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High pore pressures and porosity at 35 km depth in the Cascadia subduction zone

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ABSTRACT

In the Cascadia subduction zone, beneath southern Vancouver Island at 25–45 km depth, converted teleseismic waves reveal an ~5-km-thick landward-dipping layer with anomalously high V_p/V_s averaging 2.35 ± 0.10 (2σ), interpreted as subducted oceanic crust of the Juan de Fuca plate. This layer is observed downdip of the inferred locked seismogenic zone, in the region of episodic tremor and slip. Laboratory velocity measurements of crystalline rock samples made at 200 MPa confining pressure and elevated pore pressures demonstrate that V_p/V_s increases with increasing fluid-filled porosity. The observed high V_p/V_s values are best explained by pore fluids under near lithostatic pressure in a layer with a high porosity of 2.7%–4.0%. Such large volumes of fluid take ~1 m.y. to accumulate based on reasonable rates of metamorphic fluid production of $\sim 10^{-4}$ m³/(m² yr) in subducting Juan de Fuca crust and mantle. Accordingly, the permeability of the plate interface at these depths must be very low, $\sim 10^{-24}$ to $\sim 10^{-21}$ m², or the porous layer must have a permeability $< 3 \times 10^{-20}$ m².

INTRODUCTION

H₂O-rich fluids play a fundamental role in convergent plate boundary deformation and seismicity. Recently, hydrous fluids have been suggested to facilitate episodic tremor and slip (ETS) (see review by Gomberg et al., 2010), a phenomenon first documented in the Cascadia subduction zone (northwestern United States and southwestern Canada) by Rogers and Dragert (2003). Over subduction time scales (millions of years), large amounts of hydrous fluids are produced from subducted materials as a result of prograde metamorphic dehydration reactions (e.g., Peacock, 2004), but how much free H₂O exists at depth in subduction zones? Does H₂O produced by metamorphic dehydration reactions escape upward rapidly or does the H₂O accumulate over thousand- to million-year time scales? We combine laboratory measurements of seismic properties with published seismological observations of the Cascadia subduction zone to quantify the porosity of subducted oceanic crust, and thereby gain insight into the distribution of H₂O at depth, subduction zone seismicity including ETS, and the rheology of the subduction plate interface.

GEOLOGIC AND GEOPHYSICAL OBSERVATIONS OF THE NORTHERN CASCADIA FOREARC

In the Cascadia subduction zone, warm lithosphere of the Juan de Fuca plate subducts beneath the North American plate at ~40 mm/yr (Fig. 1). Beneath southern Vancouver Island, onshore LITHOPROBE seismic reflection lines reveal a prominent band of subhorizontal seismic reflectors at ~30 km depth, known as the E-layer (Clowes et al., 1987). The E-layer is ~5 km thick, dips to the northeast, and is interpreted as either (1) the top of the subducting Juan de Fuca plate (e.g., Green et al., 1986), or (2) within the forearc crust above the subducting Juan de Fuca plate (e.g., Hyndman, 1988). S-wave diffraction tomographic profiles (receiver functions) of the Cascadia subduction zone forearc reveal a prominent low-velocity layer that dips toward the volcanic arc (e.g., Bostock et al., 2002; Abers et al., 2009) (Fig. 2A), which we interpret as subducting oceanic crust of the Juan de Fuca plate, an interpretation consistent with that of similar structures in other subduction zones (Rondenay et al., 2008).

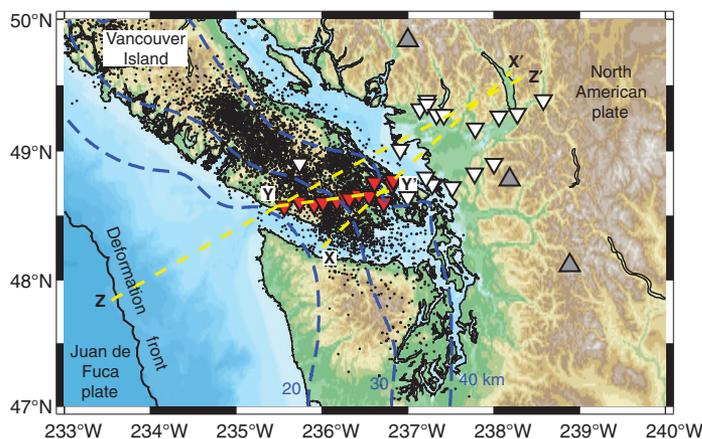


Figure 1. Geotectonic setting of Cascadia subduction zone showing depth to subducting Juan de Fuca plate (blue dashed lines; from Audet et al., 2010), location of episodic tremor and slip events (black dots; from Kao et al., 2009), Holocene arc volcanoes (gray triangles), and seismometers used for S-wave diffraction profile (white triangles) and V_p/V_s analysis (red triangles). Sections X-X' and Y-Y' refer to Figure 2, and section Z-Z' refers to Figure 4.

Audet et al. (2009) used scattered teleseismic body waves to investigate the physical properties of the prominent low-velocity layer dipping landward beneath southern Vancouver Island. Using the traveltimes of the scattered phases (Fig. 2B), the layer was found to be 5 ± 2 km thick with extremely high V_p/V_s ranging from 2.0 to 2.8 (Audet et al., 2009) (Fig. 2C). The estimated V_p/V_s of the layer beneath all 13 stations, based on a weighted average, is 2.35 ± 0.10 (2σ). These values are consistent with, but higher than, V_p/V_s values > 2.0 determined beneath southwestern Japan using seismic tomography (Kodaira et al., 2004) and V_p/V_s values of ~1.9 inferred beneath central Washington using receiver function techniques (Abers et al., 2009). As discussed by Audet et al. (2009), velocity images derived from tomography tend to be biased toward background values. Differences between the central Washington and Vancouver Island studies may reflect variations along strike (Abers et al., 2009) or increased resolution in the Audet et al. (2009) results enabled by the use of higher and broader frequency bandwidth data.

In the Audet et al. (2009) study, the amplitudes of scattered waves revealed that the boundary between the subducted oceanic crust and the overriding mafic-rich crust of Vancouver Island (i.e., the plate interface in our interpretation) exhibits a V_s contrast comparable in magnitude and opposite in sign to the lower boundary that marks the subducted oceanic Moho. These unusual observations were interpreted (Audet et al., 2009) to represent the presence of near-lithostatic pore-fluid pressures within the subducting crust that, in turn, indicates a largely impermeable plate boundary.

Subduction zone ETS tend to occur in regions possessing unusual seismic properties, such as high V_p/V_s , and observations of seismically triggered and tidally modulated tremor demonstrate that stress changes of several to tens of kilopascals can trigger ETS, implying that the plate interface at those depths is critically stressed and/or effective stresses are

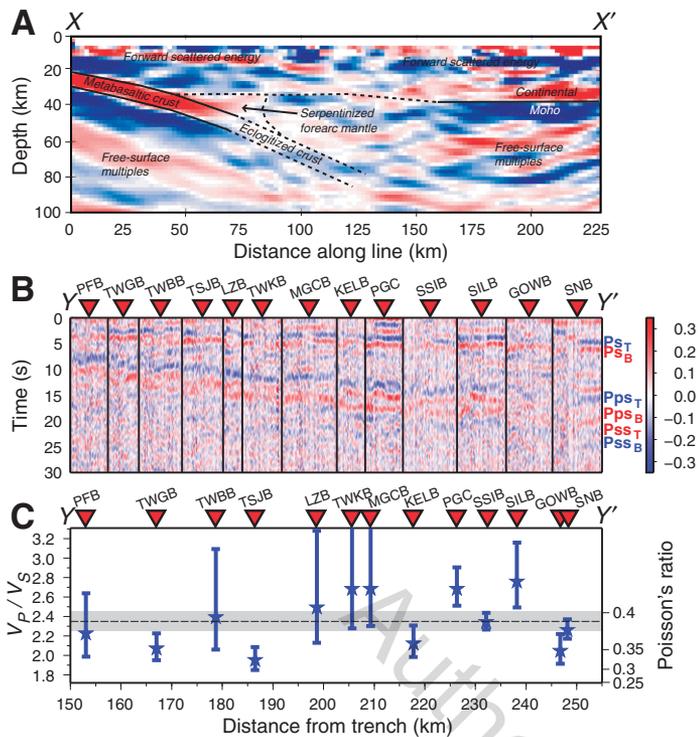


Figure 2. Seismological observations regarding low V_s layer beneath southern Vancouver Island. **A:** S-wave diffraction tomographic profile (X-X' in Fig. 1) showing east-dipping low- V_s layer, interpreted to represent subducting oceanic crust of Juan de Fuca plate, and continental Moho beneath volcanic arc (after Nicholson et al., 2005). **B:** Teleseismic receiver functions from 13 stations of the Canadian National Seismograph Network on southern Vancouver Island aligned from west to east along profile Y-Y' in Figure 1 (after Audet et al., 2009). **C:** Calculated V_p/V_s for low-velocity layer beneath each station in B with 1σ error bars. Average layer V_p/V_s of 13 stations = 2.35 ± 0.10 (2σ).

very low (see references in Gombert et al., 2010). Numerical experiments employing rate- and state-dependent friction successfully reproduce the 14 month Cascadia ETS recurrence interval when effective stresses are 2–3 MPa (Liu and Rice, 2007). In the analysis that follows, we assume that near-lithostatic fluid pressures in the ETS region reduce effective stresses to 0–5 MPa.

LABORATORY EXPERIMENTAL DATA

The observed V_p/V_s values of 2.35 ± 0.10 beneath southern Vancouver Island are higher than all laboratory V_p/V_s measurements made on an extensive range of rock types at a confining pressure of 1 GPa (Fig. 3A). The rocks measured in the laboratory span a wide range of silica contents and include all likely metamorphic and igneous rocks in this tectonic setting, such as serpentinite, amphibolite, eclogite, granulite, basalt, gabbro, peridotite, and granite. Most common rock types have measured V_p/V_s values between 1.7 and 1.9. Lizardite serpentinite, a rock that forms by hydration of ultramafic rocks ($\text{SiO}_2 \sim 40\%$) at low temperatures, has an unusually high measured V_p/V_s of 2.1, whereas the high-temperature antigorite serpentinite has a more normal V_p/V_s of 1.8. No measured dry rock type approaches the V_p/V_s values observed beneath southern Vancouver Island (Fig. 3A).

Rock porosity has an important effect on laboratory measurements of seismic velocities. In crystalline rocks, porosity is represented by cracks and grain boundaries. Studies of physical properties of relatively low porosity rocks with microcracks (e.g., Walsh, 1965; Christensen, 1989) show that crack closure depends on the differential pressure, P_{diff} , i.e.,

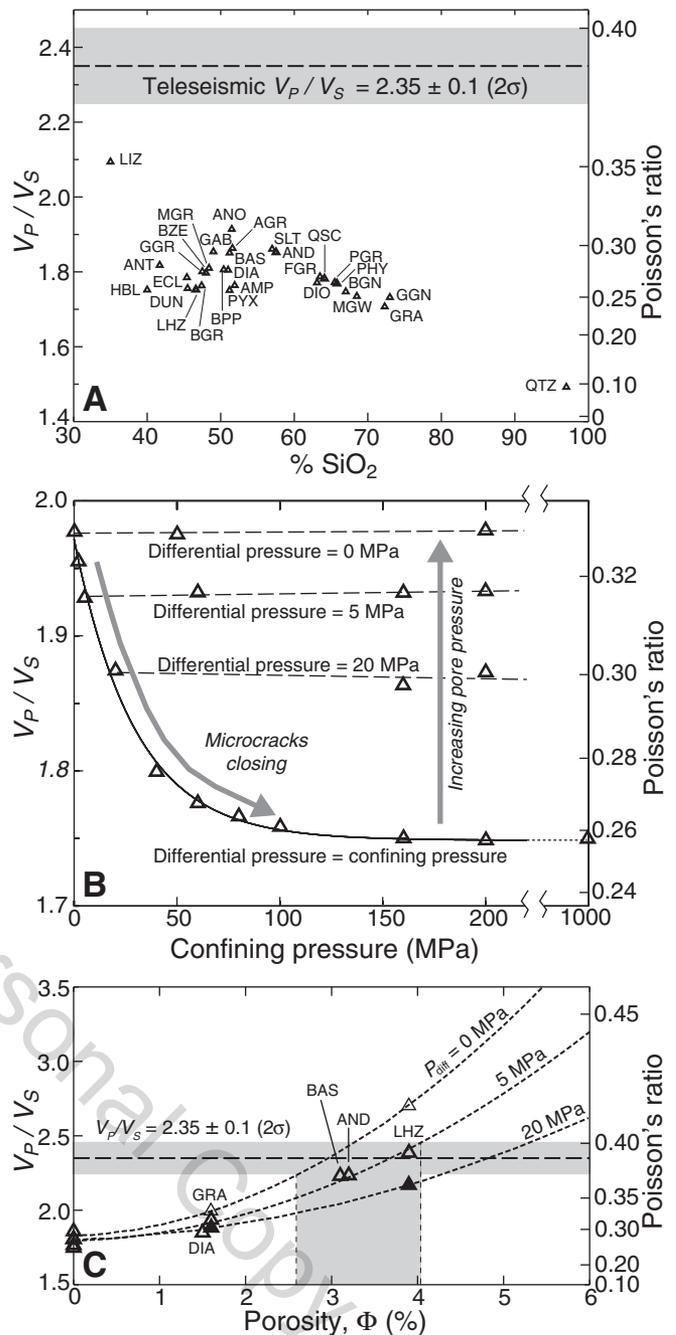


Figure 3. **A:** Lab measurements of V_p/V_s as function of SiO_2 (wt%) for wide range of igneous and metamorphic rocks at confining pressure, $P_{\text{conf}} = 1$ GPa (Christensen, 1996). Each data point represents tens to hundreds of lab measurements; typical standard deviation for $V_p/V_s < 0.1$ (Christensen, 1996). Teleseismic V_p/V_s values for subducted crust beneath southern Vancouver Island (gray band) exceed lab measurements of all dry rocks. Selected abbreviations: AND—andesite, ANT—antigorite, BAS—basalt, DIA—diabase, ECL—eclogite, GAB—gabbro, GRA—granite, LHZ—Iherzolite, LIZ—lizardite, QZT—quartzite. (For a full list of abbreviations, see the Data Repository [see footnote 1].) **B:** Lab measurements of V_p/V_s demonstrating effect of increasing P_{pore} on seismic properties of granite with 1.6% porosity (Christensen, 1989). **C:** Lab measurements of V_p/V_s as function of porosity for five crystalline samples: dredged oceanic basalt (BAS), Oman ophiolite diabase (DIA), Mount St. Helens andesite (AND), North American midcontinent granite (GRA), and a relatively high crack porosity Kilborne Hole Iherzolite xenolith (LHZ). V_p/V_s values for porosity (Φ) = 0 obtained at $P_{\text{conf}} = 1$ GPa. V_p/V_s values for $\Phi > 0$ obtained at $P_{\text{conf}} = 200$ MPa and $P_{\text{diff}} = 5$ MPa (five crystalline samples) and at $P_{\text{diff}} = 0$ and 20 MPa (GRA and LHZ only).

the difference between the confining (overburden) pressure, P_{conf} , and the internal fluid pore pressure, P_{pore} . Cracks in deeply buried rocks are likely open if P_{pore} is sufficiently high.

Christensen (1984, 1989) measured V_p and V_s as a function of P_{conf} and P_{pore} for a limited number of crystalline rocks with crack shape porosity. Values of V_p/V_s calculated from laboratory measurements of V_p and V_s in granite are shown in Figure 3B. The solid curve shows V_p/V_s values as a function of P_{conf} for a jacketed dry sample without applied pore pressure. Along this curve V_p/V_s values are quite high at low P_{conf} where microcracks remain open. As P_{conf} increases, V_p/V_s decreases dramatically; at $P_{\text{conf}} > 100$ MPa V_p/V_s remains fairly constant, reflecting crack closure. The dashed lines in Figure 3B show V_p/V_s at elevated pore pressures corresponding to $P_{\text{diff}} = 20, 5, \text{ and } 0$ MPa. Measured V_p/V_s increases with decreasing P_{diff} (increasing P_{pore}) and V_p/V_s values at $P_{\text{diff}} = 0$ are similar to the value at atmospheric pressure. This indicates that the cracks within this rock form a continuous network with fluid at a pressure equal to the P_{conf} filling the microcracks and preventing crack closure.

At high pore pressure, V_p/V_s in crystalline rocks show a positive correlation with porosity, Φ , and only a very weak dependence on pressure and temperature (Christensen, 1996). In Figure 3C the rock porosities of five crystalline samples are plotted against V_p/V_s at $P_{\text{conf}} = 200$ MPa and $P_{\text{pore}} = 195$ MPa ($P_{\text{diff}} = 5$ MPa). The zero porosity data points are the V_p/V_s values for the five samples measured from jacketed dry samples at a $P_{\text{conf}} = 1$ GPa. A quadratic fit to the data yields:

$$V_p/V_s = 0.036\Phi^2 + 0.0178\Phi + 1.79, \quad (1)$$

which can be used to estimate the porosity (in %) of crystalline rocks at depth from V_p/V_s observations. We also show quadratic fits to limited data at $P_{\text{diff}} = 0$ and 20 MPa (Fig. 3C).

POROSITY AND SOURCES OF H₂O FLUID

The observed V_p/V_s values of 2.35 ± 0.10 for the low-velocity layer beneath southern Vancouver Island are best matched by the properties of crystalline rocks with fluid-filled porosity of 2.7%–4.0%, with the fluid under near lithostatic pressure ($P_{\text{diff}} = 0$ –5 MPa) (Fig. 3C). The estimated porosity based on observed V_p/V_s is consistent with Hyndman's (1988) estimate of 1%–4% porosity based on the electrical conductivity and impedance contrast of the seismically reflective E-layer.

In a subduction zone setting, there are two potential sources of aqueous (H₂O rich) fluids: (1) evolved pore waters, and (2) metamorphic dehydration fluids. Oceanic sediments have measured porosities approaching 50%; however, sediment compaction and deformation (Moore and Vrolijk, 1992) and sedimentary offscraping and underplating (von Huene and Scholl, 1991) dramatically reduce the amount of pore H₂O contained in sediments subducted to depths >10 km. The uppermost 0.5–1 km of incoming oceanic crust has relatively high porosities of 5%–10% supported by a crystalline framework, which may persist to greater depths.

A more likely source of fluid at 25–45 km depth is metamorphic dehydration reactions. Beginning at shallow levels with the breakdown of clay minerals and extending to >100 km depth, metamorphic dehydration reactions release chemically bound H₂O from subducting sediment and variably hydrated oceanic crust and uppermost mantle. In the warm Cascadia subduction zone, thermal-petrologic models predict that $\sim 10^{-4}$ m³/(m² yr) of H₂O is produced from dehydrating oceanic crust at 35 km depth (Hyndman and Peacock, 2003), and a similar amount is expected to be produced by dehydration of partially serpentized uppermost mantle. These low fluid fluxes reflect the slow nature of subduction; in Cascadia, it takes ~ 4 m.y. for material subducted at the deformation front to reach 35 km depth. To produce enough metamorphic fluid to fill or create 3%–4% porosity, ~ 1 m.y. is required. During this time the metamorphic fluid must remain in the rock, which in turn requires that low permeability

be maintained for million-year time scales, far longer than the time scales associated with the megathrust seismic cycle (~ 500 yr) or ETS (~ 1 yr).

PERMEABILITY

In order to maintain significant porosity and high fluid pressures in the subducting oceanic crust (high V_p/V_s layer), either the layer must be capped by a very low permeability seal or the entire layer has low permeability. The permeability of a caprock or seal may be estimated from Darcy's Law, assuming a fluid production rate of 2×10^{-4} m³/(m² yr) within subducting oceanic crust and mantle and P_{pore} decreasing from lithostatic beneath the seal to hydrostatic at the top of the seal. For a seal thickness of 1–1000 m, the estimated permeability of the seal is 10^{-24} to 10^{-21} m² (see the GSA Data Repository¹ for calculation details). The plate interface may form a low-permeability seal as a result of deformation-induced grain-size reduction and the development of foliated cataclasites and mylonites containing platy hydrous minerals like talc, serpentine, and micas. A low-permeability seal may also form in the overlying forearc crust as a result of mineral precipitation from migrating fluids or volume-expanding metamorphic hydration reactions. Rather than being capped by a low-permeability seal, the high V_p/V_s layer may have low permeability. Using Manning and Ingebritsen's (1999) analysis of the conditions under which metamorphism leads to anomalous pore pressures, and a fluid production rate of 10^{-4} m³/(m² yr) within subducting oceanic crust, a layer permeability $< 3 \times 10^{-20}$ m² is necessary to retain the metamorphic dehydration fluids. These estimates are consistent with Hyndman's (1988) estimated E-layer permeability of $< 10^{-20}$ m² based on a porous media compaction calculation and Manning and Ingebritsen's (1999) permeability estimate of 10^{-19} m² for continental crust at 35 km depth based on rocks metamorphosed at slightly shallower depths.

PROPERTIES OF THE PLATE INTERFACE AND SEISMICITY AS A FUNCTION OF DEPTH

The observations presented here, combined with many previous studies, suggest that high pore pressures are a fundamental characteristic of subduction zone plate boundaries (Fig. 4). Beneath accretionary prisms, seismic reflection studies and drilling observations document high pore pressures (e.g., Moore, 1989), which promote low friction along the low-angle décollement (plate boundary). Along the seismogenic portion of the subduction plate boundary, elevated pore pressures are relieved episodically by large thrust earthquakes (e.g., Magee and Zoback, 1993). Deeper

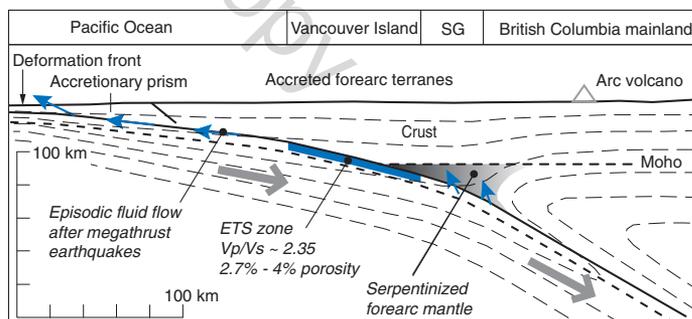


Figure 4. Schematic cross section through southern Vancouver Island (profile Z-Z' in Fig. 1) illustrating proposed porosity and fluid flow structure along Cascadia plate interface. Thin dashed lines = 200 °C isotherms. SG—Strait of Georgia; ETS—episodic tremor and slip.

¹GSA Data Repository item 2011151, permeability calculations and a full list of rock abbreviations, is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

in the Cascadia subduction zone, continuous global positioning system data reveal that slow slip events occur periodically on the plate interface downdip of the seismogenic zone (e.g., Rogers and Dragert, 2003; Brown et al., 2009). The results presented here indicate that plate-interface ETS occur in H₂O-rich, porous crystalline rocks under near lithostatic fluid pressures, supporting models in which H₂O-rich fluids released by dehydration reactions trigger or enable ETS (e.g., Kodaira et al., 2004; see references in Gomberg et al., 2010). In particular, ETS appear to occur in a relatively porous, fluid-rich environment. Unlike the shallower seismogenic zone, fluids in the ETS region appear to remain trapped (Fig. 4). Additional seismological studies are needed to determine whether the high V_p/V_s (and high porosity) layer extends updip into the seismogenic zone.

Below ~35 km depth, where the forearc Moho intersects the subducting Juan de Fuca plate, the V_s contrast marking the upper boundary of the low-velocity layer disappears (Fig. 2A). Beneath the forearc Moho, there is strong geophysical evidence for extensive serpentinization of the forearc mantle wedge in the Cascadia and many other subduction zones (e.g., Hyndman and Peacock, 2003). The evidence for pervasive serpentinization of the Cascadia forearc mantle (Bostock et al., 2002; Brocher et al., 2003) requires that substantial quantities of H₂O migrate upward through the plate boundary over the history of the subduction zone. Such migration may reflect higher vertical permeabilities of the subducting oceanic crust and/or interface beneath the forearc mantle as a result of densification eclogite-forming reactions in the subducting crust and volume expansion serpentinization reactions in the overlying forearc mantle (Audet et al., 2009).

CONCLUSIONS

Seismological observations of unusually high V_p/V_s regions in subsolidus crystalline crust are best explained by high, near lithostatic fluid pressures, and laboratory measurements of crystalline rocks permit estimates of the porosity in these regions. Applying these data to the Cascadia subduction zone, we conclude that the subducting Juan de Fuca crust beneath southern Vancouver Island contains 2.7–4.0 vol% fluid in crack porosity. The most likely sources of these fluids are metamorphic dehydration reactions in the subducting plate. Permeabilities are sufficiently low to maintain these high porosities for million-year time scales, despite this being an actively deforming plate boundary region. Porosities may be similarly high in other regions of active metamorphism, such as continent collision belts. Our results suggest caution in interpreting seismic tomographic images of crystalline rocks because, in addition to mineralogy, pressure, and temperature, seismic velocities are also a function of pore pressure and porosity.

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REFERENCES CITED

Abers, G.A., MacKenzie, L.S., Rondenay, S., Zhang, Z., Wech, A.G., and Creager, K.C., 2009, Imaging the source region of Cascadia tremor and intermediate-depth earthquakes: *Geology*, v. 37, p. 1119–1122, doi: 10.1130/G30143A.1.

Audet, P., Bostock, M.G., Christensen, N.I., and Peacock, S.M., 2009, Seismic evidence for overpressured subducted oceanic crust and megathrust fault sealing: *Nature*, v. 457, p. 76–78, doi: 10.1038/nature07650.

Audet, P., Bostock, M.G., Boyarko, D.C., Brudzinski, M.R., and Allen, R.M., 2010, Slab morphology in the Cascadia fore arc and its relation to episodic tremor and slip: *Journal of Geophysical Research*, v. 115, B00A16, doi: 10.1029/2008JB006053.

Bostock, M.G., Hyndman, R.D., Rondenay, S., and Peacock, S.M., 2002, An inverted continental Moho and serpentinization of the forearc mantle: *Nature*, v. 417, p. 536–538, doi: 10.1038/417536a.

Brocher, T.M., Parsons, T., Tréhu, A.M., Snelson, C.M., and Fisher, M.A., 2003, Seismic evidence for widespread serpentinized forearc upper mantle along the Cascadia margin: *Geology*, v. 31, p. 267–270, doi: 10.1130/0091-7613(2003)031<0267:SEFWFS>2.0.CO;2.

Brown, J.R., Beroza, G.C., Ide, S., Ohta, K., Shelly, D.R., Schwartz, S.Y., Rabbel, W., Thorwart, M., and Kao, H., 2009, Deep low-frequency earthquakes in tremor localize to the plate interface in multiple subduction zones: *Geophysical Research Letters*, v. 36, L19306, doi: 10.1029/2009GL040027.

Christensen, N.I., 1984, Pore pressure and oceanic crustal seismic structure: *Royal Astronomical Society Geophysical Journal*, v. 79, p. 411–424.

Christensen, N.I., 1989, Pore pressure, seismic velocities, and crustal structure, in Pakiser, L.C., and Mooney, W.D., eds., *Geophysical framework of the continental United States: Geological Society of America Memoir* 172, p. 783–798.

Christensen, N.I., 1996, Poisson's ratio and crustal seismology: *Journal of Geophysical Research*, v. 101, no. B2, p. 3139–3156, doi: 10.1029/95JB03446.

Clowes, R.M., Brandon, M.T., Green, A.G., Yorath, C.J., Sutherland Brown, A., Kanasevich, E.R., and Spencer, C., 1987, LITHOPROBE—Southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections: *Canadian Journal of Earth Sciences*, v. 24, p. 31–51.

Gomberg, J., and the Cascadia 2007 and Beyond Working Group, 2010, Slow-slip phenomena in Cascadia from 2007 and beyond: A review: *Geological Society of America Bulletin*, v. 122, p. 963–978, doi: 10.1130/B30287.1.

Green, A.G., Clowes, R.M., Yorath, C.J., Spencer, C., Kanasevich, E.R., Brandon, M.T., and Sutherland Brown, A., 1986, Seismic reflection imaging of the subducting Juan de Fuca plate: *Nature*, v. 319, p. 210–213, doi: 10.1038/319210a0.

Hyndman, R.D., 1988, Dipping seismic reflectors, electrically conductive zones, and trapped water in the crust over a subducting plate: *Journal of Geophysical Research*, v. 93, no. B11, p. 13,391–13,405, doi: 10.1029/JB093iB11p13391.

Hyndman, R.D., and Peacock, S.M., 2003, Serpentinization of the forearc mantle: *Earth and Planetary Science Letters*, v. 212, p. 417–432, doi: 10.1016/S0012-821X(03)00263-2.

Kao, H., Shan, S.-J., Dragert, H., and Rogers, G., 2009, Northern Cascadia episodic tremor and slip: A decade of tremor observations from 1997 to 2007: *Journal of Geophysical Research*, v. 114, B00A12, doi: 10.1029/2008JB006046.

Kodaira, S., Iidaka, T., Kato, A., Park, J.-O., Iwasaki, T., and Kaneda, Y., 2004, High pore fluid pressure may cause silent slip in the Nankai Trough: *Science*, v. 304, p. 1295–1298, doi: 10.1126/science.1096535.

Liu, Y., and Rice, J.R., 2007, Spontaneous and triggered aseismic deformation transients in a subduction fault model: *Journal of Geophysical Research*, v. 112, B09404, doi: 10.1029/2007JB004930.

Magee, M.E., and Zoback, M.D., 1993, Evidence for a weak interplate thrust fault along the northern Japan subduction zone and implications for the mechanics of thrust faulting and fluid expulsion: *Geology*, v. 21, p. 809–812, doi: 10.1130/0091-7613(1993)021<0809:EFAWIT>2.3.CO;2.

Manning, C.E., and Ingebritsen, S.E., 1999, Permeability of the continental crust: Implications of geothermal data and metamorphic systems: *Reviews of Geophysics*, v. 37, p. 127–150, doi: 1998RG900002.

Moore, J.C., 1989, Tectonics and hydrogeology of accretionary prisms: Role of the decollement zone: *Journal of Structural Geology*, v. 11, p. 95–106, doi: 10.1016/0191-8141(89)90037-0.

Moore, J.C., and Vrolijk, P., 1992, Fluids in accretionary prisms: *Reviews of Geophysics*, v. 30, p. 113–135, doi: 10.1029/92RG00201.

Nicholson, T., Bostock, M.G., and Cassidy, J.F., 2005, New constraints on subduction zone structure in northern Cascadia: *Geophysical Journal International*, v. 161, doi: 10.1111/j.1365-246X.2005.02605.x.

Peacock, S.M., 2004, Insight into the hydrogeology and alteration of oceanic lithosphere based on subduction zones and arc volcanism, in Davis, E.E., and Elderfield, H., eds., *Hydrogeology of the oceanic lithosphere*: Cambridge, UK, Cambridge University Press, p. 659–676.

Rogers, G., and Dragert, H., 2003, Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip: *Science*, v. 300, p. 1942–1943, doi: 10.1126/science.1084783.

Rondenay, S., Abers, G.A., and van Keken, P.E., 2008, Seismic imaging of subduction zone metamorphism: *Geology*, v. 36, p. 275–278, doi: 10.1130/G24112A.1.

von Huene, R., and Scholl, D., 1991, Observations at convergent margins concerning sediment subduction, sediment erosion, and the growth of continental crust: *Reviews of Geophysics*, v. 29, p. 279–316, doi: 10.1029/91RG00969.

Walsh, J.B., 1965, The effect of cracks on the compressibility of rock: *Journal of Geophysical Research*, v. 70, p. 381–389, doi: 10.1029/JZ070i002p00381.

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