The Global Range of Temperatures on Convergent Plate Interfaces

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Abstract We present accurate analytical expressions for temperatures on the upper parts of convergent plate boundaries where there are rigid plates both above and below the subduction interface. We expand on earlier formulations, which considered planar interfaces of small dip, to give expressions suitable for use on all present plate interfaces, which have both curved cross sections and maximum dips of up to 30°. We also explain the errors in studies that have asserted the inapplicability of such analytical approximations to temperatures near curved plate boundaries, or where young oceanic lithosphere is subducted. We show, by comparing these expressions with numerical solutions to the full equations, that the approximations agree with the numerical calculations to within a few percent—appreciably smaller than the uncertainties associated with the physical parameters of actual plate interfaces. The common equating of “warm” subduction interfaces with the subduction of young lithosphere, and “cold” with old lithosphere, is not valid. In the absence of dissipation, thermal gradients on the plate interface vary inversely with the product of age of the subduction ocean plate and its descent speed. Where shear stresses during slip on the plate interface exceed a ~10 MPa, the temperature gradients along the interface vary with the product of full convergence rate and shear stress during slip on the interface.

Plain Language Summary  Convergent plate boundaries host most of Earth’s great (magnitude 8 or larger) earthquakes, and conditions on those boundaries determine the degree to which sediments and fluids (in particular H₂O and CO₂) may be carried beyond the base of the plate boundary, where they can influence melting beneath volcanic arcs and may ultimately be transported to the deep mantle. The ability to calculate how temperatures on convergent plate boundaries respond to the parameters of subduction is a necessary precursor to understanding these processes. We present analytical expressions for temperatures on curved plate boundaries, such as those in most modern subduction zones, and show that recent claims of the inaccuracy of analytical solutions in such settings are erroneous. We use our expressions to demonstrate that the common practice of equating “warm” plate boundaries with subduction of young ocean floor, and conversely “cold” with old, is not valid. Temperature gradients along the plate interface depend most importantly upon the convergence rate and the shear stress during slip on the interface. Analytical approximations are to be preferred over numerical models because they provide insight into physical process, and they may be evaluated using pencil and paper.

1. Introduction

Distributions of temperature on subduction interfaces influence many important processes, including seismic and aseismic slip, dehydration of the subducting slab, and transfer of melts and fluids to the volcanic arcs. It is desirable to obtain analytical expressions for those temperatures, because they provide a transparent link between the underlying physical processes and quantities of geological, geophysical, or geochemical interest. Furthermore, such expressions permit examination of parameter space in a fashion that is much more economical than with numerical models.

The purpose of this study is to present analytical expressions for temperatures on curved plate interfaces, representative of present-day subduction zones, that may readily be used by anyone who wishes to calculate the likely range of temperatures in a situation of interest. van Keken et al. (2019) called into question the utility of such analytical approximations, in particular concluding that analytical expressions cannot be used either when subducting oceanic lithosphere is young, nor when the subduction interface is curved. On
the route to our results, we demonstrate that each of those conclusions is invalid. We use our expressions to examine thermal gradients along and beneath plate interfaces in active subduction zones.

1.1. The Plate Interface

We refer to the portion of a subduction zone that is bounded above and below by rigid plates as the “plate interface” (Figure 1). Here, the thermal regime is dominated by diffusion of heat perpendicular to the interface and advection of heat by the relative motion of the plates parallel to the interface (e.g., McKenzie, 1969; Molnar & England, 1990; Turcotte & Schubert, 1973). The plate interface extends from the trench to at least the maximum depth of thrust-faulting earthquakes on it, which varies between \( \sim 30 \) and 60 km (Hayes et al., 2012), but the presence of tremor and silent-slip events (e.g., Beroza & Ide, 2011; Ide, 2012), implies that the plate interface may extend 10–20 km deeper than this.

We begin by determining a representative shape for plate interfaces by fitting the surfaces of descending plates, as determined by Hayes et al. (2018), to a parabolic form

\[ z_f = ax^2, \]  

(1)

where \( z_f \) is the depth of the interface at horizontal distance \( x \) measured perpendicularly from the trench. The profiles used are those of SubMap 4.3 (Heuret & Lallemand, 2005; Heuret et al., 2017), excluding profiles for which convergence velocities are uncertain (principally those involving small plates in the southwest Pacific), or for which the age of ocean floor is unavailable. The profiles of Hayes et al. (2018) are fit to a depth of 60 km. We later make use of a subset of profiles, for which England (2018, Appendix C and Supporting Information) could obtain a reliable maximum depth of thrust faulting, \( z_f \); each of those profiles is fit from the surface to that depth.

As Figure 2a shows, 130 out of 154 of the larger set of profiles, and all 80 of the subset, have an RMS misfit to a parabola of 5 km or less. Those profiles with larger misfit are concentrated in Mexico, and in regions of low slab dip in northern South America. Because uncertainties in depth of slab given by Hayes et al. (2018) comfortably exceed 5 km in the depth range of interest, we do not need to seek a more complex form. Of the 154 profiles in the full set, 138 have curvature in the range \( 5 \times 10^{-4} < a < 3.5 \times 10^{-3} \text{ km}^{-1} \), and all but 6 of the profiles in the subset have curvature in that range (Figure 2b). The cosine of the dip appears in the analytical approximations for temperature; the distribution of its minimum value, at \( z_f \), is shown in Figure 2c, for the subset of profiles on which \( z_f \) has been determined (England, 2018).

Figure 1. Definition sketch of the subduction interface. This study addresses the distribution of temperatures on the plate interface, which separates two plates, each of which transfer heat by diffusion. Temperatures on the wedge-slab interface, which are influenced by advection of heat in the wedge, are not considered here.
2. Analytical Approximations to Temperatures at, Above, and Below the Plate Interface

We consider two plates that transfer heat by conduction, and which are separated by an interface on which dissipation may occur (Figures 1 and A1). Molnar and England (1990) gave approximate expressions for temperatures near planar interfaces which were elaborated, for different conductivities of upper and lower...
plates, by Tichelaar and Ruff (1993) and, for the case in which young ocean floor approaches a trench, by Molnar and England (1995). Our development follows Molnar and England (1990) (Appendix B) and Molnar and England (1995), with the exception that we allow for curved plate interfaces.

Let the speed of convergence between the two plates be $V$, and the angle between the convergence vector and the normal to the trench be $\phi$. Depending upon the application, one may wish to express $u$ either in terms of horizontal distance from a point on the profile to the nearest point on the trench ($x$, Equation 2), or in terms of the horizontal distance along a profile from its origin at the trench ($X$, Equation 3), or in terms of the depth of a point on the profile ($z$, Equation 4):

$$u(x) = \frac{x}{2} \sqrt{4a^2 x^2 + \sec^2(\phi)} + \frac{\sinh^{-1}(2ax \cos(\phi))}{4a \cos^2(\phi)},$$  \hspace{1cm} (2)$$

$$u(X) = \frac{X}{2} \sqrt{4a^2 X^2 \cos^2(\phi) + 1} + \frac{\sinh^{-1}(2aX \cos^2(\phi))}{4a \cos^2(\phi)},$$  \hspace{1cm} (3)$$

$$u(z_f) = \frac{z_f}{1 + \sec^2(\phi)} + \frac{\sinh^{-1}(2az_f \cos(\phi))}{4a \cos^2(\phi)}.$$  \hspace{1cm} (4)$$

Temperatures on the plate interface depend on the Péclet number (Pe), a dimensionless combination of parameters which is the ratio of the time scale for diffusion of heat through the thickness of the upper plate, $t_z = \frac{z_f^2}{\kappa}$, to the time $t_1$, equal to $u / V$, taken for the lower plate to travel the distance $u$ down the interface:

$$\text{Pe} = \frac{t_z}{t_1} = \frac{V z_f^2}{\kappa u}.$$  \hspace{1cm} (5)$$

The distribution of Pe for active subduction segments, evaluated at $z_f$, is shown in Figure 2d.

Molnar and England (1990) showed that, provided a time $t > (t_1 + t_z / \pi^2)$ elapses following the onset of subduction (or a major change in its rate), then steady-state conditions are obtained in which, while the temperatures of material points in the lower plate vary with time as they pass beneath the upper plate, the temperature field in a coordinate system fixed to the upper plate is independent of time. For the range of $t_1$ and $t_z$ relevant to modern subduction zones, that elapsed time is between 10 and 20 Myr, which is smaller, and usually substantially so, than the age of most volcanic arcs at subduction zones (Jarrard, 1986). We therefore concentrate on steady-state solutions. Molnar and England (1990) give expressions that may be used to estimate temperatures on plate interfaces of subduction zones that are younger than $(t_1 + t_z / \pi^2)$.

2.1. Temperatures on the Plate Interface in the Absence of Heat Sources

We neglect radiogenic heat production, which is generally unimportant in the subduction setting but may, if necessary, be treated with the expressions of England (2018). With negligible radiogenic heat production, temperature gradients within the upper plate are independent of depth, and

$$T(z) = T_f \frac{z}{z_f},$$  \hspace{1cm} (6)$$

where $T(z)$ is temperature at depth $z$ and $T_f$ is the temperature on the plate interface. Here, as in the rest of the study, we use the Centigrade scale, and approximate temperature at Earth’s surface to 0°C.

The top of the lower plate heats as it passes beneath the upper plate. Provided that diffusion of heat parallel to the plate interface may be neglected, temperatures in the lower plate are equivalent to those within a half-space whose surface temperature varies with time. In such cases, a temperature perturbation arises at the
The top of the half-space over a distance, perpendicular to the interface, that scales as $\sqrt{\kappa t}$ where $t$ is the time scale over which the surface is heated (Carslaw & Jaeger, 1959; Section 2.5).

One class of temperature variation is particularly useful for the problem under consideration: if the upper surface of a half-space $y > 0$ is maintained at $T = C t^{m/2}$ for $t > 0$, where $C$ is constant and $m$ is a positive integer, then the temperature at a distance $y$ below the surface of the half-space is

$$T(y, t) = C \Gamma(m/2 + 1)[4t]^{(m/2)} i^{m} \text{erfc} \left( \frac{y}{2\sqrt{\kappa t}} \right)$$

where $\Gamma$ is the gamma function and $i^{m} \text{erfc}$ is the repeated integral of the error function (Carslaw & Jaeger, 1959, p 63 and Appendix II). With temperature at Earth’s surface being $\sim 0^\circ C$, we may set $T(0,0) = 0$ and $T(0,t) = T_f(t)$. The heat flux at the surface of the half-space is

$$K_z \frac{\partial T}{\partial y} \bigg|_{y=0} = -b_m \frac{K_T T_f}{\sqrt{\kappa t}},$$

where $b_m = \frac{\Gamma(m/2 + 1)}{\Gamma(m/2 + 1/2)}$ and $K_z$ is the thermal conductivity of the half-space. The analytical expressions for temperatures around the plate interface are obtained by finding $T_f$ such that the difference between the heat flux out of the lower plate and the heat flux into the base of the upper plate is equal to whatever heat flux is generated on the plate interface (Molnar & England, 1990).

We consider first the case in which there is no generation of heat on the plate interface, and denote the heat flux that would have been flowing through the top of the lower plate, had it not been subducted, as $Q$.

We concentrate below on oceanic lithosphere, for which $Q$ depends on its age, but the derivation is general (Molnar & England, 1990) and we do not yet specify a form for $Q$. Equation (8) gives the diminution of heat flux from the top of the lower plate as a result of its having been heated. Equating the vertical heat fluxes either side of the interface gives

$$\frac{K_z T_f}{\zeta_f} \cos(\delta) = Q - b_Q \frac{K_z T_f}{\sqrt{\kappa u / V}},$$

where $\delta$ is the local dip of the interface and we have substituted $t = u/V$.

For the rest of this study, we consider the lower plate to be oceanic lithosphere of age $t_0$ when it enters the trench and of age $(t_0 + t_1)$ when its top is a distance $u (= V t_1)$ from the trench. The heat flux, $Q$, is calculated from the age of the ocean floor using the cooling half-space plate model

$$Q = \frac{K T_i}{\sqrt{\kappa k(t_0 + t_1)}},$$
where $T_i$ is the temperature of the mantle at the oceanic ridges. Differences between heat fluxes calculated from this expression and from plate models are negligible in the present context (e.g., Parsons & Sclater, 1977). However, an important but uncertain adjustment is needed to account for removal of heat from the upper levels of young oceanic lithosphere by hydrothermal circulation (e.g., Lister, 1972; Lowell et al., 1995; Stein & Stein, 1994; Wolery & Sleep, 1976). This process may be allowed for by multiplying the conductive heat flux (Equation 14) by a factor

$$f = 0.5 + (t_0 + t_i)/134 \text{ Myr} \quad t_0 \leq 67 \text{ Myr},$$

(Stein & Stein, 1994, Figure 4).

### 2.2. Temperatures on the Plate Interface in the Presence of Dissipative Heating

We now address the temperatures resulting from dissipative heating on the interface, without any other source of heat. This heating takes place at a rate $\tau' V$, where $\tau'$ is the average shear stress during relative motion across the interface; if that motion takes place primarily in earthquakes, then $\tau'$ may be much lower than the static shear stress (e.g., Rice, 2006). As with Equations 6–13, we equate vertical heat fluxes across the interface, but now—in the absence of other heat sources—the heat flux generated at the interface is the sum of the heat fluxes away from the interface

$$\tau' V = \frac{K_T}{z_f} \cos(\delta) + b \frac{K_T}{\sqrt{\nu u / V}}.$$

(16)

$$T_f = \frac{\tau' V z_f}{K_T S_f}.$$

(17)

$$S_f = \cos(\delta) + b \frac{K_2}{K_1} \sqrt{\frac{V z_f^2}{\nu u}}$$

(18)

$$= \cos(\delta) + b \frac{K_2}{K_1} \sqrt{\text{Pe}}.$$

(19)

### 2.3. Appropriate Values for $b_f$ and $b_s$

Following Molnar and England (1990), we note that the calculated temperatures on the interface in the case of no dissipation increase approximately linearly with time

$$T_f \propto u,$$

(20)

which suggests that, when employing Equations 13 and 19, we should use $m = 2$ and $b = 2 \sqrt{\pi}$. In considering dissipative heating, we shall assume that the effective shear stress, $\tau'$, during slip on plate interfaces is proportional to depth

$$\tau' = \mu' g \rho z_f$$

(21)

where $g$ is the acceleration due to gravity, $\rho$ is the average density of the plate above $z_f$ and $\mu'$ is an effective coefficient of friction. In this case, Equations 16–19 imply that, for small Pe ($S \sim \cos(\delta)$)

$$T_f \propto z_f^2 \propto \tau'^4,$$

(22)

equivalent to $b = 384 / 105\sqrt{\pi} \sim 2.06$, and for large Pe

$$T_f \propto z_f \sqrt{u} \propto \tau'^{5/2},$$

(23)
equivalent to $b = 15\sqrt{\pi} / 16 \sim 1.66$. For depth-dependent heating on a planar interface, as considered by Molnar and England (1990),

$$T_f \propto z_f \sqrt{u} \propto \frac{3\sqrt{\pi}}{4} \sim 1.33,$$

so the appropriate value of $b$ for that case is $3\sqrt{\pi} / 4 \sim 1.33$ (Table 1).

### 2.4. Comparison With Previous Analytical Approximations

The essential aspect of the analytical approximations is that temperatures on the interface are reduced, by a divisor $S$, from those that would be calculated for the same sources of heat ($Q$ and/or $\tau'V$), but in the absence of advection. In general,

$$S = \cos(\delta) + b \frac{K_1}{K_2} \sqrt{\text{Pe}},$$

(25)
where $t$ is the time taken for the lower plate to travel from the trench to depth $z_f$ (Molnar & England, 1990). The relation between distance-squared and time, expressed in Pe, is fundamental to the diffusion of heat, whereas $b$ and $\cos(\delta)$ arise from the shape of the plate interface, reflecting respectively the time evolution of temperature rise along the interface, and its local dip.

England (2018) and Tichelaar and Ruff (1993) dealt with curvature of the interface by using the expressions for planar interface with the sine of the dip given by $z_f/u$; the influence of this approximation is negligible, as can be seen from Equation 2, recalling that $\sinh^{-1}(x) \sim x$ for small $x$. Early applications of the analytical expressions (e.g., Molnar & England, 1990, 1995; Tichelaar & Ruff, 1993) approximated $\cos(\delta) \sim 1$, recognizing that this simplification is minor in comparison with variation in Pe along, and between, interfaces (Equation 5 and Figure 2).

2.4.1. Errors in the Analysis of van Keken et al. (2019)

van Keken et al. (2019) revisited the analysis of Molnar and England (1990) in its entirety, and concluded that neither the original expressions given by Molnar and England (1990) nor their own modifications to those expressions could “model the thermal structure for slabs with low thermal parameter due to a fundamental assumption in the derivation of the equations.” They also concluded that the analytical expressions “cannot be used for models with significant variation in dip.” Those conclusions are incorrect.

Although a fundamental error is indeed associated with the first conclusion of van Keken et al. (2019), it lies not in the derivation of the original equations but in van Keken and coworkers’ use of the age of the oceanic lithosphere at the trench to calculate its contribution to the heat flux at depth on the plate interface. As described above (Equation 14), and initially by Molnar and England (1995, Equation 6), one should use the age of the lithosphere as it passes below the point of interest. In consequence van Keken et al. (2019), overestimate temperatures on interfaces above young subducting oceanic lithosphere by up to hundreds of °C. The correct analytical expressions agree with numerical calculations to within a few percent, even for ocean floor as young as 3 Myr (see Molnar & England, 1995, Figures 3 and 4). Abers et al. (2020, Equation 7) appear to have noticed the error of van Keken et al. (2019) but, in attempting to correct it, gave the wrong sign in the denominator of the right-hand side of Equation 6 of Molnar and England (1995).

In discussing dissipation on the interface, van Keken et al. (2019) found poorer agreement than we do between numerical solutions and analytical approximations, because they used the incorrect value of $b_f$ (1, rather than $3\sqrt{\pi} / 4$, Equation 24) when evaluating Molnar and England’s $t$ (1990) expressions; this can be verified by comparing Figure 10 of Molnar and England (1990) with Figure S2.4 of van Keken et al. (2019).

van Keken and coworkers’ conclusion that analytical expressions cannot be applied to plate interfaces whose dip varies with depth is disproved in the subsection that follows. We differ from van Keken et al. (2019) on other mathematical points that are of secondary importance to the arguments we develop here; we do not discuss those issues, but our silence should not be interpreted as acquiescence.

2.5. Accuracy of the Analytical Expressions

To determine the accuracy of the analytical expressions of Equations 10–19, we compare them against numerical solutions to the full equations; the means of numerical solution are described in Appendix A. We use $a = 0.001, 0.002$, and $0.004$ km$^{-1}$, which span most of the range observed in modern plate interfaces (Figure 2b), and consider depths up to 80 km, at which the dip of the interface is 9°, 18°, and 33°, respectively.

Figures 3a, 3c and 3e compare the analytical approximation for heating from below (Equations 10–13) using $b_{Q} = 2/\sqrt{\pi}$ ($m = 2$) with the numerical solutions. For ocean floor of age 100 Myr, the differences between numerical and approximate calculations are smaller than 15°C, with the exception of a maximum difference of 18°C for a rate of 10 mm/yr, with $a = 0.0004$. For ocean floor of age 9 Myr, the differences rise to 20–30°C but remain smaller than 10% of the actual temperature.
Figure 3. Differences in temperature, $\Delta T$, between analytical approximations and numerical solutions (analytical-numerical). The shape of the plate interface is described by Equation 1 with $a = 0.001 \text{ km}^{-1}$ (a and b), $a = 0.002 \text{ km}^{-1}$ (c and d), and $a = 0.004 \text{ km}^{-1}$ (e and f). (a) No dissipative heating, with ocean lithosphere of age 100 and 9 Myr at the trench. Temperatures to right of figure are those determined numerically at the depth of 80 km for the relevant convergence rates and age of lithosphere. Numbers on curves are convergence rate, $V$ in mm/yr. (b) No heat flux through base of lower plate, dissipative heating on the interface at a rate proportional to depth: $\tau' = \mu' \rho g z_i$ with $\mu' = 0.06$. Temperatures to right of each panel are those determined numerically at the depth of 80 km for the relevant convergence rates. (c and e) As (a), with $a = 0.002, 0.004 \text{ km}^{-1}$. (d and f) As (b), with $a = 0.002, 0.004 \text{ km}^{-1}$. 

For the case in which there is dissipation on the interface but no heat supplied from below (Equations 16–19 and Figures 3b, 3d and 3f), we use \( b_1 = 15\sqrt{\pi}/16 \) \((m = 5)\). Temperatures, and hence differences between analytical and numerical solutions, are proportional to \( \mu' \); those illustrated are for \( \mu' = 0.06 \). The maximum differences between numerical and approximate calculations are, again, less than 15°C, except for a maximum difference of 22°C for a rate of 100 mm/yr, with \( a = 0.004 \).

3. Range of Temperatures at and Near Plate Interfaces

Our aim has been to provide simple and accurate expressions that may be used for quantitative analysis of a wide range of questions concerning thermal regimes on plate interfaces. We illustrate their use by applying them to the global range of subduction zones to determine the influence of their parameters on the thermal gradients along, and perpendicular to, their plate interfaces. The profiles we use are those of the 80 subduction segments for which England (2018) could reliably determine maximum depths of thrust faulting on the plate interface (the subset illustrated in Figure 2), and temperatures are calculated from the trench to those depths. Convergence rate, \( V \), and the angle, \( \phi \), between the trench and the convergence vector at the origin of the profile, are determined from NNR-MORVELS6 (Argus et al., 2011), the slab configuration is given by its best-fitting parabola (Section 1.1) and the heat flux, \( Q \), is calculated from the age of the ocean floor entering the trench, using Equations 14 and 15.

3.1. Temperature Profiles in the Absence of Dissipation on the Interface

If shear stress on the interface is sufficiently low that dissipation may be neglected in comparison with heat flux out of the lower plate, then the temperature on the interface is

\[
T_f \sim \frac{Q}{K'} \sqrt{\frac{kU}{V}} \sim \frac{Q}{K'} \sqrt{\frac{kz_f}{V \sin(\delta')}}.
\]

(27)

where the approximation consists in taking \( S_Q \gg 1 \), and we have defined an average plunge, \( \delta' \), of the interface in the direction of motion by \( u = z_f / \sin(\delta') \). Equation 27 gives a scale, \( \Theta \), for the temperature gradient

\[
\Theta = \frac{T_f}{z_f} \sim \frac{Q}{K'} \sqrt{\frac{k}{V \sin(\delta')z_f}}.
\]

(28)

Figure 4a shows profiles of temperature along plate interfaces, calculated under the assumption of no dissipative heating using Equation 11; the parameter \( \Theta \) encapsulates well the variation in temperature profiles. Making the simplification that \( Q \) is inversely proportional to the square root of the age, \( A \), of the ocean floor (Equation 14) the determining parameter, \( \Theta \), may be written

\[
\Theta \propto 1 \sqrt{AV \sin(\delta')z_f},
\]

(29)

from which it may be seen that descent speed is of equal importance to the age of the slab in determining the thermal gradient along the interface. Furthermore, the temperature gradient along the top of the slab decreases with depth, as can be seen in Figure 4a.

Although the quantity \( AV \sin(\delta') \) in Equation 29 resembles the “slab thermal parameter,” often given the symbol \( \Phi \) (Kirby et al., 1991), its physical significance is different. \( \Phi \) relates to temperatures in the interior of the slab; in particular, the maximum distance that a given isotherm is advected down-dip increases with \( \Phi \) (e.g., McKenzie, 1969; Molnar et al., 1979). In the present context, \( AV \sin(\delta') \) determines temperatures only near the top of the slab, and only when dissipative heating on the interface is negligible.

3.2. Temperature Profiles With Dissipative Heating on the Interface

It is unlikely, however, that shear stresses on the interface are negligible; estimates of shear stresses on plate boundaries from heat flux (e.g., Gao & Wang, 2014; McCaffrey et al., 2008; Molnar & England, 1990; Von Herzen et al., 2001; Yoshioka et al., 2013) and the support of topography (e.g., Cattin et al., 1997; Cubas et al., 2013; Dahlen, 1990; Davis et al., 1983; Dielforder, 2017; Lamb, 2006; Suppe, 2007) lie in the range of
Figure 4.
tens to a few hundred MPa. With the same approximations that lead to Equation 28, we obtain a scale temperature gradient associated with dissipative heating

$$\Psi = \frac{T_f}{z_f} - \frac{\mu' \rho g}{K_1} \frac{kVz_f}{\sin(\delta')}.$$  \hspace{1cm} (30)

Whereas $\Theta$ decreases with convergence rate, $\Psi$ increases, so the simple dependence of temperature gradient on slab age and descent speed is now lost. The ratio of the two scale temperature gradients is

$$\frac{\Psi}{\Theta} = \frac{V \tau'}{Q}.$$ \hspace{1cm} (31)

With $V$ varying between 10 and 150 mm/yr, and $Q$ being 100–200 mW/m$^2$ for some zones with young ocean floor, and ~40 mW/m$^2$ for old ocean floor, it is evident that the balance between basal heat flux, $Q$, and dissipation, $r'V$ must vary from zone to zone. Figures 4b and 4c, which show calculations with dissipation added on the interface, confirms that there is no simple dependence of thermal gradient along the interface on $\Psi$. Nevertheless, there is some simplicity: with the exception of a small number of profiles involving young ocean floor, temperature gradients along the interface lie in the range $6.5 \pm 2.5$°C/km for $\mu' = 0.03$ and $11 \pm 4$°C/km for $\mu' = 0.06$.

### 3.3. Temperature Gradients at the Top of the Lower Plate

We may also determine scale temperature gradients beneath the interface. In what follows, we assume that the conductivity is uniform, setting $K_1 = K_2 = K$, for which we assume a value of 3 W m$^{-1}$ K$^{-1}$. In the absence of dissipative heating, the thermal gradient beneath the interface is

$$\left.\frac{\partial T}{\partial y}\right|_{y=0} = \frac{Q}{K_1} \left(1 - \frac{b_Q \sqrt{Pe}}{1 + b_Q \sqrt{Pe}}\right);$$ \hspace{1cm} (32)

$y$ is positive downwards into the plate, so negative gradients correspond to temperatures decreasing with distance into the lower plate. The first term inside the brackets in Equation 32 arises from the thermal gradient near the top of the lower plate had it not been subducted, and the second term from the perturbation due to subduction (Section 2.1). At small and large Pe, the thermal gradient becomes

$$\Gamma_Q = \left.\frac{\partial T}{\partial y}\right|_{y=0} \sim \frac{Q}{K} \left(1 - \frac{b_Q \sqrt{Pe}}{1 + b_Q \sqrt{Pe}}\right) \quad Pe \ll 1$$

$$\sim 0 \quad Pe \gg 1.$$ \hspace{1cm} (33) (34)

The expressions for dissipative heating alone give perturbations to the thermal gradient of

$$\Gamma_D \sim -\frac{r'Vb_t \sqrt{Pe}}{K} \quad Pe \ll 1$$

$$\sim -\frac{r'V}{K} \quad Pe \gg 1.$$ \hspace{1cm} (35) (36)

We define a dimensionless temperature gradient below the interface

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Figure 4. Profiles of temperature along the plate interface, and of inverted temperature gradient beneath the interface. (a) Temperature profiles on the plate interfaces of the 80 subduction zones considered here, using Equation 11 alone (no dissipation on the interface), with the value of $Q$ appropriate to the age of the ocean floor adjacent to the beginning of the profile adjusted for hydrothermal circulation (Equation 15). Colors of line correspond to $\Theta$ (Equation 28), evaluated at $z_f$. Dashed profiles are for ocean floor younger than 20 Myr, with temperatures calculated without hydrothermal adjustment (Equation 14); influence of that adjustment is negligible for greater ages. (b) As for (a), with the addition of temperatures due to dissipative heating on the interface, with $\mu' = 0.03$ (Equation 17); (c) as for (a), with $\mu' = 0.06$. Colors of line correspond to $\Psi$ (Equation 30), evaluated at $z_f$. The end of each profile is at the maximum depth of thrust-faulting earthquakes for that interface (England, 2018), and is marked by a dot. Gray lines are labeled with their slopes in °C/km. (d) Dimensionless temperature gradient (Equation 37) immediately beneath the 80 interfaces considered here, with conditions corresponding to those in (a). (e) Scale inverted gradient for dissipative heating on the interface ($r'V / K$, Equation 36), contours in °C/km. (f) As for (d), but with dissipative heating on the interface, and $\mu' = 0.03, 0.06$; curves from the two values of $\mu'$ overlap.
\[
\Gamma' = \frac{KT_Q}{Q} + \frac{KT_r}{\tau V}. \tag{37}
\]

The scale gradient when there is no dissipative heating is \(Q / K\); for example, using the expressions of Parsons and Sclater (1977), it is \(\sim 50^\circ\text{C/km}\) for ocean floor of age 10 Myr, and \((30, 20, 15)^\circ\text{C/km}\) for ages of \((25, 60, 120)\) Myr. The thermal gradient drops from this scale value when it enters the trench to 20\% of that value when \(\sqrt{Pe} \sim 3\) and to 10\% of that value when \(\sqrt{Pe} \sim 6\) (Figure 4d).

The scale gradients in the case of dissipative heating on the interface are shown in Figure 4e for shear stress up to 100 MPa, and convergence rates up to 150 mm/yr. Figure 4f shows dimensionless gradients for the 80 subduction zones when both heating from below and on the interface are included; the curves for \(\mu' = 0.03\) and \(\mu' = 0.06\) overlap one another. In all cases, the influence of dissipation on the interface exceeds that of heating from below (gradients become negative) when \(\sqrt{Pe} \geq 1\); the negative gradients reach about half of their final value when \(\sqrt{Pe} \sim 2\) and 80\% of that value when \(\sqrt{Pe} \sim 6\).

### 3.4. “Hot,” “Cool,” and “Young” Plate Interfaces

Many discussions of the distribution of temperatures along plate interfaces posit end-member “hot” and “cool” subduction zones, and often associate the former with subduction of young ocean floor—the Nankai and Cascadia margins are common exemplars—and associate the latter with subduction of old ocean floor, such as beneath Japan and the Marianas (e.g., Peacock & Wang, 1999; Tsujimori & Ernst, 2013; van Keken et al., 2018). The discussion of Section 3.1 shows that the age of the ocean floor does not, alone, determine the temperature profile along the interface, even in the absence of dissipation. When dissipation does occur on the interface, then its influence generally outweighs that of slab age, so differences between “hot” and “cold” subduction zones reflect variation in the product of shear stress and convergence rate on the interface (Section 3.2 and Figure 4).

The notion of hot and cool subduction zones often arises in the interpretation of rocks from high-pressure-low-temperature (HPLT) terrains; here an additional problem arises because the stratigraphic position, within the interface, of the rocks is usually unknown. The temperature differences across a depth range of a kilometer near the interface are tens to about 100°C (Figures 4d–4f), which is comparable to the range of temperatures at given depths and \(\mu'\) across all the subduction segments shown in Figures 4a–4c. P-T measurements from HPLT terrains are therefore unreliable indicators of parameters of subduction such as convergence rate or the age of the descending plate.

Another tenet sometimes applied to the discussion of HPLT terrains is that temperatures in young subduction zones are hotter, at a given depth, than they are in mature subduction zones (e.g., Agard et al., 2018; Krebs et al., 2008; Peacock, 2020; van Keken et al., 2018). Expressions for the evolution of temperature on the plate interface during the initiation of subduction show that in the absence of dissipation temperatures do, indeed, drop (Molnar & England, 1990, Equations 3–9, Figure 4), but the steady-state temperatures are less than 100–200°C unless the age of the subduction ocean floor is also very young (Figure 4f) (see also numerical calculations for three individual subduction zones [van Keken et al., 2018, Figure 15]).

In contrast, if there is dissipative heating on the interface—which is required if temperatures on plate interfaces are to approach those of rocks from HPLT terrains that record equivalent pressures (\(\lesssim 1.5\text{ GPa}\)) (Kohn et al., 2018; Pennistion-Dorland et al., 2015)—temperatures rise on the interface during the early phase of subduction (Molnar & England, 1990, Equations 11 and 12, Figure 5). It therefore seems that the plate interfaces of young subduction zones are likely to be cooler, not hotter, than in their final states.

Our calculations do not address the interface between the mantle wedge and the slab, but we note that the aforementioned calculations of van Keken et al. (2018) show temperature drops of \(\lesssim 100^\circ\text{C}\) on the wedge-slab interface during the first few Myr; those drops are small in comparison with the range of temperatures calculated for mature wedge-slab interfaces (e.g., Syracuse et al., 2010, as corrected and reported by van Keken et al., 2011, Supporting Information). In contrast, Gerya et al. (2002) show temperatures in the mantle wedge increasing as the subduction zone ages. That evolution is consistent with scaling arguments (e.g.,
England & Wilkins, 2004; England & Katz, 2010): temperatures in the mantle wedge are dominated by the advection of heat toward the wedge corner, which is driven by the downward motion of the descending slab. We should therefore expect temperatures within the wedge and on the wedge-slab interface to increase as the subduction zone ages, the slab lengthens, and the intensity of flow in the wedge increases.

4. Conclusions

The shapes of most modern plate interfaces are well described by parabolas, with curvature in the range 0.0005–0.004 km$^{-1}$ (Figure 2). We give analytical expressions for temperatures on parabolic plate interfaces (Equations 10–19) and discuss, in Section 2.4, the differences between these and previous expressions, some of which (van Keken et al., 2019) are erroneous. The analytical expressions differ from numerical solutions to the full equations by a few percent (Figure 3). Such differences are negligible in comparison with the epistemic uncertainty attached to the idealization of plate interfaces by a simple mathematical model, and with uncertainties in the physical parameters. For instance, uncertainties in the relevant thermal conductivity and diffusivity are at least 20% (e.g., England, 2018, Appendix A) and uncertainties in shear stresses during motion on the interface are surely some tens of percent (Section 3.2). There is therefore no way of telling, for a given case of interest, whether a numerical model lies closer to reality than do the analytical expressions. The latter are to be preferred because they provide a simple and transparent means of evaluating temperatures, and their uncertainties, on plate boundaries.

Temperatures calculated for plate interfaces depend critically upon assumptions about the level of shear heating there (e.g., Kohn et al., 2018; Molnar & England, 1990; Peacock, 1996; Pennistur-Doland et al., 2015). The global range of temperatures at 30–60 km is ~50–200°C without shear heating (Figure 4a), ~250–350°C if the effective coefficient of friction during slip, $\mu'$, is 0.003 (Figure 4b), and ~250–600°C if $\mu'$ is 0.006 (Figure 4c). For the range of shear stresses likely to operate during slip on plate interfaces (e.g., England, 2018; Gao & Wang, 2014; McCaffrey et al., 2008; Molnar & England, 1990; Von Herzen et al., 2001; Yoshioka et al., 2013), inverted temperature gradients at the plate interface are likely to be several tens of, to about 100°C/km (Figure 4c).

The use of pressure-temperature conditions recorded by HPLT rocks to infer parameters of ancient subduction zones, such as convergence rate, ages of ocean floor, or maturity, should be approached with caution (Section 3.4).

Appendix A: Numerical Calculations

The numerical calculations employed in this work share many similarities with previous kinematically driven models of subduction zones (e.g., Currie et al., 2004; Syracuse et al., 2010; van Keken et al., 2002, 2008). In such models, the flow in the wedge is driven by a subducting plate that is prescribed in terms of a time-independent geometry and of a convergence rate $V$. The geometry of the calculations is defined by four non-overlapping domains, shown in Figure A1: the mantle wedge $\Omega_w$; the subducting plate $\Omega_s$; the over-riding plate $\Omega_r$; and a thin dissipation layer draped on top of the plate interface $\Gamma_s$. The geometry of $\Gamma_s$ is given by Equation 1. The line $\Gamma_s$ is obtained by locally projecting $\Gamma_{interface}$ a distance $\delta_s$ (the subducting plate thickness) in the direction normal to the interface. The tip of the dissipation layer, which connects to $\Omega_r$, is defined by a straight line segment and constructed such that it intersects the plate interface at 30°. We describe the full calculation scheme here, although for the present application we solve for deformation only in the slab.

The origin on the coordinate system is located at the trench (indicated by the solid gray circle in Figure A1). The overall width of the calculation domain, $L$, is set to 1,200 km to accommodate the full range of interface geometry we consider, and the thickness $\delta_s$ is set to 100 km. These lengths are comfortably large enough that the positions of the associated boundaries do not influence our calculations. For the shallowest interfaces $a = 0.001$, at the smallest convergence rates, we consider (10 mm/yr), the top of the lower plate reaches 80 km (the maximum depth over which we analyze solutions) in ~30 Myr; the diffusion length ($\sim \sqrt{kt}$) associated with this time is ~15 km.
We solve the incompressible Stokes flow equations

\[
\nabla \cdot \left( 2\eta \nabla \varepsilon (\nu) \right) + \nabla p = 0, \quad \nabla \cdot \nu = 0, \quad \forall x \in \Omega_{w,s},
\]

where \( \nu \) is the velocity, \( p \) the pressure, \( \eta \) the shear viscosity, and \( \nabla \varepsilon (\nu) \) is the symmetric gradient of the velocity (with the assumed incompressibility constraint): \( \frac{1}{2} (v_{ij} + v_{ji}) \). These equations are coupled with steady-state conservation of energy

\[
(\nu \cdot \nabla T) = \nabla \cdot (\kappa \nabla T) + \frac{1}{\rho C_p} \frac{H(x)}{\rho \delta}, \quad \forall x \in \Omega_{w,s},
\]

where \( \kappa \) is the thermal conductivity, \( C_p \) is the specific heat, \( \rho \) is the reference density, \( H(x) \) is the volumetric rate of heat production in the dissipative layer, and \( \delta \) is the thickness of the layer.

**Figure A1.** Geometry (not to scale) of the kinematically driven numerical subduction model. Refer to the text and to Table A1 for all parameter definitions and values used in the present study. The sketch shows the configuration of the domains in the general case. In the case considered here, depth of the base of the dissipation layer \( \Omega_c \) is set to be equal to that of the base of the over-riding plate \( \Omega_p \).

| Parameter       | Definition                                      | Value                  |
|-----------------|-------------------------------------------------|------------------------|
| \( C_p \)       | Specific heat                                   | 1,250 J kg\(^{-1}\) K\(^{-1}\) |
| \( D \)         | Depth of the base of the wedge domain           | 660 km                 |
| \( H \)         | Volumetric rate of heat production in dissipative layer | \( z'V' / \delta \) |
| \( K \)         | Thermal conductivity                            | 3 W m\(^{-1}\) K\(^{-1}\) |
| \( L \)         | Length of the domain                            | 1,200 km               |
| \( T_i \)       | Temperature of the mantle wedge at depth (Equation A7) | 1,300°C               |
| \( w_t \)       | Width of the velocity transition (along the slab) | 1 km                   |
| \( \delta_p \)  | Thickness of the over-riding plate              | 120 km                 |
| \( \delta_p \)  | Depth of \( p_c = \delta_p \)                  | 120 km                 |
| \( \delta_s \)  | Thickness of the subducting plate (slab)        | 100 km                 |
| \( \delta_l \)  | Thickness of the dissipation layer              | 1 km                   |
| \( \delta_{lg} \)| Depth of linear geotherm (Equation A7)         | 180 km                 |
| \( \kappa \)    | Thermal diffusivity                             | \( K / \rho C_p = 7.27 \times 10^{-7} \) m\(^2\) s\(^{-1}\) |
| \( \rho \)      | Reference density                                | 3,300 kg/m\(^3\)      |
in which the density $\rho$ and thermal diffusivity, $\kappa$, are taken to be constant at their reference values throughout the domain $\Omega = \cup_{\chi = \text{w.s.t.p}} \Omega_{\chi}$. The only source of volumetric heating, $H_{s}$, in the present calculations arises from dissipative heating, which is explicitly defined within the volume $\Omega_{s}$; the convergence rate, $V$, multiplied by the effective shear stress during slip, $\tau^\prime$, and divided by the thickness of the volume, $\delta_{i}$. In contrast to previous kinematically driven subduction zone models, we solve for the flow within in the subducting plate domain $\Omega_{p}$. This enables us to obtain incompressible flow fields within the slab for arbitrarily complex interface geometries, reducing potential modeling errors and inconsistencies which would arise in solving the energy equation which explicitly assumes that $\nabla \cdot v = 0$. The boundary conditions used for the flow problem within the slab ($\Omega_{s}$) are given by

$$v \cdot n_{b} = -V, \quad v \cdot t_{b} = 0 \quad \forall x \in \Gamma_{b},$$  
(A3)

$$v \cdot n_{s} = 0, \quad v \cdot t_{s} = V \quad \forall x \in \Gamma_{\text{interface}},$$  
(A4)

$$v \cdot n_{i} = 0, \quad t_{i} \cdot \tau \cdot n_{i} = 0 \quad \forall x \in \Gamma_{c},$$  
(A5)

$$n_{s} \cdot (\tau - \rho I) \cdot n_{s} = 0, \quad t_{s} \cdot \tau \cdot n_{s} = 0 \quad \forall x \in \Gamma_{p},$$  
(A6)

where $\tau = 2\eta \nabla \cdot v$ is the deviatoric stress, $n_{b}$ is the outward pointing normal vector to the boundary of $\Omega_{b}$ and $t_{b}$ is a unit tangent vector, defined such that $t_{b} \times n_{b} = \hat{k}$. We assume a homogenous viscosity for the slab; with this assumption, and the form of boundary conditions, the velocity within the slab is independent of $\tau$ (arbitrarily high) viscosity.

Once the flow in the slab is computed, we introduce a smooth transition in velocity along $\Gamma_{\text{interface}}$ between the points $P_{dc}$ and $P_{c}$. The modification applies the following changes: (a) left (trench-ward) of $P_{dc}$, (b) between $P_{dc}$ and $P_{c}$, we set $v(\Gamma_{\text{interface}}) = 0$; (b) between $P_{dc}$ and $P_{c}$, we set

$$v(x) = \mathcal{F}(\nu(x), x), \quad \forall x \in \Gamma_{\text{interface}}$$

where $\mathcal{F}$ is given by the cubic spline

$$\mathcal{F}(Y, x) = Y \left(3\alpha^{2} + 2\alpha - 2\right), \quad \alpha = \frac{\nu(x) - \nu(P_{dc})}{\nu_{t}},$$

in which $\nu_{t}$ defines the width of the transition along-slab, and $\nu()$ (given by Equation 2) defines the distance along $\Gamma_{\text{interface}}$ as measured from the trench (i.e., the origin). In all experiments, $P_{dc}$ was defined as the terminating point of the dissipation layer, and $P_{c}$ was taken 1 km away (in the down-dip direction).

Within the over-riding plate ($\Omega_{o}$) and the dissipation layer ($\Omega_{d}$), we impose that the velocity $v = 0$. For the purposes of this study, which considers only the distribution of temperature within two plates in relative motion, we set the thickness of the upper plate to be equal to the depth of $P_{dc}$. To ensure that advection of heat in the wedge does not influence the numerical solutions, we also set $v = 0$ in the wedge ($\Omega_{d}$).

Along the top of the domain, we impose $T = T_{0} = 0^\circ C$. Along $\Gamma_{a}$ we impose $T$ using the half-space cooling model for the age, $t_{o}$, of the ocean floor considered. Along the right-most boundary ($x = L$), we impose $T$ using a depth-limited linear geothermal gradient

$$T(x) = \begin{cases} \frac{T_{i} - T_{0}}{\delta_{tg}} z + T_{0} & \text{if } z \leq \delta_{tg}, \\ T_{i} & \text{if } z > \delta_{tg}. \end{cases}$$  
(A7)

Along all outflow boundaries ($v \cdot n \geq 0$), we impose

$$\kappa \nabla T \cdot n = 0, \quad \forall x \in \partial \Omega, \quad v \cdot n \geq 0.$$  

Note that in our models, the only segments where outflow exists is along $\Gamma_{b}$ and $\Gamma_{c}$.
The numerical simulations utilize a continuous Galerkin finite element (FE) method employing an unstructured mesh comprised of triangles. An inf-sup stable mixed formulation (Brezzi & Fortin, 1991) was used to discretize the flow problem in Equation A1 employing \( P_2 \) (quadratic) \( \times P_1 \) (linear) function space for velocity \( \mathbf{v} \) and pressure \( p \) respectively. The energy equation (Equation A2) was also solved using continuous Galerkin finite element, with temperature \( T \) discretized by a \( P_2 \) function space. The use of an analytic description of the subduction interface is particularly useful in the context of FE modeling as it provides an analytic description of the normal and tangent vectors along the slab. This is particularly useful when describing the boundary conditions for the flow problem in the slab. Additionally, since the subduction interface used here is quadratic, we can exactly represent this geometry with the quadratic function space used for velocity and temperature. Hence, in the calculations here, the finite elements along the subduction interface are actually curved and conform exactly to Equation A1. Constraints on the normal and tangential components of the velocity field along this interface are imposed by locally rotating the coordinate system associated with each FE basis function to be aligned with the geometry of the subduction interface. The numerical solution of the steady-state energy equation did not require any numerical stabilization techniques (e.g., SUPG, entropy viscosity) due to the magnitude of the Péclet numbers under consideration and the numerical spatial resolution used. Denoting the area of each finite element as \( A_n \), in the calculations shown here the meshes were constructed such that within the dissipation layer \( A_n \leq 0.0025 \) km\(^2\), while in all other regions \( \Omega \), we used \( A_n \leq 5.0 \) km\(^2\). This results in a spatial resolution of \( \approx 0.7 \) km inside \( \Omega \) and \( \approx 3 \) km elsewhere. The underlying finite element software was developed using PETSc (Balay et al., 2017).

Data Availability Statement

Data used in this study are from the published sources cited.

References

Abers, G. A., van Keken, P. E., & Wilson, C. R. (2020). Deep decoupling in subduction zones: Observations and temperature limits. Geosphere, 16, 1408–1424. https://doi.org/10.1130/GES0278.1
Agard, P., Flunder, A., Angiboust, S., Bonnet, G., & Ruh, J. (2018). The subduction plate interface: Rock record and mechanical coupling (from long to short timescales). Lithos, 320–321, 537–566. https://doi.org/10.1016/j.lithos.2018.09.029
Argus, D. F., Gordon, R. G., & Demets, C. (2011). Geologically current motion of 56 plates relative to the no-net-rotation reference frame. Geochronology, Geophysics, Geosystems, 12, Q11001. https://doi.org/10.1029/2011gc003751
Balay, S., Abhyankar, S., Adams, M. F., Brown, J., Brune, P., Buschelman, K., et al. (2017). PETSc users manual (Technical report). Argonne National Laboratory.

Beroza, G. C., & Ide, S. (2011). Slow earthquakes and nonvolcanic tremor. Annual Review of Earth and Planetary Sciences, 39, 271–296. https://doi.org/10.1146/annurev-earth-040809-152531
Brezzi, F., & Fortin, M. (1991). Mixed and hybrid finite element methods (Vol. 15). Springer Science & Business Media.

Carlson, H. S., & Jaeger, J. C. (1959). Conduction of heat in solids (2nd ed.). Oxford University Press.
Cattin, R., Lyon-Caen, H., & Chéry, J. (1997). Quantification of interplate coupling in subduction zones and forearc topography. Geophysical Research Letters, 24, 1563–1566. https://doi.org/10.1029/97gl01550
Cubas, N., Avouac, J. P., Leroy, Y. M., & Pons, A. (2013). Low friction along the high slip patch of the 2011 Mw 9.0 Tohoku-Oki earthquake required from the wedge structure and extensional splay faults. Geophysical Research Letters, 40, 4231–4237. https://doi.org/10.1002/2012gl050682

Currie, C., Wang, K., Hyndman, R. D., & He, J. (2004). The thermal effects of steady-state slab-driven mantle flow above a subducting plate: The Cascadia subduction zone and backarc. Earth and Planetary Science Letters, 223, 35–48. https://doi.org/10.1016/j.epsl.2004.04.020

Dahlen, F. A. (1990). Critical taper model of fold-and-thrust belts and accretionary wedges. Annual Review of Earth and Planetary Sciences, 18, 55–99. https://doi.org/10.1146/annurev.ea.18.050190.000415

Davis, D., Suppe, J., & Dahlen, F. A. (1983). Mechanics of fold-and-thrust belts and accretionary wedges. Journal of Geophysical Research, 88, 1153–1172. https://doi.org/10.1029/JB088iB02p01153

Diefendorf, A. (2017). Constraining the strength of megathrusts from fault geometries and application to the Alpine collision zone. Earth and Planetary Science Letters, 474, 49–58. https://doi.org/10.1016/j.epsl.2017.06.021

England, P. C. (2018). On shear stresses and the maximum magnitudes of earthquakes at convergent plate boundaries. Journal of Geophysical Research: Solid Earth, 123, 7165–7202. https://doi.org/10.1029/2018JB015907

England, P. C., & Katz, R. F. (2010). Melting above the anhydrous solidus controls the location of volcanic arcs. Nature, 467, 700–703. https://doi.org/10.1038/nature09417

England, P. C., & Wilkins, C. (2004). A simple analytical approximation to the temperature structure in subduction zones. Geophysical Journal International, 159, 1138–1154. https://doi.org/10.1111/j.1365-246X.2004.02419.x

Gao, X., & Wang, K. (2014). Strength of stick-slip and creeping subduction megathrusts from heat flow observations. Science, 345, 1038–1041. https://doi.org/10.1126/science.1255487

Gerya, T. V., Stöckhert, R., & Perczuk, A. L. (2002). Exhumation of high-pressure metamorphic rocks in a subduction channel: A numerical simulation. Tectonics, 21, 6–1–6–19. https://doi.org/10.1029/2002tc001406

Hayes, G. P., Moore, G. L., Portner, D. E., Hearne, M., Flamme, H., Furtney, M., & Smokey, G. M. (2018). Slab2, a comprehensive subduction zone geometry model. Science, 362, 58–61. https://doi.org/10.1126/science.aat4723
Wessel, P., & Smith, W. H. F. (2013). Generic mapping tools: Improved version released. *Eos, Transactions American Geophysical Union*, 94, 409–410. https://doi.org/10.1002/2013eo450001

Wolery, T. J., & Sleep, N. H. (1976). Hydrothermal circulation and geochemical flux at mid-ocean ridges. *The Journal of Geology*, 84, 249–275. https://doi.org/10.1086/628195

Yoshioka, S., Suminokura, Y., Matsumoto, T., & Nakajima, J. (2013). Two-dimensional thermal modeling of subduction of the Philippine Sea plate beneath southwest Japan. *Tectonophysics*, 608, 1094–1108. https://doi.org/10.1016/j.tecto.2013.07.003