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Subduction Zone Thermal Models: Numerical Modeling and Data Analysis

by

Saijin Huang

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

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ABSTRACT

Subduction Zone Thermal Models: Numerical Modeling and Data Analysis

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Temperature distribution in subduction zones is a key factor in understanding magma generation, earthquake occurrence, metamorphic reactions, and geochemical element recycling. In this thesis, I have developed a new numerical method to model subduction zone thermal structures with real slab geometries. The new method couples the finite-difference method with the finite-element method to solve the conduction-convection heat transfer system involved in a subduction zone. The mantle and wedge convection is simulated with the finite-element method with the incoming slab convergence rate imposed as a boundary condition at the slab surface. The heat transfer problem is solved with the finite-difference method. It uses a staggered-grid discretization approach so that the effect of the mantle and wedge convection on the thermal model can be accurately taken into account. With material averaging, the simulation method can easily handle a curved slab with a simple, structured grid. The thermal structures that I have calculated for the east Aleutian subduction zone show that a two-segment slab model can overestimate
the slab surface temperature at 100 km by up to 90 °C for a 75-km thick overriding lithosphere.

The thermal structures of ten subduction zones around the Pacific Rim have been generated including the east Aleutians, the Cascades, the Central and South America, the northeast Japan, and Mariana subduction zones. I have shown that the predicted slab surface temperature at 100 km is correlated with B/Zr ratio. As the slab surface temperature increases, the B/Zr ratio decreases systematically for two different levels of B-enrichment (5 ppm and 10 ppm). The results show that the slab tip temperatures have little correlation with the slab lengths and depths.
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CHAPTER 1

Introduction

Subduction zones (SZs) are an important class of plate boundaries where oceanic lithosphere is thrust beneath overriding plates of oceanic or continental lithosphere and penetrate deep into the upper mantle. Deep plate boundary seismicity, volcanic arc magmatism, high-pressure metamorphism, and chemical recycling take place at or near SZs. The understanding of these geological and geochemical processes strongly depends on understanding the thermal structure. This dissertation presents an attempt to refine such thermal models and to compare the structure of widely varying types of convergent margins.

Most thermal models previously reported in the literature have been derived from relatively simple geometric or numerical models of the conduction-convection heat transfer system involved in a subduction zone. McKenzie (1969), Oxburgh and Turcotte (1970), Honda (1985), Peacock (1991), and Davies and Stevenson (1992), among others, considered only simple, fixed-angle slab models to simulate the thermal fields of curved slabs. The fixed-angle models are adequate for the shallow portion of many slabs, but for most subduction zones, such models cannot accurately track slab surface temperature to depths on the order of 100 km or more. Improved models have been reported that use two or more segments to mimic a curved slab (Ponko and Peacock, 1995; Sydora, 1977). Such models
can in principle approximate a real slab geometry if a sufficiently large number of segments are used. Other approximation models have also been reported including the part-of-a-circle slab model used by van den Beukel and Wortel (1988). However, without also having an accurate simulation of mantle convection flow-field for such a slab geometry, the modeled thermal field will not be accurate. Peacock and Wang (1999) modeled Southwest and Northeast Japan SZs using curved slab for shallow depth (less than 60 km) but assumed constant slab dips for deeper depths. Wang et al. (1995) modeled a curved slab geometry, but did not consider mantle convection; hence their modeling is limited to shallow depths (less than 100 km).

In Chapter 2, I present a new numerical technique for simulating subduction zone thermal structure in which both natural slab curvature and mantle convection are considered. This new method couples the finite-difference method for simulating heat transfer with the finite-element method for calculating the mantle convection flow-velocity field. Both methods can treat a continuously curved slab. Shear heating effects are also included. A number of thermal models have been constructed for the east Aleutian subduction zone to evaluate the effects of different assumptions concerning shear heating and thickness of the overriding lithosphere - both of which are important heat sources, but poorly constrained. A comparison between the effects of curved slab and simpler slab geometries also presented.

In Chapter 3, I further investigate uncertainties in the simulated thermal model arising from errors or uncertainties in important subduction parameters. These
include radiogenic heating in the continental crust, slab age and convergence rate. This work extends the study by Peacock (1996) which was based on a fixed angle slab model. I have analyzed the sensitivity of the slab surface temperature to radiogenic heating, the slab age and the convergence rate. The results provide an assessment of the reliability of the model SZ thermal structures and, in particular, the slab surface temperature which has significance for processes of SZ seismicity and magma genesis.

Finally, in Chapter 4, I compare thermal structures of a variety of SZs characterized by different geometry, convergence rate, slab age, and other subduction parameters. Ten representative transects across the South, Central and North America, Aleutian, Kuriles, Northeast Japan and the Marianas SZs were modeled. Slab surface temperatures (SSTs) at 100 km depth vary by a factor of two, but none reach conditions whereby slab melting is likely to occur. Thus, an important influence of slab thermal structure is its impact on metamorphism and dehydration of the slab, fluid transfer of material to the mantle wedge, and possibly dehydration effects on Wadati-Benioff zone seismicity. These effects may be manifested in enrichment of water and the fluid-mobile elements (e.g. boron) in volcanic arc magmas. Interesting relations between the model slab surface temperatures (SSTs) at a reference depth of 100 km (i.e. below the volcanic fronts) and B/Zr ratio have been observed. Also discussed in the chapter is the relation between SST and distribution of SZ seismicity (e.g., 'slab length').
CHAPTER 2

Subduction Zone Thermal Modeling: A Finite-Difference and Finite-Element Approach

ABSTRACT

Accurate prediction of subduction zone temperatures is essential to evaluate the influence of subduction zone (SZ) thermal structure in controlling distribution of earthquakes, magma generation, metamorphism and other geological processes associated with SZs. I have developed a hybrid numerical technique to simulate the thermal structure of a SZ using realistic slab curvatures constrained by observed earthquake distributions. Briefly, the mantle convection velocity is computed using a finite-element method and the temperature field is then modeled using a staggered-grid finite-difference method. An effective-medium technique is used to model a curved slab with a rectangular grid. Depth- and temperature-dependent shear heating is also included in the model. Using parameters for the Eastern Aleutians island arc, the effects of slab geometry on SZ thermal models was assessed by comparing the continuously curving slab geometry with the model of Ponko and Peacock (1995) which approximates slab geometry using two linear segments.

My numerical experiments suggest that the latter model can overestimate the slab surface temperature (SST) at 100 km depth by up to 90 °C for a 75-km thick overriding lithosphere. I also show that high slab-surface shear heating results in
SSTs as much as 280 °C and 230 °C higher than those with no and with weak shear heating at 100 km, respectively. The weak and strong shear heating is described by wet quartzite and wet dunite rheologies, respectively. As shown by earlier workers (e.g. Peacock, 1996), for a given SZ, SSTs decrease with time as subduction proceeds. Blueschist metamorphic facies temperatures are likely to be reached at an early stage (i.e. near inception) of subduction assuming moderate shear heating (e.g. wet quartzite rheology as a lower limiting case). For the eastern Aleutian model, higher shear heating contributions (e.g. wet dunite rheology) could induce slab melting but it is unclear whether such shear stresses are realistic.

INTRODUCTION

Temperature distribution is a key factor in studying major geological processes associated with SZs. Generation of magma, occurrences of earthquakes, and metamorphic reactions are all controlled by temperature in one way or another. For instance, the maximum depth of SZ earthquakes depends strongly on temperature because of its influence on the brittle deformation of crust and upper mantle materials. However, the understanding of SZ thermal structure is hindered by the lack of direct measurements at depths.

In attempts to quantify the relations between thermal structure and these processes, numerous investigators have made numerical simulations of SZ thermal structure. The complexity of the problem has, until recently, limited most efforts
to consider simplified slab geometry such as fixed dip-angle slab models (McKenzie, 1969, Oxburgh and Turcotte, 1970, Honda, 1985, Peacock, 1991, Davies and Stevenson, 1992). The fixed dip angle models are adequate for gently dipping slabs. However, distributions of Benioff zone earthquake hypocenters indicate that slab dip commonly varies with depth and that geometry varies significantly from one SZ to another. A slab typically dips gently at shallow depths and bends rapidly at intermediate depths. Sydora (1977) proposed an improved model that can simulate slab geometry by using several segments to approximate the observed curvature. Similarly, Ponko and Peacock (1995) approximated a curved slab using two segments of constant dip. Furlong and Chapman (1982) also attempted to simulate a curved slab as a flat slab model by defining the temperature on the upper surface of the slab according to the actual path of a curved slab. While allowing for relatively straightforward implementation of a finite-difference numerical solution, this approach does not represent a real slab geometry and requires a top boundary condition which is not known. The part-of-a-circle model used by van den Beukel and Wortel (1988) has limitations of modeling other slab geometry. Recently, Wang et al. (1995) used a general curved slab geometry, but did not consider mantle convection, hence their simulation is limited to shallow depths (less than 100 km). Because of the variety of approaches, it has not been possible to quantitatively compare thermal structures between subduction zones exhibiting different geometries. Also, because SZ characteristics vary considerably worldwide, any generic
approach to modeling thermal structures cannot accurately reflect the wide differences observed in slab age, convergence rate, geometry, etc. To facilitate such comparisons, I have developed a flexible modeling approach that can incorporate such differences.

The aim of this study is to describe the modeling technique and to discuss my results in the context of a single SZ - the Eastern Aleutians arc which has been studied previously by several investigators. I use the finite element method to calculate mantle convection velocities for a curved slab model and then solve for the thermal structure using a staggered-grid finite-difference method. Preliminary simulation results have been presented in Huang et al. (1999). In the following sections, I discuss several issues involved in the numerical simulation including the calculation of mantle convection velocity and the treatment of a curved slab. My focus is on the influence of slab geometry, shear heating, and overriding lithosphere thickness on the slab surface temperature.

THEORY

Model Description

The model for subduction-zone temperature computation is schematically described in Figure 2.1. In the model, oceanic lithosphere subducts beneath either an oceanic lithosphere or a continental margin. Subduction induces both mantle convection beneath the subducting slab and in the wedge corner beneath the over-
riding plate. Subduction rate is assumed to be constant along the entire slab. Also shown in Figure 2.1 are the types of temperature boundary conditions and heat sources used in the numerical simulation. Details will be given in the subsequent sections.

**Temperature Computation**

The governing equation for the temperature field in a convecting material is

$$\nabla \cdot k \nabla T - \rho C v \cdot \nabla T + S = \rho C \frac{\partial T}{\partial t}. \quad (2.1)$$

where $T$ is temperature, $k$ the heat conduction coefficient, $\rho$ the rock density, $C$ the heat capacity, $v$ the material movement velocity, which includes slab movement and mantle convection, and $S$ the heat source intensity. The initial thermal structure of the subducting plate is based on the GDH1 model of Stein and Stein (1992) for an oceanic crust age. The temperature at the sea floor is taken to be that of sea-water ($0$ °C). Two types of boundary conditions have been used for the side and bottom of our model space. The first condition determines the temperature at an initial value given by

$$T(t = 0, x, z) = T_0(z). \quad (2.2)$$

where the $z$-axis points downward. Secondly, we assume that no heat flow radiates across the boundaries. The selection of initial temperature and the influence of the two different boundary conditions on the subduction zone thermal structure will be examined later.
Figure 2.1: A schematic model for subduction zone thermal modeling. $T$ and $V$ stand for temperature and subduction rate, respectively.
Now I discuss the numerical solution of equation (2.1) using the finite-difference method. The model shown in Figure 2.1 is discretized with a rectangular grid as illustrated in Figure 2.2 with \( m \) cells in the \( x \)-direction and \( n \) cells in the \( z \)-direction. A grid node \((i, j)\) is defined at the coordinates \((x_i, z_j)\) and a cell \((i, j)\) is defined by the area \((x_i \leq x < x_{i+1}, z_j \leq z < z_{j+1})\). The temperature field is defined at the center of a cell, as shown in Figure 2.3. The \( x \)-component of the mantle convection velocity, \( v_x \), is defined at the \( z \)-directed grid edge middles and the \( z \)-component convection velocity, \( v_z \), is defined at the \( x \)-directed edge middles. I assume for simplicity that the grid is equally spaced. Formulae for a non-uniform grid can be written without extra difficulty. The finite-difference equation is obtained for each cell by seeking a solution to the weak form of equation (2.1)

\[
\langle \phi_{ij}, \nabla \cdot k \nabla T \rangle - \langle \phi_{ij}, \rho C \nabla \cdot \nabla T \rangle = \left\langle \phi_{ij}, \rho C \frac{\partial T}{\partial t} \right\rangle + \langle \phi_{ij}, S \rangle. \tag{2.3}
\]

where \( \phi_{ij}(x, z) = 1 \) for \( x \) and \( z \) in cell \((i, j)\); otherwise, \( \phi_{ij}(x, z) = 0 \). The inner product in equation (2.3) is defined as

\[
\langle f, g \rangle = \frac{1}{S_{ij}} \int_{S_{ij}} fg ds. \tag{2.4}
\]

where \( S_{ij} \) is the area of the cell \((i, j)\). Equation (2.3) dictates that heat is conserved cell-wise. Now consider the cell \((i, j)\) (the shaded area in Figure 2.3). By using the Gauss theorem, the first term in equation (2.3) can be approximated as

\[
\langle \phi_{ij}, \nabla \cdot k \nabla T \rangle = \frac{1}{\Delta^2} \int_{S_{ij}} ds \nabla \cdot k \nabla T = \frac{1}{\Delta^2} \oint k \nabla T \cdot ndl
\]
Figure 2.2: A sketch of the finite-difference grid for discretizing a subduction zone.

The thick lines represent the top and bottom boundaries of the subduction zone, respectively.
Figure 2.3: A staggered grid for discretizing temperature ($T$) and mantle convection velocity ($v_x$ and $v_z$). The $x$-axis points to the right and the $z$-axis points downward. $S_{ij}$ is the area of the shaded cell.
\frac{1}{\Delta^2} \left[ \overline{\kappa}_{i-\frac{1}{2}J+\frac{1}{2}}T_{i-\frac{1}{2}J+\frac{1}{2}} + \overline{\kappa}_{i+\frac{1}{2}J+\frac{1}{2}}T_{i+\frac{1}{2}J+\frac{1}{2}} + \overline{\kappa}_{i+\frac{1}{2}J-\frac{1}{2}}T_{i+\frac{1}{2}J-\frac{1}{2}} + \overline{\kappa}_{i+\frac{1}{2}J-\frac{1}{2}}T_{i+\frac{1}{2}J-\frac{1}{2}} \right. \\
\left. - \left( \overline{\kappa}_{i-\frac{1}{2}J+\frac{1}{2}} + \overline{\kappa}_{i-\frac{1}{2}J-\frac{1}{2}} + \overline{\kappa}_{i+\frac{1}{2}J+\frac{1}{2}} + \overline{\kappa}_{i+\frac{1}{2}J-\frac{1}{2}} \right)T_{i+\frac{1}{2}J-\frac{1}{2}} \right]. \tag{2.5}

where \Delta is the grid spacing, and \( n \) the outward normal vector of the cell boundary.

In equation (2.5), \( \overline{\kappa}_{i-\frac{1}{2}J+\frac{1}{2}} \) denotes the average heat conduction coefficient of the cell \((i, j)\), and so on. The term involving \( \nu \) in equation (2.3) can be integrated as

\[ \langle \phi, \rho C \nu \cdot \nabla T \rangle = \frac{1}{\Delta^2} \int_{S_{ij}} \rho d \nu \cdot \nabla T = \frac{1}{2\Delta} \overline{\rho}_{i-\frac{1}{2}J+\frac{1}{2}} \overline{C}_{i-\frac{1}{2}J+\frac{1}{2}} \times \]

\[ \left( \left( T_{i-\frac{1}{2}J+\frac{1}{2}} - T_{i-\frac{1}{2}J-\frac{1}{2}} \right) \nu_{z(i+1,j+\frac{1}{2})} + \left( T_{i-\frac{1}{2}J-\frac{1}{2}} - T_{i-\frac{1}{2}J+\frac{1}{2}} \right) \nu_{z(i+1,j-\frac{1}{2})} \right) \]

\[ \left( T_{i-\frac{1}{2}J-\frac{1}{2}} - T_{i-\frac{1}{2}J+\frac{1}{2}} \right) \nu_{z(11-i,j+\frac{1}{2})} + \left( T_{i-\frac{1}{2}J+\frac{1}{2}} - T_{i-\frac{1}{2}J-\frac{1}{2}} \right) \nu_{z(11-i,j-\frac{1}{2})} \right]. \tag{2.6}

In the above equation, \( \overline{\rho}_{i-\frac{1}{2}J+\frac{1}{2}} \) and \( \overline{C}_{i-\frac{1}{2}J+\frac{1}{2}} \) are the average density and heat conduction coefficient of the cell \((i, j)\), respectively.

By combining equations (2.5) and (2.6) into equation (2.3), I obtain an ordinary differential equation

\[ D \frac{dT}{dt} = AT + b. \tag{2.7} \]

where \( D \) is an \( N \times N \) diagonal matrix, \( T \) the unknown temperature vector, \( A \) an \( N \times N \) matrix with at most five nonzero entries at each row, and \( b \) the heat source vector. Here, \( N \) is the number of temperature unknowns. Many different methods can be applied to solve equation (2.7), including the explicit time-stepping method and the implicit time-stepping method. The explicit time-stepping method results by replacing the time derivative in (2.7) with a forward difference whereas the
implicit time-stepping method results by replacing the derivative with a backward difference. The implicit method needs to solve a linear system of equations at each time step, which can be computationally intensive. For detailed discussion of time-stepping methods, the reader is referred to Mitchell and Griffiths (1980). For this study, we chose to use the explicit time-stepping method because of its simplicity and efficiency. By equating equation (2.7) at the time instant \(k\Delta t\), where \(k\) is the number of time steps and \(\Delta t\) is the time step size, and replacing the first derivative with the forward difference, we obtain

\[
T^{(k+1)} = T^{(k)} + \Delta t D^{-1}(AT^{(k)} + b^{(k)}). \tag{2.8}
\]

In other words, given the temperature and heat source at time \(n\Delta t\), a temperature at time \((n + 1)\Delta t\) can be updated that is in turn used to update the temperature at time \((n + 1)\Delta t\), and so on. At each time step, no matrix inversion is needed. However, to prevent the explicit time-stepping process from diverging the time step must be bounded. Considering both the conduction and convection effects, we use a time step as follows

\[
\Delta t = \min \left( \frac{\Delta_{\min}}{2 \nu_{\max}}, \frac{\rho_{\min} C_{\min} \Delta^2_{\min}}{4 k_{\max}} \right), \tag{2.9}
\]

where the subscripts \(\text{min}\) and \(\text{max}\) stand for the minimum and maximum values of the pertinent variables, respectively. The first term on the right size of equation (2.9) is the time step constrained by the mantle convection effects and the second term is the time step constrained by the heat conduction effects.
Mantle Convection Velocity

Effects of mantle convection on subduction zone thermal structure are taken into account by substituting proper mantle convection velocity in equation (2.1). The mantle convection velocity is governed by momentum and mass conservation equations

\[-\nabla \cdot \nu \nabla \mathbf{v} + \frac{1}{\rho} \nabla P = g.\]  

(2.10)

and

\[\nabla \cdot \rho \mathbf{v} = 0.\]  

(2.11)

In equation (2.10), \(\mathbf{v}\) is the velocity, \(\nu\) the viscosity coefficient, \(g\) the gravitational potential and \(P\) the pressure. Equations (2.10) and (2.11) are solved using a 2-D finite-element method (Kloucek and Rys. 1994). The convection velocities above and below the subducting slab are solved separately. Each volume is discretized with rectangular patches. All the boundaries in the model shown in Figure 2.1 are assumed to be impermeable. The subducting slab acts as a driving force with the the upper surface of the slab subducting at a constant rate. To minimize the influence of the boundary effects, I place the boundary far away from the volume of interest for temperature modeling. An example of the velocity field simulated for a straight slab is shown in Figure 2.4 as compared to that obtained with an analytical solution (Batchelor. 1967). No gravitational force has been included in numerical solutions since the analytical solution does not have this term. A very good agreement is observed between the numerical and analytical solutions.
Figure 2.4: The convection velocities calculated with (a) an analytical solution (Batchelor, 1967) and (b) the finite-element solution (Kloucek and Rys. 1994). The difference between the two is shown in (c). The slab dip angle is 50°.
Modeling a Curved Slab

When a slab curves, it becomes impossible to precisely model the slab geometry with a rectangular grid. A typical approach to such a problem is to use a stair-case approximation. That is, a curved slab boundary is approximated with piecewise straight segments. Clearly, this approximation is accurate only if the grid is sufficiently fine. Otherwise, the approximation may lead to a numerically 'broken' slab. A more effective method for modeling a curved slab is by the effective medium approach. Consider the diagram shown in Figure 2.5. Assume that point A is slightly below the slab boundary and point B is slightly above the slab boundary. Instead of setting \( v_z = v \cos \theta \) at point A and \( v_z = 0 \) at point B as for the stair-case approximation, where \( v \) is the subducting rate and \( \theta \) is the angle between the slab boundary at point B and the horizontal line. I take \( v_z \) and \( v_z \) to be the average velocities of the respective cells. For instance, \( v_z \) at point B is taken to be the average value of \( v_z \) in the shaded area. By doing so, the actual path of the slab boundary relative to the grid becomes less critical to the accuracy of the numerical results. Also, the grid spacing can be made relatively large.

In practice, the grid density should change according to the curvature of the subducting slab. A denser grid must be used where the slab curves rapidly. Based on a numerical convergence test (see Appendix), a minimum grid size of 2.5 km is used for the subsequent examples. The largest grid used is 10 km.

Initial Temperature
Figure 2.5: Diagram showing the treatment of a curved slab in the finite-difference simulation. The thick line represents the slab surface.
The initial temperature field has great impact on the calculated thermal structure of a subduction zone, especially in the mantle wedge region. Higher initial temperature will result in a higher wedge temperature. On the other hand, a larger initial temperature gradient will result in relatively higher temperature in the wedge at greater depths.

In my simulation, the initial temperature for oceanic and continental crusts are allowed to be different. The initial temperature for the oceanic crust is estimated from Stein and Stein's (1992) GDH1 model

$$T_0(z) = T_m \left( \frac{z}{a} + \sum_{n=1}^{\infty} c_n e^{-J_n z/a} \sin \frac{n\pi z}{a} \right)$$

(2.12)

where $x = \nu t$, $t$ is the age of the oceanic crust, $c_n = 2 / (n \pi)$. $J_n = (R^2 + n^2 \pi^2)^{1/2} - R$. $R = \nu a / (2 \kappa)$. $a$ is the thermal plate thickness. and $T_m$ is the basal temperature. In Stein and Stein's GDH1 model, thermal plate thickness $a$ is 95 km. basal temperature $T_m$ is 1450 °C. The initial temperature for the continental crust depends on radiogenic heat contributions. Following Fowler (1990), I use a two-layer steady-state conduction model to calculate the initial temperature for the continental crust

$$T_0(z) = -\frac{A_1}{2k} z^2 + \left[ \frac{Q_0}{k} + \frac{A_2}{k} (z_2 - z_1) + \frac{A_1 z_1}{k} \right] z.$$  

(2.13)

for $0 \leq z < z_1$ and

$$T_0(z) = -\frac{A_2}{2k} z^2 + \left( \frac{Q_0}{k} + \frac{A_2 z_2}{k} \right) z + \left( \frac{A_1 - A_2}{2k} \right) z_1^2.$$  

(2.14)

for $z_1 \leq z < z_2$. In the above equations, $z_1$ and $z_2$ are the bottom depths of the upper and lower layers in the two-layer model, $Q_0$ is the basal heat flux. $A_1$ and $A_2$
are the radiogenic heat generation in the upper and lower layers, and \( k \) is the heat conductivity.

An example of the initial temperature field for a continental crust and a 50-Ma old oceanic crust is shown in Figure 2.6. The temperature in the continental crust below 105 km is determined by mantle adiabatic gradient (i.e., 0.3 °C/km). Above that depth, the temperature varies roughly linearly with depth. On the other hand, the thermal gradient of the oceanic crust increases slightly with depth to about 95 km, and then stays at the constant adibatic gradient below 95 km.

**Heat Sources**

Two types of heat sources are considered: radiogenic heating and shear heating. Radiogenic heat derives from decaying radioactive elements which mainly reside in the continental crust. Important radioactive elements include uranium, thorium and potassium (Fowler, 1990). In this study, the radiogenic heat generation is assumed to be \( 0.6 \times 10^{-6} \) W/m\(^3\) within the entire continental crust. Lower (\( 0.5 \times 10^{-6} \) W/m\(^3\)) and higher (\( 0.68 \times 10^{-6} \) W/m\(^3\)) radiogenic heating intensities have been used by Ponko and Peacock (1995).

Shear heating results when a solid plate subducts beneath a solid lithosphere. The shear heating intensity depends on the subducting plate velocity \( (v) \) and the shear stress \( (\tau) \). \( Q = \tau v \). In the brittle regime (shallow depths), shear stress increases linearly with lithospheric pressure. However, in the ductile regime (large depths) the shear stress decreases with increasing temperature according to a power-
Figure 2.6: The initial temperature field used for the thermal modeling of the east Aleutian transect. The radiogenic heat for the continental crust (35 km and above) $A_1$ is $0.6 \, \mu Wm^{-3}$ and the basal heat flux for the continental lithosphere $Q_0$ is 40 mWm$^{-2}$. The oceanic crust is assumed to be 50 Ma old. Other parameters used in equations (2.12)-(2.14) are: $a = 95$ km, $T_m = 1450$ °C, $A_2 = 0 \, \mu Wm^{-3}$, $z_1 = 35$ km, $z_2 = 105$ km, and $k = 3.14$ Wm$^{-1}$s$^{-1}$. The dashed line indicates the slab surface.
law rheology (Kirby, 1983)

\[ \varepsilon = A_0 (\sigma_1 - \sigma_3)^n e^{-Q/RT} \]  
(2.15)

where \( \varepsilon \) is the strain rate, \( T \) the absolute temperature, \( R \) the gas constant, \( \sigma_1 \) and \( \sigma_3 \) are the maximum and minimum principal stresses, and \( Q \). \( A_0 \) and \( n \) are material constants. Here, \( \varepsilon \) is taken to be \( \nu_c/\nu_{ref} = 10^{-12} \text{s}^{-1} \) (van den Beukel and Wortel, 1988) with \( \nu_{ref} \) being a reference velocity of 8 cm/yr.

I'll consider three different rheologies. Zero shear heating represents the lowest extreme. Wet quartzite represents a relatively weak rheology and wet dunite represents a stronger rheology. The strongest possible rheology would be dry dunite. Because water from dehydration of the subducting slab will affect the overlying mantle, wet dunite rheologies are considered more realistic. Wet quartzite and wet dunite represent extreme lithologies present in subducted lithosphere with the former more closely simulating altered oceanic crust and sediments near the upper surface. are typical of portions of the oceanic crust. The flow law constants for wet quartzite are \( A_0 = 2.19 \times 10^4 \text{Kbar}^{-n} \text{s}^{-1} \), \( n = 2.44 \), \( Q = 38.2 \text{Kcal/mole} \) (Koch et al., 1980) and for wet dunite are \( A_0 = 6.31 \times 10^6 \text{Kbar}^{-n} \text{s}^{-1} \), \( n = 2.4 \), \( Q = 80 \text{Kcal/mole} \) (Carter and Ave Lallemant, 1970). Figure 2.7 shows shear stress as a function of temperature for these rheologies. The shear strain rate used is \( 6.25 \times 10^{-13} \). The shear stress intensity decreases rapidly with increasing temperature at low temperatures and becomes negligibly small at temperatures above 400 and 900 °C for wet quartzite and wet dunite, respectively. In the following section.
Figure 2.7: The ductile-regime shear stress as a function of temperature for wet quartzite and wet dunite rheology.
I demonstrate the temperature fields resulting from these different shear heating intensities.

**NUMERICAL RESULTS**

In this section, I first verify the finite-difference modeling algorithm outlined above. Then I present several examples of the subduction-zone thermal structure under different conditions.

It appears that verifying the finite-difference results is a non-trivial task because of the lack of other numerical solutions. As a first check, I have compared my model results with an approximate analytic solution (Molnar and England, 1990). This model assumes a fixed-angle, straight slab with a constant convergence velocity. Even though this model does not consider mantle wedge convection, it has been used by Harry and Green (1999) to model Cascades and Central American SZs. The temperature along the upper slab surface at steady state is given by

\[
T(z) = \frac{(Q_0 + \sigma V) z / K}{1 + b \sqrt{z V} \sin \delta / \kappa}.
\]  

(2.16)

where \(Q_0\) is the basal heat flux, \(\sigma\) the shear stress exerted on the slab, \(V\) the convergence rate, \(K\) the thermal conductivity coefficient, \(\kappa\) the thermal diffusivity and \(\delta\) the dip angle of the slab. As shown in Figure 2.8, the temperature fields calculated with the analytic model and with my finite-difference method are in excellent agreement. Ponko and Peacock (1995) also reported close agreement
Figure 2.8: Comparison of the finite-difference solution to an analytic solution (Molnar and England, 1990). The parameters used for the analytic solution [equation (2.16)] are $\delta = 6^\circ$, $Q_0 = 0.048$ W/m$^2$, $V = 5$ cm/yr, $K = 3.138$ W/mk, $b = 92$, and $\kappa = 8.05 \times 10^{-7}$ m$^2$/s.
between the analytical solution and their thermal model at shallow depth ($P < 1$ GPa). At greater depth, their temperature is slightly higher than the analytical solution and my result; this is possibly due to the fact that they did not compute for a long enough time. To mimic the analytic model, I used a straight slab and have assumed in my finite-difference model that the left grid boundary is 200 km from the junction point between the upper slab surface and the sea floor surface. All the grid boundaries are assumed to be heat insulating except the upper boundary where temperature is fixed at 0 °C. To reiterate, this simple analytic model does not include effects of convective heating in the mantle wedge - which becomes significant at sub-lithospheric depths.

I now discuss the simulated thermal structure for the eastern Aleutian subduction zone. Figure 2.9 shows the slab surface geometry and the velocity field simulated with the finite-element method. Slab geometry is derived from the Benioff zone earthquake hypocenter data. The curved slab bends most rapidly between 50 km and 100 km. The incoming slab is 46 Ma old. The convergence rate is 5 cm/yr. The overriding lithosphere is assumed to be 100 km thick. The left and right boundaries for calculating the down-going and overriding velocity fields are placed at -2000 km and 2500 km, respectively. The bottom boundary is at 2000 km. The densities of the mantle and the crust are both assumed to be 3.3 g/cm$^3$. The mantle wedge convection is clearly seen in Figure 2.9.

Figure 2.10 shows the simulated temperature data when shear heat is assumed
Figure 2.9: The mantle convection and wedge flow velocity for the Alaska subduction zone. The dashed line indicates the upper surface of the subducting slab.
to be zero. The subduction duration time is 60 Ma. Note, that below 100 km where the wedge convection starts there is a large thermal gradient right above the slab surface. The largest thermal gradient resides near the wedge corner. It is worthwhile to mention that the initial thermal model for the overriding lithosphere includes a basal heat flux of \(40 \times 10^{-3}\) W/m\(^2\) which accounts for the heat transfer from the underlying hotter convecting wedge. The same amount of basal heat flux has been added as an upward heat source in the finite-difference simulation. The addition of the basal heat flux is necessary to maintain the initial thermal gradient in the upper wedge when subduction does not take place.

The evolution of the slab surface temperature with time is shown in Figure 2.11. The slab-surface temperature decreases most rapidly in the earlier stage of subduction. By 60 Ma, the slab-surface temperature has approached the steady state. Figure 2.11 shows that the slab surface temperature barely reaches into lawsonite blueschist metamorphic facies at about 5 My after initiation of subduction for eastern Aleutian subduction zone. In this case, slab surface temperature never reaches eclogite metamorphic facies conditions which may be considered typical of subduction zone metamorphism. Clearly, for such conditions to be attained, other heat contributions are needed.

In the following subsections, I illustrate the effects of shear heating, the lithosphere thickness, and slab geometry on the thermal structure and, particularly, the temperature along the slab surface.
Figure 2.10: The temperature field of the eastern Aleutian transect. The overriding lithosphere is 100 km thick. The incoming slab is 46 Ma. old. The convergence rate is 5 cm/yr. No shear heating is included. The subduction duration time is 60 Ma.
Figure 2.11: The temperature on the slab surface of the eastern Aleutian subduction zone as a function of subduction time. WS = wet solidus and DS = dry solidus. Also shown are metamorphic facies: Am = amphibolite. EA = epidote amphibolite. EB = epidote blueschist. Ec = Eclogite, Gs = Greenschist. LB = lawsonite blueschist.
Shear Heating Effect

As discussed before, shear heating contributions in the brittle and ductile regimes depend on temperature-dependent rheologies of appropriate materials. Such contributions have been discussed by Peacock (1996). Here, I use a wet quartzite rheology to represent the realistic shear heating intensity which is in the range assumed by Peacock (1996). Additional models are shown assuming a wet dunite rheology, which may provide a realistic upper limit to shear heating intensity. The brittle to ductile transition for wet quartzite and wet dunite at the slab surface temperature after 50 Ma initiation of subduction is shown in Figure 2.12. The transition takes place at 26 km and 75 km depths for the two different rheologies, respectively. In the brittle regime, the shear stress is assumed to increase linearly with depth (van den Beukel and Wortel, 1987), i.e.,

\[ \tau = \alpha P. \]  

(2.17)

where \( P \) is the lithostatic pressure and \( \alpha \) a proportionality constant which is assumed to be 0.05 through this paper. In the ductile regime, the shear stress decreases exponentially with depth. The maximum shear stresses for wet quartzite and wet dunite at 50 Ma are about 35 MPa and 120 MPa, respectively. These values are in the range assumed by Peacock (1996). Because the transition depth is temperature-dependent, it is updated each time the temperature field is updated in the finite-difference simulation.

Figure 2.13 shows the simulated temperature fields of the eastern Aleutian sub-
duction zone with no, weak and strong shear heating, respectively. It is clear that stronger shear heating results in higher slab temperature. For instance, the maximum depth of the 500°C isotherm without shear heating is about 275 km and it is pushed up to about 250 km and 200 km with weak and strong shear heating, respectively. The P-T paths along the slab surface for different shear heating intensities are shown in Figure 2.14. The slab surface temperature at 100 km depth with strong shear heating is about 280 °C and 230 °C higher than those with no and weak shear heating, respectively. The temperature for low shear heating intensity (wet quartzite) only reaches blueschist and eclogite conditions at an early stage of subduction. In contrast, models with strong shear heating intensity (wet dunite) produce SST P-T profiles that pass through blueschist and eclogite facies at all times.

Temperatures at a short distance above or below the slab surface are also of interest. Figure 2.15 shows the P-T paths at 5 km, 10 km, and 20 km above and below the slab surface. At 20 km below the slab surface, the temperature shows little variation with depth, implying that the isotherm line is nearly parallel to the slab surface. On the other hand, at 5 km above the slab surface, the temperature initially increases with depth to about 110 km (3.56 GPa) and then decreases with depth. The maximum temperature at about 110 km is caused by the heat concentration around the wedge corner (refer to Figure 2.13).

Lithosphere Thickness Effect
Figure 2.12: The shear stress as function of depth for wet quartzite and wet dunite.

Time is 50 Ma.
Figure 2.13: The temperature fields of the eastern Aleutian transect with (a) no shear heating, (b) weak shear heating, and (c) strong shear heating added to the slab surface. Other parameters are the same as in Figure 2.10.
Figure 2.14: The slab-surface P-T paths of the eastern Aleutian transect with zero, weak (wet-quartzite rheology) and strong (wet-dunite rheology) shear heating added to the slab surface, respectively. The metamorphism facies are explained in Figure 2.11.
Figure 2.15: The steady state P-T paths of the eastern Aleutian subduction zone at various distances above and below the slab surface. This model uses weak (wet-quartzite) shear heating.
Mantle wedge convection is induced below the base of the mechanical lithosphere which is defined by a particular isotherm (Parsons and Mckenzie, 1978). Since this mechanical boundary isotherm is not very well known, I compared three different lithosphere thicknesses. Figure 2.16 shows the temperature fields for mechanical lithosphere thickness in 75 km, 100 km and 125 km, respectively. The inside-slab temperature for a 100-km lithosphere does not differ much from that for a 75-km lithosphere. It is, however, apparently higher than for a 125-km lithosphere. For instance, the maximal depth of the 500°C-isotherm for a 100-km lithosphere is at about 230 km compared to 250 km for a 125-km thick lithosphere.

The slab-surface P-T paths for the different lithosphere thicknesses as shown in Figure 2.17 show that the slab surface temperature for a 100-km mechanical lithosphere is always higher than that for a 125-km mechanical lithosphere. The 75 km thick mechanical lithosphere model has higher temperature than that of 100 and 125 km mechanical lithosphere until about 110 km (3.56 GPa). Below that depth, the slab surface temperature for a 100-km thick mechanical lithosphere becomes higher.

**Slab Geometry Effect**

Previous studies have assumed a fixed-angle, straight-slab model in calculating the thermal structure of a curved subducting slab. The straight-slab model applies where the slab dips gently over the depth interval of interest. For those subduction zones whose slabs curve significantly (such as the eastern Aleutian subduction
Figure 2.16: The temperature fields of the eastern Aleutian transect simulated with a (a) 75 km, (b) 100 km, and (c) 125 km lithosphere. Weak (wet-quartzite) shear heating is added to the slab surface.
Figure 2.17: The slab-surface P-T paths of the eastern Aleutian transect for a 75-km, a 100-km, and a 125-km lithosphere, respectively. Weak (wet-quartzite) shear heating is included. The metamorphism facies are explained in Figure 2.11.
zone), the fixed-angle slab model can be a poor representation of the actual slab geometry. To illustrate the effect of the slab geometry on the thermal structure, I show in Figure 2.18 the temperature fields of the eastern Aleutian subduction zone simulated with a curved slab and a two-segment slab (Ponko and Peacock, 1995), respectively. In both models, the mantle wedge convection is assumed to start at 75 km. The two-segment slab model assumes a dip angle of 6°C in the upper part and a dip angle of 45° in the lower part. Note that the curved slab model yields a warmer slab interior. For instance, the 500°C-isotherm line for the curved slab model reaches a maximum depth of 255 km, whereas that for the two-segment slab model reaches a maximum depth of about 290 km. However, the slab surface temperature given by the curved slab model is lower in general than that by the two-segment slab model, as shown in Figure 2.19, except at very shallow depths. The slab surface temperature given by both models is in general agreement up to 2.3 GPa pressure (70 km depth). Below that depth, the temperature given by the two-piece slab model is significantly higher than that given by the curved slab model. The reason that at shallow depth both models predict temperature in small difference while at deep depth two segments model results in higher temperature than that of curved model is that the 6° dip angle segment close to the curved slab, but deep segment in the two segments model has longer path than the curved slab model. At 100 km (3.23 GPa), the slab surface temperature given by the two-segment slab model is about 90 °C higher than that given by the curved
Figure 2.18: The eastern Aleutian transect thermal structures simulated with (a) a curved and (b) a two-segment slab model. All parameters are the same as in Figure 2.10.
slab model.

CONCLUSIONS

I have developed a hybrid numerical model to simulate the thermal structure of a subduction zone. The model allows for curved slab geometry constrained by Mantle and wedge convection is calculated with a finite-element method with the incoming plate convergence rate imposed as the boundary condition at the slab surface. The convection velocity field is then incorporated in the heat transport system to compute the temperature field around a subducting slab. A staggered-grid finite-difference method has been used to march the temperature field from an initial temperature to a desired subduction duration time. Depth and temperature-dependent shear heating has also been included in the model. My results show that a two-segment slab model may overestimate the slab surface temperature at 100 km by up to 90 °C for a 75-km thick overriding lithosphere, as compared to the use of a curved-slab model. On the other hand, the slab surface temperature at 100 km for strong shear heating intensity as represented by the wet dunite rheology can be 280 °C and 230 °C higher than that with no shear heating and for weak shear heating represented by wet quartzite, respectively.

ACKNOWLEDGMENTS
Figure 2.19: The P-T paths for the eastern Aleutian transect simulated with a curve slab and a two-segment slab model, respectively.
I would like to thank Dr. A. Lenardic for useful discussion and interest in the work and Dr. P. Kloucek for the computing facility to simulate mantle and wedge convection with the financial support by the National Science Foundation. Also support from NSF EAR-9725661.
REFERENCES


APPENDIX

FINITE-DIFFERENCE CONVERGENCE TEST

The accuracy of the finite-difference method depends on the grid size of the mesh. In general, a smaller grid size increases the accuracy of the results but meanwhile increases the computational costs. In my simulations, I am mostly concerned with the discretization of the slab surface. Therefore, I use small grids near the slab surface and larger grids for other areas. To select a proper minimum grid size, I performed a numerical convergence test. I started with a minimum grid size of 10 km and successively reduced the grid size by a factor of 2 or so. The calculated slab-surface temperatures for the east Aleutian transect are plotted in Figure 2.20. It appears that a minimum grid size of 2.5 km gives almost identical results to those with a minimum grid size of 1 km. I therefore use a minimum grid size of 2.5 km.
Figure 2.20: The slab-surface temperatures for the eastern Aleutian transect calculated using different minimum grid sizes. Note that a minimum grid size of 2.5 km appears to be adequate.
CHAPTER 3

Thermal Models of the Eastern Aleutians Subduction Zone: An Uncertainty Study

ABSTRACT

Uncertainties exist in the values of subduction zone parameters due to limited geological, geophysical, and geochemical observations. These uncertainties will convert to inaccuracies in numerically simulated subduction zone thermal models. As an attempt to evaluate the impact of the parameter uncertainties, I have studied the sensitivity of the slab surface temperature of the east Aleutians subduction zone to a number of subduction parameters such as radiogenic heating, slab age, convergence rate, and brittle-regime shear heating. I show that radiogenic heating has no effect on the SST at 100 km depth and shear heating in the brittle regime affects the SST mostly at depths above 60 km. A ±20% uncertainty in the slab age or the convergence rate may change the SST by up to 10 °C at depths above 100 km.

INTRODUCTION

In Chapter 2, I have described a coupled finite-difference and finite-element approach to modeling the thermal field of a subduction zone. There, I demonstrated the simulated thermal models for the eastern Aleutians subduction zone with differ-
ent shear heating intensity, lithosphere thicknesses and slab geometries. In reality, the thermal structure of a subduction zone is controlled by many parameters (slab age, radiogenic heating, etc.), in addition to the ones already discussed. Some of these parameters may be measured; others have to be estimated from geological, geophysical, and geochemical observations. The limited amount of observations available prevents us from very accurately determining the values of the parameters. All of the parameter values are subject uncertainties to various degrees. The objective of this chapter is to investigate the sensitivity of subduction zone thermal structure to subduction parameters. The analysis result will help estimate the reliability of the simulated temperature data. To limit the scope of my study, only four parameters will be considered, i.e., radiogenic heating, slab age, convergence rate, and shear heating.

**ANALYSIS OF THE RESULTS**

As described in Chapter 2, the temperature field in a subduction zone is modeled by solving the following equation

\[ \nabla \cdot k \nabla T - \rho C \mathbf{v} \cdot \nabla T = \rho C \frac{\partial T}{\partial t} + S. \quad (3.1) \]

where \( T \) is temperature, \( k \) the heat conduction coefficient, \( \rho \) the medium density, \( C \) the heat capacity, \( \mathbf{v} \) the convection velocity, and \( S \) the heat source intensity. The first and second terms on the left side of equation (2.1) represent heat conduc-
tion and heat convection, respectively. By using a staggered-grid finite-difference method I convert the equation into a linear system of equations

$$D \frac{dT}{dt} = AT + b,$$  \hspace{1cm} (3.2)

where $D$ is an $N \times N$ diagonal matrix containing the medium density and heat capacity data, $T$ the unknown temperature vector, $A$ an $N \times N$ matrix containing the medium heat conduction coefficient, heat capacity and convection velocity data, and $b$ the source vector. Here, $N$ is the number of unknown temperature variables. The convection velocity is solved using a finite-element method (Kloucek and Rys. 1984) with the convergence rate given as the boundary condition at the slab surface.

I will investigate the effects of radiogenic heating, shear heating, slab age, and incoming plate convergence rate on the east Aleutians subduction zone slab surface temperature (SST). The effect of the lithospheric thickness has been discussed in Chapter 2. Table 3.1 lists the nominal values of the east Aleutians subduction parameters. The temperature field simulated with the nominal subduction parameter values is shown in Figure 3.1.
<table>
<thead>
<tr>
<th>Subduction Parameter</th>
<th>Value</th>
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</thead>
<tbody>
<tr>
<td>Radiogenic heating</td>
<td>0.6 μW/m³</td>
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<tr>
<td>Stress proportionality constant (α)</td>
<td>5%</td>
</tr>
<tr>
<td>Slab age</td>
<td>50 Ma</td>
</tr>
<tr>
<td>Rheology</td>
<td>Wet quartzite</td>
</tr>
<tr>
<td>Convergence rate</td>
<td>5 cm/yr</td>
</tr>
<tr>
<td>Lithospheric thickness</td>
<td>100 km</td>
</tr>
<tr>
<td>Subduction duration</td>
<td>60 Ma</td>
</tr>
</tbody>
</table>

Table 3.1. The nominal values of the east Aleutians subduction parameters.

**Effect of Radiogenic Heating**

Radiogenic heating derives from the decay of radioactive elements in the crust. The primary radioactive elements are uranium, thorium and potassiu. Depending upon the abundance of radioactive elements, radiogenic heating can vary from \(0.5 \times 10^{-6}\) W/m³ to about \(0.7 \times 10^{-6}\) W/m³ with a typical value of \(0.6 \times 10^{-6}\) W/m³. Figure 3.2 shows the SST's for radiogenic heating values of \(0.48 \times 10^{-6}\) W/m³, \(0.6 \times 10^{-6}\) W/m³, and \(0.72 \times 10^{-6}\) W/m³. I observe that radiogenic heating has no impact on the SST. This is easily understood because radiogenic heating acts at depths above 35 km and its influence on the SST disappears at depth.

**Effect of Slab Age**

Slab age affects primarily the initial temperature of a subducting slab. An older
Figure 3.1: The thermal structure of the eastern Aleutians transect simulated using a finite-difference method. The parameters used are listed in Table 3.1.
Figure 3.2: The east Aleutians subduction zone SSTs calculated with three different radiogenic heating values: 0.48 $\mu$W/m$^3$, 0.6 $\mu$W/m$^3$, and 0.72 $\mu$W/m$^3$ with 0.6 $\mu$W/m$^3$ being the typical value.
slab is in general cooler than a younger one. The incoming oceanic slab of the east Aleutians subduction zone is about 50 My old. To evaluate the effect of potential uncertainties in the slab age on the SST, I vary the slab age by ±10 My, or ±20% variation from the nominal value of 50 My. Figure 3.3 shows a 40 My old slab will increases the SST above 100 km depth (3.23 GPa) by up to 8 °C as compared to that for a 50 My old slab. On the other hand, a 60 My old slab will decrease the SST above 100 km depth by up to 10 °C.

**Effect of Convergence Rate**

An accurate determination of the incoming plate convergence rate over a long period is difficult because the convergence rate varies with time. To account for a variable convergence rate in a numerical simulation, the mantle and wedge convection velocities must be re-calculated whenever the convergence rate changes. Given that the convergence rate may change frequently, the amount of computer time required to calculate repetitively the convection velocities can be prohibitive. In my numerical simulations, I have used a constant convergence rate along the entire slab surface.

Figure 3.4 shows the SST's for convergence rates of 4 cm/yr, 5 cm/yr, 6 cm/yr, or ±20% variation from the nominal value of 5 cm/yr. I observe that above 60 km depth (2 GPa) the convergence rate has nearly no effect on the SST. Between 60 km and 100 km depth, the slower convergence rate (4 cm/yr) increases the SST by up to 8 °C. At larger depths, the slower convergence rate decreases the SST. In
Figure 3.3: The east Aleutians subduction zone SSTs calculated with three different slab ages: 40 Ma, 50 Ma, and 60 Ma.
Figure 3.4: The east Aleutians subduction zone SSTs calculated with three different convergence rates: 4 cm/yr, 5 cm/yr, and 6 cm/yr.
contrast, a convergence rate of 6 cm/yr decreases the SST between 60 km and 100 km and increases the SST at larger depths.

**Effect of Shear Heating**

Shear heating results when a solid plate subducts beneath a solid lithosphere. The shear heating intensity is given by

\[
Q = \tau v, \tag{3.3}
\]

where \( v \) is the incoming plate convergence rate and \( \tau \) the shear stress. In the brittle regime (shallower depths), shear stress is often assumed to increase with lithospheric pressure \( P \)

\[
\tau = \alpha P, \tag{3.4}
\]

where \( \alpha \) is a proportionality constant. In the ductile regime (greater depths), however, shear stress normally decreases with increasing temperature according to the power-law rheology (Kirby, 1983)

\[
\varepsilon = A_0 (\sigma_1 - \sigma_3)^n e^{-Q/RT}. \tag{3.5}
\]

where \( \varepsilon \) is the strain rate, \( T \) the absolute temperature, \( R \) the gas constant, \( \sigma_1 \) and \( \sigma_3 \) are the maximum and minimum principal stresses. \( Q, A_0 \) and \( n \) are material constants. Here, \( \varepsilon \) is taken to be \( v_c/v_{ref} * 10^{-12} \text{s}^{-1} \) (Van Den Beukel and Wortel, 1988) with \( v_{ref} = 8 \text{ cm/yr} \).

In Chapter 2, I have shown the thermal models of the east Aleutians subduction zone with different ductile-regime shear heating rates. Here, I examine the effect
Figure 3.5: The east Aleutians subduction zone SSTs calculated with two different shear stress proportionality constant values in the brittle regime.

of the brittle-regime shear heating on the SST. The brittle-regime shear heating is controlled by the proportionality constant $\alpha$. A higher value of $\alpha$ will result in greater stress and thus higher SST, particularly in the brittle regime of the subduction zone. Figure 3.5 shows the SSTs simulated with two different $\alpha$ values, 5% and 10%. At pressures above 2 GPa (about 60 km), doubling $\alpha$ value has little or no effect on the SST. At depths above 60 km, an $\alpha$ value of 10% can increase the slab surface temperature by up to 40 °C (at about 0.5 GPa or 15 km depth) as compared to an $\alpha$ value of 5%.
CONCLUSIONS

I have studied the effects of radiogenic heating, slab age, convergence rate, and brittle-regime shear heating on the east Aleutians subduction zone SST. I show that radiogenic heating has no effect on the SST at 100 km depth and that shear heating in the brittle regime affects the SST mostly at depths above 60 km. A ±20% uncertainty in the slab age or the convergence rate may change the SST by up to 10 °C at depths above 100 km.

ACKNOWLEDGMENTS

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REFERENCES


CHAPTER 4

Subduction Zone Thermal Models: Constraints from Geochemical and Seismological Observations

ABSTRACT

By using a coupled, finite-difference and finite-element method, I have generated thermal models for ten subduction zones from the east and west Pacific Rims. These ten subduction zones cover a broad spectrum of slab age, convergence rate, duration, and slab steepness. I have shown that B/Zr ratios in the associated island arc volcanic rocks are correlated with the slab surface temperatures at 100 km. A higher slab surface temperature corresponds in general to a lower B/Zr ratio and vice versa. This result supports the hypothesis that the element boron in island arc volcanic rocks is derived largely from the subducted slabs. Another important observation is that the slab tip temperatures for different arcs are very similar. They are all about 800 °C except the South America transect at 20°S where the slab tip temperature is around 600 °C. Also, the coolest slab temperatures at the slab tip depth are all around 500 °C.

INTRODUCTION

A subduction zone is a geological setting where an old, cold oceanic lithosphere
plate descends into the Earth’s mantle. Figure 4.1 schematically shows a typical subduction zone; it has several unique units: oceanic lithosphere, slab, trench, accretionary wedge, fore-arc basin, volcanic arc and back-arc basin. The oceanic lithosphere descending into the mantle consists of up to a few hundred meters of sediments, a 6-8 km thick oceanic crust (basaltic ophiolite), and a 60-90 km thick layer of peridotite lithospheric mantle. The subduction zone is clearly defined by the distribution of earthquake forces – so called Wadati-Benioff zone earthquakes. The slab tip is defined as where the deepest earthquake occurs. The trench is the place where the oceanic lithosphere starts to bend and descend. Trenches often exceed 8 km in depth and the deepest is the Marianas Trench (Fowler, 1990). Usually, thick, less dense sediments accumulate in and adjacent to the trench in the accretionary wedge. An accretionary wedge may be poorly developed in a subduction zone involving only oceanic lithosphere or where sediment is small. The fore-arc basin and the back-arc basin are separated by the volcanic arc. The mantle in the overriding plate above the oceanic lithosphere is called the mantle wedge because of its shape and includes lithosphere of the overriding plate and asthenosphere. Slab movement during subduction induces mantle wedge convection that pushes warm asthenospheric material toward the cold slab. Convection may greatly affect the temperature distribution of a subduction zone.

Subduction of an oceanic lithosphere, which is chemically distinct in its upper part and is relatively cool compared to the surrounding mantle, will lead to many
Figure 4.1. The sketch of a subduction zone involving an oceanic lithosphere subducted beneath a continental lithosphere. (From Fowler, 1990.)
important geological phenomena. Most earthquakes, volcanos and high-pressure metamorphism take place in subduction zones, materials in ocean crust get recycled back to mantle. Hyndman and Wang (1993) showed that shallow thrust earthquake depths are controlled by 350 °C temperature limit. Slab surface temperature (SST) exceeding 350 °C will yield stable slide. Intermediate and deep depth earthquakes occur in slabs where temperature is below a critical temperature. Above the critical temperature, the subducted lithosphere can not sustain the stress necessary to generate seismic events (Molnar et al., 1979; Wortel, 1982; Hobbs and Ord, 1988).

Volcanos are believed to form either as a result of partial melting of the subducted slab and sediments (adakite magmas. Defant and Drummond, 1990) or of the overriding mantle wedge triggered by fluid which is released by metamorphic reactions when slab subducts (e.g., Schmidt and Poli, 1995; Poli and Schmidt, 1998; Peacock, 1990). In the late 1960’s and early 1970’s, it was widely believed that volcanic arc magmas were produced from slab melting. Later studies suggest that slab melting can only occur in young, unusually warm SZs. and most arc volcanism is related to wedge melting. Fluids (mainly water) released from metamorphic reactions within the subducting plate may significantly lower the melting temperature of the overriding mantle wedge and trigger mantle melting.

A common controlling factor for subduction-related seismicity, metamorphic reaction and magma generation is the SZ thermal structure. For example, the volatile inventory in subducted slab material critically depends on the P-T path
for the uppermost few kilometers (oceanic crust and sediments). In the past few decades, numerous workers have contributed to development of SZ thermal models. The pioneering work by McKenzie (1969) used an analytical model to calculate the coolest temperature at the deepest earthquake depth. The limitations of the analytical model are that it does not include shear heating and wedge convection and that it assumes that slab is immersed in an asthenosphere with constant temperature, which is unrealistic especially for the shallow lithosphere part. The first limitation prevents the model from generating accurate thermal models for most real SZs. Molnar and England (1990) derived another analytical solution that includes slab surface shear heating. Their model does not include wedge convection either and applies strictly to straight slabs, i.e., constant dip angles. Since the early 1980’s, many numerical thermal models have been developed to improve temperature estimates. To this end, Peacock and his coworkers made perhaps the most significant contributions to the subject. Peacock (1990) developed a finite-difference simulation method to compute the slab pressure-temperature-time (P-T-t) path. Using the numerical data, he was able to investigate metamorphic reactions and fluid production in subduction zones. Peacock (1991) further improved his model by including mantle wedge convection. He concluded that partial melting of the subducting oceanic crust is only possible during early stage of subduction in young (< 50 My) oceanic lithosphere. Later, Peacock (1996) showed that as either the age of the subducting lithosphere or its convergence rate increases, the slab temperature
will decrease. In contrast, slab temperature increase with decreasing convergence rate and slab age and is is further enhanced by shear stress or induced mantle convection. Although the work of Peacock and his coworkers has greatly enriched our understanding of SZ processes, these early models are most based on a simple, straight-slab geometry to represent an SZ. Such models were also presented by many other workers (Oxburgh and Turcotte, 1970; Anderson et al., 1977; Hsui and Toksoz, 1979; Honda, 1985; Davies and Stevenson, 1992; Furukawa, 1993). In the real world, the geometry of SZs varies considerable worldwide and even along strike. To better approximate actual SZ geometry, Ponko and Peacock (1995) used a two-segment geometry to model the curvature of the Alaska SZ, and van den Beukel and Wortel (1986) employed a circular-arc model to simulate large variations in slab geometry.

In addition to simplified slab geometry, many previous studies are limited in the calculation of wedge convection. Wedge convection is neglected cases (e.g., Hyndman and Wang, 1993; Oleskevich et al., 1999) or calculated for a straight-slab model using an analytical solution (cf., Peacock, 1996). The computational convenience of using such straight-slab models is perhaps the main reason for their wide-spread use. For example, Peacock and Wang (1999) utilized a hybrid-geometry slab model to simulate Southwest and Northeast Japan SZs. Their model contains a curved slab at shallow depths but a straight slab at large depths where mantle convection is important. Neglecting wedge convection will underestimate the SST.
particularly, below about 100 km depth. Also, significant differences may arise in SSTs calculated using straight-slab models vs actual slab geometry (Huang et al., 1999).

In this study, a coupled, finite-difference and finite-element approach was employed to simulate SZ thermal structures (Chapter 2, Huang et al., 2001). The finite-element method is used to simulate mantle convective flow which is then incorporated in a finite-difference solution of heat transfer across a subduction zone. In particular, the finite-difference method uses a staggered grid for discretization so that the mantle convection effect can be accurately incorporated into the model. Combination of the finite-element and staggered-grid finite-difference methods makes it feasible to model an arbitrarily curved slab geometry. This capability is lacking in most previous studies.

This approach was applied to selected SZs around the eastern and western Pacific plate boundary. Representative transects were included from South, Central, and North America, the Aleutians, Kurile, Northeast Japan, and the Mariana arcs. My primary objectives are to (1) relate the SST to geochemical tracers for slab dehydration (e.g., Boron enrichment in arc magmas), and (2) evaluate the relation between slab temperature and the distribution of intermediate-depth SZ earthquakes. The distribution of Watadi-Benioff zone earthquakes is used to develop geometrically realistic models for each transect studied.

**DESCRIPTION OF THE SUBDUCTION ZONES**
Ten subduction zones (Table 4.1) from the east and west Pacific ocean plate boundary for investigation are shown in Figure 4.2. The first selection criteria was to sample a wide variety of conditions for convergent margins with reliable information for all important subduction parameters. Certain subduction zones (e.g., Tonga-Kermadec-New Zealand and other southwest Pacific arcs) are not included because key subduction parameters are not firmly established or are highly variable along strike. A second criterion was to have representative geochemical data for the associated volcanic arc lavas. Of particular interest is the element boron, which is highly fluid-mobile and is derived largely from subducting slabs (Morris et al., 1990; Ishikawa and Nakamura, 1994; Ryan et al., 1996a; Leeman, 1996; Ishikawa and Tera, 1997 and 1999; Bebout et al., 1999). Again, southwest Pacific and Indonesian arcs were not included because the data base for B is lacking.

Figures 4.3 and 4.4 show the earthquake hypocenter locations of the subduction zones. The data in Figure 4.3 are from Engdahl (personal communication to W. Leeman). The slab geometries for Central America subduction zones - Nicaragua and Costa Rica are derived from Protti et al. (1995), as shown in Figure 4.4. The Cascades slab surface geometry is obtained by combining the earthquake data (Weaver and Baker, 1988) for shallow depths (<100 km) and tomographic velocity data (Michaelson and Weaver, 1986) for greater depths. By following Isacks and Molnar (1971), I assume that the slab surface follows closely and is about 30
Figure 4.2. The locations of 10 subduction zones chosen from the Pacific Rim

For this study, (adapted from Johnson (1986)).
km above the upper limits of the seismicity bands. The argument behind this assumption is that intermediate and deep earthquakes occur inside the downgoing slab rather than in the shear zone between the slab and the surrounding asthenosphere (Isacks and Molnar, 1969). Earthquakes probably occur within the coldest part of the slab where sufficient elastic energy can be stored (Molnar et al., 1979). Previous thermal models (Peacock, 1991; Huang et al., Chapter 2) reveal that the coolest temperature in a subduction zone is about 20 km below the slab surface which is close to Isacks and Molnar’s (1971) estimate. Isacks and Barazangi (1977) examined the geometries of the descending lithosphere for most subduction zones around the world. Their slab surface geometry data are very close to ones obtained in this study from polynomial fitting to the earthquake hypocenter data (as will be discussed shortly). For example, Mariana transect has a trench-arc distance of about 250 km in both my model and Isacks and Barazangi’s (1977) data. North Chile (20°S) transect has a trench-arc distance of about 300 km in my model and 350 km in Isacks and Barazangi’s (1977) fitting data. Projections of earthquakes along strike onto a common plane (near the selected localities) further constrain the slab geometries. Uncertainties in earthquake locations and along-strike projections are probably less than ±10-20 km and negligible to the thermal modeling. The selected subduction zones differ substantially in slab dip and length. The South America 30°S transect and the Marianas transect have the shallowest and deepest dips, respectively. The deepest earthquakes occur at depths around 600 km.
Note that tomography indicates that slabs continue to much greater depths without seismicity.

A brief description of the transects is presented below.

**East Aleutians.** The east Aleutian subduction transect is located in the eastern part of the Aleutian volcanic arc system between 50°N and 56°N. It results from subduction of the Pacific plate under the north America plate (Creager and Boyd, 1991). The strike of the east Aleutian subduction zone is NE-SW. The slab movement direction is relatively perpendicular to the trench axis, but becomes progressively more oblique toward the west. For a transect at 56 °N, the convergence rate is 5.8 cm/yr. The slab age (50 Ma) and duration of subduction (60 Ma) are long enough to approach a steady state thermal equilibrium between the oceanic lithosphere and the surrounding mantle (cf., Peacock. 1996). Thus, temperature computation is not very sensitive to small variations in slab age or subduction duration. Because this subduction zone has an unusually wide trench-arc gap (i.e., low dip angle at shallow depths), accurate temperature results are best obtained using a curved model. Ponko and Peacock (1995) approximated the actual geometry using two linear segments with results fairly similar to my geometric model.

**Cascades.** The Cascades subduction zone extends from northern California to southern British Columbia (e.g., Verplanck and Duncan, 1987). The subducting Juan de Fuca plate is relatively young, and there are no earthquakes below about 100 km depth. The earliest volcanic activity in the Cascades subduction zone
Figure 4.3. The earthquake hypocenter locations of the east Aleutians, South America, Northeast Japan, and Mariana subduction zones.
Figure 4.4. The earthquake hypocenter locations of the central America subduction zones.
occurred about 42 Ma (Lux, 1981). The crustal thickness beneath the Cascades increases eastward from 22 km to 37 km (Ganoe, 1983; Spence et al., 1985). Seismological data indicate that the Benioff zone dips nearly eastward at about 10-12° (Crosson, 1983; Taber and Smith, 1985). The combination of a young slab with slow present-day convergence (2 cm/yr) makes the Cascades one of the hottest in the world. Some suggest that slab melting may be possible in this subduction zone (e.g. Defant and Drummond, 1986). Moreover, a relatively thick (~3 km) accretionary prism sediment acts as a thermal blanket on the oceanic crust (Hyndman and Wang, 1993). This thermal blanket increases temperature at shallow depths by 100 degrees, but only a few degrees at 100 km depth in my numerical experiment. For this reason, I do not include the thick prism sediments in my model.

**Central America** The central America subduction zone can be considered as having two distinct segments separated by a distinct structural feature, the Quisada Sharp Contortion (QSC. Protti et al., 1995). To the northwest, the slab consists of a slightly older oceanic crust that dips rather steeply. To the southeast, a younger crust associated with the Nazca aseismic ridge dips more shallowly and is associated with a relatively short seismic zone. Volcanic arc magmas from these sectors display significant geochemical differences that have been attributed to differences in SZ thermal conditions (Leeman et al., 1994). Two transects across the central America subduction zone, in Nicaragua and Costa Rica, were modeled to investigate differences in thermal structure. The two transects have similar overriding
plate initial temperatures and lithosphere thicknesses (i.e., mantle convection starting depth), although the Nicaragua transect has a relatively older slab. The major difference between the two transects is in their slab geometry with the Nicaragua transect slab dipping more steeply below 100 km depth than the Costa Rica transect slab. The arc magma chemistry is distinctive between the Nicaragua and Costa Rica transects, which has been ascribed at least in part to SZ thermal conditions (Leeman et al., 1994). Results from these two transects will allow us to understand how slab geometry affects temperature.

South America. The South America subduction zone considered for this study extends from 20°S to 40°S with the subduction of the Nazca plate beneath the western South America plate. The subduction zone terminates at 45°S where the Chile Rise runs into the south America plate. The age of the subducting slab varies along strike, increasing from zero age at the Chile Rise (40°S) to 45 Ma at 30°S, and to 50 Ma at 20°S. One important feature of the South America subduction zone is that the slab dip changes significantly along the strike. North of 22°S, the subducting Nazca plate dips at about 30° and reaches a depth of about 250 km; south of 26°S and north of 34°S, the slab is subhorizontal for about 200 km at 100 km depth and then dip angle increases as at other subduction zones. This flat-slab region has been amagmatic for at least the past 10 My, whereas the adjunct sectors are characterized by active volcanism. To evaluate associated along-strike variations in SZ thermal structure, three transects at locations of 20°S, 30°S, and 40°S
were modeled. Another important feature of the South America subduction zone is that the crustal thickness also varies along the strike. According to seismological constraints, crust thickness reaches 75-80 km (Zandt et al., 1994) beneath the 20°S sector and becomes normal (30-40 km) for other areas.

**Northeast Japan.** Two transects in the northeast Japan were chosen for study. Kurile and Hokkaido. The Pacific plate subducts beneath the Eurasia plate. Seismological data (Creager and Jordan, 1986) suggest that the Kurile slab reaches 1200-1300 km depth with a dip change near 500 km. Above that depth, the dip is 25-30° and below the dip is 55-65°. The convergence rate of the subducting slab is greater than 9 cm/yr.

**SUBDUCTION PARAMETERS FOR THERMAL MODELING**

SSTs depend on several subduction parameters including slab age, convergence rate, wedge convection, and shear heating. Compilation of these parameters for the selected transects is discussed below.

**Slab age at trench.** Slab age affects the initial temperature and thermal gradient of a slab. Young slabs are initially warmer than old ones. I use the GDH1 model of Stein and Stein (1992) to simulate initial thermal profiles as a function of slab age. This model assumes a temperature of 1450 °C at the base of the lithosphere (cf., McKenzie, 1969; others). Lower values for this parameter will result in lower SSTs
at any given depth. The slab age and subduction duration data for the Cascades, Nicaragua, and Costa Rica transects are compiled from Green and Harry (1999), Mauray et al. (1995), and Zeilinga de Boer et al. (1995), respectively. The slab age and subduction duration data for the other transects are based on compilation of Jarrard (1986).

**Convergence rate.** Convergence rate is taken as the net orthogonal plate convergence at trench including the estimated component of back-arc spreading. Because back-arc spreading rates are not precisely known for all arcs (e.g., Tonga), I restrict my calculations to transects where back-arc spreading rate is generally small. The main exception is the Marianas for which back-arc spreading rate is taken to be 6 cm/yr (Creager and Jordan, 1986). I calculated the convergence rates of the selected transects calculated using the NUVEL-1a model except for the Marianas whose convergence rate was calculated using the NUVEL-1 model. The two models differ systematically by less than 5% with the NUVEL-1 model results being higher. I then increased the calculated Marianas convergence rate by 5% and added the value to the estimated back-arc spreading rate. The convergence rate for the Cascades has diminished with time; the value adopted in this study was taken from Verplanck and Duncan (1987) for the last 2 My interval.

**Duration of subduction.** Duration of subduction determines the extent to which subduction approaches a steady state condition. As shown by Peacock (1996) and my calculations, the steady state condition for the east Aleutians transect is
achieved after roughly 50 My. For shorter duration times, my models represent an intermediate stage of 'transient thermal equilibration' and SSTs are higher than those in fully equilibrated slabs. In my calculations, the durations were estimated from the ages of the oldest arc-related rocks present in each transect because arc volcanism takes place usually in less than 5 My after the initiation of subduction. However, note that volcanism was not necessarily continuous over this entire time span.

**Geometry.** Subduction angle is generally steeper for old, cold plates, and vice versa. For slabs showing strong curvature, slab surface trajectories are longer and result in a higher degree of thermal equilibration (hence, higher SSTs) compared to relatively straight slabs. This factor varies considerably between, and in some cases within, subduction zones. Here, I approximate slab surface trajectories with 4th-order polynomial fits to the WBZ earthquake distributions using cross-arc hypocenter plots from the NEIC database (R. Engdahl, pers. comm.). The results are listed in Table 4.1. Thus, my models represent site-specific rather than generalized thermal structures.
Table 4.1. Polynomial fitting to the WBZ earthquake distributions.

**Wedge convection.** Wedge convection contributes significantly to slab warming, particularly at depths below the lithosphere-asthenosphere boundary in the wedge. Because this depth is poorly constrained in most places, I use a constant depth of 100 km (Sclater et al., 1981) for all the transects. In the central Andes where crustal thickness approaches 80 km, this assumption is reasonable. Oceanic crusts usually have thinner mechanical lithospheres. For an oceanic crust about 60-70 Ma old, the mechanical lithosphere boundary layer may reach an equilibrium thickness of about 90 km (Parsons and McKenzie, 1978). Generally, the thinner the lid, the greater is the potential for convective counterflow and slab warming. For example, for the eastern Aleutians, a 75 km thick lid corresponds to SSTs almost 200 °C higher than for a 100 km thick lid (Huang et al., 2001).

The models presented also assume constant viscosity in the mantle wedge, which
clearly is an oversimplification. Although temperature-dependent viscosity models will generally result in enhanced flow and intuitively warmer conditions in the wedge (e.g., Furukawa, 1993), the study of Kincaid and Sacks (1997) indicates that approaching the relatively cool slab, viscosity increases markedly and creates an insulating barrier to heat transfer to the slab from the interior portions of the wedge. If so, my simpler model may provide a reasonable first-order approximation of convective warming of the slab. In any case, unless the non-convecting lithospheric lid is appreciably thinner than assumed, there can be only limited slab warming at depth above 100 km simply because the volume of convecting mantle is relatively small. The effect of a thinner lithospheric lid is evaluated for two SZs (Aleutians and Cascades).

**Shear (frictional) heating.** My model includes shear heating at the slab surface. Models presented assume a wet quartzite flow model (Koch et al., 1980) which is adopted to simulate the presence of a thin veneer of sediment at the slab-wedge interface. Calculated SSTs are comparable to Peacock’s (1996) models in which $\tau = 0.05P$, where $\tau$ is the shear stress at the slab surface and $P$ is the lithostatic pressure. These models are roughly 100 °C warmer at 100 km depth than models assuming zero frictional heating.

Table 4.2 lists the subduction parameters for the selected arcs. The extreme values for subduction parameters are as follows: age of subducting oceanic crust at trench (9 Ma for Cascades vs. 134 Ma for Marianas), duration of subduction (~40
Ma for Cascades and Costa Rica vs. 226 Ma for South America), convergence rate
(~2 cm/yr for Cascades vs. ~10 cm/yr for Marianas), slab length measured from
the trench to the slab tip defined by the deepest seismicity (~200 km for Cascades
and Costa Rica vs. ~1200 km for Northeast Japan).

<table>
<thead>
<tr>
<th>Subduction zone</th>
<th>Slab age (Ma)</th>
<th>Conv. rate (cm/yr)</th>
<th>Duration (Ma)</th>
<th>Overriding Plate</th>
<th>Slab length (km)</th>
<th>Tip depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>East Aleutians</td>
<td>50</td>
<td>5.8</td>
<td>60</td>
<td>Cont./oceanic</td>
<td>530</td>
<td>250</td>
</tr>
<tr>
<td>(56°N)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cascades</td>
<td>9</td>
<td>2.0</td>
<td>40</td>
<td>Cont.</td>
<td>220</td>
<td>100</td>
</tr>
<tr>
<td>(45°N)</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Nicaragua</td>
<td>25</td>
<td>8.8</td>
<td>65</td>
<td>Cont.</td>
<td>380</td>
<td>220</td>
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<td>(10°N)</td>
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<tr>
<td>Costa Rica</td>
<td>15</td>
<td>8.5</td>
<td>65</td>
<td>Cont.</td>
<td>200</td>
<td>130</td>
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<tr>
<td>Andes NVZ</td>
<td>50</td>
<td>7.8</td>
<td>226±19</td>
<td>Cont.</td>
<td>730</td>
<td>340</td>
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<tr>
<td>Andes - flat</td>
<td>45</td>
<td>8.3</td>
<td>226±19</td>
<td>Cont.</td>
<td>640</td>
<td>200</td>
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<tr>
<td>Andes SVZ</td>
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<td>8.4</td>
<td>226±19</td>
<td>Cont.</td>
<td>415</td>
<td>200</td>
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<td></td>
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</tr>
<tr>
<td>NE Japan (Hok.)</td>
<td>94</td>
<td>9.5</td>
<td>115±5</td>
<td>Cont.</td>
<td>1200</td>
<td>600</td>
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<td></td>
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<td></td>
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</tr>
<tr>
<td>NE Japan (Kur.)</td>
<td>89</td>
<td>9.2</td>
<td>82±16</td>
<td>Cont./oceanic</td>
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<td>-45±5</td>
<td>Oceanic</td>
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<td>640</td>
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<td>(20°N)</td>
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Table 4.2. The subduction parameters of ten transects from the east and west
Pacific plate boundary.

**TEMPERATURE MODELS**
In order to simulate thermal structure of the selected subduction transects, I first derive the mantle convection fields as inputs to the thermal transport system using a finite-element simulation method (Kloucek and Rys, 1994) with the convergence rate imposed as a boundary condition at the slab surface. The input parameters include slab geometry, convergence rate, slab age, and duration parameters from Table 4.1 and Table 4.2. Radiogenic heating in the continental crust is assumed to be $6 \times 10^{-7}$ W/m$^2$ in all models.

**Thermal Profiles**

Depth-distance profiles for each modeled subduction transect are shown in Figure 4.5. These diagrams show slab geometry with dash lines along the slab surface. The gray line shows slab surface above the earthquake cut-off depth. Isotherms contoured at 100 °C intervals depict the computed thermal structures. In all cases except the Andes flat transect, subduction is accompanied by volcanism and the volcanic front lies at a position about 100 km above the subducted slab surface. First, I observe from Figure 4.5 that the flat-slab segment (30 °S) of the Andes is characterized by SSTSs below 300 °C up to 100 km depth and the geometry precludes any significant warming due to mantle convection. Upper plate temperatures are also relatively cool. Thus, even if the slab is cool enough to retain a significant fraction of its original volatile inventory, temperatures are too low to generate significant partial melting of mantle or even crustal materials (cf., Peacock et al., 1994). This
Figure 4.5: Numerically simulated thermal models. The contour interval is 100 °C. The gray line represents the slab surface. The mantle wedge convection starts at 100 km depth.
sector of the Andes is of course distinguished by an absence of active volcanism. Secondly, in all transects isotherms in the mantle wedge are roughly subparallel to the slab surface at depths below 100 km. This characteristic is consistent with a dominant contribution of heating due to wedge convection. Thirdly, for slabs with steepest dip below 100 km depth (Marianas and Nicaragua) isotherms are tightly bunched with steep thermal gradient perpendicular to the slab surface. Fourthly, subduction zones with high convergence rate have cool slab surface temperature which is understandable, but have high temperature in the wedge corner area. Finally, temperatures in mantle wedge approach 'dry' basaltic liquidus value (~1300 °C) just above the slab surface at depths near 100 km or slightly greater for most arcs. To achieve higher temperatures at shallower depths requires (in my models) that wedge convection is either more robust that modeled or that wedge lithosphere is thinner than 100 km as assumed.

**Slab-Surface Temperatures**

I first evaluate the extent to which the SSTs approach steady state conditions. I then illustrate how the temperature varies with distance above and below the slab surface. This information is useful for estimating the thermal conditions slightly inside the slabs or in the overriding asthenosphere. Finally, I correlate the SSTs with several metamorphic facies.

Figure 4.6 shows the SST as a function of subduction time for each transect. Each plot shows SSTs at four different subduction times at an increment of 10
percent of the final duration time. For instance, for the East Aleutian transect
the four subduction times plotted are 42, 48, 54, and 60 Ma with the lowest curve
corresponding to 42 Ma and the uppermost curve to 60 Ma. I observe that all the
transects (except the Cascades) have approached steady state condition at 100 km
or shallower depth.

The SST gradient perpendicular to the slab surface bears important information
on how temperature varies around the slab surface. Figure 4.7 shows such SST
gradients at 100 km, 200 km, and 300 km depths along each slab surface. It appears
that transects with larger convergence rates tend to have larger SST gradients.
Examples are the central Marianas transect with a high SST gradient and the
eastern Aleutian transect with a low SST gradient. This is understandable because
a slowly converging transect has long time to homogenize the temperature around
the slab surface than a rapidly converging transect.

The SSTs for all the transects at the final subduction times are shown in Figure
4.8. It shows that the SSTs increase almost linearly with depth to about 30 km
(1 GPa). Below that depth, temperatures increase with depth but with smaller
gradients to about 100 km (~3.25 GPa). The SSTs vary by only about 100 °C at
any reference depth below 30 km, but up to 300 °C (300-580 °C) below the volcanic
front (at 100 km). Except for the Costa Rica and Mariana transects, the differences
in the SSTs at any depth are within 200 °C.

An important constraint on subduction zone thermal models are P-T conditions
Figure 4.6: Evolution of the slab surface temperatures with subduction time. The four curves in each plot show SSTs at a time increment of 10 percent of the final duration time.
Figure 4.7: The SST gradients perpendicular to the slab surfaces at 100 km, 200 km, and 300 km depths.
for metamorphic rocks, particularly blueschist rocks commonly associated with exhumed ancient subduction zones. It is expected that most subduction zones have the P-T conditions to produce blueschist. Maekawa et al. (1993) indicated that the P-T conditions for blueschist-facies rocks discovered in the forearc of the active Mariana subduction zone are 150-250 °C and 0.5-0.6 GPa (16-20 km depth), respectively, as indicated by the thick line in Figure 4.8. This P-T condition is satisfied in several subduction zones including the east Aleutians, Central Americas (10°N and 11°N), South Americas (20°S and 40°S), and Central Cascades.

At face values the thermal structures preclude direct slab melting even under water-saturated conditions except possibly the Cascades. This constraint can be reconciled with slab melting and only if there is a significant increase in wedge convection and/or shear heating contributions.

Subduction Zone Temperature Variations with Lithosphere Thickness

All the above models assume that the overriding lithosphere thickness is 100 km. While being reasonable for most subduction zones, such a large lithosphere thickness may not be true for the Cascades. Tanton et al. (2001) indicated that from west to east, the Cascades basaltic magmas segregate from mantle depth of 36 km to 66 km with temperatures of up to 1300-1400 °C. Lewis et al. (2001) also found that the shallowest melting in the Cascades is at about 50 km depth (1300-1400 °C). Such shallow melting is not predicted by my model because the highest temperature at 100 km depth from this study is around 700 °C. A thinner
Figure 4.8: The SSTs for the various transects modeled. WS = wet solidus and DS = dry solidus. Also shown are metamorphic facies: Am = amphibolite, EA = epidote amphibolite, EB = epidote blueschist, Ec = Eclogite, Gs = Greenschist, LB = lawsonite blueschist.
Figure 4.9: The central Cascades temperature fields simulated with a (a) 100 km and (b) 50 km lithosphere. Weak (wet-quartzite) shear heating is included.

lithosphere is implied to reconcile this difference. Figure 4.9 shows the model result for a 50 km thick overriding lithosphere. The highest temperature now reaches 1400 °C, which is sufficient for melting.

Comparison with Other Models

Comparison of my thermal models with previous ones is difficult because previous models are generally not site-specific. Cautions must be taken in making such comparison. Table 4.3 lists the SSTs at 100 km depth reported by other workers

The SST of 1000 °C from Honda’s (1985) model is significantly higher than that from this study. The possible reasons are as follows. First, in Honda’s model, wedge convection starts at a much shallower depth (about 35 km) than the 100 km depth used in my model. Although a 35-km thick lithosphere is not realistic, the shallow mantle convection allows deep hot mantle material to be pushed up to shallow cold slab, and increases the slab temperature. Second, Honda assumed a relatively high shear stress of 100 MPa up to 60 km depth. The above two factors combined tend to yield higher SSTs. Using a part-of-a-circle slab model, van den Beukel and Wortel (1988) assumed a shear stress range from weak limestone to strong diabase. a convergence rate from 4-12 cm/yr, an oceanic crust age from 30-150 Ma. Their SST at 100 km depth is higher than ours, likely because they included a high heat flux (80 mWm⁻²) boundary condition at the location of the volcanic front. Such high heat flux can significantly increase the computed SST.
For the Alaska subduction zone, Furukawa's (1993) and Ponko and Peacock's (1995) SST data are about 170-270 °C higher than that from this study. Furukawa's model is very similar to that of Honda's (1985) except that the former does not include shear heating. Again, Furukawa's model yields higher SST at 100 km mainly because his model used a thinner overriding lithosphere. Ponko and Peacock's model produces hotter slab surface at 100 km depth possibly because their two-segment slab model results in a longer slab path.

<table>
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<tr>
<th>Authors</th>
<th>Cascades</th>
<th>Alaska</th>
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<th>Andes Flat</th>
<th>Andes SVZ</th>
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<td>1000</td>
<td></td>
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<td>This study</td>
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<td>435</td>
<td>345</td>
<td>363</td>
<td>302</td>
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Table 4.3. SSTs at 100 km depth from previous studies compared with ones from this study.

SLAB SURFACE TEMPERATURE AND BORON ENRICHMENTS
The relatively cool P-T trajectories modeled are consistent with retention of significant fractions of original slab volatile inventories. If arc magmatism is influenced primarily by slab-derived fluid additions to the mantle wedge, then the extent of wedge modification (and degree of melting of such modified mantle) is likely to vary with the degree of slab dehydration which in turn is dependent on slab thermal structure.

To test this prediction I examine the variation in B in primitive lavas from the arcs I have modeled. It has been found that volcanic front lavas have high B contents and cross-arc transects of volcanic arcs reveal systematic decreases in B contents (Ryan and Langmuir, 1993; Edwards and Morris, 1993; Leeman et al., 1994; Ryan et al., 1995; and Ryan et al., 1996b). This suggests that the high B contents in the arc magmas most likely are melts of mantle modified by slab-derived B-rich fluids. However, the mantle and the lower continental crust have low B contents (~0.1 ppm and <2 ppm, respectively) (Donnelly et al., 1981; Seyfried et al., 1984; Taylor and McLennan, 1985; Spivack and Edmond, 1987; Leeman et al., 1992; Xu et al., 1994; Chaussidon and Marty, 1995; Smith et al., 1995; Ryan et al., 1996b; Leeman and Sisson, 1996; Staudigel et al., 1996). The primary geologic sources of B are altered oceanic crust (average 26 ppm) and particularly marine sediments (80-120 ppm for mudstones and 30-40 for sands) with additional sources coming from the upper continental crust (~15 ppm). Hence, B in the volcanic front is believed to
come largely from the subducted slab with possible contamination by the upper continental crust.

The above arguments are backed by the following observations. First, most oceanic crusts and sediments have similar B contents; however, arc magmas associated with young hot slabs lack strong B-enrichment (Leeman et al., 1990; Hochstaedter et al., 1996). This indicates that much of the B content in young hot slabs is lost during subduction. The greater loss of B in hot slabs is consistent with the observation that B concentration is strongly temperature-dependent in most prograde metamorphic suites. Moran et al. (1992). Leeman et al. (1992). Bebout et al. (1993, 1999), and Leeman and Sisson (1996) showed that average B contents and B/Zr ratios in more than 20 metamorphic suites demonstrate systematic depletions with increasing temperature of equilibration. Leeman (2001) indicates that the depletions have a linear trend. The relative B-depletion is independent of protolith variations. Secondly, B-enrichments and $^{10}$Be-richments in lava suites have strong correlation (Morris et al., 1990; Leeman et al., 1994). Because $^{10}$Be is a cosmogenic isotope and is negligible in mantle because of its short half life (1.6 my), the only source of $^{10}$Be in arc magma is subducted materials. Therefore, the correlation of B and $^{10}$Be enrichments in arc magma suggests that B, like $^{10}$Be, is derived mostly from the subducted slab. Both geochemical tracers indicate that young hot slabs experience more element depletion than old cold slabs.

Figure 4.10 shows B/Zr ratios vs slab surface temperatures at two different
B-enrichment levels, 5 ppm and 10 ppm. The ratio B/Zr rather than raw B-enrichment is shown because the ratio is less sensitive to magmatic differentiation processes and presumably reflect the magma source composition (Ishikawa and Nakamura, 1994; Leeman et al., 1994; Leeman, 1996). The use of a non-fluid mobile element Zr as the denominator increases the ratio more dramatically in the process of fluid-mitigated source enhancements. Leeman (1996) showed that the ratios at both B-enrichment levels are usually substantially higher than MORB-OIB values and differ systematically between arcs. They also correlate with a variety of subduction parameters, e.g., slab length, that are sensitive to thermal conditions (cf., Molnar et al., 1979; Jarrard, 1986; Severinghaus and Atwater, 1990). Therefore, the B/Zr ratios are considered to be closely related the slab thermal conditions. As Figure 4.10a shows, the B/Zr ratio decreases systematically with increasing slab surface temperature at 100 km depth. The B/Zr ratio at 10 ppm B-enrichment also demonstrates a similar tendency but with more scatter. The likely reason is that higher B-enrichments may have included contributions from the upper continental crust during magma-wallrock interactions, crystal fractionation, or other effects (Hildreth and Moorbath, 1988; Kelemen et al., 1990). These sources of B-enrichment do not have direct links to slab temperature. The above results have an important consequence that the numerically simulated subduction zone thermal models are consistent with previous observations that B content decreases (with a linear tendency) with increasing temperature, hence support the statement
that the element B is derived largely from the subducted slab.

**SLAB TIP TEMPERATURE AND SEISMICITY**

The slab tip is defined by the cutoff depth of seismicity beyond which no earthquake failure can be sustained. Because earthquake failure is mainly controlled by temperature, it is important to examine SSTs at slab tips or the coolest slab temperature at the earthquake cutoff depth. In general, slabs of the same length may have different tip depths, depending on the slab surface curvature. According to Molnar et al. (1979), intermediate and deep earthquakes should occur in the coldest portions of the downgoing slabs. If this is true, the coolest slab temperature at the earthquake cutoff depth should be the earthquake cutoff temperature. This will impose another constraint on calculated thermal models. As Table 4.4 and Figure 4.11 reveal, the SSTs at the tip depth are nearly independent of both tip length and tip depth. The SST at the tip depth are in general about 800-850 °C, except for the Andes flat (30°S) transects for which the tip temperature is around 590 °C. The coolest slab temperatures at the tip depth are all in the range of 425 - 580 °C. This relation has been suggested by previous workers (e.g., Wortel, 1982: Molnar et al., 1979)
Figure 4.10: B/Zr ratios vs the slab surface temperatures at two different Boron concentrations. (a) 5 ppm and (b) 10 ppm. In the plots, Ale = East Aleutian (56°N), CR = Costa Rica (11°N), Cas = Cascades (45°N), Nic = Nicaragua (10°N), Hok = Hokkaido (42°N), Kur = Kurile (50°N), Mar = Central Mariana (20°N), SAF = Anes flat (30°S), SAN = Andes NVZ (20°S), and SAS = Andes SVZ (40°S).
<table>
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<th>SST at slab tip</th>
<th>Coolest slab T at tip depth</th>
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<td>865</td>
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<td>730</td>
<td>730</td>
<td>580</td>
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<td>400</td>
<td>807</td>
<td>425</td>
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<td>490</td>
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<tr>
<td>Central Mariana (20°N)</td>
<td>355</td>
<td>858</td>
<td>480</td>
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Table 4.4. SSTs at 100 km and slab tip depths and coolest slab temperatures at the tip depths.

**CONCLUSIONS**

I have modeled ten subduction zones around the east and west Pacific plate boundary. These subduction zones cover a wide range of subduction parameters. My model results show that slab melting is unlikely except possibly for the Cascades transect. Slab melting in other transects are not predicted by the calculated thermal models. My model results compare reasonably well with several other thermal models for the Cascades, the South America transects, and the Alaska transect. The calculated P-T paths seem well constrained by reported blueschist P-T condition (150-250 °C at 0.5-0.6 GPa). The B-enrichment data from volcanic arc magmas demonstrate a clear tendency of depletion with increasing slab surface
Figure 4.11: The SSTs at slab tip depth and coolest slab temperatures vs slab tip depth.
temperature at 100 km. This is especially true at low B-enrichment level (5 ppm). This result supports that the B-enrichments in volcanic arcs are derived largely from the subducted slabs. I have also shown that the slab tip temperatures have little dependence on the slab tip lengths and depths. In most cases, the slab tip temperatures are around 800 to 850 °C. The coolest slab temperatures at the slab tip depth are all around 500 °C.

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