The thermal structure of collisional orogens as a response to accretion, erosion, and radiogenic heating

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Abstract. Thermal models of collisional orogens generally predict temperature structures that are much cooler than those recovered by thermobarometric studies. Here we demonstrate that high-temperature, low-pressure metamorphism and the development of inverted geotherms within collisional belts may be the result of accretion and erosion acting on crust enriched with heat-producing elements. A new two-dimensional finite difference model, described here, incorporates the subduction of lithosphere with heat-producing material in the upper crust, accretion of crustal material from the subducting plate to the upper plate, and surface erosion of the upper plate. These processes result in the development of a wedge of heat-producing material within the upper plate. The rate of heat production within the wedge and maximum depth of the wedge are the most important parameters controlling the magnitude of upper plate temperatures. Our model yields inverted upper plate geotherms when heat production rates exceed 0.75 μW/m² and the heat-producing wedge extends to a depth greater than 35 km. Temperatures in excess of 500°C at depths of 20-30 km are computed when heat production rates are greater than ~1.75 μW/m² and the wedge extends to a depth >50 km. Other processes, such as shear heating, fluid flow, or mantle delamination, need not be invoked to explain geologic evidence of high temperatures or inverted thermal gradients in collisional systems.

1. Introduction

Quantitative thermobarometry and geochronology have improved our ability to place constraints on the deformational and thermal evolution of collisional orogens, yet many questions remain largely unanswered. For example, collisional orogens commonly contain rocks with high-temperature metamorphic assemblages (>500°C), recording conditions several hundred degrees hotter than expected for a “normal” stable continental geotherm. In addition, many collisional zones contain anelastic melt products that are spatially and temporally associated with peak metamorphic mineral assemblages. The sources of heat responsible for high temperatures at shallow levels remain problematic, and a variety of mechanisms have been proposed, including thickening of crust enriched in heat-producing elements [England and Thompson, 1986], shear heating within fault zones [Bird et al., 1975; England et al., 1992; Molnar and England, 1990], and high levels of transient heat flow from the asthenosphere [Bird, 1978a; De Yoreo et al., 1991; Oxburgh and Turcotte, 1975].

Some orogenic belts retain metamorphic evidence of inverted geothermal gradients [Hubbard, 1989; Graham and Powell, 1984]. A variety of models have been developed to relate such inversions to orogenic processes, but none are consistent with basic geologic constraints in the Himalayan orogen, which displays the best documented example of inverted geothermal gradients within a collisional system.

For example, Jamieson et al. [1996] have shown that inverted metamorphic gradients can be the result of postmetamorphic structural restacking; however, this is not consistent with the lack of field evidence of significant postmetamorphic structures along critical transects [Hubbard, 1989; Macfarlane, 1995]. England and Molnar [1993] proposed that shear heating due to friction along the main shear boundary could heat the upper plate and cause inversion of geotherms. However, this mechanism does not yield the observed geothermal inversions at the appropriate structural position, and very high shear stresses are required (>100 MPa) to raise temperatures sufficiently. Royden [1993] suggested that erosion and accretion can play significant roles in controlling the steady state thermal regime of collisional orogens, and her model predicts upper plate geothermal inversions at high temperatures (600°C-700°C) if surface erosion is rapid and if the upper plate is composed entirely of material significantly enriched in heat-producing elements.

Our goal is to develop a first-order understanding of the thermal evolution of orogenic belts based solely on processes which are common to all collisional orogens and which are geologically reasonable. We begin by reviewing the current state of knowledge about processes that affect the thermal structure of orogenic belts. These include heat flow from the asthenosphere, crustal heat production, frictional heating, advection of rock by deforming and surficial processes, fluid flow, and magma migration.

2. Orogenic Processes, Observed and Postulated

The combination of radiogenic heat from continental crust and heat fluxed through the base of the lithosphere from the asthenosphere results in surface heat flow values in cratonic regions of 40-80 mW/m² [Slater et al., 1980].
Oceanic heat flow data indicate that the flux of heat at the base of the lithosphere is of the order of 30 mW/m² [Sclater et al., 1980, 1981]; if asthenospheric conditions beneath continental lithosphere are similar, heat fluxed through the base of the lithosphere would account for ~75% to 40% of surface heat flow and yield a background geothermal gradient of ~10°/km. The remaining 10-50 mW/m² of surface heat flow is presumably attributable to radioactive heat production within the crust. Measured values of crustal heat production vary widely, from >10 µW/m² for granites enriched in uranium and thorium to <0.1 µW/m² for tholeiitic basalts, with an average value of ~1.7 µW/m² for continental crust [Van Schmus, 1989].

While radiogenic and asthenospheric sources account for the heat within stable cratons, processes inherent to orogenesis may provide additional heat. One hypothesized process is the removal of downgoing oceanic lithosphere [Sacks and Sekor, 1990; Davies and von Blanckenburg, 1993] or continental lithospheric mantle [Houseman et al., 1981; Bird, 1978b] during subduction. Either would result in the placement of hot asthenospheric material at high levels within the lithosphere and thus raise temperatures within the crust. However, while geodynamic models suggest that such processes might occur, there is no direct evidence that detachment or convective removal of mantle lithosphere actually occurs in collisional orogens.

Another potential heat source associated with orogenesis is shear heating along the subduction boundary. Laboratory measurements of rock strength suggest differential stresses of the order of 100 MPa may be required to deform rocks within the upper crust [Brace and Kohlstedt, 1980; Kirby and McCormick, 1989], and shear heating due to fault movement associated with stresses of this magnitude is frequently invoked to explain anomalously metamorphism [Scholz, 1980; Molnar and England, 1990; England and Molnar, 1993]. However, there is no direct evidence that such high differential stresses occur in situ at plate boundaries. In fact, seismic radiation estimates and heat flow studies of the San Andreas fault suggest that shear stress on the fault is of the order of several tens of megapascals [Brace et al., 1969; Lachenbruch and Sass, 1980; Zoback et al., 1987], consistent with the low stress values predicted by mechanical models of subduction boundaries [Bird, 1978b; Barr and Dahlen, 1990]. Thermal calculations indicate that fault movement at geologically reasonable rates (tens of kilometers per million years) associated with stress levels of the order of tens of megapascals would only affect temperatures at a local scale and would not significantly perturb the thermal structure of an orogen [Scholz, 1980; Bird, 1978a].

Regardless of how heat is introduced into the crust, the overall thermal structure of orogenic belts depends on how heat is distributed through the crust. In addition to conduction, other heat transfer mechanisms play an important role. In many regions, high-grade metamorphism has been attributed to the influx of high-temperature aqueous fluids [Ferry, 1980; Hoisch et al., 1988]. However, several studies suggest that hydrothermal fluid systems may not be regionally extensive and may not significantly perturb regional thermal structures [Bunks et al., 1991; Ferry, 1994; Ferry and Dipple, 1991]. High-temperature metamorphism has also been attributed to magmatic intrusion [Yardley et al., 1987; Barron and Hanson, 1989], but some studies indicate that more than 33% of the crustal volume would have to be replaced by magma at 1000°C to raise temperatures from 300°C to 600°C [De Yoreo et al., 1991], and most collisional belts do not contain such large volumes of far-travelled magmas. The occurrence of large tracts of high-grade metamorphic rocks in orogens lacking evidence of thermal perturbation due to fluid flow or extensive magmatism indicates that such essentially convective processes may not play a fundamental role in heat transfer during collisional orogenesis.

Advection may be an important method of heat transfer. The three primary mechanisms of advection in collisional settings are subduction, erosion, and accretion. Studies of present day collisional orogens yield estimates of subduction rates up to 25 km/m.y. [Lyon-Caen and Molnar, 1985], and erosion rates up to ~5.0 km/m.y. [Li, 1976; Hubbard et al., 1991; Copeland and Harrison, 1990]. Within the Himalayas, horizontal accretion rates of ~6 km/m.y. are indicated by the presence of ~300 km of lower (Indian) plate rocks that have been accreted to the Eurasian plate over the last ~50 m.y. Results of previous thermal models indicate that the processes of erosion and accretion are important advective processes within collisional orogens and can play a critical role in the thermal structure [Barr and Dahlen, 1989; Royden, 1993; Huerta et al., 1996].

If we disregard mechanisms that are of local importance or are mostly unconstrained, we find that the gross thermal structure of orogenic belts largely depends on the redistribution of asthenospheric and crustal (radiogenic) heat through the advective processes of subduction, accretion, and erosion. Our goals are to understand the nature of these processes in a theoretical way, and to determine if they alone can be responsible for the thermal structures observed within orogens or if other processes (such as shear heating, mantle delamination, or fluid flow) must be involved. To this end, we have constructed a two-dimensional numerical model tracking the thermal history of collisional belts in which accretion and erosion act concurrently with subduction. This paper expands on the analytical study of Royden [1993] by applying a numerical method to the transient thermal evolution of collisional orogens, thus eliminating the need for the upper plate to have uniform radiogenic heat production throughout.

### 3. Model Configuration and Boundary Conditions

In this study, we simulate a collisional orogen as a slab of continental lithosphere of thickness l being subducted beneath an overriding wedge of continental lithosphere (Figure 1). The upper surface (z=0) is horizontal, the subduction boundary is assumed to have a uniform and constant dip (at angle Ω) from the surface to the base of the overriding lithosphere, and there is no internal deformation of upper or lower plates (see Table 1 for symbols, variables, and values used in this paper). A layer of crustal material enriched with heat-producing elements extends from the surface to a depth of d.

Convergence velocity (v_c) is defined as the velocity of particles in the downgoing plate with respect to particles in the upper plate, accretion rate (α) is defined as the rate at which material is transferred across the subduction contact and is measured vertically with respect to the subduction
eliminates computational complications that arise in a reference frame that is not stationary with respect to the subduction boundary. To illustrate this point, Figure 1b shows particle paths in two frames of reference, one fixed to the toe of the upper plate (Figure 1b, bottom), and a more familiar frame of reference that does not move horizontally with respect to particles in the upper plate (Figure 1b, middle). When erosion and accretion rates are zero, particle paths in the two frames of reference are identical (Figure 1b, top); particles in the upper plate are stationary, and particles in the downgoing plate move parallel to the subduction contact. However, the two frames of reference present very different views when $a$ and $e$ are nonzero. In the more familiar frame of reference (Figure 1b, middle), particles in the upper plate move vertically upwards at a rate $e$, while particles in the downgoing plate move at a velocity that is the sum of the velocity of upper plate particles plus the convergence velocity. In addition, in this frame of reference, the subduction contact moves horizontally with velocity $(a-e)/\tan \Theta$. The movement of the subduction contact makes analysis and computation complicated because the thickness of the upper plate at any location changes with time. This complication is eliminated if we use a reference frame fixed to the toe of the upper plate. In this frame of reference (Figure 1b, bottom) the subduction boundary is stationary, the velocity of upper plate particles has a horizontal component $(a-c)/\tan \Theta$ and a vertical component $e$, and the velocity of particles in the downgoing plate is the sum of the velocity of upper plate particles plus the convergence velocity.

With respect to this frame of reference, the horizontal ($u$) and vertical ($w$) velocities of particles within the upper and lower plates are given by

$$
\begin{align*}
    u &= (a-c)/\tan \Theta \\
    w &= -e \\
    u &= v_e \cos \Theta + (a-e)/\tan \Theta \\
    w &= v_e \sin \Theta - e.
\end{align*}
$$

We solve the heat flow equation in two dimensions for conduction and advection by explicit finite difference techniques, for a box of dimensions $z=2l$ by $x=x_{\tan \Theta}$, using vertical grid spacing $\Delta z=2$ km, horizontal grid spacing $\Delta x=\Delta z/\tan \Theta$, and time steps $\tau=0.05$ m.y. (Figure 1a). Boundary conditions are constant temperature of $T=0^\circ C$ at the surface ($x=0$), and constant temperature of $T=T_a$ at the “base” of the downgoing lithosphere ($z=0$, parallel to the subduction contact). Entering temperatures of the downgoing plate ($x=0$, $z<l$) are equivalent to the steady state temperatures of a lithosphere of thickness $l$ with basal temperature of $T_a$ and with an uppercrustal layer enriched in heat-producing elements to depth $d$. Boundary conditions on the right-hand edge for the upper plate ($z<l$) are also equivalent to the steady state temperatures of a lithosphere of thickness $l$ with basal temperature of $T_a$ and with an uppercrustal layer enriched in heat-producing elements to depth $d$. Right-hand boundary conditions for the downgoing plate ($z<l$) are such that horizontal thermal gradients are constant ($\partial^2T/\partial x^2=0$). Note that the boundary conditions for the base of the lithosphere and the right-hand edge have little to no effect on the results presented in this paper because of the relative time scales of advection and convection.

Figure 1. (a) Model geometry used to simulate the thermal evolution of collisional orogens. Material is accreted from the lower plate to the upper plate at rate $a$ (vertical component relative to subduction contact) and removed at the surface at rate $e$. Dip of the subduction zone, $\Theta$, and convergence velocity, $v_e$, are assumed constant. At steady state, a wedge of heat-producing crust within the upper plate has a maximum depth $d_w$ and a surface width $s_w$. (b) Particle paths in two frames of reference. (top) No accretion or erosion, particle paths are identical in both frames of reference. (middle) Frame of reference in which the subduction zone moves with respect to the frame of reference. (bottom) Frame of reference fixed to the toe of the upper plate ($a=\tan \Theta$).
Table 1. Definitions of Variables and Values Used

<table>
<thead>
<tr>
<th>Variable</th>
<th>Physical Meaning</th>
<th>Value or Units</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a$</td>
<td>accretion rate</td>
<td>km/m.y.</td>
<td></td>
</tr>
<tr>
<td>$e$</td>
<td>erosion rate</td>
<td>km/m.y.</td>
<td></td>
</tr>
<tr>
<td>$V_c$</td>
<td>convergence velocity</td>
<td>km/m.y.</td>
<td></td>
</tr>
<tr>
<td>$\Theta$</td>
<td>dip of subduction zone</td>
<td>deg</td>
<td></td>
</tr>
<tr>
<td>$A$</td>
<td>heat production rate of crust enriched in heat-producing elements</td>
<td>$\mu$W/m$^2$</td>
<td></td>
</tr>
<tr>
<td>$d_f$</td>
<td>initial thickness of crust enriched in heat-producing elements</td>
<td>18 km</td>
<td></td>
</tr>
<tr>
<td>$l$</td>
<td>thickness of foreland lithosphere</td>
<td>126 km</td>
<td></td>
</tr>
<tr>
<td>$T_b$</td>
<td>temperature at base of lithosphere</td>
<td>1260°C</td>
<td></td>
</tr>
<tr>
<td>$K$</td>
<td>thermal conductivity</td>
<td>2.5 W/mK</td>
<td></td>
</tr>
<tr>
<td>$\alpha$</td>
<td>thermal diffusivity</td>
<td>$10^{-4}$ m$^2$/s</td>
<td></td>
</tr>
<tr>
<td>$\tau$</td>
<td>time step for thermal model</td>
<td>0.05 m.y.</td>
<td></td>
</tr>
<tr>
<td>$\Delta z$</td>
<td>vertical grid spacing</td>
<td>2 km</td>
<td></td>
</tr>
<tr>
<td>$\Delta x$</td>
<td>horizontal grid spacing</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$x$</td>
<td>horizontal distance from upperplate toe</td>
<td>km</td>
<td>$x=0$ is upper plate toe</td>
</tr>
<tr>
<td>$z$</td>
<td>vertical distance from surface</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$w$</td>
<td>vertical particle velocity</td>
<td>km/m.y.</td>
<td></td>
</tr>
<tr>
<td>$u$</td>
<td>horizontal particle velocity</td>
<td>km/m.y.</td>
<td></td>
</tr>
<tr>
<td>$S_{w}$</td>
<td>maximum width of HP wedge at surface</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$d_{w}$</td>
<td>maximum depth of HP wedge</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$t_{f}$</td>
<td>time to full development of HP wedge</td>
<td>m.y.</td>
<td></td>
</tr>
<tr>
<td>$t_{i}$</td>
<td>time since initiation of collision</td>
<td>m.y.</td>
<td></td>
</tr>
<tr>
<td>$T_{ss}$</td>
<td>time to thermal steady state of orogen</td>
<td>m.y.</td>
<td></td>
</tr>
<tr>
<td>$T_{max}$</td>
<td>temperature maximum within upper plate</td>
<td>ºC</td>
<td></td>
</tr>
<tr>
<td>$x_{max}$</td>
<td>horizontal location of $T_{max}$</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$y_{max}$</td>
<td>vertical location of $T_{max}$</td>
<td>km</td>
<td></td>
</tr>
<tr>
<td>$T_{o}$</td>
<td>temperature at subduction contact at $z=60$ km</td>
<td>ºC</td>
<td></td>
</tr>
<tr>
<td>$dT/dt_{ss}$</td>
<td>near-surface thermal gradient</td>
<td>ºC/km</td>
<td></td>
</tr>
</tbody>
</table>

Initial ("precollision") geotherms are taken as the steady state temperatures of a subduction regime with accretion and erosion rates set equal to zero. The upper plate is of "continental" lithosphere with a layer of heat-producing (HP) crust between $z=0$ and $z=d_f$, and the lower plate is "oceanic" lithosphere without HP crust (or $A=0$). At $t=0$ (initiation of "collision"), continental lithosphere (with a layer of HP crust between $z=0$ and $z=d_f$) in the downgoing plate enters the orogenic system at $x=0$, and accretion and erosion are initiated. Subsequent to collision, continental lithosphere is subducted beneath the upper plate, accretion transfers material from the downgoing plate to the upper plate, and erosion removes material from the upper plate surface.

4. Redistribution of HP Crust

Of particular interest is the effect of accretion and erosion on the distribution of HP crust within the orogen. Subduction moves HP crust to depth, while accretion transfers this HP crust to the upper plate, and erosion advects the HP crust to the surface of the upper plate. These advective processes result in the development of a wedge of HP crust within the upper plate (Figure 1a). The HP wedge grows with time until a steady state geometry is attained. At this time, the surface width of the wedge is such that the amount of HP crust removed from the surface by erosion equals the amount of HP crust accreted to the upper plate. The steady state size of the wedge is a function of the initial thickness of the HP layer in the downgoing plate, convergence velocity, erosion rate, accretion rate, and subduction angle.

For cases where $\Theta\neq 0$, the HP wedge within the upper plate reaches a steady state shape (Figure 1a) by time

$$t_{o} = \left\{ \frac{d_{f}}{e} \right\} \times \left\{ 1 + \left( \frac{V_c}{a} \right) \times \sin \Theta \right\},$$

with steady state maximum depth and surface width of

$$d_{w} = d_{f} \times \left\{ 1 - \frac{a}{V_c} + \left( \frac{V_c}{a} \right) \times \sin \Theta \right\},$$

$$s_{w} = d_{w} \times \left\{ \frac{a}{\left( e \times \tan \Theta \right)} \right\}.$$  

5. Modeled Transient Thermal Structures

Three examples illustrate the transient thermal effects of accretion and erosion and the attendant redistribution of HP material, as shown by the growth of the HP wedges and associated thermal evolutions (Figure 2). In the first case (Figure 2a), we used nominal values for advection and heat production rates with $v_c=70$ km/m.y., $a=2$ km/m.y., $e=1$ km/m.y., and $A=1.75$ $\mu$W/m$^2$ (Table 2). The HP wedge reaches steady state geometry by $t=53$ km with maximum depth ($d_{f}$) of 41 km, and surface width ($s_{w}$) of 440 km. Initially ($t=0$), subduction of cold "oceanic" lithosphere cools the upper plate and inverts geotherms near the subduction contact, with temperatures of less than 300°C to depths in excess of 50 km. From $t=0$ m.y. to $t=12$ m.y., temperatures near the subduction contact increase as HP material is subducted and accreted to the upper plate. By $t=12$ m.y., HP material has been subducted to a depth of $d_{f}=33$ km, and a significant amount of HP material had been accreted to the upper plate with temperatures in the region of HP crust in
excess of 300°C at depths of ~30 km. As the wedge continues to grow laterally, temperatures within the growing wedge increase, and by t=20 m.y., the 400°C isotherm slightly inverts where it crosses from the lower plate into the upper plate. After t=20 m.y., temperatures near the foreland (from x=0 to x=150 km) change imperceptibly, while temperatures near the deepest portion of the wedge (x=200 km to x=350 km) continue to slowly rise. By t=60 m.y., the HP wedge is completely developed (t_e=53 m.y.), and maximum temperatures within the upper plate are of the order of 500°C at depths greater than 55 km. From t=60 m.y. to t=100 m.y., temperatures within the orogen do not change significantly, and by t=100 m.y., thermal steady state is achieved (t). Steady-state geotherms within the upper plate are slightly elevated (compared to typical cratonic geotherms) with temperatures of 400-500°C at depths of 20-55 km and maximum temperatures of ~500°C at depths greater than ~55 km. Lateral thermal gradients are minimal throughout the upper plate, but isotherms near the subduction contact at depths greater than 30 km are steep to inverted.

As a second example, Figure 2b displays model results using lower accretion and erosion rates (a=1.5 km/m.y., e=0.9 km/m.y.), which result in a slowly developing (t_e=72 m.y.) moderately deep HP wedge (d_e = 54 km). Early on (from t=0 to t=30 m.y.), the thermal evolution is similar to case A. However, by t=40 m.y., temperatures within the upper plate of Figure 2b are significantly higher.
(T_{max} > 500°C), and a broad region of the upper plate displays inverted geotherms associated with a local temperature maximum near the base of the HP wedge. By t=80 m.y., the HP wedge is completely developed (t_{e} = 2 m.y.). A significant portion of the upper plate has achieved temperatures > 500°C, maximum upper plate temperatures are > 600°C, and the 400°C and 500°C geotherms are inverted where they cross from the upper plate into the lower plates. Steady-state temperatures (t_{e} = 120 m.y.) within the upper plate are elevated, with temperatures in excess of 500°C at depths as shallow as z = 25 km and maximum temperatures within the upper plate of > 600°C at z = 35 km. As a third example, Figure 2c is based on the same advective rates and HP wedge geometry as Figure 2a, but the radiogenic heat production rate has been increased to 3.0 \mu W/m^3. Following initiation of collision, temperatures within the orogen increase quickly, and by t=12 m.y., temperatures near the subduction contact at depths of ~25 km are greater than 400°C. By t=30 m.y., a broad region within the upper plate displays inverted geotherms associated with a local temperature maximum in excess of 600°C. Thermal steady state has been reached by t=100 m.y., geotherms are inverted within the upper plate, and temperatures are > 600°C across the orogen for > 200 km from z=20 km to z=60 km. For all three cases, the time to thermal steady state of the orogen as a whole (t_{e}) postdates full development of the HP wedge by tens of millions of years. In general, t_{e} decreases with increasing advective rates; doubling the advective rates reduces t_{e} by a factor of ~0.6. The time to thermal steady state of a particular location is related to its distance from x=0, y=0; shallow areas close to the foreland reached thermal equilibrium by t=10 m.y., whereas deep areas and toward the hinterland equilibrated by t~ \infty.

6. Crustal Versus Asthenospheric Components of Heating

The processes of erosion, accretion and subduction not only advect heat, but they also serve to redistribute HP material within the crust. As shown in the examples above, the accumulation of HP material within the upper plate contributes significantly to the thermal budget, and the geometry of the HP wedge can exert primary control on the evolution of the thermal structure of the orogen. We next
analyze the contributions of various parameters by examining the steady state thermal structures and the associated geometry of the HP wedge for a variety of different parameter combinations.

In order to better isolate and illustrate the thermal consequences of the distribution of HP crust, we have divided steady state thermal structures into two components: the steady state structure that arises from heat conducted and advected from the asthenosphere, and the thermal structure that develops due to the production, conduction and advection of heat from HP crust. We calculate the asthenosphere-derived (AD) component by setting $A=0$ and computing the resulting temperatures, and the HP crust-derived (HPD) component by setting $T_0=0$ and computing the resulting temperatures. The total thermal structure is obtained by summing the AD and HPD components.

The AD component of the steady state thermal structure is controlled by advective rates ($v_x, a, e$), and by the temperature at the base of the downgoing plate, $T_x$. In all cases, the AD component exhibits isotherms in the upper plate that dip shallowly toward the subduction zone and lowerplate isotherms that are subparallel to the subduction zone (Figures 3-6).

The HPD component of the steady state thermal structure is controlled by the rates of heat production and advection ($A, v_x, a, e$). Since temperature scales linearly with heat production rate, the following steady state analyses (based on a heat production rate of 1 $\mu W/m^3$) can be scaled up or down to give results for any value of $A$. In all cases the HPD component produces closed isotherms encircling a local temperature maximum within the upper plate (Figures 3-6).

Below, we investigate the control on the AD and HPD components of the thermal structure exerted by the subduction angle ($\Theta$), HP wedge geometry ($d_w$ and $s_w$), and convergence rate ($v_x$). Provided that $d_w$, $\alpha$, $k$, and $T_x$ are held constant, these four variables completely specify the thermal structure of the system.

6.1. Subduction Zone Dip ($\Theta$)

Figure 3 displays the thermal structures for subduction dips varying from $\tan 0 = 0.1$ to $\tan 0 = 0.6$. Burial rate is held constant ($w = 2.9 \text{ km/m.y.}$) as are the vertical ($z$) components
Table 2. Values of Parameters Used in Computing Temperatures in Figures 2-6 and 9

<table>
<thead>
<tr>
<th>Figure</th>
<th>Convergence Rate $v_c$, km/m.y.</th>
<th>Accretion Rate $a$, km/m.y.</th>
<th>Erosion Rate $e$, km/m.y.</th>
<th>Heat Production $A$, $\mu$W/m$^3$</th>
<th>Fault Dip, $\theta$</th>
<th>Maximum Depth of Wedge $d_w$, km</th>
<th>Surface of Wedge $sw$, km</th>
<th>Time to Full Development of Wedge $t_w$, m.y.</th>
</tr>
</thead>
<tbody>
<tr>
<td>2a</td>
<td>20</td>
<td>2.0</td>
<td>1.0</td>
<td>1.75</td>
<td>tan$\theta$=0.2</td>
<td>44</td>
<td>440</td>
<td>53</td>
</tr>
<tr>
<td>b</td>
<td>20</td>
<td>1.5</td>
<td>0.9</td>
<td>1.0</td>
<td>54</td>
<td>450</td>
<td>72</td>
<td>53</td>
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<tr>
<td>c</td>
<td>50</td>
<td>2.0</td>
<td>1.0</td>
<td>3.0</td>
<td>tan$\theta$=0.2</td>
<td>44</td>
<td>147</td>
<td>53</td>
</tr>
<tr>
<td>d</td>
<td>39.5</td>
<td>2.0</td>
<td>1.0</td>
<td>0 and 1.0</td>
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<td>387</td>
<td>33</td>
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<tr>
<td>3a, c</td>
<td>20.0</td>
<td>2.0</td>
<td>1.0</td>
<td>0 and 1.0</td>
<td>tan$\theta$=1.2</td>
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Unless otherwise specified, the following values are used in all figures: thermal diffusivity, $\alpha$=10$^{-6}$ m$^2$/s; thermal conductivity, $K$=2.5 W/mK; lithospheric thickness, $L$=126 km; initial thickness of HPF-enriched layer, $d_p=18$ km; basal lithospheric temperature, $T_0=1760^\circ$C.

of all other parameters ($d_p$=44 km, and $w_c$ = -1 km/m.y.), while horizontal components ($x_n$, $u_n$, and $u_e$) are scaled with $1/\tan \theta$.

6.1.1. HPD component, (Figures 3a, 3b, 3c, and 3d, and Table 3). For 0.1$\tan \theta$<0.6, varying the subduction angle has little effect on the thermal structures; thermal structures for subduction angles from $\tan \theta$=0.1 to $\tan \theta$=0.4 are nearly identical, with temperatures differing by less than 10% (Table 3). For $\tan \theta$=0.6, temperatures within the upper plate are up to 15% lower than for $\tan \theta$=0.1 and temperatures within the lower plate are up to 15% higher, reflecting the reduction of lateral temperature gradients by the increased horizontal conduction.

6.1.2. AD component (Figures 3e, 3f, 3g, and 3h, and Table 3). Temperatures do not vary significantly for $\tan \theta$<0.4. For $\tan \theta$<0.6 temperatures increase slightly (<10%) in the upper plate, due to horizontal conduction from the right-hand boundary conditions.

6.1.3. Total thermal structures (HPD plus AD). Both the AD and HPD components of the thermal structure are fairly insensitive to changes in subduction zone dip for $\tan \theta$<0.6. Thus, total thermal structures are also be unchanged, while at higher subduction dips ($\tan \theta$>0.6) the increased horizontal conduction reduces lateral temperature gradients, resulting in a slightly cooler upper plate, and a warmer lower plate.

6.2. Maximum Depth of the HP Wedge ($d_w$)

The maximum depth of the HP wedge is primarily a function of accretion rate; lower accretion rates result in deeper wedges (equation (2)); for constant $v_c$, $\theta$, and $x_n$. In order to maintain constant surface width of the wedge we scale erosion rates using the relationship derived from equation (3),

$$\frac{e}{a} = \frac{d_e}{x_n \sqrt{\tan \theta}}$$

6.2.1. HPD component (Figures 4a, 4b, and 4c, and Table 3). Increasing the maximum depth of the HP wedge ($d_w$) significantly raises upper plate maximum temperatures, increases temperatures somewhat throughout the hinterland, but has little effect on temperatures near the foreland. For example, increasing the maximum depth by ~50% raises the upper plate maximum temperature ($T_{max}$) by a factor of two, but temperatures near the foreland ($x$>100 km) are almost unchanged. In addition to controlling the magnitude of $T_{max}$ within the upper plate, the maximum depth of the HP wedge controls the location of $T_{max}$. Deeper wedges have maximum temperatures located deeper and farther away from the toe of the upper plate.

6.2.2. AD component, (Figures 4d, 4e, and 4f, and Table 3). Temperatures are cooler throughout most of the orogen with decreasing $a$ and $e$ (increasing $d_w$). The lower values of $a$ and $e$ result in steeper particle trajectories in the lower plate and consequently steeper lower plate isotherms and cooler temperatures near the subduction contact ($T_{max}$). Accretion of this cool material to the upper plate results in lower temperatures throughout the upper plate and lower near-surface thermal gradients (d$T$/dx$_{max}$).

6.2.3. Total thermal structures (HPD plus AD). The AD and HPD components of the thermal structure respond in opposite ways to the maximum depth of the HP wedge; deeper wedges lead to higher temperatures for the HPD component, but lower temperatures for the AD component. Whether or not total temperatures (HPD plus
Figure 3. Steady state temperatures as a function of subduction angle (θ) for constant vertical component of lower plate velocity (w=3.9 km/m.y.). Heavy lines show position of subduction contact. (a)-(d) The component of thermal structure due to heat production within the crust (with Tc=0). Dashed lines show geometry of wedge of heat production material. (e)-(h) The component of thermal structure due to heat from the asthenosphere (with Tc=0). Dashed line shows representative particle path with arrows scaled to legend. See text for further discussion. (Parameters are listed in Table 2.)
AD) increase or decrease with deeper wedges depends directly on the level of radiogenic heat. For low levels of AD (<0.5 μW/m²), the AD component dominates, and deeper wedges are associated with lower temperatures overall. For higher heat production rates, the HPD component dominates, and temperatures within the upper plate increase with deeper wedges.

6.3. Surface Width of HP Wedge ($s_w$)

Holding $v_\alpha$, $d_\alpha$, and $\theta$ constant, the surface width of the HP wedge ($s_w$) is primarily a function of erosion rate as given by the expanded version of equation (3): 

\[
\frac{d \theta}{\sigma} = \frac{v_\alpha \sin \theta - 1}{\frac{d\theta}{ds_w}}.
\]

In order to maintain constant maximum depth of the wedge, we scale accretion rate to erosion rate using the relationship derived from equation (2): 

\[
\frac{d \theta}{\sigma} = \frac{v_\alpha \sin \theta - 1}{\frac{d\theta}{ds_w}}.
\]

6.3.1. HPD component (Figures 5a, 5b, and 5c, and Table 3). Increasing the surface width of the HP wedge ($s_w$) enlarges the expanse of heated lithosphere but does not significantly affect maximum temperatures. The surface width of the HP wedge controls the horizontal position of the maximum temperature: $T_{\text{max}}$ is closer to the foreland, whereas $T_{\text{max}}$ for wider wedges is farther toward the hinterland. Temperatures near the foreland ($x<100$ km) are fairly insensitive to changes in the surface width of the HP wedge.

6.3.2. AD component (Figures 5d, 5e, and 5f, and Table 3). Changes to $e$ and $a$ associated with increasing $s_w$ result in lower temperatures in the hinterland ($x>250$ km), while temperatures near the foreland are not significantly affected. The lower erosion rates associated with wider wedges advect hot material from deeper levels more slowly, resulting in lower temperatures and near-surface thermal gradients ($T_{\text{mo}}$ and $T/ds_w$).

6.3.3. Total thermal structures (HPD plus AD). The two components of the thermal structures respond in opposite ways to variations in the surface width of the HP wedge: increasing $s_w$ raises maximum upper plate temperatures slightly for the HPD component, while leading to cooler upper plate temperatures for the AD component. Thus, for low levels of AD (<1.0 μW/m²) the AD component dominates and wider wedges are associated with slightly cooler temperatures. For higher heat production rates, the HPD component dominates and temperatures within the upper plate increase slightly with wider wedges.

6.4. Convergence Velocity ($v_\alpha$)

Varying the convergence velocity alone changes the geometry of the HP wedge (equations (2) and (3), constant $\theta$). Thus, in order to maintain constant $d_\alpha$ and $s_w$, erosion rate and accretion rate are scaled with $v_\alpha$ such that $v_\alpha$, $d_\alpha$, and $v$ are held constant.

6.4.1. HPD component: (Figures 6a, 6b, 6c, 6d, and 6e, and Table 3). Higher convergence velocities result in slightly lower temperatures throughout the orogen as both material and heat is moved through the system more rapidly. For example, comparison of Figures 6a and 6c shows that doubling $v_\alpha$ (along with $a$ and $e$) decreases maximum temperatures by less than 20%, while temperatures near the foreland ($x<100$ km) are not significantly affected. Upper plate maximum temperature are attained at shallower levels and farther from the toe of the upper plate for higher convergence velocities.

The relative insensitivity of the thermal structure to convergence velocity is independent of wedge geometry. Doubling $v_\alpha$ lowers maximum temperatures by <10% for very narrow HP wedges ($a=\tan \theta/2e=1$) and decreases maximum temperatures by ~20% for infinitely wide HP wedges ($\sigma=0$).
6.4.2. AD component (Figures 6f, 6g, 6h, 6i, and 6j, and Table 3). Increasing convergence velocity lowers temperatures slightly throughout much of the upper plate, raises the maximum near-surface thermal gradient, but has little effect on lower plate temperatures. Temperatures are cooler at the subduction contact due to the increased efficiency of subduction of cold crust with respect to conductive thermal relaxation, while higher maximum near-surface thermal gradients are due to higher erosion rates (\(T_{se} \) and \(T_{dz_{crust}}\)).

6.4.3. Total thermal structures (HPD plus AD). Higher convergence velocity results in lower temperatures for both the HPD and AD thermal structures; thus temperatures for the total thermal structure will also decrease with increasing convergence velocity.

7. Synthesis of Results

Subduction, accretion, and erosion redistribute HP material, resulting in the formation of an HP wedge within the upper plate of collisional orogens. Maximum depth of the HP wedge and the rate of heat production are the primary factors controlling thermal structure. High temperatures at shallow levels (>500°C at depths of 20-30 km) result from the combination of moderately deep wedges and moderate to high rates of heat production; i.e., \(d_p > 30 \text{ km} \) and \(A > 2.5 \mu \text{W/m}^2 \) or \(d_p > 65 \text{ km} \) and \(A > 1.25 \mu \text{W/m}^2 \) (Figures 7a and 7b). Combinations of shallow wedges and low heat production rates result in thermal structures that are cooler than the foreland geotherm, indicating that conductive heat loss to the downwelling plate exceeds heating from the HP wedge. Inverted geotherms develop over a broad range of combinations of \(d_{W} \) and \(A \), for example, \(d_p > 35 \text{ km} \) with \(A > 2.5 \mu \text{W/m}^2 \) and \(d_p > 65 \text{ km} \) with \(A > 0.75 \mu \text{W/m}^2 \) (Figure 8), while low values of heat production (\(A < 0.5 \)) and/or shallow wedges (\(d_{W} < 30 \text{ km} \)) do not produce temperature inversions. Higher convergence velocities reduce maximum temperatures; for example, doubling \(v_c \) while holding \(A \) and \(d_c \) constant, reduces temperatures by about 20%.

Varying the surface width of the HP wedge changes the horizontal location of upper plate maximum temperatures and thus their vertical proximity to the underlying subduction boundary. For example, given \(d_p = 44 \text{ km} \) and \(s_c = 440 \text{ km} \), \(T_{\text{max}} \) is located at \(z = 34 \text{ km} \) and \(x = 270 \text{ km} \), 20 km above the subduction boundary. For the same wedge maximum depth and \(s_c = 220 \text{ km} \), \(T_{\text{max}} \) is located at \(z = 32 \text{ km} \) and \(x = 190 \text{ km}, 6 \text{ km} \) above the subduction boundary.

Inverted geotherms develop within tens of millions of years, with the time required depending on the time needed to accumulate significant amounts of HP crust at intermediate depths (>35 km) within the upper plate, roughly ~20 m.y. for \(v_c = 20 \text{ km/m.y.} \) and \(v_c = 40 \text{ km/m.y.} \). High temperatures at shallow levels (>500°C at depths of 20-30 km) also occur within tens of millions of years, with the time required depending on both the rate of growth of the HP wedge and the magnitude of the heat production rate; for example, 10-15 m.y. for \(v_c = 40 \text{ km/m.y.} \) and \(A = 2.5 \mu \text{W/m}^2 \), and 45-55 m.y. for \(v_c = 20 \text{ km/m.y.} \) and \(A = 1.5 \mu \text{W/m}^2 \).
8. Geologic Applications

A broad range of paleotemperatures and pressures are observed within orogenic systems. If, as is suggested by this study, only a few fundamental processes control the primary thermal structure of orogenic belts, then the relative importance of these processes must vary from belt to belt. Inspection of Figure 7 suggests that variations in $A$ and $d$, can lead to large differences in the maximum temperatures obtained at shallow levels, providing a reasonable explanation of the observed spectrum of temperature structures within orogenic belts.

For example, one of the best documented occurrences of inverted metamorphic field gradients is in the central Himalayan orogen, where steep to inverted isograds above a Miocene intracontinental subduction boundary (the Main Central thrust zone) are associated with maximum temperatures in excess of 600°C at paleodepths of 15-30 km [Hodges et al., 1988; Hubbard, 1989; Pecher, 1989]. The development of steep to inverted field gradients, accompanied by significant crustal anatexis, began about 20-25 Ma in the central Himalayas, approximately 25-35 m.y. after the initiation of collision between India and Eurasia [Hubbard and Harrison, 1989; Hodges et al., 1994; Searle et al., 1988]. Rocks now at the surface were at depths of 20-40 km during Miocene metamorphism [Hodges et al., 1988; Pecher, 1989] and were subsequently denuded at average rates of 1-2 mm/yr. Additionally, extensive tracts of lower plate (Indian) rocks with an exposed surface width of the order of 300 km have been accreted episodically onto the upper (Eurasian) plate since the time of collision, yielding a vertical accretion rate of ~1.5 km/m.y. (for subduction angle of ~15°). Paleosubduction rates are unconstrained, but modern rates of convergence across the Himalayas are ~10-25 mm/yr. [Lyon-Caen and Moinar, 1985]. Heat production rates for metamorphic and igneous rocks of the Himalayas range from 1.5 to ~6 µW/m², with a significant proportion (>25%) of reported values in excess of 4 µW/m² [Scullet al. et al., 1990, Vidul et al., 1982, Macfarlane, 1992].

Using these estimates as constraints on model parameters, model results at ~32 m.y. show the development of inverted geotherms within the upper and lower plates with maximum temperatures in excess of 600°C at depths from ~20 km to 40 km (Figure 9). A detailed discussion of metamorphism resulting from the model thermal structure is beyond the scope of this paper and is the focus of a separate paper (A. Huerta et al., The effects of accretion, erosion, and radiogenic heat on the metamorphic evolution of collisional orogens, submitted to Tectonics, 1998). However, Figure 9 shows that because particle paths are parallel to the solid arrows, a vertical column of rock passing through the region of high temperatures in the upper plate preserves an inverted geotherm as an inverted metamorphic gradient. The model thermal structure is compatible with observations of metamorphism from the Himalayas, and it preserves inverted geotherms, thus demonstrating that it is probably not necessary to appeal to postulated transient heat sources (e.g., frictional heating along faults like the Main Central thrust zone) or postmetamorphic structures to explain the Miocene metamorphism of the Himalayas.
Figure 6. Steady state temperatures as a function of convergence velocity \( (v_c) \). Heavy lines shows position of subduction contact. (a)-(e) The component of thermal structure due to heat production within the crust (with \( T_c = 0 \)). Dashed lines show geometry of wedge of heat production material. (f)-(j) The component of thermal structure due to heat from the asthenosphere (with \( A - 0 \)). Dashed line shows representative particle path with arrows scaled to legend. See text for further discussion. (Parameters are listed in Table 2.)
9. Discussion

Unlike the transient heat sources postulated by other model studies, accretion and erosion are well-defined geologic processes known to operate in convergent orogens. When coupled with moderate rates of crustal heat production, these processes can result in high-grade metamorphism, partial melting, and the development of inverted geotherms at midcrustal levels. In addition, the broad range of metamorphic conditions associated with orogenic belts can be explained by variations in the magnitudes of these processes. Although other processes, such as shear heating, magma migration, and fluid flow, may have an effect on the thermal evolution of the crust during orogenesis, model results show that thermal regimes consistent with observations can be reproduced solely by redistributing HP crustal material through accretion and erosion. Because surface erosion and basal accretion exert such a strong control on temperatures, we feel that thermal models that do not include these processes are inadequate representations of orogenic systems.

In the approach taken here, intraplate deformation is neglected, a process which could increase the maximum depth of the HP wedge in the upper plate and result in higher temperatures than predicted by our model. In addition, we have assumed that accretion is spatially and temporally continuous throughout the orogenic cycle, although accretion in real orogenic belts may occur episodically along a series of discrete thrust fault systems. Incorporation of these processes will not change the basic conclusions of this study, although they are aspects of orogenic systems worthy of investigation in the future.

Our results may also yield insight into the development of extensional structures within collisional orogens. Such structures have been recently recognized in a number of collisional systems, but the mechanisms which trigger episodic extension or “extensional collapse” remain unknown [Northrup, 1996; Burchfield et al., 1992, and references therein; Gee et al., 1994; Hodges and Walker, 1992; Carmignani and Klugfeld, 1990]. Results of this

As another example, the Franciscan complex represents a low-temperature high pressure accretionary prism formed during Jurassic/Cretaceous subduction along the western coast of North America [Hamilton, 1969; Blake et al., 1988]. Composed of a number of tectonostratigraphic terranes, metamorphic grade increases from west to east, with estimated P-T conditions of ~250°C and 600 MPa in the west to maximum conditions of about 345°C and 800 MPa in the east [Blake et al., 1988, and references therein]. Accreted material consists of metasediments and basaltic metavolcanics, interpreted to be fragments of oceanic crust.

Using a low heat production rate to simulate subduction of oceanic crust (A=0.5 µW/m²) model temperatures consistent with observed conditions (of the order of 300°C to 400°C at z=20-30 km) are computed for a broad range of d_w from <30 km to >65 km (Figures 7a, and 7b). There are numerous combinations of erosion and accretion rates which can result in wedge depths ranging from 30 to 65 km, indicating that in this situation, the thermal evolution is relatively insensitive to the magnitudes of erosion and accretion. These results emphasize the fact that the primary control on the thermal evolution of convergent orogens exclusive of arc settings is the redistribution of crust enriched in heat-producing elements into the upper plate by accretion and erosion.

Figure 7. Maximum upper plate temperatures as a function of heat production rate (A) and depth to the base of the heat-producing wedge (d_w) (a) at z=20 km, and (b) at z=30 km. Hatched area indicates that temperatures within the orogen are cooler than foreland temperatures. (Parameters are listed in Table 2.)

Figure 8. Maximum upper plate temperatures for combinations of heat production rate (A) and depth to the base of the heat-producing wedge (d_w) which result in inverted geotherms. Hatched area indicates conditions that do not yield inverted geotherms within the upper plate. See text for further discussion.
study suggest that orogens in which significant amounts of HP material are accreted to the upper plate may experience temperatures high enough for thermally induced weakening and/or melting of upper plate rocks, conditions which may precipitate extensional collapse. Thus accretion and erosion may indirectly provide the rheologic conditions necessary for development of the extensional features observed within many collisional orogens.

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