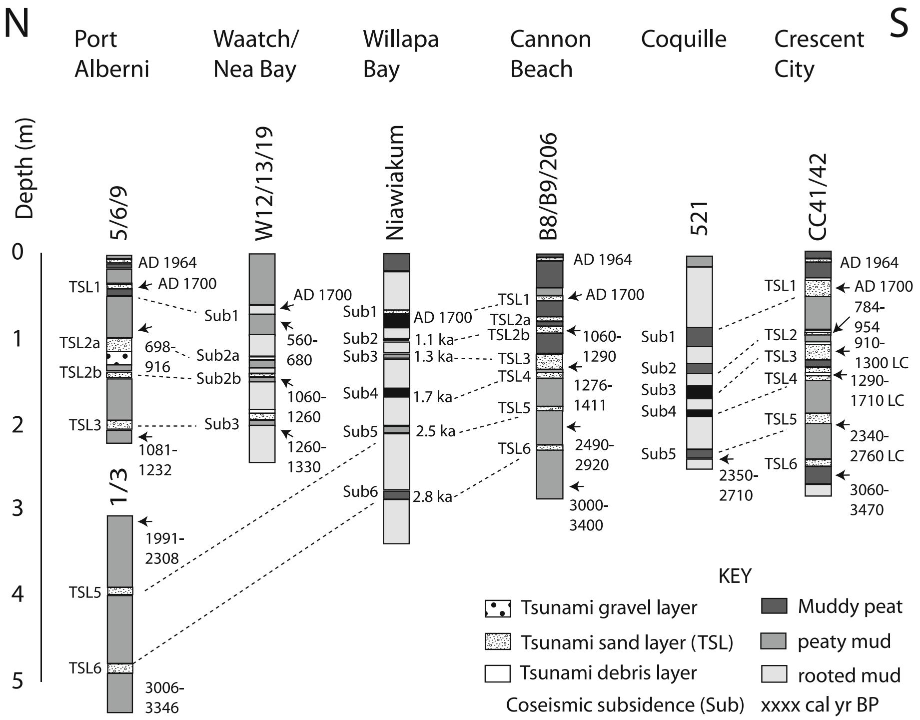

Fig. 7 Records of (*1*) regional coseismic subsidence at Waatch/Neah Bay, Willapa Bay and Coquille and (*2*) major paleotsunami sand sheets (TSL) at Port Alberni, Cannon Beach and Crescent City, from north (N) to south (S)

margin occurred at ~1.7 ka, as recorded at Willapa Bay and other central Cascadia localities (Fig. 7). The paleotsunami associated with the Cascadia rupture at 1.7 ka is likely represented at Discovery Bay (Fig. 3) and it might be confirmed at Port Alberni with additional work. Two older major tsunami inundations at Port Alberni, bracketed by dates of 2308 and 3006 years BP, could correlate with regional ruptures Sub5 and Sub6, as recorded at Willapa Bay.

Although there are long records of tidal marsh accretion in the Elk River valley at Crescent City (Fig. 3), no abruptly buried marsh horizons have been observed there (Peterson et al. 2011). Crescent City might lie too close to the 1st zero-isobase in the southern Cascadia margin (Figs. 1, 2) to directly record coseismic subsidence events there. However, a geologic record of five abruptly buried wetland horizons, extending back to 2350–2710 years BP (Fig. 7), does occur at Coquille, about 150 km north of Crescent City (Fig. 1). Coseismic subsidence events are also intermittently recorded near the mouth of the Coquille River at the coast (Witter et al. 2003; Peterson et al. 2012). The sequence of five regional subsidence events since 2350–2710 years BP at Coquille is similar to the sequence of the last five major paleotsunami inundations that occurred at Crescent City since 2340–2760 years BP (Fig. 6). The last five major paleotsunami sand sheets at Crescent City thus can be accounted for by the same number of apparent ruptures of the south-central Cascadia margin during the past ~2.7 ka.

3.3 Potential for far-field major tsunami

With one possible exception, there are no major paleotsunami sand sheets at Port Alberni, Cannon Beach, or Crescent City that cannot be accounted for by dated ruptures of the Cascadia margin, as established by regional coeismic subsidence events during the past 2.6–2.8 ka (Fig. 7). The one exception is the TSL2a event, which is tentatively tied to a weak subsidence event (Sub2a) in the Waatch wetlands (Peterson et al. 2013). The Sub2a subsidence event has yet to be confirmed elsewhere along the northern Cascadia margin. All other major tsunami inundations during the past 3,000 years at the three localities are assumed to have originated from long ruptures of the Cascadia megathrust.

An alternative method of evaluating the relative importance of far-field tsunamis along the Cascadia margin involves extrapolating the historic tsunami record. The historic record of six damaging far-field tsunamis along the Cascadia margin during the past ~150 years yields one damaging far-field tsunami per ~25 years (Fig. 4; Landers et al. 1993). During the 3,000-year period of the prehistoric record, as many as 120 damaging far-field paleotsunamis in Cascadia could have been generated from subduction zones around the Pacific Rim. It is evident, however, that such far-field paleotsunamis did not reach the threshold of major tsunami inundations, as defined above, in the study areas. In other words, nearly all of the transoceanic tsunamis, as extrapolated from the historic record, did not exceed the magnitude of the 1964 Alaskan far-field tsunami, in terms of sand sheet inland extent, continuity, and/or thickness at the three study localities. Only the Gulf of Alaska subduction zone had the capability to produce a far-field tsunami sand sheet (1964 tsunami) that could be recognized in the proximal core sites of the three study areas. Presumably, only the largest Gulf of Alaska ruptures (M_w 9) produce tsunamis large enough to be detected by sand sheets along the Cascadia margin.

Rare sand laminae (~0.5 cm thick) occur between some of the prominent sand sheets in the most proximal core sites at the Crescent City wetlands (Carver et al. 1998). These sand laminae are discontinuous between the proximal core sites. It is not known whether the laminae were derived from coseismic paleoliquefaction, severe wind-storms, or minor tsunami from local fault ruptures (Valentine et al. 2012). As the laminae do not reach the thickness or extent of the 1964 tsunami sand sheet at Crescent City, their possible origins are not addressed in this study of major paleotsunamis along the Cascadia margin.

3.4 Cascadia margin paleotsunami hazard: recurrence and evacuation imperative

If we assume that the major paleotsunami inundations recorded at Port Alberni, Cannon Beach and Crescent City result from regional ruptures of the Cascadia margin, then the rupture ages can be used to refine major tsunami recurrence intervals. For example, the mean recurrence interval of seven major paleotsunamis between AD 1700 and 2.8 ka at Port Alberni and Cannon Beach along the northern and central Cascadia margins is 416 years. The mean recurrence interval of six major tsunamis between AD 1700 and 2.8 ka at Crescent City at the southern end of the Cascadia margin is 500 years. The mean recurrence interval of the three-event clusters of major paleotsunamis, between assumed rupture dates of 1.1 and 1.7 ka, along the central Cascadia margin is 300 years. It has been 313 years since the last Cascadia rupture at AD 1700, but there is uncertainty about whether the Cascadia margin is now within or between clustered rupture events. As it might not be possible to predict whether the next rupture will occur within or between event cluster(s), a prudent approach is to provide a range of estimated mean recurrence intervals (300–500 years) to the general public that span both (1) clustered event groupings and (2) a mixture of clustered and nonclustered event groupings. While not the same as formal probability analyses, these mean recurrence interval estimates are easily and effectively communicated to the 'at risk' populations.

To avert a catastrophe like the recent tsunami in Japan in 2011 (Ando et al. 2011), coastal residents and tourists at the Cascadia margin must evacuate to high ground without delay following a megathrust earthquake. The 'at risk' populations must not wait for official warnings, directions, or permissions to self-initiate the life-saving evacuations. To reinforce this imperative, the following 15:15:15 evacuation rule could be repeated with every public discussion and/or media presentation of tsunami hazards in the Cascadia margin. The basic rule is as follows: (1) at least 15 s of strong ground shaking could be followed by a major tsunami that (2) could arrive within 15 min after the earthquake, with (3) waves possibly reaching 15 m in elevation relative to NAVD88, or the heights of 3–4 story buildings built on beach plains along the open Pacific Ocean coast. Some variations in these parameters are expected between different localities and different rupture scenarios, but the 15:15:15 rule, if adhered to in self-initiated evacuations, could save tens of thousands of lives along the Cascadia margin. Additional roles of regional government agencies could include providing support for dedicated evacuation routes or vertical evacuation structures and the provisioning of evacuation centers for post-event relief.

Educating and training 'at risk' populations to discriminate between related but different hazards and to take the appropriate life-saving precautions could have relevance in other multiple hazard settings around the world.

4 Conclusions

Coastal residents of Cascadia are confused about the hazards posed by two related but different types of tsunami events—near-field and far-field tsunamis. The different hazards are compared in this study based on expected magnitudes of tsunami flooding and recurrence interval, as calibrated by a damaging historic tsunami (1964) in the region. Six or seven paleotsunami sand sheets are recorded at our three study localities—Port Alberni, British Columbia, Canada, and Cannon Beach, Oregon, and Crescent City, California, USA during the past 3,000 years. The paleotsunami sand sheets exceeded in thickness, continuity, and extent, those left by the historic far-field 1964 tsunami at the same sites, and thus, they are defined as major paleotsunami inundations. The number and approximate ages of the major paleotsunami inundations are similar to the number and reported ages of coseismic plate-boundary ruptures that are recorded by coseismic coastal subsidence events on the central Cascadia margin. The combinations of prolonged seismic shaking, damaged transportation and/or communications infrastructure, and short transit times for the associated near-field tsunami are linked by the near 1:1 correspondence between the megathrust rupture events and major tsunami inundations. This apparent direct relation on the Cascadia margin provides the impetus for (1) focusing public attention on infrequent, but devastating, near-field tsunami and (2) repeatedly reminding the 'at risk' populations that they are entirely responsible for their own safety and they must promptly self-evacuate during strong prolonged ground shaking.

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