NATURAL RESOURCES

TSUNAMI HAZARD MAPS OF THE SAN JUAN ISLANDS, WASHINGTON—MODEL RESULTS FROM A CASCADIA SUBDUCTION ZONE EARTHQUAKE SCENARIO

by Timothy J. Walsh, Edison Gica, Diego Arcas, Vasily V. Titov, and Daniel W. Eungard

WASHINGTON
DIVISION OF GEOLOGY
AND EARTH RESOURCES
Map Series 2016-01
October 2016

[Partially Superseded by Map Series 2021-01]

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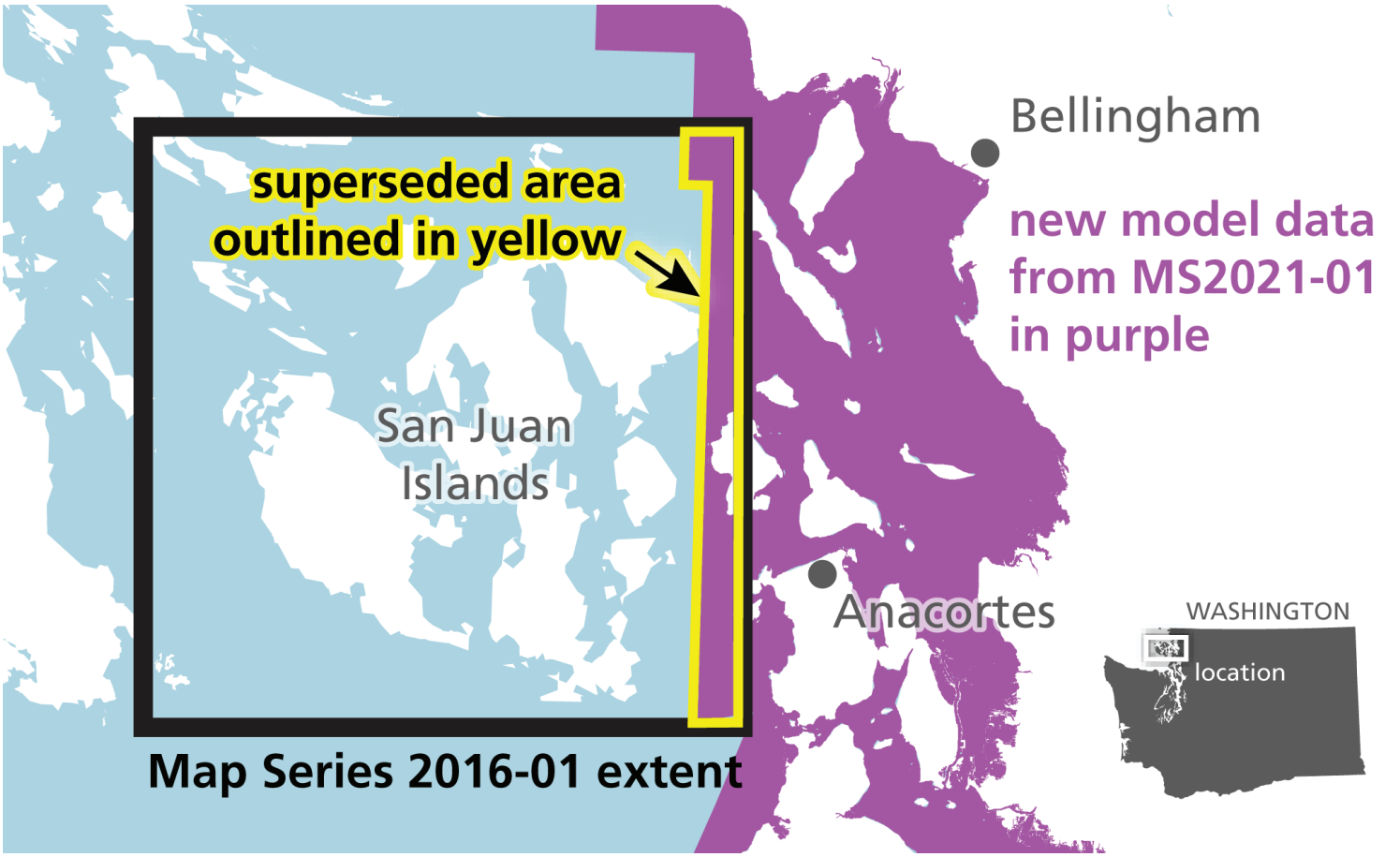

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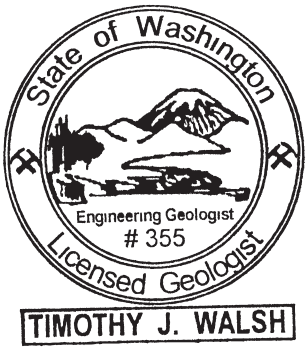

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Tsunami Hazard Maps of the San Juan Islands, Washington—Model Results from a Cascadia Subduction Zone Earthquake Scenario

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INTRODUCTION

In 1995, Congress directed the National Oceanic and Atmospheric Administration (NOAA) to develop a plan to protect the West Coast from tsunamis generated by the nearby Cascadia subduction zone (CSZ). A panel of representatives from NOAA, the Federal Emergency Management Agency (FEMA), the U.S. Geological Survey (USGS) and the five Pacific coast states wrote the plan and submitted it to Congress, which created the National Tsunami Hazard Mitigation Program (NTHMP) in October of 1996. The NTHMP is designed to reduce the impact of tsunamis through warning guidance, hazard assessment, and mitigation. A key component of the hazard assessment for tsunamis is delineation of areas subject to tsunami inundation. Because local tsunami waves may reach nearby coastal communities within minutes of the earthquake, there will be little or no time to issue formal warnings; evacuation areas and routes will need to be planned well in advance.

This map is part of a series of tsunami inundation maps produced by the Washington Division of Geology and Earth Resources, in cooperation with the Washington Emergency Management Division, as a contribution to the NTHMP (Table 1). These maps are produced using computer models of earthquake-generated tsunamis from nearby seismic sources. We model two main types of tsunami inundation in Washington—inundation caused by local crustal faults in the Puget Sound (such as the Tacoma or Seattle faults), and inundation caused by the Cascadia subduction zone. Our previous inundation modeling of an event on the CSZ used the 1A and 1A with asperity scenarios (Myers and others, 1999; Priest and others, 1997), but more recent studies (Witter and others, 2011) have shown that the L1 scenario is a more conservative choice, that is, it is less likely to be exceeded. See *Earthquake Magnitudes and Slip Distributions* for more information on model scenarios. The modeling for this map was done by the NOAA Center for Tsunami Research at NOAA's Pacific Marine Environmental Laboratory in Seattle.

CASCADIA SUBDUCTION ZONE

Research over the last few decades about the occurrence of great earthquakes off the British Columbia, Washington, Oregon, and northern California coastlines and resulting tsunamis (Atwater, 1992; Atwater and others, 1995) has led to concern about locally generated tsunamis that will leave little time for response. Numerous workers have found geologic evidence of tsunami deposits attributed to the CSZ in at least 59 localities from northern California to southern Vancouver Island (Peters and others, 2003).

Table 1. Completed tsunami hazard maps for Washington. CSZ, Cascadia subduction zone.

Location	Reference	Modeled Scenario
San Juan Islands	This study	CSZ L1
southern Washington coast	(Walsh and others, 2000)	CSZ 1A and 1A with asperity
Port Angeles	(Walsh and others, 2002a)	CSZ 1A and 1A with asperity
Port Townsend	(Walsh and others, 2002b)	CSZ 1A and 1A with asperity
Neah Bay	(Walsh and others, 2003a)	CSZ 1A and 1A with asperity
Quileute area	(Walsh and others, 2003b)	CSZ 1A and 1A with asperity
Seattle	(Walsh and others, 2003c)	Seattle fault
Bellingham	(Walsh and others, 2004)	CSZ 1A and 1A with asperity
Anacortes–Whidbey Island	(Walsh and others, 2005)	CSZ 1A and 1A with asperity
Tacoma	(Walsh and others, 2009)	Tacoma fault
Everett	(Walsh and others, 2014)	Seattle fault

Figure 1. Location map of Washington, showing San Juan Islands study area, Cascadia subduction zone, major offshore channels, and major crustal faults discussed in the text.

While most of these locations are on the outer coast, inferred CSZ tsunami deposits have been identified as far east as Discovery Bay (near Port Townsend)(Williams and others, 2005), and on the west shore of Whidbey Island (Williams and Hutchison, 2000)(Fig. 1). Heaton and Snavely (1985) report that Makah stories may tell of a tsunami washing through Waatch Prairie near Cape Flattery. Ludwin (2002) has found additional stories from native peoples up and down the coast that appear to corroborate this and also include apparent references to associated strong ground shaking.

Most-Recent Large Earthquake and Tsunami

Additionally, correlation of the timing of the last CSZ earthquake by high-resolution dendrochronology (Jacoby and others, 1997; Yamaguchi and others, 1997) with Japanese historical records (Satake and others, 1996) demonstrate that a distant-source tsunami in A.D. 1,700 almost certainly came from the CSZ. This tsunami may have lasted as much as 20 hours in Japan and contributed to a shipwreck about 100 km (62 mi) north of Tokyo (Satake and others, 2003; Atwater and others, 2005).

Recurrence Intervals

Estimates of the frequency of occurrence of Cascadia subduction zone (CSZ) earthquakes are derived from several lines of evidence: coastal submergence events, tsunami deposits, and turbidite deposits. Great subduction zone earthquakes commonly cause coseismic subsidence (Plafker, 1969; Plafker and Savage, 1970). Where this subsidence occurs in coastal marshes, marsh deposits may be abruptly overlain by estuarine mud, indicating sudden submergence and drowning of upland surfaces (Atwater,

1992). Atwater and Hemphill-Haley (1997) reported six sudden submergence events in Willapa Bay over the last 3,500 years. Their data imply an average recurrence interval of about 500–540 years, but individual intervals vary from as little as one to three centuries to as much as 1,000 years.

Researchers working farther south have found a somewhat different record. Using marsh stratigraphy and inferred tsunami deposits, Kelsey and others (2002) found a 5,500-year record at Sixes River in southern Oregon (Fig. 2) that included an abrupt subsidence event not observed on the southern Washington coast. Kelsey and others (2005) examined Bradley Lake on the southern Oregon coast near Bandon and found that it recorded inferred tsunami deposits with an average recurrence interval of ~390 years. This discrepancy implies that tsunamis were generated by earthquakes on the CSZ that did not produce abrupt subsidence in southern Washington. A possible explanation is that the earthquake did not rupture the entire length of the subduction zone and may represent an additional event not seen in the southern Washington record. Nelson and others (2006) examined the degree of overlap and amount of abrupt subsidence at eight sites along the Oregon and Washington coasts and concluded that rupture lengths—and therefore earthquake magnitudes—varied; ruptures along the northern CSZ are generally long; ruptures along the southern CSZ are more variable in both length and recurrence interval.

Another approach to inferring recurrence intervals is the correlation of turbidites—deposits of sediment gravity flows, or turbidity currents—below the continental shelf. Adams (1990) inferred that turbidite deposits in Cascadia Channel and Astoria Canyon (Fig. 1) were triggered by great earthquakes. If turbidity currents are triggered independently, at different times, and at multiple submarine canyon heads that merge with a main channel, then they should be additive in the main channel. For example, if a channel has three tributaries, each of which has ten independent turbidites, there would be 30 turbidites in the main channel. However, if the turbidites are triggered simultaneously—which would likely be the case if they were triggered by a great earthquake—they should coalesce. In this case, the maximum number of turbidites in the main channel would be no more than the maximum number found in any individual channel (Fig. 3). In both Cascadia Channel and Astoria Canyon, Adams (1990) reported that Oregon State University researchers logged 13 turbidites in multiple deep sea cores that were stratigraphically above the Mazama ash (radiocarbon dated at about 6,845 radiocarbon years BP [calibrated to about 7,700 cal yr BP]). Adams (1990) therefore inferred an average recurrence interval of 590 ±170 years.

Goldfinger and others (2012) tested Adams' (1990) hypothesis by collecting numerous additional cores along the Cascadia continental margin. Their effort greatly expands the geographic and chronologic range of observation, and increases observation density. Goldfinger and others (2012) inferred from their record of turbidite deposits that the Cascadia subduction zone is segmented, with full-length ruptures having a recurrence interval similar to those estimated by Adams (1990) and Atwater and Hemphill-Haley (1997), but with additional partial-length ruptures offshore Oregon and northern California.

Combining full-length and partial ruptures on Cascadia, they estimated a recurrence interval of ~240 years for earthquakes off Oregon and northern California, but still 500–530 years offshore Washington and British Columbia. Earthquakes that rupture only the northern part of the CSZ are also a possibility, although there is less direct evidence for that. In Discovery Bay and in the northeast of the Olympic Peninsula, Williams and others (2005) observed nine muddy sand beds bearing marine diatoms that interrupt a 2500-yr-old sequence of peat deposits beneath a tidal marsh. If all of these are tsunami deposits, then it is likely that some of them record events that are not full-length ruptures of the CSZ. The ages of four of these beds overlap with known late-Holocene tsunamis generated by full-length ruptures of the CSZ. Diatom assemblages in peat deposits bracketing these four beds do not indicate a concurrent change in elevation at Discovery Bay.

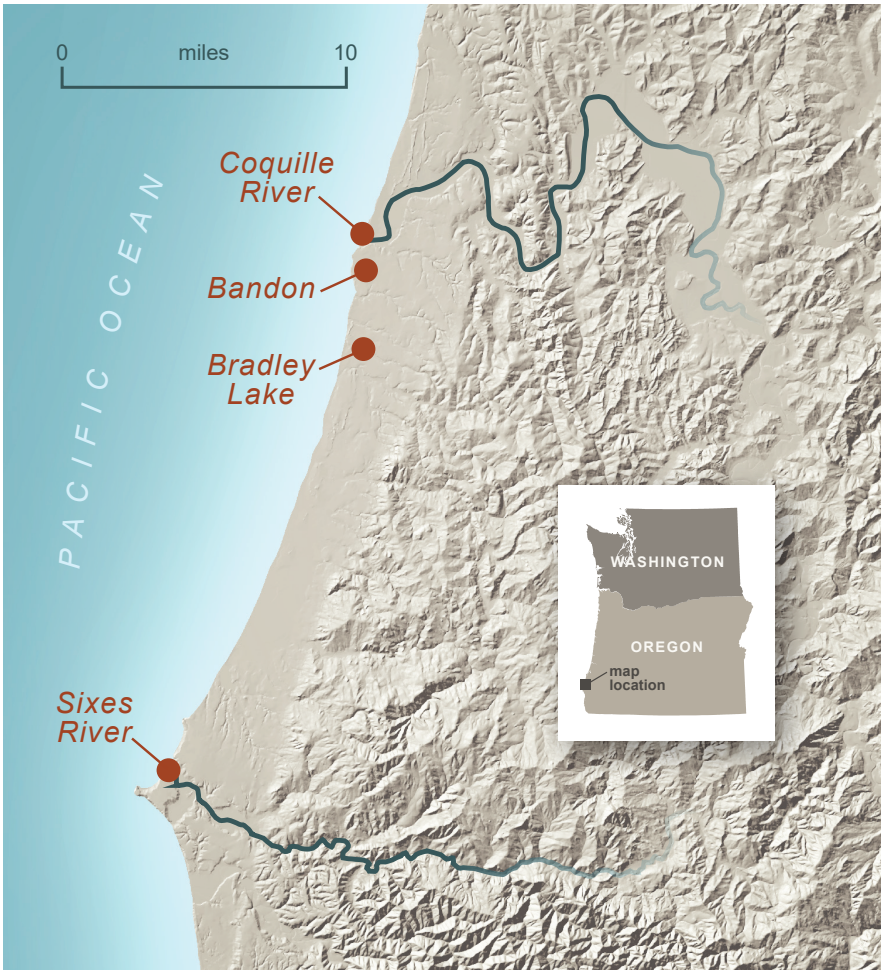

Figure 2. Map of southwest Oregon, showing tsunami deposits and abrupt subsidence locations used by Kelsey and others (2002; 2005) to determine CSZ earthquake recurrence intervals.

Table 2. Estimates of earthquake recurrence on the Cascadia subduction zone.

Events over time interval	Average recurrence interval; range if given	Section of CSZ	References	Major evidence
6 submergence events in 3,500 years	500–540 average, 100–300 to 1,000	northern	Atwater and Hemphill-Haley (1997)	submergence events
11 submergence events in 5,500 years	510 years	southern	Kelsey and others (2002)	marsh stratigraphy and tsunami deposits
13 tsunamis, 17 disturbances in 7,000 years		southern	Kelsey and others (2005)	marine incursions and disturbance events in Bradley Lake
	variable	whole	Nelson and others (2006)	multiple
	590 ±170	northern	Adams (1990)	turbidites in Astoria Canyon and Cascadia Channel
19 or 20 full margin turbidites in 10,000 years; 22 turbidites restricted to the south	500–530 average for full margin rupture; ~240 full margin plus southern only	whole and partial	Goldfinger and others (2012)	turbidites along Cascadia margin

This suggests that coseismic subsidence has been negligible as far east as Discovery Bay, which is due south of the community of Eastsound on Orcas Island and approximately equidistant from the CSZ. However, one inferred tsunami deposit is accompanied by several decimeters of abrupt subsidence, which is interpreted as the result of deformation associated with an upper plate fault. Other sand sheets in the sequence may represent tsunamis generated by partial ruptures of the CSZ restricted to the north that did not generate or preserve turbidites, tsunamis generated by upper plate fault earthquakes, or by landslides. This implies that either some CSZ earthquakes do not leave turbidite deposits in Cascadia Channel, or that some tsunami deposits were generated by other events such as local earthquakes or landslides.

Atwater and others (2014) also questioned whether the absence of turbidites along the northern CSZ necessarily proves the absence of ground shaking, or rather is influenced by differences in sediment supply and in the differences in flow paths down tributary channels. They also questioned some of the correlations among widely spaced sites—used to infer the length of fault rupture—that were used by Goldfinger and others (2012).

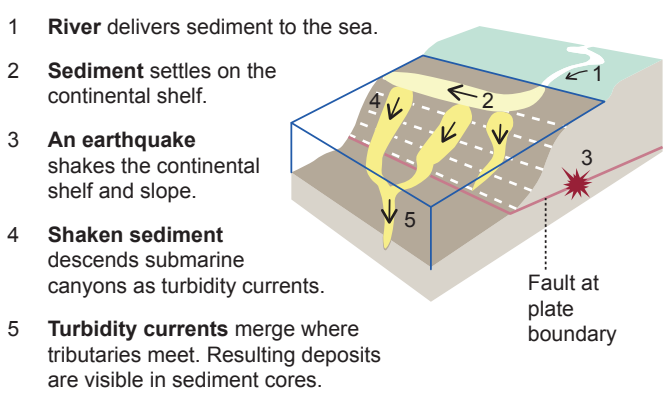

Figure 3. Schematic view of the confluence test for extensive seismic shaking, first used as a guide to fault rupture length by Adams (1990). Adams assumed that extensive shaking enables turbidity currents to descend different submarine channels at the same time, and to merge below channel confluences. Atwater and others (2014), dispute the reliability of this indicator.

Earthquake Magnitudes and Slip Distributions

A.D. 1,700 EARTHQUAKE

It is believed that the last earthquake on the Cascadia subduction zone was about magnitude (M_w)9 (Satake and others, 1996; 2003). Satake and others (2003) tested various rupture lengths, slip amounts, and observed tsunami wave heights in Japan for the A.D. 1,700 event. They estimated that the A.D. 1,700 event had a rupture length of ~1,100 km (684 mi) and 19 m (62 ft) of coseismic slip on an offshore, full-slip zone with linearly decreasing slip on a down-dip partial-slip zone, suggesting a magnitude of 8.7 to 9.2. They inferred that the most likely magnitude was 9.0 based on the correlation between estimates of coseismic subsidence from paleoseismic studies and the subsidence predicted by the dislocation models of their scenarios.

PRE-A.D. 1,700 EARTHQUAKES

Partial-Length Rupture Models

The magnitudes and slip distributions of earlier CSZ earthquakes are less well constrained. Inferences of shorter ruptures that affect only the southern part of the CSZ generally imply smaller magnitude earthquakes. Tsunamis from postulated various shorter ruptures limited to the southern part of the CSZ were modeled by Priest and others (2014), who concluded that the tsunamis they generated were significantly smaller in Washington than those generated by full-length ruptures and they will not be considered further here. A partial CSZ rupture restricted to the north was suggested by Goldfinger and others (2013), but evidence for it is insufficient to generate a tsunami model.

Full-Length Rupture Models

Witter and others (2012) combined: (1) turbidite data from Goldfinger and others (2012); (2) correlation of inferred tsunami deposits with turbidites in Bradley Lake; and (3) inferred tsunami deposits in the Coquille River estuary at Bandon, Oregon that extend as much as 10 km (6.2 mi) farther inland than the A.D. 1,700 tsunami deposits (Witter and others, 2003). They inferred from this that tsunamis generated by Cascadia over the last 10,000 years had been highly variable, with some larger than the one in A.D. 1,700. They constructed 15 scenarios of full-length ruptures

defining vertical seafloor deformation used to simulate tsunami inundation at Bandon, Oregon. Rupture models include slip partitioned to a splay fault in the accretionary wedge and models that vary the up-dip limit of slip on a buried megathrust fault. Slip estimates were made from several sources. Total turbidite volume was estimated from the thickness averaged over all the paleoseismic records, which Goldfinger and others (2012) correlated to earthquake magnitude. This was combined with estimates of the convergence rate for different segments of the subduction zone multiplied by the time since the last previous event to estimate total slip (Witter and others, 2012). Tsunami simulations at Bradley Lake (Witter and others, 2011) and at Bandon (Witter and others, 2012) were compared with the distribution of inferred tsunami deposits there and were broadly compatible with their larger scenarios. They performed numerical simulations of the tsunamis generated by each scenario and evaluated them using a logic tree that ranked model consistency with geophysical and geological data (Table 2).

Witter and others (2011) concluded that scenario L1—a splay fault model with a maximum slip of 27 m (88.6 ft) and an average slip of 13 m (42.6 ft)—produced a tsunami that equaled or exceeded 95 percent of the variability in their simulations (Fig. 4). Other 'L' earthquake scenarios (L2 and L3) have the same amount of slip but somewhat different distributions across strike of the subduction zone. In other words, the L1 scenario produces tsunamis as big as, or bigger than most other models. Witter and others (2011) also estimated the size of the earthquakes that generated turbidites along the full length of the CSZ. They concluded that three earthquakes in the last ~10,000 years were probably similar to scenario L and only one was larger (table 1 in Witter and others [2011]). The inter-event times between pairs of inferred L earthquakes are ~1,800 and ~4,600 years. Another way to estimate recurrence frequency is that if three earthquakes in the last 10,000 years are L, then these type of events have an average recurrence interval between 2,500 and 5,000 years. If this truly represents 95 percent of the hazard over a 10,000-year period, then scenario L earthquakes have a long recurrence interval and likely is of a similar probability of occurrence as the International Building Code seismic standard of 2 percent probability of exceedance in 50 years, or colloquially, a 2,500-year event.

MODELING APPROACH AND RESULTS

The tsunami inundation model is based on a numerical model of waves generated by a scenario earthquake. We use the finite difference model of Titov and Synolakis (1998), also known as the Method of Splitting Tsunami (MOST) model (Titov and González, 1997). The model uses a grid of topographic and bathymetric elevations and calculates a wave elevation and velocity at each gridpoint at specified time intervals to simulate the generation, propagation, and inundation of tsunamis down the Strait of Juan de Fuca and into the San Juan Islands. The model is calculated for mean high water and does not include tidal effects.

Our model results do not include potential tsunamis from landslides, which in the 1964 Alaska earthquake caused 85 of the 106 tsunami deaths in Alaska (Lander, 1996), or nearby crustal faults, which are generally not well enough understood to be modeled. Tsunami deposits that are apparently locally generated have been found on Whidbey Island (Williams and Hutchinson,

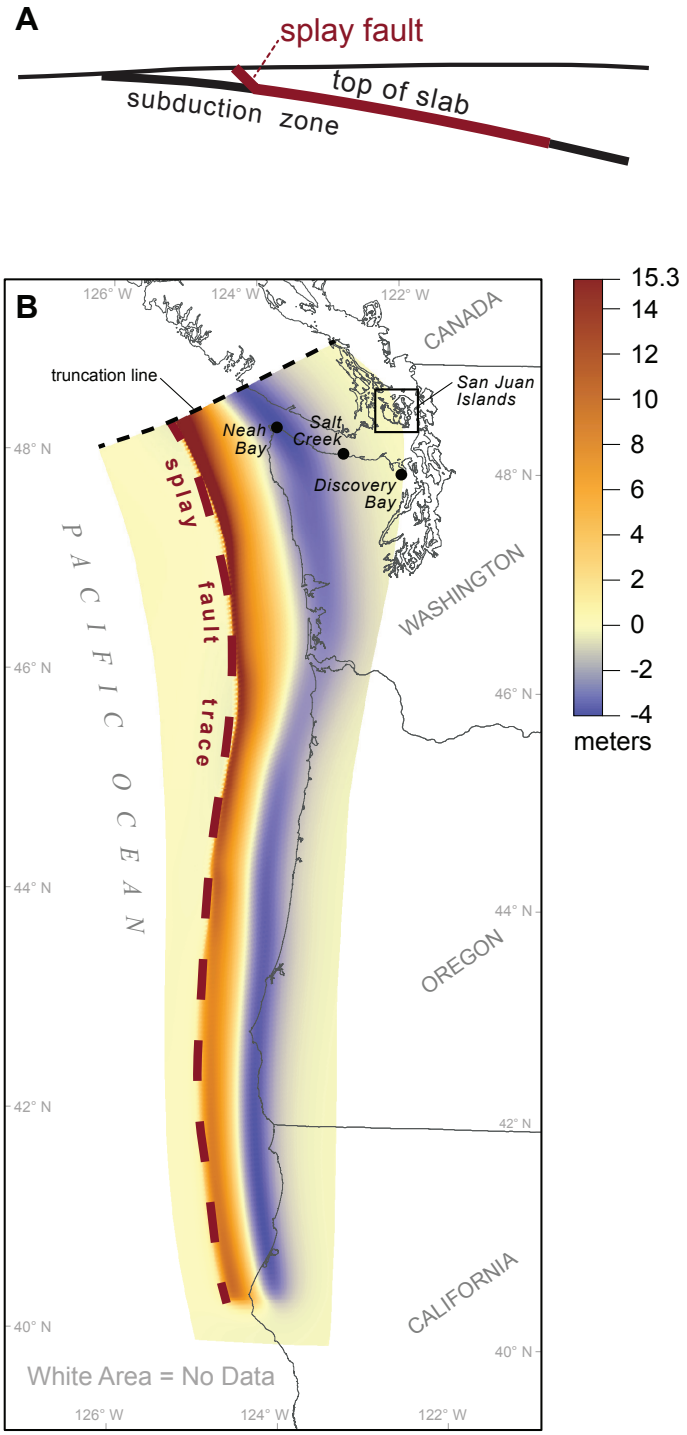

Figure 4. L1 splay fault model diagram (A) and map of vertical deformation (B) during a great Cascadia earthquake in the L1 scenario of Witter and others (2011). The northern part of the scenario is truncated south of where the Juan de Fuca plate is beginning to break up into microplates. Subsequent modeling (Edison Gica and Diego Arcas, Pacific Marine Environmental Laboratory, 2015, written commun.) shows that tsunami energy farther north is directed to the northeast, away from the Strait of Juan de Fuca, and would have limited effect in the San Juan Islands. Paleoseismic investigations have suggested approximately 0.5 meters or less of coseismic subsidence at Salt Creek (Hutchinson and others, 2013) and little or no subsidence at Discovery Bay (Williams and others, 2005), similar to this model. However, the L1 model has more subsidence farther west than has been confirmed for the A.D. 1,700 Cascadia earthquake at Neah Bay (Peterson and others, 2013).

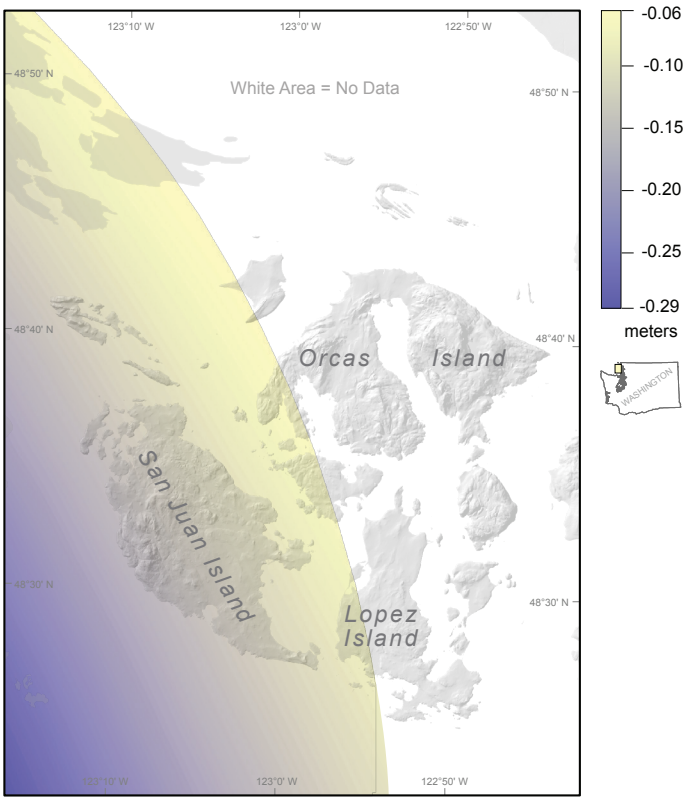

Figure 5. Closeup of initial deformation from the L1 scenario in the San Juan Islands. Note that tectonic subsidence in the model is less than 20 cm (7.9 in.) along the west shore of San Juan Island, tapering to 0 farther east.

2000; Atwater and Moore, 1992), in Discovery Bay near Port Townsend (Williams and others, 2005), in the Snohomish delta near Everett (Bourgeois and Johnson, 2001), and at West Point near Seattle (Atwater and Moore, 1992). González (2003) summarizes the evidence for tsunamis generated within the Puget Lowland by local earthquakes and landslides and estimates their probabilities.

The selected scenario is a splay-fault model in which all slip is partitioned into a thrust fault in the accretionary wedge that has an approximate 30º landward dip and the same sense of movement as the megathrust; this results in a much higher, narrower area of uplift than a fault rupture on the megathrust which dips landward much more shallowly and reaches farther seaward than the splay fault. The land surface in the San Juan Islands is modeled to subside minimally or not at all during ground shaking (Fig. 5), although there are no paleoseismic investigations in the islands to validate this inference in the San Juan Islands. However, as noted above, Discovery Bay displays little or no coseismic subsidence and is situated at the same contour of coseismic deformation in this scenario (Fig. 5).

Inundation

The modeled tsunami inundation depths are shown on the map in three ranges: 0–2.5 ft, 2.5–6 ft, and greater than 6 ft (Map Sheets 1, 2, and 3). These ranges were chosen because they are about knee-high or less, knee-high to head-high, and more than head-high, and so approximate the degree of hazard for life safety. The greatest amount of tsunami flooding from this scenario is expected to occur on Lopez Island, southeastern San Juan Island, southwestern Decatur Island, near the community of Eastsound on Orcas Island, and on parts of Shaw Island (Map Sheets 1, 2, and 3). Elsewhere, tsunami flooding is expected only in the immediate vicinity of the shoreline where evacuation to higher ground would be relatively straightforward.

Current Speed

The modeled current speed (Map Sheet 4) is shown in four ranges: less than 3 knots, 3–6 knots, 6–9 knots, and greater than 9 knots, following the port damage categorization of Lynett and others (2014). These zones represent, respectively, no expected damage, minor/moderate damage possible, major damage possible, and extreme damage possible. Modeled current speed locally exceed 30 knots in narrow passages between islands, where tidal currents are strongest and tsunami-tide interactions are likely to be most significant. These larger values are not depicted separately on the map because the variability due to tidal interactions is very large.

Timing of Tsunami and Initial Water Disturbance

Figure 6 shows simulated tide gauge records at Friday Harbor off San Juan Island, and at East Sound off Orcas Island. The initial water disturbance is a wave trough arriving about 1.5 hours after the earthquake. This suggests that the first local sign of a tsunami may be water receding, resembling a large outgoing tide. The trough has a magnitude of about 3 ft (0.9 m) at Friday Harbor and about 6 ft (1.8 m) at East Sound. A wave crest follows about 15 minutes later (1.75 hours after the earthquake) and allows some time for evacuation to high ground.

LIMITATIONS OF THE MODEL

Because the characteristics of the tsunami depend on the initial deformation of the earthquake, which is poorly understood, the largest source of uncertainty is the input earthquake. The

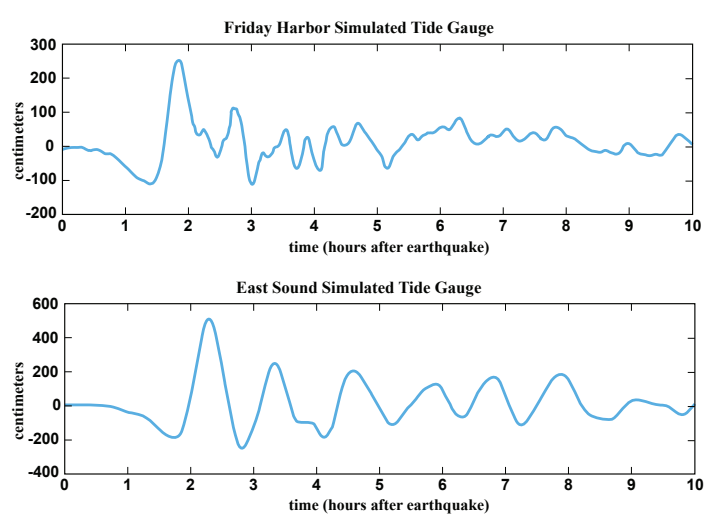

Figure 6. Modeled tsunami amplitude time history for harbors near the towns of Friday Harbor and Eastsound. (see Map Sheet 2 for locations).

earthquake scenario used in this model was selected to approximate the 2 percent probability of exceedance in 50 years, but the next earthquake may have a more complex slip distribution than the simplified scenario we used and thus the ensuing tsunami may differ in detail. Witter and others (2011) suggest that the most likely full-length CSZ earthquake will have an average slip of about ⅔ of the L1 scenario and generate a smaller tsunami.

Another significant limitation is that the resolution of the modeling is limited by the bathymetric and topographic data used to make the model grid. Lidar data with 3-ft grid cells were used to build the topographic model, but the bathymetric model in the vicinity of the islands has grid cells of 33 ft (10 m). The coarser the grid, the fewer number of small topographic features that can influence the tsunami locally are included.

The model does not include the influences of changes in tides and is referred to mean high water. The tide stage can amplify or reduce the impact of a tsunami on a specific community. For example, the diurnal range (the difference in height between mean higher high water and mean lower low water) is 7.76 ft (2.36 m) at the Friday Harbor tide gage (http://www.tidesandcurrents.noaa.gov). The model also does not include interaction with tidal currents, which can be additive, or if in opposite directions, can steepen the tsunami wave front and cause a breaking wave. This means that while the modeling can be a useful tool to guide evacuation planning, it is not of sufficient resolution to be useful for site-specific land-use planning.

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