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Key Points:

- Extraordinary cluster of M >~6.5 crustal quakes in Cascadia Puget Sound region
- The 1200-900 cal year B.P. cluster not triggered by great Cascadia thrust quake
- Great Cascadia subduction thrusts may have triggered a few crustal earthquakes

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Crustal earthquake triggering by pre-historic great earthquakes on subduction zone thrusts

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Abstract Triggering of earthquakes on upper plate faults during and shortly after recent great ($M > 8.0$) subduction thrust earthquakes raises concerns about earthquake triggering following Cascadia subduction zone earthquakes. Of particular regard to Cascadia was the previously noted, but only qualitatively identified, clustering of $M >$ ~6.5 crustal earthquakes in the Puget Sound region between about 1200–900 cal years B.P. and the possibility that this was triggered by a great Cascadia thrust subduction thrust earthquake, and therefore portends future such clusters. We confirm quantitatively the extraordinary nature of the Puget Sound region crustal earthquake clustering between 1200–900 cal years B.P., at least over the last 16,000. We conclude that this cluster was not triggered by the penultimate, and possibly full-margin, great Cascadia subduction thrust earthquake. However, we also show that the paleoseismic record for Cascadia is consistent with conclusions of our companion study of the global modern record outside Cascadia, that $M > 8.6$ subduction thrust events have a high probability of triggering at least one or more $M >$ ~6.5 crustal earthquakes.

1. Introduction

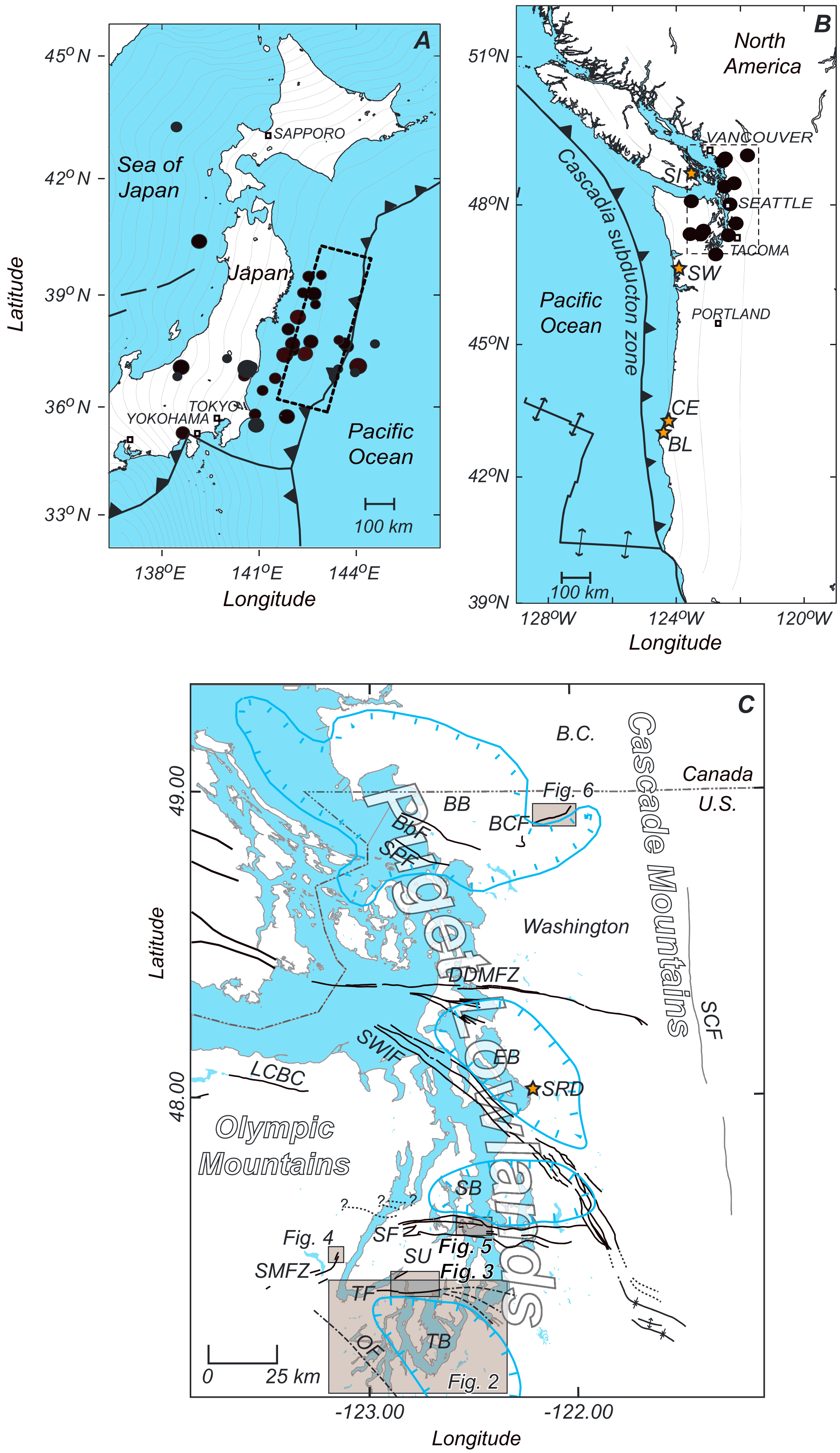
The most recent great ($M > 8.0$) subduction thrust earthquakes, particularly the 2011 *M*9.0 Tohoku-oki Japan earthquake [*Asano et al.*, 2011; *Ishibe et al.*, 2011; *Kato et al.*, 2011; *Nettles et al.*, 2011; *Yoshida et al.*, 2011], raised concerns about the potential for triggered crustal earthquakes to threaten major population centers or power-generating facilities in subduction zones that include Tokyo, Jakarta, Manila, Lima, Santiago (Chile) [*Bilham*, 2009], and in the Cascadia subduction zone (CSZ) of the Pacific Northwest, U.S.. These concerns motivated us to study possible triggering of crustal earthquakes by subduction thrust events using the modern global earthquake record, as described in a companion paper [*Gomberg and Sherrod*, 2014]. We also became more concerned about the possibility that a great subduction thrust earthquake triggered an alleged cluster of $M >$ ~6.5 paleo-earthquakes on multiple crustal faults in the greater Puget Sound region within the same few centuries or less, about 1200–900 cal years B.P. (i.e., sidereal years before A.D. 1950) [*Bucknam et al.*, 1992; *Sherrod*, 2001; *Sherrod et al.*, 2004; *Barnett*, 2007; *Blakely et al.*, 2009] (Figure 1), thus portending similar paroxysms accompanying future great earthquakes. We independently approached questions of immediate triggering [see *Gomberg and Sherrod*, 2014] and longer term triggering (herein), but discuss if and how results of both may relate to one another in section 6.

We examine the question of whether CSZ thrust earthquakes are likely to trigger crustal events purely observationally, in part, because simple models fail to explain recent damaging crustal earthquakes clearly associated with great subduction thrust earthquakes. For example, several $M > 6$ and numerous smaller shallow crustal earthquakes were triggered by the *M*8.8 2010 Maule Chile and *M*9.0 2011 Tohoku-oki Japan subduction thrust events (Figure 1), and in both cases the crustal earthquakes had normal faulting mechanisms despite being located within compressive tectonic stress fields (and oriented suitably to relax compressive stresses) [*Ryder et al.*, 2012; *Kato et al.*, 2011].

It should also be noted that triggered (crustal) and triggering (subduction thrust) faults may be loaded by different deformation fields, as would be the case for Cascadia crustal and subduction zone thrust faulting [*Mazzotti et al.*, 2002]. All that may be required is that triggering deformations move a triggered fault to failure, so that very small perturbations may be effective triggers for already critically stressed faults (regardless of how they were brought to a critical state) [*Stein*, 1999; *Freed*, 2005; *Steacy et al.*, 2005].

2. Tectonic Setting

In this section we summarize briefly the tectonic setting of crustal faults in the Puget Sound region. Models of the Cascadia fore arc show a series of migrating, clockwise-rotating fore arc blocks driven by regional-scale


A
45° N
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Japan
TOKYO
YOKOHAMA
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141°E
144°E
Longitude
B
51°N
48°N
45°N
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39°N
North America
VANCOUVER
SI
SEATTLE
TACOMA
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Fig. 6
BB
BCF
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Puget Lowlands
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Washington
DDMFZ
SWIF
EB
SRD
LCBC
SCF
Olympic Mountains
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Figure 1

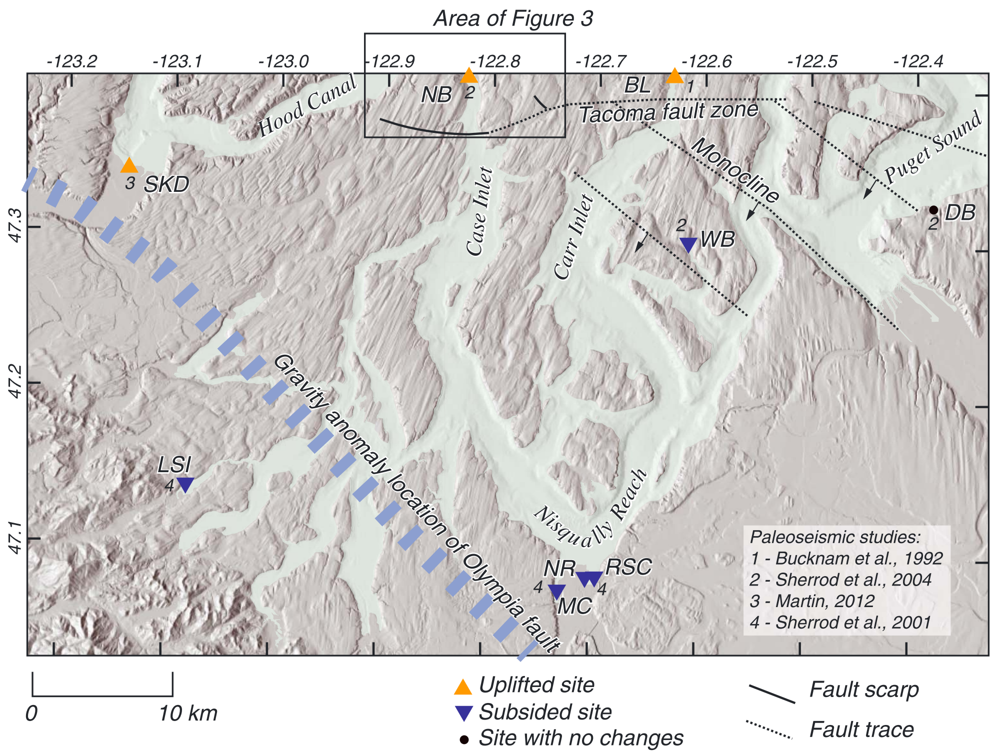

Figure 2. Shaded relief LiDAR image of the Olympia fault and Tacoma fault zones showing locations of paleoseismic features. Locations of sites described in the text and shown on figure: SKD = Skokomish River delta [*Martin*, 2012], NB = North Bay, BL = Burley Lagoon, DB = Dumas Bay, WB = Wollochet Bay, LSI = Little Skookum Inlet, MC = McAllister Creek, ND = Nisqually River, and RSC = Red Salmon Creek.

rotation from Pacific-North America shear and extension in the Basin and Range [*Wells and Simpson*, 2001; *Wells et al.*, 1998; *Wells and McCaffrey*, 2013]. This clockwise rotation causes higher rates of convergence in western Washington than elsewhere in Cascadia where the Oregon Coast Range block impinges on Tertiary volcanic rocks and sediments, compressing these Tertiary rocks against the southern edge of the British Columbia Coast Mountains. This north-south contraction results in series uplifts separating of structural basins in the Puget Sound region (123.6°W to 121.5°W, 46.5°N to 49.0°N) of the northern fore arc (Figure 1c). Fore arc basins and uplifts in western Washington—from south to north, the Tacoma, Seattle, Everett, and Bellingham basins—are defined mainly on the basis of high-amplitude geophysical anomalies, with faults hypothesized where the gradients are strongest [*Kelsey et al.*, 2012; *Blakely et al.*, 2002; *Brocher et al.*, 2001; *Danes et al.*, 1965]. Many of the faults found between the basins and the adjacent uplifts are active and present substantial seismic hazards to the region (Figures 1c and 2–6). Geodetic studies show that the Puget Sound region is undergoing north-south contraction averaging ~3 mm/yr to 4.4 mm/yr, rates are higher in the southern fore arc (6–10 mm/yr) and fall off to near zero in British Columbia, Canada. [*Wells et al.*, 1998; *Wells and Simpson*, 2001; *Mazzotti et al.*, 2002; *Hyndman et al.*, 2003; *McCaffrey et al.*, 2007; *McCaffrey et al.*, 2013].

Figure 1. (a) Map of Japan showing location of Japan Trench (hachured curve) and locations of triggered upper plate earthquakes (black circles, sizes scaled to magnitudes) following the 11 March 2011 *M*9 Tohoku-oki earthquake (dashed rectangle outlines approximate rupture plane). Thin lines are contours of the depths to the plate interface, at 20 km intervals. (b) Map of the Cascadia subduction zone showing locations of known late Holocene upper plate earthquakes. Dashed rectangle shows location of enlarged map the Puget Sound region. (c) Enlarged map of Puget Sound region showing locations of active faults and principal geographic features. Tinted boxes show locations of Figures 2–6. Orange stars in Figures 1b and 1c show locations of sites described in the text and shown on Figure 7: SI = Saanich Inlet, SW = southwest Washington, CE = Coquille estuary, BL = Bradley Lake, and SRD = Snohomish River delta. Black curves (dashed where inferred) are surface traces of fault zones with Holocene earthquakes, with names labeled (ending F or FZ refers to fault or fault zone): BBF = Birch Bay, SPF = Sandy Point, BCF = Boulder Creek, DDM = Darrington Devils Mountain, SWIF = South Whidbey Island, SF = Seattle, LCBC = Lake Creek/Boundary Creek, SMFZ = Saddle Mountain, TF = Tacoma, and OF = Olympia. Fore arc basins (enclosed by blue hatched curves) are labeled: BB = Bellington, EB = Everett, SB = Seattle, and TB = Tacoma basins; fore arc uplifts are labeled SU = Seattle uplift.

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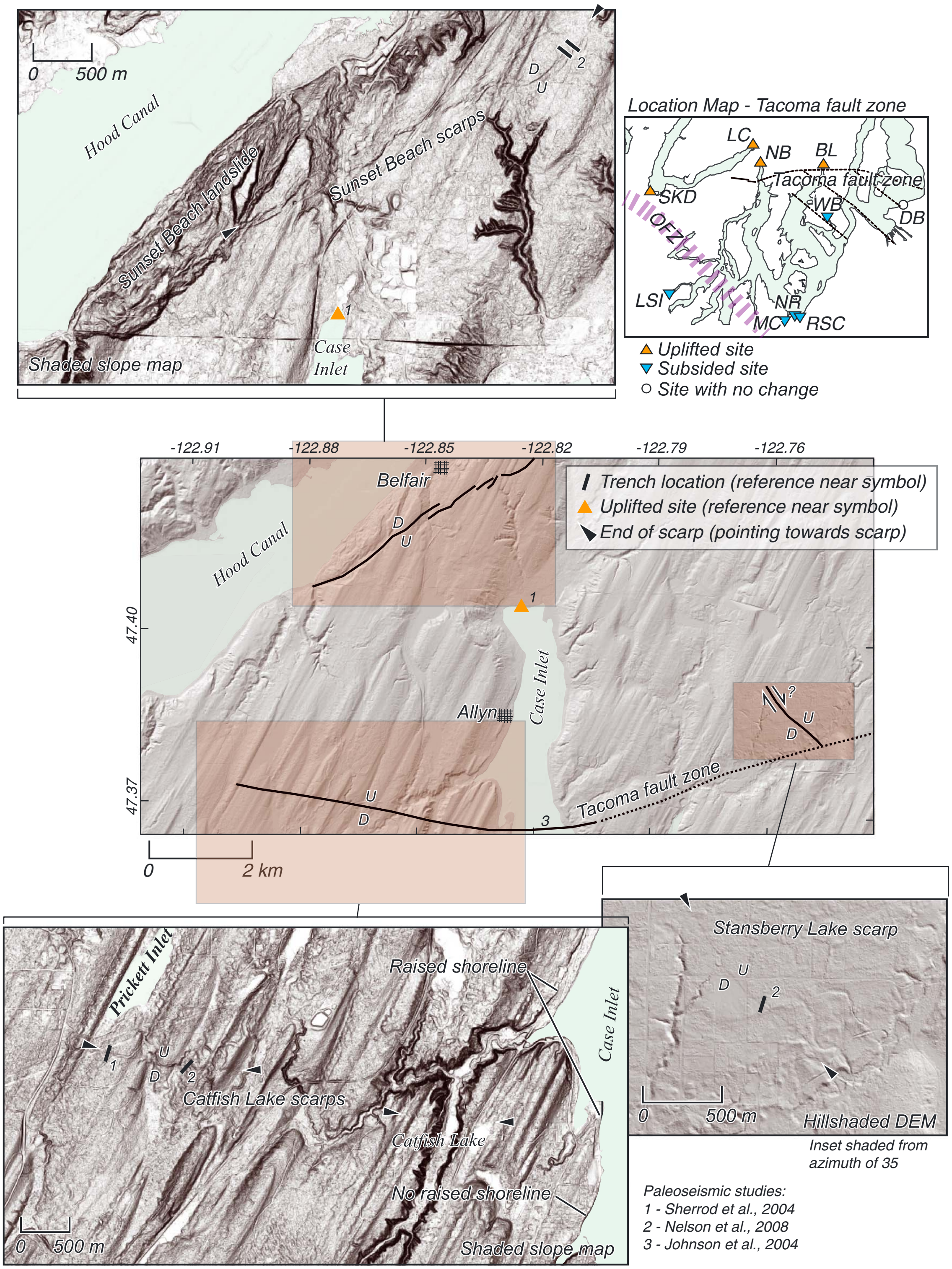

Figure 3. Shaded relief LiDAR. image of the Tacoma fault zone showing locations of paleoseismic features. Locations of sites described in the text and shown on small location map: LC = Lynch Cove, SKD = Skokomish River delta [*Martin*, 2012], NB = North Bay, BL = Burley Lagoon, DB = Dumas Bay, WB = Wollochet Bay, LSI = Little Skookum Inlet, MC = McAllister Creek, ND = Nisqually River, and RSC = Red Salmon Creek.

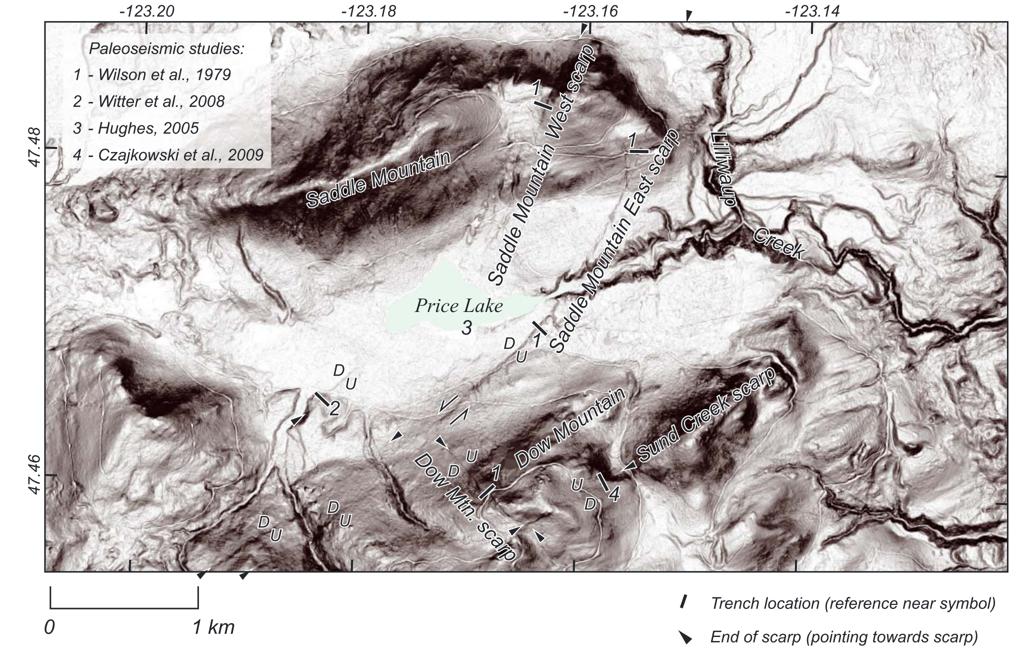

Figure 4. Shaded LiDAR slope map of the Saddle Mountain fault zone showing locations of scarps and previous paleoseismic studies.

3. Cascadia Paleoseismic Records

Before considering the possibility that a great subduction thrust earthquake triggered the 1200–900 cal years B.P. cluster, we first establish the existence and significance of this cluster. Prior to this study, an apparent surge in crustal earthquake rate was based only on qualitative perception, so to establish the existence of the cluster rather than rely on the published record alone, we conducted a more rigorous assessment of the earthquakes with age ranges within the last 2000 years. *Bucknam et al.* [1992] first noted evidence for

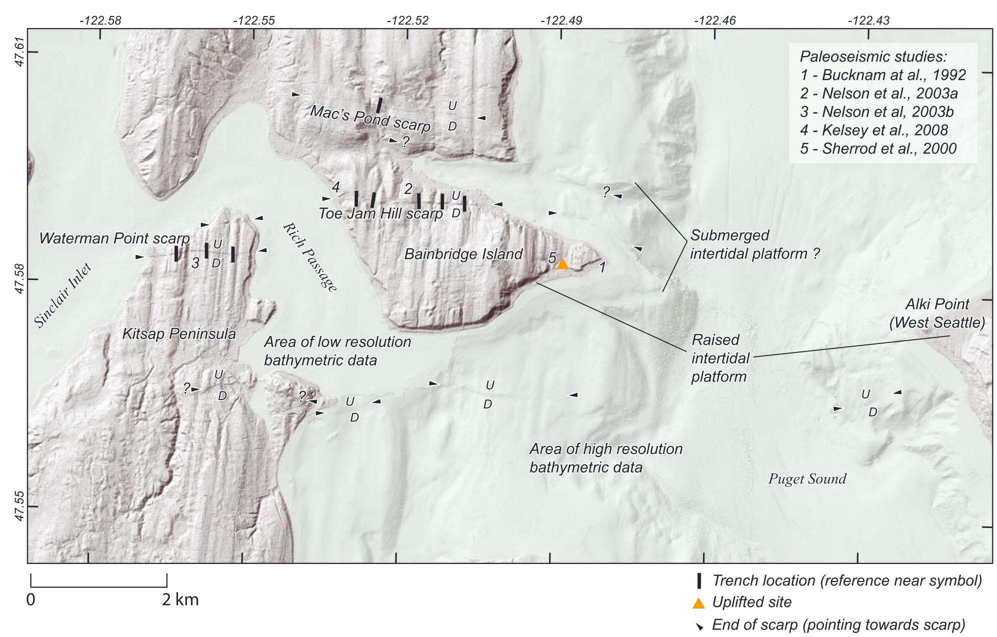

Figure 5. Shaded relief LiDAR and bathymetric image of the Seattle fault zone showing locations of scarps and previous paleoseismic studies.

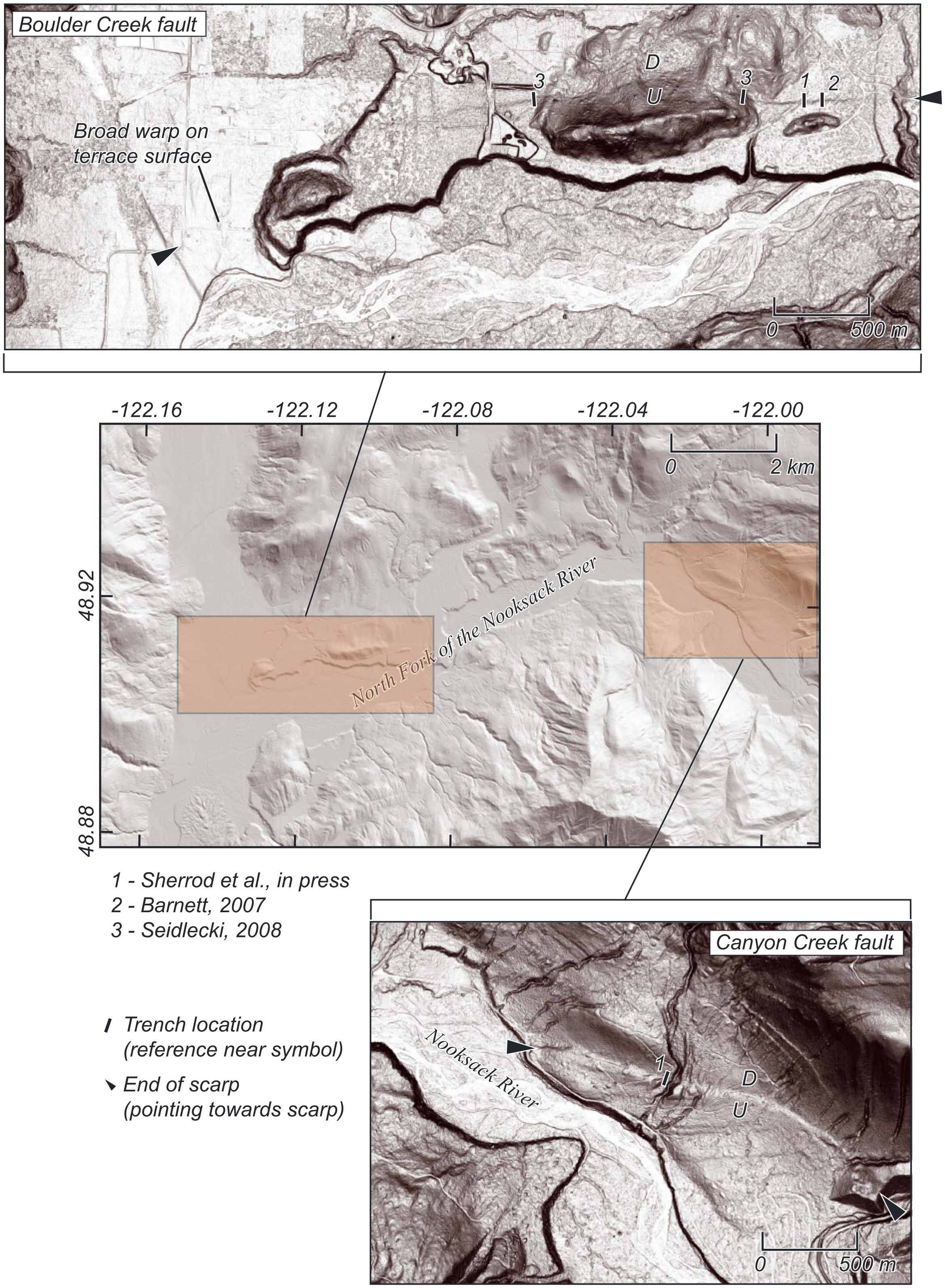

Figure 6. Shaded LiDAR slope map of the image of the Boulder Creek and Canyon Creek fault zones showing locations scarps and previous paleoseismic studies.

Table 1. Paleo-Earthquakes on Upper Plate Faults in Western Washington[a]

Fault Zone/Strands	Name	EQ Age (cal years B.P.)	Scarp Height (m)	Motion	Reference
		Seattle FZ			
(1) Master	RP–7500 yr B.P.	7200–6900	U	D	*Sherrod* [2000]
(1) Master	RP–1100 yr B.P.	1050–1020	U	D	*Bucknam et al.* [1992], *Nelson et al.* [2003a], and *Atwater* [1999]
(2) Vasa Park	Vasa	12,880–12,310	2.0	D	*Sherrod* [2002]
(3) Mac's Pond	Spotted Frog	<1020	2.9	D	*Nelson et al.* [2003b]
(4) Waterman	Madrone	1420–1710	3.9	D	*Nelson et al.* [2003b]
(5) Toe Jam	Mossy Lane	~1200	3.5	D	*Nelson et al.* [2003a]
(5) Toe Jam	Crane Lake	6000–2500	5.2	D	*Nelson et al.* [2003a]
(5) Toe Jam	Crane Lake	2500–1900	5.2	D	*Nelson et al.* [2003a]
(5) Toe Jam	Crane Lake	1700–1200	5.2	D	*Nelson et al.* [2003a]
(5) Toe Jam	Saddle et al.	~1500	??	D	*Nelson et al.* [2003a]
(5) Toe Jam	Bears's Lair	~15,000	??	D	*Nelson et al.* [2003a]
		South Whidbey Island FZ			
(1) Brightwater	Beef Barley	~15,000	<1m	O?	*Sherrod et al.* [2008]
(2) Crystal Lake	Mtn Beaver	<11,690	1.9	D	*Sherrod et al.* [2008]
(3) main	Whidbey Isl.	3200–2800	U/S	D	*Kelsey et al.* [2004]
Olympia FZ	Little Skookum	1150–980	S		*Sherrod* [2001]
Tacoma FZ	Case Inlet	1235–815	U	D	*Sherrod et al.* [2004]
		Boulder/Canyon Creek FZ			
Boulder Creek	Hornet	1165–699	3.4	D	*Sherrod et al.* [2014], *Barnett* [2007], and *Seidlecki* [2008]
Boulder Creek	Hornet	3344–2990	3.4	D	*Sherrod et al.* [2014], *Barnett* [2007], and *Seidlecki* [2008]
Boulder Creek	Hornet	~7700	3.4	D	*Sherrod et al.* [2014], *Barnett* [2007], and *Seidlecki* [2008]
		Bellingham Coastal FZ			
(1) Sandy Point	Sandy Point	>2100	U	D	*Kelsey et al.* [2012]
(1) Sandy Point	Sandy Point	~2100	U	D	*Kelsey et al.* [2012]
(1) Sandy Point	Sandy Point	<2100	U	D	*Kelsey et al.* [2012]
(2) Birch Bay	Birch Bay	1280–1070	U/S	D	*Kelsey et al.* [2012]
		Saddle Mtn/Canyon River FZ			
(1) Frigid Creek	Hummingbird	3800–415	2.9	D	*Blakely et al.* [2009]
(2) Saddle Mountain East	Quarry	~1100	8.0	O	*Wilson et al.* [1979], *Witter et al.* [2008], and Barnett et al. (2014)[b]
(3) Canyon River	Canyon River	<1630	~8	O	*Walsh and Logan* [2007]
Boundary/Lake Creek FZ	Nelson	2000–600 (2 eqs)	1.0	O	*Nelson et al.* [2007]
		Darrington-Devils Mtn FZ			
(1) Darrington-Devils Mtn	Lake Cavanaugh	<2200	1.0	O	*Personius et al.* [2009]
(2) Utsalady Pt.	Duffers	2200–1200	4.3	O	*Johnson et al.* [2004b]
(2) Utsalady Pt.	Duffers	500–100	4.3	O	*Johnson et al.* [2004b]

[a]Fault zones and their respective segments or strands are labeled as in Figure 8. Strands with the same name and/or number may have multiple entries, corresponding to multiple earthquakes. In column "Scarp Height", U and S indicate sites of coastal uplift and subsidence, respectively. Motion designations D and O refer to dip-slip or oblique-slip motion, respectively.
[b]E. A. Barnett et al., unpublished manuscript, 2014.

similarly-aged paleo-earthquakes on the Seattle and Saddle Mountain faults and what was later to be called the Tacoma fault, in the southern Puget Sound region, building on an earlier work of *Wilson et al.* [1979]. Subsequent studies provided evidence confirming activity on these and on the Boulder Creek and Olympia faults, evidence that we summarize in section 3.1. Although not ascribable to specific causative faults, *Bourgeois and Johnson* [2001] found evidence for liquefaction at the Snohomish River delta at the northern end of the

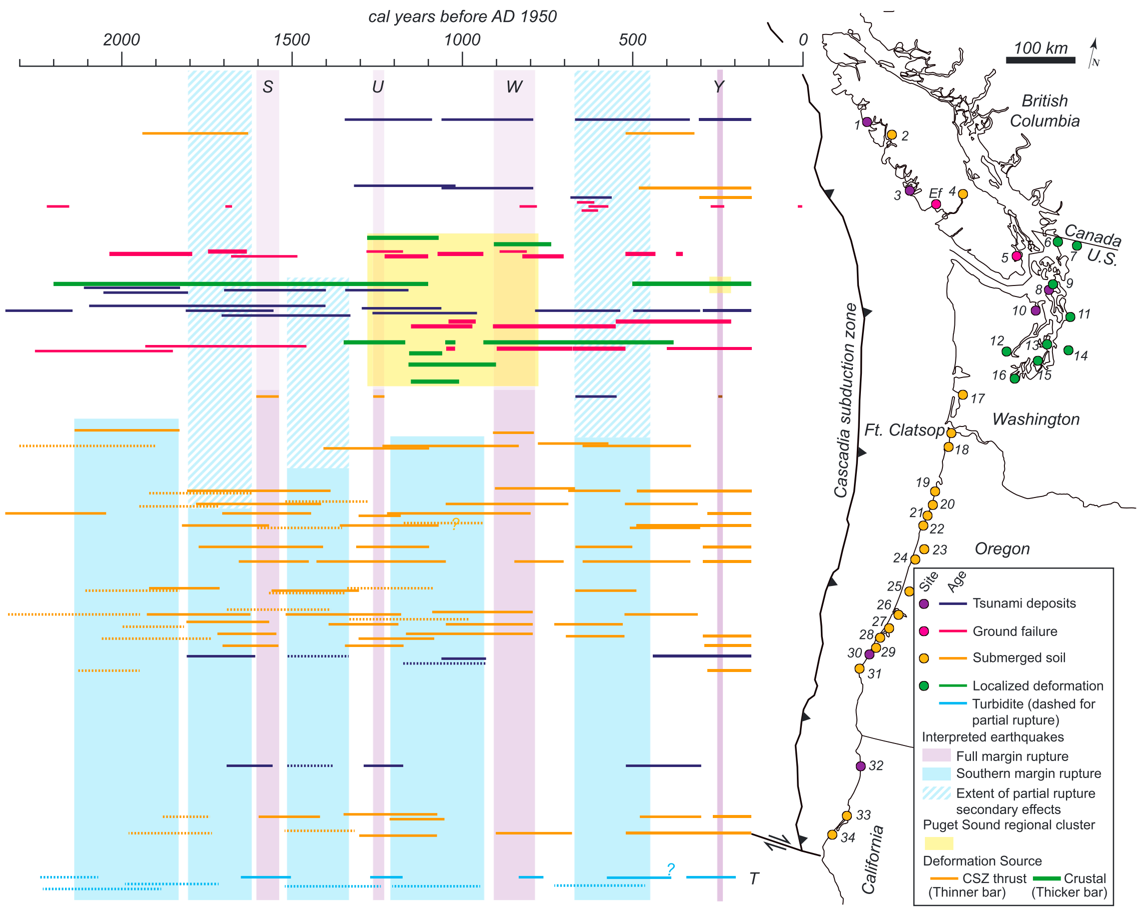

Figure 7. Estimated paleo-earthquake ages in the Puget Sound region during the last 2000 cal years B.P. and of Cascadia subduction zone paleo-earthquakes. (right) Index map showing the mapped location of observations constraining earthquake dates, with numbers indicating relevant references listed in section A. (left) Horizontal lines span estimated ages (horizontal axis) distributed according to the latitude of the observations constraining the ages. Thinner and thicker lines correspond to CSZ thrust and crustal earthquakes, respectively, colors indicate the type of observations (see legend), and solid and dashed lines denote inferred association with full and partial margin ruptures, respectively. Age ranges for turbidites at the bottom of the figure show no geographic significance, and are taken from *Goldfinger et al.* [2012]. Vertical bands in purple span inferred age ranges and latitudinal extent of great CSZ earthquakes. Purple bands correspond to full margin ruptures, referred to as events Y, W, U, and S, with lighter purple shading highlighting an absence of subsidence observations (orange age intervals) and a possible partial margin rupture interpretation. Light blue bands correspond to inferred partial margin ruptures and light blue hachured areas show extent of secondary effects from partial ruptures (see text). The vertical darker yellow band spans 1200–900 cal years B.P. when Puget Sound crustal earthquakes clustered, and lighter yellow bands denote possible triggering of just one or two crustal events. Age ranges of Puget Sound region crustal earthquakes are shown for the following fault zones (labels refer to Figure 1): Birch Bay = site 6, BB; Boulder Creek = site 7, BCF; Darrington-Devils Mountain = site 9, DDMFZ; Seattle = sites 12, 13 SFZ; Saddle Mountain = site 14, SMFZ; Tacoma = site 15, TFZ; Olympia = site 16, OFZ; and unknown sources in the Snohomish River delta (liquefaction features, SRD) and possibly near Saanich Inlet (debris flows) = sites 11 and 5.

Puget Sound region resulting from ground shaking in a series of large earthquakes in the last 2000 years, including earthquakes in the cluster (Figure 1c). The cluster may eventually encompass a few more faults—late Holocene earthquakes are noted on several other fault zones but as of now, the ages are too poorly constrained to include these earthquakes in the 1200–900 cal years B.P. cluster [*Barnett et al.*, 2010; *Gower et al.*, 1985].

Establishing the significance of the cluster and a causal connection with a CSZ thrust earthquake requires demonstrating that clusters do not occur frequently, thereby making a cluster quickly following a CSZ thrust event an improbable chance occurrence. To demonstrate this, in the second part of our analysis, we estimate pseudo rates for paleo-earthquakes in the Puget Sound region over the past 16,000 years. We use probability density functions (PDFs) describing the ages of paleo-earthquakes from published sources for nine active fault zones in the Puget Sound region (section 3.2, Table 1). We discuss the completeness of this 16,000 year-long record in section 5.

After assessing the existence and uniqueness of the 1200–900 cal years B.P. cluster, we then compare ages and rates of 1200–900 cal years B.P. paleo-earthquakes identified in the Puget Sound region to ages of CSZ thrust earthquakes (section 3.3) to determine whether the CSZ earthquakes pre- or post-dated the Puget

Sound region crustal cluster. We consider published estimates of ages of CSZ earthquakes constrained by a variety of data types and analyses, and note that our goal is to present a comprehensive compilation. New radiocarbon ages that better constrain the timing of the great CSZ thrust earthquake temporally closest to the crustal cluster particularly facilitate this comparison [*Atwater and Griggs*, 2012].

3.1. Puget Sound Region Crustal Earthquake Record

The earthquake age data we analyzed and show in Figure 7 represent the major crustal paleo-earthquakes in the Puget Sound region that led to the initial qualitative identification of clustering between 1200 and 900 cal years B.P. We constrain ages of the subset of the paleo-earthquakes documented on crustal faults within the Puget Sound region within the last 2000 years using a Bayesian analysis of radiocarbon ages in the Oxcal radiocarbon calibration program [*Bronk Ramsey*, 1995; *Lienkaemper and Bronk Ramsey*, 2009]. Bayesian analysis in paleoseismology is particularly powerful because it incorporates prior chronologic information such as stratigraphic order, laboratory uncertainty, ages of known events, and historical constraints, to derive probability distributions describing the likely ages of past events (entered in the sequences in the Oxcal model as boundaries).

In Figure 7 we show the results of our Oxcal Bayesian analysis as a simple sequence of geographically ordered calibrated radiocarbon estimated age ranges of Puget Sound region earthquakes. Ages of samples from pre-earthquake soils in the trenches and wetlands are entered as phases (a set of related samples) because their true stratigraphic order is not known. We show calibrated radiocarbon ages as the 2 standard deviation range in cal years B.P. relative to A.D. 1950 (green horizontal bars in Figure 7), and round to the nearest multiple of five where reported standard error is below 50 years and to nearest multiple of 10 where standard error is between 50 and 100 years [*Stuiver and Polach*, 1977]. We provide details about the data constraining the paleo-earthquakes on each fault in the subsections that follow. The age range of each paleo-earthquake on the left side of Figure 7 is plotted at latitude of the site providing the constraining observations, which is plotted on the map on the right, and numbers next to each site correspond to relevant publications keyed to the reference list in the section A.

3.1.1. Olympia Fault Zone

The Olympia fault zone separates 3.5–6.0 km of sedimentary strata in the Tacoma basin from Eocene volcanic rocks located in the Black Hills to the south [*Pratt et al*., 1997; *Brocher et al*., 2001; *Clement et al*., 2010]. *Sherrod* [2001] found submerged forests and high marsh soils suggesting abrupt changes in relative sea level at four coastal localities along the trace of the Olympia fault zone (Figures 1 and 2). At localities along McAllister Creek and the Nisqually River, tidal-flat mud buries high marsh soils. Sand dikes cutting marsh soils and sand volcanoes lying on buried marsh surfaces attest to strong ground shaking at the same time as submergence of a high marsh soil at McAllister Creek. Dramatic changes in plant macrofossil and diatom assemblages across these sharp stratigraphic contacts confirm rapid submergence. At both of these localities, low-marsh and tidal-flat diatoms in laminated mud overlying high marsh peat also indicate a rapid environmental change from high marsh to tidal flat at the same time as strong ground shaking. At Little Skookum Inlet and Red Salmon Creek, salt marsh peat buries Douglas-fir stumps in growth position. Salt-marsh peats immediately above the buried forest soil contain diatoms indicative of low marsh and tidal-flat environments.

At all of the above sites, inferences from modern diatom assemblages suggest at least 1 m of subsidence at each site. Up to 3 m of submergence at Skookum Inlet is possible given that the lowest elevation of modern trees at the site is 3 m above the buried forest soil. Coseismic subsidence best explains abrupt burial of Puget Sound region soils along the north side of the Olympia structure between 1160 and 1010 cal years B.P., estimated in our Oxcal model using a wiggle-matched pair of radiocarbon samples from a single Douglas-fir stump killed by subsidence and burial by tidal flat muds and salt marsh peat (site 16 in Figure 7).

3.1.2. Tacoma Fault Zone

The Tacoma fault zone separates the southern edge of the Seattle uplift from the adjacent edge of the Tacoma basin. Inferences from seismic tomography and seismic reflection data suggest up to 6–7 km of north-side-up structural relief on the top of Eocene basalt along the western end of the Tacoma fault zone [*Sherrod et al*., 2004; *Johnson et al*., 2004a]. *Johnson et al*. [2004a] show that a north dipping thrust fault, the Tacoma fault, bounds the northwestern margin of the Tacoma basin, that a south-dipping monocline bounds the northeastern margin of the Tacoma basin (Figures 1–3), and estimate Quaternary structural relief as much as 350–400 m and thus a minimum Quaternary slip rate of ~0.2 mm/yr.

Geologic evidence for past activity of the Tacoma fault includes uplifted tidal flat deposits and shorelines along Hood Canal, Case Inlet, and Carr Inlet (Figures 1c and 3). Radiocarbon ages of peat and delicate plant fossils suggest that freshwater peat began forming over tide flat muds between 1300 and 900 cal years B.P., indicating uplift of the tidal flats in that time period [*Bucknam et al*., 1992; *Sherrod et al*., 2004]. Light detection and ranging (LiDAR) surveys along the Tacoma fault zone reveal faults scarps near Belfair and Allyn, Washington. These scarps, as high as 4 m in places, suggest surface rupture from past earthquakes on the Tacoma fault in the recent past. Trenches across the Catfish Lake scarp showed folding of glacial deposits and recent soils from a late Holocene earthquake, likely the same earthquake associated with locally uplifted shorelines along Case Inlet and Hood Canal (Figure 3), raised as much as 4 m in the late Holocene between 1240 and 850 years B.P. [*Sherrod et al*., 2004]. Additional trenches across two other scarps, both located in the up-thrown block of the Tacoma fault zone, provide evidence of right lateral oblique and normal faulting between 1300 and 600 years B.P. [*Nelson et al*., 2008a]. All of these ages are consistent with a large regional earthquake on the Tacoma fault zone between 1170 and 900 cal years B.P. (site 15 on Figure 7).

3.1.3. Saddle Mountain Fault Zone

The Saddle Mountain fault zone (SMFZ) is a ~45 km long zone of left-lateral, oblique-reverse deformation (Figure 1c) that accommodates northward shortening of crust beneath the Puget Sound region inland of the Olympic massif [*Blakely et al*., 2009]. A series of three parallel scarps at the western edge of the Puget Sound region attest to past large earthquakes on the SMFZ (Figure 4).

Paleoseismic observations suggest at least two large earthquakes on the SMFZ, the latest of which was about 1100 cal years B.P. [*Wilson et al*., 1979; *Witter et al*., 2008; *Hughes*, 2005]. A trench across the Saddle Mountain East fault revealed a 1 m wide zone of sheared and brecciated basalt, with an ~3.5 m of left-lateral-oblique-reverse offset on the overlying late Quaternary glacial till (~16 ka old). The best evidence for the age of the last earthquake comes from Price Lake, a small lake formed when the Saddle Mountain East fault scarp blocked a tributary to Lilliwaup Creek (Figure 5). Ages of wood samples from submerged stumps in Price Lake are 1370–1060 cal years B.P. and 7160–4650 cal years B.P. [*Wilson et al*., 1979], and samples of wood from stumps in nearby marshes and ponds along the same scarp are 1990–1090 cal years B.P., 1270–930 cal years B.P., and 10,240–9410 cal years B.P. [*Wilson et al*., 1979]. Taken as a group, the stump samples suggest at least one large earthquake on the Saddle Mountain East fault strand between 1990 and 930 cal years B.P., and an earlier possible earthquake, albeit poorly constrained, sometime after 4650 cal years B.P.

Hughes [2005] uses submerged tree stumps in Price Lake and marsh stratigraphy in adjacent wetlands to unravel the postglacial history of the SMFZ, relying on radiocarbon analyses of delicate plant remains from deposits above and below a submerged forest soil. Conifer cones and needles from a submerged forest soil between the two scarps yielded ages between 1290 and 1080 cal years B.P., while a western red cedar cone from a submerged forest soil west of the Saddle Mountain West scarp had a similar age of 1270–1050 cal years B.P.. We use a sequence of radiocarbon samples from submerged stumps and wetland samples in our Oxcal model to estimate the age of the last earthquake on the SMFZ between 1160 and 1060 cal years BP (E. Barnett, personal communication, 2013) (site 12 on Figure 7).

3.1.4. Seattle Fault Zone

Geological and geophysical studies identify an east trending zone of active but mostly concealed reverse faulting bisecting the Puget Lowland at the latitude of Seattle, Washington. Uplifted Tertiary volcanic rocks to the south of the Seattle fault zone and down-dropped Tertiary and Quaternary sediments to the north in the Seattle basin create one of the strongest gravity anomalies in the continental United States [*Danes et al*., 1965]. Studies relying on high-resolution aeromagnetic, gravity, geologic mapping, and seismic reflection data locate several subparallel fault strands within an east trending zone, referred to as the Seattle fault zone, and interpret these strands as a south-side-up set of imbricated thrust faults or passive roof duplex [*Blakely et al*., 2009; *Johnson et al*., 1994, 1999; *ten Brink et al*., 2002; *Brocher et al*., 2004].

The best geological evidence for a large earthquake on the Seattle fault about 1100 years ago is a conspicuous platform bordering the shoreline of southern Bainbridge Island in the central Puget Sound region (Figure 5). *Bucknam et al*. [1992] documents evidence that this platform is an intertidal wave cut platform, cut on Oligocene marine sedimentary rocks and Miocene nonmarine sedimentary rocks, and uplifted as much as 8 m in a single earthquake about 1100 years ago. Cores from a former intertidal marsh sitting on the uplifted platform show an abrupt shift from tide flat diatoms to freshwater taxa resulting from ~7 m of uplift during an earthquake on the Seattle fault about 1100 cal years B.P. [*Sherrod*, 2000]. Brackish water diatom and plant fossils indicate that the marsh lay near sea level between ~7500 and ~1100 cal years B.P., suggesting that no other uplift similar to that in ~1100 cal years B.P. occurred in the past 7500 years.

Secondary effects of this large earthquake ~1100 cal years B.P. include a Puget Sound tsunami, landsliding into Lake Washington, and liquefaction, all of which are consistent with a blind reverse rupture on the main strand of the Seattle fault [*Atwater and Moore*, 1992; *Bucknam et al*., 1992; *Jacoby et al*., 1992; *Karlin and Abella*, 1996; *Huntley and Clague*, 1996; *Bourgeois and Johnson*, 2001; *Karlin et al*., 2004; *Arcos*, 2012a, 2012b; *Martin and Bourgeois*, 2011]. Tree ring width pattern matching showed that the tsunami and the landslides occurred in the same several months; the rings of a tsunami-deposited log at West Point, Seattle, matched those of slide-block trees found standing on the bottom of the lake [*Jacoby et al*., 1992]. *Atwater* [1999] narrowed the time of the earthquake to 1050–1030 cal years B.P. using a wiggle-match analysis of calibrated radiocarbon ages on wood samples from a Douglas-fir log found in a tsunami deposit (site 13 on Figure 7).

LiDAR images reveal several fault scarps along the Seattle fault. Paleoseismic trenches across these scarps show evidence for up to five surface-rupturing earthquakes in the past 2500 years, with two earthquakes close together in time about 1100 cal years B.P. [*Nelson et al*., 2003a, 2003b]. These scarps, all north-side-up (opposite to the south-side-up evidence for the buried master fault), likely result from ruptures on backthrusts to the south-dipping master or on flexural slip faults in a north dipping monocline [*Nelson et al*., 2003a; *Kelsey et al*., 2008]. *Kelsey et al*. [2008] infer that the backthrusts/flexural slip faults root at shallow seismogenic depths and do not require simultaneous movement on the south-dipping master fault.

3.1.5. Boulder Creek Fault Zone

LiDAR images over most of Whatcom County reveal two fault scarps along the North Fork of the Nooksack River (Figure 6; site 7 on Figure 7). One scarp lies on the mapped trace of the Boulder Creek fault (BCF) and the other scarp lies along a fault near Canyon Creek (informally referred to as the Canyon Creek fault, or CCF). The BCF scarp is ~4.3 km long, south-side-up, and has a maximum preserved height of ~4 m. Sediments preserved in a small scarp-dammed wetland along the BCF record each earthquake, containing delicate plant remains required for accurately dating past earthquakes. The CCF scarp is ~2 km long, south-side-up, and has a maximum height of ~4 m. Four trenches across the BCF scarp, one trench across the CCF scarp, and the scarp-dammed wetland show evidence for three Holocene earthquakes, the youngest of which occurred between 910 and 740 cal years B.P. [*Sherrod et al*; 2013; *Barnett*, 2007; *Seidlecki*, 2008].

3.2. Estimating the Paleoseismic Crustal Earthquake Rate

Inference of triggering typically is based on the occurrence of two improbable or infrequent events within close temporal and/or spatial proximity of one another, because the joint probability of both being nearly concurrent and/or coincident is exceedingly low. Triggering often is inferred from an extraordinary increase in seismicity rate that follows an infrequent posited triggering event [*Stein*, 1999; *Freed*, 2005; *Steacy et al*., 2005], in this study we test for triggering by asking if a cluster of crustal earthquakes, manifest as an extraordinary increase in the crustal earthquake rate (effectively a PDF describing the probability of earthquake occurrence) and follows an infrequent great CSZ thrust earthquake. Estimates of modern seismicity rates have uncertainties that arise from assumptions about how to treat the temporal and spatial variability but not from imprecision in origin times. We estimate a pseudo-rate of paleo-earthquakes because their occurrence times are not precisely known but instead, as noted in section 3, are described in terms of PDFs. To estimate this pseudo rate, we use a more complete catalog and extending over a much longer time interval than analyzed using the Oxcal model and shown in Figure 7, compiled from this and other published studies (see Table 1). Several of the Puget Sound region fault zones contain multiple strands or segments, with evidence of slip events on one at times that may or may not overlap with events on other strands (Figure 8). Following *Kelsey et al*. [2008], we assume in our rate calculations that each strand may slip independently of the others.

For some events, only a lower or upper bound, or an educated guess of a mean and uncertainty may be available based on indirect evidence. A Gaussian PDF often is assumed in radioisotope laboratory age dates, each reported as mean event date and standard deviation, σ. We combine the Oxcal model results of section 3.1 with age dates from the published literature, and approximate PDFs with Gaussian functions, combined with uniform distributions when only a bounding estimate is known (Figure 8). This simplification captures the salient features of the relative uncertainties and makes little difference to our results. When only an approximate mean date exists, we assume σ ~300 years, guided by a median and mean σ values of 105

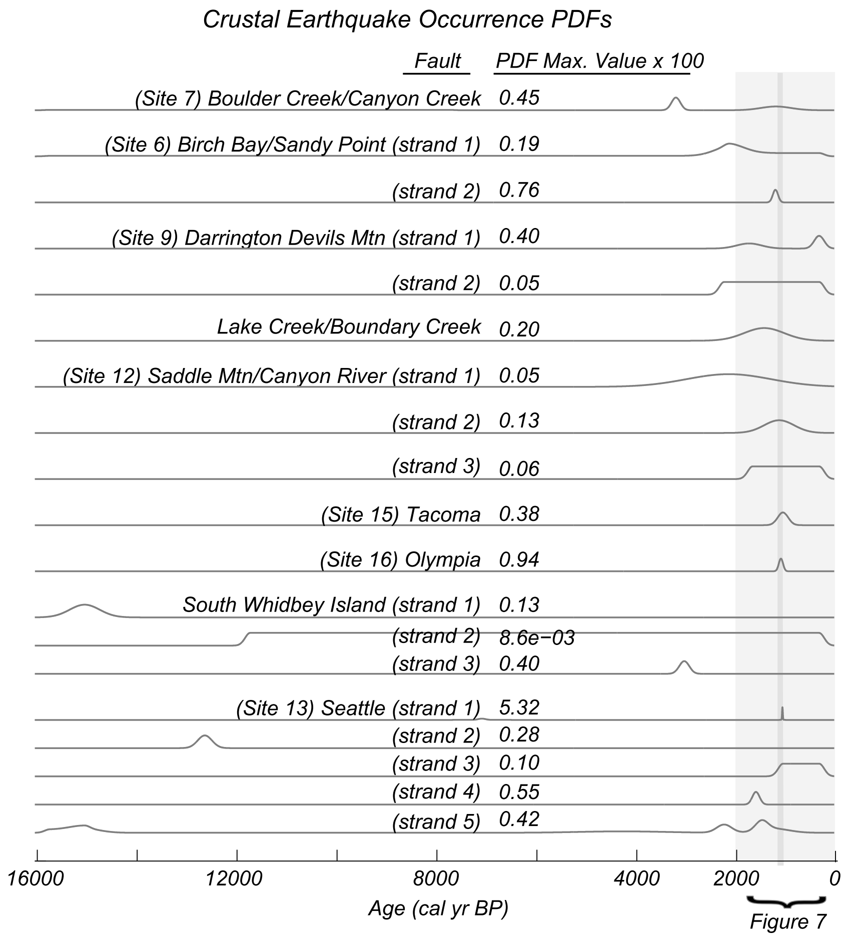

Figure 8. Probability density function time histories for Puget Sound region fault paleo-earthquakes. Location of each fault system is shown in Figure 1 and corresponding site numbers from Figure 7 are in parentheses. Each row corresponds to a single fault strand of the system, showing the PDFs describing the likelihood of each earthquake occurrence identified in the paleoseismic record. PDFs are normalized so that the integrated probability of each equals one, but for plotting, are scaled for clarity so that the peak probability for each segment has the same height. The true amplitude of the peak probability is listed to the right of the name of each fault strand. The vertical gray line indicates the time of the maximum earthquake rate, and the gray shaded region, the time interval of Figure 7.

and 225 years, respectively, for the set of measured dates. When only a lower or upper bound exists, we construct a PDF comprised of two half-Gaussians with one peak at the bounding age and σ ~100 years, joined by a uniform (flat) distribution. If a lower bound, the other half-Gaussian peak is at 300 cal years B.P., since the event has not occurred in the last few hundred years. If an upper bound, the other half-Gaussian peak is at the start of the record at 16,000 cal years B.P.

We estimate a pseudo rate of paleo-earthquakes in moving time windows by summing the probability that an earthquake occurred on each fault segment in each window. Figure 9a shows the resulting rates. What is the likelihood that the observed earthquake rate between 1200 and 900 cal years B.P. occurred simply by chance, i.e., if the earthquakes occurred randomly in time? We find the distribution of earthquake rates expected if earthquake occurrence times were randomly distributed but known with the same uncertainties, so the reference chance rates also reflect our ability to measure them in the same way as the true rates. For a single such reference rate, we calculate a probability for each fault in a randomly selected time interval and as for the unscrambled rate history, these probabilities are then summed. This is repeated for numerous trials and the resulting distribution shows that the observed rate between 1200 and 900 cal years B.P. is highly improbable if the earthquakes were independent uniformly distributed events (Figures 9a and 9b). Specifically, the maximum rate at ~1100 cal years B.P. clearly exceeds the rates estimated by the randomized

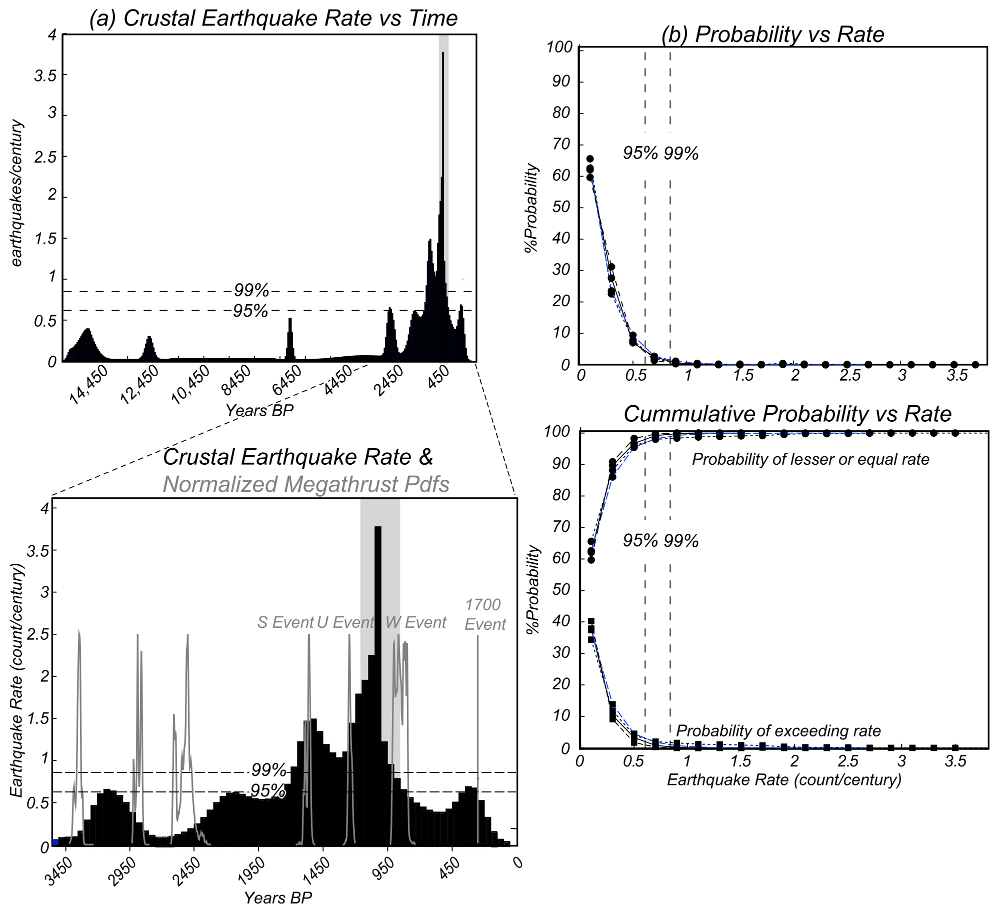

Figure 9. Puget Sound region crustal earthquake rate and megathrust time histories. (a, top) Rate of Puget Sound crustal earthquakes with time, calculated as a moving average for the entire 16,000 year record (see text). (a, bottom) Expanded view of the crustal rate for the last 3500 years with the probability density functions (PDFs) describing the dates of earthquakes on the Cascadia plate interface (gray lines); the PDFs are scaled to plot with the same amplitude for clarity. In both frames, the dashed horizontal lines show the rates having 95% and 99% chance of being observed if the times of earthquakes were randomly distributed (see Figure 9b and text). Shaded bar shows the 1200–900 cal years B.P. interval previously identified as the Puget Sound crustal paroxysm. (b, top) Probabilities of crustal earthquake rates if earthquake occurrence on each fault was independent with a uniform distribution process derived from the distribution of rates estimated by randomly sampling the PDFs in Figure 8. The window length used to derive the rates makes insignificant difference; rates derived for 100, 200, and 400 year windows are shown by different lines. (b, bottom) Results are plotted as cumulative probabilities and as the remainder. Dashed vertical lines are plotted where the probabilities are 5% and 1% such that all rates estimated to the left of the lines have 95% and 99% chance of being observed.

sampling that have a 5% and 1% chance of occurring by chance (Figure 9a). The previously identified 1200–900 cal years B.P. interval qualitatively adequately captures the width of the peak of the rate estimates centered at the ~1100 cal years B.P. maximum (Figure 9a).

The errors required to change these results are unlikely. The clustering largely results from the relatively precise and overlapping date ranges of earthquakes on one segment of the Seattle fault (the master fault), and the Olympia and Tacoma faults, which collectively account for about 80% of the total rate. Adding to these other segments of the Seattle, Saddle Mountain, Boulder, and Boundary/Lake Creek fault systems each contribute just over 5%, accounting for over 95% of the total rate. Another interval with a rate comparable to that during this cluster missing from the record requires the unlikely omission of several earthquakes dated within uncertainties $<$ ~200 year and that occurred within a few hundred years or less of one another.

3.3. Cascadia Subduction Zone Thrust Earthquake Record

The CSZ, extending 1100 km from Vancouver Island to Cape Mendocino, offers tens of inferred earthquake histories with which the dating of the Puget Sound cluster can be compared. Herein, we summarize the histories of late Holocene CSZ thrust earthquakes from published studies (for details, we refer the reader to the references provided). We attempt to acknowledge objectively, discordant estimates or interpretations by simply ascribing greater uncertainty to them, as reconciliation is well beyond the scope of this study. Thus, to assess the timing of significant CSZ earthquakes, it seemed most appropriate to create 2 σ bar plots using the time interval preferred for each earthquake in their respective published chronology (Figures 7 and 9a), rather than recreate radiocarbon models in Oxcal. In addition to the age ranges of Puget Sound region crustal earthquakes discussed in section 3.1, Figure 7 also summarizes onshore evidence of great CSZ thrust earthquakes temporally and geographically, with references for the sources of the observations shown keyed to the numbers next to each site on the figure's map. Onshore evidence of CSZ thrust earthquakes includes submerged marsh and estuary soils (orange lines/circles in Figure 7), ground failure (thin red lines/circles in Figure 7), and tsunami deposits (purple lines/circles in Figure 7). For reasons noted below, we show only the temporal offshore record (cyan lines in Figure 7).

Estimates of the magnitudes of past CSZ thrust earthquakes are required for comparison with inferences about triggering from the modern record and other studies [*Gomberg and Sherrod*, 2014] but are perhaps even more uncertain than their ages. Thus, as for the ages, we attempt simply to capture the range of possibilities based on published results and clearly stated assumptions. We do know, with a high degree of certainty, that the CSZ is capable of producing $M \sim 9$ earthquakes, noting that an abundance of data and studies demonstrate the most recent great CSZ earthquake of 26 January 1700 ruptured the entire zone [*Satake et al.*, 1996; *Jacoby et al.*, 1997; *Yamaguchi et al.*, 1997; *Satake et al.*, 2003; *Atwater et al.*, 2005; *Thrush and Ludwin*, 2007; *Wang et al.*, 2013]. Additional evidence bearing on the potential for $M \sim 9$ CSZ ruptures is the absence of any documentation or physical manifestations of $M8+$ CSZ thrust earthquakes since 1811 when record keeping Europeans began occupying the region. This contrasts with plate boundary sequences in which great subduction plate interface earthquakes reruptured in several smaller events within a few hundred years. For example, in the Nankai Trough, the rupture area of a great earthquake in 1707 [*Furumura et al.*, 2011] broke in two pieces in 1854 and again in the 1940s [*Ando*, 1975], and offshore Colombia and Ecuador, the long rupture area of 1906 unzipped next in the series of three shorter ruptures of 1942, 1958, and 1979 [*Kanamori and McNally*, 1982]. Finally, we also know that at least today, the entire Cascadia plate boundary appears to be strongly coupled [*McCaffrey et al.*, 2007; *Burgette et al.*, 2009].

Despite the above observations, whether $M \sim 9$ events recur with any regularity or at all, and/or are punctuated by earthquakes in the $M8+$ range are questions that are well beyond the scope of this study. We simply assume that the geographic distribution of measurable deformation scales with the rupture dimensions of a CSZ thrust earthquake, and thus, that if plausibly temporally coincident, deformation observed with a greater variety of signatures and over a more widespread area indicates a larger earthquake. The signatures of CSZ thrust earthquakes comes from both offshore and onshore geologic records, and while we consider both, we focus more on the onshore record because of its more precise ages and size metrics (e.g., fault offset, uplift, and subsidence). Thus, in Figure 7 we plot both the geographic and temporal components of the onshore record, and only the temporal record of the offshore record. Offshore evidence of fault rupture comes from deep-sea turbidites that record sediment density flows that likely began as slides and debris flows in submarine canyons [*Goldfinger et al.*, 2012]. In Figure 7, vertical bands represent the age ranges and latitudinal extents of individual CSZ plate interface events inferred in other studies. Purple bands are interpreted as full margin ruptures, referred to as events "Y", "W", "U" and "S," after the soils they submerged [*Atwater and Hemphill-Haley*, 1997; *Atwater and Griggs*, 2012; *Atwater et al.*, 2004a]. No evidence exists for subsidence during these earthquakes north of southwest Washington (indicated in Figure 7 by vertical bands shaded a lighter purple, containing no orange bars that denote submerged soils), and it is permissible that these earthquakes ruptured only the southern half of the margin. Partial margin ruptures inferred in the literature extend from southwest Washington southward (indicated in Figure 7 by solid light blue vertical bands).

An important posited triggering earthquake in our comparison is the penultimate great CSZ earthquake, event "W" [*Atwater and Hemphill-Haley*, 1997] that is plausibly recorded in both the onshore and offshore geology from southern Oregon to British Columbia. The primary evidence for earthquake W comes from estuaries in southwestern Washington, where burial of soil W was originally estimated to be 1190–780 cal years B.P., derived from three ages of a buried shrub root obtained from a marsh outcrop at Fort Clatsop (site 17 on Figure 7) near the mouth of the Columbia River [*Atwater and Hemphill-Haley*, 1997; *Atwater et al.*, 2004a], similar in age to the Puget Sound region cluster between 1200–900 cal years B.P. New, more precise ages for the W event were obtained by averaging calibrated ages obtained on three samples of the same shrub roots places the onland age for the earthquake between 915 and 790 cal years B.P. (purple band labeled W in Figure 7 [*Atwater and Griggs*, 2012]). The ages of event W inferred from the estuarine data may correlate with the causative event of turbidite T3 (Figure 7, bottom, cyan line) inferred in *Goldfinger et al*. [2012], although the dates and uncertainties associated with the latter appear still to be matters of debate [see *Atwater and Griggs*, 2012]. Event W also, plausibly, was the same earthquake causing tsunami deposits (sites 1 and 3 in Figure 7), and some of the debris flow deposits in Saanich Inlet, British Columbia (site 5 in Figure 7; some inferred to result from shaking caused by local crustal earthquakes, as thicker lines indicate).

If all these deformation signatures record a single W event, it would represent a full-margin rupture. However, this inference requires an explanation of why subsidence of soil W was modest relative to that associated with the CSZ Y earthquake of 1700 (e.g., no more than half at Willapa Bay [*Atwater and Hemphill-Haley*, 1997]). This is exacerbated by poor preservation of soil W at many sites, most likely due to decomposition of the A horizon after burial in the weathering profile of soil Y [*Atwater and Hemphill-Haley*, 1997].

Like the W event, the U event [*Atwater et al*., 2004a] also may have left a record in both the offshore and onshore geology, spanning most of the CSZ margin. Subsidence of buried soil U has been dated concordantly from samples from Grays Harbor, Willapa Bay, and the Columbia River estuaries, and if combined, yield age for event U is 1265–1230 cal years B.P. [*Atwater et al*., 2004a]. Soil U ages correspond with ages ascribed to turbidite T4 in *Goldfinger et al*. [2012], although the latter are questioned in *Atwater and Griggs* [2012]. Event U also plausibly could have generated debris flow deposits in Saanich Inlet [*Blais-Stevens et al*., 2011] and tsunami deposits in Bradley Lake [*Kelsey et al*., 2005], as well as buried soils at Coquille estuary in southern Oregon [*Witter et al*., 2003] (sites 5, 30, and 29, respectively; Figure 7). If true, event U would be comparable in size to the 1700 earthquake, although a series of smaller earthquakes also would explain these observations.

4. Cascadia Subduction Zone Thrust Earthquake Triggering of Puget Sound Region Seismicity

Was the 1200–900 cal years B.P. clustering of crustal earthquakes triggered by a $M > 8.6$ paleo-earthquake on the Cascadia plate boundary, as the modern record from other subduction zones would suggest [*Gomberg and Sherrod*, 2014]? Could smaller CSZ thrust paleo-earthquakes have triggered the cluster? Were isolated $M > \underset{\sim}{}6.5$ crustal earthquakes triggered by CSZ thrust paleo-earthquakes? We addressed these questions in order, by comparing both the ages and rates of Puget Sound region paleo-earthquakes to the dates of late Holocene CSZ thrust earthquakes.

We conclude that the Puget Sound region cluster was not triggered by the CSZ thrust earthquakes most likely to have been $M > 8.6$ events. For reasons noted in section 3.3, the W and U events may have been full-margin ruptures, although partial margin rupture or a series of smaller distributed CSZ earthquakes cannot be ruled out. CSZ thrust event W falls between 915 and 790 cal years B.P. [*Atwater and Griggs*, 2012], suggesting that the 1200–900 cal years B.P. Puget Sound region cluster is older by as much as a few centuries. Age estimates on soils defining the CSZ U event of 1265–1230 cal years B.P. [*Atwater et al*., 2004a] clearly predate the cluster by almost 100 years (Figure 7). The Puget Sound cluster's most precisely dated member, the Seattle fault rupture of 1100–1070 cal years B.P., post and predates these two CSZ thrust earthquakes by more than a century (Figure 7). Figure 9 corroborates these interpretations, comparing the rate of crustal earthquakes, or the PDF describing their likelihood, with the dates of the likely largest CSZ thrust events. Figure 9a shows that the likely date of the W event in the range 915–790 cal years B.P. [*Atwater and Griggs*, 2012] postdates the peak in the crustal earthquake rate sufficiently to conclude that event W postdated the cluster of crustal earthquakes. This observation supports the findings of *Hagstrum et al*. [2004], who also used paleomagnetic measurements to infer that the Seattle earthquake of 1100–1070 cal years B.P. preceded event W.

The uniqueness of the 1200–900 cal years B.P. Puget Sound region crustal earthquake cluster within the long history of recurrent great interplate Cascadia earthquakes further argues against a causal relationship. Recurrence intervals of $M \sim 9$ interplate earthquakes averaging about 500 years have been inferred over the last 10,000 years or more from deep-sea turbidites [*Goldfinger et al*., 2012] and over the last 3500 years from subsided estuarine soils [*Atwater and Hemphill-Haley*, 1997]. If instead ancient earthquakes were smaller, they would recur even more frequently, strengthening this argument.

While a causal connection between the Puget Sound region crustal cluster and a smaller CSZ event(s) between the U and W earthquakes cannot be ruled out, we consider this implausible. The Puget Sound region cluster correlates with the southern margin rupture identified in *Goldfinger et al*. [2012] and possibly with observations that could be signatures of even smaller CSZ thrusts (e.g., at Bradley Lake/site 30 and at Saanich Inlet/site 5 in Figure 7, although the latter observations are attributed to a local crustal event and denoted as such [*Blais-Stevens et al*., 2011]). Inference of a connection between these and the cluster would then lead to unanswered questions about why the much larger and closer U event did not trigger the cluster earlier instead, noting that the 100 years between represents a tiny fraction of the crustal fault recurrence intervals of thousands of years and thus their proximity to failure and ripeness for triggering.

Finally, although the evidence suggests a CSZ thrust event did not trigger the Puget Sound region crustal cluster, it does suggest that thrust events may have triggered isolated crustal earthquakes in the region (lighter yellow bands in Figure 7). Figure 7 shows that several crustal earthquakes younger than the cluster may correlate temporally with CSZ thrust event W, including the youngest earthquake on the Boulder Creek fault (site 7), a poorly dated earthquake on the Seattle fault zone (site 12), and several ground failure observations attributed to crustal events at widely-separated locations (sites 5 and 11). The U event occurred within an age range consistent with triggered earthquakes on the Seattle fault zone and Birch Bay fault (site 6 in Figure 7).

The rates shown in Figure 9a are consistent with CSZ thrust events triggering individual or a few crustal earthquakes for a longer, 3500 year interval than in Figure 7, but do not require this. Figure 9a shows that two of the seven date ranges of CSZ thrust earthquakes coincide with a peak in the Puget Sound region crustal rate since 3550 cal years B.P., including the 1700 CSZ earthquake within one standard deviation of the estimated date of a crustal earthquake on the Darrington-Devils Mountain fault zone (Figure 8) (site 9 in Figure 7; the Utsalady Point strand, [*Johnson et al*., 2004b]). However, the rates do not require triggering, given that the crustal rate peaks span ~30% of the total interval (assuming each has a width of ~200 years) so that two of the seven CSZ events should occur near a crustal rate peak simply by chance, assuming uniform probabilities. This would likely be true even assuming smaller peak widths, because the dates of all but one of the great CSZ earthquakes are known only within ranges of 50–100 years.

In summary, although the 1200–900 cal years B.P. Puget Sound region cluster likely was not triggered by the likely largest, and potentially full-margin CSZ thrust events, the data permit the possibility that these and/or smaller such events have triggered individual crustal earthquakes. If so, Cascadia would be consistent with the global study of *Gomberg and Sherrod* [2014].

5. Comparing the Paleoseismic and Modern Cascadia Records

We assess the completeness of the paleoseismic record by comparing it to the regional shortening rate and the modern earthquake record in Cascadia. Noting that the rates of strain accumulation and release may differ significantly over multiple earthquake cycles [*Weldon et al*., 2004; *Dolan et al*., 2007], knowledge of the shortening rate does not strongly constrain the recurrence rates on individual faults, although the range of ~3 mm/yr to 4.4 mm/yr (see section 2) is at least consistent with the paleoseismic record. The most well-studied slip history exists for the Seattle fault zone, which contains multiple strands and a main thrust cut by opposite-dipping backthrusts [see *Mace and Keranen* [2012] for a summary of prevailing models). The postglacial slip rate for this fault zone varies temporally; e.g., for the Toe Jam backthrust from 0.2 to 0.3 mm/yr over the entire postglacial period and 0.8–3.4 mm/yr for the last 3000 years [*Nelson et al*., 2003a; *Kelsey et al*., 2008], and for the entire Seattle fault zone 0.7–1.1 mm/yr (postglacial) [*Johnson et al*., 1999]. The latter rate exceeds that obtained assuming shortening was accommodated uniformly across most of the Puget Sound region faults (e.g., a uniformly distributed average shortening rate would be approximately 0.4 mm/yr). Thus, in addition to varying temporally, we infer that shortening also may not be uniformly accommodated spatially, that most of the other Puget Sound region faults have lower slip rates than the Seattle fault and may be consistent with recurrence intervals for $M > \sim 6.5$ earthquakes in excess of 16,000 years (Figure 8). At slip rates of just a few tenths of a mm/yr over 16,000 years, a few meters of slip would accrue, just enough to produce a $6.0 < \sim M < 6.5$ earthquake assuming standard scaling relations [*Wells and Coppersmith*, 1994; *Blaser et al*., 2010].

The historic and recent crustal earthquake rates also are consistent with the recurrence intervals inferred from the paleoseismic record. Since instrumental monitoring began ~42 years ago, the largest cataloged earthquakes are three $5.0 < M < 5.5$ crustal earthquakes in the inset map area of Figure 1. The historic record includes one $6.0 < M < 6.5$ event in the last 150 years and none with $M > 6.5$ (http://old.pnsn.org/HIST_CAT/PUGET/). (The historic catalog dates back to 1872 and contains three $M > 6.0$ potentially crustal earthquakes in the Puget Sound region. However, two of these having depths estimated at 31 km and lacked aftershocks, indicating they were probably intraslab events (C. Weaver, personal communication, 2012) These two records are roughly consistent with a Gutenberg-Richter scaling; e.g., assuming the instrumentally-derived rate has been constant approximately 10 ~ *M*5.25 earthquakes should have occurred in 150 years and 1 ~ M6.25, the latter number observed in the historic record. These also are consistent with the paleoseismic record, noting that extrapolating to a duration of ~15,000 years would predict ~30 *M*6.75 earthquakes, roughly comparable to the number in the paleoseismic catalog.

6. Discussion and Conclusions

The modern record of earthquakes globally, in subduction zones other than Cascadia, suggests that the next $M > 8.6$ CSZ thrust earthquake will trigger $M > 5.5$ crustal earthquakes on faults in the overriding plate [*Gomberg and Sherrod*, 2014]. We have shown that a cluster of $M >$ ~6.5 crustal earthquakes during 1200–900 cal years B.P. in Puget Sound region was real and extraordinary, at least in this region and over the last 16,000 years. However, although we have shown that the cluster probably was not triggered by a great CSZ thrust earthquake, a causal connection between Puget Sound region crustal faults seems inescapable. Although the historic and instrumental records of Puget Sound region crustal earthquakes in the overriding plate contain no $M > 6.5$ events, clustering like that in the paleoseismic record for the Puget Sound region may be common elsewhere in Cascadia (e.g., crustal M_s6.9 and M_s7.3 earthquakes occurred just beneath Vancouver Island in 1918 and 1946, respectively [*Cassidy et al.*, 1988]). Globally, nearly synchronous (geologically) crustal earthquakes are not unique to the Puget Sound region. Clusters of earthquakes separated by a few hundred years or less, punctuated by lulls lasting several thousands of years, are documented on the Velino–Magnola normal fault in Italy [*Schlagenhauf et al.*, 2011] and the Levant transform fault system in Lebanon [*Daeron et al.*, 2007], over the last ~15,000 and ~12,000 years, respectively. A similar "flurry of four events" within a 300 year interval and periods of sluggish activity lasting many hundreds of years have been documented on one segment of the San Andreas fault near Wrightwood, California [*Weldon et al.*, 2004]. Clustering has also been documented on networks of faults by *Dolan et al.* [2007], who noted that fault systems in the Los Angeles region of California and the eastern California shear zone both exhibit temporal clustering over a 12,000 year period, with anticorrelated periods of activity on each system.

One model proposed to explain clustering appeals to theories of synchronization, or coupled oscillators, applied to other natural phenomena [*Scholz*, 2010]. This "phase-locking" requires slip rates that differ by less than 30%, becomes more likely for near-parallel or conjugate faults [*Scholz*, 2010], and has been applied to explain repeated clusters of moderate to large earthquakes in other regions globally [*Scholz*, 2010]. As noted above, the slip rates for most Puget Sound region faults are highly uncertain or unknown [*Witter et al.*, 2008; *ten Brink et al.*, 2002; *Johnson et al.*, 2004a; *Sherrod et al.*, 2004], and at least on the Seattle fault system may vary temporally [*Nelson et al.*, 2003a]. Of the six faults that may have contributed to the 1200–900 cal years B.P. cluster (each >5% of the total rate), two (the Seattle main thrust and Boulder Creek faults) have had previous earthquakes within the 16,000 year record apparently without any on the other four. Thus, the relevance of this type of synchronization model to this study remains unclear. This model is consistent with our conclusion that the Puget Sound region cluster was independent of CSZ thrust earthquakes, because together, the two systems do not satisfy the model requirements for synchronization.

While we focused on triggering of a cluster of paleo-earthquakes in the Puget Sound region's crust by CSZ thrust events, a variety of other manifestations of causal connections between faulting environments may exist. For example, *Perez and Scholz* [1997] examined $M > 7$ earthquakes with depths <70 km and suggest that $M > 7$ clusters occur more often in the decades just prior to $M > 8.5$ subduction thrust earthquakes (they do not distinguish plate boundary from intraplate events). However, while the Puget Sound region 1200–900 cal years B.P. cluster may have preceded the temporally nearest largest subduction thrust by decades, the lag most probably was longer. The modern record in Cascadia also suggests that subduction thrust earthquakes may trigger earthquakes within the subducting plate; e.g., the 1992 *M*7.1 Petrolia subduction thrust triggered two *M*6.6 earthquakes within the subducting Gorda Plate [*Oppenheimer et al.*, 1993]. (*McCrory et al*. [2012] suggest the *M*7.1 event may have occurred above the plate interface, but still with a thrust mechanism consistent with plate convergence.) Finally, *Valentine et al*. [2012] found evidence in stratigraphic data for at least two local *M*6.5–7.2 crustal earthquakes in southern Cascadia in the last 2000 years and curiously, although they do not suggest a causal connection, they infer that one of these crustal events (1300–1050 cal years B.P.) was either coincident or actually preceded a $M > 8.2$ subduction thrust earthquake (1450–1200 cal years B.P.).

Appendix A: Sites Used in Figure 7

1. Catala Lake—*Clague et al*. [1999];
2. Deserted Lake—*Hutchinson et al*. [2000];
3. Kanim Lake—*Hutchinson et al*. [1997];
4. Port Alberni and Tofino—*Clague and Bobrowsky* [1994];
5. Saanich Inlet—*Blais-Stevens et al*. [2011]
6. Birch Bay—*Kelsey et al*. [2012]
7. Boulder Creek fault—*Sherrod et al*. [2013]
8. Swantown—*Williams and Hutchinson* [2000]
9. Utsalady Point fault—*Johnson et al*. (2004)
10. Discovery Bay—*Williams et al*. [2005];
11. Snohomish River—*Bourgeois and Johnson* [2001];
12. Seattle fault—*Atwater and Moore* [1992], *Bucknam et al*. [1992], *Kelsey et al*. [2008], *Nelson et al*. [2003a], and *Sherrod* [2000];
13. Lake Washington—*Karlin et al*. [2004]
14. Saddle Mountain fault—*Wilson et al*. [1979], *Witter et al*. [2008], and E. A. Barnett et al. (unpublished manuscript, 2014)
15. Tacoma fault—*Bucknam et al*. [1992] and *Sherrod et al*. [2004];
16. Olympia fault—*Clement et al*. [2010] and *Sherrod* [2001];
17. SW Washington/NW Oregon—*Atwater et al*. [2004b, 2004c], *Atwater and Hemphill-Haley* [1997], and *Shennan et al*. [1996];
18. Necanicum River—*Darienzo et al*. [1994];
19. Netarts Bay— *Darienzo and Peterson* [1990] and *Shennan et al*. [1998];
20. Nestucca Bay—*Darienzo et al*. [1994];
21. Salmon River—*Nelson et al*. [2004];
22. Siletz Bay—*Darienzo et al*. [1994] and *Peterson et al*. [1996];
23. Yaquina Bay—*Darienzo et al*. [1994];
24. Alsea Bay—*Peterson and Darienzo* [1991] and *Nelson et al*. [2008b];
25. Siuslaw River—*Nelson* [1992a] and *Briggs* [1994];
26. Umpqua River—*Briggs* [1994];
27. N. Coos Bay—*Nelson* [1992a] and *Briggs* [1994];
28. S. Coos Bay—*Briggs* [1994] and *Nelson et al*. [1996, 1998];
29. Coquille River—*Nelson* [1992a, 1992b] and *Witter et al*. [1997, 2003];
30. Bradley Lake—*Kelsey et al*. [2005];
31. Sixes River—*Kelsey et al*. [1998, 2002];
32. Lagoon Cr.—*Garrison-Laney et al*. [2006];
33. Humboldt Bay—*Vick* [1988], *Carver et al*. [1992], *Clarke and Carver* [1992], *Valentine et al*. [2012], *Pritchard* [2004], and *Patton and Witter* [2006];
34. Eel River—*Li* [1992];
35. Ef. Effingham Inlet—*Enkin et al*. [2013].

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