

# Thermal Constraints on the Emplacement Rate of a Large Intrusive Complex: The Manaslu Leucogranite, Nepal Himalaya

C. ANNEN<sup>1\*</sup>, B. SAILLET<sup>2</sup> AND R. S. J. SPARKS<sup>3</sup>

<sup>1</sup>SECTION DES SCIENCES DE LA TERRE, UNIVERSITÉ DE GENÈVE, 13 RUE DES MARAÎCHERS, 1205 GENÈVE, SWITZERLAND

<sup>2</sup>INSTITUT DES SCIENCES DE LA TERRE D'ORLÉANS, UMR 6113 CNRS-UO, 1A RUE DE LA FÉROLLERIE, 45071 ORLÉANS CEDEX 2, FRANCE

<sup>3</sup>DEPARTMENT OF EARTH SCIENCES, UNIVERSITY OF BRISTOL, WILLS MEMORIAL BUILDING, QUEENS ROAD, BRISTOL BS8 1RJ, UK

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*The emplacement of the Manaslu leucogranite body (Nepal, Himalaya) has been modelled as the accretion of successive sills. The leucogranite is characterized by isotopic heterogeneities suggesting limited magma convection, and by a thin (<100 m) upper thermal aureole. These characteristics were used to constrain the maximum magma emplacement rate. Models were tested with sills injected regularly over the whole duration of emplacement and with two emplacement sequences separated by a repose period. Additionally, the hypothesis of a tectonic top contact, with unroofing limiting heat transfer during magma emplacement, was evaluated. In this latter case, the upper limit for the emplacement rate was estimated at 3.4 mm/year (or 1.5 Myr for 5 km of granite). Geological and thermobarometric data, however, argue against a major role of fault activity in magma cooling during the leucogranite emplacement. The best model in agreement with available geochronological data suggests an emplacement rate of 1 mm/year for a relatively shallow level of emplacement (granite top at 10 km), uninterrupted by a long repose period. The thermal aureole temperature and thickness, and the isotopic heterogeneities within the leucogranite, can be explained by the accretion of 20–60 m thick sills intruded every 20 000–60 000 years over a period of 5 Myr. Under such conditions, the thermal effects of granite intrusion on the underlying rocks appear limited and cannot be invoked as a cause for the formation of migmatites.*

KEY WORDS: granite emplacement; heat transfer modelling; High Himalayan Leucogranite; Manaslu; thermal aureole

## INTRODUCTION

The melting of continental crust can be achieved via two fundamental mechanisms: either by advective heating resulting from basalt underplating or intrusion (Huppert & Sparks, 1988; Bergantz, 1989; Raia & Spera, 1997) or via thickening of crust rich in heat-producing elements (England & Thompson, 1984), assisted or not by shear heating (Le Fort, 1975; England *et al.*, 1992; Harrison *et al.*, 1998) or isothermal decompression (Harris & Massey, 1994). The Himalaya range is a collisional belt characterized by the production of crustal melts without the involvement of basalt (Le Fort *et al.*, 1987). Thus, it is a place where models belonging to the second category can be investigated.

Study of the High and North Himalayan Leucogranites (HHL and NHL) has been pivotal for understanding the structural and thermal evolution of thickened crust in collisional settings. The HHL are generally located on top of a high-grade metamorphic unit—the Precambrian Higher Himalayan Crystallines (HHC)—which forms the backbone of the range, and below a thick Paleozoic to Mesozoic sedimentary sequence, termed the Tethyan Sedimentary Series (TSS). The HHC is truncated at its base by the Main Central Thrust (MCT), a 10 km thick shear zone that has accommodated part of the Himalayan Miocene–Pliocene shortening. The top of the HHC is defined by the South Tibetan Detachment (STD), a low-dipping extensional, ductile to brittle

\*Corresponding author. Telephone: ++41 22 379 66 23. Fax: ++41 379 32 10. E-mail: Catherine.Annen@terre.unige.ch

crustal-scale shear zone that is believed to have been episodically active since the Miocene, perhaps in response to gravitational instability of the Himalayan topographic front (Burg *et al.*, 1984; Royden & Burchfiel, 1985). The HHC is believed to have been extruded by simultaneous motion along the MCT and STD (Hodges *et al.*, 1992). The HHL may be related to the development of either the MCT or STD and their dating has been used to constrain the age of fault movement (Guillot *et al.*, 1994; Coleman, 1998; Dèzes *et al.*, 1999; Murphy & Harrison, 1999).

Geochronological data show that most of the HHL were produced in the time interval 24–19 Ma (Harrison *et al.*, 1999b). The onset of the India–Tibet collision is dated at ~55 Ma in the west of the range (Rowley, 1996; Guillot *et al.*, 2003), with a probable younging eastward because of the obliquity of convergence between India and Tibet. The time lag of 20–30 Myr between crustal thrust stacking and crustal melting is significantly shorter than results from numerical simulations of the thermal evolution of the thickened crust, which predict that melting occurs some 50 Myr after thickening (England & Thompson, 1984). Therefore, modelling studies on the thermal evolution of the Himalayan orogenic belt have looked for either additional sources of heat, or alternative mechanisms of crustal melting, to explain the time lag and its relationship to the tectonometamorphic evolution of the orogen, in particular the well-known inverted metamorphic gradient that affects the HHC (Molnar *et al.*, 1983; Pinet & Jaupart, 1987; England *et al.*, 1992; Henry *et al.*, 1997; Harrison *et al.*, 1998, 1999a). Alternative mechanisms include shear heating (England *et al.*, 1992; Harrison *et al.*, 1998), accretion of radiogenic crustal layers (Huerta *et al.*, 1999; Guillot & Allemand, 2002), thermal conductivity contrast between the crystalline basement and its sedimentary cover (Pinet & Jaupart, 1987), large-scale fluid infiltration that lowers the melting point of overthrust terranes (Le Fort, 1975), slab break-off during collision (Kohn & Parkinson, 2002) and fast decompression of hot metamorphic rocks (Harris & Massey, 1994). Although differing in many respects, most recent models propose that both compressional and extensional faults exert a central role in the orographic development of the orogen (Harris & Massey, 1994; Grujic *et al.*, 1996; Nelson *et al.*, 1996; Beaumont *et al.*, 2001; Vannay & Grasemann, 2001).

Recent studies suggest that exhumation of HHC results from the competing effects of the ductile extrusion of a crustal channel (HHC), bounded by low-angle normal (STD) and thrust (MCT) faults. Focused surface denudation (erosion) is localized at the southern edge of the Tibetan Plateau (i.e. the Himalaya). This hypothesis requires that both MCT and SDT detachments were simultaneously active. Several geophysical surveys have

indicated the presence of partial melt in the Tibetan middle crust, immediately north of the Himalayan range (Pham *et al.*, 1986; Brown *et al.*, 1996; Nelson *et al.*, 1996), which has led to the suggestion that the Himalaya range could correspond to the southward extrusion of the partially molten crust underlying southern Tibet, and that contemporary melting beneath south Tibet could be the continuation of the dynamical processes that formed the HHL during the Miocene (Wu *et al.*, 1998; Hodges *et al.*, 2001; Beaumont *et al.*, 2004). This channel flow model seeks to unify geological, geochronological and geophysical observations gathered in the Himalayan–Tibetan area (Hodges *et al.*, 2001; Beaumont *et al.*, 2004).

Interest in the tectonic exhumation concept has also been triggered by the suggestion that production of high reliefs, following the India–Tibet collision, could have affected global climate (Raymo & Ruddiman, 1992). Abrupt changes in world seawater chemistry in the Miocene (Edmond, 1992; Harris *et al.*, 1995), as well as the sedimentary record of the Bengal Fan (France-Lanord *et al.*, 1993), have been explained by Miocene land exposure and subsequent erosion of HHC and HHL rocks. Because Himalayan metamorphism and magmatism are essentially Miocene in age, their exhumation during the Miocene requires a fast mechanism of rock uplift. Tectonic unroofing of thermally weakened (i.e. partially melted) crust is commonly advocated (Hodges *et al.*, 1998; Searle & Godin, 2003). However, the mechanisms leading to crustal melting, as well as the role of partial melt in the development of the Himalayan orogen, are still controversial (Harrison *et al.*, 1999a). It is still unclear to what extent crustal melting is related to crustal faults and, if so, whether it is the cause or the consequence of crustal-scale compressional (MCT) or extensional (STD) faulting. Both the orogenic wedge (e.g. Harris & Massey, 1994) and channel flow (e.g. Beaumont *et al.*, 2004) models require that motions on both the MCT and STD were broadly contemporaneous with the formation of the HHL. Although available timing constraints are consistent with this idea, it is not the only possibility (Murphy & Harrison, 1999). Models that do not invoke normal faulting can also reproduce the principal tectono-metamorphic and geochronological constraints of the Himalaya (Harrison *et al.*, 1998).

In this study, we model the thermal evolution of individual leucogranite bodies to constrain emplacement rates of the HHL plutons. From inferred emplacement rates, the residence time of anatectic melts in their source regions can be constrained. The study also sheds light on the potential role of extensional faulting on crustal magmatism, as coeval STD slip and HHL emplacement should affect the cooling regime of the HHL. Our model only concerns the period of pluton emplacement. We combine these results with fluid dynamic considerations

to constrain the conditions under which HHL cooling could be achieved in a purely conductive regime. The geochemical characteristics of the HHL suggest emplacement with no large-scale convection within the magma bodies (Deniel *et al.*, 1987).

Modelling the thermal evolution of the HHL plutons is relevant to the following Himalayan issues.

(1) We explore heat advection within the Himalayan crust to test whether the increase in degree of migmatization observed up-section in the HHC (i.e. the so-called inverse metamorphism) could be a result of HHL intrusion, as has been proposed in Bhutan by Davidson *et al.* (1997).

(2) The HHL have been alternatively interpreted as plutons emplaced *in situ* close to their level of production (Visona & Lombardo, 2002), and as intrusive bodies in cold upper crust detached from their source, with vertical transport distances of ~8–10 km (Copeland *et al.*, 1990; Inger & Harris, 1992; Guillot *et al.*, 1995*b*; Scaillet *et al.*, 1996; Walker *et al.*, 1999). The level of pluton intrusion has a major impact on its subsequent thermal evolution (Davidson *et al.*, 1992), and our modelling results aim to distinguish between these hypotheses.

(3) Our model provides a framework for interpreting geochronological data for the HHL, which have been difficult to date accurately. Progress has been made with ion probe studies (Harrison *et al.*, 1999*b*). However, the accessory minerals used for dating can be inherited (Copeland *et al.*, 1988), so even these methods are not straightforward (Harrison *et al.*, 1999*b*).

### General modelling approach

Our thermal models place constraints on the emplacement of granite intrusions. We apply the model results to the Manaslu leucogranite, the best known of the HHL (Le Fort *et al.*, 1987; France-Lanord *et al.*, 1988; Guillot *et al.*, 1995*a*). As a result of the highly dissected nature of the Himalaya range, field observations provide a continuous record of the nature of the host rocks of the HHL, both above and beneath the intrusions, down to levels where crustal melting occurred. These data, combined with the wealth of geological, petrological and geochemical data summarized below, place tight constraints on key parameters that affect the thermal evolution of the cooling plutons, including the  $P$ - $T$ - $H_2O$  conditions of the magma and its surroundings during injection and the country rock properties.

Geological and geophysical data indicate that many granitic plutons, including the HHL, are tabular, low-aspect ratio bodies fed by thin vertical conduits (Le Fort, 1981; Scaillet *et al.*, 1995*a*; McCaffrey & Petford, 1997; Cruden, 1998; Searle, 1999; Petford *et al.*, 2000; Haederle & Atherton, 2002). Field observations

combined with detailed isotopic studies suggest that the HHL are an amalgamation of numerous accreted intrusions (Deniel *et al.*, 1987), as has been inferred for many other plutons (Evans *et al.*, 1993; Wiebe & Collins, 1998; Glazner *et al.*, 2004).

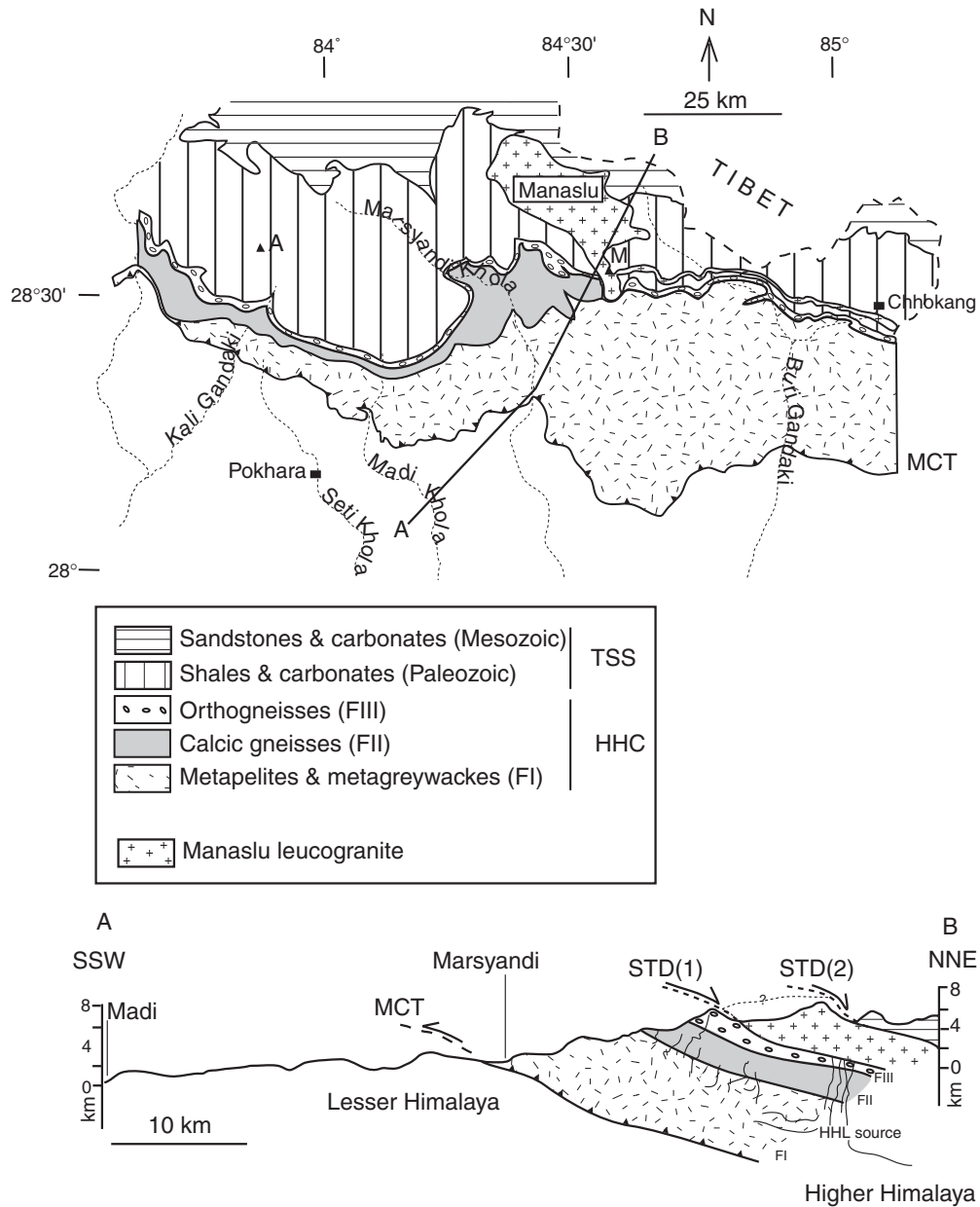
The basic modelling concept is the successive emplacement of numerous thin sills, which grow into a large granite intrusion. The intrusions are sufficiently wide in comparison to their thickness that a 1-D model can be justified. We have modelled the heat transfer between the growing intrusion and its country rocks. Our objective is to constrain the thermal regime that leads to the emplacement of a lensoid igneous body several kilometres wide in the upper crust that, once solidified, is characterized by two critical features: (1) a strong Sr-isotopic heterogeneity that precludes wholesale convection causing magma mixing and homogenization; (2) a thin, no more than 100 m thick, top contact aureole.

## THE MANASLU LEUCOGRANITE AND ITS HOST ROCKS

### Bottom and top host rocks

The Manaslu pluton is exposed in central Nepal, covers an area of 400 km<sup>2</sup> and has an estimated volume of 3000 km<sup>3</sup> (Le Fort *et al.*, 1987). It is a laccolithic-shaped body with a thickness of up to 5–6 km and a lateral extent of 30–40 km (Fig. 1).

The Manaslu pluton is located on top of the HHC, which is here divided into three sub-units of variable thickness: FI, FII and FIII (Colchen *et al.*, 1986). The basal unit, FI (thickness 2–10 km), is made of alternating metapelitic and metagreywacke gneisses metamorphosed within kyanite (bottom) to sillimanite (top) grades. FI is truncated at its base by the Main Central Thrust (Fig. 1). Migmatization is prominent in FI and increases up-section (Barbey *et al.*, 1996; Coleman, 1998). U–Pb geochronology on monazite and zircon gives migmatization ages of 18–22.5 Ma for FI migmatites (Hodges *et al.*, 1996; Coleman, 1998). The FI unit is overlain by the FII unit of calcic gneisses; this is up to 3.5 km thick but displays strong lateral variations in thickness (Fig. 1). The FIII unit is made of variably migmatized porphyritic orthogneisses, 1 km thick, that are interpreted as the metamorphosed equivalent of Cambro-Ordovician metaluminous to weakly peraluminous granites occurring in the Lesser Himalaya (Le Fort, 1986). U–Pb dating of monazite suggests that the age of migmatization of the FIII orthogneisses is older than the FI migmatites, at around 36 Ma, or early Oligocene (Hodges *et al.*, 1996; Coleman, 1998). Thermobarometric studies have shown that peak metamorphic pressures at the base of the HHC (close to the MCT) are in the range 8–10 kbar, and decrease regularly up-section down to about 5 kbar at



**Fig. 1.** Generalized geological map of the Manaslu area in central Nepal, after Colchen *et al.* (1986), and NE-SW cross-section of the High Himalayan range showing the main geological units discussed in the text (adapted from Colchen *et al.*, 1986). Only the main lithological units are shown, together with the position of the Main Central Thrust (MCT). On the map, other tectonic contacts (STD) are omitted for clarity. On the cross-section, two positions for the South Tibetan Detachment (STD) are shown: STD (1) is according to Guillot *et al.* (1993) (i.e. below the pluton); STD(2) is according to Searle & Godin (2003) (above the pluton). M, Manaslu peak; A, Annapurna peak.

the top of the HHC (Pêcher, 1989; Coleman & Hodges, 1998; Guillot, 1999; Macfarlane, 1999). Peak temperature varies between 580 and 700°C within the HHC in this area (Pêcher, 1989; Coleman & Hodges, 1998; Guillot, 1999; Macfarlane, 1999). Temperatures up to 800°C have been estimated along other transects, such as in the Langtang section, some 75 km west of Manaslu (Inger & Harris, 1992), whereas other studies suggest a

maximum temperature of 750°C even in rocks that are partially melted (Ayres & Harris, 1997; Harris *et al.*, 2004). However, these thermobarometric constraints provide only minimum temperatures, as the metamorphic grade increases northward. Thus, peak temperatures reached in the source region of the HHL, which is presumably located beneath the HHL or further north, must be somewhat higher than those recorded in

more distal (i.e. southward) sections of the HHC thrust sheet.

Overlying the HHC is the Tethyan (or Tibetan) sedimentary series (TSS), which forms a near-continuous, weakly to non-metamorphosed, Paleozoic to Mesozoic (up to the Cretaceous) sequence, in which MgO-poor limestones dominate over sandstones and shales (Colchen *et al.*, 1986). The top of the granite is in contact with progressively higher stratigraphic units westward, from lower Paleozoic levels up to Upper Triassic shales in the middle part of the lens (Fig. 1). The restored thickness of the sedimentary sequence in the area of the Manaslu pluton reaches 11 km (Colchen *et al.*, 1986). However, Tertiary folding or nappe emplacement prior to granite intrusion is believed to have significantly increased the overburden depth at the time of granite emplacement (Guillot *et al.*, 1995a).

### Top contact

The exact location of the STD is contentious in the Manaslu area (Coleman, 1996; Searle & Godin, 2003). Some workers (Caby *et al.*, 1983; Pêcher, 1989, 1991; Coleman, 1998) consider that the STD lies structurally below the Manaslu pluton, which has a top intrusive contact, whereas others (Searle & Godin, 2003) consider that the STD lies above the granite, as most often observed, and view the top contact of the granite as a tectonic one. For Guillot and collaborators (Guillot, 1993; Guillot *et al.*, 1993, 1995a), the normal shearing observed at the Manaslu top contact is local and related to the ballooning effect of the intrusion process.

Given the importance of this contact for the purpose of thermal modelling, we review the observations pertinent to these alternative interpretations. Le Fort (1981), Colchen *et al.* (1986) and Guillot *et al.* (1993, 1995a) reported that the top of the pluton develops a thin metamorphic aureole whose thickness depends on the lithology of the local host rocks. At limestone contacts, the aureole only extends 2–3 m and is marked by clusters of radiating wollastonite (Colchen *et al.*, 1986). At sandstone or shale contacts, the aureole is 30–50 m thick, and the sedimentary rocks are transformed into muscovite–garnet quartzite and biotite–muscovite–garnet–staurolite micaschists, respectively (Colchen *et al.*, 1986). Thermobarometric studies of the contact metamorphic conditions yielded  $P$ – $T$  conditions of 3–4 kbar/550 ± 40°C for the top of the pluton and 5–6 kbar/580 ± 40°C for the base, in agreement with  $P$ – $T$  conditions estimated for the top part of the HHC (Guillot, 1999; Macfarlane, 1999). The difference in barometric estimates between the top and the bottom of the pluton is in good agreement with the lithostatic pressure difference of ~2 kbar, corresponding to the structural thickness of the pluton of 5–6 km (Guillot *et al.*, 1995a). In addition, Ar–Ar ages of minerals in the top aureole (Guillot *et al.*, 1994)

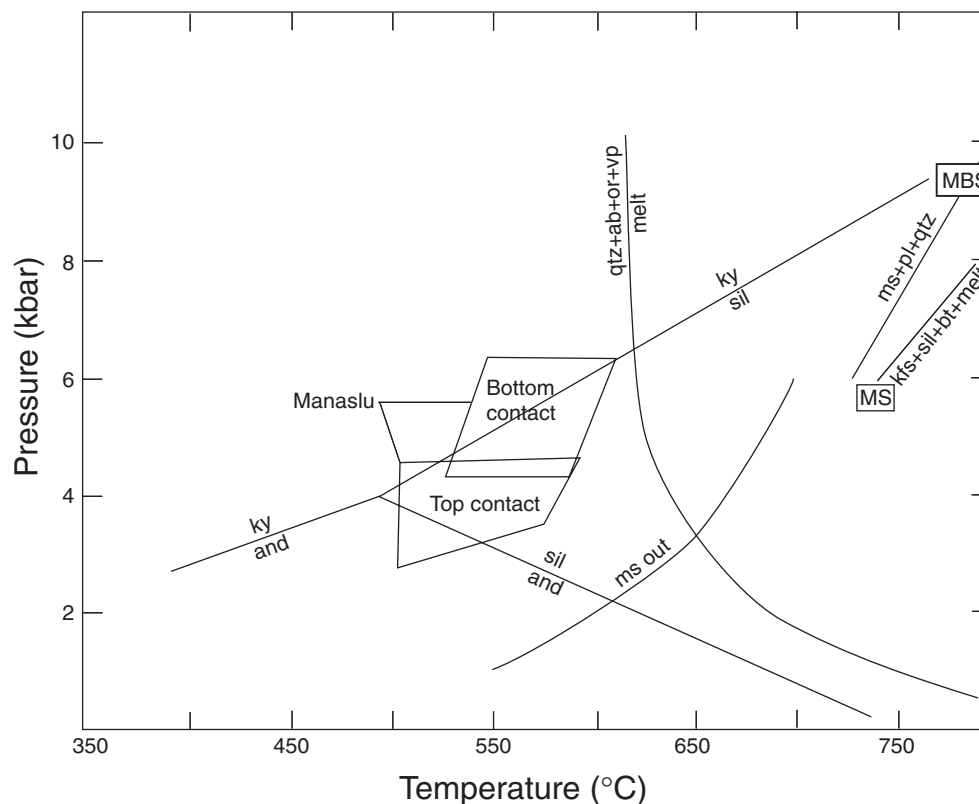
are similar to the ages of the nearby leucogranite (Copeland *et al.*, 1990). However, Searle & Godin (2003), working on the northwestern part of the Manaslu area, observed a 400 m thick shear zone of high-strain mylonites structurally above the pluton, with the upper part displaced to the north. This shear zone is reported to wrap around the northern contact of the pluton, although Searle & Godin (2003) explored only three valley transects west of the pluton, and extrapolated the shear zone eastward to the area in which Guillot *et al.* (1995a) worked. We have no reasons to dismiss the general structural interpretation of Searle & Godin (2003). However, the field and petrological observations of the topmost contact by Guillot *et al.* (1993, 1995a) cannot be discounted either: in particular, the very thin aureole described in the limestone layers, the presence of host rock xenoliths in the granite (see photo 54 of Colchen *et al.*, 1986), and the age correspondence between the leucogranite and its host rock all suggest that in this area, the intrusive relationship of the granite has been locally preserved and that the  $P$ – $T$  conditions derived by Guillot *et al.* (1995a, 1995b) reflect the conditions of aureole metamorphism. In the eastern area, where the contact aureole has been studied, the STD does not occur at the granite contact, suggesting that the tectonic boundary is located further north than shown in the map of Searle & Godin (2003). Geochronological data constrain fault movement along the STD in the interval 14–17 Ma (Harrison *et al.*, 1999a), making the STD younger than the Manaslu crystallization age of 19–24 Ma (Coleman, 1998; Harrison *et al.*, 1999b). In addition, although detailed structural studies of large HHL are scarce, those available, including at Manaslu, have shown that the large plutons are, in general, characterized by the absence of strong penetrative deformation, which led early workers to infer the late kinematic character of the pluton intrusion with respect to the regional deformation (Le Fort, 1981). Magmatic lineations in the HHL plutons trend E–W (Guillot *et al.*, 1993; Scaillet *et al.*, 1995a), and not N–S, as would be expected if intrusions occurred during a regional phase of N–S extension.

In this study, we have considered two possible geometric relationships between the Manaslu granite and its top contact, either intrusive or faulted, to evaluate whether different models relating STD motion to HHL emplacement are supported by thermal constraints.

### Conditions during magma intrusion

The HHL are characterized by the occurrence of magmatic muscovite, in addition to tourmaline and biotite (e.g. Le Fort *et al.*, 1987). Experimental work by Bénard *et al.* (1985) on the Manaslu granite shows that at 3 kbar, muscovite is not stable in fluorine-poor HHL compositions under magmatic conditions. Scaillet *et al.* (1995b) defined a stability field for muscovite at 4 kbar.





**Fig. 2.**  $P$ - $T$  diagram showing the locations of the main dehydration melting reactions (MS, muscovite schist and MBS, muscovite–biotite schist involving muscovite, obtained for two Himalayan metapelites by Patiño Douce & Harris (1998), together with the  $P$ - $T$  conditions inferred for the top and bottom contact aureoles of Manaslu (Guillot *et al.*, 1995*a*). Melting conditions of HHL are inferred to occur in the pressure range 6–10 kbar, whereas emplacement of HHL takes place at 3–4 kbar. Also shown are the haplogranite wet solidus (qtz + ab + or + 3vp) after Luth (1976) and the muscovite out (ms-out) curve of Chatterjee & Johannes (1974). Note that both curves intersect at around 3 kbar, in agreement with phase equilibria constraints for the HHL (Benard *et al.*, 1985; Scaillet *et al.*, 1995*b*).

The stability curve of pure OH–muscovite intersects the wet haplogranite solidus in the pressure range 3–4 kbar (Fig. 2). Therefore, if devoid of fluorine, crystallization of muscovite in HHL constrains a minimum emplacement pressure of  $\sim 3.5$  kbar. However, HHL do have some fluorine, which will extend the magmatic stability field of muscovite to lower pressures (Pichavant *et al.*, 1988). HHL muscovites have 0.5–0.9 wt % F (Scaillet, unpublished data) which would allow crystallization of muscovite down to 2.8 kbar (Pichavant *et al.*, 1988), matching the pressure constraint derived by Guillot *et al.* (1995*a*, 1995*b*) for the top aureole. For  $T$  and melt  $H_2O$  content, petrological and experimental data show that the magma was near the liquidus on emplacement, with intrusion temperatures of 750–800°C and 5–7 wt % dissolved  $H_2O$  (Montel, 1993; Scaillet *et al.*, 1995*b*, 1996).

The High Himalayan Crystallines, in particular the metapelites and metagreywacke of FI, are believed to be the protholiths of the HHL, based on a variety of geological and geochemical arguments (Deniel *et al.*, 1987; Le Fort *et al.*, 1987; France-Lanord *et al.*, 1988; Harris & Inger, 1992; Inger & Harris, 1993; Guillot &

Lefort, 1995; Harris *et al.*, 1995; Searle *et al.*, 1997). Harris & Inger (1992) proposed, based on trace element modelling, that the HHL were produced by dehydration melting of muscovite-bearing metasediments, with melt fractions <10 wt %. This hypothesis is supported by the experimental work of Patiño Douce & Harris (1998), who showed that dehydration melting experiments on Himalayan metapelites (muscovite + plagioclase + quartz assemblage) in the pressure range 6–10 kbar yield melts that are virtually indistinguishable from the HHL (Fig. 2). We note that, in addition, the temperature and water contents of the experimental melts are the same as those inferred from near-liquidus phase equilibria of HHL (Scaillet *et al.*, 1995*b*). These results support the view that the magma migrated upward with limited, if any, chemical or thermal modifications. There is field evidence that the HHL magmas moved upward through narrow dykes (Le Fort, 1981; Inger & Harris, 1992; Scaillet *et al.*, 1996; Walker *et al.*, 1999), some of which connect to the base of the pluton, such as in the Garwhal Himalaya (Searle *et al.*, 1993; Scaillet *et al.*, 1995*a*).

### Age constraints

Although homogeneous in terms of their major element compositions (Le Fort *et al.*, 1987), the HHL have Rb–Sr and  $^{87}\text{Sr}$ – $^{86}\text{Sr}$  heterogeneities that have prevented accurate dating using the Rb–Sr system, and which are believed to reflect heterogeneity in the source region (Vidal *et al.*, 1982; Deniel *et al.*, 1987; Inger & Harris, 1993; Guillot & Le Fort, 1995). Rb–Sr heterogeneities indicate that the granites were not well mixed internally and that conditions for large-scale magma convection (which would homogenize the magma) were not reached (Deniel *et al.*, 1987). Using U–Pb monazite ages, Coleman (1998) proposed that magmatic activity in the area of Manaslu occurred in two main pulses at 18 and 22 Ma. Harrison *et al.* (1999b) determined 158 Th–Pb ages for monazite from 12 samples of the Manaslu granite. Based on the peak of the age distribution obtained for each sample, Harrison *et al.* (1999b) defined crystallization ages ranging from  $19.0 \pm 0.4$  to  $24.0 \pm 0.8$  Ma. The histogram of all the Th–Pb monazite ages shows two distinct peaks, one around  $19.3 \pm 0.3$  Ma and the other centred at  $22.9 \pm 0.6$  Ma. The oldest ages correspond to samples from near the top of the body, whereas the youngest are from samples at the bottom. These results led Coleman (1998) and Harrison *et al.* (1999b) to conclude that the Manaslu granite was emplaced in two magma pulses separated in time by 4 Myr. The lack of intermediate ages between 19.6 and 22.0 Ma might, however, reflect a lack of sampling in the middle part of the intrusion. The age distribution shows a tail towards older ages which falls within the interval 24–37 Ma, and an even older interval, between 100 and 600 Ma. The older interval can be ascribed to inheritance from a 500–600 Ma crustal protolith, probably generated during the Cambro-Ordovician thermal event that produced the Lesser Himalayan granites and their metamorphosed equivalents in the HHC. Harrison *et al.* (1999b) interpreted the 24–37 Ma population as reflecting Eo-Himalayan metamorphism, i.e. the metamorphic phase that affected the HHC during the early stages of crustal stacking. However, this population might also record incipient stages of magma intrusion. We conclude that the time interval of laccolith growth of 5 Myr should be viewed as a minimum value.

### THE NUMERICAL MODEL

We simulated the growth of a laccolith-shaped granite body by the successive emplacement of discrete sills at magmatic temperature. In the Manaslu pluton, the youngest ages are stratigraphically located below the oldest (Harrison *et al.*, 1999b) and so, in our model, each successive sill is emplaced below the previous one. The room for the sills is made by uplifting the overlying sediments or by moving the sequence below the sills

downward. This latter case is an approximation for isostatic equilibration. In terms of computation, these two kinds of room accommodation correspond to two different bottom boundary conditions. For a room accommodation by isostatic equilibration, the bottom temperature is constant and at a fixed depth (Fig. 3b), whereas for a room accommodation by overlying sediments uplifting, the bottom temperature is also fixed but its depth is moving downwards relative to the surface level (Fig. 3c). We verified that the results were not sensitive to our assumptions about room accommodation.

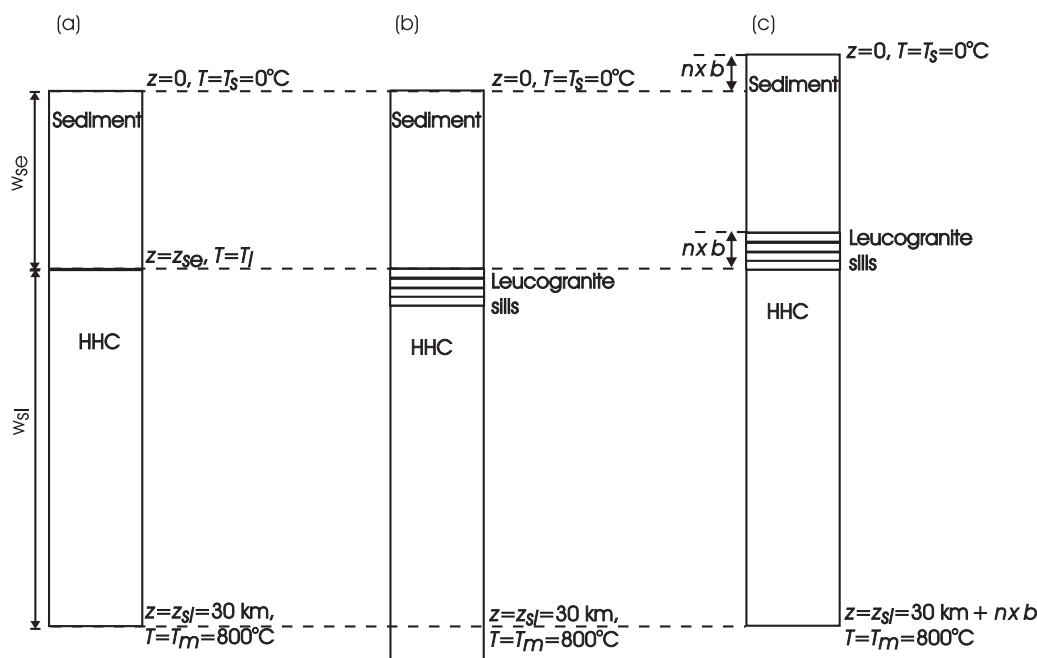
Two main models were tested. In the first one, magma sills are regularly injected throughout the entire magmatic event leading to the emplacement of 5 km of granite. In the second model, the same thickness of magma is emplaced in two pulses separated by a repose period of 4 Myr, as proposed by Harrison *et al.* (1999b). The sills are accreted in the same way as in the first model until the body is 2.5 km thick, then magma injection stops. After 4 Myr of no magmatism, intrusions resume and the remaining 2.5 km are emplaced at the same rate as in the first pulse. It should be noted that the thickness of the Manaslu granite has been estimated to reach 8 km in some areas (Guillot *et al.*, 1995a), so that the 5 km total thickness is conservative.

The media, i.e. the granite and the country rock layers, are discretized and the temperature evolution is computed using the finite difference expression of the equation of heat balance:

$$\rho C_p \frac{\delta T}{\delta t} = k \nabla T + \rho L + A \quad (1)$$

where  $\rho$  is density,  $C_p$  is specific heat capacity,  $T$  is temperature,  $t$  is time,  $k$  is thermal conductivity,  $L$  is latent heat of fusion or crystallization, and  $A$  is radioactive heat production. Values of these parameters used in the model are given in Table 1.

Models were tested for sill thicknesses from 5 to 1000 m. Preliminary results showed that only thin sills produced results consistent with the geological observations (small aureole size). Additionally, Deniel *et al.* (1987) found that the scale length of the Rb–Sr heterogeneities was 100 m. Thus, most simulations were run with sills of thickness 50 m. The convergence of the results was tested for different node spacing. For models with sills 50 m thick or more, the node spacing was 25 m. For thinner sills, the node spacing was equal to the thickness of the sill. The top of the intrusion was emplaced at a pressure of 3–4 kbar (Guillot *et al.*, 1995a). We tested our model with granite top depth, i.e. depths of the first intruded sill, at 10, 12.5 and 15 km. We assumed that the granite magma temperature at 30 km depth was 800°C and that the adiabatic drop between 30 km and the emplacement level is about 20°C, resulting in a modelled leucogranite emplacement temperature of 780°C. We also ran some



**Fig. 3.** One-dimensional model of emplacement of a large igneous body by discrete sill accretion. (a) Initial conditions; (b) and (c) after the intrusion of  $n$  sills. In (b) the volume of the sills is accommodated by 'isostasy' (sagging) and the bottom boundary  $T = 800^\circ\text{C}$  is at a fixed depth ( $z = 30\text{ km}$ ); in (c) the volume of the sills is accommodated by roof uplift: the bottom boundary  $T = 800^\circ\text{C}$  is at  $30\text{ km} + nb$ , with  $b$  the sill thickness. For abbreviations see Table 1.

tests with an emplacement temperature of  $750^\circ\text{C}$  to check the influence of lower temperature on the results. We did not consider temperatures lower than  $750^\circ\text{C}$  because melt fractions in partially melted Himalayan metapelites are below 5 wt % in the temperature range  $720\text{--}750^\circ\text{C}$  (Patiño Douce & Harris, 1998) and it is unlikely that such partial melts will easily move out of their source but, instead, will remain trapped within it.

### Conductivities

Heat transfer in a material is governed by the material heat diffusivity,  $\kappa$ :

$$\kappa = \frac{k}{\rho C_p}. \quad (2)$$

Density and specific heat capacity change only slightly with rock composition. Conductivity strongly depends on rock composition and on pore water or air content (Robertson, 1988). Rock conductivity is especially sensitive to quartz content, because of its high conductivity ( $7.69\text{ W/m/K}$ ). Depending on quartz content, sedimentary and felsic igneous rocks have conductivities ranging from  $1.5$  to  $>6\text{ W/m/K}$  (Robertson, 1988). If the mode of the rock is known, its conductivity can be estimated as follows (Robertson, 1988):

$$k = n_1 k_1 + n_2 k_2 + n_3 k_3 + \dots \quad (3)$$

where  $n_1, n_2, n_3$  are fractional volumes of mineral phases 1, 2, 3 and  $k_1, k_2, k_3$  are conductivities of mineral 1, 2, 3.

Conductivities decrease with increasing temperature (Robertson, 1988; Chapman & Furlong, 1992; Clauser & Huenges, 1995), but this decrease is poorly constrained and might be balanced by an increase in radiative heat transfer (Jaupart & Provost, 1985).

The average HHL leucogranite quartz content is 32% (Le Fort *et al.*, 1987). At the pressure of emplacement (3–4 kbar), porosity is assumed to be negligible. We estimate a granite conductivity of  $3.2\text{ W/m/K}$  at room temperature, in agreement with data for other gneisses and granites (Roy *et al.*, 1968; Wenk & Wenk, 1969; Jaupart & Provost, 1985; Robertson, 1988). The quartz content of Himalayan metasedimentary rocks varies from  $<20\%$  to  $>70\%$  (France-Lanord, 1987), corresponding to conductivities in the range of  $2\text{--}6\text{ W/m/K}$  (Robertson, 1988). The TSS are characterized by an alternation of limestones, sandstones and shales (Colchen *et al.*, 1986). Sandstones should have the highest conductivity, because of their high quartz content, followed by limestones and then shales. Conductivities of representative Himalayan metapelites of the HHC have room temperature conductivities in the range  $3.2\text{--}5.3\text{ W/m/K}$  with an average of  $3.7 \pm 0.8\text{ W/m/K}$  (Table 2). The conductivity of calcite is lower than that of quartz ( $3.57\text{ W/m/K}$  at room  $T$ ) and is taken as the value for limestone. At 10 km depth, the conductivity of shale is estimated to be around  $1.7\text{ W/m/K}$  (Guéguen & Palciauskas, 1992). Forty-seven per cent of the TSS is made of limestones, 23% of sandstones and the remaining 30% of shales



Table 1: Model input parameters

Fixed parameters		Values	Units
$\rho_{se}$	Sediment density	2700	kg/m <sup>3</sup>
$\rho_g$	Granite density	2300	kg/m <sup>3</sup>
$\rho_{sl}$	HHC density	2700	kg/m <sup>3</sup>
$C_{pse}$	Sediment-specific heat capacity	1000	J/kg/K
$C_{pg}$	Granite-specific heat capacity	1600	J/kg/K
$C_{psl}$	HHC-specific heat capacity	1000	J/kg/K
$k_g$	Granite conductivity	3	W/m/K
$A_{se}$	Sediment radioactive heat production	$1.4 \times 10^{-6}$	W/m <sup>3</sup>
$A_g$	Granite radioactive heat production	$5.3 \times 10^{-6}$	W/m <sup>3</sup>
$A_{sl}$	HHC radioactive heat production	$2.7 \times 10^{-6}$	W/m <sup>3</sup>
$\alpha_v$	Magma volumetric coefficient of thermal expansion	$2 \times 10^{-5}$	/K
wt % H <sub>2</sub> O	Melt water content	5	%
Tested parameters		Ranges	Units
$k_{se}$	Sediment conductivities	2–5	W/m/K
$k_{sl}$	HHC conductivities	2–3	W/m/K
$T_i$	Initial temperatures at emplacement level	250–350	°C
$b$	Sill thickness	20–100	m
$Q$	Magma emplacement rates	0.1–4.0	mm/year
$w_{se}$	Sediment thickness = granite roof depth	10–15	km

(Colchen *et al.*, 1986), suggesting an average conductivity of 3.6 W/m/K. Limestones are dominant at the base of the pile (Paleozoic), shales are more abundant upsequence, and sandstones are more or less evenly distributed throughout (Colchen *et al.*, 1986). The TSS is thus more conductive at its base than at its top. Local conductivity variations are expected at the contact with the granite, depending on the local lithology. Although realistic values for TSS conductivities are in the range of 2.5–4 W/m/K, we have tested thermal conductivities of TSS spanning a large range of 2–5 W/m/K in order to constrain the role of this parameter better.

Hydrothermal convection can increase heat transfer and the apparent conductivity. A minimum permeability,  $K$ , is required for the onset of convection. According to the theory of fluid convection in a porous medium (Turcotte & Schubert, 1982),

$$K = \frac{Ra\mu_f k_{se}}{\alpha_f g \rho_f^2 c_{pf} w_{se} \Delta T} \quad (4)$$

where  $\mu_f$ ,  $\rho_f$ ,  $\alpha_f$  and  $c_{pf}$  are the fluid viscosity, density, volumetric coefficient of thermal expansion and specific

heat capacity;  $k_{se}$  and  $w_{se}$  are the sediment conductivity and thickness. Assuming conservative values of  $Ra$  (Rayleigh number) =  $4\pi^2$ ,  $\mu_f = 1.33 \times 10^{-4}$  Pa s,  $k_{se} = 2$  W/m/K,  $\alpha_f = 10^{-3}$ /K,  $\rho_f = 1000$  kg/m<sup>3</sup>,  $c_{pf} = 4200$  J/kg,  $w_{se} = 10$  000 m and  $\Delta T = 500^\circ\text{C}$ , we obtain a minimum value for  $\log K$  of  $-16.3$ . The log of rock permeability at 10 km depth can be estimated from experimental data (Shmonov *et al.*, 2003) at  $-18.1$  or from geothermal data at  $-17.2$  (Ingebritsen & Manning, 1999). We conclude that permeabilities will be too low for hydrothermal convection to occur. The absence of mineralization and veining close to the HHL plutons also suggests that there was no pervasive hydrothermal system surrounding the granite.

### Boundary and initial conditions

The top boundary condition at the Earth's surface is a fixed temperature of  $0^\circ\text{C}$ . The bottom boundary condition is a magmatic temperature of  $800^\circ\text{C}$ , located at a depth of 30 km (Fig. 3). The initial temperature in the crust is determined by the geothermal gradient. A geothermal gradient at equilibrium can be calculated, where surface heat flux balances the heat flux from the base of the HHC and the internal heat production. The equations describing this steady-state geothermal gradient are given in the Appendix. The equilibrium geothermal gradient and the consecutive crust total heat content are controlled by the HHC and sediment conductivities (Fig. 4). Because, in an active orogenic belt, the geothermal gradient is probably not at equilibrium, we tested the model with initial temperature at emplacement level  $T_i$  of 250, 300 and  $350^\circ\text{C}$ . In this case, the initial geothermal gradient is linear between the magmatic temperature at depth ( $T_m = 800^\circ\text{C}$  at 30 km) and the temperature,  $T_b$ , at the granite emplacement level, and between  $T_i$  and the surface temperature ( $0^\circ\text{C}$ ). With a linear geothermal gradient, the temperature of the crust evolves with time and tends toward the equilibrium geothermal profile, which is controlled by rock conductivities. As discussed later, this drift toward the equilibrium profile affects the results. The results presented below show that the thermal evolution during granite emplacement is governed by the equilibrium temperature at the locus of intrusion. Most results are in fact weakly dependent on the initial temperature. This, in turn, shows that they are not strongly dependent on the extent of thermal disequilibrium of the geotherm because of orogenic process.

### Latent heat of crystallization and metamorphism

The latent heat released during crystallization is derived from phase equilibrium experiments performed on HHL at 4 kbar (Scaillet *et al.*, 1995b). Closed-system

Table 2: HHC metasediment and Manaslu leucogranite modal mineralogy and conductivities

	$k_i$ (W/m/K)	Density (g/cm <sup>3</sup> )	NA410 vol %	NA411 vol %	MS vol %	MBS vol %	HS-1 vol %	HS-2 vol %	Manaslu wt %
Quartz	7.69	2.65	63	36	43	38	35	40	32
Mica	0.4	3	11	40	24	42	30	35	10
Plagioclase 2	2.7	25	21	28	11	15	30	37	
KFd	2.4	2.55	—	—	—	—	—	—	21
$k$ (W/m/K)			5.3	3.2	3.9	3.2	3.0	3.7	3.2

All conductivities are calculated using equation (3).  $k_i$  is the mineral conductivity. Metasediment compositions are from: NA410, NA411: Colchen *et al.* (1986); MS, MBS: Patiño Douce & Harris (1998); HS-1, HS-2: (Harris & Inger, 1992); the Manaslu leucogranite composition is from Guillot *et al.* (1995a). The modal compositions of NA410 and NA411 are calculated by mass balance using the bulk-rock compositions listed by Colchen *et al.* (1986), and the mineral compositions of Patiño Douce & Harris (1998).

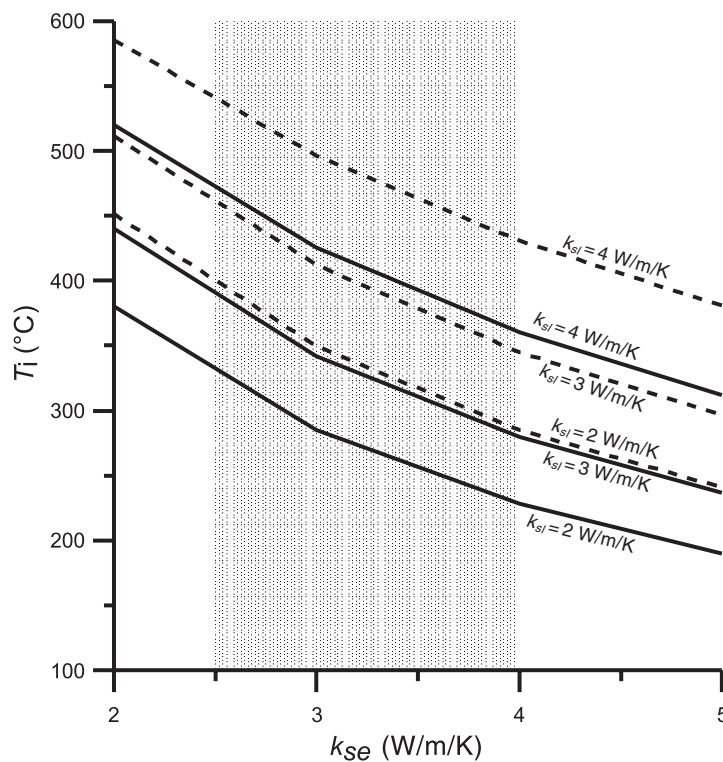


Fig. 4. Equilibrium temperature  $T_i$  at the HHC–sediment interface for different sediment conductivities  $k_{se}$  and HHC conductivities  $k_{sl}$ .  $T_i$  is calculated with the equation in the Appendix for a steady-state geothermal gradient. The grey shaded area indicates the range of realistic values for the sediment conductivities. The continuous and dashed curves are for HHC–sediment interfaces at 10 and 12.5 km depth respectively.

crystallization is assumed and the cumulated latent heat is calculated by estimating quartz, plagioclase and alkali feldspar proportions as a function of temperature and H<sub>2</sub>O in the melt. The role of accessory phases such as micas and tourmaline, which amount to less than 5% of the minerals, is neglected.

Because the HHL are highly evolved magmas, it follows that their curves of crystallization and, thus, of latent heat release are highly non-linear with temperature

(Scaillet *et al.*, 1997). Based on fits to experimental data, the expressions for HHL latent heat used in the modelling are

$$L(T) = \begin{cases} 4.9204 \times 10^5 - 463.72(T + dT) & : 928 \text{ K} < T + dT < 1061 \text{ K} \\ 2.5636 \times 10^7 - 2.7571 \times 10^4(T + dT) & : 900 \text{ K} < T + dT < 928 \text{ K}. \end{cases} \quad (5)$$

where  $L(T)$  is cumulated latent heat in J/kg;  $T$  is in K. The term  $dT$  is a correction for pressures different from 4 kbar and is equal to  $-15(P - 4)$  with  $P$  in kbar. The latent heat released by the magma cooling from temperature  $T_1$  to  $T_2$  is  $L(T_2) - L(T_1)$ . As  $T_2$  depends on the latent heat released, the solution to equation (1) is found numerically by iteration.

To evaluate the effect of latent heat of metamorphic reactions on the aureole size and temperature, some runs have been performed by including the latent heat absorbed by muscovite and biotite dehydration between 400 and 600°C using a latent heat of 600 J/kg/K (Furlong *et al.*, 1991).

### Comparison with nature: convection and thermal aureole

The constraints imposed by observations on the HHL are the absence of large-scale convection within the magma body and a thermal aureole less than 100 m thick. We carried out a parametric study to determine the conditions that are consistent with these characteristics.

Convection within an individual sill will happen on short time-scale before it crystallizes. We considered the convection extending to more than two sill thicknesses. We assume that the conditions for convection are reached in the leucogranites when the Rayleigh number,  $Ra$ , exceeds 3000 (Sparks *et al.*, 1984):

$$Ra = \frac{\rho g \alpha_v \Delta T w_m^3}{\mu \kappa} \quad (6)$$

where  $\rho$  is density,  $\alpha_v$  is the volumetric coefficient of thermal expansion,  $\Delta T$  is temperature difference between the top and bottom of the fluid layer,  $w_m$  is the thickness of the magmatic layer, and  $\mu$  is magma viscosity. The melt viscosity of HHL has been experimentally determined between 800 and 1100°C, between 3 and 8 kbar, for H<sub>2</sub>O contents between 3.98 and 6.66 wt % (Scaillet *et al.*, 1996):

$$\log(\mu_0) = \frac{16280}{T} - 7.5461 + \left(0.59784 - \frac{1235.4}{T}\right) X_{H_2O} \quad (7)$$

where  $\mu_0$  is melt viscosity in Pa s,  $T$  is in K and  $X_{H_2O}$  is the H<sub>2</sub>O content expressed as a weight percentage of the melt. In our model, the H<sub>2</sub>O content is 5 wt %.

Below the liquidus, the presence of crystals modifies the magma bulk viscosity (Roscoe, 1952; Lejeune & Richet, 1995):

$$\mu = \mu_0 \left(1 - \frac{\Phi}{\Phi_m}\right)^{-n} \quad (8)$$

where  $\Phi$  is crystal fraction,  $\Phi_m$  is the critical crystal fraction beyond which flow is prevented, and  $n$  is a constant. The form of equation (8) reflects the large

increases in viscosity that occur when crystallization is sufficient for the crystals to develop a touching framework in the transition between a crystal-rich magma and a partially molten rock. Based on experimental and empirical evidence (Marsh, 1981; Lejeune & Richet, 1995), we take  $\Phi_m = 0.6$  and  $n = 2.5$  as typical values. The viscosity becomes infinite as  $\Phi$  tends to 0.6. The model results are not sensitive to the shape of the viscosity curve because, in the case of the HHL, the curve of crystallization is close to the eutectic and  $\Phi$  increases from 0 to 0.75 over a temperature range of about 20°C. The crystal fraction at any given temperature between the solidus and the liquidus is estimated from the experimental results of Scaillet *et al.* (1995b) and has been given by Scaillet *et al.* (1997).

The modelling is based on the assumption that large-scale convection within the granite would have homogenized Rb–Sr. This assumption is supported by the results of Jellinek *et al.* (1999), who showed that convection efficiently mixes fluids with low viscosity contrasts, even at low Reynolds number. Moreover, as discussed above, the magma has an eutectic behaviour. As a consequence, rapid changes of viscosity occur over a very narrow temperature window and the viscosities become relatively low just above the eutectic temperature; thus, the sluggish convection regime is expected to be confined to very restricted conditions.

Application of equation (6) using a characteristic HHL magma viscosity of 10<sup>5</sup> Pa s and a  $\Delta T$  of 2°C, using the input parameters listed in Table 1, shows that  $Ra$  exceeds 3000 for any sill thicker than 20 m. For instance, for a sill 100 m thick,  $Ra = 3.7 \times 10^5$ , which indicates that convection can be attained in relatively thin HHL sills, even when the thermal gradient across the magma layer is very small. In this paper, we distinguish between small-scale convection in an individual sill and large-scale convection on a scale of more than two sill thicknesses. In all our models, the time-scale for a thin sill to come to thermal equilibration with its surroundings is much shorter than the repose period between intrusions. Thus, it is possible for individual sills to convect during their cooling, but large-scale convection does not develop. Bergantz and Dawes (1994) argued that with a realistic set of parameters and assumptions, convective models of heat transfer in magma do not give fundamentally different results from purely conductive models. Thus, local convection should not increase significantly the heat transfer. The possible slight increase of heat transfer that would be related to small-scale convection is accounted for by the wide range of thermal conductivity that we explored.

The other observation we used to constrain the magma emplacement rate is the thermal aureole thickness,  $w_t$ . In the model,  $w_t$  was defined as the thickness of country rock whose temperature has exceeded 400°C during the magmatic event.

The two constraints (absence of large-scale convection and a narrow thermal aureole) were tested independently. No assumption was made about the size of the thermal aureole when the conditions for the absence of convection were explored and, similarly, no assumption was made about the occurrence of convection when we studied the size of the thermal aureole.

### Model limitations

We developed a one-dimensional model, which allowed us to test a large set of parameters. Such a detailed parameter study would not have been possible with a two-dimensional model. However, using a one-dimensional model implies that on long time-scales, the temperatures might be overestimated because horizontal heat loss is neglected. This time-scale can be approximated as follows (Turcotte & Schubert, 1982):

$$t = \frac{L^2}{\kappa}. \quad (9)$$

For a body that is 10 km wide, the time-scale for lateral heat flow is about 4.5 Myr. If the time-scale of the granite emplacement is no more than 5 Myr, as suggested by Harrison *et al.* (1999b), the lateral heat flow can be neglected. At lower granite emplacement rates, we will show later that the thermal evolution of the system is controlled by individual sills cooling on time-scales much shorter than 4.5 Myr and thus that the lateral heat flow is insignificant.

Conductivity anisotropy is not captured by our one-dimensional model. We have modelled the country rock as homogeneous but, as stressed above, the sediments and crystalline basement surrounding the HHL are layered and probably have conductivity variations.

The heat advected by the fluids exsolved by the crystallizing granite is not included in the model. Fluid advection would increase the size and temperature of the thermal aureole. Thus, the results we obtained are conservative.

## RESULTS

### The thermal evolution of the system

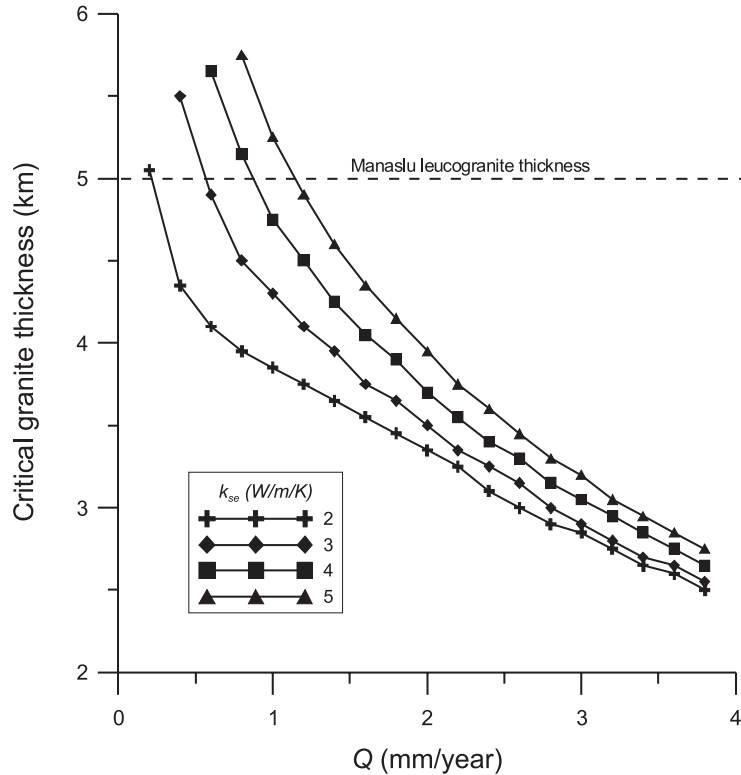
To constrain the conditions that give rise to a thin thermal aureole and the absence of large-scale convection, we made a parametric study on rock conductivities, magma emplacement rates, emplacement depths, initial temperatures and individual sill thicknesses. Early sills injected into cold crust solidify and transfer their heat to the country rock. The consecutive temperature evolution in the growing granitic body and in the surrounding crust is the result of a competition between the sensible and latent heat advected by the magma, and the heat dissipated

toward the surface by conduction. Thus, the parameters that control the thermal evolution of the system are the magma emplacement rate and rock thermal conductivity. If more heat is advected than can be conducted away, the temperatures increase in the granite and country rocks, resulting in the growth of a contact thermal aureole. Eventually, newly injected sills equilibrate at temperatures above the leucogranite solidus. The intruding sills do not completely solidify any more and a zone of granite containing residual melt starts to accumulate. If the conditions described by equation (6) are satisfied, the magma convects and the intrusions begin to amalgamate and homogenize.

### Conditions for convection

We define a critical granite thickness, which corresponds to the total granite thickness that is accreted before large-scale convection initiates. We first illustrate how magma emplacement rate and country rock conductivity influence the critical granite thickness (Fig. 5). For the Manaslu body, no large-scale convection happened; thus, the critical thickness was larger than the thickness of the body. For the calculation conditions shown in Fig. 5—a granite roof at 12.5 km depth and no repose period—emplacement rates decrease with sediment conductivity. An average TSS conductivity of 4 W/m/K or less constrains the maximum emplacement rate to <0.8 mm/year; this value corresponds to the emplacement of 5 km of granite in 6.25 Myr.

In the parametric study, we use the critical emplacement rate,  $Q_c$ , which is the maximum emplacement rate allowing 5 km of leucogranite to be emplaced without onset of large-scale convection. Figure 6 shows the relationships between  $Q_c$  and the sediment conductivities, with and without the repose period, for injection depths of 10, 12.5 and 15 km and for HHC conductivities of 2 and 3 W/m/K. As shown before, high sediment conductivities allow fast cooling and high  $Q_c$ . High  $Q_c$  is also favoured by low HHC conductivities, because less heat comes from depth, and by shallow injection depth (Fig. 6). The initial temperature,  $T_b$ , at the locus of magma emplacement does not influence  $Q_c$  significantly. With a shallow granite roof depth (10 km), a 4 Myr period of repose allows  $Q_c$  for each magma pulse to be higher than  $Q_c$  for continuous emplacement, because the crust is cooling down during the repose period. At greater depths, the difference in emplacement rate between the two models is small (Fig. 6). According to these results, the emplacement rate of the Manaslu granite is estimated to be less than 1.6 mm/year if emplacement is continuous and <2.3 mm/year with a 4 Myr repose. The granite emplacement duration is obtained by dividing 5 km by the emplacement rate. To this duration, 4 Myr must be added in the case of a repose period. Thus, although



**Fig. 5.** Critical granite thickness that needs to be emplaced before conditions for convection are reached as a function of magma emplacement rate ( $Q$ ), for different overlying sediment conductivities ( $k_{se}$ ). The critical emplacement rate for the Manaslu ( $\sim 5$  km thick) is defined by the intersection of the curves with the dashed line. The sill thickness is 50 m, the granite conductivity is 3 W/m/K, the HHC conductivity is 2 W/m/K, the initial HHC–sediment boundary is at 12.5 km, and the initial geothermal gradient is linear with the HHC–sediment interface temperature  $T_i = 300^\circ\text{C}$ .

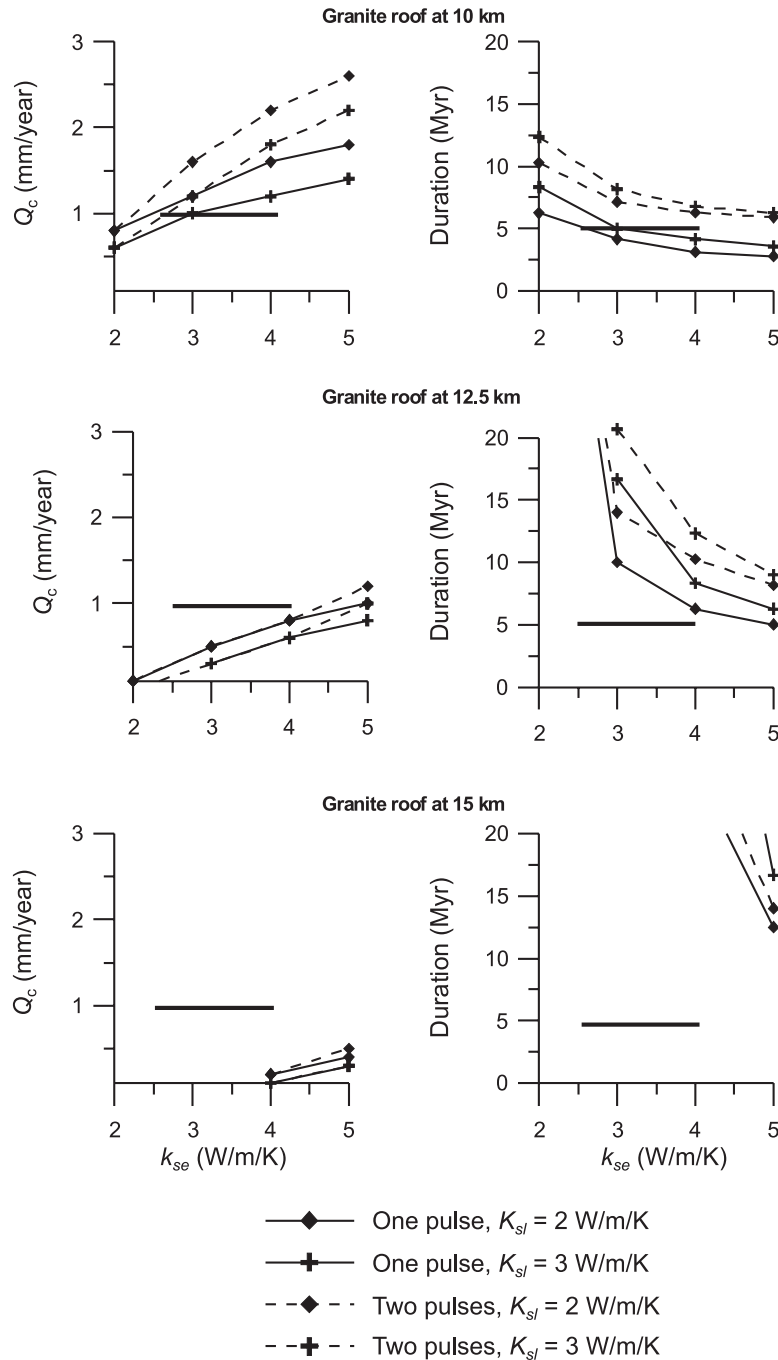
each magma pulse emplacement rate is higher in the case of two magma pulses, the averaged emplacement rate is lower and the emplacement duration is longer (Fig. 6).

According to Harrison *et al.* (1999b), the age difference between the Manaslu top and bottom is 5 Myr. Figure 6 suggests that this emplacement duration is consistent with the absence of convection if the emplacement depth is shallow (granite roof at 10 km). For a granite roof at 12.5 km, emplacement durations of 5 Myr are possible only with unrealistic sediment conductivities ( $>4.5$  W/m/K). At depths of 15 km or more, forming a 5 km thick pluton in 5 Myr results in convection, using reasonable values of sediment conductivities. This suggests that the preservation of isotopic heterogeneities within the thickest HHL is not consistent with a deep intrusion level, assuming a static overlying column. Although we have not tested initial temperatures,  $T_i$ , at the locus of magma intrusion hotter than  $350^\circ\text{C}$ , the above results do not substantiate the case for *in situ* partial melting to construct the plutons. Intrusion in a deep environment (i.e. middle crust) will make it very difficult to avoid convection and mixing within thick HHL laccoliths.

### The thermal aureole

Figure 7 shows the width of the thermal aureole,  $w_t$ , in the granite overlying sediments for a granite roof at 10 km and sill thickness of 50 m, for a total granite thickness of 5 km. Note that in Fig. 7, part of the results involves emplacement rates that are above the critical emplacement rate for convection as discussed above. As our model of heat transfer is conductive only, the aureole thicknesses shown in Fig. 7 are minimum values. For most emplacement rates and country rock conductivities investigated, the thickness of the thermal aureole depends on both of those parameters (Fig. 7). For instance, at an emplacement rate of 1 mm/year, any sediment conductivity lower than 4 W/m/K yields an aureole thickness of at least 1 km. However, there is a domain at low emplacement rates and/or high country rock conductivities, where the aureole thickness is nearly independent of emplacement rate and only depends slightly on conductivities (Fig. 7). Under such conditions, heat is conducted away through the rocks overlying the granite sufficiently rapidly to balance the heat advected by magma input and conducted through the formerly emplaced granite pile. We define a critical emplacement rate,  $Q_a$ , for the growth of the thermal aureole. Below  $Q_a$ , the thermal aureole

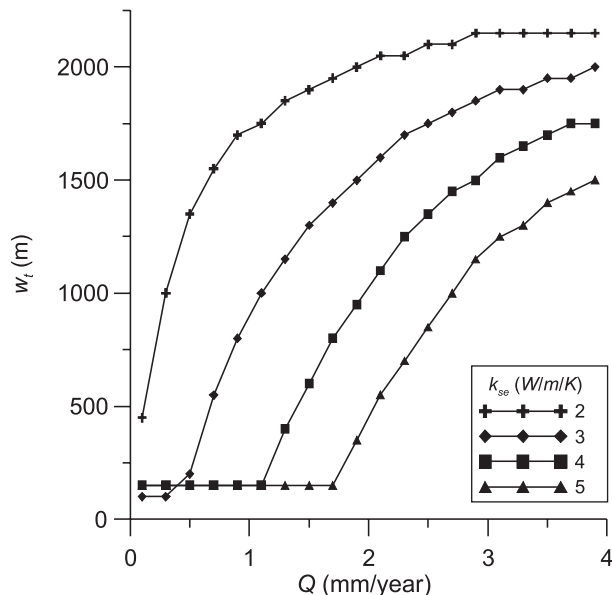




**Fig. 6.** Critical emplacement rate for convection ( $Q_c$ ), and total emplacement duration, corresponding to a critical granite thickness of 5 km, for different sediment conductivities  $k_{se}$ , HHC conductivities  $k_{hs}$ , and initial HHC–sediment interface depths. Continuous lines: sills are regularly injected during the emplacement of 5 km of granite; dashed lines: sills are emplaced in two magmatic pulses separated by a repose period of 4 Myr as suggested by Harrison *et al.* (1999b). The thick horizontal line marks the range of realistic sediment conductivities and Manaslu emplacement rates or duration inferred from geochronological data (Harrison *et al.*, 1999b). These results are for a leucogranite injection temperature of 780°C. The maximum difference in  $Q_c$  for a lower injection temperature of 750°C is 0.2 mm/year.

reaches its maximum thickness after the first sill emplacement, and its size depends on the individual sill thickness (Fig. 8) and not on the total thickness of granite.  $Q_a$  is strongly dependent on country rock conductivity (Fig. 9).

$Q_a$  is smaller than  $Q_c$ , the critical emplacement rate for convection (Fig. 6). Whatever the initial temperature, the system tends toward an equilibrium geothermal gradient and  $Q_a$  can be higher if the equilibrium temperature at



**Fig. 7.** Top thermal aureole thickness  $w_b$ , defined as the thickness of country rock heated above  $400^\circ\text{C}$ , as a function of the magma emplacement rate  $Q$ , for different sediment conductivities  $k_{se}$ . The total accumulated granite thickness is 5 km; the granite top is at 10 km, the sill thickness is 50 m, the granite conductivity is  $3\text{ W/m/K}$ , the HHC conductivity is  $2\text{ W/m/K}$ , and the initial geothermal gradient is linear with HHC–sediment interface temperature  $T_i = 300^\circ\text{C}$ . Here  $w_t$  is calculated with a precision of 50 m, which is why the curves are not smooth.

the granite emplacement depth is low. The equilibrium temperature in the system is controlled by the HHC, granite and sediment conductivities, and by the emplacement depth of the granite (Fig. 4). Thus, high values of  $Q_a$  are favoured by shallow emplacement depth, low HHC conductivities and high overlying sediment conductivities (Fig. 9). Figure 4 shows that there is a domain at high HHC conductivities, and low sediment conductivities, where the equilibrium temperature is above  $400^\circ\text{C}$ . In this domain, the thermal aureole cannot remain thin.

We found that for  $Q_a$  to be above  $0.1\text{ mm/year}$  (the lowest tested emplacement rate), the magma injection depth must be  $<15\text{ km}$ . For emplacement at  $12.5\text{ km}$  depth, critical emplacement rates of more than  $0.1\text{ mm/year}$  are possible, but require unrealistically high sediment conductivities. For  $10\text{ km}$  depth and for sediment conductivities in the range  $3\text{--}4\text{ W/m/K}$ , it is possible to build a  $5\text{ km}$  thick lens with a corresponding small contact aureole (Fig. 9). For this range of sediment conductivity, greater depths of intrusion increase the period of pluton growth to at least  $10\text{ Myr}$ , i.e. significantly beyond the geochronologically constrained interval of  $5\text{ Myr}$ . Below  $Q_a$ , the aureole thickness is two to six times the sill thickness, depending on the initial crustal temperature (Fig. 8). Our computation suggests that to

form a thermal aureole less than  $100\text{ m}$  thick, the typical sill thickness should be  $<60\text{ m}$ , which is about the scale-length of the Rb–Sr heterogeneity at Manaslu (Deniel *et al.*, 1987).

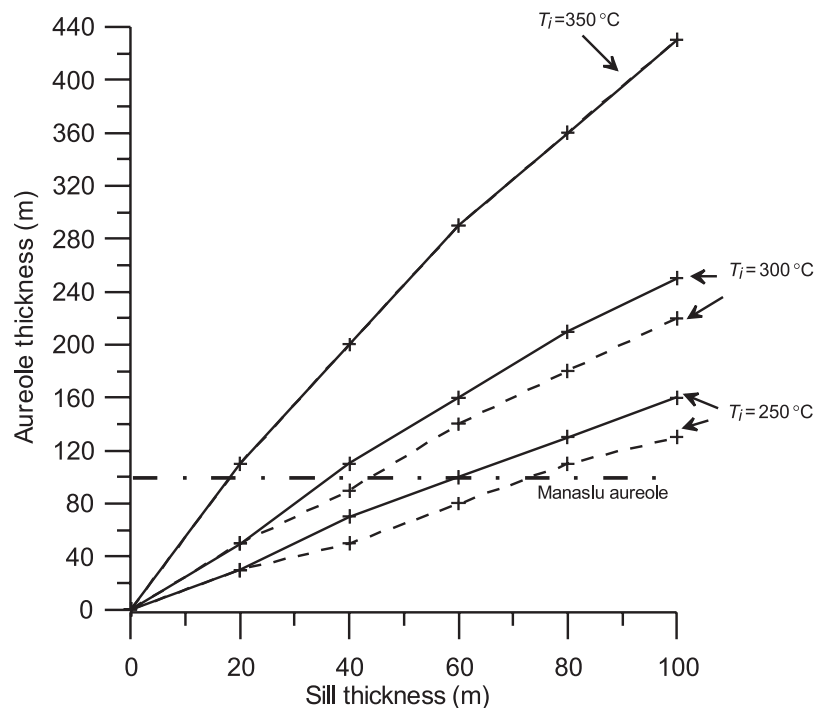
We have computed the temperature in the country rock at every  $1\text{ m}$  over a distance to the contact of  $50\text{ m}$ . The thermal aureole peak temperature is a maximum at the contact with the granite and decreases further away (Fig. 10). The slope of this temperature decrease is steeper with thin sills. Below  $Q_a$ , the aureole peak temperature is not sensitive to the country rock conductivity, or to the injection depth, but depends on the initial country rock temperature, magma temperature and the sill thickness (Fig. 11). Our results suggest that the peak temperature of  $550 \pm 40^\circ\text{C}$  inferred from thermobarometric studies on the Manaslu top aureole (Guillot *et al.*, 1995a) is consistent with an initial country rock temperature at the onset of injection of no more than  $300^\circ\text{C}$ , in agreement with previous estimates (Copeland *et al.*, 1990), and a sill thickness of  $<50\text{ m}$ .

### Tectonic unroofing and tectonic boundary

A tectonic top contact implies that the thermal aureole thickness is not a relevant constraint on the emplacement rate because of the renewal of the country rock. We modelled the process of tectonic unroofing by keeping the roof temperature,  $T_b$ , constant during emplacement as a consequence of fault movement. In this model, the maximum emplacement rate for  $5\text{ km}$  of leucogranite to emplace without onset of large-scale convection,  $Q_c$ , is independent of the overlying sediment conductivity.  $Q_c$  depends on  $T_i$  and not on the emplacement depth. As expected,  $Q_c$  is higher than in the case of a statically heating roof (Fig. 12).  $Q_c$  is about the same, with or without a repose period; thus, allowing for the repose period adds only  $4\text{ Myr}$  to the total emplacement time (Fig. 12). The highest  $Q_c$  is  $3.4\text{ mm/year}$  with  $T_i = 250^\circ\text{C}$ . In this latter case, the Manaslu granite could have been emplaced in  $1.3\text{ Myr}$  if its emplacement was not interrupted. In the case of a  $4\text{ Myr}$  repose period, the total duration is at least  $5.3\text{ Myr}$ .

### Thermal effects of HHL intrusion

Figure 13a shows the temperature profile in the crust after the emplacement of  $5\text{ km}$  of granite between  $10$  and  $15\text{ km}$  depth at a rate of  $1\text{ mm/year}$ . The case for tectonic unroofing is shown in Fig. 13b. In the case of tectonic unroofing, the emplacement level is deeper ( $12.5\text{ km}$ ) and the emplacement rate is higher ( $2.5\text{ mm/year}$ ). However, because the constraints of no large-scale convection apply in both cases, the maximum temperatures allowed in the granite body and the bottom contact temperature are the same. In the HHC, close to



**Fig. 8.** The thermal aureole thickness for conditions below the critical emplacement rate, as a function of sill thickness for different HHC–sediment interface temperatures  $T_i$ . Continuous lines: the latent heat of sediment metamorphism is neglected; dashed lines: the sediment latent heat of metamorphism is 600 J/kg/K between 400°C and 600°C (Furlong *et al.*, 1991). HHC and sediment conductivities are 2 and 3.5 W/m/K respectively. The horizontal dot-dashed line marks the maximum aureole thickness overlying the Manaslu leucogranite. The aureole thickness is calculated with a precision of 5 m.

the bottom granite contact, the temperature reaches 700°C when the magma is intruded. However, this peak in temperature is transient and restricted to a few tens of metres close to the contact. On a larger scale, the HHC temperature equilibrates at 600°C in the proximity of the granite (Fig. 13). Under the lithostatic pressure at 15 km depth and below (pressure > 4 kbar), this temperature is below the metapelite solidus temperature (Clemens & Vielzeuf, 1987; Patiño Douce & Harris, 1998). The dehydration melting temperature of pelites (about 700°C) is reached more than 8 km below the granite bottom contact. Thus, our results suggest that the increase in migmatization toward the thickest HHL cannot be the effect of HHL cooling.

### Cooling of the leucogranite

The temperature evolution at depths of 10, 11 and 12.5 km during the emplacement of the granite and until 3 Myr after the end of intrusion is plotted in Fig. 14. The leucogranitic body extends from 10 to 15 km. We assumed here that the first sill was injected at 24 Ma at an emplacement rate of 1 mm/year and that intrusion stopped at 19 Ma. The temperature evolution at 10 km depth corresponds to the temperature evolution of the first emplaced sill. The thickening of the 10 and

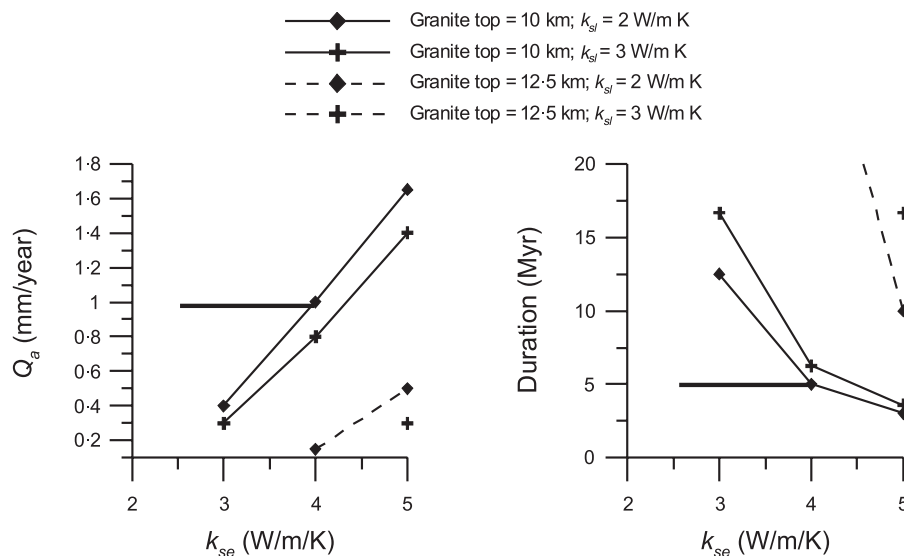
12.5 km curves at 23 and 21.5 Ma, respectively, corresponds to temperature oscillation because of the injection of sills at or close to this level.

The Ar<sup>40</sup>–Ar<sup>39</sup> data for the top of Manaslu granite indicate a closure age of muscovite in the range of  $17.1 \pm 0.1$  to  $18.4 \pm 0.1$  Ma (Copeland *et al.*, 1990), suggesting that the upper part of the granite cooled below  $357 \pm 38^\circ\text{C}$ , 1–2 Myr after the end of intrusion. The temperatures we computed for the top of the granite are in good agreement with the Ar–Ar data. For the lower part of the body, Ar–Ar ages range from  $14.7 \pm 0.2$  to  $16.6 \pm 0.1$  Ma. At this time, our computed temperatures at 12.5 km depth, in the middle of the body, are above the muscovite closure temperature (Fig. 14). This discrepancy can be explained if the granite emplacement rate was less than the modelled rate of 1 mm/year, or if granite cooling was accelerated by the post-emplacement slip of the STD.

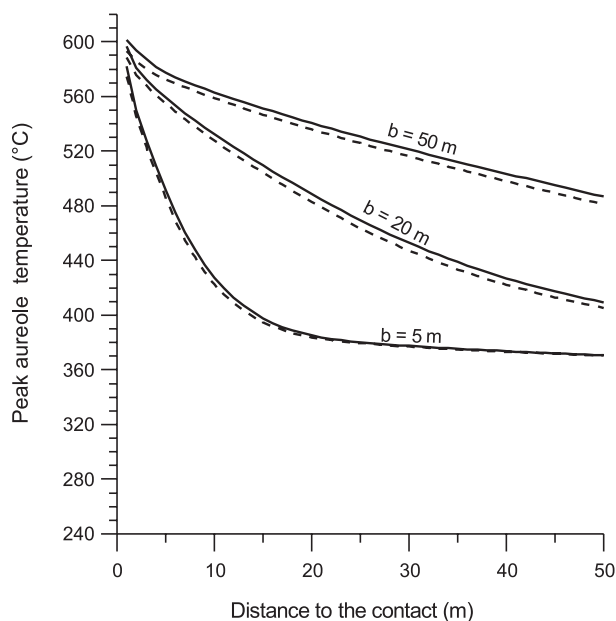
## DISCUSSION

### The emplacement of the Manaslu granite

Heat transfer modelling provides constraints on the upper limit of emplacement rates for the HHL. Geochronological data suggest that the emplacement of the



**Fig. 9.** Critical emplacement rate  $Q_a$  limiting the aureole growth and corresponding emplacement duration for different sediment conductivities  $k_{se}$ , HHC conductivities  $k_{gf}$ , and initial HHC–sediment interface depths. The bold horizontal line marks the range of realistic sediment conductivities and the Manaslu emplacement rate or duration inferred from geochronological data (Harrison *et al.*, 1999b). These results are for a leucogranite injection temperature of 780°C. The maximum difference in  $Q_a$  for an injection temperature of 750°C is 0.2 mm/year.

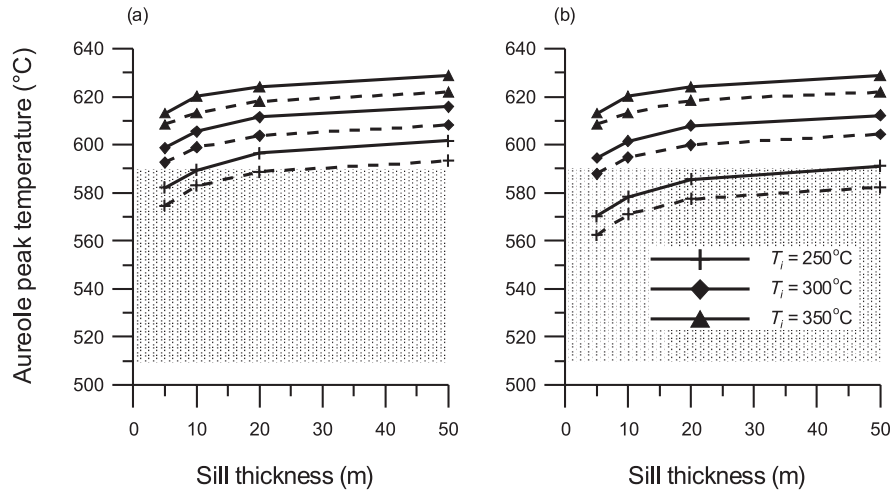


**Fig. 10.** Profiles of the peak temperature in the top thermal aureole as a function of distance to the contact, for different sill thickness  $b$ , for injection temperatures of 780°C (continuous lines) and 750°C (dashed lines). The sediment latent heat of metamorphism is 600 J/kg/K between 400°C and 600°C (Furlong *et al.*, 1991). Initial temperature  $T_i$  is 250°C, emplacement depth is 10 km,  $k_{se}$  is 3.5 W/m/K, emplacement rate is below  $Q_a$  (0.5 mm/year).

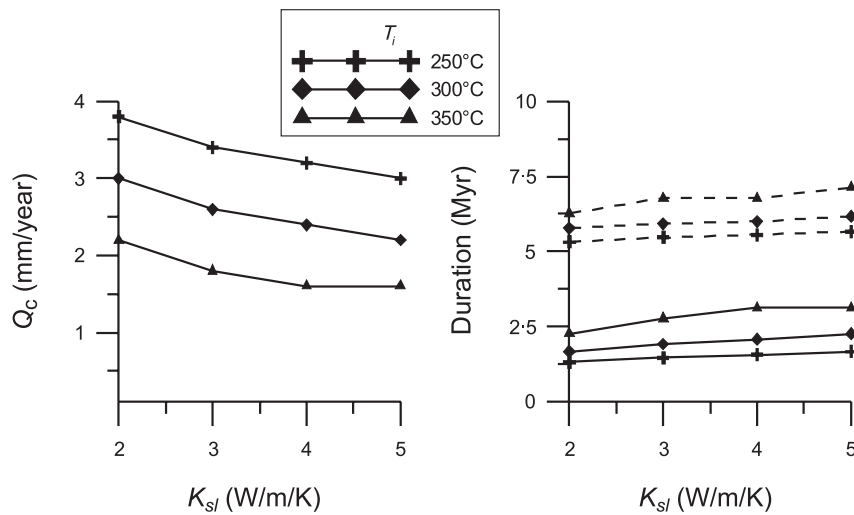
Manaslu intrusion took place over no more than 5–6 Myr. If the intrusion rate was steady, then these data suggest that the intrusion rate could not have been lower than 1 mm/year. Our modelling suggests that many

parameter ranges, consistent with the observations of no convection and a thin thermal aureole, fall below this lower limit. However, model results with shallow emplacement (roof at 10 km) are at, or above, 1 mm/year, and are, therefore, consistent with the geochronological data. The thickness and temperature of the Manaslu thermal aureole can be explained by the intrusion of sills 20–60 m thick, in broad agreement with the size of individual magma batches inferred from Rb–Sr systematics (Deniel *et al.*, 1987). A 1 mm/year emplacement rate implies that the average time interval between injections was of the order of 20 000–60 000 years. Two magma pulses separated by a repose period of 4 Myr, as proposed by Harrison *et al.* (1999b), permit a higher emplacement rate for each pulse. However, when the repose time is included in the total emplacement duration, this duration is more than 5 Myr. Thus, a better fit between the thermal results and the geochronological data is found with a continuous emplacement, rather than with a repose period.

If the top contact is not tectonic, for the leucogranite to avoid large-scale convection and thermal aureole growth, the model requires the HHC conductivity to be lower than that of the sediments, in contrast to the conclusions of Jaupart & Provost (1985) and Pinet & Jaupart (1987). Those workers proposed a model of heat refraction at a high conductivity HHC–low conductivity sediment interface; heat focusing causes partial melting at this interface. Such an approach assumes that melt generation and leucogranite emplacement have occurred at the same crustal level, as it seems to have done along some transects such as in the Everest–Makalu area (Visona &



**Fig. 11.** Peak temperatures in the sediments at distance of 1 m from the granite contact for different sill thickness and  $T_i$ . (a) The latent heat of sediment metamorphism is neglected. (b) The sediment latent heat of metamorphism is 600 J/kg/K between 400°C and 600°C (Furlong *et al.*, 1991). The emplacement rate is below  $Q_a$  (0.5 mm/year). The continuous lines are for a sill intrusion temperature of 780°C, the dashed lines are for a sill intrusion temperature of 750°C. The peak temperatures range inferred from thermobarometry (Guillot *et al.*, 1995) is shown by the grey shaded area.



**Fig. 12.** Critical emplacement rate for convection  $Q_c$  and total emplacement duration, corresponding to a critical granite thickness of 5 km, in the case of unroofing, for different HHC conductivities  $k_{sl}$  and temperatures  $T_i$ . The sediment temperature at the granite contact is kept constant and equal to  $T_i$ .  $Q_c$  with or without a repose period is the same. In the case of a repose period, the duration of the repose period must be added to the total duration time (dashed lines).

Lombardo, 2002; Searle *et al.*, 2003). However, in order to avoid convection, *in situ* models are constrained to produce HHL at small rates and to allow the first-intruded magma layer to cool before the next sill is intruded. The latter condition seems particularly difficult to realize, as melting driven by thermal refraction is likely to be a continuous process, and the melt fraction should increase with time. We conclude that such models, although conceptually viable, are not applicable to the HHL whenever they exhibit isotopic heterogeneities.

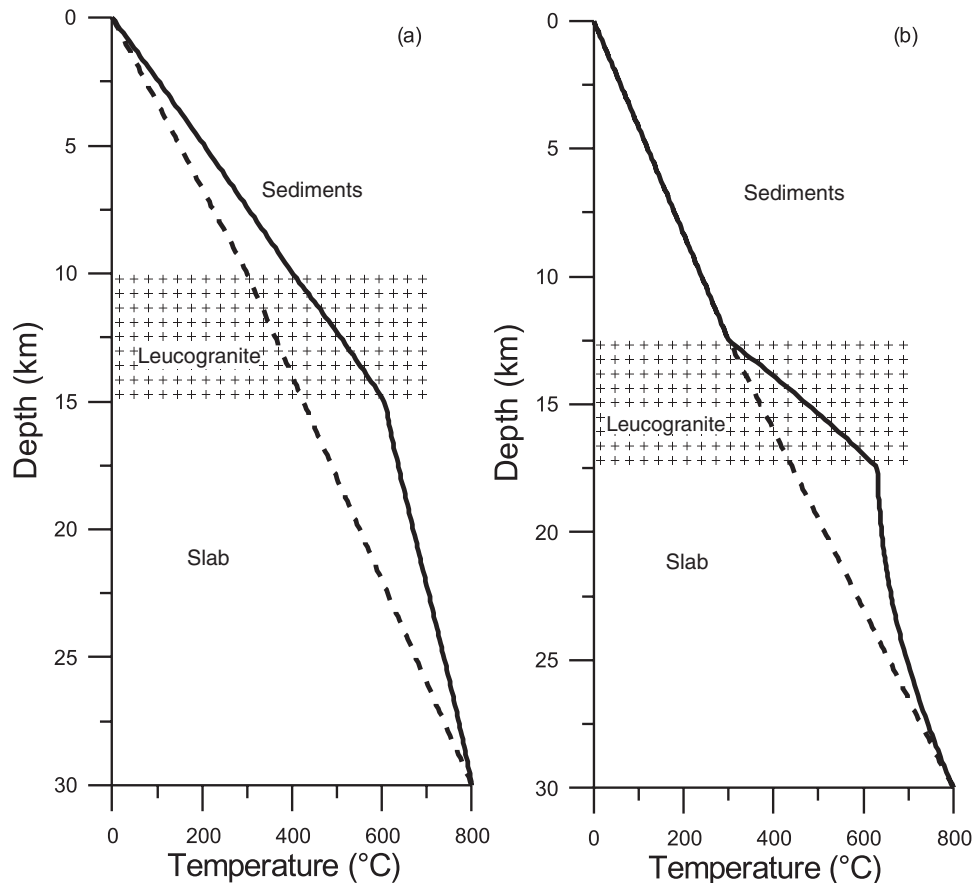
Because of the slow accretion rate, the amount of heat available at any time at the intrusion locus to establish a

hydrothermal convective system is small. Stable isotope studies on the Manaslu granite (France-Lanord *et al.*, 1988) also indicate that there has not been any large-scale hydrothermal interaction of the granite with meteoric water.

### Were HHL intrusion and STD motion coeval?

If the upper contact is tectonic in origin, then the existence of a thin aureole is not a constraint; high emplacement rates become feasible even for an emplacement

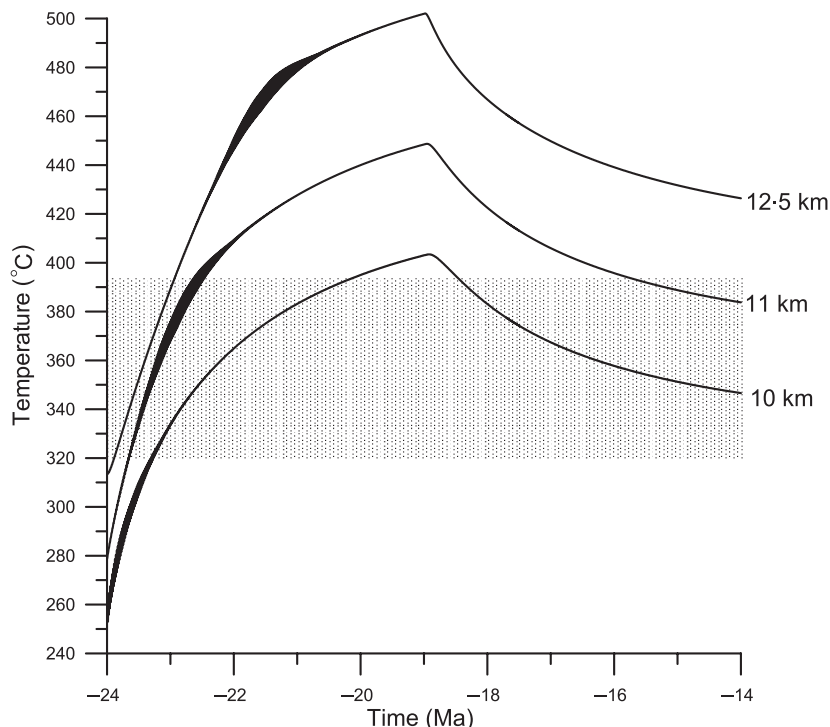




**Fig. 13.** Profiles of crustal temperature after the emplacement of 5 km of leucogranite at a rate close to  $Q_a$  (continuous line). The dashed line shows the initial geothermal gradient. (a) The leucogranite roof is allowed to heat; depth of emplacement is 10 km,  $Q$  is 1 mm/year, and the sediment conductivity is 4 W/m/K. (b) The roof of the leucogranite is at a fixed temperature of 250°C (unroofing case); the emplacement depth is 12.5 km,  $Q$  is 2.5 mm/year. Sills are 15 m thick and the HHC conductivity is 2 W/m/K in both (a) and (b).

level between 15 and 20 km. We note, however, that everywhere in the Himalaya, the HHL contain magmatic muscovite, which restricts pressure to above  $\sim 2.8$  kbar during crystallization. This does not preclude that solidified parts of the granite went outside the muscovite stability field during pluton growth, but it does imply that its molten parts remained in the  $P$ – $T$  field of muscovite crystallization. This suggests that pressure conditions remained more or less constant during the intrusive stage, at least at the locus of magma intrusion. If initial intrusion was at 3–4 kbar, then motion on the STD cannot have been extensive during most of the period of pluton growth. For a pluton emplaced at deeper levels, the possibility of STD slip during laccolith growth can be evaluated using the simple model illustrated in Fig. 15, where a pluton intruded at a depth of 15 km is intersected by the STD, which dips north at an angle of 30°, although we note that the dip angle of the STD can be lower (e.g. Burchfiel *et al.*, 1992). Assuming a slip rate on the STD of 20 mm/year (Dèzes *et al.*, 1999; Harrison *et al.*, 1999a), and that muscovite cannot crystallize in the HHL

at depths shallower than  $\sim 10$  km, then the geometrical relationships show that it would take 0.5 Myr for the pluton to rise to a depth of 10 km by fault slip alone (see Fig. 15 caption). Starting at 20 km depth, with the same dip angle and slip rate, gives a duration of 1 Myr to unroof the granite from 20 to 10 km, which is about the time required to intrude 5 km of leucogranite. However, both the dip angle and the slip rate of the STD are likely to vary along strike; thus, these estimates are only indications. If tectonic unroofing indeed took place during HHL intrusion and crystallization, then it requires a slip rate  $< 20$  mm/year and a dip angle no higher than 30° such that none of the HHL so far documented along the 2500 km strike of the Himalaya range reached a depth shallower than 10 km while still molten. Because deformation related to the STD has demonstrably affected the HHL once crystallized (Burg *et al.*, 1984; Searle *et al.*, 1997), the timing of STD slip during and after HHL intrusion should have followed a precise temporal pattern so as to allow first magmatic crystallization of muscovite and then subsolidus deformation. A detailed study of the



**Fig. 14.** Temperature evolution in the system at 10, 11 and 12.5 km depth. The leucogranite intrusion starts at 24 Ma and stops at 19 Ma according to geochronological data (Harrison *et al.*, 1999b). The curve for 10 km corresponds to the temperature evolution of the first emplaced sill. The apparent thickening of the curves is due to temperature oscillations corresponding to sill emplacement at or close to the plotted level. The magma is injected at 750°C with an emplacement rate of 1 mm/year. The grey shaded area represents the closure temperature of muscovite.

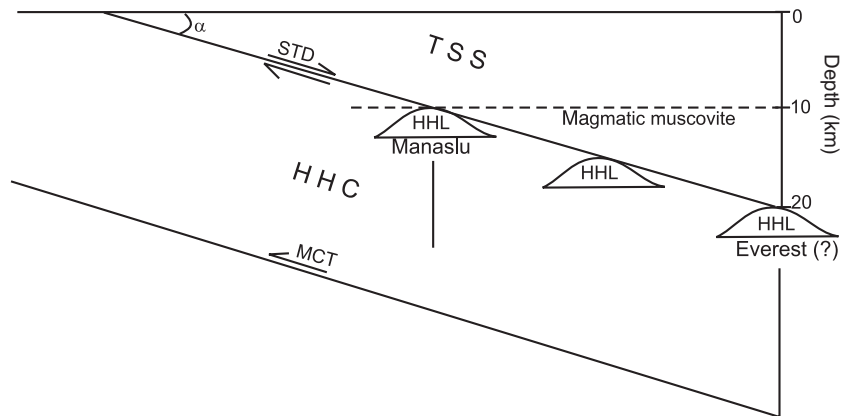
possible effects of extensional fault activity on the cooling of the HHL is presented elsewhere (Annen & Scaillet, in preparation).

An observation that argues against a close relationship between active magmatism and extensional faulting is that the deformation related to the STD is concentrated above and not within the HHL. If the STD was active during HHL emplacement, then we would expect that each newly injected thin sill would act as a preferential décollement layer, as it represents the weakest rheological level of the deforming upper pile (Davidson *et al.*, 1994). The very fact that the largest HHL have mostly isotropic magmatic fabrics strongly suggests that, for the most part, they were fully crystallized when STD motion started, acting as rigid objects, and that deformation was concentrated above them in the more ductile sedimentary layers. This view is supported by the fact that, at least at Manaslu, the barometric difference between the top and bottom contact of the intrusion fits with the present-day structural thickness (Guillot *et al.*, 1995a), implying that the entire body has not been significantly shortened or thinned by regional deformation. This does not preclude the possibility that smaller intrusions could have been actively deformed while partly liquid, as observed in some instances (Dèzes *et al.*, 1999; Murphy & Harrison,

1999) and, in fact, the protracted time interval of HHL crystallization documented at many places along the range (Coleman, 1998) offers room for small, late-stage, intrusive stocks to be affected by the STD. However, the field evidence argues against the idea that the main phase of extension was coeval with the main phase of Manaslu intrusion. In particular, Guillot *et al.* (1993) reported the Manaslu granite as cross-cutting the Naïke fold—a large anticline structure that they related to the first phase of extensional tectonics having affected this area. We thus contend that Manaslu intrusion probably occurred before the main phase of STD slip. We recognize, however, that the timing constraints suggest STD motion immediately after the main phase of HHL intrusion.

### Implications for magma production

The zircon and monazite saturation systematics of the HHL show that the residence time of the parental anatectic melts in their source was short, of the order of a few thousand years, and certainly shorter than 50 000 years (Ayres *et al.*, 1997). This agrees well with our model results of thin sills intruded at intervals of 20 000–60 000 years. For a given emplacement rate, thicker sills require a longer time interval between melt injections.



**Fig. 15.** Schematic N–S cross-section showing the geometrical relationships between the HHL plutons and the South Tibetan Detachment (STD), assuming that intrusion and extension processes were coeval. TSS is the Tethyan Sedimentary Series, HHC is Higher Himalayan Crystallines, MCT the Main Central Thrust. The dashed horizontal line marks the upper possible limit for magmatic crystallization in the HHL. Three possibilities of HHL emplacement are shown: the shallowest corresponds to the Manaslu, whereas the deepest could correspond to intrusives in the Everest–Makalu area. Depending on the dip angle ( $\alpha$ ) of the STD and on the slip rate along the STD, the time allowed for pluton intrusion to reach the limiting pressure below which no muscovite can crystallize in the HHL, owing to STD slip, may vary along strike. This time  $t$  (in years) is given by the equation  $t = [(D - 10)/\sin(\alpha)]/(V \times 10^{-6})$  where  $V$  is the slip rate in mm/year and  $D$  is the intrusion starting depth in km.

Short residence times suggest that whatever the physical processes of melt extraction and collection from the source, these are not the rate-limiting factors of pluton growth (Harris *et al.*, 2000). Similarly, rates of magma ascent through the dyke network are expected to be geologically instantaneous (Scaillet *et al.*, 1996), and cannot be the controlling parameter of long-term intrusion rates. As explained in the Introduction, in the Himalaya range, the main heat source was internal radioactive decay following crustal thickening, perhaps helped by frictional heating along large thrust faults (England *et al.*, 1992; Harrison *et al.*, 1998). Previous thermal modelling has shown that the time-scale of thermal excursion of thickened crust into the melting range ( $>650^\circ\text{C}$ ) is of the order of 10–20 Myr, or more, depending on erosion–decompression rates (England & Thompson, 1984), in broad agreement with our findings for the time needed for the upper crustal assembly of anatectic melts into sizeable plutonic bodies. We thus conclude that melt availability for pluton growth is fundamentally controlled by the rate of heat supply into the protoliths—a conclusion also reached by Harris *et al.* (2000). Thus, although the very process of magma ascent within the crust can be very fast, the integrated time to build a pluton of batholithic dimensions in collisional orogens with no mantle supply is several Myr.

## CONCLUSIONS

We have shown that it is possible to emplace a thick magma body in the upper crust that has a rather limited thermal effect on its surroundings (i.e. a small contact

aureole one order of magnitude smaller than the thickness of pluton). Our results indicate that for an igneous body with a thickness of several kilometres, the absence of convection and presence of a narrow thermal aureole can be explained by the slow accretion of separate magma batches, injected as thin sills.

Application of our data to the Manaslu pluton shows that, with the most conservative assumptions, the maximum emplacement rate of this pluton was 3.4 mm/year if unroofing was active during emplacement, and 1 mm/year if no unroofing took place. First-order geological observations, however, seem to preclude strict contemporaneity between STD slip and HHL intrusion, and thus our preferred scenario is that intrusion of the HHL occurred before the main phase of extensional tectonics. In such a case, the best fit with available geochronological data, which suggest an emplacement duration of 5 Myr, and geochemical and thermobarometric constraints, is found for 20–60 m thick sills emplaced at relatively shallow depth (granite roof at about 10 km) in a crust with an initial temperature of no more than  $300^\circ\text{C}$ . A steady emplacement rate is in better agreement with the geochronological data than several short magma pulses separated by long repose periods. The results agree with the view that the time-scales of emplacement of the granite at shallow level are commensurate with the time-scale of melt generation at depth. In collisional settings with no mantle contribution, melt production and pluton emplacement must take place over several million years—a view that contrasts with recent proposals that most granitic plutons are emplaced in less than 100 000 years, irrespective of tectonic setting (Petford *et al.*, 2000).

From the perspective of the mechanisms of exhumation of deep-seated crustal rocks in collisional ranges, the available data at Manaslu do not support the idea that the MCT and the STD were strictly synchronous in this area, but instead suggest that STD initiated after crystallization of the pluton emplacement. Future work is needed to evaluate whether such diachronicity is specific to Manaslu or applies to other HHL and, if so, to explore to what extent this diachronicity is compatible with the concept of crustal channel flow as currently envisioned for the Himalayas.

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## APPENDIX

For the geothermal gradient to be at equilibrium, the heat flux at the Earth's surface must balance the heat flux input at the bottom of the HHC and the radioactive heat production. The equations of heat balance have been given by Turcotte & Schubert (1982):

$$T(z) = \begin{cases} T_s + \frac{q_{se}}{k_{se}}z - \frac{A_{se}}{2k_{se}}z^2 & 0 < z < z_{se} \\ T_i + \frac{q_i}{k_{sl}}(z - z_{se}) - \frac{A_{sl}}{2k_{sl}}(z - z_{se})^2 & z_{se} < z < z_{sl} \end{cases} \quad (\text{A1})$$

The subscript *se* is for sediments and the subscript *sl* is for the HHC (Figs 1 and 2).  $A_{se}$  and  $A_{sl}$  are the radioactive heat production in the sediments and the HHC, respectively.  $z$  is depth,  $z_{se}$  and  $z_{sl}$  are, respectively, the sediment–HHC interface and the HHC bottom depths,  $q_{se}$  is the surface heat flux, and  $q_i$  is the heat flux through the sediment–HHC interface:

$$q_{se} = q_i + w_{se}A_{se} \quad (\text{A2})$$

$$q_i = k_s \frac{T_m - T_i}{w_{sl}} + \frac{w_{sl}}{2} A_{sl} \quad (\text{A3})$$

$$T_i = \frac{w_{sl}}{k_s w_{sl} + k_s w_{se}} \left( k_s T_s + \frac{k_s w_{se}}{w_{sl}} T_m + \frac{A_{sl} w_{se} w_{sl}}{2} + \frac{A_{se}}{2} \right) \quad (\text{A4})$$

where  $T_s$  is surface temperature,  $T_m$  is temperature at the HHC bottom, and  $w_{se}$  and  $w_{sl}$  are sediment and HHC thickness, respectively.