Numerical modelling:

The governing equation used in this study is:

$$\nabla \bullet (K\nabla T) - \rho c v \bullet \nabla T + H = 0,$$

where K is thermal conductivity, T is temperature, ρ is density, c is specific heat, ν is velocity of the subducting plate, and H represents sources of heat. Both solid material and fluid in pore space of the oceanic plate are advected at the convergence rate. The model does not include coupled fluid and heat transport; rather, we use a high Nusselt number proxy to simulate the thermal effects of fluid circulation. It can be difficult to achieve numerical stability in models of coupled fluid and heat transport for extremely permeable (~10^{-10} - 10^{-9} m^2) basaltic basement aquifer of oceanic crust. In such high permeability systems, the efficiency of convective heat transfer is very high; the Nusselt number (*Nu*), quantifying the efficiency of convective heat transfer, is greater than 100 (Davis et al., 1997; Spinelli and Fisher, 2004; Kummer and Spinelli, 2008; Spinelli and Wang, 2008). In systems with hydrothermal circulation vigorous enough to have *Nu* in excess of 20, a high-conductivity proxy for the thermal effects of the fluid circulation is accurate (Davis et al., 1997). In this approach, a Rayleigh number (*Ra*) is defined for each aquifer element using the permeability, temperature-dependent fluid density and viscosity, and local heat flux:

$$Ra = \frac{\alpha g k L^2 \rho_f q}{\mu \kappa K},$$

where α is fluid thermal expansivity, g is gravity, k is permeability, L is aquifer thickness, ρ_f is fluid density, q is conductive heat flux, μ is fluid viscosity, and κ is thermal diffusivity. In our simulations, aquifer permeability decreases with depth; in all simulations the logarithm of permeability decreases by 0.01 per kilometer depth (Fig. 2). An elemental *Nu* is derived from *Ra* using the empirical *Ra-Nu* relationship found by comparing results from a coupled fluid and heat transport simulation to a conductive proxy of the same problem (Kummer and Spinelli, 2008; Spinelli and Wang, 2008).

$$Nu = 0.08Ra^{0.89}$$

This *Ra-Nu* relationship is within the range of the host of published *Ra-Nu* relationships (Wang, 2004). Multiplying the intrinsic thermal conductivity of the aquifer by *Nu* simulates the thermal effects of hydrothermal circulation. Adjacent aquifer elements with elevated thermal conductivity facilitate lateral heat transport, simulating the effect of heat transport via wide convection cells in the basement aquifer of oceanic crust (Wang, 2004; Spinelli and Fisher, 2004; Kummer and Spinelli, 2009). Because of nonlinear feedbacks between temperature and *Nu*, an iterative procedure is used. Starting with a temperature field found using a conductive model (i.e., no fluid circulation; $Nu = 1$), a convergent solution (temperature variation of <1 ˚C in each element) is obtained usually within 10 iterations.

For each of the four transects examined, the geometry of the subduction zone is constrained by seismic reflection and refraction data, earthquake hypocenters, and teleseismic travel-time data (Fuis, 1998; Gerdom et al., 2000; Flueh et al., 1998; Gulick et al., 1998; McCrory et al., 2006). The slab is coupled to the mantle wedge, inducing mantle wedge flow, only after the top of the subducting slab has reached a depth of 70 km. Thus, where the plate boundary is shallower than 70 km depth, the overlying tip of the mantle wedge is stagnant. This is consistent with slab-mantle decoupling to 70-80 km depth found by Wada and Wang (2009) for Cascadia and numerous other subduction zones. We use an isoviscous mantle wedge rheology. The regions of interest for this study, the seismogenic portion of the plate boundary fault and the subducting crust to the depth of the basalt-to-eclogite transition, are >60 km updip of the region of mantle wedge flow. Temperatures and surface heat flux in these forearc regions are primarily controlled by advection of the subducting slab, heat conduction, and hydrothermal circulation; they are minimally influenced by mantle wedge flow (Currie et al., 2004). We do not use our modeled temperatures to evaluate sub-arc or back-arc processes, where the thermal field is significantly influenced by mantle wedge rheology (Currie et al., 2004).

Temperature at the upper boundary of the model is 0 ˚C, and at the base of the oceanic lithosphere it is 1400 ˚C. We use a geotherm consistent with a back-arc setting at the landward boundary. At the seaward boundary (200 km seaward of the trench), we use geotherms for conductively cooled oceanic lithosphere. We use an effective friction coefficient in the seismogenic portion of the plate boundary fault of 0.03, consistent with numerous subduction zone thermal models (e.g., Wada and Wang, 2009; Harris et al., 2010). Mechanical and thermal studies indicate that the effective friction coefficient is commonly <0.05 (Wang et al., 1995; Wang and He, 1999). Finally, we do not consider the small fluid sources due to dehydration reactions in the slab. They are important for fluid flow only if the permeability is extremely low, in which case the slow fluid flow is not thermally significant.

Surface heat flux observations:

Surface heat flux observations are from seafloor probe measurements, temperature gradient estimates from the depth to a gas hydrate related bottom simulating reflector, and temperature gradients in deep Ocean Drilling Program sites, exploration wells on the continental shelf, and boreholes on land for the Vancouver Island (Lewis et al., 1988; Davis et al., 1990; Hyndman and Wang, 1993; Hyndman and Wang, 1995), Washington (Blackwell et al., 1990; Hyndman and Wang, 1995; Booth-Rea et al., 2008), Oregon (Hyndman and Wang, 1995; Trehu et al., 1995; Trehu, 2006), and California (Lachenbruch and Sass, 1980; Mase et al., 1982; Shipboard Scientific Party, 1997; Villinger et al., 2010) transects. Heat flux estimates from the deep boreholes are the most robust, with uncertainties of approximately 5-10% (e.g., Davis et al., 1990). For the shallow probe measurements, error estimates primarily results from uncertainty in the thermal conductivity; the average uncertainty for the probe measurements is ±11% (Davis et al., 1990). The scatter in the seafloor probe measurements (particularly on the accretionary prism) may results from localized advection of fluid and heat from the dewatering sediment. The greatest uncertainty (10-55%) is associated with heat flux estimates derived from the depth to a bottom-simulating reflector (BSR) (Grevemeyer and Villinger, 2001; Villinger et al., 2010). Errors in calculating heat flux from the BSR result from uncertainty in: 1) the depth to the BSR on seismic reflection lines, 2) hydrocarbon composition, 3) pore water chemistry, 4) seafloor temperature, and 5) thermal conductivity (e.g., Davis et al., 1990; Grevemeyer and Villinger, 2001; Villinger et al., 2010). Uncertainties in BSR derived heat flux estimates are smallest if the temperature at the BSR can be constrained by probe or borehole measurements (Grevemeyer and Villinger, 2001; Villinger et al., 2010). Where probe or borehole observations are not co-located with the seismic reflection data used to derive heat flux estimates from BSR depth (e.g., the Washington transect), uncertainties may be large.

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