A new view into the Cascadia subduction zone and volcanic arc: Implications for earthquake hazards along the Washington margin

Tom Parsons U.S. Geological Survey, M.S. 999, 345 Middlefield Road, Menlo Park, California 94025
Anne M. Trehu College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, Oregon 97331
James H. Luetgert U.S. Geological Survey, M.S. 977, 345 Middlefield Road, Menlo Park, California 94025
Kate Miller
Fiona Kilbride Department of Geological Sciences, University of Texas at El Paso, El Paso, Texas 79968
Ray E. Wells U.S. Geological Survey, M.S. 975, 345 Middlefield Road, Menlo Park, California 94025
Michael A. Fisher U.S. Geological Survey, M.S. 999, 345 Middlefield Road, Menlo Park, California 94025
Ernst Flueh GEOMAR Research Center for Marine Sciences, Wischhofstr. 1-3 24148 Kiel, Germany
Uri S. ten Brink U.S. Geological Survey, Quissett Campus, 384 Woods Hole Road, Woods Hole, Massachusetts 02543
Nikolas I. Christensen Department of Geology and Geophysics, University of Wisconsin, Madison, Wisconsin 53706

ABSTRACT

In light of suggestions that the Cascadia subduction margin may pose a significant seismic hazard for the highly populated Pacific Northwest region of the United States, the U.S. Geological Survey (USGS), the Research Center for Marine Geosciences (GEOMAR), and university collaborators collected and interpreted a 530-km-long wide-angle onshore-offshore seismic transect across the subduction zone and volcanic arc to study the major structures that contribute to seismogenic deformation. We observed (1) an increase in the dip of the Juan de Fuca slab from 2°–7° to 12° where it encounters a 20-km-thick block of the Siletz terrane or other accreted oceanic crust, (2) a distinct transition from Siletz crust into Cascade arc crust that coincides with the Mount St. Helens seismic zone, supporting the idea that the mafic Siletz block focuses seismic deformation at its edges, and (3) a crustal root (35–45 km deep) beneath the Cascade Range, with thinner crust (30–35 km) east of the volcanic arc beneath the Columbia Plateau flood basalt province. From the measured crustal structure and subduction geometry, we identify two zones that may concentrate future seismic activity: (1) a broad (because of the shallow dip), possibly locked part of the interplate contact that extends from ~25 km depth beneath the coastline to perhaps as far west as the deformation front ~120 km offshore and (2) a crustal zone at the eastern boundary between the Siletz terrane and the Cascade Range.

INTRODUCTION

The Juan de Fuca plate subducts beneath North America at a rate of ~40 mm/yr directed N68°E (e.g., DeMets et al., 1990). This oblique subduction has created a complex, geologically diverse, and potentially hazardous region, the Cascadia subduction zone and volcanic arc. No great Cascadia subduction zone earthquakes have been recorded in written history, and much of the region is relatively quiet seismically (e.g., Dewey et al., 1989). However, global comparisons indicate that the Cascadia subduction zone has many characteristics in common with those that produce great interplate earthquakes (e.g., Heaton and Kanamori, 1984; Heaton and Hartzell, 1987).

Recent studies of the Holocene geologic record have shown consistent indications that great subduction-zone and/or large upper-plate earthquakes have affected the Washington coastal margin. Interpretation of geologic evidence (subsidence, tsunami deposits) along the coast has suggested that great earthquakes (M ≥8) have occurred in the Cascadia subduction zone on a recurrence interval of hundreds of years (e.g., Atwater, 1996). Turbidite sequences cored off the Washington-Oregon margin revealed as many as 13 events that may have resulted from great earthquakes in the Cascadia subduction zone between ~7500 and 300 yr ago (Adams, 1990; Nelson et al., 1996). Eight buried shorelines were identified near Willapa Bay, Washington, that are thought to have resulted from instantaneous submersion during great subduction zone earthquakes between 5800 and 300 yr ago (Meyers et al., 1996). Satake et al. (1996) limited the ~300 yr old Cascadia event to January 1700, on the basis of tsunami records in Japan. Geodetic surveys have shown that the Washington coastal region is actively rising relative to sea level and accumulating strain (e.g., Adams, 1984, Savage et al., 1991), an indication that the subduction zone is locked offshore. Although the uplift of the Coast Ranges results in part from the variably buoyant subducting Juan de Fuca plate, some of the uplift appears to be caused by interseismic strain accumulation (Kelsey et al., 1994), supporting the suggestion that the subduction zone is locked offshore.

While the threat of dangerous earthquakes may loom over the Pacific Northwest, very little is known about the deep geologic structure beneath the continental margin where the Juan de Fuca slab is most likely to be locked and accumulating seismic stress (e.g., Heaton and Hartzell, 1987; Savage et al., 1991; Hyndman and Wang, 1993). To provide a model for lithospheric structure of the Cascadia subduction zone, the U.S. Geological Survey (USGS) in cooperation with collaborators from the Research Center for Marine Geosciences (GEOMAR), Oregon State University, and the University of Texas at El Paso, acquired wide-angle seismic data in southern Washington (Fig. 1). The model resulting from these data (Fig. 2) provides an interpretation of the structure in the offshore, potentially locked part of the subduction zone where few earthquakes occur, as well as an image of the crust and upper mantle of western continental Washington.

NEW SEISMIC IMAGE OF THE CASCADIA SUBDUCTION ZONE AND ARC

In 1995, wide-angle seismic data were collected on land along an east trending profile (~1500 instruments at 200 m spacing) by recording 17 large explosive sources. In 1996, the German research vessel Sonne conducted an extensive investigation of the offshore Oregon and Washington margins in a joint GEOMAR-USGS effort (Flueh et al., 1997). About 2000 offshore air-gun sources (~25 m spacing) were recorded on 35 instruments located on the sea floor and on land (~5 km spacing) along an extension of the 1995 profile. Traveltimes from refracted arrivals (crustal and upper mantle; Pg and Pn) were merged and inverted for velocity structure, and wide-angle reflected arrivals (base of oceanic and continental crust; PmP) were forward modeled following the methods outlined by Hole and Zelt (1995) and Parsons et al. (1996). The resulting velocity-structure model is shown in Figure 2. The root-mean-squared traveltime misfits for refracted arrivals were less than 0.1s, whereas misfits for all reflected arrivals were less than ±0.15s. The velocity model was used as input for a Bouguer gravity model,

Data Repository item 9822 contains additional material related to this article.

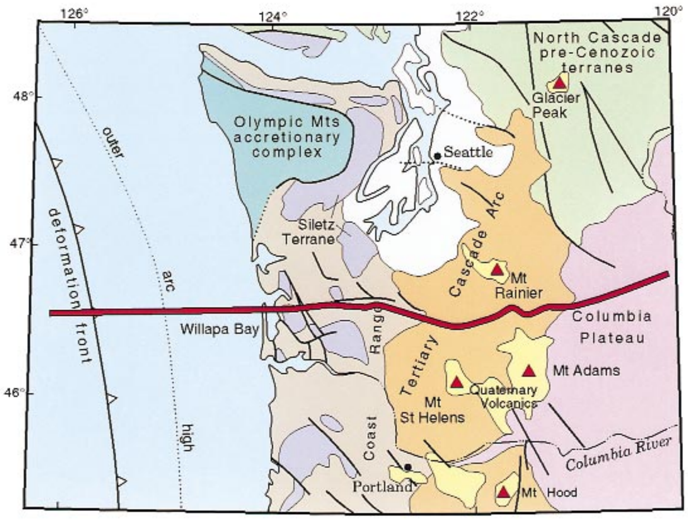

Figure 1. Location of seismic traverse in relationship to major geologic units, faults, and volcanoes in southwest Washington.

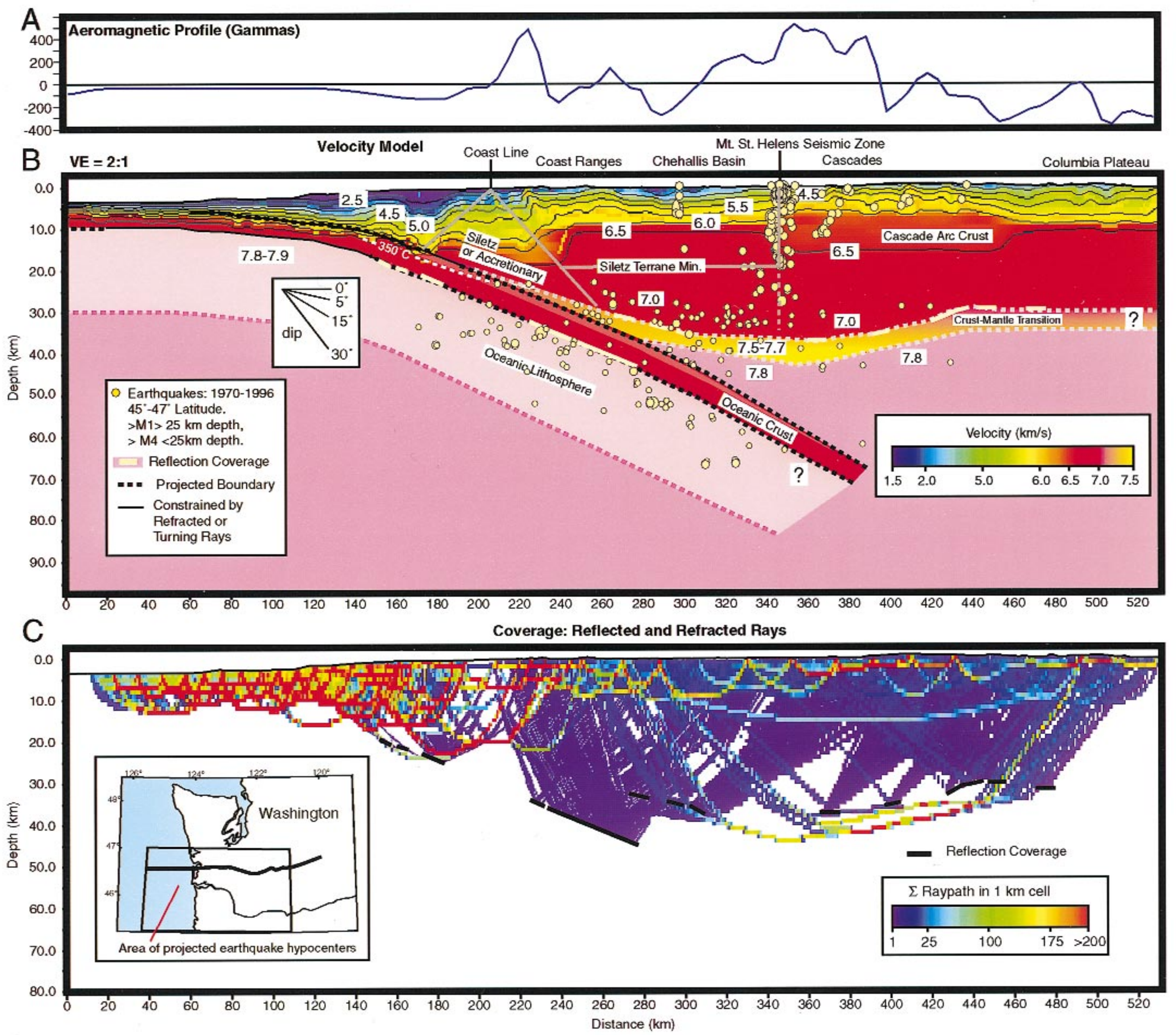

Figure 2. A: Aeromagnetic profile coincident with seismic profile. B: Seismic velocity model across Cascadia subduction zone and volcanic arc. Model features are discussed in text. C: Coverage diagram showing where model is constrained.

and primary model features fit well with the observed gravity data. Data plots with phase interpretations are archived.[1]

Juan de Fuca Slab

The descending Juan de Fuca slab was imaged by refractions and reflections from offshore air-gun blasts and reversing reflections recorded from chemical explosions on land. Seismic data coverage extends about 200 km offshore (Fig. 2). Normal-thickness (6 km) subducting oceanic crust encounters the deformation front at about km 60, where its dip increases from 0°–2° to 3°–5°. Coincident multichannel reflection data show a series of landward-verging thrusts and fluid-diapirs beginning at this point, implying low coupling stress along the interplate contact (Flueh et al., 1997). About 35 km offshore, the Juan de Fuca slab steepens from a 5°–7°dip to a 12° dip, as determined from overlapping and reversed wide-angle reflections from the base of oceanic crust recorded from land and marine sources (Fig. 2). Our observed 12° dip, slightly steeper than the 9° dip of Taber and Lewis (1986), persists to at least 50 km depth, 75 km east of the coastline (Fig. 2). Slab earthquakes are located exclusively in the mantle part of the slab, counter to the global observations for young, warm slabs of Kirby et al. (1996). Continuing the slab downward along a 12° dip places it at rather shallow depths (60–70 km) beneath the Cascade arc volcanoes. Either the slab dip steepens eastward of where it reaches the 50 km depth at which we last measure it, or it is young and warm enough to produce arc volcanism at 60–70 km depths. Seismicity appears to coincide with the Moho boundary in Fig. 2b, but this location is a result of our choice of events to plot; we restricted events to M ≥4.0 above 25 km depth, but allowed all events M ≥1.0 beneath 25 km depth to enable the best combination of highlighting zones of seismicity without obscuring the model with too many events. Smaller earthquakes are actually quite widely distributed throughout the continental crust (e.g., Stanley et al., 1996).

Accretionary Complex and Siletz Terrane

At the edge of the continental shelf (km 120), the thickness of accumulated sediment scraped from the descending plate reaches 7 km. A 3-km-thick, low-velocity (2.0–3.0 km/s) sequence fills a basin just inboard (east) of the continental slope edge. A basement high intervenes between this and a second, deeper (5 km of sub-3.0-km/s sedimentary rocks) basin at about km 170. This part of the shelf and upper slope is undergoing extension and collapse as shown by widespread normal faulting (McNeill et al., 1997).

Near the coastline lies the surface contact between accretionary rocks and the Siletz terrane, a late Paleocene-early Eocene age accreted oceanic block that forms the basement to much of the Cascadia forearc. Snavely and Wagner (1982) map this contact from well and seismic reflection data as an east-dipping thrust that partially cuts a thin Quaternary unit about 1 km offshore near Greys Harbor (~20 km north of our transect). The Siletz terrane is associated with a strong aeromagnetic signal along our profile where it outcrops; this signal tapers off abruptly at the coastline (Fig. 2A).

Projecting the Siletz-accretionary boundary to depth based on the velocity model is problematic; outcrops of the Siletz terrane (Walsh et al., 1987) (Fig. 1) correspond with the 5.0 km/s contour in our model where it approaches the surface, and independent laboratory velocity measurements from the Siletz terrane taken from samples along the seismic profile show P-wave velocities between 4.5 and 5.2 km/s for near-surface pressures. However, similar velocities would be expected for metamorphosed sedimentary rocks of the accretionary complex. High-velocity rocks (6.5–7.7 km/s) are imaged in the lower continental crust beneath the surface outcrops of the Siletz volcanics in the Coast Ranges (km 190–230) and extend to the top of the descending slab, persisting to ~20 km offshore. If these high-velocity rocks are interpreted as Siletz terrane, then the total thickness offshore is 20 km, and the Siletz-accretionary boundary dips west from the coastline (Fig. 2B). If instead the high-velocity lower crustal rocks offshore are more recently accreted oceanic crust, then the Siletz-accretionary boundary may dip east (e.g., Snavely and Wagner, 1982). There is a lateral velocity change at km 240 in the model evidenced by a sudden shallowing of turning rays (Pg) (Fig. 2C), where higher-velocity rocks (>6.5 km/s) are found at shallower depth (10–11 km as compared with 17–18 km to the west). This shallowing of higher-velocity rocks to the east might result from an east-dipping Siletz boundary (Fig. 2). The maximum thickness of the Siletz volcanics beneath the Coast Ranges could reach 35 km if they extend to the base of the crust (Fig. 2B).

Cascadia Volcanic Arc and Back Arc Crust

The velocity model shown in Figure 2 provides a cross section through the Cascade Range and transition from the primarily collisional forearc to the extensional back arc. We observe a significant lateral change in upper-crustal velocity across the Mount St. Helens seismic zone (km 340); the 6.5 km/s contour is deeper on the east side of the seismic zone, whereas the 6.0 km/s contour is shallower (Fig. 2B). We interpret this boundary as the eastern extent of the Siletz terrane and transition into Cascade arc crust, and we interpret the 6.0–6.5 km/s zone beneath the volcanic arc as silicic intrusive rocks on the basis of laboratory velocity measurements made on plutonic inclusions taken from the Mount St. Helens lava dome (Paine, 1982). This seismically determined position for the eastern extent of the Siletz terrane and transition into the Cascade arc is in reasonable agreement with the boundary modeled from gravity data (Finn, 1990). East of the Cascade Range, the combined thickness of Columbia River Basalt flows and sub-basalt sedimentary units ranges from 6 to 8 km, in reasonable agreement with the more detailed studies of Saltus (1993) and Jarchow et al. (1994).

In the region east of about km 220, turning rays extend only to about 10–12 km depth. Thus crustal velocity beneath that depth was not directly measured by refracted waves but was instead modeled from arrivals reflected off the Moho (PmP). East of km 220, turning rays were refracted at about 10–16 km depth along a horizon that represents rocks having ≥6.5 km/s velocities; beneath the 6.5 km/s contour, we show a velocity gradient from 6.5 km/s to 7.0 km/s, and a crust-mantle transition from 7.5 to 7.7 km/s (Fig. 2B). We tested a number of crustal velocity gradients ranging from a uniform 6.5 km/s crust to a 6.5–7.5 km/s gradient, but did not introduce low-velocity zones. We found that the 6.5–7.0 km/s gradient produced the smallest travel time residuals for continental PmP reflections (traveltime misfit to within ±0.15 6s). The crust-mantle transition zone was required to fit reflection traveltimes from the descending slab beneath the crust and is consistent with low observed amplitudes of the continental PmP and Pn phases. The model shown in Figure 2 assumes no lateral lower crustal velocity contrasts, thus changes in crustal thickness could also be modeled by lateral velocity changes. Our attempts to model a flat Moho by inserting lateral velocity contrasts led to arbitrary and in some cases impossible crustal velocities.

Continental crustal thickness increases from ~20 km at the coastline to ~30 km beneath the Coast Ranges and Chehalis basin. The crust is thickest (40–45 km) beneath the Cascade Range where there is a broad root. The crust thins to about 30–38 km beneath the Columbia Plateau, slightly thinner than in the model of Catchings and Mooney (1988). A weak, but measurable Pn phase was observed from three explosive sources and shows upper-mantle velocities of about 7.8 km/s at the base of the crust-mantle gradient zone (Fig. 2). Upper mantle velocities were measured at 7.8–7.9 km/s beneath the oceanic crust offshore.

IMPLICATIONS FOR EARTHQUAKE HAZARDS

The image of the subduction process emerging from our model is one of a young (~10 m.y. old) and buoyant (e.g., Heaton and Hartzell, 1987), shallow-dipping (12°) Juan de Fuca plate that appears to bend as a result of overriding contact with rocks accreted to the North American plate. This observed subduction geometry concentrates the potential locked interplate

[1]GSA Data Repository item 9822, onshore-offshore wide-angle seismic sections, gravity model, and methodology, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301. E-mail: editing@geosociety.org.

contact offshore; the shallow dip broadens the zone out to the deformation front (~120 km offshore), and the steeper part of the slab encounters 350 °C temperatures at about 25–30 km depth (350 °C is thought to be the temperature where stable sliding begins; Hyndman and Wang, [1993]) near the coastline (Fig. 2B). A similar seismogenic depth limit was observed beneath Mexico, where the comparably young Cocos plate subducts (Tichelaar and Ruff, 1993). The interplate contact above the approximate depth of the 350 °C contour is devoid of even very small earthquakes (Fig. 2B). The low coupling stress along the interplate contact—as inferred from the presence of landward-dipping thrusts, high pore-fluid pressures (Flueh et al., 1997), and low frictional heating—implies that on the time scale of the seismic cycle, earthquakes are likely to be associated with a complete stress drop (Wang et al., 1995).

In the upper plate, the strong, mafic Siletz terrane rocks apparently play an important role in delimiting earthquakes along their eastern boundary. The correspondence between the eastern edge of the Siletz block and the onset of seismicity in the Mount St. Helens zone (Fig. 2B) may result from clockwise rotation (determined from paleomagnetics) of the coherent Siletz block that concentrates seismicity at its edges and limits internal seismogenic deformation (e.g., Wells, 1990; England and Wells, 1991; Trehu et al., 1994; Stanley et al., 1996). At present, the western edge of the Siletz terrane is not a focus of seismic activity. Based on the apparent influence of this boundary on accretionary complex deformation offshore of Oregon, where the western edge is overlain by short wavelength folds and faults (e.g., Trehu et al., 1995), we suspect that an analysis of sedimentary structures in this region would reveal similar effects; however, the details of this process likely depend strongly on the dip of Siletz terrane and on the nature of the high velocity material at depth beneath the shelf (Byrne et al., 1993).

ACKNOWLEDGMENTS

This work was supported by the National Earthquake Hazards Reduction Program; the German Ministry of Education, Research, Science, and Technology; and the USGS Deep Continental Surveys and Coastal and Marine Geology Programs. We thank Guy Cochrane, Michael Ellis, Harvey Kelsey, and Walter Mooney for helpful reviews.

REFERENCES CITED

Adams, J., 1984, Active deformation of the Pacific Northwest continental margin: Tectonics, v. 3, p. 449–472.

Adams, J., 1990, Paleoseismicity of the Cascadia subduction zone: Evidence from turbidites off the Oregon-Washington margin: Tectonics, v. 9, p. 569–583.

Atwater, B. F., 1996, Coastal evidence for great earthquakes in western Washington: U.S. Geological Survey Professional Paper 1560, p. 77–90.

Byrne, D. E, Wang, W., and Davis, D. M., Mechanical role of backstops in the growth of forearcs: Tectonics, v. 12, p. 123–144, 1993.

Catchings, R. D., and Mooney, W. D., 1988, Crustal structure of the Columbia Plateau: Evidence for continental rifting: Journal of Geophysical Research, v. 93, p. 459–474.

DeMets, C., Gordon, R. G., Argus, D. F., and Stein, S., 1990, Current plate motions: Geophysical Journal International, v. 101, p. 425–478.

Dewey, J. W., Hill, D. P., Ellsworth, W. L., and Engdahl, E. R., 1989, Earthquakes, faults, and the seismotectonic framework of the contiguous United States, *in* Pakiser, L. C., and Mooney, W. D., eds., Geophysical Framework of the continental United States: Geological Society of America Memoir 172, p. 541–576.

England, P. C., and Wells, R. E., 1991, Neogene rotations and quasicontinuous deformation of the Pacific Northwest continental margin: Geology, v. 19, p. 978–981.

Finn, C., 1990, Geophysical constraints on Washington convergent margin structure: Journal of Geophysical Research, v. 95, p. 19533–19546.

Flueh, E., Fisher, M., Scholl, D., Parsons, T., ten Brink, U., Klaeschen, D., Kukowski, N., Trehu, A., Childs, J., Bialas, J., and Vidal, N., 1997, Scientific teams analyze earthquake hazards of the Cascadia subduction zone: Eos (Transactions, American Geophysical Union), p. 153–157.

Heaton, T. H., and Kanamori, H., 1984, Seismic potential associated with subduction in the northwestern United States: Seismological Society of America Bulletin, v. 74, p. 933–941.

Heaton, T. H., and Hartzell, S. H., 1987, Earthquake hazards on the Cascadia subduction zone: Science, v. 236, p. 162–168.

Hole, J. A., and Zelt, B. C., 1995, 3-D finite-difference reflection travel times: Geophysical Journal International, v. 121, p. 427–434.

Hyndman, R. D., and Wang, K., 1993, Thermal constraints on the zone of major thrust earthquake failure: The Cascadia subduction zone: Journal of Geophysical Research, v. 98, p. 2039–2060.

Jarchow, C. M., Catchings, R. D., and Lutter, W. J., 1994, Large-explosive source, wide-recording aperture, seismic profiling on the Columbia Plateau, Washington: Geophysics, v. 59, p. 259–271.

Kelsey, H. M., Engebretson, D. C., Mitchell, C. E., and Ticknor, R. L., 1994, Topographic form of the Coast Ranges of the Cascadia margin in relation to coastal uplift rates and plate subduction: Journal of Geophysical Research, v. 99, p. 12245–12255.

Kirby, S., Engdahl, E. R., and Denlinger, R., 1996, Intermediate-depth intraslab earthquakes and arc volcanism as physical expressions of crustal and uppermost mantle metamorphism in subducting slabs, *in* Bebout, G. D., Scholl, D., Kirby, S., and Platt, J., eds., Subduction from top to bottom: American Geophysical Union, Geophysical Monograph, v. 96, p. 195–214.

McNeill, L., Piper, K. A., Goldfinger, C., Kulm, L. D., and Yeats, R. S., 1997, Listric normal faulting on the Cascadia continental margin: Journal of Geophysical Research, v. 102, p. 12123–12138.

Meyers, R. A., Smith, D. G., Jol, H. M., and Peterson, C. D., 1996, Evidence for eight great earthquake-subsidence events detected with ground-penetrating radar, Willapa Barrier, Washington: Geology, v. 24, p. 99–102.

Nelson, C. H., Goldfinger, C., Vallier, T. L., McGann, M. L., Kashgarian, M., 1996, North to south variation in Cascadia basin turbidite event history: Implications for paleoseismicity: Geological Society of America Abstracts with Programs, v. 28, no. 5, p. 95.

Paine, J. G., 1982, Crustal structure of volcanic arcs based on physical properties of andesites, volcaniclastic rocks, and inclusions in the Mt. St. Helens lava dome [M.S. thesis]: Seattle, University of Washington, 138 p.

Parsons, T., McCarthy, J., Kohler, W. M., Ammon, C. J., Benz, H. M., Hole, J. A., and Criley, E. E., 1996, The crustal structure of the Colorado Plateau, Arizona: Application of new long-offset seismic data analysis techniques: Journal of Geophysical Research, v. 101, p. 11173–11194.

Saltus, R. W., 1993, Upper-crustal structure beneath the Columbia River basalt group, Washington: Gravity interpretation controlled by borehole and seismic studies: Geological Society of America Bulletin, v. 105, p. 1247–1259.

Satake, K., Shimazaki, K., Tsuji, Y, and Ueda, K., 1996, Time and size of a giant earthquake in Cascadia inferred from Japanese tsunami records of January 1700: Nature, v. 379, p. 246–249.

Savage, J. C., Lisowski, M., and Prescott, W. S., 1991, Strain accumulation in western Washington: Journal of Geophysical Research, v. 96, p. 14493–14507.

Snavely, P. D., and Wagner, H. C., 1982, Geologic cross section across the continental margin of southwestern Washington: U.S. Geological Survey Open File Report 82-459, 10 p.

Stanley, W. D., Johnson, S. Y., Qamar, A. I., Weaver, C. S., and Williams, J. M., 1996, Tectonics and seismicity of the southern Washington Cascade Range: Seismological Society of America Bulletin, v. 86, p. 1–18.

Taber, J. J., and Lewis, B. T. R., 1986, Crustal structure of the Washington continental margin from refraction data: Seismological Society of America Bulletin, v. 76, p. 1011–1024.

Tichelaar, B. W., and Ruff, L. J., 1993, Depth of seismic coupling along subduction zones: Journal of Geophysical Research, v. 98, p. 2017–2037.

Trehu, A. M., Asudeh, I., Brocher, T. M., Luetgert, J., Mooney, W. D., Nabelek, J. L., and Nakamura, Y., 1994, Crustal architecture of the Cascadia forearc: Science, v. 266, p. 237–243.

Trehu, A. M., Lin, G., Maxwell, E., and Goldfinger, C., 1995, A seismic reflection profile across the Cascadia subduction zone offshore central Oregon: New constraints on methane distribution and crustal structure: Journal of Geophysical Research, v. 100, p. 15101–15116.

Walsh, T. J., Korosec, M. A., Phillips, W. M., Logan, R. L., and Schasse, H. W., 1987, Geologic map of Washington, southwest quadrant: Washington Division of Geology and Earth Resources Geologic Map GM-34, scale 1:250 000.

Wang, K., Mulder, T., Rogers, G. C., and Hyndman, R. D., 1995, Case for very low coupling stress on the Cascadia subduction fault: Journal of Geophysical Research, v. 100, p. 12907–12918.

Wells, R. E., 1990, Paleomagnetic rotations and the Cenozoic tectonics of the Cascade arc, Washington, Oregon, and California: Journal of Geophysical Research, v. 95, p. 19409–19417.

Manuscript received July 22, 1997
Revised manuscript received November 12, 1997
Manuscript accepted December 22, 1997