Extreme heterogeneity in Sr isotope systematic in the Himalayan leucogranites: A possible mechanism of partial melting based on thermal modeling

Dilip K Mukhopadhyay

Department of Earth Sciences, University of Roorkee, Roorkee 247667, Uttaranchal, India
e-mail: dilipfes@rurkiu.ernet.in

The small leucogranite plutons occurring in linear belts in the Higher Himalayas have formed due to post-collision partial melting within the Himalayan crust. Several studies have documented that the Sr isotopic ratios in the granite bodies show chaotic variation and meaningful Rb-Sr isochron ages are difficult, if not impossible, to obtain. In tectonically overthickened crust, the depth-temperature profile (geotherm) remains strongly transient for the first tens of millions of years. It is proposed here that the intersecting relations between the transient geotherms and activity-dependent solidus/melting curves may generate small pods of magma at different depths and at different times. Each of these pods will have its unique Sr isotopic ratios. Coalescence of these small pods of magma without any effective homogenization due to deformation-induced fast segregation, ascent and emplacement may lead to pluton-wide extreme heterogeneity in Sr isotopic ratios.

1. Introduction

The Indus - Tsangpo Suture Zone (ITSZ, figure 1) in the Himalayan orogenic belt marks the site of a 50–65Ma old (e.g. Klootwijk *et al* 1992) collision between the Indian and the Eurasian continents. South of the ITSZ, a number of small leucogranitic plutons occur in two linear belts either close to the contact between the High Himalaya Crystalline Zone (HHCZ) and the overlying High Himalaya Sedimentary Zone (HHSZ) or within the HHSZ (figure 1). Field relations and isotope geochemistry suggest that these leucogranites have formed by post-collision partial melting within the Himalayan crust (e.g. Le Fort 1981; Le Fort *et al* 1987). Indeed, Le Fort (1981) considers these granites as a signature of the collision event. A significant character of these leucogranite plutons is that the *intrapluton* Rb-Sr isotopic ratios are so heterogeneous that definitive Rb-Sr isotopic ages for individual plutons are very difficult, if not impossible, to obtain. In a detailed study on the Manaslu granite (see figure 1 for location), Deniel *et al* (1987) documented that the pluton-wide Rb - Sr isotopic ratios show a chaotic variation (figure 2). In order to check the scale of heterogeneity, Deniel *et al* (1987) also analyzed 17 samples collected from an approximately 80 meter long exposure. Of these 17 samples, 11 samples fall on an isochron that gives an age of 18.1 ± 0.5 Ma with initial isotopic ratio (Sr_i) at 0.7470 ± 0.0005. The other 6 samples do not fall on the isochron; they define a poor isochron with approximately similar age but with lower Sr_i. Similar difficulty in obtaining a reliable Rb-Sr age in other leucogranite plutons has been reported by several workers (e.g. Dietrich and Gansser 1981; Stern *et al* 1989).

Since these leucogranites are post-collisional, their ages cannot be more than 50Ma. Consequently, the Rb-Sr isotopic heterogeneity can not be due to variation in age. The calculated Sr_i for 20 Ma for all the samples analyzed by Deniel *et al* (1987) show a large variation between 0.74013 and 0.76205. Therefore, one of the important reasons for the chaotic variation in the Sr isotopic ratios in these leucogranites is obviously isotopic heterogeneity in the source region. However, the large variation in $^{87}Sr/^{86}Sr$ ratios in a scale of less than a hundred meter is rather unusual.

In this article a mechanism of partial melting in tectonically overthickened crust is proposed that can explain, even in outcrop scale, the observed

Keywords. Himalayan leucogranites; partial melting; thermal model; Sr isotope ratio.

Proc. Indian Acad. Sci. (Earth Planet. Sci.), **110**, No. 2, June 2001, pp. 161–169

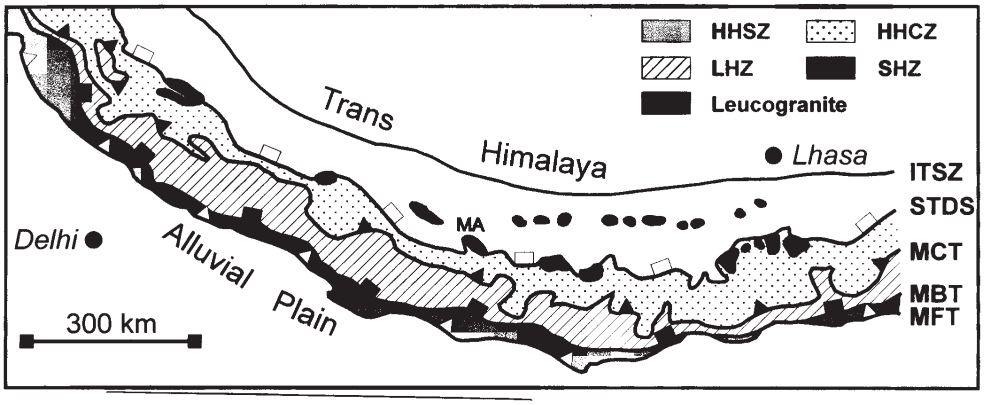

Figure 1. Geological sketch map of the Himalayas showing the locations of leucogranite plutons (dark) in relation to different lithotectonic zones (simplified from Dietrich and Gansser 1981). Note that the leucogranite plutons occur close to the contact between HHSZ and HHCZ or within HHSZ. **SHZ**: Sub-Himalaya Zone; **LHZ**: Lesser Himalaya Zone; **HHCZ**: High Himalaya Crystalline Zone (also called the Central Crystallines or the Tibetan Slab); **HHSZ**: High Himalaya Sedimentary Zone (also called the Tethys Sedimentary Zone); **MFT**: Main Frontal Thrust; **MBT**: Main Boundary Thrust; **MCT**: Main Central Thrust; **STDS**: South Tibet Detachment System (a system of low angle normal faults); **ITSZ**: Indus Tsangpo Suture Zone; **MA**: Manaslu leucogranite (discussed in the text).

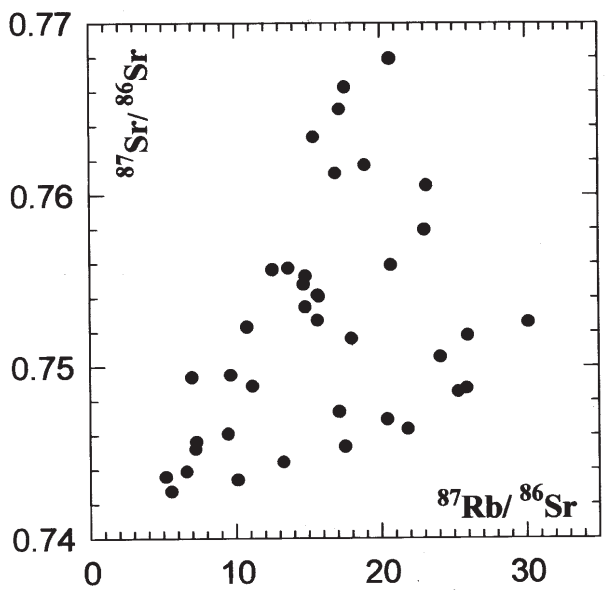

Figure 2. Rb-Sr isotopic diagram for the Manaslu pluton (MA in figure 1). Data taken from table 5 of Deniel *et al* (1987); sample no DK152 with $^{87}Rb/^{86}Sr = 58.4$ plots outside the diagram. From the same pluton 11 of the 18 samples collected from an approximately 80 meter long exposure yield a Rb-Sr age of 18.1 Ma (see Deniel *et al* 1987, figure 2).

Rb-Sr isotopic heterogeneity in the Himalayan leucogranites. The model is based on the geometric relationship between the transient nature of the geotherms (depth-temperature profiles) in tectonically active crust and the variably displaced solidus/melting curves. The proposed model can be important only as a contributing factor in inducing pluton-scale Rb-Sr isotopic heterogeneity in granites, rather than the only means for providing the variation. This is because the petrogenesis of granitic systems is usually complex owing to contributions from too many end-member factors such as source rock heterogeneity, restite-melt segregation/unmixing, P-T-volatile-phase composition, assimilation etc.

The main purpose of the calculated thermal structures presented in this article is to *demonstrate* the transient nature of the geotherm in a tectonically overthickened crust such as the Himalayan orogenic belt. Therefore, relatively simple tectonic and thermal models have been chosen. The calculations are neither novel nor unique, even for Himalayan leucogranites (cf. Oxburg and Turcotte 1974; Molnar *et al* 1983; England and Thompson 1984; Molnar and England 1990; England *et al* 1992; England and Molnar 1993).

2. Tectonic and thermal models

The geophysical data suggest that the doubling of the crustal thickness to about 70–80 km under the Himalayas is due to overthrusting of the HHCZ (also called the Central Crystallines or Tibetan Slab) along a very gently dipping detachment thrust (figure 3(a)). The HHCZ is actually the frontal part of the subducting Indian plate containing the Precambrian continental crust, thrust back on to itself. The MCT and other thrusts towards the foreland are splay faults from this detachment, which reaches a mid-crustal depth of about 30km under the High and Tethys Himalayas (figure 3(a)). Therefore, a tectonic model with a horizontal thrust at a depth of 30km is a good approximation for the purpose of calculating the transient geotherms. With this geometry, the temperature at any given depth and time can be calculated using

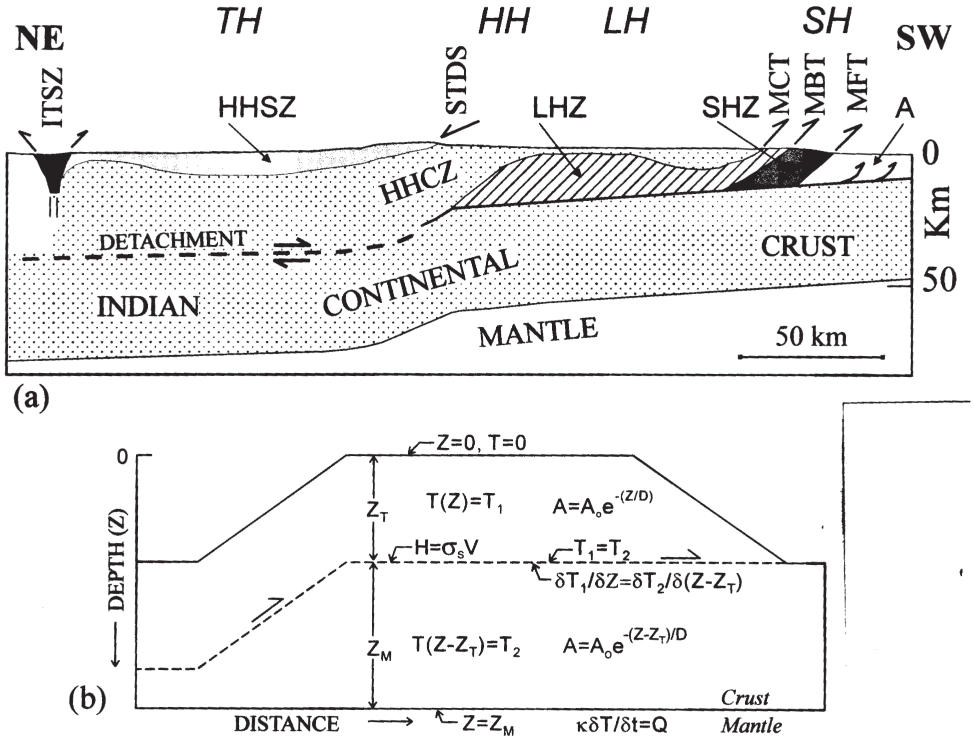

Figure 3. **(a)** Generalized north-south cross section of the Himalayan orogenic belt (adapted from Seeber *et al* 1981; Hirn *et al* 1984; Ni and Barazangi 1984). Physiographic zones: **TH**: Tethys Himalayas; **HH**: High Himalayas; **LH**: Lower Himalayas; **SH**: Sub Himalayas. Other abbreviations are as in figure 1. The ITSZ marks the site of a 50–65 Ma collision between the northern Eurasian plate and the southern Indian plate. The 35–40 km thick Indian continental crust is doubled up under the Tethys Himalayas due to overthrusting along a gently dipping detachment that reaches a mid-crustal depth of about 30 km below the High Himalayas. Leucogranite plutons (not shown) occur close to the STDS or within the HHSZ. **(b)** Thrusting model and boundary conditions used for calculating transient geotherms. Symbols are as in the table 1. Note that the upper flat corresponds with the detachment in (a).

a one-dimensional time-dependent heat conduction equation for a vertically moving medium (Carslaw and Jeager 1959),

$$\frac{\partial T(z,t)}{\partial t} = \kappa \frac{\partial^2 T(z,t)}{\partial z^2} - v_z \frac{\partial T(z,t)}{\partial z} + \frac{H(z)}{\rho c_p} \tag{1}$$

where the symbols are as in table 1. The solutions to the heat-conduction equation involving second-order partial derivatives can be obtained using the method of separation of variables, with boundary conditions as shown figure 3(b). The final equations that give the solutions to equation (1) have been adopted from Molnar *et al* (1983). This relatively simple thermal model (figure 3(b)) is adequate to show how geotherm varies with time because thrusting velocity is much faster than the rate of conductive heat transfer in rocks (Oxburgh and Turcotte 1974). Since the leucogranite plutons are present in the Higher Himalayas where the detachment is nearly flat, mathematically more involved heat conduction equations (e.g. Shi and Wang 1987) appropriate for complex tectonic set up are possibly not required. In any case, detailed thermal modeling is beyond the scope of this paper.

There are three mechanisms by which the crust can be heated up in order to initiate partial melting, viz., conductive heating, advective heating and heating from internal source(s) represented by the three terms on the right-hand side of equation (1), respectively. The two universally important heat sources in the crust, i.e., conductive heat flux from the mantle and radioactive heating within the crust have been considered to calculate the initial steady-state geotherm. The other forms of heating that act as a modifier in restricted areas/tectonic setup have been ignored for model calculations. Since much of the Himalayan crust is actually the frontal part of the Precambrian Indian crust, the heat source and other physical parameters appropriate for Precambrian shield areas (Lachenbruch and Sass 1977; Sclatter *et al* 1980) have been chosen (table 1). The radioactive heat generation in surface rocks (A_o) in shield areas varies between 3 and 6 hgu and, therefore, steady-state geotherms have been calculated for these two extreme values.

Table 1. *Symbols and preferred values of physical parameters used for calculating geotherms.*

Symbol	Parameter	Preferred value
A	Radioactive heat generation	
A_0	Radioactive heat generation in surface rock	3 and 6 hgu (hgu = 10^{-13} cal/cm^3 sec)
D	Characteristic depth for radioactivity	15 km
H	Heat generation pet unit volume	
k	Thermal diffusivity	0.01 cm^2/sec
K	Thermal conductivty	0.005 cal/°C sec cm
T	Temperature, °C	
Q	Vertical conductive heat flux	0.6 hfu (hfu = 10^{-6} cal/cm^2 sec)
v	Thrusting velocity	18 mm/year
Z	Depth (increasing downward)	
Z_M	Depth of mantle	80 km
Z_T	Depth of thrust plane	30 km
σ_S	Shear stress on fault	300 bars

3. Transient geotherms

As the rocks are bad conductors of heat, the advection of heat due to overthrusting of the hot hangingwall block will result in perturbation of the steady-state geotherm. In figures 4 and 5, the curves labeled "0 m.y." are the initial (i.e., perturbed) geotherms immediately after thrusting. The 'saw-tooth' shape of the curves is a consequence of the model chosen, i.e., a fast thrusting velocity as compared to the rate of heat conduction. The heat will conduct from the hot hangingwall block to the colder footwall block and the perturbation of the geotherm will decay with time. Consequently, the geotherms will remain strongly transient for a few tens of million years after thrusting. The transient geotherms have been calculated to 50 m.y. after thrusting since the terminal collision occurred at about 50 Ma. It is to be noted that the time of the formation of the detachment is not known. There are suggestions that the MCT, which is a splay fault from the detachment, was active at about 20Ma (Hubbard and Harrison 1989) and therefore, the detachment may have formed sometime between 50 and 20Ma. The 'present-day geotherm' in this simple model should be in between the curves labeled 20 and 50 m.y.

With normal heat sources, i.e. radioactive heating and mantle heat flux, partial melting could take place only in the deepest part of the Himalayan crust (figure 4). In order to trigger partial melting at mid-crustal depth additional heat source(s) would be required. There are several possible heat sources, such as mantle delamination (Bird 1978), heat-focussing by HHSZ (Jaupart and Provost 1985; Pinet and Jaupart 1987), shear heating at detachment (Molnar *et al* 1983), abnormally high concentration of radioactive elements in the crust, advective heating through circulation of hot fluids, magma emplacement or exothermic metamorphic reactions. All of these additional heat sources can potentially shift the transient geotherms to the higher temperature side. A combination of these heat sources together with the normal heat sources and the age of the detachment should determine whether or not partial melting could be triggered at a particular point in the crust at a specific time after collision. Given the number of variables involved even for a relatively simple one-dimensional heat-conduction equation together with uncertainties in each of the variables, an exact solution to the problem of source region for partial melting in the Himalayas would be difficult, if not impossible, to obtain. It is also possible that different heat sources or a combination of heat sources are important at different places. However, for the purpose of demonstration, the shear heating on the detachment has been considered because, if important, this should affect the entire orogen. The amount of shear heating generated on a thrust depends on the shear stress and thrusting velocity which have been taken as 300 bars (Chander and Gahalaut 1995) and 18 mm/year (Molnar 1987) respectively. Figure 5 shows how transient geotherms can be modified by the addition of shear heating. The important point to note here is that irrespective of the thermal model adopted and the values of the physical parameters chosen, the geotherms would always be strongly transient for the first few tens of millions of years after overthrusting (figures 4 and 5).

4. Partial melting

Partial melting in the crust leading to the generation of granitic melt depends on several factors but an important requirement is the availability of 'fertile' rocks. The dominant rock types in the HHCZ are low-to medium-grade pelitic schists and granite gneisses; both these rock types are potential fertile rocks for the generation of granitic magma. In these rocks both vapour-phase present and vapour-phase absent melting are possible. The second important

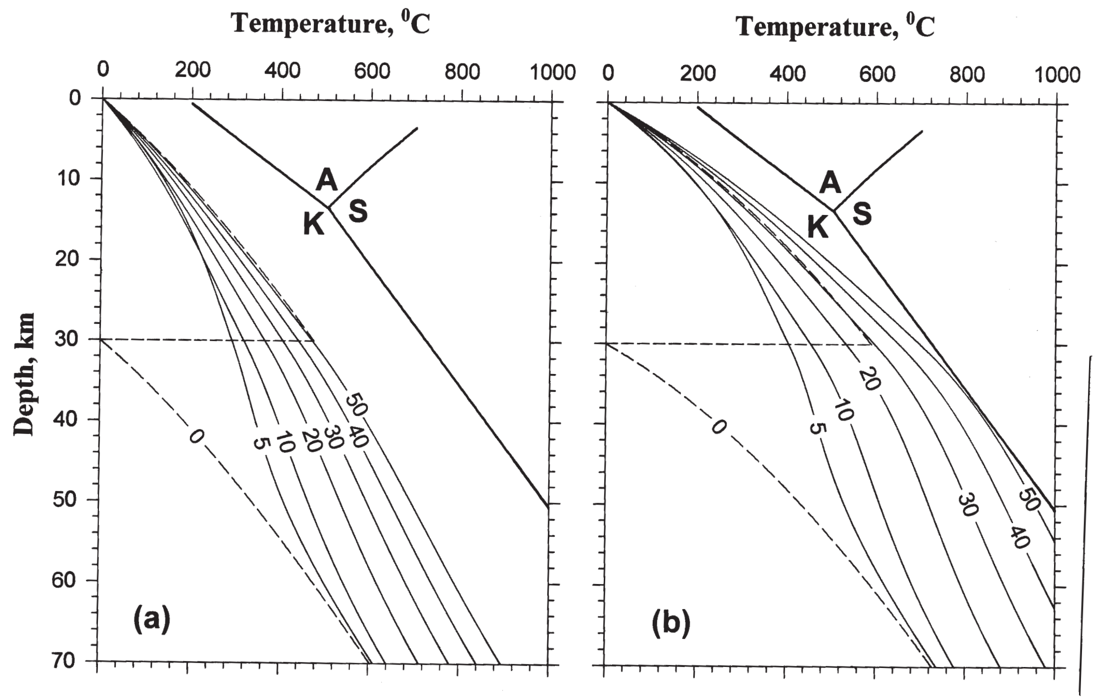

Figure 4. Calculated transient geotherms (depth-temperature profiles) for A_0 values of 3 hgu **(a)** and 6 hgu **(b)** and without any additional heat source. Transient geotherms (solid lines) are labeled with time in m.y. after thrusting. The dashed lines (labeled "0" m.y.) are post-thrusting initial geotherms. The geotherms are regression lines through data points calculated at depth interval of 1 km. The kyanite (K)-sillimanite (S)-andalusite (A) phase diagram (Holdaway and Mukhopadhyay 1993) is for reference.

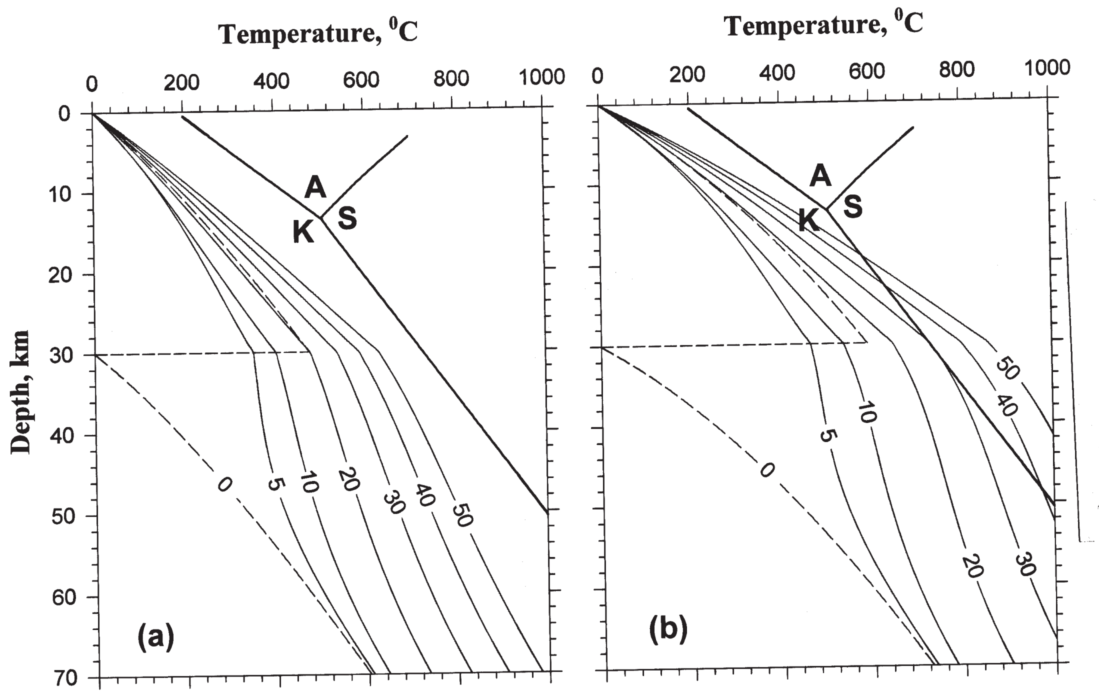

Figure 5. Same as figure 4 but with shear heating added to the total heat budget. **(a)** $A_0 = 3$ hgu. **(b)** $A_0 = 6$ hgu.

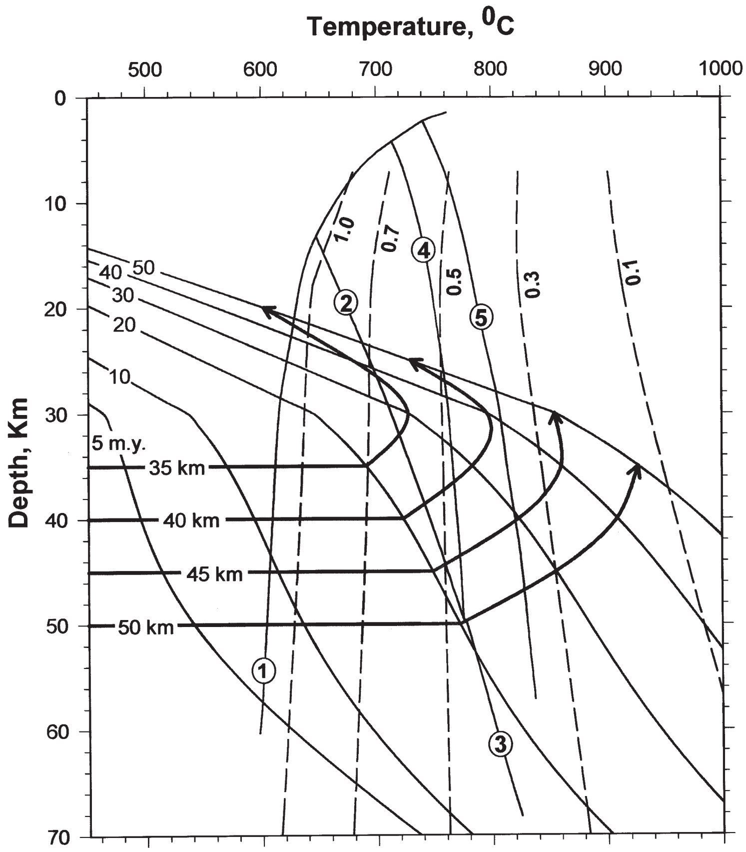

Figure 6. The relationships between transient geotherms (taken from figure 5(b)) and some of the solidus/melting curves appropriate for granitic and pelitic rocks. The dashed curves are minimum granite solidus for different values of activity of H_2O, as labeled (data extrapolated from Ebadi and Johannes 1991). Curves labeled 1–5 are (Spear 1993): **1**: H_2O-saturated pelite solidus; **2**: muscovite + plagioclase + quartz = aluminosilicate + K-feldspar + melt; **3**: muscovite + plagioclase + quartz = biotite + aluminosilicate + K-feldspar + melt; **4**: biotite + aluminosilicate + plagioclase + quartz = garnet + K-feldspar + melt; **5**: biotite + plagioclase + quartz = orthopyroxene + K-feldspar + melt. Heavy grey lines are P-T-t paths for rocks at initial depths of 35, 40, 45 and 50 km. The P-T-t paths are calculated assuming no erosion for the first 20 m.y. followed by long-term average erosion rate of 0.5 mm/year (cf. Thompson 1981).

requirement is that the fertile rocks have to be heated to a temperature above the solidus/melting curve. This depends on the geothermal gradient which in the present case is not in steady state.

The relationships between some of the possible solidus/melting curves together with transient geotherms (taken from figure 5(b)) are shown in figure 6. Let us first consider the relation between H_2O-saturated solidus curve for granite (dashed curves in figure 6). The transient geotherms intersect the solidus curve at different depths. Therefore, as transient geotherms sweep across, partial melting should take place at different depths at different times but the minimum depth of partial melting will in general migrate to a shallow depth with the progress of time. It is unlikely, however, that at deep crustal levels the Precambrian crystalline rocks would be pervasively saturated with H_2O since rocks are not likely to have significant porosity (for a discussion see Spear 1993). The solidus curve will be displaced to the higher temperature side by variable amounts depending on the a_{H_2O}. Also it is reasonable to assume that at different places in the crust (both along and across depth axis) different a_{H_2O}-dependent solidus curve would be appropriate. This should lead to partial melting at different places in the crust at different times. Whatever little H_2O is present in pores and along grain boundaries will help to produce a melt but the amount of melt produced will be small. As melt formed, the H_2O would preferentially get into the melt phase depleting the rocks of H_2O and the solidus curve would move to the higher temperature side leading to cessation of melting. The H_2O-saturated solidus curve for pelitic rocks (curve 1 in figure 6) also gets displaced depending on the a_{H_2O} as in the case with granite. Similarly, the fluid-absent muscovite and biotite breakdown curves leading to the formation of a melt (curves 2–5 in figure 6) are also affected by displaced equilibria. Therefore, the melting relations as discussed for granite are also applicable for pelitic rocks.

Further, in a tectonically overthickened crust a rock occurring at a particular depth is progressively unroofed due to erosion. It is well known (e.g. England and Thompson 1984) that for the situation described here the P-T-t paths for rocks originally at different depths will be counterclockwise in temperature-depth space. The P-T-t paths have been calculated assuming no erosion for the first 20 m.y. (cf. Thompson 1981) followed by long term average erosion rate of 0.5 mm/year (figure 6). The P-T-t path of a particular rock intersects different melting curves at different times (figure 6). Therefore, consideration of P-T-t paths also suggest that partial melting should take place at different depths at different times.

From the above description two important points emerge. Firstly, transient geotherms, P-T-t paths and variably displaced solidus curves should lead to partial melting at different depths at different times. In this scenario, partial melting can affect a large part of the crust and need not be confined to a particular stratigraphic horizon or tectonic plane (e.g. Le Fort *et al* 1987). This conclusion is perfectly general because the overall geometric relation between the transient geotherms and the solidus curves will remain similar although both of them can vary from place to place due to variation in modal mineralogy and values of heat flow parameters. Secondly, since the source region is made up of Precambrian crystalline rocks it should not be pervasively saturated with water nor should the activity of H_2O be held constant. As melt formed, the H_2O would preferentially partition into the melt phase depleting the rocks of H_2O and the solidus curve would move to the higher temperature side leading to cessation of melting for no substantial increase in T (or isothermal decrease of P). Consequently, as the geotherm sweeps across, the total amount of melt at a particular depth would be small and highly leucocratic with unique Sr isotopic ratio. The small pods of magma formed at different places may coalesce to form a larger magma body. As the granitic magma is very viscous, homogenization of the coalesced magma body is likely to be inhibited. Consequently, the small magma pods within the larger coalesced magma body may retain their individual Sr-isotope characteristic.

If the coalesced magma pods had to retain their unique Sr isotopic characters, then magma ascent and emplacement must have been rapid. In an experimental study on the viscosity of Himalayan leucogranites, Scaillet *et al* (1996) have shown that rates of magma flow was very fast and dominantly laminar which suggest that any chemical heterogeneity inherited from the source could have survived from aggregation to the emplacement stage. The aggregation, segregation, ascent and emplacement of granitic magma are complex processes. The buoyancy (i.e. gravity)-induced diapiric rise of small granitic magma pods should be agonizingly slow; the small magma pods should freeze not far from the source forming migmatite rather than ascent to the higher crustal level (e.g. Clemens and Mawer 1992). It is now increasingly recognized that deformation plays an important role in the extraction, ascent and emplacement of granitic magma (see review by Brown 1994 and references therein). The terminal collision between India and Asia took place at about 50Ma (e.g. Klootwijk *et al* 1992). However, from the time of terminal collision to the present day, the Himalayan rocks are being continuously deformed due to continued convergence and crustal shortening. The HHCZ, the source for partial melting, is highly deformed with extensive development of foliation surfaces along which deformation-induced magma migration could take place. A large number of leucogranite plutons occur in the close vicinity of the STDS which is major tectonic plane. Further, leucogranite plutons in the Himalayas are usually underlain by a system of feeder dikes (e.g., Le Fort 1981; Scaillet *et al* 1995, 1996). These lines of evidence suggest that deformation-induced magma transport and emplacement is the appropriate mechanism for the Himalayan leucogranite.

Figures 4 and 5 also show that the transient geotherms can overshoot the initial steady-state geotherms depending on the amount of extra heat sources. Consequently, partial melting and transfer of magma to shallow depths are probably continuing to the present time. The measured high heat flow in the Himalayan region (Francheteau *et al* 1984; Shankar 1989) and very young (2–7 Ma) ages of some of the leucogranitic dikes (e.g. Zeitler and Chamberlain 1991) support this contention.

5. Discussion and conclusions

The model of partial melting for the Himalayan leucogranites outlined in this paper depends on the geometric relationship between the transient geotherms in overthickened crust due to thrusting and variably displaced solidus/melting curves. As the geotherm sweeps across, partial melting leading to generation of granitic melt can take place at different depths at different times. The amount of melt at any point in the crust is expected to be small and leucocratic. The small pods of magma thus generated will have its own unique initial Sr isotope ratio. The deformation-induced fast segregation, ascent and eventual emplacement along feeder dikes can lead to the formation of small plutonic bodies. Each of the magma pods may retain its original Sr isotopic signature leading to pluton-wide heterogeneity in Sr isotopic ratios. It is emphasized that the proposed model of partial melting should be considered as one of the contributing factors rather than only means of inducing pluton-scale Rb-Sr isotopic heterogeneity.

The upper age limit for these granites is about 50 Ma which is the time of the terminal collision. Within this short time the initial Sr isotope ratio will not change very significantly (Faure 1986). Consequently, the Sr isotope ratios measured in each sample should be very close to the initial Sr ratio. Therefore, any colinearity of data points in an Rb-Sr diagram should be viewed with extreme caution.

Acknowledgements

I am grateful to the journal reviewers for critical and constructive suggestions that led to significant improvement of the manuscript. Thanks are due to Dr. P K Gupta for help with the mathematical formulations and to the CSIR for financial support.

References

Bird P 1978 Initiation of intracontinental subduction in the Himalaya; *J. Geophys. Res.* **83** 4975–4987

Brown M 1994 The generation, segregation and emplacement of granite magma: the migmatite-to-crustally-derived granite connection in thickned orogens; *Earth Sci. Rev.* **36** 83–130

Carslaw H S and Jeager J C 1959 *Conduction of Heat in Solids*; Oxford University Press, 510p.

Chander R and Gahalaut V K 1995 A simulation of upper crustal stresses for great and moderate thrust earthquakes of the Himalaya; *Proc. Indian Acad. Sci. (Earth Planet. Sci.)* **104** 115–129

Clemens J D and Mawer C K 1992 Granitic magma transport by fracture propagation; *Tectonophysics* **204** 339–360

Deniel C, Vidal P, Fernandez A, Le Fort P, and Peucat J-J 1987 Isotopic study of the Manaslu granite (Himalaya, Nepal): Inferences on the age and source of Himalayan Leucogranites; *Contrib. Mineral. Petrol.* **96** 78–92

Dietrich V and Gansser A 1981 The leucogranites of the Bhutan Himalaya (crustal anatexis versus mantle melting); *Schweiz. Mineral. Petrogr. Mitt.* **61** 177–202

Ebadi A and Johannes W 1991 Beginning of melting and composition of first melts in the system Qz-Ab-Or-H_2O-CO_2; *Contrib. Mineral. Petrol.* **106** 286–295

England P C and Molnar P 1993 The interpretation of inverted metamorphic isograds using simple physical calculations; *Tectonics* **12** 145–157

England P C and Thompson A B 1984 Pressure-temperature-time paths of regional metamorphism of the continental crust, I, Heat transfer during the evolution of regions of thickened continental crust; *J. Petrol.* **25** 894–928

England P, Le Fort P, Molnar P and Pecher A 1992 Heat sources for tertiary metamorphism and anatexis in the Annapurna-Manaslu region, central Nepal; *J. Geophys. Res.* **97** 2107–2128

Francheteau J, Jaupart C, Jie S X, Wen-Hua K, De-Lu L, Jia-Chi B, Hung-Pin W and Hsia-Yeu D 1984 High heat flow in southern Tibet; *Nature* **307** 32–36

Faure G 1986 *Principles of Isotope Geology*; (John Wiley & Sons)

Hirn A, Lepine J-C, Jobert G, Sapin M, Wittlinger G, Xin X Z, Yuan G E, Jing W X, Wen T J, Bai X S, Pandey M R and Tater J M 1984 Crustal structure and variability of the Himalayan border of Tibet; *Nature* **307** 23–25

Holdaway M J and Mukhopadhyay B 1993 A reevaluation of the stability relations of andalusite: Thermochemical data and phase diagram for aluminium silicates. *Amer. Min.* **78** 298–315

Hubbard M and Harrison T M 1989 ^{40}Ar/^{39}Ar age constraints in deformation and metamorphism in the Main Central Thrust Zone and Tibetan Slab, Eastern Nepal Himalaya; *Tectonics* **8** 865–880

Jaupart C and Provost A 1985 Heat focussing, granite genesis and inverted metamorphic gradients in continental collision zones; *Earth Planet. Sci. Lett.* **73** 385–397

Klootwijk C T, Gee J S, Peirce J W, Smith G M and McFadden P L 1992 An early India-Asia contact: paleomagnetic constraints from Ninetyeast Ridge, ODP Leg 121; *Geology* **20** 395–398

Lachenbruch A H and Saas J H 1977 Heat flow in the United States and the thermal regime of the crust; In *The Earth's Crust* (ed.) J G Heacock (AGU, Washington: *Geophys. Mon. Sr.* 20) pp. 626–675

Le Fort P 1981 Manaslu leucogranite: a collision signature of the Himalaya. A model for its genesis and emplacement; *J. Geophys. Res.* **86** 10545–10568

Le Fort P, Cuney M, Deniel C, France-Lanord C, Sheppard S M F, Upreti B N and Vidal P 1987 Crustal generation of Himalayan leucogranites; *Tectonophysics* **134** 39–57

Molnar P 1987 Inversion of profiles of uplift rates for the geometry of dip-slip faults at depth, with examples from Alps and Himalaya; *Annales Geophysicae* **5B(6)** 663–670

Molnar P, Chen W-P and Padovani E 1983 Calculated temperatures in overthrust terranes and possible combination of heat sources responsible for the Tertiary granites in the greater Himalaya; *J. Geophys. Res.* **88** 6415–6429

Molnar P and England P 1990 Temperatures, heat flux, and frictional stress near major thrust faults; *J. Geophys. Res.* **95** 4833–4856

Ni J and Barazangi M 1984 Seismotectonics of the Himalayan collision zone: geometry of the underthrusting beneath the Himalaya; *J. Geophys. Res.* **89** 1147–1163

Oxburgh E R and Turcotte D 1974 Thermal gradients and regional metamorphism in overthrust terrains with special reference to eastern Alps; *Schweiz. Mineral. Petrogr. Mitt.* **54** 641–662

Pinet C and Jaupart C 1987 A thermal model for the distribution in space and time of the Himalayan granites; *Earth Planet. Sci. Lett.* **84** 87–99

Scaillet B, Pecher A, Rochette P and Champenois M 1995 The Gangotri granite (Garhwal Himalaya): laccolithic emplacement in an extending collisional belt; *J. Geophys. Res.* **100** 585–607

Scaillet B, Holtz F, Pichavant and Schmidt M 1996 Viscosity of Himalayan leucogranites: Implications for mechanisms of granitic magma ascent; *J. Geophys. Res.* **101** 27691–27699

Sclatter J G, Jaupart C and Galson D 1980 The heat flow through oceanic and continental crust and heat loss of the earth; *Rev. Geophys. Space Phys.* **18** 269–311

Seeber L, Ambruster J G and Quittmeyer R C 1981 Seismicity and continental subduction in the Himalayan arc; In *Zagros, Hindu-Kush, Himalaya Geodynamic Evolution* (eds.) H K Gupta and F M Delany (AGU, Washington: Geodynamic Sr. 3) pp. 215–242

Shanker R 1989 Heat flow regime in Himalaya; *Geol. Surv. India Sp. Pubn.* **26** 181–192

Shi Y and Wang C-Y 1987 Two-dimensional modeling of the P-T-t paths of regional metamorphism in simple overthrust terrains; *Geology* **15** 1048–1051

Spear F S 1993 Metamorphic phase equilibria and pressure-temperature-time paths; *Min. Soc. Amer. Monograph* 799p

Stern C R, Kligfield R, Schelling D, Virdi N S, Futa K and Peterman Z E 1989 The Bhagirathi leucogranite of High Himalaya (Garhwal, India): age, petrogenesis, and tectonic implications; *Geol. Soc. Am. Sp. Paper* **232** 33–45

Thompson A B 1981 The pressure-temperature (P,T) plane viewed by geophysicists and petrologists; *Terra Cognita* **1** 11–20

Zeitler P K and Chamberlain C P 1991 Petrogenetic and tectonic significance of young leucogranites from the northwestern Himalaya, Pakistan; *Tectonics* **10** 729–741

MS received 25 March 1998; revised 10 May 2000