Shear heating reconciles thermal models with the metamorphic rock record of subduction

Matthew J. Kohn\textsuperscript{a,1}, Adrian E. Castro\textsuperscript{b}, Buchanan C. Kerswell\textsuperscript{a}, César R. Ranero\textsuperscript{c,d}, and Frank S. Spear\textsuperscript{b}

\textsuperscript{a}Department of Geosciences, Boise State University, Boise, ID 83725; \textsuperscript{b}Department of Earth and Environmental Sciences, Rensselaer Polytechnic Institute, Troy, NY 12180; \textsuperscript{c}Instituto de Ciencias del Mar, Spanish National Research Council, 08003 Barcelona, Spain; and \textsuperscript{d}Institució Catalana de Recerca i Estudis Avançats, 08010 Barcelona, Spain

Edited by Peter B. Kelemen, Lamont-Doherty Earth Observatory, Palisades, NY, and approved October 4, 2018 (received for review June 14, 2018)

Some commonly referenced thermal-mechanical models of current subduction zones imply temperatures that are 100–500 °C colder at 30–80-km depth than pressure-temperature conditions determined thermobarometrically from exhumed metamorphic rocks. Accurately inferring subduction zone thermal structure, whether from models or rocks, is crucial for predicting metamorphic reactions and associated fluid release, subarc melting conditions, rheologies, and fault-slip phenomena. Here, we compile surface heat flow data from subduction zones worldwide and show that values are higher than can be explained for a frictionless subduction interface often assumed for modeling. An additional heat source—likely shear heating—is required to explain these forearc heat flow values. A friction coefficient of at least 0.03 and possibly as high as 0.1 in some cases explains these data, and we recommend a provisional average value of 0.05 ± 0.015 for modeling. Even small coefficients of friction can contribute several hundred degrees of heating at depths of 30–80 km. Adding such shear stresses to thermal models quantitatively reproduces the pressure-temperature conditions recorded by exhumed metamorphic rocks. Comparatively higher temperatures generally drive rock dehydration and densification, so, at a given depth, hotter rocks are denser than colder rocks, and harder to exhumation through buoyancy mechanisms. Consequently—versus previous proposals—exhumed metamorphic rocks might overrepresent old-cold subduction where rocks at the slab interface are wetter and more buoyant than in young-hot subduction zones.

Determination of the pressure–temperature (P–T) conditions of subduction zone metamorphism is key for understanding rock rheology, fluid-mass transfer and geochemical cycling of carbon, trace elements, and other chemical species between Earth’s crust and mantle (e.g., refs. 1–3), genesis and chemistry of arc volcanoces (e.g., ref. 4), origin of shallow interplate seismicity (e.g., refs. 5–7), and H\textsubscript{2}O release vs. storage in the mantle (e.g., refs. 8–10). Interpretations of these processes all depend strongly on thermal structure, yet at depths of 30–80 km, a 100–500 °C discrepancy occurs between many commonly cited, frictionless thermal models (11) and P–T conditions of subduction zone metamorphic rocks, as determined either from metamorphic facies (12, 13) or from thermobarometry (14–16). In this context, we make several comparisons to the most recent and comprehensive data compilation for rocks (16), using the term “PD15” to refer to the reported dataset and average geotherm. As shown previously (16), bias to the calculated P–T conditions through postpeak processes (transfer to hanging wall, residence times, isothermal exhumation, etc.) does not likely explain model-data discrepancies because prograde P–T paths show the same P–T distribution as thermobarometry. Rather, many numerical models predict slab-top geotherms that are <5 °C/km and enter a region of P–T space that metamorphic petrologists sometimes denote as the “forbidden zone” (17). No known rocks up to 4 GPa record such cold P–T conditions (14, 15).

Explaining the large temperature discrepancy between rocks and models could reveal either omission or underestimation of possible heat sources in models (16), or bias in the exhumation of subduction zone metamorphic rocks (10). Indeed, it was argued recently that all rocks exhumed from subduction zones reflect only warm subduction conditions (10), for example from subduction of anomalously young crust or from subduction initiation. In fact, dehydration forms dense garnet and pyroxene, so warmer and comparatively drier rocks from the slab interface should be denser (14, 18) and possibly more difficult to exhumation than cooler, wetter rocks. Nonetheless, if widely accepted cold models (11) are correct, exhumed rocks provide few examples of P–T conditions predicted by these models.

To reassess temperatures along the subduction interface, we evaluated constraints on shear heating derived from global heat flow datasets and site-specific studies (see SI Appendix for sources). We then modeled the slab-top thermal structure using a simple but highly flexible analytical model (19, 20) that accounts for the effects of shear heating, thermal weakening, and other key subduction parameters (Methods and SI Appendix). We emphasize shear heating because it is a large potential contributor to slab-top temperature increases: Heat sources from reactions, fluid flow, radioactive heating, and convection have been estimated to be 2–10 times smaller (e.g., ref. 21). More recent work (22, 23) has shown that fluid flow along the top of the subducting slab may reduce temperatures by up to ~100 °C, which is comparable in magnitude to the effects of the lowest proposed coefficients of friction. Insofar as shear heating could theoretically increase slab-top geotherms by several hundred degrees Celsius (13, 21, 24), quantifying typical coefficients of friction and their thermal consequences is crucial to the reliability of any petrologic or geochemical interpretations.

Significance

Thermal structure controls numerous aspects of subduction zone metamorphism, rheology, and melting. Many thermal models assume small or negligible coefficients of friction and underpredict pressure–temperature (P–T) conditions recorded by subduction zone metamorphic rocks by degrees Celsius. Adding shear heating to thermal models simultaneously reproduces surface heat flow and the P–T conditions of exhumed metamorphic rocks. Hot dry rocks are denser than cold wet rocks, so rocks from young-hot subduction systems are denser and harder to exhumation through buoyancy. Thus, the metamorphic record may underrepresent hot-young subduction and overrepresent old-cold subduction.


The authors declare no conflict of interest.

This article is a PNAS Direct Submission. Published under the PNAS license.

1To whom correspondence should be addressed. Email: mattkohn@boisestate.edu.

This article contains supporting information online at www.pnas.org/lookup/suppl/doi:10.1073/pnas.1809962115/-/DCSupplemental.

www.pnas.org/cgi/doi/10.1073/pnas.1809962115

PNAS Latest Articles | 1 of 6
Fig. 1. Heat flow data (39) normalized to expected heat flow of incoming plate (25) vs. distance normalized to arc–trench distance. Model curves are labeled with assumed apparent coefficients of friction ($\mu^*$). Large dots with error bars are medians of binned data and their errors. Gray regions are affected by corner flow and are not considered in this study. (A) Raw data excluding Cascadia. Models are calculated using analytical approach of refs. 19 and 20, and assume global average modern subduction parameters: plate age = 50 Ma, convergence rate = 6 cm/yr, subducting plate geometry = central Chile, thickness of crust in overriding plate = 20 km. About 6% of data fall above upper heat flow limit of 4.0. (B) Raw data for Cascadia. Models use average Cascadia subduction parameters: plate age = 8.5 Ma, convergence rate = 3.5 cm/yr, subducting plate geometry = central Chile, thickness of crust in overriding plate = 35 km. About 33% of data fall above upper heat flow limit of 4.0. Thin solid line is from ref. 31; thin dashed line is from refs. 32 and 33. (C) All data between normalized distances of 0.02 and 1.0 for well-sampled subduction zones, randomly sampled in proportion to trench length. Arrow indicates region of anomalously low heat flow. Orange and blue medians show two random samples of the same data. Models use global average subduction parameters. About 5% of data fall above upper heat flow limit of 4.0.

As a complement to the global heat flow database, we also compiled estimates of the coefficient of friction in specific subduction zones. Last, we calculated densities for typical subducted rocks along three representative cold, intermediate, and hot P–T trajectories to identify whether cold vs. hot subduction might preferentially induce exhumation of rocks through buoyancy (Methods and SI Appendix).

Our overall intent is to integrate disparate models and observations—many made by others—to evaluate whether the exhumed metamorphic rock record reflects typical vs. atypical metamorphic P–T conditions expected in modern subduction environments. In this context, we hypothesize that integration of rock P–T conditions, geophysical observations of heat flow and rock strength, and thermal modeling reconcile thermal models with petrologic observations. We emphasize temperatures along the subduction interface (the boundary between the subducting slab and overlying plate) because it is thought to represent the region of maximum geothermal exchange, and to be the source of exhumed metamorphic rocks. We refer to the temperature distribution along this interface as the slab-top geotherm.

**Context of Thermal Models and the Petrologic Record**

Numerous studies have investigated key components of subduction zones that must be mutually integrated to evaluate rock vs. model P–T conditions. These include compilations of heat flow and estimated coefficients of friction; compilations of the peak P–T conditions of exhumed, subduction zone metamorphic rocks; calculations of rock density; experimental determination of rock strengths; and numerical and analytical thermal models of subduction zones. Much previous work addresses these topics. Our contribution is to expand and update the global heat flow dataset across current subduction systems compared with ref. 25, which provides constraints on realistic coefficients of friction; to calculate subduction zone thermal structure and rock densities along specified P–T paths using parameters that we argue more accurately represent typical thermal conditions along the top of the slab; and to critically review previous interpretations in the context of revised coefficients of friction.

**Heat Flow and Friction.** Heat flow measurements are typically unevenly distributed (SI Appendix, Fig. S1) and scattered (e.g., Fig. 1), so many studies focus on specific subduction zones with large numbers of measurements (see SI Appendix for data sources; Fig. 2). This approach elucidates some processes well, but also emphasizes a few subduction zones to the exclusion of others. Frictional heating is commonly calculated from a friction coefficient, which is multiplied by normal stress to determine shear stress. Intrinsic ($\mu$) vs. apparent ($\mu^*$) coefficients of friction are determined from mechanical experiments on rocks vs. geophysical data (e.g., heat flow). $\mu^*$ is lower than $\mu$ because other factors reduce shear strength, such as high pore fluid pressure or rapid shear rate.

**P–T Compilations.** For comparison with models, identification of typical P–T distributions from exhumed subduction zone rocks has been based on metamorphic facies (12, 13) (see also reviews of refs. 26 and 27 and recent work of ref. 28) and thermobarometry (14–16). Quantitative P–T conditions fall toward the hotter side of the fields that delimit blueschist and eclogite facies (Fig. 3). For example, the PD15 dataset shows that P–T conditions for blueschist-facies rocks are on average $\sim$100 °C higher than the midpoint of the blueschist-facies field (Fig. 3).

**Fig. 2.** Published values of $\mu^*$ (SI Appendix), with mean $\pm$ 2 SE (blue band) and median $\pm$ 2 $\mu^*$ (red band) as calculated in this study.
Rock strength depends most on the weakest
Thermal models generally fall between two
0.10 at subduction zone temperatures. Shear weakening
\geq \mu \geq 0.050 \geq 0.00 = 0.00 implies low-T
plus low-T models (34) with shear heating (solid lines). Dashed lines and double arrows
endmember types. Kinematic models fix subduction geometries,
and highly variable input parameters complicate isolation of key
factors to explain observations.

In this study, we use analytical models because they elucidate
key parameters more simply than other types of models, espe-
cially the effects of shear heating. In that sense, they parallel
earlier studies (13, 21, 24), but we compare them to more
comprehensive datasets and consider combinations of para-
ters more representative of subduction zones (11). Thus, while
our models do not supersede previous studies, they provide more
focused examination of likely parameters in the context of
quantitative data that have been compiled subsequently. We
particularly emphasize that although many thermal models in-
clude shear heating (e.g., refs. 13, 21, 24, 26, and 33–36), many
recent calulations of mineralogy, petrology, and geochemistry
(2, 4, 9, 37) are either directly based on models that do not in-
clude shear heating or reference frictionless models as prima
facie standards. If shear heating is important, as we argue (and as
have several modelers), a large fraction of published geochemical
and petrologic interpretations ranging from metamorphism and
fluid release to arc volcanism must be reevaluated. Compilations
of modern geophysical and geometric data (11), heat flow measure-
ments (refs. 38 and 39 and Fig. 1), and the PD15 dataset afford the
opportunity to reassess model accuracy with its implications for
petrology, geophysics, and geochemistry.

Metamorphic Rock Density. Seminal calculations of rock density
relevant to subduction zones (14, 18) were based on P–T fields
Corresponding with metamorphic facies, assuming an idealized
chemical system or distribution of rocks across different P–T
fields. These calculations do not account for continuous changes
in mineral assemblage and mineral chemistry, but they do identify
the most important shifts in density associated with (nearly) dis-
continuous reactions. In general, for a specified pressure, meta-
morphosed basalt is denser at higher temperatures, while hydrated
peridotite shows a dramatic increase in density (∼0.25 g/cm³) at
~500 °C (14).

Rock Strength. Rock strength depends most on the weakest
interconnected minerals. Rock mechanics measurements on
weak sheet silicates ranging from talc to serpentine (40) ex-
trapolated to geologic strain rates imply coefficients of friction
(μ) ≥ 0.10 at subduction zone temperatures. Shear weakening
can occur at seismic slip rates due to transient phenomena, al-
though coefficients of friction are still typically ≥0.15 (41).

Thermal Models. Thermal models generally fall between two
different endmember types. Kinematic models fix subduction geometries,
Fig. 4. Calculated density along representative hot, moderate, and cold P-T paths (Inset) for hydrated mid-ocean ridge basalt (MORB) and peridotite. Thin vertical lines bound regions of maximum differences between commonly accepted frictionless models and rock P-T conditions. Densities are always greatest for hot model and lowest for cold model.

close to the trench, then a rise toward the arc. For raw data (Fig. 1A) the rise is considerably steeper than for Cascadia (Fig. 1B) or for trench-length–normalized data (Fig. 1C). Cascadia shows high heat flow anomalies seaward of the trench associated with hydrothermal systems (e.g., ref. 43) and especially low normalized heat flow (<0.5) landward (Fig. 1B). Data that have been trench-length–normalized (Fig. 1C) show unusually low values close to the trench. Slabs and overriding plates seismically couple to 50–80-km depth (ref. 44; i.e., up to normalized distances ≥0.7 for an average subduction geometry), and viscously couple to drive mantle wedge convection deeper than ∼80-km depth (33).

Relative to analytical models of slab-top geotherms, the magnitude of the heat flow minimum and rate of rise for the raw data imply an apparent coefficient of friction (μ*) ≥ 0.1 (Fig. 1A). A similar comparison for Cascadia implies a lower coefficient of friction of ∼0.05 (Fig. 1B), but models are insensitive to μ* (note close spacing of model curves) and sensitive to the assumed heat flux of the incoming plate and subduction geometry. Fully parameterized models (e.g., ref. 34) may provide better estimates of μ* for Cascadia (Fig. 2). Trench-length–normalized data cannot be fit well for normalized distances of up to 0.2 but a μ* value of ∼0.035–0.06 would fit most of the data for normalized distances up to 0.5. With the caveat that the analytical models are highly simplified, we take a value of ∼0.05 as best representing the data. Compilation of published coefficients of friction for specific subduction zones (see SI Appendix for data sources) implies similar ranges of 0.02–0.13 (Fig. 2), with mean and median values of 0.055 ± 0.013 (2 SE) and 0.062 ± 0.021 (2σm), respectively. Values of 0.05–0.06 for μ* are higher than typically assumed in previous thermal models, but far lower than values for μ of 0.2–0.5 determined from deformation experiments of weak sheet silicates (40).

Models. Analytical models show that the slab-top geotherm is most sensitive to μ*, slab angle, and temperature of thermal weakening (SI Appendix, Figs. S3 and S4) consistent with previous studies (21). Models that combine realistic bounds for these parameters have likely slab-top geotherms that span the PD15 dataset (Fig. 3 and SI Appendix, Figs. S3 and S4). In fact, a geotherm that closely parallels the PD15 slab-top geotherm results from using a value for μ* of 0.05, comparable to estimates from global heat flow (Fig. 1C) and studies of specific subduction zones (Fig. 2); an average subduction rate and age of subducting plate as estimated from global compilations [6 cm/yr, 50 Ma (11)]; a typical geometry (central Chile); and a moderate temperature of thermal weakening (400 °C; Fig. 3A).

Recent thermal models that include shear heating (34) predict slab-top geotherms that parallel ours for low pressures but steepen considerably at a temperature of ∼400 °C (Fig. 3C). Consequently, these geotherms overlap the lower-T and lower-P portion of the PD15 dataset, but fall toward the colder side of the distribution at higher pressures. This behavior reflects a relatively low temperature for the onset of thermal weakening of 300 °C. This temperature is the minimum bound that others have considered (24) (as do we: See SI Appendix, Fig. S4). Different numerical models that are parameterized differently, especially such that shear heating was included (34) vs. excluded (11), show quite different slab-top temperatures. Models that include shear heating show much closer correspondence with PD15, at least at low pressure and low temperature (Fig. 3C).

Rock Density. Density increases with pressure (and temperature; Fig. 4), with distinct step-ups as dehydration reactions are crossed. Rocks that follow higher-temperature P-T paths consistently show higher densities. The disparity is greatest at a pressure of ∼1.75 GPa (∼60-km depth), where the densities of high-temperature metabasalt and hydrated peridotite exceed those of their low-temperature counterparts by 0.25–0.55 g/cm³. Disparities are evident between 1.25 and 1.75 GPa for cold/ moderate- vs. hot P-T paths but disappear at ∼2.5 GPa where P–T paths converge.

Discussion

Heat Flow and Friction Are Relatively High. Previous regional studies (SI Appendix) and our compilation are congruent in demonstrating that heat flow in the forearc is higher than can be explained by μ* ≤ 0.02. The negative heat flow anomaly close to the trench possibly reflects hydrothermal cooling of the oceanic plate (23, 45) rather than an absence of friction. Both the global datasets (Fig. 1C) and specific studies (Fig. 2) converge on average and median values for μ* of 0.04–0.065, which exceed more typically assumed values of 0.0–0.03 (e.g., refs. 9–11, 31, and 33). Provisional values of average μ* = 0.05 ± 0.015 and minimum μ* = 0.03 satisfy observations and may be preferred for future modeling, at least until more comprehensive analysis is undertaken that eliminates biases on a case-by-case basis. The disparity between coefficients of friction determined from natural data (low μ*) vs. experiments (high μ) may reflect high pore fluid pressures or dynamic weakening at seismic slip rates, due to flash melting, pressurization of pore fluids, formation of weak silica gel, or loss of grain-to-grain contact (see summary of ref. 41).

Shear Heating Reconciles Thermal Models with Metamorphic P–T Conditions. The difference in predicted temperature between models that include shear heating (34) (Fig. 3) vs. those that omit shear heating (9–11) is especially striking: even the smallest proposed coefficient of friction (0.02–0.03) (34) raises slab-top temperatures by ≥100 °C at pressures of 1.5–2.0 GPa and permits predicted slab-top geotherms to intersect a substantial portion of the PD15 data set. Use of our preferred value for μ* of 0.05 repositions average slab-top geotherms from near the boundary of the forbidden zone to the average PD15 geotherm (Fig. 3A). On a case-by-case basis, numerical models that include vs. exclude shear heating show temperature differences of 100–250 °C (Fig. 3C), although other modeling parameters besides μ* also differ and could contribute to the difference. Overall, a broad range of friction coefficients inferred from forearc heat flow implies that a range of slab-top geotherms comparable to natural data are possible, and arguably likely. Many numerical models that predict metamorphic facies distributions have
assumed $\mu^*$ values as low as 0.00–0.02 (e.g., refs. 9, 13, 24, 26–28, and 31), although values of 0.03–0.05 have also been considered (46, 47). A recent study (34) did not assume a value of $\mu^*$, rather derived values over a range of subduction zones based on heat flow, but did not evaluate the implications for metamorphic conditions. While many of these studies have presented $P$–$T$ distributions of slab-top geotherms, comparison with the rock record has been lacking. Fig. 3 represents a comparison between quantitative thermobarometric $P$–$T$ conditions and models that include $\mu^*$ over a range of values consistent with heat flow. Overall, models that include shear heating (Fig. 3A and B) are much more consistent with the PD15 dataset. Inasmuch as shear heating is inevitable, this correspondence implies that petrologic data are not restricted to subduction of anomalously young lithosphere or to subduction initiation (10). Consequently, we recommend that calculations that require a typical slab-top geotherm use the average PD15 geotherm ($T$ in °C)

$$ P (\text{GPa}) = 1.533 \times 10^{-3} \cdot T + 5.07 \times 10^{-6} \cdot T^3. $$  \[[1]\]

In that context, the ±2σ bounds on the PD15 distribution (Fig. 3) reflect the range of expected subduction conditions. Data and geotherms outside those bounds may be viewed as unusual. In reference to disparities between PD15 and many thermal models, other sources of heat along the subduction interface are possible, for example heat advection and production via rock convection, fluid flow, and hydration reactions (see summary of ref. 16). Nonetheless, shear heating provides a simple explanation to reconcile differences between rock $P$–$T$ conditions and thermal model predictions.

**Exhumation of Subduction Zone Rocks.** Recent models have been used to argue that buoyancy favors rock exhumation from subduction of extremely young crust or subduction initiation (10), i.e., hot subduction settings. However, from a petrologic perspective (14), higher temperatures enhance densification compared with lower temperatures. All slabs undergo hydrothermal alteration at the ridge axis, so arguably they should all have similar water content upon subduction. If so, hotter subduction zones should be less amenable to buoyant exhumation because they densify through eclogitization and deserpentinitization at a lower temperature (Fig. 4). For example, in a hot subduction zone, at a depth of ~60 km (1.75 GPa), metabasalt and hydrated metamorphic peridotite are ~10% and ~15% denser, respectively, than in a cold subduction zone. Reaction kinetics can also retard eclogitization of drier plutonic rocks (e.g., ref. 48), but insofar as higher temperatures speed kinetics, transformation of anhydrous oceanic crust to eclogite should occur at shallower levels in hot subduction zones than in cold subduction zones. That is, hotter subduction zones may well have denser rocks throughout the oceanic crust—not only in hydrated metabasalts and metaperidotite but also in anhydrous gabbros—and their rocks may be less amenable to exhumation than in cold subduction zones.

**Implications for Rock Strength and Temperatures of Thermal Weakening.** A temperature of 300 °C marks a transition from brittle to plastic deformation behavior in quartz at low geologic strain rates [$10^{-15}$ s$^{-1}$ (49)], and is commonly assumed to represent conditions of thermal weakening. Similarly, field studies of quartz microstructures in mylonites often return lower low differential stresses of only 10–20 MPa (e.g., ref. 50). For a typical subduction zone geotherm of ~10 °C/km and a coefficient of friction of 0.05, these values imply a transition from brittle to ductile behavior at depths of 20–40 km (~1.0 GPa; see ref. 50), which is too shallow to explain the large number of rock $P$–$T$ conditions at 1.5–2.0 GPa and 500–600 °C. In part these discrepancies may reflect strain rate. For a subduction rate of 6 cm/y distributed over a 1-km-thick shear zone, the strain rate is very high, ~$2 \times 10^{-12}$ s$^{-1}$. At high strain rate, rocks can sustain greater shear stresses and produce more heat, although increasing water fugacity at higher pressures mitigates this effect. Regardless, to control the strength of the subduction interface, weak quartzites would have to occur as continuous sheets. Bending-related horst-and-graben formation likely disrupts the thin sedimentary carapace of the oceanic crust, while trench turbidite infill is clay-rich and discontinuous both spatially and temporally. Both factors make strength continuity unlikely.

**Implications for Earthquakes and Arc Volcanism.** Seismic and geodetic networks define the seismogenic zone as a region where interplate locking promotes elastic energy buildup extending to 40–60-km depth (7). This region grades into the interplate fault segment that moves by slow-slip phenomena for 20–40 km down-dip (51). Intermediate-depth intraslab seismicity extends another 200–300-km down-dip. Thermally controlled processes possibly trigger the form of failure (5–7), but the uncertainties described above have hindered definitive identification of rocks containing the mineral assemblages and (micro)structures required to identify the metamorphic facies, rheological properties, and fault mechanics of each of the different-depth slip phenomena. In light of revised $\mu^*$, generally hotter models should be considered to identify the deformation phenomena at play.

Although we view the thermal structure of subduction zones above 80-km depth as considerably hotter than many recent models, inferences of the mineralogical drivers of arc volcanism remain robust. Because full mantle wedge convection at depths ≥80 km (~2.5 GPa) controls heat budgets, the slab-top thermal structure of the deeper portions of subduction systems in these recent models (10, 11, 31) should converge with the $P$–$T$ conditions of rocks that record pressures of ~2.5 GPa (16), as observed (Fig. 3). Because it is the deep, not shallow thermal structure that defines hydrous mineral stability at depths ≥80 km, dehydration of hydrous peridotites still could catalyze arc formation (4).

**Methods**

We obtained heat flow data (39) from W. Gosnold and limited data to within 350 km of subduction zone plate boundaries (52). We added data from Tonga/Kermadec (53), which were otherwise missing. We did not intend to seek out every heat flow measurement, but rather to compile sufficient data to gain insights into heat flow distributions. We eliminated ~1,000 measurements that were from extensional zones, and separated Cascadia (~4,000 measurements) from other subduction zones (~5,000 measurements). We accounted for the effect of plate age (11) on heat flow by normalizing heat flow data set to the expected heat flow for the age of the subducting plate (25). We did not account for convergence rate because modeling shows that it does not have a major effect on thermal structure (SI Appendix, Figs. S3 and S4). Distances were normalized for each datum by ratioing the distance perpendicular to the trench with the arc-trench distance (11). In this scheme, a heat flow value of 1.0 represents the expected heat flow for the age of the subducting plate, a distance of 8.0 represents the position of the trench, and a distance of ~1.0 represents the position of the arc.

Data are scattered (Fig. 1), so we binned data and used medians. Errors are two times the median absolute deviation divided by the square root of the number of data in the bin. To address data bias, we randomly subsampled data between the arc and the trench (normalized distances between 0.02 and 1.0) such that the number of data points was proportional to the length of the subduction zone trench. Our choice of a lower bound of 0.02 reflects slight disparities between the tabulated positions of the plate boundary (52) vs. the topographically identified position of the trench. Subsampling allowed us to include observations from both high- (e.g., Cascadia) and sparsely- (e.g., Andes) sampled subduction zones.

Because of its flexibility, we used an analytical model that includes shear heating to model the thermal structure of the subduction interface (Fig. 3 and Figs. S19 and 20). We calculate the slab-top geotherm for a subduction rate of 20 km/My and 466 km because it conveniently corresponds with a pressure of ~2.0 GPa, and because corner flow significantly impacts thermal structure at this level and deeper. The analytical model requires specifying numerous subduction
parameters as described in SI Appendix, most importantly slab geometry, coefficient of friction, and temperature of thermal weakening. Slab geometry was taken from a bathymetric map of Chile, which has an intermediate geometry down to depths of ~100 km of the subduction zones considered in this study (SI Appendix, Fig. 52). For modeling, we assumed coefficients of friction from 0 (frictionless) to 0.1. The upper bound is within error of the highest values that data suggest (Fig. 2), and suffices to illustrate the impact of coefficients of friction on thermal structure. Temperatures of thermal weakening were assumed to range from 300 to 600 °C (24, 42). The effect of shear zone thickness data suggest (Fig. 2), and suffices to illustrate the impact of coefficients (coefficient of friction, and temperature of thermal weakening. Slab geom-


dependent flux of H2O from subducting slabs worldwide.


dependent flux of H2O from subducting slabs worldwide. J. Geophys Res. 109:G03007.
9. van Keken PE, Hacker BR, Syracuse EM, Abers GA (2011) Subduction zone fluid flow. 1. Intermediate-depth earthquakes in subducting slabs linked to metamorphic, dehy-


ACKNOWLEDGMENTS. We thank P. Molnar and S. Roecker for comments prior to submission, W. Gosnold and X. Gao for providing the global heat flow database and the slab-top geotherms from numerical models, L. Spatea for help parsing the global heat flow database, and P. Kelemen, T. Gerya, and four anonymous reviewers whose incisive comments helped improve the manuscript substantially. Support was provided by NSF Grants EAR1419865 and OISE1545903 (to M.J.K.), EAR1447468 and EAR1750674 (to F.S.S.), and the Spanish Ministry of Science under “Acción de Programación Conjunta Internacional” Grant PON-2015-053 (to C.R.R.).
Methods. As described in the main text, research on the Cascadia subduction zone strongly biases the global heat flow database: of the c. 10,000 measurements that we considered (Fig. S1), c. 4000 are from Cascadia. Another c. 1000 measurements are from extensional zones proximal to Colombia and Ecuador (i.e. they are not relevant to this study). Consequently, we eliminated the Colombia – Ecuador data completely, and plotted Cascadia separately of other subduction zones for comparison. Each subducting slab has a different age, convergence rate and dip angle, but overall the global-minus-Cascadia dataset of c.5000 measurements over-represents young slab (≤25 Ma) subduction zones (17%) compared to the global distribution [13%; (1)].

Slab-top temperatures and surface heat flow were calculated using a simple spreadsheet, based on analytical solutions to the heat flow equation for planar (2) and curved (3) geometries. This approach was used because it is computationally simple, can readily accommodate a large range of input parameters, and is sufficient to illustrate the effect of shear heating on subduction zone thermal structure. We performed calculations to a depth of 66 km (P=2.0 GPa) to illustrate better the effects of thermal weakening, but temperatures become increasingly inaccurate (too low) at pressures of c. 1.5 GPa because mantle convection increasingly affects temperatures at higher pressures. Shear heating may also be important for understanding deeper earthquakes (4), but these levels are not especially relevant to our interests. We chose not to use a community benchmark numerical model (5) that has been used in other petrologic studies (6, 7) because it does not account for shear heating or the progressive decrease in shear heating due to thermal weakening, and it assumes an unusually shallow depth for seismic decoupling [50 km vs. 80 km; (8)]

These analytical solutions require specifying a subduction geometry, basal heat flux (Q_b), convergence velocity (1-12 cm/yr), effective coefficient of friction (μ* = 0.0 to 0.1), thickness of the overlying crust (20 km), radiogenic heat production (1μW/m³) in the overlying crust, thermal conductivity of the mantle (3.1 W/m·K), thermal conductivity of the crust (2.5 W/m·K), thermal diffusivity (1x10⁻⁶ m²/s), and conditions of thermal weakening of the subducting slab. These choices reflect typical values long used for thermal modeling (9). Our reference subducting slab geometry corresponds with central Chile (angles from 10 to 26°), which is close to, albeit somewhat shallower than, the median of the global range of subducting slab geometries (14-16) (Fig. S2). We tested the sensitivity of models to variability in numerous parameters, but basal heat flux was always based on the age of the subducting plate (10). Heat flow for young oceanic crust, especially ≤10 Ma, is lower than predicted from thermal models (11), which likely reflects the effects of hydrothermal cooling at the ridge axis (12, 13). In that context, the parameterization of heat flow for young modern plates (10) probably includes effects of hydrothermal cooling, and accurately represents basal heat flux into the subduction interface for short times. But over timescales of many Ma, continued thermal relaxation of the subducting plate might increase basal heat flux. Any modeling of such complexities would necessarily be arbitrary, as the prior thermal structure of the lithosphere would have to be specified. But in general, our models for younger subducting lithosphere may underestimate temperatures, so that the modeled effect of plate age on thermal structure is smaller than may occur in nature (Fig. S3-S4). For the overriding crust, we also assume a constant thickness (20 km) and heat production
(1 μW/m²), but slab-top geotherms are highly insensitive to these parameters. Variations in geometry of curved surfaces were tested simply by stretching or compressing distances by factors ranging from 1/3 angle (3-9°) to 1.5x angle (15-39°). For slab-top geotherms with planar surfaces, dips ranged from 2 to 25°. Reported coefficients of friction range from 0.02 to 0.13 (main text Fig. 2); we assumed a minimum µ* reference model of µ* = 0.025, but explored the implications of µ* = 0.0 to 0.1; we chose the upper bound somewhat arbitrarily as reflective of the highest coefficients of friction within error for the Hikurangi, Kuriles, and Kamchatka subduction zones. For thermal weakening, we assumed a critical temperature (T_{crit}) that ranged from 300 °C to 600 °C, reflecting weakening of wet quartz-rich vs. mafic bulk compositions (17), as well as estimates of the range of earthquake depths in the interiors of oceanic plates (18). Reduction in strength was assumed to follow an exponential dependence on temperature (9), with an e-folding ΔT of 60 °C, consistent with typical activation energies for rock strength (9).

The sensitivity of heat flow to variations in input parameters (Fig. S5) shows that heat flow is highly sensitive to shear heating, moderately sensitive to convergence rate and angle of subduction, and relatively insensitive to plate age (albeit with caveats relating to prior hydrothermal cooling).

Estimates of µ* from the literature for 9 specific subduction zones were based on (3, 19-28). These values are not exhaustive, but serve to illustrate ranges of values.

For calculating densities, we used a standard thermodynamic modeling approach (29, 30) with the HP662 thermodynamic dataset to calculate mineral abundances and densities along prescribed P-T paths (main text Fig 4). These simplified P-T paths were based on #1: the PD15 average rock geotherm (31), #2: a numerical model passing through the middle of the predictions of (20), and #3: an analytical model that closely matches one of the warmest numerically modeled subduction zones (20) (Manila) and with the hottest P-T conditions for rocks that appear to have been exhumed from subduction systems that include old crust subducted at high rates (31). We chose not to use the warmest bound on the PD15 dataset to avoid the risk of biasing results towards subduction of unusually young crust or subduction initiation (16, 32, 33). We focused on representative bulk rock compositions for MORB (34) and mantle peridotite (35). These compositions, in wt%, are:

MORB:
\[ Na_2O = 2.71 \]
\[ MgO = 7.84 \]
\[ Al_2O_3 = 15.47 \]
\[ SiO_2 = 50.57 \]
\[ K_2O = 0.18 \]
\[ CaO = 11.52 \]
\[ TiO_2 = 1.43 \]
\[ FeO = 9.35 \]

Peridotite:
\[ MgO = 41.4 \]
\[ Al_2O_3 = 2.27 \]
\[ SiO_2 = 44.0 \]
We used the following solution models:
Olivine: O(JH)
Cpx: Cpx(JH)
Clinoamphibole: cAmph(G)
Chlorite: Chl(LWV)
Spinel: Sp(JH)
Garnet: Grt(JH)
Anthophyllite: Anth
Clinohumite: Chum
Antigorite: Atg(PN)
Brucite: B
Talc: T
Biotite: Bi(W)
Omphacite: Omph(GHP)
Di-octahedral mica: Mica(W)
Plagioclase: Pl(JH)
Garnet: Gt(W)
Carpholite: Carp(SGH)
Stilpnomelane: Stlp
Ilmenite: Ilm(WPH0)
Pumpellyite: Pu

Model-model and model-data comparisons. In comparing similarly parameterized numerical (20) and analytical (this study) models, both predict similar temperatures to within a few km, but direct comparison is imperfect for two reasons. First, the real trench is many km below Earth’s surface, whereas the analytical solution assumes the “trench” is at the surface. Consequently, our models are slightly hotter over the first c. 50 °C, which changes the curvature slightly. However, deeper P-T conditions are comparable. Second, the numerical models include lateral heat flow due to forced mantle convection (corner flow) that is ignored in the analytical models. Corner flow begins to affect temperatures at depths of 40-50 km (36), and dominates at depths >65 km, so we do not consider results quantitatively at greater depths. The sensitivity of thermal models to input parameters (Figs. S3-5) shows that slow subduction of young crust results in high slab-top geotherms, similar to previous results (37), in part because slow subduction is less effective at advecting heat, so the slab and overlying mantle remain relatively warm. Conversely, in models with moderate $\mu^*$, subduction rate makes little difference because shear heating is proportional to convergence velocity, whereas diffusive heating (juxtaposition of subducting slab against hotter mantle) is inversely proportional. Similarly, the age of the subducting crust affects geotherms far less because total heat flux includes shear heating, and differences in basal heat flux associated with different crustal ages affect temperatures proportionately less.

Geophysical and petrologic discussion. Several geophysical observations provide useful perspective in interpreting subduction thermal structure. Clearly, in reference to heat flow across Earth, low fore-arc heat flow reflects subduction of cold oceanic lithosphere and has been
interpreted to imply no shear heating (an effective coefficient of friction, \( \mu^* \), of zero) along the subduction interface (16). However, for non-dimensional distances of 0.1 to 0.5, global fore-arc heat flow is either constant or increases towards the arc (main text Fig. 1), and has long been interpreted to reflect non-zero \( \mu^* \) (3, 19-28) (main text Figs. 1-2). Our work follows the logical implications of these latter interpretations. Low seismic attenuation (high stiffness) has been interpreted to reflect extremely cold conditions in the fore-arc mantle (38), but in well-studied systems, such as in northern Japan, low attenuation is not uniformly distributed, and instead exhibits complex geometries and attenuation values (39). In addition, the seismic velocities of serpentine along the two crystallographic axes of the basal plane are comparable to or even higher than different directions in olivine (40). Thus, discussions of velocity and attenuation are incomplete without reference to seismic transport directions and seismic anisotropy. Many studies have inferred large quantities of serpentine in the fore-arc mantle to explain not only seismic velocities (41), but also seismic anisotropy (39, 41).

Petrologic observations provide additional insight. Metamorphic P-T conditions and paths imply much warmer conditions (on average 10 °C/km) than common models (4-5 °C/km), with a large majority of paleodepths of exhumed rocks corresponding to P<2.5 GPa (<80 km depth) (31). Some have argued that unusually hot conditions and correspondingly greater buoyancy favor exhumation of subduction zone rocks (33), i.e. the petrologic record is substantially or perhaps even wholly biased to subduction of young oceanic crust or subduction initiation settings. This interpretation was considered and discounted in (31) based on tectonic histories of ancient subduction systems (42), and warmer rocks from the subducting slab are generally denser, not more buoyant, than colder rocks (43) (main text Fig. 4). We acknowledge that some bias to the rock dataset is possible, and this is why we chose not to use the warmest bound on the compilation of rock P-T conditions to explore magnitudes of shear heating. However, the present study suggests that the main discrepancy between rocks and (several) models does not reflect warm subduction, but rather incomplete parameterization of some numerical models.

Studies of mechanical behavior of minerals imply an intrinsic coefficient of friction (\( \mu \)) greater than ~0.10. Talc is the weakest of sheet silicates (44) and arguably could control the frictional behavior of subduction zones at the subduction interface (45) at depths typically >30-40 km. Hydrous experiments at 100-400 °C and confining pressures >0.1 GPa (46) imply \( \mu \geq 0.12 \) for displacement rates of 32 cm/yr. Insofar as talc maintains its strength with increasing temperature and that \( \mu \) decreases by ~0.02 for each order of magnitude decrease in shearing rate (46), a shearing rate of 3.2 cm/yr implies \( \mu \geq 0.10 \). Water-saturated experiments on serpentine, micas, and mineralogically heterogeneous rocks imply higher coefficients of friction, typically 0.2 to 0.5 (44). Thus, natural systems must experience a reduction in the coefficient of friction compared to the intrinsic rock strength at low pore fluid pressure.

**Effect of shear zone thickness.** In our models (and many others), shearing is assumed to occur along an infinitely thin planar interface. Shear zones must occupy some thickness, so to evaluate the effects of shear zone thickness on temperature distributions, we solved the heat flow equation analytically to describe temperatures across a shear zone of arbitrary thickness. Solutions show that bias to models is very small for likely shear zones ranging in thickness from hundreds of meters to kilometers.

We first integrate the heat flow equation:
\[
k \frac{dT}{dz} = -Q(z) \tag{Eq. S1}
\]

where \( k \) is thermal conductivity, \( T \) is temperature, \( z \) is depth (\( z_o \) at the base of the shear zone, \( z_o-d \) at the top of the shear zone, for a shear zone thickness of \( d \)), and \( Q \) is heat flux through a surface of unit area of interest. Integration of Eq. S1 for a point at depth \( z \) within the shear zone yields:

\[
T(z) = T(z_o) - \frac{1}{k} \int_z^{z_o} Q(z) dz \tag{Eq. S2}
\]

where \( T(z_o) \) is the temperature at the base of the shear zone, and \( 0 \leq z_o - z \leq d \). The following solution to Eq. S2 results if \( Q \) depends linearly on depth (i.e., \( Q = Q_o + \Delta Q \cdot (z_o - z) / d \), where \( Q_o \) is basal heat flux, and \( \Delta Q \) is the increase in heat flux over distance \( d \):

\[
T(z) = T(z_o) - \frac{Q_o}{k} (z_o - z) - \frac{\Delta Q}{zd} (z_o - z)^2 \tag{Eq. S3}
\]

We compared temperature differences between an infinitely thin shear zone vs. a shear zone whose thickness ranges from \( d = 500 \) to \( 5000 \) m, where the infinitely thin shear zone is positioned in the center of the thick shear zone (i.e. at \( d/2 \)). We assumed \( Q_o \) and \( \Delta Q \) values that were reflective of our models at a depth of \( \sim 40 \) km, where the heat contribution associated with the incoming slab is \( \approx 0.01 \) W/m\(^2\) and heat flux from shear heating is \( 0.02 \) mW/m\(^2\). Note that it might at first appear that heat flow should be higher because the incoming plate has higher heat flow. In our reference model, basal heat flux is \( 0.062 \) mW/m\(^2\), while the contribution of shear heating is the coefficient of friction times the shear rate, which can be much larger. But for the Molnar and England model (2), the temperature at the subduction interface is calculated from the heat flux divided by the subduction parameter \( S \). At a depth of \( 40 \) km, \( S \) is \( \sim 6.0 \), so calculated heat flux is very low, e.g. the incoming basal heat flux of \( 0.062 \) W/m\(^2\) is reduced to \( \sim 0.01 \) W/m\(^2\). For these parameters, the deviation between an infinitely thin shear zone vs. a distributed shear zone is small: \( 0.6 \) °C for a \( 500 \) m-thick shear zone, and \( 6 \) °C for a \( 5 \) km-thick shear zone (Fig. S6). It is important, too, that the temperatures at the top and the bottom of the shear zone are identical, so the only differences in model temperatures occur within the shear zone itself.

Because these temperature differences are so small, we chose to ignore contributions of shear zone thickness to thermal structure.
Table S1. Parameters used in modeling [principally from (20)]

<table>
<thead>
<tr>
<th>Age, t (Ma)</th>
<th>$Q_b$ (W/m$^2$)</th>
<th>$V$ (cm/yr)</th>
<th>$\mu^*$ (dimensionless)</th>
<th>$\rho_{crust}$ (kg/m$^3$)</th>
<th>$\rho_{mantle}$ (kg/m$^3$)</th>
<th>d (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10-200</td>
<td>0.047-0.090</td>
<td>1-12</td>
<td>0.0 – 0.1</td>
<td>2750</td>
<td>3300</td>
<td>20000</td>
</tr>
<tr>
<td>A (W/m$^3$)</td>
<td>$k_c$ (W/m·K)</td>
<td>$k_m$ (W/m·K)</td>
<td>$\kappa$ (m$^2$/s)</td>
<td>$T_{crit}$ (°C)</td>
<td>$\Delta T$</td>
<td>$z$ (m)</td>
</tr>
<tr>
<td>1x10$^{-6}$</td>
<td>2.5</td>
<td>3.1</td>
<td>1x10$^{-6}$</td>
<td>300-600</td>
<td>60</td>
<td>0-6600</td>
</tr>
</tbody>
</table>

$Q_b$ = basal heat flux (derived from age of oceanic plate)
= (67-0.1·t)/1000 for $t \geq 27.1$ Ma;
= (105-1.5·t)/1000 for $27.1$ Ma > $t > 7.3$ Ma

$V$ = convergence velocity
$\mu^*$, $\mu^*_{BP}$ = coefficient of friction, coefficient of friction at the brittle-plastic transition
$\rho_{crust}$ = density of the crust
$\rho_{mantle}$ = density of the mantle
d = thickness of overriding crust
A = heat production in overriding crust
$k_c$ = thermal conductivity of the crust
$k_m$ = thermal conductivity of the mantle
$\kappa$ = thermal diffusivity
$T_{crit}$ = critical temperature (temperature of the onset of thermal weakening)
$\Delta T$ = exponential constant describing the decrease in coefficient of friction with increasing $T$
$z$ = depth to the subduction interface

Equation describing thermal weakening (Peacock et al., 1994):
At temperatures above $T_{crit}$:
$$\mu^* = \mu^*_{BP} e^{-\frac{(T-T_{crit})}{\Delta T}}$$

Equations describing temperature along the subduction interface (3):
For $z > d$, temperature at the subduction interface is given by:
$$T = \frac{Ad^2 + \sigma V + Q_b}{2k_c} \cdot \left[ z + \left( \frac{k_m}{k_c} - 1 \right) \cdot d \right]$$

where $Q_b$ is positive [unlike the published sign convention (3)] and $\sigma$ is given by:
$$\sigma = \mu \cdot P$$
where $P$ is pressure, and $S$ is given by:
$$S = 1 + \frac{b}{\sqrt{kt}} \cdot \left[ z + \left( \frac{k_m}{k_c} - 1 \right) \cdot d \right]$$

where $b \approx 1$, and $t'$ is the time for the oceanic lithosphere to move from the trench to the depth of interest.
For a planar geometry, time is easily determined trigonometrically from depth, but for curved interfaces it must be determined incrementally along the subduction interface.

For $z \leq d$, temperature at the subduction interface is given by
$$T = \frac{\frac{z}{k_c} \cdot \left[ \frac{(Az_f)}{2} + \sigma V + Q_b \right]}{S} \cdot d$$

where $S$ is given by:
$$S = 1 + \frac{b}{\sqrt{kt}} \cdot \frac{k_m}{k_c} \cdot z$$
**Table S2. Subduction interface geometry**

<p>| | | | | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>x (m)</td>
<td>z (m)</td>
<td>x (m)</td>
<td>z (m)</td>
<td>x (m)</td>
<td>z (m)</td>
</tr>
<tr>
<td>1</td>
<td>1</td>
<td>106500</td>
<td>24000</td>
<td>157000</td>
<td>46000</td>
</tr>
<tr>
<td>13500</td>
<td>2000</td>
<td>112250</td>
<td>26000</td>
<td>161250</td>
<td>48000</td>
</tr>
<tr>
<td>25000</td>
<td>4000</td>
<td>117750</td>
<td>28000</td>
<td>165000</td>
<td>50000</td>
</tr>
<tr>
<td>35750</td>
<td>6000</td>
<td>122000</td>
<td>30000</td>
<td>169000</td>
<td>52000</td>
</tr>
<tr>
<td>47000</td>
<td>8000</td>
<td>126500</td>
<td>32000</td>
<td>173500</td>
<td>54000</td>
</tr>
<tr>
<td>57500</td>
<td>10000</td>
<td>132000</td>
<td>34000</td>
<td>177500</td>
<td>56000</td>
</tr>
<tr>
<td>66000</td>
<td>12000</td>
<td>136000</td>
<td>36000</td>
<td>181750</td>
<td>58000</td>
</tr>
<tr>
<td>74250</td>
<td>14000</td>
<td>140000</td>
<td>38000</td>
<td>186250</td>
<td>60000</td>
</tr>
<tr>
<td>82750</td>
<td>16000</td>
<td>144000</td>
<td>40000</td>
<td>190250</td>
<td>62000</td>
</tr>
<tr>
<td>89000</td>
<td>17750</td>
<td>148000</td>
<td>42000</td>
<td>195000</td>
<td>64000</td>
</tr>
<tr>
<td>95500</td>
<td>20000</td>
<td>152500</td>
<td>44000</td>
<td>199000</td>
<td>66000</td>
</tr>
<tr>
<td>101750</td>
<td>22000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

x is the distance from the trench, and z is the depth to the subduction interface.
Captions for SI Datasets.

SI Dataset 1. Spreadsheets contain the following information:
- Thermal modeling computations
- Trench metadata
- Heat flow data excluding the Cascades
- Heat flow data for the Cascades
- Randomized heat flow data sampled according to trench length
- Heat flow datasets on a trench-by-trench basis
- Sources and values of published apparent coefficient of friction (μ*)

References:


Figure legends

**Fig. S1.** World map showing locations of heat flow data used in this study. Data are culled from (47) based on 300 km proximity to subduction zones.

**Fig. S2.** Average slab dip for subduction systems calculated from the arctangent of 100 km (the approximate depth to the top of the subducting slab beneath the arc) divided by the arc-trench distance. Blue symbols are for slabs considered by (15) and (16). Red dot is for the central Chile subducting slab. An increase in slab dip from 20.7° (Chile) to the median of subduction zones considered (21.8°) does not significantly change our model results or interpretations.

**Fig. S3.** Sensitivity analysis of planar model geotherms without consideration of a critical temperature of weakening (i.e. $T_{\text{crit}}$ is implicitly infinite). Successively holding several parameters constant, the sensitivity of planar models is shown with respect to age of the subducting crust (which affects basal heat flow), convergence rate, coefficient of friction ($\mu^*$), and angle of subduction. Blue field spans P-T conditions of rocks exhumed from subduction zones (PD15 dataset). Coldest and warmest geotherms shown in blue and red. Lavender and pink curves represent average published model (16) and PD15 geotherms. Thick gray lines show bounds on metamorphic facies (48).

**Fig. S4.** Sensitivity analysis of curved model geotherms. Same as Figure S2, but for curved subduction geometry and including thermal weakening ($T_{\text{crit}}$). A composite of different possible geotherms shows that typical values reported in the literature for these parameters span the PD15 dataset.

**Fig. S5.** Calculated heat flow for thermal models corresponding to the top 4 panels in Fig. S4.

**Fig. S6.** Comparison of the analytical solution to thermal structure across a shear zone of arbitrary thickness (blue curves) vs. assumption that shearing occurs along a single plane (orange lines). Thickness of shear zone = d, depth of shear zone base = $Z_0$, temperature at base of shear zone = $T_{Z0}$. (A) Shear zone thickness of 500 m. (B) Shear zone thickness of 5 km.