Land-level changes from a late Holocene earthquake in the northern Puget Lowland, Washington

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ABSTRACT

An earthquake, probably generated on the southern Whidbey Island fault zone, caused 1–2 m of ground-surface uplift on central Whidbey Island ~2800–3200 yr ago. The cause of the uplift is a fold that grew coseismically above a blind fault that was the earthquake source. Both the fault and the fold at the fault's tip are imaged on multichannel seismic refection profiles in Puget Sound immediately east of the central Whidbey Island site. Uplift is documented through contrasting histories of relative sea level at two coastal marshes on either side of the fault. Late Holocene shallow-crustal earthquakes of M_w = 6.5–7 pose substantial seismic hazard to the northern Puget Lowland.

Keywords: paleoseismology, relative sea level, blind faults, seismic reflection, Puget Sound.

INTRODUCTION

Late Holocene earthquakes accompanied by ground-surface displacement indicate a significant seismic hazard in the central and southern Puget Lowland of Washington. Earthquakes with surface rupture or regional coseismic uplift, associated with the Seattle and Tacoma fault zones (Fig. 1), have affected the Seattle and Tacoma urban region approximately three or four times in the late Holocene (Johnson et al., 1999; Brocher et al., 2001; Blakely et al., 2002; Nelson et al., 2003).

In contrast, less is known of the late Holocene activity of faults in the northern Puget Lowland, a region that has not been subject to damaging earthquakes in historic time (Ludwin et al., 1991). Northern Puget Lowland faults include the southern Whidbey Island, Utsalady Point, Strawberry Point, and Devils Mountain fault zones (Johnson et al., 1996, 2001a) (Fig. 1). The southern Whidbey Island fault zone, although lacking a Holocene surface trace on Whidbey Island, is a major basin-bounding structure that has a pronounced gravity and aeromagnetic signature and deforms Quaternary sediment visible on offshore seismic reflection profiles (Johnson et al., 1996, 2001b; Blakely and Lowe, 2001). The Utsalady Point fault in northern Whidbey Island has late Holocene fault scarps, and trenching investigations indicate one or two earthquakes in the late Holocene (Johnson et al., 2003). In this paper we provide data that suggest that an earthquake occurred ~3000 yr ago on a buried fault beneath central Whidbey Island, resulting in 1 m or more of vertical, north-side-up displacement. The fault is likely part of the southern Whidbey Island fault zone (Fig. 1). Taken together, these studies indicate that the northern Puget Lowland was host to several shallow-crustal, surface-deforming magnitude 6.5–7 earthquakes in the past 3000 yr.

LATE QUATERNARY TECTONIC DEFORMATION ON CENTRAL WHIDBEY ISLAND

Multichannel seismic reflection data in Holmes Harbor immediately east of central Whidbey Island (Fig. 2) prompted us to investigate the possibility of late Holocene fault-related ground-surface deformation. Identification of the Pleistocene section is derived from seismic facies analysis, projection from nearby boreholes (e.g., Standard Engstrom, Fig. 2), and iterative correlation and mapping from a large suite of U.S. Geological Survey and industry seismic reflection profiles (Johnson et al., 1996, 2001a, 2001b; Rau and Johnson, 1999). The seismic data show a prominent asymmetric syncline, steeper to the north, developed in Pleistocene sediment. Holocene strata at the top of the section also appear to dip to the south. In the Pleistocene section, the strata thicken in the fold axis and thin on the limbs, suggesting that the fold has been growing in the Quaternary. Along the profile of the section, the modern bathymetry mimics the underlying structure; a bathymetric high in Holmes Harbor overlies the uplifted northern fold limb. We infer that the syncline is cut by a steep, north-side-up fault on the basis of truncation of seismic reflections and inferred offset of the base of the Quaternary section.

The deformed Pleistocene marine section is ~2–4 km southeast along regional strike from the narrow central neck of Whidbey Island, where both island topography and geology show a geomorphic and structural setting similar to that in the offshore bathymetry and seismic stratigraphy. Light detection and ranging (lidar) imagery from central Whidbey Island shows a pronounced lowland in which Hancock marsh is at the lowest point, situated on a northeast projection of the syncline (Fig. 2). Similarly, the bathymetric high in Holmes Harbor (between shotpoints 700 and 800, Fig. 2) projects to the upland that rises abruptly just to the north of Hancock marsh. The Hancock marsh lowland is bordered by steep (10 percent) slopes to the north and gradual slopes to the south (Fig. 2), similar to the asymmetry of the folded Pleistocene sediment offshore to the southeast.

Although the Pleistocene section appears to be faulted, no late Holocene fault scarp is evident on the lidar image in the area where a fault would propagate to the surface (Fig. 2). We infer therefore that if a fault was active in the late Holocene, surface deformation would be folding above a blind fault tip.

RESEARCH APPROACH

If a fold in the study area above a fault tip has been active in the late Holocene, raising the upland to the north relative to the Hancock lowland, then relative sea levels at Hancock marsh should be different from those recorded in a coastal marsh north of the projection of the fault at the surface. The Crockett marsh wetland is ~1 km north of the projected fault, and the Hancock marsh wetland is ~1 km south of the projected fault (Figs. 1B and 2). Barring any Holocene fault displacement between the wetlands, they should have identical relative sea-level histories because they are close (8 km) to each other. Any glacio-isostatic influence on relative sea level should be reflected identically in both wetlands, and any tectonic influence on relative sea-level resulting from strain accumulation and release on the underlying subduction zone similarly should result in identical relative sea-level perturbations at both sites. Therefore, different relative sea-level histories implicate late Holocene tectonic displacement caused by folding above an active fault (Fig. 2).

If relative sea-level curves for the two marshes are not the same, then the divergence of the two curves would indicate the timing

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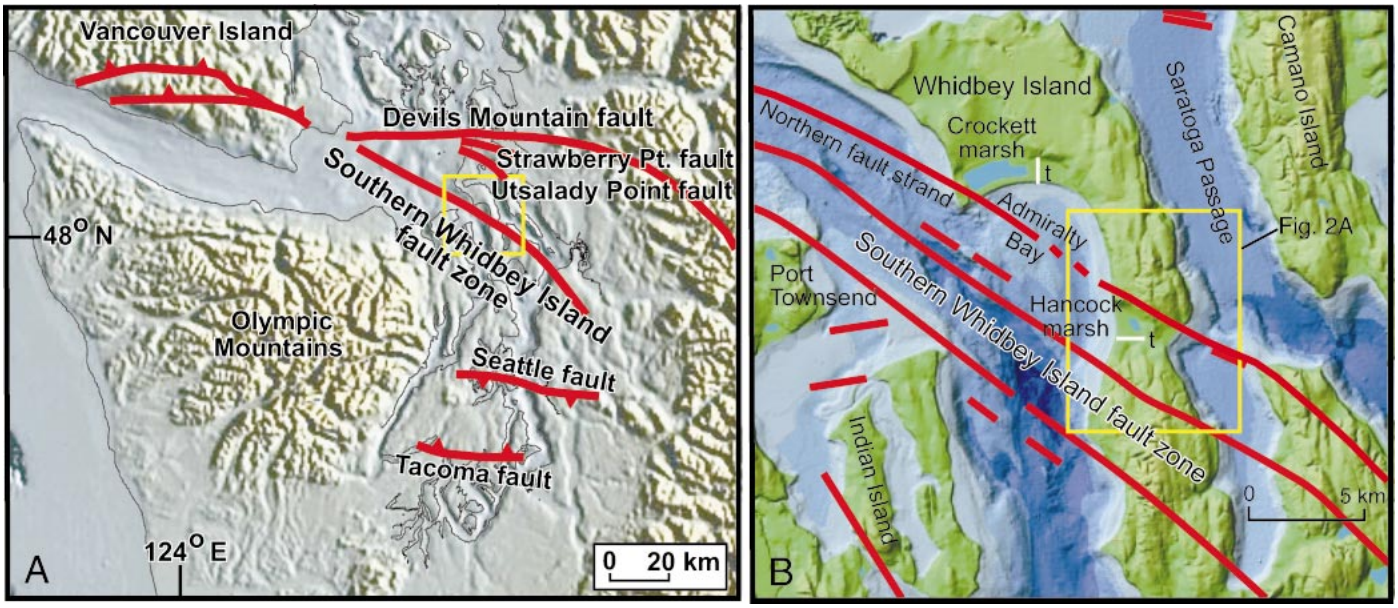

Figure 1. A: Faults in Puget Lowland of northwest Washington and southern Vancouver Island, modified after Johnson et al. (1996, 2001b). B: Three main fault traces of southern Whidbey Island fault zone, adapted from Johnson et al. (1996, 2001b). Crockett marsh and Hancock marsh straddle trace of northern fault strand. In area of incomplete seismic line coverage in Admiralty Bay, gap is shown in northern strand. U.S. Geological Survey (USGS) 1995 seismic reflection line shows fault in Admiralty Bay between Crockett and Hancock marshes (Johnson et al., 2001b). Earlier industry seismic reflection line east of USGS line does not show fault in Admiralty Bay (Johnson et al., 1996); however, industry line may not have extended far enough north to cross fault. Yellow rectangle represents area of Figure 2A. Two short white lines (labeled t) show locations of two transects in Figure 3.

and magnitude of the vertical displacement caused by folding. In order to determine sea-level histories at the two marshes, cores were taken in a transect across each wetland perpendicular to the barrier sand bar that separates the Puget Sound (Admiralty Bay) from the wetland. Each of the two core transects starts at the sand bar and proceeds inland across the wetland; cores were taken every 50 m until the dry upland was reached.

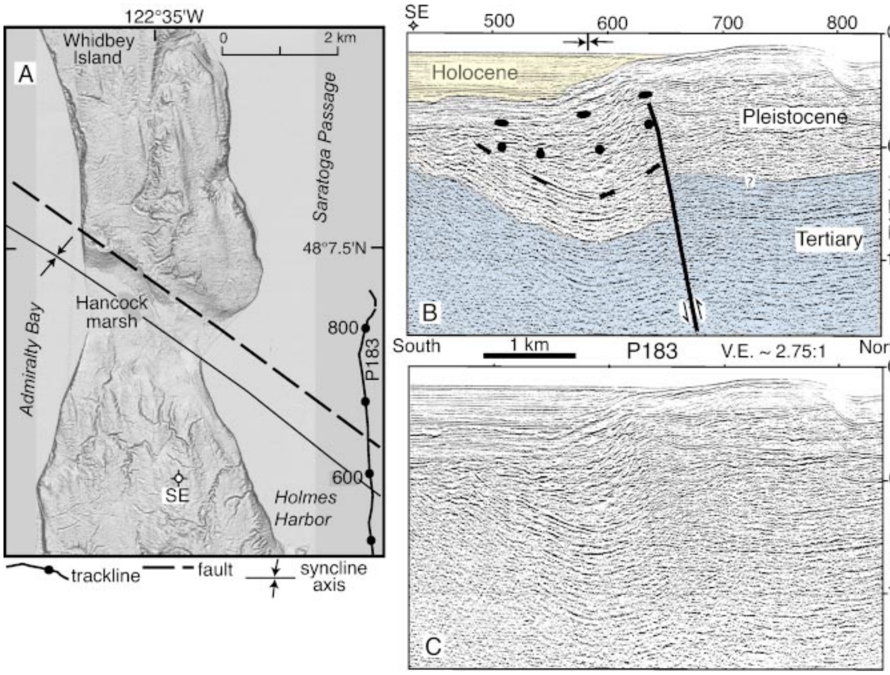

Figure 2. A: Light detection and ranging (lidar) image of central Whidbey Island, location of offshore seismic line number P183 (irregular line in lower right with solid dots representing every 100 shotpoints), and projection along regional strike of synclinal fold and fault (dashed line) from seismic line. SE—Standard Engstrom borehole (Rau and Johnson, 1999). B: Multichannel seismic reflection profile in Holmes Harbor; data collected in 1995 with processing described in Johnson et al. (2001a). TWT—two-way traveltime. Three sets of distinctive solid black shapes track three different reflectors along trend of profile. C: Same as B except uninterpreted seismic profile.

GEOLOGIC CROSS SECTIONS AND RELATIVE SEA-LEVEL CURVES

At each marsh, the geologic cross section (Fig. 3) indicates relative sea-level rise over time because the sand barrier seaward of each marsh has built upward during the late Holocene. As the barriers built upward, peat wetlands aggraded behind them (Fig. 3).

Paleo–sea-level index points define a relative sea-level curve at each site. Sea-level index points are at the contact of the peat with the underlying substrate (Fig. 3), which is beach sand, sandy gravel, or a thin veneer of beach sand over Pleistocene glacial sediment. The sand-peat contact is a paleo–sea-level indicator because the underlying sand has marine diatoms and the overlying peat has freshwater and brackish-water plant macrofossils. The elevation of the sand-peat contact in the modern marsh, based on level surveys, is the same at each marsh within 0.2 m (see Fig. DR1[1]) and is at the approximate level of mean high water (MHW) (Fig. 3).

The sea-level curves for the two marshes (Fig. 4A) are based on the age and depth of paleo–sea-level index points relative to mod-

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[1] GSA Data Repository item 2004081, Figure DR1, survey data, and Table DR1, radiocarbon ages, is available online at www.geosociety.org/pubs/ft2004.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

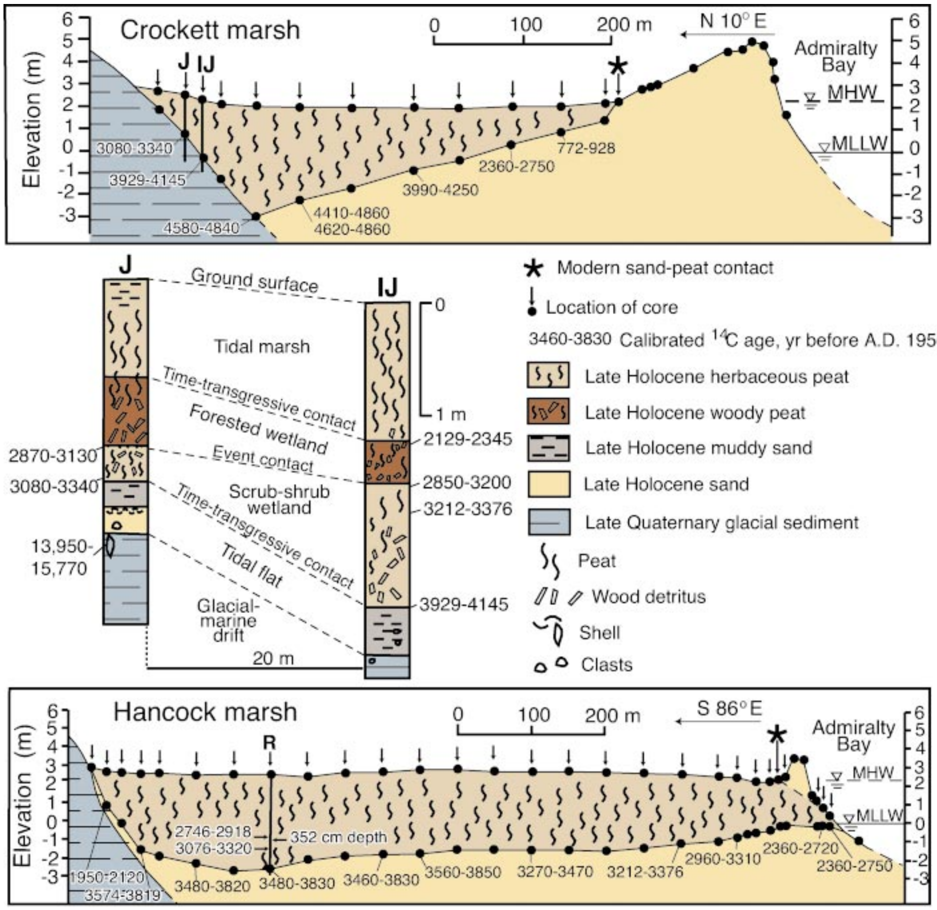

Figure 3. Cross sections of Crockett and Hancock marshes and stratigraphic data from two cores at Crockett marsh. Hancock marsh differs from Crockett marsh in that in Hancock traverse, herbaceous peat is exposed in intertidal zone seaward of sand-barrier berm. At Hancock, sand-barrier berm is underlain by peat because sand barrier first built upward and prograded seaward; then, while sand barrier continued to build upward, barrier started to retreat landward, eroding peat that had been previously deposited behind it. At Crockett, sand barrier built upward and seaward as relative sea level rose. MHW—mean high water; MLLW—mean lower low water.

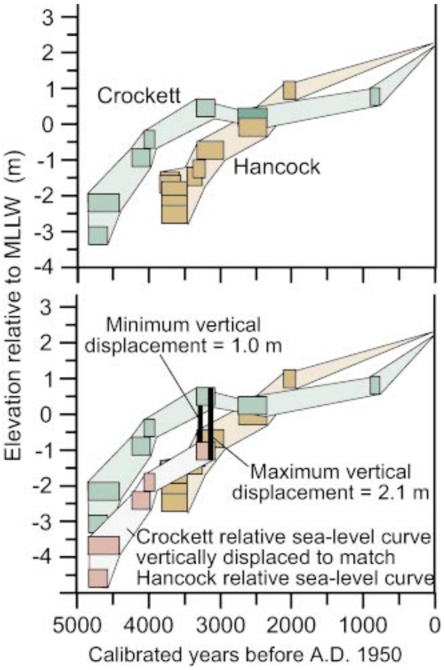

Figure 4. Top: Relative sea-level curves for Crockett and Hancock marshes. Bottom: Combined relative sea-level curves for Crockett and Hancock marshes, showing reconstruction of single sea-level curve (Hancock curve with superposed, displaced Crockett curve) if there had been no vertical displacement of Crockett vs. Hancock marshes. MLLW—mean lower low water.

ern mean lower low water (MLLW). Accelerator mass spectrometry ^{14}C dating of plant macrofossils (mainly seeds: Table DR1; see footnote 1) in the peat immediately above the contact provides age control. Age ranges in Figures 3 and 4 are years before A.D. 1950 and were converted from the laboratory-reported 2σ ^{14}C age ranges by using the method of Stuiver et al. (1998) (Table DR1; see footnote 1). The height and width of the rectangles (data points, Fig. 4) express the magnitude of the error in locating the relative sea-level curve in time versus depth space. The width of the rectangle is the age range of the sand-peat contact (Fig. 3). The height of the rectangle is the uncertainty in elevation of the sand-peat contact, ±0.3 m, calculated as the square root of the sum of the squares of the three variables that determine elevation relative to MLLW: survey error (±0.01 m), tidal measurement at site (±0.20 m), and variation in elevation of the modern sand-peat contact (±0.20 m).

NORTH-SIDE-UP DISPLACEMENT 3000 YR AGO

The two relative sea-level curves are not the same. Superposition of the Crockett and Hancock curves (Fig. 4B) shows that, for ages older than ~3200 yr, equivalent-age sea-level index points at Crockett are 1–2 m higher than at Hancock. Also, there is an ~700 yr interval of time, after 3200 yr, when the Crockett site shows no change in relative sea level, or a relative sea-level fall, whereas there is no period of time at Hancock when relative sea level is not rising. The two relative sea-level curves in the past 2500 yr are similar within limits of sea-level resolution.

The apparent upward shift of the Crockett relative sea-level curves is caused by upward vertical displacement at Crockett relative to Hancock ~3200–2800 yr ago (Fig. 4). Reconstructing one relative sea-level curve common to both sites requires vertically lowering the Crockett curve onto the Hancock curve by 1.0–2.1 m (Fig. 4), which implies that the amount of vertical displacement is 1.5 m ± 0.5 m up to north.

Paleoecologic data from cores at the two marshes provide corroborative evidence for upward displacement at the Crockett marsh relative to the Hancock marsh ~3000 yr ago. At the Crockett marsh, a 2-cm-thick, clean, fine sand layer occurs within the peat sequence in core J ~25 cm above the sand-peat contact; the sand layer contained 2870–3130-yr-old detrital conifer leaves (Fig. 3). The thin sand layer separates an underlying herbaceous and detrital peat from an overlying peat containing woody detritus from a forested wetland. Similarly, in adjacent core IJ, 25 m farther north, a contact of the same age as the sand layer separates an underlying herbaceous peat from an overlying peat containing coarse woody detritus (Fig. 3). At the time of deposition of the thin sand layer, the wetland at Crockett emerged suddenly, changing from a scrub-shrub wetland to a relatively drier forested wetland at core site J and from a peat wetland to a forested wetland at core site IJ (Fig. 3). Although not convincing because of limited core observations, the thin sand may have been deposited by a tsunami generated by seafloor displacement offshore in the Puget Sound; the same displacement would have produced the rapid emergence, and a change to drier site conditions, observed at the Crockett wetland relative to the Hancock wetland.

The drier environment at Crockett persisted for ~700–800 yr, on the basis of ^{14}C ages, at which time the Crockett marsh again responded to relative sea-level rise through peat aggradation. This ~750 yr interval is roughly equivalent to the time interval when relative sea level at Crockett was static or fell slightly (Fig. 4).

Hancock marsh shows a profound change in the diatom assemblages deposited ~3000 yr ago (352 cm depth, core R, Fig. 3), which is synchronous, within limits of the calibrated radiocarbon ages, with the time of abrupt relative sea-level fall at Crockett marsh. Prior to 3000 yr ago, the Hancock diatom assemblage was dominated by the intertidal diatom flora *Paralia sulcata* and *Trachyneis aspera*. At ~3000 yr ago, the assemblage became dominated by *Fragilaria construens*, a cosmopolitan diatom flora that characteristically pioneers disturbed wetland sites. The change to *Fragilaria construens* does not require a change in relative sea level, but it does reflect a sustained change in site conditions, such as a change in vascular flora or slightly reduced salinity (Krammer and Lange-Bertalot, 1991).

DISCUSSION

The best explanation for the dissimilarity of relative sea-level curves is vertical tectonic displacement on a blind fault between the two wetlands, a fault inferred to be part of the southern Whidbey Island fault zone. The surface deformation is a fold above the buried fault tip; the seismic reflection profile 2–4 km to the southeast along regional strike images this fold (Fig. 2). The buried fault is situated within the approximately located northern strand of the southern Whidbey Island fault zone (Figs. 1 and 2). Because the fault is buried, the fault trend, dip, and sense of slip are unclear. We infer from the asymmetry of the folded section (Fig. 2) that the fault dips steeply to the north. If the fault dips 60° and accommodated 1–2 m of vertical displacement, then the net slip on the fault was a minimum of ~1.2–2.3 m. Net slip could have exceeded 2.3 m if there was a component of strike slip. Empirical scaling relationships of vertical displacement to M_w for historic strike-slip and reverse faults (Wells and Coppersmith, 1994) suggest an earthquake of $M_w = 6.5–7.0$.

CONCLUSION

About 2800–3200 yr ago, a shallow-crustal earthquake of probable $M_w = 6.5–7$ deformed the ground surface of central Whidbey Island in the northern Puget Lowland. Multichannel seismic reflection profiles in Holmes Harbor east of central Whidbey Island show folded Pleistocene, and possibly Holocene, strata. Folding appears to be at the tip of a high-angle fault that displaces the base of Quaternary sediment. The axis of the syncline, projected 4 km northwest along regional strike to central Whidbey Island, underlies the Hancock marsh coastal wetland. On the basis of relative sea-level analysis of Hancock marsh in comparison to Crockett marsh, a coastal wetland ~8 km northwest across the projected trace of the buried fault, Crockett marsh underwent a 1–2 m abrupt uplift relative to sea level 2800–3200 yr ago at a time of unchanging relative sea level at Hancock marsh. Because there is no late Holocene fault scarp between the two marshes, we infer that the uplift is a result of coseismic reactivation of the buried fault and overlying fold evident in the reflection profiles. The fault beneath Holmes Harbor is within the mapped area of the southern Whidbey Island fault zone; this fault zone is the likely source of the earthquake 2800–3200 yr ago. Our research demonstrates the utility of relative sea-level investigations to identify active blind faults.

The earthquake ~3000 yr ago in central Whidbey Island and the one to two late Holocene earthquakes related to the Utsalady fault on northern Whidbey Island indicate that the northern Puget Lowland is vulnerable to surface-deforming, shallow earthquakes accompanied by strong ground motions. Such earthquakes pose significant hazard to the 85-km-long northern Washington urbanized coastal corridor extending from Everett, Washington, north to the Canadian border.

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REFERENCES CITED

Blakely, R.J., and Lowe, C., 2001, Aeromagnetic anomalies of the eastern Juan de Fuca strait region, *in* Mosher, D.C., and Johnson, S.Y., eds., Neotectonics of the eastern Strait of Juan de Fuca: A digital geological and geophysical atlas: Geological Society of Canada Open-File 3931, CD-ROM.

Blakely, R.J., Wells, R.E., Weaver, C.S., and Johnson, S.Y., 2002, Location, structure and seismicity of the Seattle fault zone, Washington: Evidence from aeromagnetic anomalies, geologic mapping and seismic reflection data: Geological Society of America Bulletin, v. 114, p. 167–177.

Brocher, T.M., Parsons, T., Blakely, R.J., Christensen, N.I., Fisher, M.A., Wells, R.E., and the SHIPS Working Group, 2001, Upper crustal structure in Puget Lowland, Washington: Results from the 1998 seismic hazard investigation in Puget Sound: Journal of Geophysical Research, v. 106, p. 13,541–13,564.

Johnson, S.Y., Potter, C.J., Armentrout, J.M., Miller, J.J., Finn, C., and Weaver, C.S., 1996, The southern Whidbey Island fault: An active structure in the Puget Lowland, Washington: Geological Society of America Bulletin, v. 108, p. 334–354.

Johnson, S.Y., Dadisman, S.V., Childs, J.R., and Stanley, W.D., 1999, Active tectonics of the Seattle fault and central Puget Sound, Washington—Implications for earthquake hazards: Geological Society of America Bulletin, v. 111, p. 1042–1053.

Johnson, S.Y., Dadisman, S.V., Mosher, D.C., Blakely, R.J., and Childs, J.R., 2001a, Active tectonics of the Devils Mountain fault and related structures, northern Puget Lowland and eastern Strait of Juan de Fuca region, Pacific northwest: U.S. Geological Survey Professional Paper 1643, 45 p.

Johnson, S.Y., Mosher, D.C., Dadisman, S.V., Childs, J.R., and Rhea, S.B., 2001b, Tertiary and Quaternary structures of the eastern Juan de Fuca Strait: Point map, *in* Mosher, D.C., and Johnson, S.Y., eds., Neotectonics of the eastern Strait of Juan de Fuca: A digital geological and geophysical atlas: Geological Society of Canada Open File 3931, CD-ROM.

Johnson, S.Y., Nelson, A.R., Personius, S.F., Wells, R.E., Kelsey, H.M., Sherrod, B.L., Okumura, K., Koehler, R., Witter, R.C., and Bradley, L., 2003, Evidence for one or two late Holocene earthquakes on the Utsalady Point fault, northern Puget Lowland, Washington: Geological Society of America Abstracts with Programs, v. 34, no. 7, p. 479.

Krammer, K., and Lange-Bertalot, H., 1991, Bacillariophyceae: 3. Teil: Centrales, Fragilariaceae, Eunotiaceae, *in* Ettl, H., et al., eds., Swasserflora von Mitteleuropa, Volume 2/3: Stuttgart, Jena, Gustav Fischer Verlag, 576 p.

Ludwin, R.S., Weaver, C.S., and Crosson, R.S., 1991, Seismicity of Washington and Oregon, *in* Slemmons, D.B., et al., eds., Neotectonics of North America: Boulder, Colorado, Geological Society of America Decade of North American Geology Decade Map Volume, p. 77–98.

Nelson, A.R., Johnson, S.Y., Kelsey, H.M., Wells, R.E., Sherrod, B.L., Pezzopane, S.K., Bradley, L., Koehler, R.D., and Bucknam, R.C., 2003, Late Holocene earthquakes on the Toe Jam Hill fault, Seattle fault zone, Bainbridge Island, Washington: Geological Society of America Bulletin, v. 115, p. 1388–1403.

Rau, W.W., and Johnson, S.Y., 1999, Well stratigraphy and correlations, western Washington and northwest Oregon: U.S. Geological Survey Map I-2621, 3 sheets, 31 p.

Stuiver, M., Reimer, P.J., Bard, E., Beck, J.W., Burr, G.S., Hughen, K.A., Kromer, B., McCormac, G., Plicht, J.V., and Spurk, M., 1998, INTCAL98 radiocarbon age calibration, 24,000–0 cal BP: Radiocarbon, v. 40, p. 1041–1084.

Wells, D.L., and Coppersmith, K.J., 1994, New empirical relations among magnitude, rupture length, rupture width, rupture area and surface displacement: Seismological Society of America Bulletin, v. 84, p. 974–1002.

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