Chapter 5

The Thermal Evolution of an Earth with Strong Subduction Zones

Abstract. It is commonly supposed that plate tectonic rates are controlled by the temperature-dependent viscosity of Earth's deep interior. If this were so, a small decrease in mantle temperature would lead to a large decrease in global heat transport. This negative feedback mechanism would prevent mantle temperatures from changing rapidly with time. We propose alternatively that convection is primarily resisted by the bending of oceanic lithosphere at subduction zones. Because lithospheric strength should not depend strongly on interior mantle temperature, this relationship decreases the sensitivity of heat flow to changes in interior mantle viscosity, and thus permits more rapid temperature changes there. The bending resistance is large enough to limit heat flow rates for effective viscosities of the lithosphere greater than about 10^{23} Pa s, and increases with the cube of plate thickness. As a result, processes that affect plate thickness, such as small-scale convection or subduction initiation, could profoundly influence Earth's thermal history.

5.1 Introduction

Plate velocities are typically thought to depend on the viscosity of the mantle's deep interior, which depends strongly on temperature. In this view, as the earth cools, mantle viscosity increases, which forces slower convection and less efficient heat transport. This slows the mantle's cooling rate and stabilizes its internal temperature [e.g., *Davies,* 1980; *Tozer,* 1972]. Mantle temperatures can change more rapidly with time if convection is influenced by the lithosphere, which is colder, and therefore stiffer, than the underlying mantle. In simple convection calculations that include temperature-dependent viscosity, *Christensen* [1985] found surface heat flow to depend more on the viscosity of the cold thermal boundary layer than on that of the underlying mantle. Because lithospheric viscosity depends on surface temperatures, which are thought to have changed little since the Archean, *Christensen's* [1985] calculations predict nearly constant heat flow. The decoupling of heat flow from temperature prevents heat loss rates from slowing as the interior cools. This leads to greater present-day cooling rates in thermal evolution models that reproduce current mantle temperatures [*Christensen*, 1985].

The notion that the viscosity of the lithosphere limits mantle heat flow has received some criticism, particularly because *Christensen's* [1985] convection models do not produce realistic subduction zones. Instead, these models describe convection beneath a "stagnant lid," where only a small portion of the upper thermal boundary layer participates in downwelling beneath nearly stationary surface plates. *Gurnis* [1989] produced more realistic subduction and plate motions by including "weak zones" of low viscosity fluid at plate boundaries. These weak zones diminish the role of the lithosphere in controlling plate velocities. Because interior mantle viscosity remains as the controlling parameter, mantle temperatures are again stabilized.

The geometry of Earth's subduction zones requires cold, strong, lithosphere to deform as it bends and descends into the mantle. *Conrad and Hager* [1999a] show that this bending deformation provides at least as much resistance to plate motions as viscous deformation of the underlying mantle. Thus, the rheological properties of

the bending lithosphere should play an important role in determining plate velocities. If this is the case, the negative feedback mechanism that stabilizes temperatures is disrupted while plate motions and subduction are maintained. In this study, we include the bending resistance in studies of parameterized convection to show how plate bending at subduction zones could play an important role in Earth's thermal evolution.

5.2   Parameterized Convection Models

The relationship between convective heat transport and interior temperature is typically investigated in the context of Rayleigh-Bernard convection, where a constant temperature boundary condition is imposed at the base of the system. For this scenario, the dimensionless heat flux is given by the Nusselt number, Nu, which is the ratio of the heat flow due to convection to that due to conduction alone. For steady convection, Nu is often related to the Rayleigh number of the system, Ra_m, by a power law [e.g., *Davies,* 1980]:

$$Nu \sim Ra_m^{\beta} \qquad \text{where} \qquad Ra_m = \frac{\rho g \alpha T D^3}{\eta_m \kappa} \tag{5.1}$$

where ρ, g, α, and κ are the density, gravitational acceleration, thermal expansivity and diffusivity, and T is the temperature difference across the system's depth, D. For a system with constant viscosity η_m and constant basal temperature, boundary layer theory gives $\beta = 1/3$ [e.g., *Turcotte and Oxburgh,* 1967]. For internal heating, both the definition of Nu and the value of β predicted by boundary layer theory differ [e.g., *O'Connell and Hager,* 1980].

The strength of the negative feedback mechanism that stabilizes mantle temperatures depends on β. As the earth cools, Ra_m decreases because η_m depends on temperature. If $\beta > 0$, this decreases Nu, and slows the mantle's rate of temperature change. Conversely, knowledge of the mantle's cooling rate allows us to distinguish among convection models that yield different values of β [e.g., *Christensen,* 1985].

The earth's present rate of secular cooling is obtained from estimates of the Urey ratio, which is the ratio of current mantle heat production to total heat flow. The latter has been estimated at about 39×10^{12} W, after the contribution from radioactivity in the continental crust, 5×10^{12} W, is subtracted from the worldwide total heat flow [e.g., *O'Connell and Hager,* 1980; *Sclater et al.,* 1980; *Stein,* 1995]. The rate of internal heating can be estimated from the abundances of the heat producing elements, U, Th, and K, in the mantle. Their relative abundances are relatively well constrained to a ratio of about 1:4:10000 [e.g., *Jochem et al.,* 1983], but their absolute abundance is more uncertain. Geochemical models of the primitive mantle yield 21 ppb for the present abundance of U, giving a current mantle heat production of 15×10^{12} W [*Jochem et al.,* 1983]. For comparison, if the mantle were solely composed of the material sampled by mid-ocean ridges (MORB), with a concentration of 3 ppb for U, the mantle heat production would be 2.4×10^{12} W [*Kellogg et al.,* 1999]. These estimates, when compared with the mantle heat flow, yield a Urey ratio of 0.4 for models using a primitive mantle and 0.06 for a mantle composed of MORB source. Thus, 60% or more of the current mantle heat flow may represent secular cooling.

A thermal history for the earth is obtained by assuming a heating history from estimates of the Urey ratio and integrating heat flow rates from (5.1) backward in time from Earth's present thermal state. For $\beta = 1/3$, *Christensen* [1985] finds that the mantle cools from a molten state to its present one in significantly less than the expected 4.5 billion years if more than about 15% of current mantle heat flow is due to secular cooling. Thus, to allow Urey ratios less than 0.85, the feedback mechanism that stabilizes mantle temperatures must be diminished, which implies a decrease in β. If $\beta < 0.1$, *Christensen* [1985] obtains plausible thermal histories for Urey ratios between 0.3 and 0.6. These values are consistent with the above estimates of the Urey ratio, but, as described above, *Christensen's* [1985] convection models that yield $\beta < 0.1$ do not produce realistic plate motions or subducting slabs. In what follows, we introduce an analysis based on plate bending in subduction zones that produces small β while still allowing realistic plate and slab behavior.

5.3 Plate Bending and Mantle Heat Flow

The mantle loses much of its heat through the formation of oceanic lithosphere. The total heat flow from an oceanic plate, expressed here as N, a dimensionless number analogous to a Nusselt number, is given by:

$$N = 2D \left(\frac{v_p}{\pi \kappa L} \right)^{(1/2)} = \frac{4D}{\sqrt{\pi} h_s} \tag{5.2}$$

where v_p is the plate velocity, L is the ridge to trench plate length, and $h_s = 2\sqrt{\kappa L / v_p}$ is the plate thickness at the time of subduction [e.g., *Turcotte and Schubert*, 1982, p. 280-3]. It has been suggested that oceanic plates only thicken for 80 Ma, the age at which seafloor flattening is observed to begin [e.g., *Sclater et al.*, 1980]. If additional heat transport for older plates limits their thickness to some maximum value $h_m < h_s$, then N should be larger than that given by (5.2). For Earth, the additional heat flow is less than 10%, and is important only for the oldest plates [*Sclater et al.*, 1980].

The velocity of a plate can be estimated by considering the energy budget of a single convecting cell [e.g., *Conrad and Hager*, 1999a; *Richter*, 1984; *Turcotte and Oxburgh*, 1967]. Convection is driven by the potential energy release of a descending slab, given by $\Phi^{\mathrm{pe}} = \rho g \alpha T v_p l_s h_s / \sqrt{\pi}$ [*Conrad and Hager*, 1999a] where l_s is the effective slab length. Slab reheating is accompanied by cooling of the surrounding mantle, so a slab's net buoyancy change as it descends should be negligible. Thus, l_s is determined by the depth of convective circulation, and can be taken as a constant. Potential energy release is typically balanced by viscous dissipation within the shearing mantle, given by *Conrad and Hager* [1999a] as $\Phi^{\mathrm{vd}}_m = 3\eta_m v_p^2 (L/D + 2)$. *Conrad and Hager* [1999a] also derive the expression $\Phi^{\mathrm{vd}}_l = 2\eta_l v_p^2 (h_s/R)^3$ for the viscous dissipation within the bending subducting slab and note that Φ^{vd}_l could be as large as, or even much greater than, Φ^{vd}_m. Here η_l is the effective viscosity of the lithosphere and R is the radius of curvature of the subducting plate. Setting $\Phi^{\mathrm{pe}} = \Phi^{\mathrm{vd}}_m + \Phi^{\mathrm{vd}}_l$

and solving for v_p yields:

$$v_p = \frac{\rho g \alpha T l_s h_s / \sqrt{\pi}}{3\eta_m \left(L/D + 2\right) + 2\eta_l \left(h_s/R\right)^3} \tag{5.3}$$

We now solve for plate thickness $h_s = 2\sqrt{\kappa L/v_p}$ using (5.3):

$$h_s = \left(\frac{\eta_m L}{\eta_l l_s} \frac{12\sqrt{\pi} R^3 \left(L/D + 2\right)}{Ra_l - 8\sqrt{\pi} L/l_s}\right)^{(1/3)} \tag{5.4}$$

where we have defined a dimensionless "lithospheric" Rayleigh number:

$$Ra_l = \frac{\rho g \alpha T R^3}{\kappa \eta_l} \tag{5.5}$$

that is a measure of convective instability resisted by the bending lithosphere.

It is clear from (5.4) that as Ra_l approaches $8\sqrt{\pi}L/l_s$, the thickness h_s becomes infinite. This occurs because a plate slowed by lithospheric bending will thicken due to increased cooling, and thus slow even further because a thicker plate is more difficult to bend. This runaway process leads to infinitely thick plates, which are not plausible for the earth, so we impose a maximum plate thickness h_m. Thus, if h_s in (5.4) is larger than h_m, we obtain an expression for N by setting $h_s = h_m$ in (5.3) and applying the resulting expression for v_p to (5.2). In terms of Ra_m and Ra_l, this yields:

$$N = \left(\frac{4l_s h_m}{\pi^{3/2} L D} \frac{Ra_m Ra_l}{3 \left(L/D + 2\right) Ra_l + 2 Ra_m h_m^3/D^3}\right)^{(1/2)} \tag{5.6}$$

We now study limiting cases for weak and strong lithosphere.

5.3.1   Weak Bending Lithosphere

If lithosphere is weak or does not bend sharply, Ra_l is large. If $Ra_l >> 8\sqrt{\pi}L/l_s$, applying (5.4) to (5.2) yields:

$$N = \left(Ra_m \frac{16 l_s}{3\pi^2 L \left(L/D + 2\right)}\right)^{(1/3)} \tag{5.7}$$

which applies if Ra_m is sufficiently large that $h_s < h_m$. In this case, convection is adequately described by boundary layer theory, giving $\beta \sim 1/3$. If, however, Ra_m is sufficiently small that $h_s > h_m$, plates reach their maximum thickness before subducting and (5.6) yields:

$$N = \left(Ra_m \frac{4l_s h_m}{3\pi^{3/2} LD\left(L/D + 2\right)} \right)^{(1/2)} \tag{5.8}$$

In this scenario, plate velocities are slowed by resistance in the mantle until the plate thickness saturates. Because slab thickness is constant, faster plates are no longer slowed because their slabs are thinner. As a result, N is more sensitive to changes in Ra_m than it is for simple boundary layer theory, as shown by the power-law exponent of $\beta = 1/2$.

5.3.2 Strong Bending Lithosphere

A strong resistance to bending occurs if Ra_l is less than or about $8\sqrt{\pi}L/l_s$, in which case plates always reach their maximum thickness, so (5.6) applies. If Ra_m is small so that the denominator of (5.6) is dominated by Ra_l, then (5.6) reduces to (5.8). Otherwise, if Ra_m is large, (5.6) becomes:

$$N = \left(Ra_l \frac{2l_s D^2}{\pi^{3/2} Lh_m^2} \right)^{(1/2)} \tag{5.9}$$

Here the bending dissipation Φ_l^{vd}, overwhelms the dissipation in the underlying mantle Φ_m^{vd}, and thus controls plate velocities. This causes N to be sensitive only to Ra_l. If Ra_l is independent of changes in Ra_m, $\beta \sim 0$ in (5.1) is implied.

5.4 Application to Earth

We now determine how N varies with Ra_m for a set of parameter values that characterize the earth. We use $\rho_m = 3300$ kg m^{-3}, $g = 10$ m s^{-2}, $\alpha = 3 \times 10^{-5}$ K^{-1}, $T = 1200$ K, $\kappa = 10^{-6}$ m^2 s^{-1}, $D = 2500$ km, $R = 200$ km [*Bevis*, 1986], and
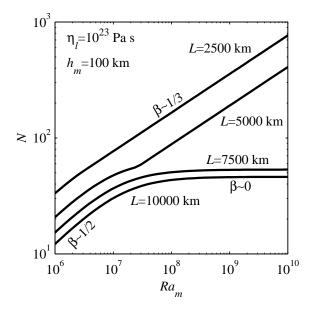
Figure 5.1: A log-log plot of N vs. Ra_m for four different plate lengths, L, using the "preferred" values of $\eta_l = 10^{23}$ Pa s and $h_m = 100$ km. The slopes, β, are the exponent of the $N \sim Ra_m^{\beta}$ relation.

$l_s = 1000$ km, and assume that these values remain constant throughout Earth history. In doing so, we assume that the dynamical effect of any change in T, by at most 50%, will be overwhelmed by the accompanying orders of magnitude change in mantle viscosity. Because these temperature-induced changes in η_m should have the largest effect on the mantle Rayleigh number, we vary η_m to generate a range of values for Ra_m.

The maximum plate thickness and the lithosphere viscosity are somewhat difficult to estimate and may have been different in the past. To see how heat transport is affected by each of these quantities, we plot N as a function of Ra_m for various combinations of h_m and η_l (Figures 5.1 to 5.3), but show results for "preferred" values in Figure 5.1. We estimate $h_m = 100$ km by assuming plates achieve their maximum thickness after 80 Ma. The effective viscosity of the bending lithosphere is likely the result not only of viscous flow, but also of plastic and brittle deformation associated with non-Newtonian creep and faulting. *Conrad and Hager* [1999a] estimate this effective viscosity to be of order $\eta_l = 10^{23}$ Pa s, a value a few orders of magnitude

stiffer than estimates for the mantle. We vary L between 2500 and 10000 km, the approximate range of plate lengths on Earth. To calculate N, we first use (5.4) to calculate h_s. If $h_s > h_m$, we use (5.6) to calculate N, otherwise, we use (5.2).

The functional dependence of N on Ra_m varies differently for different plate lengths, L. For small Ra_m, the N versus Ra_m curve has a slope of $\beta \sim 1/2$ (Figure 5.1), as predicted by (5.8). For large Ra_m, long plates show a slope of $\beta \sim 0$ while short plates give $\beta \sim 1/3$ (Figure 5.1). The divergence between these two behaviors occurs because short plates do not thicken sufficiently for the bending resistance to become large. Long plates, on the other hand, do have time to thicken, so their bending resistance becomes large enough to limit plate velocities if $Ra_l < 8\sqrt{\pi}L/l_s$. Thus, for weak lithosphere leading to large Ra_l (Figure 5.2a), plates of all lengths are governed by $\beta \sim 1/3$, as predicted by (5.7) for weak bending resistance. If the lithosphere is strong (Figure 5.2b), plates of all lengths are dominated by the bending resistance, which yields $\beta \sim 0$, as predicted by (5.9). Intermediate behavior occurs for $\eta_l = 10^{23}$ Pa s, for which only long plates show $\beta \sim 0$ (Figure 5.1).

If the mantle cools, the negative buoyancy of the lithosphere decreases (assuming constant surface temperatures). As a result, the lithosphere should become less unstable; if small-scale convection erodes the lithospheric base, it should become less effective as the mantle cools, increasing h_m. Thicker plates produce smaller heat flow, so larger h_m should lead to smaller N (compare N for long plates in Figures 5.1 and 5.3). Thus, if h_m increases with decreasing Ra_m, β must be greater than zero. By comparing the values of N for $h_m = 50$ and 100 km (Figures 5.3a and 5.1), we estimate that if h_m were to double due to a decrease of Ra_m by two orders of magnitude, we would obtain $\beta \sim 0.15$. Such a large increase in h_m seems unlikely, so β should probably be somewhat less than 0.15.

Cooler mantle temperatures produce larger mantle viscosities, but may also increase the effective viscosity of the lithosphere. If bending stresses control convection, N decreases with increasing η_l, leading to $\beta > 0$ (compare N for long plates in Figures 5.1 and 5.2b). An increase in η_l could also cause bending stresses to suddenly become important if Ra_l becomes smaller than $8\sqrt{\pi}L/l_s$ (compare curves for long
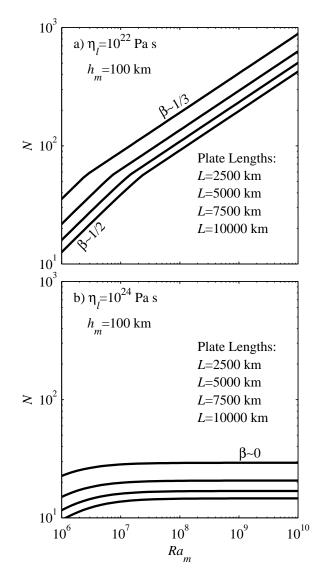
Figure 5.2: N vs. Ra_m curves for $h_m = 100$ km and (a) $\eta_l = 10^{22}$ Pa s or (b) $\eta_l = 10^{24}$ Pa s. To examine a full range of lithospheric viscosities, η_l, compare to Figure 5.1.
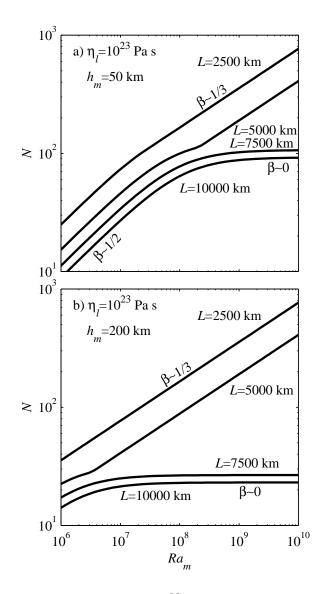
Figure 5.3: N vs. Ra_m curves for $\eta_l = 10^{23}$ Pa s and (a) $h_m = 50$ km or (b) $h_m = 200$ km. To examine a full range of plate thicknesses, h_m, compare to Figure 5.1.

plates in Figures 5.2a and 5.1). Thus, mantle cooling could cause bending slabs to become more effective at resisting convection than the shearing mantle, causing β to drop from 1/3 to 0 at some time.

5.5   Conclusions

We have shown that if plate bending in subduction zones is important in controlling plate velocities, heat flow is less sensitive to interior viscosities and temperatures, so more rapid changes in mantle temperature are permitted. Thus, plate bending could be a mechanism by which values of β less than 0.1 can be obtained in the relationship $Nu \sim Ra^{\beta}$. As discussed above and by *Christensen* [1985], β must be small to reconcile parameterized convection models with the observed secular cooling of the mantle, expressed by estimates of Urey ratios near 0.4. The plate bending mechanism is preferable to other models of convection that yield $\beta < 0.1$ because it arises naturally from subduction zone geometry, and thus produces realistic plate and slab behavior.

We conclude that small values of β are only achieved if bending subducting slabs have an effective viscosity of order 10^{23} Pa s or more. For viscosities close to this value, only an earth with large, Pacific-sized plates that can grow sufficiently thick before they subduct will experience a decrease in β. Because the plate thickness is so influential in determining plate velocities, any process that affects plate thickness could be an essential aspect of plate tectonics, and could greatly influence Earth's thermal evolution. Such processes may include small-scale convection beneath plates, which may limit the plate thickness through basal erosion, or the physical details of subduction initiation, which may control how large, and thus how thick, plates can become.