A model for the origin of Himalayan anatexis and inverted metamorphism

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Abstract. The origin of the paired granite belts and inverted metamorphic sequences of the Himalaya has generally been ascribed to development of the Main Central Thrust (MCT). Although a variety of models have been proposed that link early Miocene anatexis with inverted metamorphism, recent dating studies indicate that recrystallization of elements of the MCT footwall occurred in the central Himalaya as recently as ~6 Ma. The recognition that hanging wall magmatism and footwall metamorphism are not spatially and temporally related renders unnecessary the need for exceptional physical conditions to explain generation of the High Himalayan leucogranites and North Himalayan granites, which differ in age, petrogenesis, and emplacement style. We suggest that their origin is linked to shear heating on a continuously active thrust that cuts through Indian supracrustal rocks that had previously experienced low degrees of partial melting. Numerical simulations assuming a shear stress of 30 MPa indicate that continuous slip on the Himalayan decollement beginning at 25 Ma could trigger partial melting reactions leading to formation of the High Himalayan granite chain between 25 and 20 Ma and the North Himalayan belt between 17 and 8 Ma. The ramp-flat geometry we apply to model the Himalayan thrust system requires that the presently exposed rocks of the hanging wall resided at middle crustal levels above the decollement throughout the early and middle Miocene. Late Miocene, out-of-sequence thrusting within the broad shear zone beneath the MCT provides a mechanism to bring these rocks to the surface in their present location (i.e., well to the north of the present tectonic front) and has the additional benefit of explaining how the inverted metamorphic sequences formed beneath the MCT. We envision that formation of the MCT zone involved successive accretion of tectonic slivers of the Lesser Himalayan Formations to the hanging wall and incorporate these effects into the model. The model predicts continued anatexis up to 400 km north of the Himalayan range, consistent with the timing and geochemistry of leucogranites exhumed on the flank of a south Tibetan rift.

1. Introduction

The signature geological association of the Himalayan range is the juxtaposition of inverted metamorphic sequences in the footwall of the Main Central Thrust (MCT) with a belt of early Miocene leucogranites emplaced above the fault (Figure 1). Understanding how these rocks relate to one another has long been considered key to comprehending the tectonic evolution of this mountain belt, and numerous models have been advanced to explain the igneous and metamorphic development of the MCT, largely on the basis of the assumption that the leucogranite and inverted metamorphism are temporally related. However, difficulties in accurately determining both granite crystallization ages and the timing of recrystallization of the footwall have until recently precluded direct tests of these hypotheses. The development of Th-Pb ion microprobe dating of monazite and its application to this problem [Harrison et al., 1995a, 1997a] produced results that challenge the above assumption and indicate that the inverted metamorphism formed largely during the late Miocene-Pliocene as footwall rocks were accreted to older metamorphic rocks in the MCT hanging wall. In this paper, we advance a model for the development of both the inverted metamorphism and paired granite belts that is consistent with their age relationships and established constraints on the tectonic history of the Himalaya.

2. Background

Prior to the collision of the Indian Shield and southern Tibet, which began at ~55 Ma [Rowley, 1996], the northern Indian margin was probably composed of a thinned cratonic wedge over which was draped both Proterozoic clastic deposits and the Cambrian-Eocene Tethyan shelf sequence [Le Fort, 1996]. The south directed thrust faults of the Himalaya, principally the MCT and the Main Boundary Thrust (MBT), sole in a common decollement at depth termed the Main Himalayan Thrust (MHT) [Brown et al., 1996]. In general, the MCT places the high-grade gneisses of the Greater Himalayan Crystallines atop schists and phyllites of the Lesser Himalayan Formations, and the MBT juxtaposes those rocks against unmetamorphosed Sub-Himalayan Neogene molasse (Figure 2). Protoliths of the Lesser Himalayan Formations
and Greater Himalayan Crystallines are interpreted to be Middle and Late Proterozoic clastic rocks, respectively [Parrish and Hodges, 1996].

Thermobarometry of the Greater Himalayan Crystallines in the central Himalaya indicate that pressures of ~8 kbar were achieved adjacent to the MCT, whereas values at the structurally highest levels were more typically 3–4 kbar [e.g., Hodges et al., 1988]. Deformation and anatexis within the MCT hanging wall were occurring at 22±1 Ma, while cooling ages at the top of the Greater Himalayan Crystallines suggest that ductile deformation had ceased there by ~18 Ma [e.g., Copeland et al., 1991; Hodges et al., 1996]. Juxtaposition of the Greater Himalayan Crystallines atop the Lesser Himalayan Formations is associated at most locations with an increase from chlorite- to kyanite-grade metamorphism with higher structural position (Figure 2). The region approximately bounded by the garnet isograd in the Lesser Himalayan Formations and the MCT is characterized by a highly sheared, 4–8 km thick zone of distributed deformation with a top-to-the-south shear sense, referred to as the “MCT Zone” [Pécher, 1989] (Figure 2).

Tertiary magmatism in the Himalaya is largely confined to two parallel granite belts: the High Himalayan leucogranites (HHL) and the North Himalayan granites (NHG) (Figure 1).
Table 1. Crystallization Ages of Himalayan Granites

<table>
<thead>
<tr>
<th>Pluton</th>
<th>Age, Ma</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zanskar</td>
<td>20.0±0.5</td>
<td>Noble and Searle [1995]</td>
</tr>
<tr>
<td>Gangotri</td>
<td>22.4±0.5</td>
<td>Harrison et al. [1997b]</td>
</tr>
<tr>
<td>Shivling</td>
<td>21.9±0.5</td>
<td>Harrison et al. [1997b]</td>
</tr>
<tr>
<td>Manaslu</td>
<td>23.0±0.4</td>
<td>Harrison et al. [1998a]</td>
</tr>
<tr>
<td>Shisha-Pangma</td>
<td>19.3±0.3</td>
<td>Harrison et al. [1998a]</td>
</tr>
<tr>
<td>Nyalam</td>
<td>20.2±0.2</td>
<td>Searle et al. [1997]</td>
</tr>
<tr>
<td>Nyalam</td>
<td>17.3±0.2</td>
<td>Searle et al. [1997]</td>
</tr>
<tr>
<td>Makalu</td>
<td>17.2±0.9</td>
<td>Schärer et al. [1986]</td>
</tr>
<tr>
<td>Makalu</td>
<td>23.0±1.0</td>
<td>Schärer [1984]</td>
</tr>
</tbody>
</table>

**North Himalayan granites**

<table>
<thead>
<tr>
<th>Pluton</th>
<th>Age, Ma</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dolpo</td>
<td>17.6±0.3</td>
<td>Harrison et al. [1997b]</td>
</tr>
<tr>
<td>Lhagoi Kangri</td>
<td>15.1±0.5</td>
<td>Schärer et al. [1986]</td>
</tr>
<tr>
<td>Maja</td>
<td>9.5±0.5</td>
<td>Schärer et al. [1986]</td>
</tr>
<tr>
<td>Khula Kangri</td>
<td>12.5±0.5</td>
<td>Edwards and Harrison [1997]</td>
</tr>
</tbody>
</table>

The HHL chain is exposed adjacent to, and on either side of, the South Tibetan Detachment System (STDS), which separates the Greater Himalayan Crystallines from lower-grade Tethyan shelf deposits in the hanging wall (Figure 7). In Table 1, we list what we consider to be reliable crystallization ages of Himalayan leucogranites. The eight dated pluons in the High Himalayan leucogranite belt range from 24.0 to 17.2 Ma (Table 1) and experienced peak melting temperatures of ~700°-750°C [Montel, 1993]. The NHG (Figure 1) are incised into the Tethyan cover rocks generally well above the STDS [Le Fort, 1986] and differ from the HHL in their emplacement style, younger ages (17.9 to 9.0 Ma; Table 1), and apparently higher melting temperatures [see Harrison et al., 1997b].

Four classes of models have been proposed to explain the relationship of inverted metamorphism and/or anatexis to large-scale faulting within the Himalaya. Although models in the first category differ in important respects, they share the assumption that anatexis and inverted metamorphism are spatially and temporally related to each other and result from slip on the Himalayan thrust system. These models include fluid influx from the subducting footwall coupled with lateral heat flow across the fault [Le Fort, 1975], frictional heating during thrusting [Arita, 1983; Molnar and Englund, 1990], delamination of the mantle lithosphere during continental subduction [Bird, 1978], radioactive heating alone or combined with other sources under prolonged deep crustal residence [Molnar et al., 1983], and accretion of highly radioactive crust to the MCT hanging wall coupled with high denudation [Royden, 1993]. A second type of model advanced by Nelson et al. [1996] proposes the opposite (i.e., thrusting within the Himalaya is caused by melting rather than being a consequence of it). No predictions are advanced regarding inverted metamorphism. The third class of models [e.g., Harris and Massey, 1994] are also unconcerned with inverted metamorphism and link anatexis to decompression melting related to slip on the STDS rather than to thrusting. The final category of models is primarily concerned with the inverted metamorphism and assumes no genetic relationship between anatexis and faulting. These models infer that the distribution of metamorphic assemblages resulted from subsequent deformation: folding of preexisting isograds [Searle and Rex, 1989], imbricate thrusting [Brunel and Kienast, 1986], and transposition of a normally zoned metamorphic sequence [Hubbard, 1996]. Because of their distal location, the origin of the North Himalayan granites has not been so inextricably tied to the development of Himalayan faulting. Their relative youth has been ascribed to a low rate of fluid infiltration across the MCT [Le Fort, 1986] and heat focusing by thermal refraction off low thermal conductivity Tethyan metasdomes [Pinet and Jaupart, 1987].

Despite the common assumption of a causal link between anatexis and inverted metamorphism, the timing of footwall recrystallization had not, until recently, been directly dated. Harrison et al. [1997a] utilized the fact that detrital monazite is destabilized in pelitic rocks during diagenesis but reappears under lower amphibolite grade conditions [e.g., Kingsbury et al., 1993] to establish that garnet-grade metamorphism affecting the Lesser Himalayan Formations occurred between about 8 and 6 Ma. They proposed that following termination of an early Miocene phase of slip along the MCT, the thrust reactivated at ~8 Ma followed by development of the broad shear zone underlying the MCT. In this view, the inverted metamorphism largely reflects juxtaposition of two tectonically unrelated, right-way-up metamorphic sequences. Numerical simulations of Harrison et al. [1997a] demonstrated that the above history met all isotopic and petrologic constraints, assuming a fault-bend-fold geometry for the MCT. An implication of recognizing that recrystallization of the MCT footwall is not temporally related to leucogranites production is that anatexis of the Greater Himalayan Crystallines need not be restricted to the Main Central Thrust ramp [cf. England et al., 1992].

Harrison et al. [1997b] explored an alternate model that ascribed the spatial and temporal variations of granite emplacement to continuous slip on a shallowly dipping decollement cutting through crust previously metamorphosed during the Neo-Himalayan phase (circa 55-35 Ma) of collision [Le Fort, 1996]. They assumed that, immediately prior to collision, the northern Indian margin resembled Figure 3a, and that during the initial phase of collision, the Greater Himalayan Crystallines protholith underwent dehydration and partial melting reactions [e.g., Pécher, 1989; Hodges et al., 1994, 1996] that produced a stratified paragenetic sequence in which metamorphic grade increased regularly with depth (Figure 3). Assuming that partial melting was characterized
the ice sheet and glacial isostatic adjustment (ISIA) of the Greenland Ice Sheet. The modeling results show that the Greenland Ice Sheet has been significantly affected by past ice sheet history and that the current state of the ice sheet is a result of the complex interactions between the ice sheet and the underlying bedrock.

The authors present a new model that incorporates the influence of past ice sheet history on the current state of the Greenland Ice Sheet. The model is based on a combination of ice sheet and bedrock deformation equations and includes the effects of past ice loading and unloading.

The results of the model show that the current state of the Greenland Ice Sheet is influenced by past ice sheet history, and that the current ice sheet thickness is significantly different from the initial state. The model also shows that the current state of the ice sheet is sensitive to changes in the climate and that the ice sheet is likely to respond to future climate changes.
ing physical mechanisms, we have been able to constrain a more complex model by taking advantage of details of the thrust history, geometry, and displacement rate that have only recently become available [e.g., Schelling and Arita, 1991; Brown et al., 1996; Blichar et al., 1997].

3. Numerical Simulation

3.1 Model Description

Our objective is to assess whether it is feasible to generate the temporally distinct Himalayan granite belts and inverted metamorphic sequences by continuous slip on the Main Himalayan Thrust. In doing this, we are constrained by geometric relationships (thrust dip angle, location of the granite belts with respect to each other, and the inverted metamorphic sequence), timing (granite crystallization ages, age of MCT footwall recrystallization, mineral age patterns), kinematics (shear sense, slip rates from active tectonics), thermal structure (heat flow observations, measured radioactive, thermomagneto), melting conditions (experimental studies of anatexis of appropriate lithologies), and physically imposed restrictions (ductile shear stress of ≤ 50 MPa). Within these constraints, we have chosen input parameters to a numerical model to create a thermotectonic simulation that is consistent with the known location and timing of igneous, metamorphic, and structural events.

Below, we describe a numerical model that relates granite formation, large-scale thrusting, and inverted metamorphism within the Himalaya. The model focuses upon the interval of time in question. While we key upon the spatial and temporal characteristics of melt generation within the hanging wall of the entire Himalayan thrust system from 25 to 8 Ma, we narrow our scope to examine deformation and recrystallization within the MCT footwall at later times. We refer to the 25-8 Ma and 8-2 Ma models below as phase I and phase II, respectively.

3.1.1 Eo-Himalayan crustal thickening. Accumulating evidence [e.g., Hodges et al., 1994, 1996; Coleman and Parrish, 1995; Parrish and Hodges, 1996; Edwards and Harrison, 1997] indicates that Proterozoic slope deposits overlying northern India were transformed into the Greater Himalayan Crystallines during Eo-Himalayan high grade recrystallization and anatexis that attended crustal thickening early in the collision. As discussed below, we allow for limited partial melting during this period in specifying the initial conditions at 25 Ma.

As indicated in Figure 3, we initiate our model using an assumed crustal geometry we believe appropriate for the mid-way point for the collision of India with Asia [see Harrison et al., 1997b]. We assume that the Himalayan thrust system initiated immediately above the craton and as a result employ a ramp-flat geometry to simulate crustal thickening in the north. We then model the evolution of the thrust system by assuming a fixed fault geometry with respect to the surface and flexural-bending deformation in both hanging wall and footwall [Suppe, 1983].

3.1.2 Displacement history. Estimates of the total amount of shortening based on balanced cross-section reconstructions average ~110 km within the Lesser and Sub-Himalaya, and the magnitude of thrusting of the Greater Himalayan Crystallines atop the Lesser Himalaya varies from >140 km to ~500 km [Schelling and Arita, 1991; Schelling, 1992; Srivastava and Mitra, 1994; DeCelles et al., 1998]. Thus perhaps half of the ~1000 km of shortening between the Indian Shield and southern Tibet [Chen et al., 1993; Paces et al., 1996] may have been absorbed within the Himalaya, with the remaining deformation taken up along Tethyan thrust systems [Ratsekhachev et al., 1994]. Assuming that the present convergence rate across the Himalaya of 20 mm/yr [Bilham et al., 1997] reasonably approximates the long term rate, and that this motion is equally partitioned into the hanging wall and footwall, the minimum ~250 km of shortening in the MHT hanging wall requires 75 Ma of slip.

Early displacement of the MHT hanging wall appears to have been accommodated by the MCT. While the timing of slip initiation is poorly known, several lines of evidence, including the accommodation of convergence elsewhere in the orogen prior to ~25 Ma [Yin et al., 1994], evidence of initial exposure of the GHC at circa 20 Ma [Richter et al., 1992], and the lack of pre-25 Ma sedimentary, metamorphic, or igneous products ascribable to slip on the MCT, all imply that significant displacement did not occur on the MCT prior to the early Miocene (see summary in Harrison et al. [1998b]). Cessation of anatexis and simple shear deformation in the MCT hanging wall by ~17 Ma suggests abandonment of the thrust at this time. Slip may have been transferred to the Main Boundary Thrust, a north-dipping fault marking the contact between the Lesser Himalaya and the underlying Miocene-Pliocene Siwalik Group [Johnson et al., 1982] (Figure 1). Changes in sedimentation patterns within the Himalayan foreland indicate that slip on the MBT began during the middle Miocene [Burbank et al., 1996]. The MCT ramp was reactivated at circa 8-4 Ma, with activity shifting progressively southward across the MCT Zone [Harrison et al., 1997a]. Although it is unclear whether synchronous (i.e., late Miocene-Pliocene) activity occurred along the MHT, the ramp has been recently active [Nakata, 1989]. Presently, the frontal ramp of the Himalayan thrust system (Figure 2) is represented by the active Main Frontal Thrust (MFT), which places Siwalik Group sediments atop the Quaternary Ganga basin [e.g., Schelling and Arita, 1991].

The slip history we employ for the Himalayan thrust system is summarized in Table 2 and schematically depicted in Figure 3. Thrusting begins after the crust has been thickened in response to ~25 m.y. of Eo-Himalayan crustal shortening (Figure 3a). We assume that the MHT is continuously active between 25 and 0 Ma at a rate of 20 mm/yr. The resulting 250 km of hanging wall shortening is accommodated as follows: slip occurs on the MHT/MCT ramp between 25 and 15 Ma (Figure 3a), on the MHT/MBT ramp between 15 and 8 Ma (Figure 3b), and on the MHT and various fault ramps defining the MCT Zone between 8 and 2 Ma (Figures 3c-3e). The MCT ramp that was abandoned at 15 Ma is then rotated by ramp-flat deformation associated with the subsequently activated thrusts (i.e., MBT and MCT-I). As will be discussed later in detail, this causes the dip of the MCT ramp to increase from 7° to 14° between 15 and 8 Ma, and to 30° between 8 and 2 Ma (Figures 3b-3c). New north-dipping thrust faults formed in the vicinity of the abandoned MCT strand (Figures 3c-3e) were also assigned the present observed inclinations of ~30° [e.g., Schelling, 1992; Hubbard, 1989].

The sequence of faulting outlined in Figures 3c-3e for the interval between 8 and 2 Ma is intended to simulate the pattern of out-of-sequence thrusting indicated by recent tectono-stratigraphic results [Harrison et al., 1997a; Catlos et al., 1998].
Table 2. Slip History of Himalayan Thrusts

<table>
<thead>
<tr>
<th>Time, Ma</th>
<th>MHT, km</th>
<th>MCT, km</th>
<th>MBT, km</th>
<th>MCT Zone, km</th>
<th>MFT, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>25-15</td>
<td>100</td>
<td>100</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>15-8</td>
<td>70</td>
<td>0</td>
<td>70</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>8-2</td>
<td>60</td>
<td>0</td>
<td>0</td>
<td>60</td>
<td>0</td>
</tr>
<tr>
<td>2-0</td>
<td>20</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>20</td>
</tr>
</tbody>
</table>

As indicated, we accommodate slip from 8 and 6 Ma along a fault equivalent to Arita's [1983] MCT-I that cuts the original MCT at depth. We position this new fault subparallel to, but ~6 km outboard from the preexisting MCT ramp (Figure 3d). The northward decrease in calculated pressures from the MCT hanging wall [e.g., Hodges et al., 1988; Hubbard, 1989; Inger and Harris, 1992; Kaneko, 1995; MacFarlane, 1995] indicates a ramp-flat geometry for this new structure. The observed contrast in structural level (from ~8 to 4 kbar) is best explained by positioning the flat of this thrust during the late Miocene at 25 km (7 km above the MHT). At 6 Ma, we introduce a second, geometrically similar, out-of-sequence thrust 13 km structurally outboard of the original MCT ramp (Figure 3d) to reproduce the presently observed metamorphic inversion within the MCT Zone [e.g., Hodges et al., 1988].

Note that we have not accounted in our model for activity on the STDS and the north-directed Renbu Zedong Thrust (RZT), both of which were active during the Miocene (Figure 2). In order to obtain an upper bound on the conditions necessary to produce the observed pattern of Himalayan anatectic and inverted metamorphism by shear heating alone, we have not incorporated tectonic denudation produced by normal slip on the STDS. Moreover, activity on the RZT between 17 and 11 Ma [Ratschbacher et al., 1994; Quideau et al., 1997] would not have influenced the thermal structure of the Greater Himalayan Crystallines in the region represented by our model.

3.1.3. Partial melting. For simplicity, our earlier investigation of the feasibility of shear heating producing the paired Himalayan granite belts assumed that the HHL and NHG were produced by minimum melting and muscovite dehydration melting, respectively [Harrison et al., 1997b]. In light of several emerging lines of evidence which appear to rule out wet melting as a source of the High Himalayan leucogranites evidence [e.g., Harris and Massey, 1994; Patiño Douce and Harris, 1998], we have chosen to consider only dehydration melting reactions in the present study. Experimental studies of muscovite and biotite dehydration melting indicate that multivariant reactions are involved, particularly for biotite breakdown [Vielzeuf and Holloway, 1988]. Solid solution involving sodium and Tschermak exchange have relatively minor effects upon the position of muscovite dry melting equilibria [Huang and Wyllie, 1973]. In contrast, dehydration melting of biotite-bearing assemblages involves either multivariant reactions that span a temperature range of more than 100°C for intermediate Fe/Fe+Mg compositions [Vielzeuf and Holloway, 1988; Le Breton and Thompson, 1988; Patiño Douce and Johnston, 1991; Patiño Douce and Harris, 1998] or a series of discontinuous reactions distributed over a similar temperature range for Mg-rich compositions [Carrington and Harley, 1995]. The lowest temperature granitic melts produced from biotite breakdown require that quartz, plagioclase, and aluminosilicate be present [e.g., Gardien et al., 1995]. Given a suitable phase assemblage, initial melting

Figure 4. Plot of melt fraction versus temperature for dry melting experiment performed on several different pelitic bulk compositions (see AKNa and AFM projections in inset). The gray solid line represents the bulk melting relationship used in the model.
triggered by biotite breakdown can take place below 800°C at 10 kbar [Le Breton and Thompson, 1988].

Rather than invoking a complex series of reactions to represent the dehydration melting process in our model, we have instead utilized simple, experimentally determined relationships between melt fraction and temperature [Gardien et al., 1995]. Such relationships for compositions representative of pelites in the Greater Himalayan Crystallines are depicted in the inset of Figure 4. The specific relationship used in our model is indicated by the thick line in Figure 4. Such a curve is appropriate for an intermediate pelitic bulk composition that would also adequately describe an intercalated sequence of muscovite-rich (comparatively fertile) and plagioclase-poor (comparatively infertile) lithologies. Specifically, muscovite-rich bulk compositions such as those studied by Vielzeuf and Holloway [1988] are characterized by a higher melt fraction at a given temperature, while plagioclase-poor assemblages such as those investigated by Pattino Douce and Johnston [1991] yield less melt at a given temperature.

3.2. Thermal Model

Thermobarometric histories are calculated using a two-dimensional 2-D finite-difference solution to the diffusion-advection equation. The region of calculation during phase I is enclosed in a 500 km wide x 60 km deep grid. The positions of faults active within the grid at various times are indicated schematically in Figure 4. In the model, only one fault is active at a given time, with the 20 mm/yr slip rate partitioned equally between the hanging wall and footwall. Fault displacement is varied linearly across a 1 km shear zone intended to simulate the basal decollement. As discussed above, dip angles of active fault ramps were maintained at 7° throughout phase I and at 30° during phase II. For the dip angle used in phase I, it was convenient to set the grid spacing to 900 x 876 such that fault planes coincided with individual grid points. For the same reason, the grid spacing for phase II was 900 x 187. Important model parameters are summarized in Table 3.

Constraints are imposed where melting can take place in the model. We adopt our previous approach [Harrison et al., 1997b] of assuming that rocks underlying the basal decollement consisted of refractory Indian basement that was not susceptible to partial melting at the temperatures realized in the model. This assumption limits anatexis to hanging wall rocks overlying the basal decollement. We further imposed an initial condition of 12% melting within these hanging wall rocks at all positions to the right of 400 km in order to simulate prior anatexis resulting from Eo-Himalayan crustal thickening and metamorphism that attended the early stages of the collision of India with Asia.

Table 3. Model Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>σ</td>
<td>shear stress = 10, 30, 50 MPa</td>
</tr>
<tr>
<td>V_h</td>
<td>hanging wall velocity = 10 mm/yr</td>
</tr>
<tr>
<td>V_f</td>
<td>footwall velocity = 10 mm/yr</td>
</tr>
<tr>
<td>V</td>
<td>[V_h - V_f] = 20 mm/yr</td>
</tr>
<tr>
<td>K</td>
<td>thermal conductivity = 2.5 W/(m-K)</td>
</tr>
<tr>
<td>κ</td>
<td>thermal diffusivity = 8 x 10^-7 m²/s</td>
</tr>
<tr>
<td>c_p</td>
<td>heat capacity = 1 kJ/(kg-K)</td>
</tr>
<tr>
<td>L</td>
<td>latent heat of fusion = 400 kJ/kg</td>
</tr>
<tr>
<td>Q_s</td>
<td>initial surface flux = 70 mW/m²</td>
</tr>
<tr>
<td>Q_b</td>
<td>basal heat flux = 30 mW/m²</td>
</tr>
<tr>
<td>z</td>
<td>radioactive scale length =15 km</td>
</tr>
</tbody>
</table>

![Figure 5](image_url)

Figure 5. Sequence of isothermal sections at (a) 15 Ma, (b) 8 Ma, (c) 6 Ma, and (d) 2 Ma indicating the position of the HHL and NHG source regions with time. Active and abandoned faults are represented by thick solid and dashed lines, respectively (see Figure 3). The dotted line indicates symmetry plane for ramp-flat fault geometry. Displacement histories of samples originating from within the Greater Himalayan Crystallines (GHC) and lower Lesser Himalayan Formations (LHF) are shown by thin solid lines. Open and solid symbols indicate initial and final positions in a given time frame. The isothermal contour interval is 100°C.
Frictional heating within the shear zone is approximated by the viscous dissipation between parallel walls (i.e., Couette flow). We assume that rock shear stress is largely unaffected by melting, as indicated by experiments using texturally equilibrated materials [e.g., Dell’Angelo and Tullis, 1988; Rushmer, 1996]. A constant stress of 30 M Pa [e.g., Stüwe and Sandiford, 1994; Rutter, 1997; Harrison et al., 1997b] and shear zone width of 1 km were used in our preferred model, but we also investigated somewhat broader zones (2 and 4 km), different shear stress values (10 and 50 M Pa), and temperature variation in the melting model (Figure 4) to investigate the sensitivity of the variation in these parameters on the space, time, and volumetric characteristics of anatexis. Thermal energy produced by dissipation was partitioned between heating the rocks and melting when conditions depicted by the continuous melting curve shown in Figure 5 were achieved. Conversion of thermal energy into melting was carried out in proportion to the latent heat of fusion (\(H_f = 400 \text{ kJ/kg}\)). Note that because the thermal energy required for melting is restricted to the shear zone, melt fractions are calculated relative to the width of this zone.

The model has zero-flux lateral boundaries, constant surface temperature (25°C), and a constant basal flux of 30 mW/m² [Englund et al., 1992] (Table 3). Implicit in the lateral boundary conditions is the assumption that the same characteristics specified for rocks at the right boundary (500 km) also apply for rocks extending at least 170 km beyond this boundary, reflecting the cumulative displacement along the decollement. The initial geotherm is calculated as the steady state condition resulting from a surface flux of 70 mW/m² and a distribution of radioactive heat sources within the crust. The latter was estimated by assuming an exponential distribution of radioactive elements with depth, a surface heat generation of 2.5 \(\mu\text{W/m²}\), and a characteristic scale length of 15 km as indicated by heat flow measurements from the northern Indian craton [Rao et al., 1976].

The model has zero-flux lateral boundaries, constant surface temperature (25°C), and a constant basal flux of 30 mW/m² [Englund et al., 1992] (Table 3). Implicit in the lateral boundary conditions is the assumption that the same characteristics specified for rocks at the right boundary (500 km) also apply for rocks extending at least 170 km beyond this boundary, reflecting the cumulative displacement along the decollement. The initial geotherm is calculated as the steady state condition resulting from a surface flux of 70 mW/m² and a distribution of radioactive heat sources within the crust. The latter was estimated by assuming an exponential distribution of radioactive elements with depth, a surface heat generation of 2.5 \(\mu\text{W/m²}\), and a characteristic scale length of 15 km as indicated by heat flow measurements from the northern Indian craton [Rao et al., 1976].

4. Results

4.1. Phase I (25-8 Ma)

 Isothermal distributions produced by the phase I thermal model, positions of active faults, and locations of source regions for HHL and NHG intrusions are shown in Figures 5a and 5b for times of 15 and 8 Ma. The cumulative melt fraction produced within the shear zone by the end of phase I (8 Ma, or 17 m.y. after slip began) is plotted as a function of distance from the left boundary in Figure 6a. Note that in Figure 5, we have plotted only positions of source regions that have experienced at least 2% partial melting. At 8 Ma, hanging wall source regions for lower temperature melts related to muscovite breakdown (i.e., <750°C; Figure 4) are situated between 140 and 230 km while low-percentage melt fractions related to higher-temperature anatexis involving biotite breakdown (i.e., >785°C; Figure 4) occur beyond 250 km. Note that the abrupt truncation of the low-temperature source region along its northern boundary at 230 km is simply a consequence of the initial condition of 12% low-temperature partial melting we have imposed for all grid positions further to the right (see section 3.2).

Age contours superposed onto Figure 6a indicate the progress of melt production. Because hanging wall temperatures along the flat are reduced as the ramp is approached, the
amount of melt generated at a given time increases systematically to the north. The time that melting ceased at a given position in the hanging wall above the decollement is shown in Figure 6b. Contours indicate the melt fraction produced as a function of position and time. As shown, granites generated via lower temperature anatexis in the model formed largely within a 2 m.y. interval following the onset of slip at 25 Ma. In comparison, igneous bodies produced by higher-temperature, biotite dehydration melting in the model began to aggregate 7 m.y. after slip began (i.e., at 18 Ma) and were produced steadily thereafter until displacement was terminated on the 35 km flat at 8 Ma (Figure 3). Were slip to have been maintained at this level beyond this time, melting would have continued indefinitely.

The rate of progress of the melt cessation front (~16 km/m.y. to the north; Figure 6b) is similar to, though greater than, the rate at which hanging wall positions along the flat approach the ramp vertex in the south (10 km/m.y.). This difference is largely due to conductive heat loss to the subducting footwall, which is most pronounced in the south. Contours indicating the time and position of a given melt fraction in Figure 6b reflect the interplay between shear heating and heat transfer to the subducting footwall. As indicated by these contours, temperatures required for a given melt fraction occur first in the north and then rapidly propagate southward only to be slowed by heat conduction to the footwall. The latter ultimately reduces temperatures below those required for melting.

As discussed above, low-temperature melting in our model begins almost immediately after slip is initiated along the decollement at 25 Ma. This outcome is a consequence of the fact that the initial temperature at the base of the shear zone was only 8°C below that required for the onset of melting (Figure 4, 740°C). Although the onset of melting could have been delayed by placing the flat at a shallow depth, the square root dependence of the heating rate is an important limiting factor [see Harrison et al., 1997b]. For example, in the present case where the base of the shear zone is initially at 732°C, ~7 m.y. are required before the first melt related to initial biotite breakdown reaches 785°C appears. Setting the initial temperature of the thrust flat 10°C lower further delays the appearance of initial high temperature melt by an additional 3 m.y.

The effect of varying key model parameters on the melting history is shown in Figure 7. The integrated melt area produced by the relationship shown in Figure 4 is 24 km². By displacing the melting relationship (Figure 4) downward 10°C, the integrated melt area increases to 34 km² (Figure 7a). Because low temperature melting is limited to 12%, this increase largely reflects greater degrees of high-temperature melting and an overall expansion of the high-temperature melt field. Note, for example, that the loci of low- and high-temperature melting along the decollement can no longer be recognized at the 2% level. Moreover, because melting is possible at lower temperatures, the first appearance of low-temperature melt in the south shifts ~5 km closer to the ramp apex. In contrast, displacing the melting relationship to high temperatures by 10°C causes overall melt production to decline to 15 km². Low-temperature melt production dips below 12% for positions to the south of 200 km, while the appearance of the high-temperature melt field is delayed by 3 m.y. and contracts such that the separation of the source regions increases to ~100 km at the 2% level.

The effect of increasing the dimensions of the shear zone demonstrate the ability of the model to melt increasingly larger volumes of rock (Figure 7b). For example, at 500 km, the degree of partial melting produced within 1, 2, and 4 km thick shear zones is 12%, 9%, and 7%, respectively. The integrated melt areas corresponding to 1, 2, and 4 km thick shear zones are 24, 40, and 63 km³. Note that because we have maintained the base of the shear zone at the same depth in performing these calculations, the effects observed do not reflect the influence of increased temperature with depth.

Varying the flow stress within the shear zone indicates that the minimum shear stress required to cause melting by biotite breakdown is >10 MPa (Figure 7c). At 10 MPa, the maximum amount of low-temperature melt produced is ~4% at 230 km. No high-temperature melting has occurred 17 m.y.

Figure 7. Variation in melt fraction versus source region position in the hanging wall caused by changes in the model parameters. (a) Effect of shifting the melting relationship of Figure 4 downward 10°C in temperature (dashed line). A positive 10°C shift produces the opposite effect (dotted line). The solid line represents no shift in temperature. (b) Effect of increasing the thickness of the shear zone from 1 km (solid line) to 2 km (dashed line) and 4 km (dotted line). Note that although the fraction of melt produced decreases, the overall melt volume increases significantly. (c) Effect of changing shear stress from 50 MPa (solid line) to 10 MPa (dashed line) and 5 MPa (dotted line).
4.2. Phase II (8-2 Ma)

The predicted tectonothermal evolution of the MCT Zone and adjacent regions is shown in Figures 5c and 5d. Note that the position of original MCT ramp, which was active from 25 to 15 Ma, coincides with the granite source regions. To illustrate the contrasting thermal histories of the major lithotectonic units, we monitor the position of five samples that reach the presently exposed surface. Samples at progressively greater structural distances above the MCT ramp are referred to as lower, middle, and upper Greater Himalayan Crystallines and correspond to samples NL25, U129, and U752 of Harrison et al. [1997a]. Sample DH71 [Catlos et al., 1997] is located within the upper Lesser Himalayan Formations (i.e., between the MCT and MCT-I), while AP332 [Harrison et al., 1997a] is positioned within the lower Lesser Himalayan Formations beneath the MCT-I. Note that AP332 first subsducts beneath the MCT (8-6 Ma) and is then transported toward the surface in the hanging wall of the active ramp from 6 to 2 Ma.

The temperature-time histories experienced by the above mentioned samples are shown in Figure 8a. Samples situated at progressively higher structural levels above the MCT experienced systematically lower temperatures. Because of the inactivity of the MCT ramp since 15 Ma, the thermal history of DH71 from the MCT footwall achieves a thermal maximum at this time owing to accretion of the upper Lesser Himalayan Formations to the MCT hanging wall. Although the subsequent T-I path of DH71 largely parallels those traced by Greater Himalayan Crystallines samples, it experiences higher temperatures between 15 and 6 Ma owing to its greater depth in that time interval (Figure 5). Later accretion of the lower Lesser Himalayan Formations to the upper Lesser Himalayan Formations/Greater Himalayan Crystallines at 6 Ma produces a thermal maximum for AP332 at this time. Subsequent cooling experienced by AP332 essentially follows the same path defined by the overlying samples.

Calculated K-Ar biotite and Th-Pb monazite ages obtained from the model thermal histories and daughter product kinetics are plotted for comparison in Figures 8b and 8c as a function of structural distance from the MCT. Positions of the MCT, MCT-I, and 6 Ma thrust ramps are indicated. The width of each region corresponds to the structural thickness between the bounding faults. The predicted trend of *Ar*-Ar biotite ages (Figure 8b) assumes an infinite cylinder geometry, an effective diffusive radius of 350 μm [e.g., Copeland et al., 1987], and the *Ar* diffusion parameters of Grove and Harrison [1996]. Samples situated >10 km structurally above the MCT are expected to yield cooling ages of >10 Ma, whereas those >13 km beneath the MCT yield protolith ages largely unaffected by Miocene heating. In contrast, biotites located at <10 km from the MCT are predicted to drop to ages of 3 to 2 Ma because of rapid exhumation of the MCT Zone and overlying Greater Himalayan Crystallines at that time (Figures 5c and 5d).

Th-Pb ages as a function of structural position relative to the MCT (Figure 8c) are calculated for Neogenic monazites using Pb diffusion data [Smith and Gillett, 1997], a slab geometry, and a domain size of 15 to 50 μm, typical of monazite included in garnet porphyroblasts in the MCT Zone [e.g., Harrison et al., 1997a]. Separate calculations were performed for the Greater Himalayan Crystallines, upper Lesser Himalayan Formations, and lower Lesser Himalayan Formations to illustrate likely differences in monazite crystallization ages. We assume that newformed monazites within the rele-
vanl portion of the Greater Himalayan Crystallines formed at 25 Ma (the initiation age of the model) but that crystallization of neofomed monazite in both the upper (~15 Ma) and lower (~6 Ma) Lesser Himalayan Formations could take place only between the attainment of 500°C and the thermal maximum [Kingsbury et al., 1993].

As shown in Figure 8c, calculated Th-Pb monazite ages begin to be reduced because of Pb loss within 5 km of the MCT. Immediately above MCT, 15 μm monazites are predicted to yield ~15 Ma ages. Similarly sized monazites are also expected to have experienced significant Pb loss throughout the upper Lesser Himalayan Formations. The results indicate that monazite formed at 25 Ma will yield Th-Pb ages of 12-16 Ma. Because much lower temperatures (~550°C) attend accretion of the lower Lesser Himalayan Formations rocks at the exposed erosion level, monazite Th-Pb essentially records the time of crystallization (~6 Ma).

5. Discussion

We show in this section that our model reproduces known structural, petrological, and timing constraints on evolution of the Himalayan thrust system. To do so, however, requires imposition of specific geometric constraints that while consistent with existing knowledge, are not otherwise supported. Because small changes in model inputs (e.g., fault geometry, shear stress, slip rate) can produce large changes in predicted thermal histories, there is little point in exhaustively exploring model sensitivity across the full range of input parameters [cf. Ruppel and Hodges, 1994]. An implication of this is that having found conditions that yield good correspondence between model and observation, there will be very few alternative permutations of input parameters that will satisfy the constraints.

5.1. Spatiotemporal Predications of Himalayan Petrogenesis

5.1.1. Paired granite belts. In the absence of pronounced shear heating, hanging wall rocks juxtaposed against colder subducting materials across thrust ramps seem an unlikely source for granite formation. Although refrigeration due to underthrusting is important along a steeply dipping ramp, its effect is sufficiently diminished on the thrust flat such that relatively low shear stresses (i.e., 20-30 MPa) can generate enough heat to produce pelitic anatectic, even under vapor-absent conditions. The simple shear heating model we have outlined in this paper not only demonstrates this basic idea but predicts general features of the two Tertiary granite belts of the Himalaya.

Before describing the degree to which the model matches observation, we first state the assumptions in our interpretation. First, the parameter predicted by the model (i.e., the timing of melting) necessarily predates that yielded by U-Th-Pb monazite ages (the time of effective retention of Pb which likely approximates the time the coalesced magma crystallized at midcrustal depths). This is due to both the time required for monazite to crystallize from the melt and the appreciable Pb diffusion in monazite [Smith and Gillett, 1997] expected at magmatic temperatures [Montel, 1993; Guillot et al., 1995]. Second, as the timescale of emplacement must be very short relative to the duration of thrusting (in order to preserve inherited ages in restitic monazite cores), the magnitude of the lag between the onset of melting and granite emplacement is dictated by the time required to achieve the minimum melt fraction required for segregation of granitic liquids in an actively deforming and reerystallizing system. Recent work suggests that relatively low melt fractions (<10%) probably only apply [e.g., Dell’Angelo and Tullis, 1988; Rushmer, 1996]. In interpreting our model results, we have arbitrarily considered the critical melt fraction required to form mobile igneous bodies at 2% and assume a <=1 m.y. temporal lag between anatexis and granite emplacement into middle crust. We point out that while melt extraction at the 2% level may seem unrealistic, the melt fraction produced in the model (Figure 6 and 8) is based upon the entire width of the shear zone and average melting properties depicted in Figure 4. Locally within the shear zone, more fertile horizons (dashed line in Figure 4) produce appreciably higher melt fractions at a given temperature.

The restricted time interval indicated in Figure 6 for low-temperature melting agrees well with the observed span of HHL ages (Table 1). Although HHL ages shown there vary from 24 to 17 Ma, the vast majority of the melt was produced at 23±1 Ma [Le Fort et al., 1987] (Table 1). While comparison with the NHG ages is somewhat complicated by their relative paucity (Table 1), the correspondence of the four available ages (~17-10 Ma) with the model prediction is also promising. The 8 Ma position of hanging wall source regions shown in Figure 5 is critically dependent upon our assumptions regarding the position of the ramp/flat and the prior history of Eo-Himalayan melting. Thus it is unsurprising that we have been able to match the observed separation distance between the HHL and NHG at ~80 km (Figure 1 and 5). Note, however, that differences in the emplacement style and level of exposure preclude drawing an accurate division between the two belts. This is particularly true for lower-temperature IIIIL melts which are unlikely to have been sufficiently thermally energetic to ascend to the shallow crustal levels presently exposed in the Tethyan Himalaya.

The physical attributes of the two magmatic belts agree well with the melting model we have employed (Figure 4) and the calculated volume of magma is broadly consistent with that inferred from the present outcrop pattern (~3%; Le Fort et al. [1987]). The relatively viscous melts produced by lower-temperature melting reaction (ca. 740°C) are more likely to be emplaced as sills than diapirs. The HHL, which corresponds to the lower temperature portion of the melt generation curve (Figure 4), appear to have formed at temperatures between 700° and 750°C [Copeland et al., 1988; Montel, 1993; Scaillet et al., 1995] and are largely emplaced as syntectonic sills [Le Fort, 1986; Searle et al., 1993]. Although they may contain less water, the magmas formed by the higher-temperature reactions (>790°C) are more likely to be sufficiently buoyant and thermally energetic to ascend as diapirs into the middle crust. In fact, the higher temperature (>750°C) NHG melts, as indicated by noneutectic compositions and high light rare earth contents coupled with low monazite inheritance [Debon et al., 1986; Schärer et al., 1986; Montel, 1993], are generally exposed as semicircular bodies emplaced into relatively low grade Tethyan metasediments.

Finally, we point out that our model (Figure 3) predicts that the Greater Himalayan Crystallines immediately above the present exposure of the MCT did not experience temperatures high enough to cause widespread melting. This is consistent with the observation that the source rocks of the
presently exposed High Himalayan leucogranites cannot be traced to sillimanite migmatites immediately above the MCT which remain fertile at many locations for muscovite dehydration melting [e.g., Harris and Massey, 1994; Barbeys et al., 1996].

After abandonment of the MCT ramp at 15 Ma (Figure 5b), rocks in the hanging wall of the MBT remained at middle crustal levels (Figures 5b, 5c, and 8a). Exposure of the HHL at the surface in their present location (i.e., well to the north of the MBT) requires a subsequent out-of-sequence thrust (Figure 5d). As discussed below, recognition of the pattern of out-of-sequence thrusting involving the MCT Zone (Figure 3d and 3e) has the additional benefit of explaining key features of the inverted metamorphism beneath the MCT.

5.1.2. Inverted metamorphism. The absence of a pronounced structural or metamorphic break across the broad MCT Zone makes it difficult to recognize which surfaces accommodated significant displacement. As such, we have used recent geochronological results to aid our reconstruction of the tectonic evolution of the MCT (Figure 3). On the basis of results of Th-Pb dating of monazite inclusions within garnet and available thermobarometry, Harrison et al. [1997a] proposed that late Miocene reactivation of the MCT Zone caused subduction of lower Lesser Himalayan Formations rocks which reached 550°C at 25 km depths just prior to their accretion to the upward moving MCT hanging wall at ~6 Ma.

Additional Th-Pb monazite age and thermobarometric results from the upper Lesser Himalayan Formations [Cato et al., 1997] allow us to build upon the Harrison et al. [1997a] model. Cato et al. [1997] examined garnet-grade, lower Lesser Himalayan Formations rocks from the Darodi Khola and obtained late Miocene ages (7.9 Ma) from monazite inclusions in garnet that were similar to those described by Harrison et al. [1997a]. Interestingly, however, appreciably older (15-16 Ma) Th-Pb ages were yielded by garnet-hoisted monazites sampled from structurally higher, kyanite-grade rocks of the upper Lesser Himalayan Formations in the same drainage. The new results suggest possible modifications to the Harrison et al. [1997a] model for the evolution of the MCT Zone.

The thermal evolution predicted for the Lesser Himalayan Formations/Greater Himalayan Crystallines from the tectonic model outlined in Figure 3 is shown in Figure 8. This scenario has (1) kyanite-grade upper Lesser Himalayan Formations rocks accreting to the MCT hanging wall at 15 Ma after slip ceased along the original MCT ramp and (2) rocks of the lower Lesser Himalayan Formations accreting to the hanging wall at 6 Ma when slip was transferred from the MCT-I [Arresta, 1983] to a more southerly thrust ramp (Figures 3c and 3e). The predicted distribution of peak temperature conditions (~550°C for the lower Lesser Himalayan Formations and ~600°C for the upper Lesser Himalayan Formations; see Figure 8a) resembles available geothermometry for the central Himalaya [e.g., Hodges et al., 1988; Kaneko, 1995].

In addition to satisfying available geothermometry for the MCT Zone, we point out that the model also accounts for variation of metamorphic conditions within the MCT hanging wall. Immediately above the MCT, garnet-biotite pairs record temperatures up to 700°C [e.g., Hodges et al., 1988; Kaneko, 1995], similar to the model (Figure 8a). Thermobarometry of rocks collected in the Greater Himalayan Crystallines indicate that pressures as high as ~8 kbar were achieved adjacent to the MCT, whereas peak pressures at the top of the section were only ~3-4 kbar [e.g., Hodges et al., 1988]. As shown in Figure 5d, our model predicts that hanging wall rocks from the range of depths corresponding to these pressures would today be exposed at the surface together as a result of rotation associated with their sequential transport up the ramp.

5.2. Age Patterns Across the MCT Zone

Because of their contrasting daughter product retention properties, biotite K-Ar and monazite Th-Pb ages yield complementary thermochronologic information. Measured ages, from the Darodi Khola, Marsyandi, and Burhi Gandaki drainages (Figure 1), are projected onto Figures 8b and 8e as a function of structural distance from the MCT. Calculated structural distances above the MCT in these northerly-southwesterly sections may be complicated by STDs faulting. However, the master detachment at this location may predate the 23-19 Ma Manaslu intrusive complex [Harrison et al., 1995a; 1998a] in which case the section would have remained structurally intact since the early Miocene.

Biotite 40Ar/39Ar ages [Copeland et al., 1991; Edwards, 1995] along central Himalayan transects decrease from ~16 Ma in the upper Greater Himalayan Crystallines to as young as 2-3 Ma within the MCT Zone (Figure 8b). Below the MCT-I, biotite ages increase rapidly to pre-Tertiary ages. Samples within the MCT Zone yield scattered ages, with older samples generally exhibiting systematics suggestive of excess radiogenic "Ar." In interpreting these results, we therefore place highest reliance on the minimum ages obtained at a given location. We recognize, however, that real differences in mica closure ages between adjacent drainages (reflecting variable tectonic histories, mineral chemistry) are likely. In spite of these complexities, the distribution of mineral ages in the combined data set reveals three distinct thermal regimes: (1) a 16-2 Ma age range across the Greater Himalayan Crystallines southward into the MCT Zone, (2) a zone of scattered ages within the MCT Zone with minimum ages of ~2 Ma, and (3) a rapid rise of mica ages south of the MCT Zone to values as old as 900 Ma [see Copeland et al., 1991]. Correspondence between the pattern of biotite age predicted by the model and the measured values (Figure 8b) is reasonably good, particularly when allowances are made for excess "Ar" and interlayer variability. The ~16 Ma apparent ages calculated for positions high within the Greater Himalayan Crystallines record closure as these samples encountered the MCT-I ramp and were transported upward (Figure 5). The drop in age down to 2 Ma seen in samples within ~7 km of the MCT reflects differential exhumation produced by the late Miocene-Pliocene thrust ramps.

The distribution of Th-Pb monazite ages within and adjacent to the MCT Zone appears broadly compatible with the spatial distribution of thermal regimes proposed above. As described earlier, monazites situated structurally high above the MCT yield early Miocene ages (e.g., the Manaslu granite; Harrison et al. [1998a]). The expectation from the model is that monazites from these positions should yield Th-Pb ages essentially similar to the time of crystallization (Figure 8c). In contrast, monazites from kyanite-grade rocks from the upper Lesser Himalayan Formations and the structurally deeper portions of the Greater Himalayan Crystallines yield 16-11 Ma ages that appear to reflect diffusive loss (Figure 8c).
Support for such an interpretation comes from depth-profiling measurements made on monazite grains separated from a pegmatite immediately above the MCT. Th-Pb ages systematically increase from 4 to 15 Ma, indicating the existence of a Pb* diffusion profile within ~2 μm of grain surfaces [Catlos et al., 1997]. In the model, significant Pb loss (Figure 8b) results from ambient ~600°C middle to late Miocene conditions that attended isobaric transport of the upper Lesser Himalayan Formations subsequent to its accretion to the MC1 hanging wall at ~15 Ma (see LHF upper in Figure 5). Finally, neoformed Late Miocene monazites appear in lower Lesser Himalayan Formations rocks metamorphosed to garnet-grade yield garnet growth ages of 6–8 Ma [Harrison et al., 1997a; Catlos et al., 1997]. Both our model results and peak temperatures estimated from thermobarometry indicate that minimal Pb loss is likely to have occurred from these grains.

5.3. Comparison With Other Models

As summarized earlier, four distinct types of models have been proposed to explain the relationship of inverted metamorphism and/or anatexis to large-scale faulting within the Himalaya: (1) anatexis and inverted metamorphism are spatially and temporally related by thrusting; (2) thrusting results from anatexis; (3) anatexis results from normal faulting; and (4) inverted metamorphism is produced by progressive accretion of the footwall to the hanging wall. Our model is a hybrid of types 1 and 4. Specifically, while we view anatexis as resulting from thrusting in type 1, we maintain that the inverted metamorphism largely results from later deformation as in type 4. Below, we explore how our model compares with those previously advanced. We then provide additional comments regarding a possible role for tectonic denudation (i.e., type 3) and evidence bearing on thrusting produced by anatexis (i.e., type 2).

In seeking a causal relationship between Greater Himalayan Crystallines anatexis and inverted metamorphism, type 1 models generally required an extraordinary source of heat to maintain high temperatures in the Greater Himalayan Crystallines for melting while subduction of India refrigerates the system. For example, one of the earliest models of the thermal evolution of the MCT was that of Le Fort [1975], who proposed that thermal relaxation following nappe emplacement heated the footwall sufficiently to induce dehydration reactions. When introduced into the still hot hanging wall, these fluids flushed the Greater Himalayan Crystallines gneisses, producing the leucogranite melts. However, it was widely recognized that this mechanism alone was insufficient either to generate the high lateral thermal gradient needed to produce the inverted metamorphic sequences or to maintain temperatures in the hanging wall that were sufficiently high to permit anatexis. Le Fort’s [1975] suggestion that shear heating along the thrust could sustain the temperatures required by these petrogenetic processes subsequently received a great deal of attention [e.g., Graham and England, 1976; Arita, 1983; Molnar and England, 1990; England et al., 1992, England and Molnar, 1993]. For the Himalayan case, it was argued that the observed thermal structure requires flow stresses in the range 100-1100 MPa [e.g., England and Molnar, 1993]. This range far exceeds an upper bound for ductile shearing of ~10-30 MPa inferred from laboratory deformation experiments [e.g., Engelder, 1993] or tectonic modeling [e.g., Kong and Bird, 1996].

Other models have employed mantle lithosphere delamination [Bird, 1978], high radioactive [Molnar et al., 1983], and rapid erosion [Royer, 1993] to create temperature conditions appropriate to both anatexis and inverted metamorphism. The recognition that recrystallization of the MCT footwall is a late Miocene phenomenon, and thus not temporally related to production of the Himalayan leucogranites, and the corollary that Greater Himalayan Crystalline anatexis need not be restricted to the MCT ramp, has largely removed the need for exceptional heat sources of the types outlined above. Moreover, while all models are capable of generating granite in the hanging wall, none can account for hanging wall anatexis and footwall metamorphism separated in time by more than 10-15 Ma [e.g., Harrison et al., 1997a; Catlos et al., 1997]. Thus a type 4 mechanism seems required to produce the inverted metamorphism within the footwall and juxtapose it against the granite source region in the hanging wall. Although models based on folding of preexisting isograds [Searle and Rex, 1989], imbricate thrusting [Brunel and Kienast, 1986], or ductile shearing of an existing zone of right-way-up metamorphism [Hubbard, 1996] are capable of producing this outcome, poor exposure and the lack of dateable products with which to establish the timing of deformation has previously limited their popular appeal. New techniques to constrain the timing of recrystallization defining the inverted metamorphism [e.g., Harrison et al., 1997a; Catlos et al., 1997] largely remove this criticism.

5.3.1 The role of tectonic denudation. The greatly contrasting structural levels juxtaposed across the STDs [Burchfiel et al., 1992] together with recognition of the importance of dehydration melting reactions in the Himalayan anatexis [e.g., Harris et al., 1993; Harris and Massey, 1994] leave open the possibility that decompression melting could have played a role in granite genesis within the Himalaya. We have not included such an effect in our model for three reasons. First, since our goal is to assess the feasibility of shear heating in producing all the igneous and metamorphic products of the Himalaya, ignoring the effects of other anatectic mechanisms provides us with an upper bound on the requirements for dissipative heating. Second, the pressure dependence for univariant muscovite dehydration melting at 35 km depth (~3.7°C/km; Huang and Wyllie [1973]) is small relative to the ambient geotherm expected at this depth (~15°C/km). A simple calculation incorporating decompression into the melting relationship (Figure 4) indicates that the 1.2 mm/yr denudation rate produced by 10 mm/yr slip on a 7° ramp (assuming uplift equals denudation) yields <<1% additional melting. Third, decomposition melting fails to predict significant features of Himalayan anatexis. For example, Harrison et al. [1998a] used numerical simulations of tectonic denudation due to extensional faulting to address the feasibility of decomposition melting in generating the Himalayan leucogranites. They concluded that slip rates nearly twice as high as the current rate of convergence across the Himalaya were required for this mechanism to generate even a single generation of anatexis, let alone the multiple phases documented. Maintaining these rates on 1 to 4 m.y. timescales seems unrealistic, particularly since mineral assemblages at the exposed erosion level of the Greater Himalayan Crystallines provide no indication of the ~10 12 kbar parageneses required by that model. Indeed, the magnitude of melting from even extraordinary decompression (~20-30 km)
is matched by only ~40 km of subhorizontal displacements along a thrust assuming modest (<50 MPa) shear stress [Harrison et al., 1998a].

5.3.2. Does thrusting result from anatexis? On the basis of seismic reflection profiling along a southern Tibet rift, Nelson et al. [1996] advocate a reversal of roles for thrusting and anatexis. They characterize the middle Tibetan crust as extensively partially molten today, and the region between the MCT and STDs as an earlier extruded equivalent (Figure 2). The HHL and NHG represent to them "progressively younger, frozen, snapshots of the partially molten mid-crustal layer" [Nelson et al., 1996, p. 1687] and they view the initiation of the MCT as being caused by melting rather than being a consequence of it. A potential test of the hypothesis is made possible by the exposure of middle crustal rocks in southern Tibet.

The western margin of the rift studied by Nelson et al. [1996] is bounded by a ductile normal fault (Figure 1) that was active between ~9 and 3 Ma and produced ~15-20 km of tectonic exhumation [Harrison et al., 1995b]. The result is that footrock wall exhumations are a record of the state of the middle crust at circa 9 Ma. Leucogranites yield U/Pb zircon ages at least as young as ~12 Ma [Xu et al., 1985]. The timing of ductile shearing (~29 Ma; Harrison et al. [1995b]) constrains the leucogranites to between 12 and 9 Ma. Tracer isotope studies of these rocks [Harris et al., 1988; X. Zhou, personal communication, 1997] suggest mixing between the local basement (Late Cretaceous-Eocene Gangdese batholith) and a Proterozoic component, similar to the Greater Himalayan Crystallines. Interestingly, Owens and Zondet [1997] interpreted the seismic velocity structure and magnitude of Poisson's ratio beneath southern Tibet as indicating that the Indian craton currently underthrusts southern Tibet. Hence we argue that the leucogranite exposures along the rift flank reflect the northward extension of melting predicted by our model.

6. Conclusions

We propose a thermotectonic model that unifies the formation of all key elements of Himalayan tectonics (the paired granite belts, large-scale thrusting, and inverted metamorphism) in terms of a constant displacement rate on the basal decollement. The numerical simulation we present is constrained by a variety of geological, geochronological, and geophysical observations. The model examines the consequences of the Himalayan thrust supporting a shear stress of 30 MPa during continuous slip over the past 25 Ma. In contrast to earlier models that linked early Miocene anatexis with inverted metamorphism, our starting point is the recognition that hanging wall magmatism and footwall metamorphism are not temporally related. The model we propose is capable of reproducing thermochronologic age constraints within and above the MCT Zone. In fact, the model also predicts that leucogranite generation occurred as much as 400 km north of the Himalayan range, consistent with the timing and geochemistry of leucogranites exhumed on the flank of a south Tibetan rift.

While the model predicts the timing and location of both the paired granite belts and inverted metamorphic sequences with high fidelity, it requires a highly specific chain of events in order to do so. Although these conditions are in reasonable agreement with the current knowledge of Himalayan tectonics, they underscore the model's sensitivity to small changes in certain parameters (e.g., location of isograds, initial temperature distribution). However, this may be one of the model's strengths, as the pair of contrasting age granite belts appears to be a unique feature of the Himalayan orogen. Had we found this feature to be an inescapable consequence of continental collision, we would be hard pressed to document a second example.

Our model results also clearly illustrate that rocks defining the inverted metamorphism on either side of the MCT may be temporally and spatially unrelated to each other. For example, rocks within the Greater Himalayan Crystallines just above the MCT achieve peak temperatures of ~750°C at 20 Ma, while samples from the upper and lower Lesser Himalayan Formations experienced peak temperature-time conditions of 600°C at 15 Ma and 550°C, respectively, at 6 Ma. Because of the 20 Myr convergence rate, these samples necessarily originated from positions that were separated by ~150 km at 25 Ma. An important implication of the model presented is that the structural feature recognized by Arita [1983] as the MCT-I may represent an important late Miocene structural break between the earlier metamorphosed upper Lesser Himalayan Formations and the latter recrystallized lower Lesser Himalayan Formations.

Acknowledgments. This research was sponsored by grants from the National Science Foundation and the UCPR program of Lawrence Livermore National Laboratory. Ho L. Ratcliffe is thanked for useful comments regarding likely temporal variation in the geometry of elements of the Himalayan thrust system and F. Spear is thanked for ideas relating to the continuous melting approach we employed. We thank C. Miller, F. Spear, and an anonymous reviewer for helpful comments.

References

HARRISON ET AL.: HIMALAYAN ANATEXIS AND INVERTED METAMORPHISM


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(Received November 19, 1997; revised June 8, 1998; accepted July 20, 1998)