Origin and Episodic Emplacement of the Manaslu Intrusive Complex, Central Himalaya

T. MARK HARRISON1*, MARTY GROVE1, KEVIN D. McKEEGAN1, C. D. COATH1, OSCAR M. LOVERA1 AND PATRICK LE FORT2

1DEPARTMENT OF EARTH AND SPACE SCIENCES AND INSTITUTE OF GEOPHYSICS AND PLANETARY PHYSICS, UNIVERSITY OF CALIFORNIA, LOS ANGELES, CA 90095-1567, USA
2CNRS, INSTITUT DOLOMIEU, 38301 GRENOBLE, FRANCE

RECEIVED JANUARY 20, 1998; REVISED TYPESCRIPT ACCEPTED MAY 14, 1998

INTRODUCTION

The Manaslu granite is the most studied of the dozen or so plutons that make up the High Himalayan leucogranite belt. The inferred relationship of the Manaslu granite with important Himalayan tectonic structures has inspired repeated attempts to determine its crystallization age, but several geochemical controls have hampered these geochronological investigations. These limitations are circumvented by Th-Pb ion microprobe dating of monazite. We have determined 158 Th-Pb ages on monazites separated from 11 samples of the Manaslu leucogranite. Together with an earlier published datum, these results indicate that the Manaslu intrusive complex was constructed during pulses of magmatism at 22±9 ± 0.6 Ma (Larkya La phase) and 19±3 ± 0.3 Ma (Bimtang phase). Monazite Th-Pb results for a single sample of the Dolpo-Mugu granite, located 75 km NW of the Manaslu intrusive complex, yield an age of 17±6 ± 0.3 Ma and suggest a protracted (~0.5 my) magmatic history that appears unrelated to the two phases of Manaslu magmatism. Age constraints on the emplacement of the Manaslu intrusive complex require that extensional structures cut by the Manaslu intrusive complex be 23±19 Ma or older. Geochemical constraints are consistent with the Larkya La phase of magmatism resulting from muscovite dehydration melting. The available timing and geochemical constraints are consistent with the Bimtang phase being produced from a higher-temperature pulse of dehydration melting. Thermo-kinematic modeling indicates that (1) two phases of leucogranite magmatism separated by ~4 my can be produced by shear heating along the Himalayan décollement assuming a shear stress of 45 MPa, and (2) scenarios involving only decompression to achieve the same result appear to require extreme conditions not permitted by available geological constraints.

KEY WORDS: Himalaya; Manaslu; Th-Pb; monazite; leucogranite

The High Himalayan leucogranites are a discontinuous chain of sheet-like plutons and dikes that extend along the crest of the Himalayan range (Fig. 1). Their possible interrelationship with both the Main Central Thrust (MCT) and South Tibetan Detachment System (STDS), arguably the two most significant tectonic features of that mountain belt (Fig. 1), has inspired several competing petrologic models. The classical interpretation that the High Himalayan leucogranites formed from fluid-induced anatexis of MCT hanging-wall rocks, triggered by devolatilization of overthrust footwall rocks in concert with shear heating (e.g. Le Fort, 1975; England et al., 1992), has been challenged by the view that the leucogranites instead formed by dehydration melting induced during decompression caused by slip on the STDS (e.g.
Harris et al., 1993; Guillot & Le Fort, 1995). This paradigm shift carries the implication that anatexis immediately followed substantial displacement along the STDS, but may not have been synchronous with motion on the MCT. If anatexis resulted from slip on the MCT, leucogranite crystallization ages provide a lower bound on the initiation of Himalayan thrusting. Although geochronological investigations suggest that the MCT hanging wall was actively deforming at ~22 Ma (e.g. Hodges et al., 1996), the age of MCT initiation is poorly known (England et al., 1992). Recent evidence indicates that the MCT ramp was reactivated during the late Miocene (Harrison et al., 1997a) and that early Miocene anatexis occurred immediately above the basal décollement of the Himalayan thrusts rather than along the ramp (Harrison et al., 1997b). If instead anatexis was driven by decompression melting caused by tectonic denudation, establishing the timing of emplacement of these intrusions constrains the displacement history of the STDS. Determining the deformation histories of these Himalayan master faults should permit better assessment of contrasting mechanical models that causally relate MCT thrusting and STDS extension (e.g. Burg et al., 1984; Burchfiel & Royden, 1985; Yin, 1993).

Although the tectonic significance of the High Himalayan leucogranites motivated numerous geochronological studies, these rocks have proven challenging to date. Because of the paucity of reliable crystallization ages, the few that have been obtained have been assumed to be representative of relatively large plutons that may in fact be composite bodies. Our goal in this paper is to exploit the recently developed monazite-Th-Pb ion microprobe dating method (Harrison et al., 1995b) to obtain the first detailed geochronology of a Himalayan leucogranite. Our results for the Manaslu granite document two phases of magmatism, constrain the ages of internal deformation that may be related to STDS activity, and provide the basis with which to investigate petrogenetic models of anatexis.

THE MANASLU GRANITE

Geological setting

The Manaslu granite, the most thoroughly studied of all the High Himalayan leucogranites (e.g. Le Fort, 1975, 1981; Vidal, 1978, 1982, 1984; Le Fort et al., 1982, 1987; Deniel et al., 1987; France-Lanord & Le Fort, 1988; France-Lanord et al., 1988; Copeland et al., 1990; Pécher, 1991; Guillot et al., 1993, 1994, 1995; Guillot & Le Fort, 1995; Harrison et al., 1995b; Chen et al., 1996), is a lenticular body of ~10 km maximum thickness that gently dips to the NNE with a relatively thin sheet, the Chhokang Arm, extending ~60 km to the east of the main body (Fig. 2). The granite is emplaced into, and cuts, a north-verging recumbent anticline developed within Tethyan
metasediments (Le Fort, 1981). The base of that fold is truncated by an STDS structure that is itself cut by granitic dikes. There is localized development within the pluton of ductile normal shear with a right lateral component indicating that internal deformation was occurring while the leucogranite was still at high temperature (Guillot et al., 1993, 1995). Emplacement appears to be largely by dike intrusion with the depth of the roof zone estimated to be ~12 km (Guillot et al., 1993, 1995). Beneath the STDS are gneisses of the Greater Himalayan Crystallines (GHC) that are themselves juxtaposed atop lower-grade schists of the Lesser Himalayan Formations by the south-directed MCT (Fig. 1). Based on its compositional and isotopic characteristics, the lowermost unit of the GHC (Formation I) is widely believed to represent the unmelted equivalent of the leucogranite source region (e.g. Le Fort, 1981; Barbey et al., 1996).

The Manaslu granite is composed of two leucogranite types; a two-mica variety and a less abundant (~20%) tourmaline leucogranite (Guillot & Le Fort, 1995). The generally lower Rb/Sr and initial $^{87}$Sr/$^{86}$Sr ratio ($^{87}$Sr/$^{86}$Sr) of the two-mica leucogranite is broadly similar to that of peraluminous metagraywackes in the Greater Himalayan Crystallines and suggested to Guillot & Le Fort (1995) that these rocks were the source of melting. Those workers further proposed that the tourmaline leucogranites, which appeared to them to have higher Rb/Sr and $^{87}$Sr/$^{86}$Sr, than the two-mica varieties, were derived via $\text{H}_2\text{O}$-undersaturated melting of GHC metapelites.

In general, the southwestern side of the pluton is characterized by a relatively fine grain size whereas the central and northeastern portions are generally coarser grained (Fig. 2). Although possibly a reflection in part of different levels of exposure of the laccolith, the textural contrasts between the northeastern and southwestern portions of the Manaslu granite are generally mimicked by several geochemical signatures. The coarser-grained central and northeastern portions of the pluton (Fig. 2) are associated with generally higher Rb/Sr (Fig. 3; see Table 2, below) higher $^{87}$Sr/$^{86}$Sr, ($^{87}$Sr/$^{86}$Sr, >0.752; Deniel et al., 1987), and lower Th/U ratios (<0.7) relative to the
southwestern side (Guillot, 1993; see Table 2). An exception to this trend is a restricted region on the southwestern side of the pluton which, in contrast to the encompassing region, is characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of $>0.752$ (Deniel et al., 1987). We note that these spatially, texturally, and geochemically distinct units contain both two-mica and tourmaline leucogranite varieties.

Fig. 3. Contour plot showing variation of $\text{Rb/Sr}$ within the Manaslu granite. It should be noted that the southwestern portion of the pluton is characterized by lower $\text{Rb/Sr}$ than the central and northeastern portions. Data are taken from Guillot (1993).

Prior age constraints

Problems inherent in dating High Himalayan leucogranites are in general due to the minimum melt character of these magmas. Heterogeneous $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ initial ratios (e.g. Vidal et al., 1982; Le Fort et al., 1987) preclude whole-rock isochron dating of the Manaslu granite except in rare cases (e.g. Deniel et al., 1987). U–Pb accessory mineral dating is complicated by the low solubilities of these phases in leucogranite magmas (Harrison & Watson, 1983; Montel, 1993) and the likelihood of their containing an inherited component (Copeland et al., 1980). In general, leucogranite zircons are highly contaminated by inherited Pb* and only rarely appear to yield meaningful crystallization ages (e.g. Schärer et al., 1986). Igneous monazites commonly incorporate significant $^{230}\text{Th}$ during crystallization; this results in the production of unsupported $^{206}\text{Pb}^*$ and obviates the advantages of U–Pb concordia. $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Manaslu granite and aureole yield either ~15–19 Ma cooling ages or older dates which reflect contamination by excess $^{40}\text{Ar}$ (e.g. Copeland et al., 1990; Guillot et al., 1994; Harrison & Mahon, 1995). Deniel et al. (1987) obtained a single conventional U–Pb monazite age for one Manaslu granite sample. The datum was slightly reversely discordant, because of the presence of unsupported $^{206}\text{Pb}^*$, and thus the $^{206}\text{Pb}/^{238}\text{U}$ age of 25.7 Ma was interpreted as indicating the age of crystallization (Deniel et al., 1987), a view that was later widely accepted (France-Lanord & Le Fort, 1988; Hodges et al., 1988a; Le Fort, 1988; Pécher, 1991; Chen et al., 1994; Guillot et al., 1994; Villa, 1996).

Harrison et al. (1995a) used $^{30}\text{Pb}/^{232}\text{Th}$ ion microprobe dating of monazite to constrain the crystallization age for one sample from the southeastern portion of the Manaslu pluton. Because of the short half-lives of the intermediate daughters of $^{232}\text{Th}$, this approach has the advantage of being insensitive to problems of disequilibrium Pb*. Moreover, the spatial resolution of ion
microprobe analysis allows measurements of sub-domains of individual monazite crystals so that the question of homogeneity of calculated ages can be explicitly examined. The measurements reported by Harrison et al. (1995a) revealed the presence of a single inherited monazite crystal which yielded ages of ~600 Ma. This result cast doubt on the reliability of the Deniel et al. (1987) datum as representing the emplacement age and led Harrison et al. (1995a) to instead conclude that this rock crystallized at 22.4 ± 0.5 Ma. Despite the seeming clarity of this interpretation, several important questions remain. These include: Is the Manaslu granite homogeneous with respect to crystallization age? Could the single inherited grain be a laboratory contaminant? Is the 22.4 Ma age actually dating metamorphic monazite in the protolith rather than magmatic crystallization? In plane are a function of initial energy and angle of ion emission (Montel, 1993; Scaillet et al., 1995), the di(Scaillet et al., 1995), the di
differences in the average surface density distributions of Th+ and Pb+ in these planes and therefore affect the relative sensitivity factor (RSF) for these species. Because sample charging and crater development are believed to be major contributors to these variations, care is taken to control the surface potential and crater depth during analysis. The depth/diameter ratios of the sputtered craters were maintained at <0.1 and monitoring of energy spectra indicates that changes in the average surface potential of the analyzed area under our measurement conditions were <1 eV. In addition, an aperture was inserted into the beam path at an image plane to reject secondary ions emitted from near the edge of the crater, where the extraction field is far from uniform. A further benefit of this optical gating is a substantial reduction of common Pb associated with surface contamination.

The energy distributions of Pb+ and Th+ secondary ions differ markedly, with Pb+ behaving in a fashion similar to a molecular ion (Harrison et al., 1995b). A plot of 205Th+ / 232Th+ vs 208Pb+ / 232Th+ for a standard monazite yields a linear array (the calibration curve) allowing the RSF to be derived by dividing the measured 208Pb+ / 232Th+ of the standard at a reference 264ThO2+ / 232Th+ value by its known 208Pb+ / 232Th (Fig. 4). The age of an unknown, measured under identical conditions, can then be determined by applying this RSF. For ~20 Ma monazites, the precision of the method is not limited by counting statistics but by the reproducibility of the calibration curve, which is typically ±2%. The accuracy of the method depends critically on usage of a well-characterized standard monazite.

Monazite standard 554

Our primary monazite standard is separated from sample 554, a mildly deformed peraluminous granodiorite from...
Whole-rock isotopic and trace element analysis

Previous work (e.g. Deniel et al., 1987; Guillot, 1993) has shown that the Manaslu granite is characterized by spatial variability in trace element concentrations and isotopic compositions over multiple length scales. To relate trends observed in the Th-Pb monazite dating to key geochemical indicators, we provide several trace element and isotopic results for the whole-rock samples used in the age determinations. Guided by the earlier investigations, we chose Rb/Sr, 87Sr/86Sr, and Th/U as these parameters. Whole-rock Sm-Nd and Pb isotopic results were also obtained for four samples. Data shown in bold were obtained at UCLA using the methods described by Nelson & Davidson (1993). All other isotopic results were undertaken at the University of Clermont-Ferrand and are compiled from Deniel et al. (1987) and Guillot (1993). U and Th were analyzed at CRPG Nancy using the methods of Govindaraju et al. (1976).

**RESULTS**

**Manaslu granite**

In an earlier paper we reported Th-Pb ion microprobe dating results on sample U315, located near the base of the pluton where the Chhokang Arm meets the main body of the Manaslu granite (Fig. 2). Two distinct age populations were recognized; the majority of ages defined a normal distribution with a mean age of 22.4 ± 0.5 Ma (±2 SE) that was interpreted to be the age of crystallization. A single inherited grain yielded ages of ~600 Ma (Harrison et al., 1995b).

High-density concentrates were obtained from an additional 11 spatially separated samples (~1 kg each) of the Manaslu granite (Fig. 2) from which monazites were hand picked. Each concentrate yielded ~40–60 monazite grains of ~30–80 μm in size. Twenty grains from each sample were mounted together with the 554 standard in an epoxy disk of 1 inch diameter, ground flat using 1200 SiC grit, polished using 1–1.4 μm diamond paste, cleaned, and sputter coated with ~200 Å of Au. Our initial analyses were undertaken using a 2–3 nA primary beam of O− with a spot size of ~30 μm diameter. All monazite ages reported have been corrected for common Pb based on the average Manaslu whole-rock 208Pb/204Pb ratio of 39.4 ± 0.1 (Table 2; Vidal et al., 1982). Results of all 186 Th-Pb age measurements are available from our web page (http://oro.ess.ucla.edu/ionprobe/manaslu.html), and are summarized in Table 1.

With the exception of XG43, all monazite samples analyzed under the analysis conditions described above yielded an early Miocene age peak with a distribution of individual ages similar in form to that previously seen for U315 (see Table 1). A substantial inherited component...
was present in most of the samples with prominent ages of ~30-40 Ma and ~100-600 Ma (Table 1). Age assessment of sample XG43 was initially precluded as portions of the pluton. We have previously noted (e.g. Edwards & Harrison, 1997) that the shape of restitic monazite cores inferred from ion microprobe analysis, not otherwise easily imaged by conventional methods, is substantially more irregular that that routinely observed for zircon. All eight samples in the older group (Table 1) yield a mean square weighted deviation (MSWD) of 1.4, consistent at the 2σ level with all four samples belonging to a single population. The MSWD of 3.7 for the older cluster is higher for these older rocks average 17.6 ± 2.7 Ma, with none in excess of 18.6 Ma (Copeland et al., 1990). Because no 40Ar/39Ar results exceed the interpreted Th–Pb age of the younger intrusive rocks, we infer that rocks in the older group remained at sufficient depth (>10 km; Guillot et al., 1995) to remain open to argon loss or to be reset (i.e. 23.8 to ~500 Ma) reflecting the common occurrence of restitic monazite, which precludes obtaining an estimate of the timing of crystallization, although we interpret the minimum age of 23.8 ± 0.6 Ma to be an upper age limit. With the view that increasing spatial resolution might overcome this problem, we repolished the sample mount and undertook a detailed characterization of a single grain (XG43f) using an analysis spot of ~5 μm diameter. Eighteen Th–Pb analyses of this grain (Fig. 5) revealed an inherited core with ages varying from 102 to 533 Ma and a magmatic rim with an age of 22.7 ± 0.5 Ma (Table 1), consistent with the seven other results from the central and northeastern portions of the pluton. We have previously noted (e.g. Edwards & Harrison, 1997) that the shape of restitic monazite cores inferred from ion microprobe analysis, not otherwise easily imaged by conventional methods, is substantially more irregular than that routinely observed for zircon. All eight samples in the older group (Table 1) yield a mean age of 22.9 ± 0.3 Ma, which we interpret as the crystallization interval. The 40Ar/39Ar mica ages for these older rocks average 17.6 ± 0.8 Ma, with none in excess of 18.6 Ma (Copeland et al., 1990). Because no 40Ar/39Ar results exceed the interpreted Th–Pb age of the younger intrusive rocks, we infer that rocks in the older group remained at sufficient depth (>10 km; Guillot et al., 1995) to remain open to argon loss or to be reset by the associated thermal pulse. It should be noted from Table 2 that this group is characterized by relatively high Rb/Sr (4-11) and 87Sr/86Sr (>0.752) ratios, suggestive of a derivation from a source containing a mica-rich, metapelitic component. It should be noted also, however, that the seven samples of this age group, although dominantly two-mica leucogranites, also contain tourmaline and two-mica tourmaline varieties. The four rocks that define the 19.3 ± 0.3 Ma phase are spatially restricted to the southwestern side of the

<table>
<thead>
<tr>
<th>Sample</th>
<th>x/y*</th>
<th>z (m)</th>
<th>Age (Ma, ± 2 SE)†</th>
<th>n/no. used‡</th>
<th>MSWD</th>
<th>Inheritance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Manaslu intrusive complex</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Larkya La phase</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DK89</td>
<td>408.20/38.10</td>
<td>4400</td>
<td>21.9 ± 0.5</td>
<td>9/8</td>
<td>2.0</td>
<td>36 Ma</td>
</tr>
<tr>
<td>U315§</td>
<td>428.25/21.97</td>
<td>3878</td>
<td>22.4 ± 0.5</td>
<td>28/24</td>
<td>1.8</td>
<td>584-613 Ma</td>
</tr>
<tr>
<td>XK43</td>
<td>416.96/36.78</td>
<td>4980</td>
<td>22.7 ± 0.5</td>
<td>38/9</td>
<td>2.6</td>
<td>26-533 Ma</td>
</tr>
<tr>
<td>XK46</td>
<td>416.91/35.81</td>
<td>4980</td>
<td>23.6 ± 0.7</td>
<td>13/5</td>
<td>1.0</td>
<td>29-413 Ma</td>
</tr>
<tr>
<td>XK56</td>
<td>414.83/35.04</td>
<td>5440</td>
<td>23.2 ± 0.5</td>
<td>11/10</td>
<td>5.7</td>
<td>38 Ma</td>
</tr>
<tr>
<td>XK102</td>
<td>414.14/34.89</td>
<td>5140</td>
<td>24.0 ± 0.8</td>
<td>12/5</td>
<td>1.7</td>
<td>5@33 Ma, 99-423 Ma</td>
</tr>
<tr>
<td>XK270</td>
<td>420.39/30.50</td>
<td>4548</td>
<td>22.9 ± 0.5</td>
<td>13/10</td>
<td>0.4</td>
<td>3@31 Ma</td>
</tr>
<tr>
<td>XP130</td>
<td>425.36/28.19</td>
<td>4090</td>
<td>23.0 ± 0.5</td>
<td>10/10</td>
<td>7.7</td>
<td>none</td>
</tr>
<tr>
<td>Dolpo-Mugu granite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LO68</td>
<td>83°48'/29°07'</td>
<td>17.6 ± 0.3</td>
<td>20/20</td>
<td>1.0</td>
<td>none</td>
<td></td>
</tr>
</tbody>
</table>

*Lambert grid map reference, Geological Survey of India.
†Interpreted crystallization age based on peak of the age distribution.
‡Number of spot analyses/number used in age calculation (i.e. number of inherited grains).
§Datum from Harrison et al. (1995a).
Table 2: Whole-rock isotopic and trace element data for Manaslu granite samples used for Th–Pb geochronology

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type*</th>
<th>Age (Ma)</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>87Rb/86Sr</th>
<th>87Sr/86Sr</th>
<th>Th (ppm)</th>
<th>U (ppm)</th>
<th>Th/U</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Larkya phase</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DK99</td>
<td>2M</td>
<td>21.9 ±0.7</td>
<td>311</td>
<td>87.0</td>
<td>10.35</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td></td>
</tr>
<tr>
<td>U315</td>
<td>2M</td>
<td>22.4 ±0.5</td>
<td>393</td>
<td>51.4</td>
<td>22.2</td>
<td>0.759326</td>
<td>n.d.</td>
<td>n.d.</td>
<td></td>
</tr>
<tr>
<td>XG43</td>
<td>T</td>
<td>22.7 ±0.5</td>
<td>438</td>
<td>47.0</td>
<td>25.63</td>
<td>0.773818</td>
<td>3.40</td>
<td>9.88</td>
<td>0.34</td>
</tr>
<tr>
<td>XG46</td>
<td>2MT</td>
<td>23.6 ±0.7</td>
<td>306</td>
<td>47.0</td>
<td>22.14</td>
<td>0.771027</td>
<td>2.89</td>
<td>11.9</td>
<td>0.24</td>
</tr>
<tr>
<td>XG56</td>
<td>2M</td>
<td>23.2 ±0.5</td>
<td>433</td>
<td>41.6</td>
<td>30.27</td>
<td>0.765090</td>
<td>3.51</td>
<td>13.6</td>
<td>0.26</td>
</tr>
<tr>
<td>XG102</td>
<td>2M</td>
<td>24.0 ±0.8</td>
<td>470</td>
<td>44.7</td>
<td>30.55</td>
<td>0.761066</td>
<td>4.85</td>
<td>15.9</td>
<td>0.31</td>
</tr>
<tr>
<td>XG270</td>
<td>2M</td>
<td>22.9 ±0.5</td>
<td>349</td>
<td>65.0</td>
<td>15.55</td>
<td>0.760331</td>
<td>3.00</td>
<td>9.35</td>
<td>0.32</td>
</tr>
<tr>
<td>XP130</td>
<td>2M</td>
<td>23.0 ±0.5</td>
<td>369</td>
<td>49.8</td>
<td>21.52</td>
<td>0.752333</td>
<td>3.00</td>
<td>9.35</td>
<td>0.32</td>
</tr>
<tr>
<td><strong>Bimtang phase</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DK203</td>
<td>2M</td>
<td>19.3 ±0.4</td>
<td>303</td>
<td>105</td>
<td>8.38</td>
<td>0.748998</td>
<td>6.50</td>
<td>13.8</td>
<td>0.47</td>
</tr>
<tr>
<td>DK208</td>
<td>2M</td>
<td>19.6 ±0.4</td>
<td>316</td>
<td>109</td>
<td>8.44</td>
<td>0.748189</td>
<td>8.83</td>
<td>3.94</td>
<td>2.24</td>
</tr>
<tr>
<td>XG162</td>
<td>2M</td>
<td>19.3 ±0.4</td>
<td>114</td>
<td>114</td>
<td>2.90</td>
<td>0.746491</td>
<td>5.17</td>
<td>4.82</td>
<td>1.07</td>
</tr>
<tr>
<td>XL24</td>
<td>2MT</td>
<td>19.0 ±0.4</td>
<td>276</td>
<td>83.0</td>
<td>9.63</td>
<td>0.744458</td>
<td>3.57</td>
<td>4.81</td>
<td>0.74</td>
</tr>
</tbody>
</table>

*2M, two-mica granite; T, tourmaline granite; 2MT, two-mica + tourmaline granite.

Data in bold were produced at UCLA. See section ‘Whole rock isotopic and trace element analysis’ for other sources.

Fig. 5. Photomicrograph of XG43 grain f showing location of spot analyses together with calculated ages in Ma. Although the Au coating is removed from a larger area, the actual size of analyzed region is ~5 μm in diameter.

pluton, where Deniel et al. (1987) obtained an Rb–Sr whole-rock isochron relationship on a suite of fine-grained samples (Fig. 2). Recalculation of Rb/Sr data using the algorithm of Mahon (1996) yields an age of 18.3 ± 0.5 Ma. Mica 40Ar/39Ar ages are distinctly younger (15.6 ± 0.6 Ma; Copeland et al., 1990) than those yielded by the older, coarser-grained leucogranite in the central and northeastern portions of the Manaslu granite and are consistent with evidence for greater emplacement depths (~20 km; Guillot et al., 1995). Compared with these rocks (Fig. 2), the ~19 Ma samples have a higher average Th/U (1.1 vs 0.35), and lower Rb/Sr (1.3 vs 4.11) and 87Sr/86Sr (<0.749 vs >0.752) (Table 2; Guillot, 1993). If we assume that these textural and geochemical distinctions are intrinsic to the younger phase and that we can therefore use them as a basis for estimating the fraction of the exposed pluton that was emplaced at 19.6–19.0 Ma, then the ~19 Ma phase represents about one-third of the present exposure of the Manaslu granite. Although this estimate is opposite to that inferred from the bimodal distribution of 87Sr/86Sr ratios (Guillot & Le Fort, 1995) that are dominated by the <0.752 group, geochemical sampling coverage is biased toward the southwestern side of the pluton (Guillot, 1993). Again, we note that the four samples contain both two-mica and two-mica tourmaline
leucogranite types. Lastly, we point out that the younger phase yielded no pre-Tertiary monazite ages (Table 1), consistent with much lower levels of restitic monazite compared with the older phase. Because the parameter most sensitive to monazite dissolution is melting temperature, this probably reflects higher peak temperatures for the younger granite. However, other effects, such as a decreased water activity, or a lower abundance of restitic monazite (or a large average grain size) in the source, could also potentially influence monazite dissolution.

The evidence presented for episodic injection of discrete magmas clearly indicates that the Manaslu granite is an intrusive complex and requires us to distinguish between the two magmatic phases thus far recognized. Henceforth, we shall refer to the younger, ~19 Ma, apparently lower \(^{87}\)Sr/\(^{86}\)Sr granitoid as the Bimtang phase, and the ~23 Ma, higher \(^{87}\)Sr/\(^{86}\)Sr intrusion as the Larkya La phase (see Fig. 3). Because these two plutons contain a complex mixture of two-mica and tourmaline leucogranites, petrologic models emphasizing the need for contrasting physical conditions in the genesis of the different granite types appear not to be supported by our observations (Guillot & Le Fort, 1995).

**Dolpo-Mugu granite**

The North Himalayan granite belt runs parallel to, and ~80 km to the north of, the High Himalaya (Fig. 1). In contrast to the generally sheet-like High Himalayan leucogranites, the North Himalayan granites belt is composed of ~16 elliptical plutons totaling 4000 km² in area that intrude into Tethyan metasedimentary rocks (Debon et al., 1986; Le Fort, 1986). They differ from the High Himalayan leucogranites in their emplacement style, younger ages (15–9 Ma), and higher melting temperatures as evidenced by noneutectic compositions, lower monazite inheritance, and a monazite saturation temperature of >760°C (Debon et al., 1986; Schaerer et al., 1986; Montel, 1993). Their relative youth with respect to the High Himalayan leucogranites has been ascribed to a low rate of fluid infiltration (Le Fort, 1986) and thermal refraction from the Tethyan sedimentary rocks coupled with low heat production (Pinet & Jaupart, 1987).

The Dolpo-Mugu granite (Le Fort & France-Lanord, 1995) is a 1600 km² pluton that straddles the High and North Himalayan belts (Fig. 1). The adjacent Mustang granite appears to have been circular in cross-section before development of the Thakkhola graben, which bisected the pluton and buried its eastern half (Fig. 1). The inferred circular shape of the Mustang granite and its alignment along strike with the northerly belt (Fig. 1) distinguishes it as a North Himalayan granite. Because the Dolpo-Mugu granite straddles the region separating the two belts and is relatively inaccessible, it is difficult to similarly characterize. We undertook Th–Pb monazite dating on a sample of the Dolpo-Mugu granite with the goal of distinguishing with which plutonic series it is most appropriately associated.

Twenty analyses of monazite separated from sample LO68 yield a single population (MSWD = 0.97) with a Th–Pb age of 17.6 ± 0.3 Ma (Table 1). Two splits of the LO68 monazite separate yield isotope dilution Th–Pb ages of 17.13 ± 0.12 and 17.77 ± 0.12 Ma (F. Oberli & M. Meier, personal communication, 1996). Although the isotope dilution ages are separated from each other at the 5σ level, their average value of 17.5 Ma is indistinguishable from the ion microprobe result of 17.6 Ma. Inheritance and/or Pb* loss could explain the difference between the isotope dilution ages, but the normal distribution and MSWD of the ion microprobe age distribution of unity tend to rule out these effects. More likely, this discordance represents a minimum crystallization interval of 0.4–0.6 m.y. It should be noted that the uncertainty on individual ion microprobe spot ages is typically ±0.5 Ma and thus our method is insufficiently precise to distinguish between a single event at 17.6 ± 0.3 Ma and a protracted magmatic history extending over 0.5 my. None the less, this result further demonstrates our ability to accurately date ~17 Ma monazites using our 45 Ma 554 monazite standard.

**DISCUSSION**

**Manaslu granite**

A plot of all 186 Th–Pb spot ages of Manaslu monazites (Fig. 6) clearly reveals two magmatic pulses at ~23 and ~19 Ma, and also shows the spectrum of ages resulting from the incorporation of restitic monazite containing inherited Pb. Interpreting this broad age spectrum in terms of protolith ages is complicated by the potential effects of Pan African magmatism, which is well documented to have affected the northern margin of the Indian craton, the inferred protolith of the Himalayan gneisses (Le Fort et al., 1982; Scharer & Allègre, 1983;
The behavior of monazite during anatexis is well understood in terms of its solubility in the melt (Montel, 1993). Once the melt reaches saturation, restitic monazite cores are stabilized in the melt and later provide a site for nucleation of new magmatic monazite. The kinetics of monazite dissolution suggests that crystals of ~50 μm will dissolve relatively rapidly to the point of saturation (Rapp & Watson, 1986) under conditions expected in High Himalayan leucogranite melts (i.e. ~5% H₂O, ~700–775°C; Scaillet et al., 1995). While exposed to the melt, restitic cores passively record the duration of that interval via the diffusive loss of their Pb* to the liquid. In conjunction with accessory mineral thermometry (Harrison & Watson, 1983; Montel, 1993; Ayres et al., 1997), which provides an estimate of peak melting temperature, we can use the ion microprobe to investigate the distribution of Pb* within inherited cores and recover this temporal information. Because of the high Pb* concentrations within inherited cores in these rocks, we can also utilize the U±Th±Pb systems to determine the degree of isotopic re-equilibration.

Le Fort et al., 1986; Copeland et al., 1988; Parrish & Hodges, 1996).

Assuming that the crystallization ages given in the caption of Fig. 1 are representative of the plutons from which they were sampled, it appears that about three-quarters of the ~8000 km² of leucogranite exposed along the crest of the High Himalaya (Le Fort et al., 1987), including the Larkya La phase of the Manaslu granite, was emplaced in the relatively brief interval between 24 and 22 Ma.

The Bimtang phase of the Manaslu granite, apparently spatially restricted to the structurally deeper, southwestern side of the intrusive complex, was emplaced at 19.3 ± 0.3 Ma, ~4 my after the Larkya La phase. A leucogranite stock intruded into the Greater Himalayan Crystallines ~30 km to the southwest of the Manaslu granite also yields ~19 Ma U–Pb monazite ages (Coleman & Parrish, 1995) and another small leucogranite body near Nyalam (Fig. 1), ~200 km to the east, yields a similar age (Scharer et al., 1986). Although it is at least arguable that magmatism in the High Himalaya was dominantly bimodal in age with peaks at 23 and 19 Ma, other small bodies yield ages of ~20 and ~17 Ma (Noble & Searle, 1995; Searle et al., 1997), leaving open the possibility that there may have been a continuum of magmatic activity from 24 to ~18 Ma. However, it is clear that models of High Himalayan leucogranite magmatism must account for a minimum duration of 5 my.

The persistence of the ~500 Ma protolith ages in samples with demonstrable ~23 Ma magmatic rims effectively rules out sample contamination as the source of this component. This same grain rim-core age relationship also tends to rule out the possibility that the ~23 Ma ages reflect metamorphic recrystallization. However, monazite that pre-dates magmatism by ~10–15 my is recognized and appears to be related to prograde crystallization of monazite during the Eo-Himalayan phase of collision (Le Fort, 1996).
Dolpo-Mugu granite
Crystallization between 17.9 and 17.1 Ma places this granite at the upper end of the range of $^{207}\text{Pb}/^{235}\text{U}$ monazite ages previously determined for North Himalayan granites by Schärer et al. (1986), but below the lower bound for the voluminous High Himalayan leucogranites (Harrison et al., 1997a). Although its structural location, age, and apparent lack of inheritance suggest that the Dolpo-Mugu granite is more akin to that belt than the High Himalayan chain (Harrison et al., 1997a), two issues raised by the present study condition that conclusion. The Dolpo-Mugu granite is one of the largest Tertiary plutons of the Himalayan range (Fig. 1) and, as we have found in the case of the Manaslu massif, one sample is insufficient to characterize a large body. Our results also raise the question of whether Himalayan granitoid magmatism is confined to two separate belts or is in fact a spatiotemporal continuum produced by progressive dehydration melting (Harrison et al., 1999).

Implications for the timing of deformation
Because the Manaslu intrusive complex crosses extensional structures in the hanging wall of the detachment, we infer that these features must be older than 19.23 Ma. The phase of syn-magmatic oblique-slip normal shearing within the Larkya La phase must have been active between 23.5 and 22.3 Ma. However, leucogranite dikes observed to crosscut the detachment immediately beneath the southeastern side of the pluton have not been dated.

U–Th–Pb monazite dating of other leucogranite bodies indicates a protracted slip history for the STDS throughout the Himalaya. In the western Himalaya, a small leucogranite in the Zanskar region beneath (and not cut by) the STDS yields an age of 19 ± 1 Ma (Noble & Searle, 1995). Footwall $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages between 18 and 15 Ma (T. M. Harrison, unpublished data, 1997) suggest that activity on the basal detachment was probably limited to the interval 20–17 Ma. Near Shisha Pangma, Searle et al. (1997) concluded that the age of normal faulting must largely post-date a leucogranite body dated at 17.3 ± 0.3 Ma. Lastly, in the eastern Himalaya (Fig. 1), the detachment remained active until at least 12.5 ± 0.4 Ma (Edwards & Harrison, 1997).

There are at least three possible explanations for this apparent diachronicity (see Edwards & Harrison, 1997). The first is that the STDS represents a single, northward propagating normal fault system, in which case the variations in age along strike are due to the rollback of the location of the basal detachment. Second, a topographic gradient capable of driving an STDS-type normal fault system may have initiated much later in the eastern Himalaya relative to locations further west. Third, the north-directed thrusts in southeastern Tibet (the Renbu Zedong Thrust System; Quidelleur et al., 1997) may have accommodated STDS displacement in the eastern Himalaya much later than in the western Himalaya.

A mechanism for diachronous leucogranite production from a single source region
Our results present a new challenge to the problem of leucogranite magma genesis. Specifically, how can hundred-km$^2$-sized bodies that differ in age by 4 my be generated from the same crustal source region during continental collision? Below we evaluate whether available models are individually capable, or able when acting in concert, to explain this phenomenon. Our criteria to judge the success of each model in the present context are its ability to predict (1) the emplacement of juxtaposed granite plutons of ~200 km$^2$ separated in emplacement age by 4 my, and (2) a decrease in $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr and concomitant increase in melting temperature for the younger phase relative to that of the older leucogranite.

Le Fort (1975) proposed that fluids introduced into the GHC from the dehydrating footwall following nappe emplacement fluxed the gneisses to produce the High Himalayan leucogranite melts via minimum melting reactions. Alternatively, Harris et al. (1993) proposed that the leucogranites were produced by higher-temperature, muscovite dehydration melting (e.g. Thompson, 1982) and became mobile only after tectonic decompression (via the STDS) markedly increased the melt fraction. The high Rb/Sr ratios observed in the Himalayan leucogranites seem to preclude the former and heavily favor dehydration melting. Specifically, measured leucogranite Rb/Sr values are typically 2–6 whereas those determined for the GHC source rocks lie in the range 0.7–1.4 (Vidal et al., 1982; Harris et al., 1993). Because the relative proportion of mica (the principal host of Rb in the protolith) to feldspar (the main repository for Sr) consumed in water-saturated melting is small, Rb/Sr ratios are not enriched in minimum melts. Conversely, as breakdown of mica during dehydration melting preferentially releases Rb with respect to Sr, this mechanism has the potential to produce the high Rb/Sr ratios observed for the leucogranites. Furthermore, recent experimental evidence indicates that water-saturated melting of Himalayan source rocks produces melt of tonalitic rather than leucogranitic composition (Fatíño Douce & Harris, 1996).

Although such arguments make it seem probable to us that dehydration melting was responsible for Himalayan anatexis, we consider it unlikely that both melting episodes were triggered by tectonic decompression. First, to induce even one melting event requires very high and

---

13
et al. (1997a, 1999) suggested that the origin of both the High and North Himalayan granite belts could be linked to shear heating on a continuously active décollement that cuts through Indian supracrustal rocks that were transformed into basement during the initial stages of the Indo-Asian collision. Harrison et al. (1999) approximated the deformation history of the Himalaya and then calculated the distribution of melting that would result from shear heating along the MCT décollement. Assuming a melting relationship that yields pulses of magma at temperatures of 740–750°C and >780°C (Fig. 7a), they found that two temporally distinct granite belts separated by ~80 km (corresponding to the High and North Himalayan chains) could form from shear stresses along the thrust as low as 20–30 MPa. Although High Himalayan leucogranite anatexis as a result of muscovite dehydration melting along the décollement was largely confined to a ~2 my interval in this model, Harrison et al. (1999) demonstrated that an increase in shear stress from 30 MPa to 50 MPa would result in additional, delayed melting from the same source region. Below we explore the extent to which the progressive melting model of Harrison et al. (1999) is capable of explaining the geochemical characteristics of the temporally distinct magmatic phases within the Manaslu intrusive complex.

Harrison et al. (1999) represented the integrated melting properties derived from melting of a source containing equal quantities of metapelite (long-dashed line) and metapsammite (short-dashed line). (b) Calculated melt fraction in the source region as a function of time since initiation of thrusting. The position of the source region is initially 230 km from the left edge of the numerical grid used by Harrison et al. (1999). The positions on the curves corresponding to the two melting pulses at 11% and 15% total melting are indicated as ‘Larkya La phase’ and ‘Bimtang phase’, respectively. The associated error bars reflect the ±2 SE uncertainty on the ages determined in this study. It should be noted that the contrast in age between the two melting peaks of ~4 my is the same as the difference between emplacement of the Larkya La (~22.9 ± 0.6 Ma) and Bimtang (~19.3 ± 0.3 Ma) phases. The duration of the melting event leading to emplacement of the Larkya La phase, estimated from consideration of the length scale of Pb diffusion in a restitic monazite core, is indicated.

Fig. 7. (a) Melting relationships of intercalated pelitic and psammitic source rocks used in calculations. The bold curve represents the bulk melting properties derived from melting of a source containing equal quantities of metapelite (long-dashed line) and metapsammite (short-dashed line). (b) Calculated melt fraction in the source region as a function of time since initiation of thrusting. The position of the source region is initially 230 km from the left edge of the numerical grid used by Harrison et al. (1999). The positions on the curves corresponding to the two melting pulses at 11% and 15% total melting are indicated as ‘Larkya La phase’ and ‘Bimtang phase’, respectively. The associated error bars reflect the ±2 SE uncertainty on the ages determined in this study. It should be noted that the contrast in age between the two melting peaks of ~4 my is the same as the difference between emplacement of the Larkya La (~22.9 ± 0.6 Ma) and Bimtang (~19.3 ± 0.3 Ma) phases. The duration of the melting event leading to emplacement of the Larkya La phase, estimated from consideration of the length scale of Pb diffusion in a restitic monazite core, is indicated.

sustained denudation rates. For example, the melting model of Harris & Massey (1994) requires >25 km of denudation to produce 8% melting because of decomposition, a value which exceeds the maximum structural thickness of the Greater Himalayan Crystallines (e.g. Hodges et al., 1988b). Although other melting models (e.g. Gardien et al., 1995; Fig. 7a) require less exhumation to achieve a similar effect, the issue remains as to whether this effect is capable of generating two magmatic pulses separated temporally by 4 my. We return to this question later, but conclude that tectonic decomposition by itself is not a viable mechanism.

A number of models have linked granite formation to shear heating along the Himalayan thrust system (e.g. Le Fort, 1973; England et al., 1992). Recently, Harrison et al. (1997a, 1999) suggested that the origin of both the High and North Himalayan granite belts could be linked to shear heating on a continuously active décollement that cuts through Indian supracrustal rocks that were transformed into basement during the initial stages of the Indo-Asian collision. Harrison et al. (1999) approximated the deformation history of the Himalaya and then calculated the distribution of melting that would result from shear heating along the MCT décollement. Assuming a melting relationship that yields pulses of magma at temperatures of 740–750°C and >780°C (Fig. 7a), they found that two temporally distinct granite belts separated by ~80 km (corresponding to the High and North Himalayan chains) could form from shear stresses along the thrust as low as 20–30 MPa. Although High Himalayan leucogranite anatexis as a result of muscovite dehydration melting along the décollement was largely confined to a ~2 my interval in this model, Harrison et al. (1999) demonstrated that an increase in shear stress from 30 MPa to 50 MPa would result in additional, delayed melting from the same source region. Below we explore the extent to which the progressive melting model of Harrison et al. (1999) is capable of explaining the geochemical characteristics of the temporally distinct magmatic phases within the Manaslu intrusive complex.

Harrison et al. (1999) represented the integrated melting properties of the source region by averaging experimentally determined melting relationships for various metasedimentary bulk compositions (Gardien et al., 1995). In the present discussion, we can consider the melt fraction vs temperature relationship as representing the contributions of two sub-equal interlayered components in the GHC source region: muscovite-rich, plagioclase-poor pelitic gneisses and muscovite-poor, plagioclase-rich psammitic rocks (Table 3). The relationships between melt fraction and temperature appropriate for each of these bulk compositions, based upon muscovite and biotite dehydration melting (Le Breton & Thompson, 1988; Gardien et al., 1995), are shown in Fig. 7a. It should be noted that the extent of lower-temperature melting (740–755°C) in each of the lithologies is dictated by muscovite abundance whereas biotite dehydration melting (>785°C) in the metapelite is limited by the complete consumption of plagioclase at higher temperature. Although certain lithologies present in the GHC will be more refractory to melting than indicated by these curves (e.g. Patiño Douce & Harris, 1998), we can rule out candidate source rocks that require melting temperatures >800°C (Montel, 1993).

Following the approach of Harrison et al. (1999), but assuming a 50% higher shear stress of 45 MPa, we calculated the melt fraction as a function of time for a given source region position (Fig. 7b). Although the higher shear stress could reflect the dehydration of the
Table 3: Trace element partitioning and isotopic variation during batch melting

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Unmelted source</th>
<th>Residual solid</th>
<th>Integrated</th>
<th>Residual solid</th>
<th>Integrated</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pelite</td>
<td>Psammite</td>
<td>Pelite</td>
<td>Psammite</td>
<td>Melt 1*</td>
</tr>
<tr>
<td>Quartz</td>
<td>27</td>
<td>31</td>
<td>21</td>
<td>30</td>
<td>26</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>6</td>
<td>10</td>
<td>1</td>
<td>9</td>
<td>0</td>
</tr>
<tr>
<td>K-feldspar</td>
<td>0</td>
<td>20</td>
<td>12</td>
<td>21</td>
<td>15</td>
</tr>
<tr>
<td>Muscovite</td>
<td>32</td>
<td>3</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Biotite</td>
<td>10</td>
<td>30</td>
<td>10</td>
<td>30</td>
<td>11</td>
</tr>
<tr>
<td>Garnet</td>
<td>20</td>
<td>6</td>
<td>20</td>
<td>6</td>
<td>26</td>
</tr>
<tr>
<td>Sillimanite</td>
<td>5</td>
<td>0</td>
<td>16</td>
<td>1</td>
<td>20</td>
</tr>
<tr>
<td>Melt</td>
<td>0</td>
<td>0</td>
<td>20</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>Total</td>
<td>100</td>
<td>100</td>
<td>100</td>
<td>100</td>
<td>100</td>
</tr>
<tr>
<td>Rb</td>
<td>109</td>
<td>81</td>
<td>182</td>
<td>73</td>
<td>172</td>
</tr>
<tr>
<td>Sr</td>
<td>75</td>
<td>254</td>
<td>52</td>
<td>97</td>
<td>57</td>
</tr>
<tr>
<td>Rb/Sr</td>
<td>1.1</td>
<td>0.32</td>
<td>3.5</td>
<td>0.75</td>
<td>3.0</td>
</tr>
<tr>
<td>$^{87}$Sr/$^{86}$Se</td>
<td>0.765</td>
<td>0.740</td>
<td>0.765</td>
<td>0.740</td>
<td>0.765</td>
</tr>
</tbody>
</table>

*Residual bulk partition coefficients: pelite Rb 0.50, Sr 1.53; psammite Rb 1.12, Sr 2.64.
†Residual bulk partition coefficients: pelite Rb 0.46, Sr 1.52; psammite Rb 1.05, Sr 2.68.

thrust surface following the initial phase of melting, variations of other melting parameters [see fig. 7 of Harrison et al. (1999)] could produce the required melting with a lower shear stress. Although we will later conclude that decompression alone cannot trigger the two episodes of leucogranite magmatism, this mechanism could work in concert with frictional heating to reduce the required flow stress to a few hundred bars, which we have previously argued could be maintained along a thrust surface at ~750°C (Harrison et al., 1997a). We also note that because both phases of the Manaslu intrusive complex are emplaced within the hanging wall of the STDS, the source of the ~19 Ma magmatic pulse could be significantly further north within the GHC than the ~23 Ma phase. The higher temperatures further down dip along the MCT would be conducive to the second melting episode appearing earlier than our model predictions and thus be consistent with a lower shear stress.

Coupled to the melting relationship (Fig. 7a), our thermal model predicts that ~11% melt is generated from the heat produced during the initial 1–2 my of slip along the décollement, and an additional 4% is produced after 4–5 my (Fig. 7b). It should be noted that the intervening ~4 my between the initial melting pulse and higher-temperature anatexis is comparable with the interval between emplacement of the Larkya La and Bimtang phases. In addition, the fraction of lower-temperature melt exceeds that produced at higher temperature by a factor of two to three, similar to the inferred proportions of the ~23 Ma and ~19 Ma leucogranite phases. Dropping the flow stress by a factor of two significantly increases the time before appearance of the second melt phase.

Assuming that the leucogranite source region can be associated with a lower shear stress. Although we will later conclude that decompression alone cannot trigger the two episodes characterized by the present exposure of Formation I within the GHC (Guillot & Le Fort, 1995), the melting history (Fig. 7) can be used to predict the trace element and isotope geochemistry of the Manaslu magmas. Rb/Sr and $^{87}$Sr/$^{86}$Se correlate well with bulk composition in these rocks (Guillot & Le Fort, 1995). On the basis of the data of Vidal et al. (1982) and Deniel et al. (1987), we characterize the metapelites as having Rb/Sr = 0.8 and $^{87}$Sr/$^{86}$Sr = 0.765 and the metapsammitic rocks by Rb/Sr = 0.8 and $^{87}$Sr/$^{86}$Sr = 0.740 (Table 3). Experimentally determined diffusion properties of Sr (e.g. Cherniak & Watson, 1992) of the constituent phases in the ~700°C source region lead us to assume that all phases within a given compositional band are equilibrated with respect to $^{87}$Sr/$^{86}$Sr at the time of anatexis. Using the expression for element partitioning during batch melting together with trace element partition coefficients (Henderson, 1982; Nash & Crecraft, 1985; Harris et al., 1995) yields $^{87}$Sr/$^{86}$Sr = 0.763 and Rb/Sr = 3 for the initial phase of 11% melting (Fig. 7 and Table 3) and $^{87}$Sr/$^{86}$Sr = 0.746 and Rb/Sr = 1.2 for the second, higher-temperature episode that produced an additional 4% melt (Fig. 7 and Table 3). The modal changes during partial melting are dictated by the stoichiometry of reactions specified by equations (2) and (3) of Harris et al. (1995), and the melt contributions from the intercalated
pelitic and psammitic sources are integrated to yield the above values (see Table 3). The values of $^{87}\text{Sr}/^{86}\text{Sr}$ predicted by the above calculations agree well with the bimodal distribution observed in the Manaslu complex (Guillot & Le Fort, 1995). Although the calculated values of Rb/Sr are at the lower end of those observed (Guillot & Le Fort, 1995), use of a continuous extraction melting model and different parameters (relative abundance of phases, melt fraction, partition coefficients) can easily generate the range of observed values (see Harris et al., 1993).

The progressive melting model accounts for the $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr systematics of the two intrusive phases, the contrast in emplacement ages, and potentially explains the observed lack of pre-Tertiary monazite inheritance in the Bimtang phase (Table 1) by virtue of its higher-temperature genesis.

Viability of decompression melting for diachronous magmatism

Recognition that most Himalayan anatexis occurred by dehydration melting reactions (e.g. Harris & Inger, 1992) stimulated increasing support for the view that melting was induced by decompression via slip along STDS (e.g. Harris et al., 1993; Harris & Massey, 1994; Guillot & Le Fort, 1995; Searle et al., 1997). Although this enthusiasm is understandable given the positive $dP/dT$ of vapor-absent equilibria, particularly for reactions involving muscovite, the physical viability of this mechanism has not been directly addressed. Data which document diachronous anatexis (Searle et al., 1997; this study) place more specific tectono-thermal requirements upon decompression melting as an explanation for Himalayan anatexis.

To address the feasibility of decompression melting to produce diachronous anatexis, we have undertaken numerical simulations of tectonic denudation owing to extensional faulting. Using the thermal model of Harrison et al. (1995a), $P$–$T$–$t$ paths for footwall source regions at differing horizontal distances from a normal fault dipping at 30° were calculated (Fig. 8). Although the present observed dip of the STD is much shallower (Burchfiel et al., 1992), it is unlikely that slip occurred at dip angles much lower than 30° (e.g. Jackson & McKenzie, 1983), and shallower inclinations would require sustained slip rates that are much higher than the total present convergence across the Himalaya (Bilham et al., 1997). To facilitate comparison with the shear-heating model just presented, we used an identical geothermal structure, monitored samples originating from the same initial depth of 35 km (i.e. 10 kbar), and interpret results based upon the same melting model (Fig. 7a). Calculated trajectories of footwall positions at differing distances from the normal fault are shown on a pressure–temperature diagram in Fig. 8 assuming a vertical component of slip of 10 mm/yr, 2.5 mm/yr, and 10 mm/yr for the time intervals 0–1 my (i.e. 24–23 Ma), 1–3 my (i.e. 23–21 Ma), and 3–4 my (i.e. 21–20 Ma), respectively. Numbers adjacent to the 20 km result are in millions of years and apply to all three curves.

Fig. 8. Pressure–temperature diagram showing calculated trajectories of footwall positions at differing distances from a normal fault dipping at 30° (see Harrison et al., 1995a). Melting conditions shown are: wet melting of metapelites, Le Breton & Thompson (1988); muscovite–quartz–albite vapor-absent melting, Peto (1976). Emplacement conditions of the Manaslu magmas are from Scaillet et al. (1995) and Guillot et al. (1995). Vertical component of slip for footwall positions for the various time intervals are: 10 mm/yr, 2.5 mm/yr, and 10 mm/yr for 0–1 my (i.e. 24–23 Ma), 1–3 my (i.e. 23–21 Ma), and 3–4 my (i.e. 21–20 Ma), respectively. Numbers adjacent to the 20 km result are in millions of years and apply to all three curves.
of a significantly lower geotherm (e.g. Harris & Massey, 1994) together with higher slip rates (>20 mm/yr) and total magnitude of denudation (>30 km) could produce better agreement with the known emplacement conditions of the Manaslu intrusive complex (Fig. 8). Despite our reservations regarding the viability of decompression melting, the two models make significantly different, testable predictions that should permit a clear selection between them. For example, a fundamental consequence of the decompression model is that melting occurs at progressively lower pressures and temperatures. In contrast, shear heating along a sub-horizontal décollement results in progressively higher temperature melting. Our inferences from accessory mineral saturation levels and inheritance patterns that the Bimtang phase achieved higher peak temperatures relative to those of the Larkya La phase argues against decompression melting acting alone. Finally, we wish to emphasize that the magnitude of melting from extraordinary decompression (>20–30 km) is matched by only ~40 km of sub-horizontal displacements along the basal décollement assuming modest (<50 MPa) shear stress.

CONCLUSIONS

Th–Pb monazite ages determined by the ion microprobe technique on 12 samples from the Manaslu intrusive complex are consistent with two episodes of intrusion at 22.9 ± 0.6 Ma (the Larkya La phase) and 19.3 ± 0.3 Ma (the Bimtang phase). Correlated textural and geochemical relationships suggest that the Bimtang phase represents a one-third of the present leucogranite exposure. Because the Manaslu intrusive complex crosscuts a collapse structure in the hanging wall of a detachment immediately beneath the pluton, this tectonic feature must be older than ~23 Ma. The associated symmetrical extensional deformation within the Larkya La phase must have occurred between 23.5 and 22.3 Ma. Inherited monazite is observed in 10 of the 12 samples, with Th–Pb ages interpreted to reflect contamination by both Eo-Himalayan (~45 Ma) and Pan-African (~500 Ma) components. The scale of isotopic disequilibrium preserved within restitic monazites indicates that the Larkya La phase was at magmatic temperatures for <2 my. Dating results for a single sample from the nearby Dolpo-Mugu granite yield an age of 17.6 ± 0.3 Ma, but suggest a protracted magmatic history extending over 0.5 my that cannot at this time be related to the Manaslu intrusive complex.

The episode of anatexis at ~23 Ma appears to have resulted from vapor-absent muscovite dehydration melting of the Greater Himalayan Crystalline (GHC) whereas the magmatic pulse at 19.3 Ma is interpreted to result from higher-temperature dehydration melting of the same source region that produced the earlier magma. Thermo-kinematic modeling indicates that: (1) two phases of leucogranite magmatism separated by ~4 my can be produced by shear heating along the Himalayan décollement assuming a shear stress of 45 MPa, and (2) scenarios involving only decompression to achieve the same result appear to require extreme conditions not permitted by available geological constraints.

ACKNOWLEDGEMENTS

We thank F. Oberli, M. Maier, M. Tsutsumoto, G. Gehrels, R. Parrish, and T. Ireland for sharing their unpublished U–Th–Pb results on 554 and LO68 monazites with us, and F. Ramos for providing Rb–Sr, Sm–Nd, and Pb isotopic analyses. The manuscript greatly benefited from constructive comments by reviewers Nigel Harris, Alberto Patiño Douce, and Peter Copeland. This research was sponsored by a grant (EAR-9614869) from the National Science Foundation. Support for the ion microprobe facility was obtained from NSF’s Instrumentation and Facilities Program.

REFERENCES


