PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Granites in the Tectonic Evolution of the Himalaya, Karakoram and Southern Tibet [and Discussion]

P. Le Fort and V. S. Cronin

Phil. Trans. R. Soc. Lond. A 1988 **326**, 281-299
doi: 10.1098/rsta.1988.0088

Phil. Trans. R. Soc. Lond. A **326**, 281–299 (1988) [281]
Printed in Great Britain

Granites in the tectonic evolution of the Himalaya, Karakoram and southern Tibet

By P. Le Fort†

Centre de Recherches Pétrographiques et Géochimiques, B.P. 20, 54501 *Vandoeuvre-lès-Nancy, France*

Four major plutonic belts are related to the Meso-Cainozoic orogenic evolution of the Himalaya–Transhimalaya–Karakoram realm: the Transhimalaya belt and its satellite Kohistan arc, the Karakoram batholith, the High Himalaya belt and the North Himalaya belt. A fifth one results from the lower Palaeozoic epirogenic events: the 'Lesser Himalaya' belt. The tectonic settings of their production and emplacement are successively reviewed. Among the first four, two result from oceanic subduction along an Andean margin locally branching into an island arc and two result from intracontinental subduction after closure of the oceanic realm. Both Andean belts are made up of very large quantities of highly diversified granitoids produced more or less continuously during 70 Ma at least, whereas the intracontinental ones are limited to a small volume of very uniform anatectic granite produced during a 10–15 Ma period. The production and emplacement in the Andean belts is partly controlled by the obliquity of the convergence between India and Eurasia. The emplacement of the intracontinental belts is even more dependent on the regional tectonic setting. These contrasting belts are case studies probing the depths and mechanisms of their production and giving adequate models for older geodynamic frames.

0. Introduction

Five large plutonic belts occur in the Karakoram–Transhimalaya–Himalaya region (Debon *et al.* 1981) (see figure 1).

1. The Karakoram belt extending for some 800 km to the north of the western syntaxis of the Himalaya and made up of a large variety of calc-alkaline to sub-alkaline quartz diorite to granite.

2. The Transhimalaya belt lying just north of the Indus–Tsangpo Suture Zone (ITS) and forming a nearly continuous 2500 km long batholith made up of gabbros to granites. To the west lies the large Kohistan Zone, comprising a wide variety of mainly mafic plutonic and volcanic rocks.

3. The North Himalaya belt running some 50 km south of the suture and forming a series of some 20 domes of two-mica adamellite.

4. The High Himalaya belt intruding the High Himalayan sedimentary series, between 100 and 150 km south of the suture and forming a series of a dozen or so main lenticular slabs and innumerable dykes made up of muscovite, biotite and/or tourmaline leucoadamellite (Le Fort *et al.* 1987).

5. The 'Lesser Himalaya' belt lying mostly in the southern part of the Himalaya but extending widely over the entire Himalaya and south Tibet realms, forming discontinous masses of both gneissic and non-gneissic granites (Le Fort *et al.* 1986).

All first four belts are related to the Mesozoic–Cainozoic evolution of the India and Eurasia

---

† Present address: Institut Dolomieu, 15 rue Maurice Gignoux, 38031, Grenoble, France.

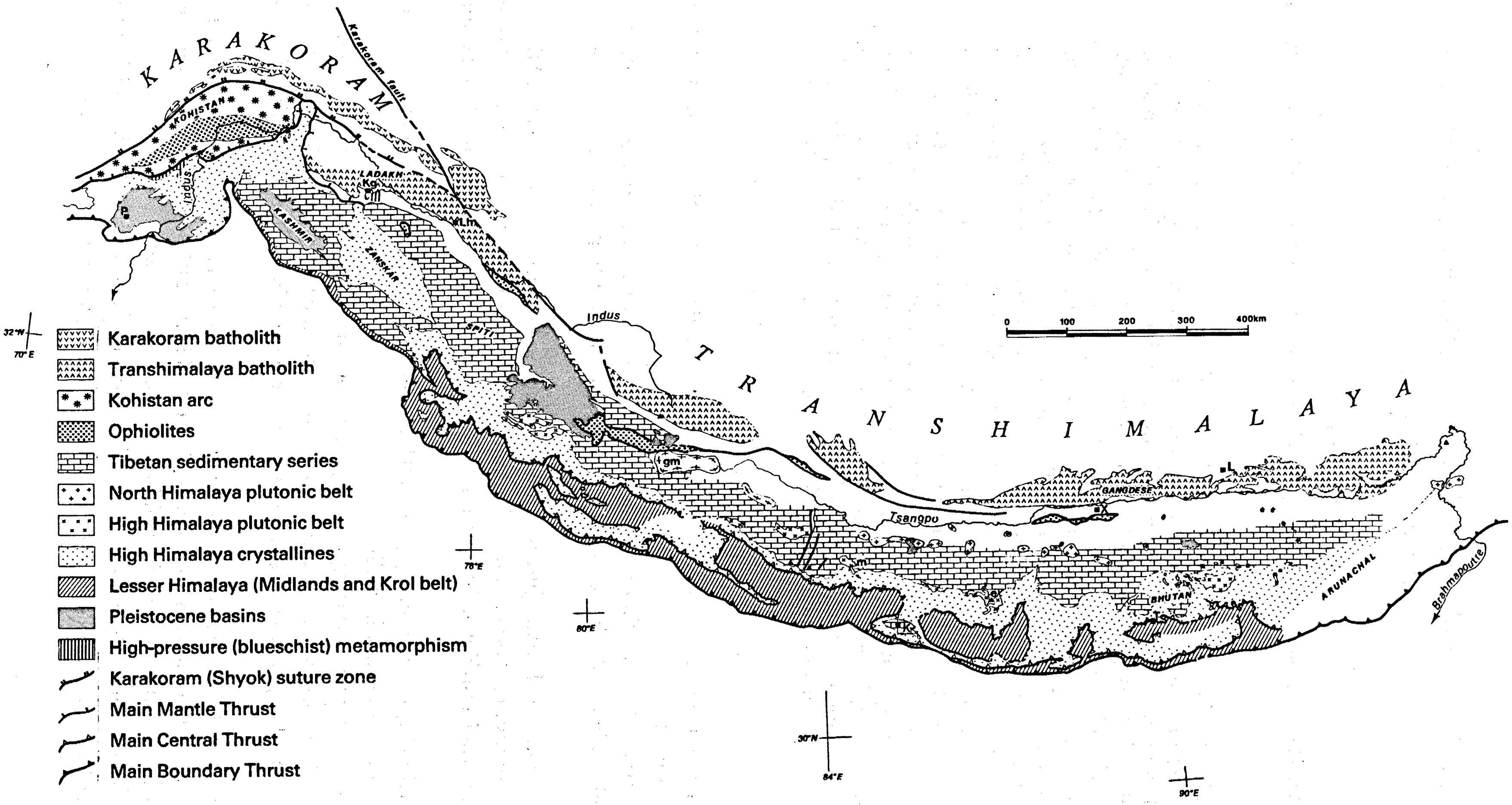

FIGURE 1. Main structural divisions and Meso–Cainozoic plutonic belts of the Himalaya–Transhimalaya–Karakoram. Based on Desio (1964), Gansser (1977), Gamerith (1979), Bard *et al.* (1980), Stöcklin & Bhattarai (1980), Academia Sinica (1981), Bordet *et al.* (1981), Fuchs (1981), Polino (1981), Valdiya (1981), Gansser (1983), Colchen *et al.* (1986*b*) and Le Fort (1986). The zone shown as ophiolitic in Kohistan corresponds to the lowermost part of the island arc. 'Lesser Himalaya' granites and Subhimalaya series are not represented; e = Everest, gm = Gurla Mandata, K = Kathmandu, Kg = Kargil, L = Lhasa, Lh = Leh, m = Manaslu, P = Peshawar, T = Thimpu, X = Xigaze.

Plates. Only the fifth one of Lower Palaeozoic age is not directly connected to the system as its granites had long ago crystallized and cooled down. However, their presence modifies the rheological properties of the Himalayan deforming crust and in some cases they are probably responsible for the shape of the Himalayan structures.

Let us review the major characteristics of the five belts before trying to relate more precisely their generation and emplacement to the tectonic framework, and to evaluate their bearing on the evolution of the Himalayan Orogeny *s.l.*

1. Geological setting and description

Transhimalaya batholithic belt

The Transhimalaya batholith, nearly continuous for some 2500 km, with a width of about 50 km, has been studied mainly along two segments: some 200 km in Ladakh (Honegger *et al.* 1982; Sharma & Choubey 1983), and another 200 km in southern Xizang south of Lhasa (Academia Sinica 1980; Tu *et al.* 1981; Debon *et al.* 1982, 1986*a*), where it is also known as the Gangdese (or Kangdese) batholith. Characteristics are broadly similar in both segments. As summarized by Debon *et al.* (1986*a*), the batholith is composite, made up of numerous plutonic bodies often continuous with gradational contacts. Postmagmatic cleavage is often superimposed on flow structures. Compositions range from noritic gabbro to adamellite through quartz monzonite and granodiorite. Contrary to rather common statements, true tonalite has not been described. Most rocks are typically metaluminous and subalkaline but not calcalkaline as mostly considered (figure 2). It has been known to intrude Mesozoic rocks since Hayden (1905) described it intruding Jurassic and Cretaceous series.

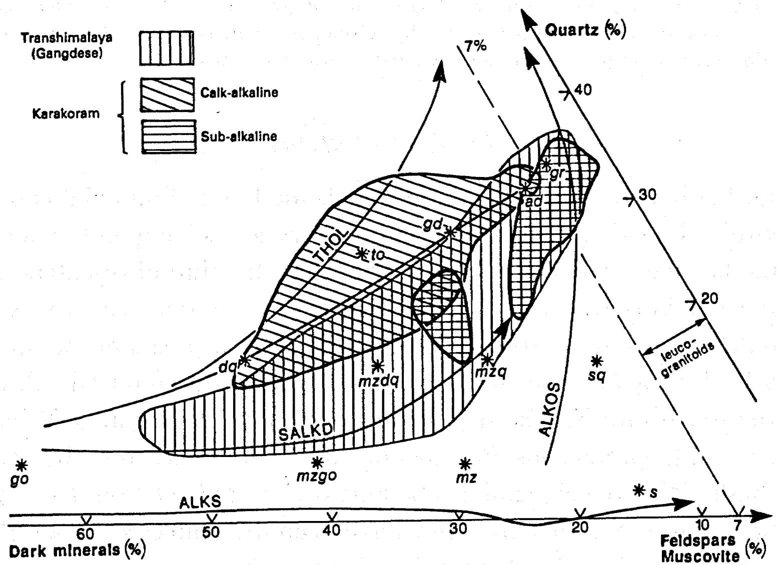

Figure 2. Distribution of the Transhimalaya and Karakoram plutonic belts in the triangular quartz–dark minerals–feldspars + muscovite diagram ('Q–B–F' diagram) (from Debon *et al.* 1986*a*, 1987). The parameters in percentage (by mass) are directly calculated from chemical analysis (La Roche 1964; Debon & Le Fort 1982). Different subtypes among the three main types of cafemic and alumino–cafemic associations are distinguished: THOL (tholeiitic), CALK (calc-alkaline), SALKD (dark coloured sub-alkaline, i.e. monzonitic), ALKS and ALKOS (alkaline saturated and oversaturated, respectively). Representative points of the different petrographic types are also shown. The 120 analyses of the Transhimalaya (Gangdese region) fall in a typical sub-alkaline field. For the 60 analyses from Karakoram, part of them are calc-alkaline while others fall in the same sub-alkaline field.

Age determinations from various methods mostly range from 120 to 40 Ma (Lower Cretaceous to Upper Eocene). There is no obvious quiescence in the magmatic activity at the scale of the entire range although, locally, ages generally cluster around one or two values.

Initial ratios of $^{87}Sr/^{86}Sr$ are moderately low and present a general tendency to increase from west to east (from 0.704 to 0.707) and when available, from south to north (figure 3). This can be related to an increasing contribution of continental crust toward the east during the subduction of the Tethys ocean floor.

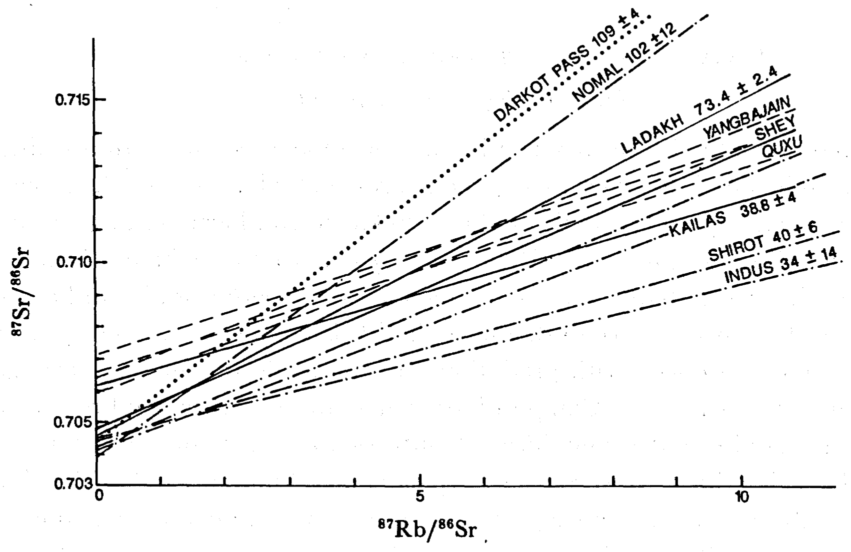

FIGURE 3. Whole-rock Rb–Sr isochrons for some plutonic (and volcanic) bodies of the Karakoram, Kohistan and Transhimalaya showing a general increase in $^{87}Sr/^{86}Sr$ initial ratio from west to east. Lines: dots, Karakoram (Debon *et al.* 1987); broken lines and dots, Kohistan (Petterson & Windley 1985; Debon *et al.* 1987); continuous lines, western Transhimalaya (Ladakh) (Honegger *et al.* 1982; Schärer *et al.* 1984); broken lines, eastern Transhimalaya (Gangdese) (Debon *et al.* 1982; Xu *et al.* 1985).

Karakoram batholith

In Karakoram, the backbone of the main range is made up of an axial composite batholith intruding Palaeozoic–Triassic (Desio 1964) sedimentary series lying on the northern side of the range. During the last few years it has been shown that the plutonism extended as far back as middle Cretaceous (Le Fort *et al.* 1983*b*; Debon *et al.* 1987) and as recently as Upper Miocene (Debon *et al.* 1986*b*; Searle *et al.* 1988). In between, several pulses of magmatism have been dated, particularly during Palaeocene and Eocene, with light-coloured sub-alkaline plutons that extend to the south into Kohistan (Debon *et al.* 1987; Petterson & Windley 1985).

The chemistry and Cretaceous–Palaeogene ages obtained on the Karakoram and Transhimalaya batholiths are very similar (Le Fort *et al.* 1983*b*; Debon *et al.* 1987) and suggest that they both belong to the same belt. They have been disconnected by a right-lateral strike-slip movement of some 300 km along the still-active Karakoram Fault (figure 4). Accordingly, the Shyok ophiolitic zone (or northern suture zone, NSZ, of some authors) is only a branch of the Indus–Tsangpo Suture (Stöcklin 1977; Sengör 1984, 1985), the back-arc suture.

To the southwest, the Karakoram Zone corresponds with the Central Mountains of Afghanistan through the eastern Hindu Kush (Le Fort *et al.* 1983*b*) (figure 4). From the Lhasa Block to the Central Mountains, one can thus follow the same Tethyside continental strip

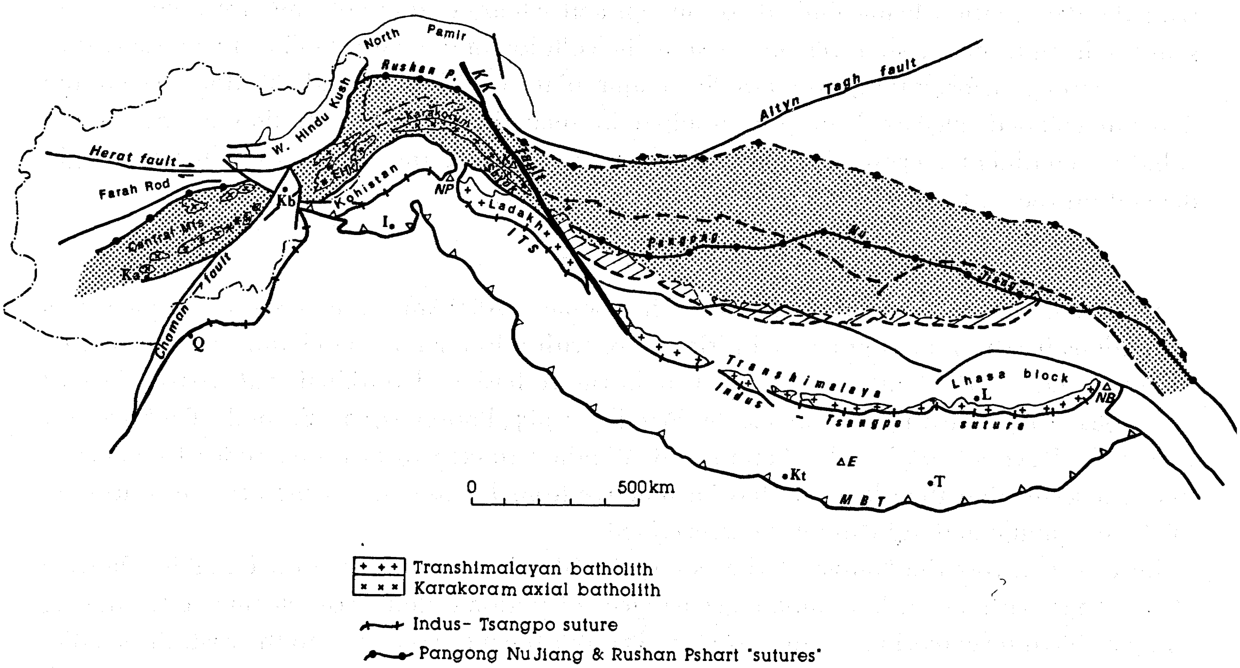

FIGURE 4. Global structural map of Afghanistan, Hindu-Kush, Karakoram, Himalaya and Tibet after various sources. A right-lateral strike-slip movement of some 300 km along the Karakoram Fault has been restituted (broken lines) to illustate the previous geodynamic setting where the Karakoram prolongs the Transhimalaya, the Rushan Pshart prolongs the Pangong NuJiang Sutre Zone and the Shyok Zone is only a branch of the Indus Tsangpo Suture (ITS). The Tethyside south Tibetan (Lhasa) Block, stippled, extends from the Central Mountains of Afghanistan, through Karakoram, to eastern Transhimalaya. To the SE, the Karakoram Fault seems to merge into the ITS. E = Everest, EHK = Eastern Hindu-Kush, I = Islamabad, ITS = Indus Tsangpo Suture, K = K2, Ka = Kandahar, Kb = Kabul, KK = Karakoram, Kt = Kathmandu, L = Lhasa, NB = Namche Barwa, NP = Nanga Parbat, Q = Quetta, T = Thimpu.

through Karakoram. The simple Transhimalayan arc of southern Xizang prolongs westward into the double-arc system of Kohistan–Ladakh to the south and Karakoram to the north.

Magmatically and tectonically, two major differences appear between Karakoram and Transhimalaya belts. In Karakoram the plutonic activity outlasts the Eocene timing of India–Eurasia collision, and continues at least up to Miocene, when the Baltoro pluton and numerous granitic dykes and pods are emplaced (Debon *et al.* 1986*b*; Searle *et al.* 1988), whereas, in Transhimalaya, granitoid production seems to stop rather abruptly some 10 Ma after collision. Actually, upper Miocene volcanic activity has been recently dated in the Kangdese region (Maqiang andesites and acid ignimbrites, 15–10 Ma; Coulon *et al.* 1986), and the presence of magma is now suspected at shallow depth in the same region (Pham *et al.* 1986). Thus the difference between the two batholiths may be more in the rates of uplift and erosion, otherwise known to be very fast in the Karakoram region (Zeitler 1985).

The other difference appears in the deformation pattern of Karakoram when compared to that of the Transhimalaya. In Karakoram, Cretaceous plutons such as the Hunza have been intensely sheared toward the south under high-grade regional metamorphic conditions before the middle Eocene (Le Fort *et al.* 1983*b*; Bertrand & Debon 1986; Coward *et al.* 1986; Debon *et al.* 1987), whereas Transhimalayan plutons remain almost unaffected. Such a deformation may result from an Upper Cretaceous–Palaeocene collision of the Kohistan arc after subduction of the northern Neo-Tethys oceanic crust along the so-called NSZ. This explanation agrees with the deformation being limited to the segment where a large volcanic arc occurs to the south of it. But, the question remains of how the collision of this rather minor mass resulted in such a strong deformation, whereas the major plate collision around 50 Ma affected the Tibetan collided edge so little and resulted in such dramatic deformation of the Indian colliding continental crust. Actually, the characteristics of the deformation may strongly depend on the strike-slip component of the convergence.

Kohistan arc

In this large oval-shaped region covering some $36000\ \text{km}^2$, the observed sequence of formations has been attributed to an island arc with a lower magma chamber succeeded by plutonic and volcanic suites and topped by intra-arc basins (Tahirkheli *et al.* 1979; Coward *et al.* 1982, 1986; Bard 1983; Petterson & Windley 1985; Pudsey 1986). From the few ages now available (Reynolds *et al.* 1983; Petterson & Windley 1985; Coward *et al.* 1986; Debon *et al.* 1987) it seems that the island arc has been active from Jurassic to Eocene but the continuity of the magmatic activity cannot be ascertained.

East of it, across the Nanga–Parbat spur, the island arc continues into Ladakh (Dietrich *et al.* 1983) with somewhat similar occurrences of plutonic and two volcanic suites having comparable time extension (Reuber *et al.* 1987). But the more one goes to the east, the smaller abundance of volcanics one meets. The Dras volcanics, for example, laterally pass into a flysch formation (Nindam flysch of Bassoullet *et al.* 1978 and Colchen *et al.* 1986*a*). Concurrently, as noted above, the $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios present a general tendency to increase (figure 3). These two facts can be related to the increasing contribution of continental crust material in the magmas and to the screen effect of this material and of the water that it contains against the ascent of the magma to the surface.

High Himalaya belt

Numerous studies have been devoted to the High Himalaya granitoid belt during the past 15 years (see Le Fort *et al.* 1987 for an extensive set of references) although they cover only a few percent of the Himalaya and account for not more than 0.5‰ of its volume. The dozen or so main plutonic bodies have a lenticular shape. They are accompanied by a dense network of aplopegmatitic dykes that extends on a much larger area than the plutons themselves and

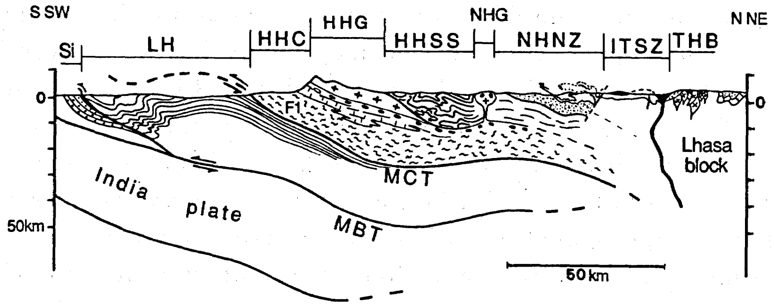

FIGURE 5. Schematic cross section giving the present structure of the Himalaya in Central Nepal, the disposition of the different tectonic units and the localization of three of the plutonic belts discussed here (from France-Lanord & Le Fort 1988). HHC = High Himalaya Crystalline (Tibetan Slab), HHG = High Himalaya leucogranite (Manaslu), HHSS = High Himalaya sedimentary series, ITSZ = Indus-Tsangpo Suture Zone, LH = Lesser Himalaya, MBT = Main Boundary Thrust, MCT = Main Central Thrust, NHG = North Himalaya granite, NHNZ = North Himalaya Nappe Zone, Si = Siwaliks, THB = Transhimalaya batholith.

that often connects them. The granitic rocks occur close to the limit between the crystallines (HHC) and the sedimentary series (HHSS) of the High Himalaya (figure 5). The plutons may intrude the HHSS up to the Cretaceous.

The High Himalaya granites have a very homogeneous composition of muscovite, biotite and/or tourmaline-bearing leucoadamellite reflected in their major element chemical composition. This composition is close to minimum melt composition in the haplogranitic system under addition of B and F, and variable water saturation of the magma as revealed by the slight variations of the Na/K ratio (Le Fort 1981, 1986; Le Fort *et al.* 1987; France-Lanord & Le Fort 1988). Trace element and isotopic characteristics are for most of them inherited from the source rock in the HHC, shown to be represented in Central Nepal by the quartzo-pelitic Formation 1 of the Tibetan Slab (Le Fort 1981; Vidal *et al.* 1982; Cuney *et al.* 1984; Deniel *et al.* 1987; France-Lanord *et al.* 1988, France-Lanord & Le Fort 1988).

From field, petrographical and geochemical studies I have proposed a model of generation of the High Himalaya granites (Le Fort 1975, 1981, 1986; Vidal *et al.* 1982). It associates the major zone of thrusting at continental scale along the Main Central Thrust (MCT), the generation of inverted metamorphism in the underthrusted Midlands formations, the liberation of large quantities of fluids from these formations, and the begetting of anatectic melts in the hot gneisses of the overthrusting HHC. The degree of migmatization lies around 10–12% of the total volume of underlying gneisses as estimated from the projection of their field extension. Thus collection of the melts is a slow process that proceeds in a discontinuous way, and releases multiple successive batches of leucogranitic magmas rising through the overlying crystallines.

The different batches emplace at the limit between the HHC and the HHSS, which corresponds to the disharmonic boundary between the infrastructure and the superstructure. This emplacement occurs at syn- to late-metamorphic and deformation time as evidenced by the variable intensity of deformation taken up by the granite.

Isotopic (cooling) ages by various methods span from 25 Ma by U–Pb on monazite from the Manaslu (Deniel *et al.* 1987) to around 10 Ma by K–Ar on micas from plutons such as in Bhutan (Dietrich & Gansser 1981). The magmatic part of the evolution may have lasted for more than 10 Ma.

The North Himalaya belt

Mentioned as such for the first time by Academia Sinica (1980), the North Himalaya belt groups, in a series of dome, Lower Palaeozoic porphyritic granite and Cainozoic two-mica adamellite (Debon *et al.* 1981, 1985, 1986*a*; Burg *et al.* 1984*b*; Le Fort 1986). The two types of rock may be associated or independently present in the above-mentioned plutonic domes. I have suggested (Le Fort 1986) that these plutons were diapirically emplaced and had carried away in their ascent portions of the cap of porphyritic granite lying above their migmatitic zone of production.

Compared to the HHG, the diapirically emplaced NHG necessitates a much higher degree of melting in the migmatitic production zone, probably several tens percents. Actually, the two-mica adamellite rock is similar to the two-mica leucoadamellite of the High Himalaya in its major and trace element characteristics, heterogeneous Sr isotopic ratios, high Sr initial ratios, Pb isotope compositions and high $\delta^{18}O$ values. Its age seems to be slightly younger than the average age of the High Himalaya plutons, which agrees with its production being linked to the same mechanism but its localization being more to the north in the overthrusted slab.

'Lesser Himalaya' belt

The 'Lesser Himalaya' granitic belt (Le Fort *et al.* 1980, 1983*a*, 1986; Debon *et al.* 1986*a*) is made up of some fifteen independent plutons that appear in the HHC but at a short distance to the north of the Main Boundary Thrust (figure 6). The plutons are generally composed of porphyritic peraluminous quartz-rich granite, rich in dark igneous and metasedimentary inclusions. Isotopic ages have cofirmed the identity of the belt emplaced around the Cambro-Ordovician boundary and showing a high $^{87}Sr/^{86}Sr$ initial ratio.

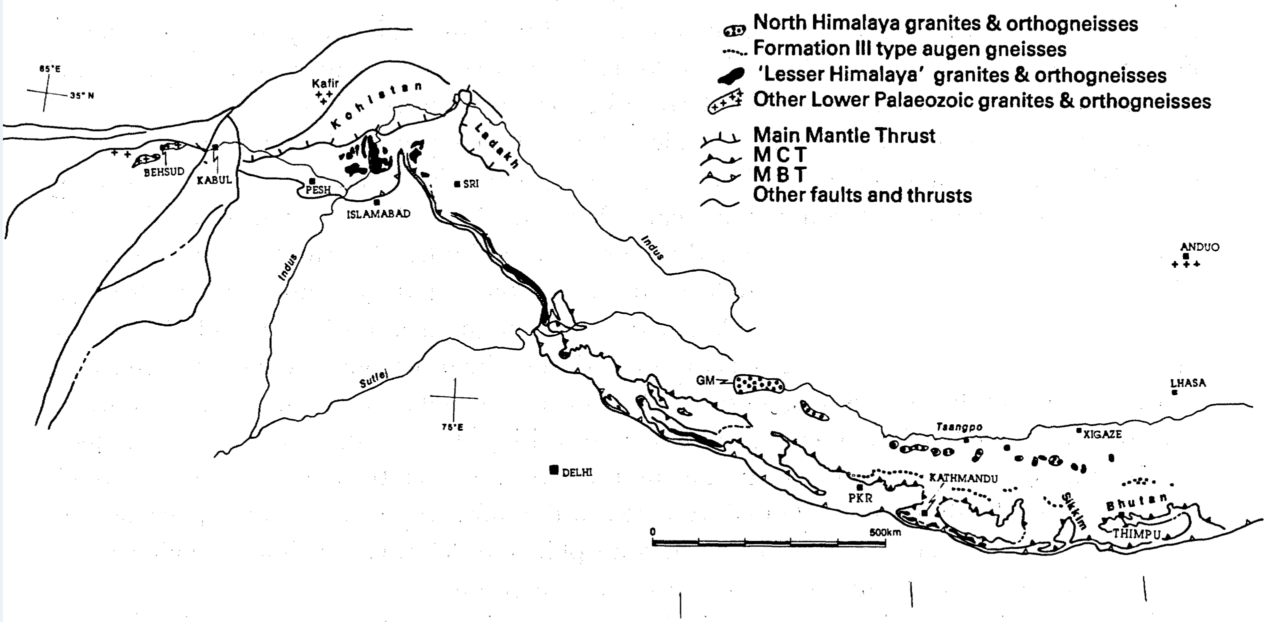

FIGURE 6. Geological sketch map showing the location of the 'Lesser Himalaya' granitic belt and the extension of the magmatic formations with an isotopic age around 500 Ma (from Le Fort *et al.* 1986). GM = Gurla Mandata, PESH = Peshawar, PKR = Pokhra, SRI = Srinagar.

This magmatism that has still preserved its plutonic characteristics along the Lesser Himalaya belt, is part of a much larger ensemble extending in a region exceeding 10^6 km^2 (figure 6). In most of this region it has been affected by the Mesozoic–Cainozoic deformations resulting in the production of large augen gneiss formations such as the Formation III of the HHC of central Nepal. In western Himalaya it may be associated with some peralkaline plutonism (Rafiq 1987). It also forms the masses of porphyritic granite and gneisses of the North Himalaya belt.

These granites are the last plutonic generation before the Himalayan times. As such they have conferred fundamental rheological and geochemical characteristics to the Himalayan basement. In particular, I have suggested (Le Fort *et al.* 1983*b*, 1986) that their southern limit has induced the position of the Main Central Thrust (MCT) of the Himalaya.

2. Tectonic setting of their production

In the overall convergent framework of India and Eurasia, the different belts of granitoid relate to two main types of subduction: oceanic and continental. The youngest plutons are more ambiguous.

Oceanic subduction: Transhimalaya and Karakoram belts

From Jurassic to Eocene, the drift of the Indian Plate has progressively closed the Neo-Tethys, its oceanic floor being subducted to the north under the southern Eurasian margin and producing the Transhimalaya and Karakoram belts. The cafemic to alumino-cafemic character, the calc-alkaline to sub-alkaline nature and some rather low $^{87}Sr/^{86}Sr$ initial ratios of the first stage of magmatism support the inference that magmas were generated from oceanic subduction of oceanic material along an Andean-type margin involving some sialic crust (Debon *et al.* 1981, 1987). This margin shows lateral variations. From east to west, the Transhimalaya Andean margin progressively passes into the Kohistan island arc, branching off the south Tibet continental crust in the Ladakh region (figure 7). According to Debon *et al.* (1986*a*), the island arc has extended, during part of its existence, for 1400 km further eastward, at least as far as Quxu where the Transhimalaya batholith has emplaced simultaneously in the continental margin and anphibolites of the arc, originally built onto its side. To the northwest, the early evolution of the Karakoram prolongs that of the Transhimalaya.

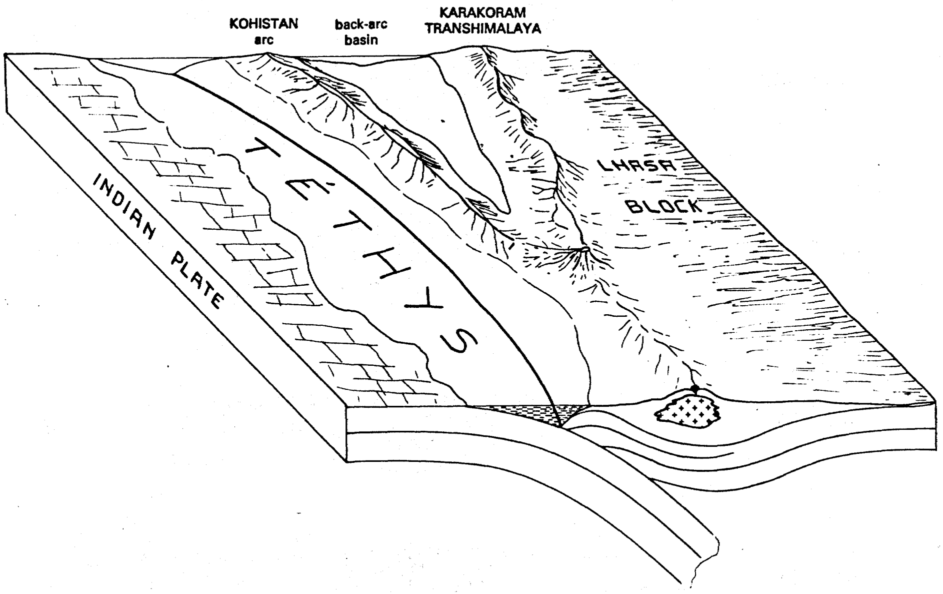

FIGURE 7. Tentative block diagram showing the relation between Transhimalaya Karakoram, Kohistan and India around Upper Cretaceous time. The Kohistan island arc branches off the Transhimalaya belt in the Ladakh region.

Thus the oceanic subduction of the Tethys occurs either below continental crust or below oceanic crust. We have seen that the increasing contribution of the continental crust in the production of the Transhimalayan granitoids was already visible in the $^{87}Sr/^{86}Sr$ initial ratio variations (figure 3). The results from Pb isotope geochemistry (Gariépy *et al.* 1985) also suggest the contribution of lead derived from the crust of the Lhasa Block in the granitoids of the Kangdese region, the contribution being higher in the more acid types. In Karakoram, inherited Precambrian components are found in the zircons of the only analysed pluton, the Baltoro one (Searle *et al.* 1988; Rex *et al.*, this symposium), and the $\delta^{18}O$ values for eastern Karakoram increase with time and evolution of the granitoids (Srimal *et al.* 1987).

We have already noted the subalkaline frequent character of the Transhimalaya and

Karakoram plutons. Debon *et al.* (1986*a*) have suggested that such trends could be related to a strike-slip component along the subduction zone. In the case of the Karakoram-Transhimalaya batholiths, the character is, in fact, met along their entire length.

Intracontinental subduction: High and North Himalaya

Once the India–Eurasia collision occurred, the building of the Himalaya proceeded in two main steps (Mascle 1985; Le Fort 1988). During the first, a system of nappes developed from the suture zone, thickening the crust by at least 15 km (figure 8*b*, *c*). The second step succeeded with the birth of intracontinental thrusting along a thick ductile zone known as the Main Central Thrust Zone (MCT) (figure 8*d*). In between these steps, carbonatite sheets have been emplaced around 31 Ma in the SW part of the Himalayan realm, SW of the northwestern syntaxis (Le Bas *et al.* 1987).

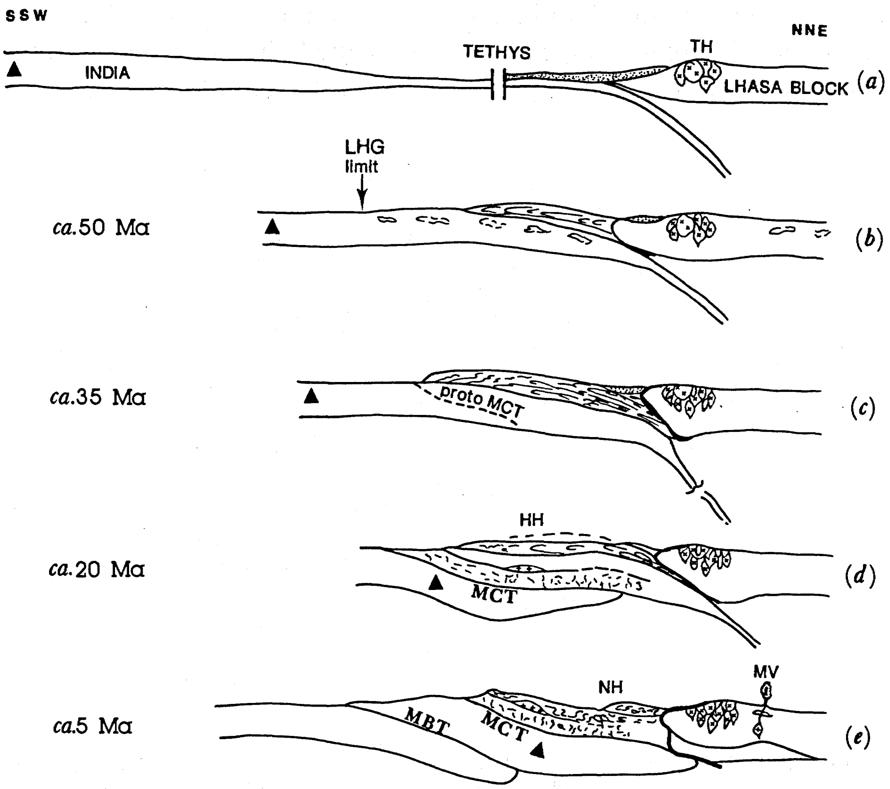

FIGURE 8. Schematic evolution of the Transhimalaya and Himalaya regions from Mesozoic to Recent. (*a*) Mesozoic. The Tethys oceanic crust is subducted under the Lhasa Block Andean margin; sedimentation occurs on both sides of the ocean: dominantly carbonated on India passive margin, of flysch-type south of the active margin. (*b*) Eocene. Collision of India and Eurasia (Lhasa Block) Plates induces the nappe thrusting on the northern margin of India Plate; these nappes include ophiolitic material mélanges, fore-arc sediments and flyschs; the extent of the 500 Ma plutons is shown in both continental plates. (*c*) Eocene–Oligocene. The North Himalaya nappe system had its largest extension; subduction of Indian crust becomes impossible because of buoyancy forces; MCT is about to break out, probably on a former limit of the 500 Ma magmatism; Transhimalaya becomes quiescent. (*d*) Lower Miocene. Underthrusting of the Lesser Himalaya sediments and Indian crust along the MCT releases large quantities of fluids, induces anatectic melting in the High Himalaya crystallines, produces the High Himalaya granites that emplace at the limit between the HHC and the sedimentary series, and slightly later the North Himalaya granites that emplace diapirically. (*e*) Mio–Pliocene. MCT has been relayed by MBT; High and North Himalaya plutonic belts are cooling; to the north of the Transhimalaya, Indian crustal material and or sediments trapped below the Lhasa Block has produced the Maqiang volcanics; erosion in the Himalaya continues at a rate of about 1 mm a^{-1}. HH = High Himalaya, LH = Lesser Himalaya, MBT = Main Boundary Thrust, MCT = Main Central Thrust, MV = Maqiang volcanics, NH = North Himalaya, TH = Transhimalaya.

The first step, of Alpine type, occurred shortly after collision, around 50 Ma and lasted for some 20 Ma. The second step, purely Himalayan, probably started around 30 Ma and lasted for another 20 Ma. It was relayed by another thrusting movement more to the south along the Main Boundary Thrust (MBT) some 10 Ma ago.

The time elapsed between collision and the initiation of the MCT movement enabled the crust already thickened by the nappes, to heat up. Thus the future HHC reached a level of temperature where anatexis was only prevented by the dryness of this basement already granitized during early Palaeozoic time.

The production of High and North Himalayan melts accompanied the movement along the MCT. During this movement, the hotter High Himalaya crystallines (HHC) were thrusted over the weakly metamorphosed Lesser Himalaya. The latter underwent dehydration and released large quantities of fluids that rose into the HHC, and thereby in turn, induced partial melting of them. This kind of process resembles that of dehydration and melting in oceanic subduction such as that of the Transhimalaya. But in the present case, the intracontinental subduction generates an entirely crustal magma with a very narrow range of composition obtained by close to minimum melt anatectic conditions.

In addition, what we know from field and geochemical studies shows that the source rocks of the granite were rather homogeneous along the entire Himalaya and suffered similar conditions of metamorphism and melting.

The mineralogical and geochemical differences between the High Himalaya and the Cainozoic North Himalaya plutons seem to lie mostly in the degree of anatexis; higher anorthite contents of the plagioclase and higher dark-mineral content of the North Himalaya adamellites being related to the higher temperature and therefore the higher percentage of melting attained in the source migmatites.

Dating of the Himalayan granites has proved to be difficult and hazardous. Actually, the magmatism lasts over a period of 10 Ma at least, as shown by U–Pb determinations from 25 to 14 Ma on monazites and zircons (Schärer 1984; Schärer *et al.* 1986; Deniel *et al.* 1987). Such determinations need to be multiplied on the High Himalaya and even more on the North Himalaya plutons to ascertain the slight difference in age (a few million years) that seems to exist globally between the two belts. Taking an average velocity of around 2 cm a^{-1} on the MCT and a distance of some 40–60 km between the two belts, the North Himalaya should be younger by 2–3 Ma, everything remaining similar otherwise.

The late magmatism

In Karakoram (Baltoro granite) as in the Lhasa region of the Transhimalaya (Maqiang volcanites), Miocene magmatism has occurred at a time when the Himalayan magmatism was already active and the MCT was about to be relaid by the MBT. Searching for the cause of this magmatism, one may look some 20–30 Ma back, the time necessary to thermally re-equilibrate a portion of crust (England & Thompson 1984), at the epoch when the system of nappes was replaced by the intracontinental thrusting. The Alpine system probably became stuck by the buoyancy of the marginal terranes underplating the Lhasa Block *s.l.* These terranes, composed of portions of the volcanic arc and back-arc fillings to the west and more flyschoid to the east, dumped below the Karakoram–Transhimalaya range, around Oligo-Miocene time would have undergone melting during Upper Miocene (figure 8*e*).

The production of this late magmatism illustrates rather well the time gap existing between the tectonic impulse and the magmatic response. This gap represents the time necessary for the crust to re-equilibrate the thermal disorders induced by the tectonics. It varies with the physical properties of the crust and the thermal characteristics of the surrounding media but remains in the order of 20–30 Ma. Similarly, one may predict that responding to the blocking of movement along the MCT, around 10–15 Ma ago, crustal melting of the tip of the underplated material (the equivalents of the Lesser Himalaya) should occur within a few million years from now. The present-day activity observed in southern Xizang with abundance of hot springs with high chemistry (Zhang *et al.* 1981), high heat flow measurements (Francheteau *et al.* 1983) and the possible presence of isotropic material at shallow depth shown by magnetotelluric methods (Pham *et al.* 1986) may be precursors of this coming magmatic activity.

3. Tectonic setting of their emplacement

Our knowledge of the emplacement of the intracontinental subduction related plutonism is quite different from that of the oceanic related one. The latter is less well known because the characteristics of this enormous quantity of granitoid rocks as well as of their surrounding formations has, in very few regions, been studied in detail. The former, the High Himalaya and North Himalaya belts, although more inaccessible in a way, are less extensive and have been paid more attention.

Emplacement of the Transhimalaya and Karakoram belts

In both belts, emplacement of granitoid plutons has been going on for several tens of million years between around 100 Ma and 40 Ma. Whether the production and emplacement of magmas has been a more or less continuous process or whether they occurred as discontinuous pulses, remains conjectural. But the more isotopic ages obtained, the smaller the time gap between the different plutons seems to be. This is especially true if one considers the entire belt and not only a small portion of it. At this scale, there seems to be no systemic migration of the magmatic activity with time (Debon *et al.* 1986*a*).

In the Transhimalaya, but probably also in the Karakoram, it seems that the rate of granitoid emplacement and volcanic production is triggered by the collision with the Indian continent (and Koshistan island arc) (Debon *et al.* 1986*a*; Le Fort 1986). Two factors may have contributed to this increase; the dragging of water-rich sediments along the subduction zone during the final stages of India–Eurasia convergence (Debon *et al.* 1986*a*) and the deformation of the collided margin facilitating the ascent of magma in a weakened crust (Debon *et al.* 1985; Le Fort 1988).

Few plutons have been studied with enough detail to appreciate the characteristics of the deformation of the rocks in which they have been emplaced. To our knowledge, only four such plutons, in the Lhasa–Quxu region, have been studied (Brun *et al.* 1983). From the shape and internal asymmetry of the microgranular mafic inclusions, these authors have deduced that the plutons have been emplaced in a portion of the crust undergoing strong left-lateral wrenching. A similar conclusion is reached on geochemical grounds (Afzali *et al.* 1979; Debon *et al.* 1986*a*) to explain the subalkaline nature of the magmatism (cf. figure 2). Strike-slip movement tends to generate tension fractures down to very deep levels in the lithosphere and thus promote alkaline to subalkaline magmatism.

In the Karakoram belt, such strike-slip movement has only been documented at post-magmatic stage (Coward *et al.* 1986) when the mid-Cretaceous plutons were entirely crystallized. In the southern part of the Hunza pluton, for example, a right-lateral strike-slip movement has been documented (Coward *et al.* 1986).

In the Kohistan arc, a first stage of plutonic intrusions has been folded together with the surrounding volcanic and sedimentary rocks. This deformation is related by Petterson & Windley (1985) to the formation of the northern suture between Kohistan and Karakoram, around Upper Cretaceous.

In the best exposed sections of the two belts, the visible thickness of granitoid material frequently exceeds 3000 m as in the Hunza Valley, the Ladakh batholith along the Indus and the Kangdese along some north–south valleys. The floor of the plutons never seems to be exposed. Inclusions of the surrounding rocks in which the plutons are intruded are particularly abundant towards the borders of the more mafic plutons such as Hunza (Debon *et al.* 1987), Ladakh (Sharma & Choubey 1983) and Kangdese (Debon *et al.* 1984). There are more sediments in the surrounding rocks of the Karakoram batholith but more volcanics for the Transhimalaya batholith (see Rai 1983). The inclusion of such metamorphosed shales, limestones, quartzites and volcanics indicates a somewhat high level of emplacement.

The crust of the Transhimalaya was likely to be thick and elevated at the time of India–Eurasia collision. In fact, the Cainozoic compressive deformation of the Gangdese range of southern Tibet has been reported to be rather limited (Tapponnier *et al.* 1981; Burg 1983). According to England & Searle (1986), a precollision elevation of 3000 m in the Lhasa Block would have limited its post-collisional thickening and favoured the deformation and thickening of the suture zone and northern margin of the Indian Plate. A smaller elevation contrast could have prevailed in Karakoram during Upper Cretaceous and would partly explain the high thickening strain of the Karakoram active margin when it collided with the Kohistan arc around Cretaceous–Tertiary boundary.

The level of intrusion being already quite high in the crust, the mean rate of denudation for the Transhimalaya must have been relatively small, in the order of a few kilometres for a minimum of 40 Ma (around 0.1 mm a^{-1}). The same rate estimated for the Karakoram is much higher as some of the plutons are much more recent. However, this rate does not have to be continuous. In fact, a recent study (Copeland *et al.* 1988) along a section southwest of Lhasa, with the dating techniques of $^{39}Ar/^{40}Ar$, has shown a strong acceleration in the rate of uplift in the interval 20–17 Ma, from 0.07 to around 3.7 mm a^{-1}.

Emplacement of the High and North Himalaya plutons

Resulting from the same process of production, the two belts differ in their mode of emplacement.

The High Himalaya granites (HHG)

Mapping of the High Himalaya plutons clearly shows that they start to emplace along the main boundary between the isoclinally folded High Himalaya crystallines, or Tibetan Slab, and the openly folded High Himalaya sedimentary series (Le Fort 1973, 1981, 1986) (figure 5). This is particularly conspicuous when the pluton is thin as for the 300 m thick and 50 km long 'arm of Chhokang', stretching eastward of the Manaslu pluton (Colchen *et al.* 1986*b*). There, the slab of granite corresponds to the disharmonic boundary between the infrastructure and the superstructure, it also corresponds to a regional metamorphic gap, the sillimanite grade below the granite giving way to a mild epizonal metamorphic assemblage only a few hundred metres above the slab in many cases (Manaslu, Shisha Pangma, Everest, etc.), this boundary remains the location of the floor of the main pluton that extends higher up, intruding the sedimentary series from Lower Palaeozoic up to Triassic, Jurassic and even Cretaceous.

Geochemical studies have shown that emplacement of the HHG is a slow process that operates by incremental gathering of small batches of magma (Deniel *et al.* 1987). It probably spreads on more than 10 Ma, the average volume of magma emplaced per year being less than 0.1 km^3 for the entire Himalayan range.

During this very slow process, the HHG do not rise diapirically as they never form a large mass of molten magma. We will see that this is different from the North Himalaya granites (see below). The room necessary for the magma may be produced by two main mechanisms: tectonic movements and fluid convection around the pluton.

The relation of the HHG to regional tectonics may be considered at several scales. On the outcrop, the granite often presents a magmatic layering underlined by micas and tourmaline, superimposed by a tectonic cleavage especially well developed towards the edges of the pluton or in smaller lenses and sheets where it can lead to the formation of an augen gneiss (Le Fort *et al.* 1987). The emplacement of the granite is syn-kinematic to late-kinematic (Le Fort 1975, 1981, 1986). When the pluton is not so big that the entire shape cannot be appreciated (as for the Manaslu and Everest), it appears that it develops a pinch and swell kilometric structure (for the Garhwal plutons; Scaillet *et al.* 1988). Although this structure had developed when the granite was already much more competent than the surrounding sedimentary rocks, it may have offered part of the room necessary for the additional magma to coalesce. In fact the flow and deformation structures vary in intensity, sometimes within very short distances. Thus it is not unlikely that, during deformation, the stretching of the competent granite creates the room for the ascending magma to emplace within it. But this remains difficult to observe in the field as the different batches of leucogranite are generally very much alike in their mineralogy, colour and grain size, and boundaries are extremely hard to follow.

Another way of providing some room for the granite is by arching the roof of the pluton. Actually there is a definite ballooning effect at the top of the granite. Longitudinal sections show this very broad anticlinal doming (France-Lanord *et al.* 1988), which is probably not entirely caused by the boudinage effect mentioned above, but, cutting across the general sedimentary bedding and warping it, may be ascribed to the buoyancy of the granitic magna.

Finally and especially in the case of large plutons, Le Fort (1981) has suggested that the ‘caving out’ of the sedimentary country rocks, when dominantly carbonated, by fluid convection and dissolution around the pluton, could help to emplace significant quantities of magma and make the granite pluton grow. A frozen image of this process is given by the very dense network of granitic and aplo-pegmatitic dykes that crosscut the country rocks of the plutons for several kilometres around them (Le Fort 1981).

Thermobarometric data on the contact aureole towards the roof of major plutons such as the Manaslu (Roy-Barman & Le Fort 1988) indicate that the pressure at the time of the maximum temperature was not less than 270 MPa and could have reached 400 MPa at somewhat deeper levels. Such rather high values indicate that the granite did not reach a much higher level than 10–15 km below the surface. This is a major difference with some of the other leucogranites to which the Himalayan ones are often compared, such as the Hercynian leucogranites of Western Europe, for which high level of emplacement have been obtained especially for the most specialized ones such as Echassières (Cuney & Autran 1987), although they are around 300 Ma old.

For the HHG, such deep level of emplacement and young ages, between 25 and 14 Ma (cf. Le Fort *et al.* 1987), suggest very rapid denudation, in the order of 1 mm a^{-1}. Some values calculated from closure temperatures of different minerals from the Manaslu and Everest plutons give the same order of magnitude: between 0.6 and 0.8 mm a^{-1} for 4–10 Ma (Krummenacher *et al.* 1978; Kai 1981; Le Fort 1988). As suggested by Caby *et al.* (1983), such rapid denudation is probably not only by erosion but implies tectonic denudation. Burg (1983) and more recently Burg *et al.* (1984*a*), Gapais *et al.* (1984) and Herren (1987), for example, have described large longitudinal normal faults lying generally north of the HHG but deforming by northward shear some of them and having a down dip movement of several kilometres, up to 25 km. Burg (1983) mapped several of these north of Everest, one of which is cut by a two-mica leucogranite dated by U–Pb on two fractions of monazite at 25–22 Ma (Copeland *et al.* 1987). Such a relatively old age suggests that tectonic denudation has been active since almost the beginning of the granite emplacement in the High Himalaya.

The North Himalaya granites (*NHG*)

Satellite pictures have shown, before mapping, the elliptical shape of the North Himalaya plutons (Gansser 1977; Academia Sinica 1980; Burg 1983). Their emplacement is not controlled by the HHC–HHSS disharmony as for the HHG. I have suggested (Le Fort 1986) that the NHG belt could be best explained by diapiric emplacement of magma. The NHG plutons are probably intermediate between the two extreme cases of Brun (1981), the gneiss domes and the diapiric plutons, and have risen for more than 10 km through the crust.

The generally good alignment of the belt some 50 km south of the trace of the suture probably corresponds to a broad anticline of the entire continental crust that can be compared with the outer rise observed in oceanic domain at a constant distance from the subduction trench.

To the west, the NHG seems to disappear, although a number of domes such as the Gurla Mandata (figures 1 and 6) and Rupshu are more or less aligned with it and could prolong it westwards all the way to Zanskar and southern Ladakh. However, the interruption of the regularly spaced plutons corresponds with a sharp bend of the MBT trace and its merging with the MCT; one may relate it to the former configuration of India and Eurasia Plate margins (Le Fort 1986).

The presence of chlorite, chloritoid, garnet and staurolite ± andalousite and cordierite (Burg *et al.* 1987) suggests that the NHG emplaced at a roughly similar level to the HHG and that the rate of uplift was of the same order of magnitude (around 1.5 mm a^{-1}). There also, discontinuous normal faulting may be responsible for a good part of this fast denudation. But more-thorough field, petrographical and geochemical work must be collected to constrain better the model of emplacement of the NHG.

Conclusion

The four belts of the Himalayan Orogeny *s.l.* divide in two very contrasting groups (Blattner *et al.* 1983; Pitcher 1983), related to the Tethys oceanic and the India intracontinental subductions respectively. They contrast notably by the duration of their production (more than

70 Ma against about 10 Ma), by the volume of granitoid rock produced (several 10^5 km^3 against a few 10^4 km^3), by the origin of the melt (dominantly oceanic against almost entirely continental), by the petrographic and geochemical nature of the granitoids (diverse against very homogeneous) and by their relation to tectonics (generally little affected against syntectonic).

Their bearing on the tectonic evolution of the region is also quite contrasted. In Karakoram and Transhimalaya, the tectonic activity seems to be low during the emplacement of the plutons and, in turn, these do not seem to foster the deformation or to weaken the crust of the active margin (but they probably form no more than 10% of the crust). On the contrary, in the High Himalaya there is an intricate link between the granite and the migmatitic zone of production as well as in the disharmonic zone of emplacement. It is likely that the periods of normal faulting that occur during the overall compressive régime are made possible by the presence of leucogranitic mushes not yet entirely crystallized. The granites of the Himalaya are an intrinsic part of the tectonic evolution.

All the characteristics of the different belts are in a way the signature of the tectonic situation that has generated them. Time passing and erosion getting to deeper levels, the contrast between the different signatures will even increase. The patterns observed in this region help to interpret similar situations from the past, specially from Precambrian shields where granitic belts are so profuse and other evidences so limited.

These belts are among the best to work on. In fact, deep sections enable us to study and sample at crustal scale, and their relatively recent age has preserved many details and optimizes the geochronological studies. In addition, the absence of orogenic activity in the region for most of the Palaeozoic and Mesozoic, that is for about 400 Ma, clarifies the situation and greatly enhances the legibility of the tectonic and petrologic evolution. But not even half of these belts have yet been geologically walked across and studied on a reconnaissance basis. Thus the hope is great that, with an average amount of work, we will gain an unrivalled knowledge of the most important orogenic processes.

Field work has been supported mostly by the GRECO Himalaya–Karakoram (CNRS). Laboratory work has been supported by the Centre de Recherches Pétrographiques et Géochimiques (Nancy). Fruitful discussions, suggestions and comments from many colleagues have been greatly appreciated. Among them J. M. Bertrand, F. Debon, C. France-Lanord, G. Mascle, P. Molnar, A. Pêcher and S. M. F. Sheppard are specially thanked. Technical assistance from J. Gerbaut and A. Legros is gratefully acknowledged. This is CRPG contribution no. 747.

REFERENCES

Academia Sinica 1980 *Symp. on Qinghai-Xizang* (*Tibet*) *plateau*. (104 pages.)

Afzali, H., Debon, F., Le Fort, P. & Sonet, J. 1979 *C. r. hebd. Séanc. Acad. Sci., Paris* D, 287–290.

Bard, J. P. 1983 *Earth planet. Sci. Lett.* **65**, 133–144.

Bard, J. P., Maluski, H., Matte, Ph. & Proust, F. 1980 Proc. Int. Comm. Geodyn., spec. issue, *Geol. Bull. Univ. Peshawar* (*Pakistan*) **13**, 87–94.

Bassoullet, J. P., Colchen, M., Marcoux, J. & Mascle, G. 1978 *C. r. hebd. Séanc Acad. Sci., Paris* **286**, 563–566.

Bertrand, J. M. & Debon, F. 1986 *C. r. hebd. Séanc. Acad. Sci., Paris* (II) **303**, 1611–1614.

Blattner, P., Dietrich, V. & Gansser, A. 1983 *Earth planet. Sci. Lett.* **65**, 276–286.

Bordet, P., Colchen, M., Le Fort, P. & Pêcher, A. 1981 In *Zagros–Hindu Kush–Himalaya geodynamic evolution* (ed. H. K. Gupta & F. M. Delaney), *Am. Geophys. Union, Geodyn, ser.* **3**, 1149–168. Washington, D.C.: American Geophysics Union.

Brun, J. P. 1981 Thèse Sci., University of Rennes, France. (197 pages.)

Brun, J. P., Pons, J. & Wang, F. P. 1983 *Terra Cogn.* **3**, 264.

Burg, J. P. 1983 Thèse Doct. ès-Sci., University of Montpellier. (368 pages).

Burg, J. P., Brunel, M., Gapais, D., Chen, G. M. & Liu, G. H. 1984*a* *J. struct. Geol.* **6**, 535–542.

Burg, J. P., Guiraud, M., Chen, G. M., & Li, G. C. 1984*b* *Earth planet. Sci. Lett.* **69**, 391–400.

Burg, J. P., Leyreloup, A., Girardeau, J. & Chen, G. M. 1987 *Phil. Trans. R. Soc. Lond.* A **321**, 67–86.

Caby, R., Pêcher, A. & Le Fort, P. 1983 *Rev. Géogr. phys. Géol. dyn.* **24**, 89–100.

Colchen, M., Mascle, G. & Van Haver, T. 1986*a* In *Collision tectonics* (ed. M. P. Coward & A. C. Ries), Geol. Soc. London, spec. publ. no. 19, pp. 173–184.

Colchen, M., Le Fort, P. & Pêcher, A. 1986*b* In *Recherches géologiques dans l'Himalaya du Népal*, éd. Cent. Natl. Rech. Sci., Paris. (138 pages.)

Copeland, P., Harrison, T. M., Parrish, R., Burchfield, B. C., Hodges, K. & Kidd, W. S. F. 1987 *Eos, Wash.* **68**, 1444.

Copeland, P., Harrison, T. M., Kidd, W. S. F. & Xu, R. H. 1988 *Earth planet. Sci. Lett.* **86**, 240–252.

Coulon, C., Maluski, H., Bollinger, C. & Wang, S. 1986 *Earth planet. Sci. Lett.* **79**, 281–302.

Coward, M. P., Jan, M. Q., Rex, D., Tarney, J., Thirlwall, M. & Windley, B. F. 1982 *Nature, Lond*, **295**, 22–24.

Coward, M. P., Windley, B. F., Broughton, R. D., Luff, I. W., Petterson, M. G., Pudsey, C. J. & Rex, D. C. 1986 In *Collision tectonics* (ed. M. P. Coward, & A. C. Ries), Geol. Soc. London, spec. publ. no. 19, 203–219.

Cuney, M., Le Fort, P. & Wang, Z. X. 1984 In *Proc. Symp. on Geology of granites and their metallogenetic relations, Nanjing University, China, 1982* (ed. K. Xu & G. Tu), pp. 853–873. Beijing: Science Press.

Cuney, M. & Autran, A. 1987 *Géol. Fr.* **2–3**, 7–24.

Debon, F. & Le Fort, P. 1982 *Trans. R. Soc. Edinb.* **73**, 135–149.

Debon, F., Le Fort, P. & Sonet, J. 1981 *Symp. Geol. Ecol. Stud. of Qinghai Xizang plateau*, vol. 1, *Beijing, May 1980*, pp. 395–405. Beijing: Science Press.

Debon, F., Sonet, J., Liu, G. H., Jin, C. W. & Xu, R. H. 1982 *C. r. hebd. Séanc. Acad. Sci., Paris* **295**, 213–218.

Debon, F., Sonet, J., Liu, G. H., Jin, C. W. & Xu, R. H. 1984 In *Etude géologique et géophysique de la croûte terrestre et du manteau supérieur du Tibet et de l'Himalaya* (éd. J. L. Mercier & L. Li), CNRS-Acad. Sci. Géol. Chine rédaction, pp. 309–317.

Debon, F., Zimmermann, J. L., Liu, G. H., Jin, C. W. & Xu, R. H. 1985 *Geol. Rdsch.* **74**, 229–236.

Debon, F., Le Fort, P., Sheppard, S. M. F. & Sonet, J. 1986*a* *J. Petr.* **27**, 219–250.

Debon, F., Zimmermann, J. L. & Bertrand, J. M. 1986*b* *C. r. hebd. Séanc. Acad. Sci., Paris* (II) **303**, 463–468.

Debon, F., Le Fort, P., Dautel, D., Sonet, J. & Zimmermann, J. L. 1987 *Lithos* **20**, 19–40.

Deniel, C., Vidal, Ph., Fernandez, A., Le Fort, P. & Peucat, J. J. 1987 *Contr. Miner. Petr.* **96**, 78–92.

Desio, A. 1964 Geological tentative map of the Western Karakoram. Scale 1/500.000. University of Milan: Institute of Geology.

Dietrich, V. J. & Gansser, A. 1981 *Schweiz. miner. petrogr. Mitt.* **61**, 177–202.

Dietrich, V. J., Frank, W. & Honegger, K. 1983 *J. volcan. geotherm. Res.* **18**, 405–433.

England, P. C. & Thompson, A. B. 1984 *J. Petr.* **25**, 894–928.

England, P. C. & Searle, M. 1986 *Tectonics* **5**, 1–14.

France-Lanord, C. & Le Fort, P. 1988 *Trans. R. Soc. Edinb.* **79**. (In the press.)

France-Lanord, C., Sheppard, S. M. F. & Le Fort, P. 1988 *Geochim. cosmochim. Acta.* **52**, 513–526.

Francheteau, J., Jaupart, C., Kang, W. H. & Shen, H. C. 1983 *Terra Cogn.* **3**, 266.

Fuchs, G. 1981 *Mitt. öst. geol. Ges.* **74/75**, 101–127.

Gamerith, H. 1979 *Geologische Karte von Gilgit–Chitral–Wakhan* (*Nord Pakistan und Ost Afghanistan*). Austromineral, Vienna (Austria).

Gansser, A. 1977 Structural map of Himalaya and South Tibet. Scale 1:3.500.000. *Colloq. int. 268. Ecologie et géologie de l'Himalaya, Paris*, (éd. Cent. Natl. Rech. Sci., vol. Sci. de la Terre), pp. 181–191.

Gansser, A. 1983 *Denkschr. Schweiz. naturf. Ges.* **96**, 181.

Gapais, D., Gilbert, E. & Pêcher, A. 1984 *C. r. hebd. Séanc. Acad. Sci., Paris* (II) **299**, 179–182.

Gariépy, C., Allègre, C. J. & Xu, R. H. 1985 *Earth planet. Sci. Lett.* **74**, 220–234.

Hayden, H. H. 1905 *Rec. geol. Surv. India* **32**, 160–174.

Herren, E. 1987 *Geology*, **15**, 409–413.

Honegger, K., Dietrich, V., Frank, W., Gansser, A., Thöni, M. & Trommsdorff, V. 1982 *Earth planet. Sci. Lett*, **60**, 253–292.

Kai, K. 1981 *Geochem. J.*, **15**, 63–68.

Krummenacher, D., Basett, A. M., Kingery, F. A. & Layne, H. F. 1978 In *Tectonic geology of the Himalaya* (ed. P. S. Saklani), pp. 151–166. New Delhi; Today & Tomorrow Publishers.

La Roche, H. de 1964 *Sciences Terre* **9**, 293–337.

Le Bas, M. J., Mian, I. & Rex, D. C. 1987 *Geol. Rdsch.* **76**, 317–323.

Le Fort, P. 1973 *Bull. Soc. géol. Fr.* **7**, 555–561.

Le Fort, P. 1975 *Am. J. Sci*, **275** A, 1–44.

Le Fort, P. 1981 *J. Geophys. Res.* (Red Ser.) **86** B (11), 10545–10568.

Le Fort, P. 1986 In *Collision tectonics* (ed. M. P. Coward & A. C. Ries), Geol. Soc. London, spec. publ. no 19, 159–172.

Le Fort, P. 1988 In *Tectonic evolution of the Tethyan regions* (ed. A.M.C. Sengör) *Proc. NATO ASI meet., Istanbul, Oct. 1985*, Dordrecht: Reidel. (In the press.)

Le Fort, P., Debon, F. & Sonet, J. 1980 In Proc. of the Int. Geodyn. conf., Peshawar, Nov.–Dec. 1979, *Geol. Bull. Univ. Peshawar* (*Pakistan*) **13**, 51–61.

Le Fort, P., Debon, F. & Sonet, J. 1983*a* *Granites of Himalayas, Karakorum and Hindu Kush* (ed. F. A. Shams), pp. 235–255. Lahore, Pakistan: Institute of Geology, Punjab University.

Le Fort, P., Michard, A., Sonet, J. & Zimmermann, J. L. 1983*b* *Granites of Himalayas, Karakorum and Hindu Kush* (ed. F. A. Shams), pp. 377–387. Lahore, Pakistan: Institute of Geology, Punjab University.

Le Fort, P., Debon, F., Pêcher, A., Sonet, J. & Vidal, Ph. 1986 In *Evolution des domaines orogéniques d'Asie méridionale* (*de la Turquie à l'Indonésie*) (ed. P. Le Fort, M. Colchen & C. Montenat) *Mém. Sci. Terre*, **47**, 191–209.

Le Fort, P., Cuney, M., Deniel, C., France-Lanord, C., Sheppard, S. M. F., Upreti, B. N. & Vidal, Ph. 1987 *Tectonophysics* **134**, 39–57.

Mascle, G. H. 1985 *Bull. Soc. géol. Fr.* **8**, 289–304.

Petterson, M. G. & Windley, B. F. 1985 *Earth planet. Sci. Lett.* **74**, 45–57.

Pham, V. N., Boyer, D., Therme, P., Xue, C. Y., Li, L., & Guo, Y. J. 1986 *Nature, Lond.* **319**, 311–314.

Pitcher, W. S. 1983 *Mountain building processes* (ed. K. Hsü), pp. 19–40. London: Academic Press.

Polino, R. 1981 *Geological map of the upper Imja Khola* (*Mount Everest region, eastern Nepal*). Scale 1/25000. S.E.L.C.A., Firenze.

Pudsey, C. J. 1986 *Geol. Mag.* **123**, 405–423.

Rafiq, M. 1987 Ph.D. dissertation, Centre of excellence in geology, University of Peshawar (273 pages.)

Rai, H. 1983 In *Geology of Indus Suture Zone of Ladakh* (ed. V. C. Thakur & K. K. Sharma), pp. 79–97. Dehra Dun: Wadia Institute of Himalayan Geology.

Reuber, I. 1988 *Tectonophysics*. (In the press.)

Reuber, I., Colchen, M. & Mevel, C. 1987 *Geodin. Acta* **1**, 283–296.

Reynolds, P. H., Brookfield, M. E. & Mcnutt, R. H. 1983 *Geol. Rdsch.* **72**, 981–1004.

Roy-Barman, X. & Le Fort, P. 1988 (In preparation.)

Scaillet, B., Dardel, J., Pêcher, A. & Le Fort, P. 1988 In *12ème réunion Sci. Terre, Lille*, p. 120. Soc. géol. France.

Schärer, U. 1984 *Earth and planet. Sci. Lett.* **67**, 191–204.

Schärer, U., Hamet, J. & Allègre, C. J. 1984 *Earth planet. Sci. Lett.* **67**, 327–339.

Schärer, U., Xu, R. H. & Allègre, C. J. 1986 *Earth planet. Sci. Lett.* **77**, 35–48.

Searle, M. P., Rex, A. J., Tirrul, R., Rex, D. C. & Barnicoat, A. 1988 *Bull. geol. Soc. Am.* (In the press.)

Sengör, A. M. C. 1984 *Geol. Soc. Am. spec. paper*, no. 195. (82 pages.)

Sengör, A. M. C. 1985 *Episodes*, **8**, 3–12.

Sharma, K. K. & Choubey, V. M. 1983 In *Geology of Indus Suture Zone of Ladakh* (ed. V. C. Thakur & K. K. Sharma), pp. 41–60. Dehra Dun: Wadia Institute of Himalayan Geology.

Srimal, N., Basu, A. R. & Kyser, T. K. 1987 *Tectonics*, **6**, 261–273.

Stöcklin, J. 1977 *Mém. Soc. géol. Fr.* **8**, 333–353.

Stöcklin, J. & Bhattarai, K. D. 1980 *Geological map of Kathmandu area and central Mahabharat range*. Scale 1/250000. H. M. Govt of Nepal, Dep. min. and geol.

Tahirkheli, R. A. K., Mattauer, M., Proust, F. & Tapponnier, P. 1979 In *Geodynamics of Pakistan* (ed. A. Farah & K. A. De Jong), pp. 125–130. Quetta: Geol. Survey Pakistan.

Tapponnier, P. *et al.* 1981 *Nature, Lond.* **294**, 405–410.

Tu G. C., Zhang, Y. Q., Zhao, Z. H. & Wang, Z. G. 1981 In *Geol. Ecol. Stud. of Qinghai–Xizang plateau*, vol. 1, pp. 353–361. Beijing: Science Press.

Valdiya, K. S. 1981 In *Zagros–Hindu Kush–Himalaya geodynamic evolution* (ed. H. K. Gupta & F. M. Delany), *Am Geophys. Union, Geodyn. Ser.* **3**, 87–110. Washington, D.C.: American Geophysics Union.

Vidal, Ph., Cocherie, A. & Le Fort, P. 1982 *Geochim. cosmochim. Acta*, **46**, 2279–2292.

Xu, R. H., Schärer, U. & Allègre, C. J. 1985 *J. Geol.* **93**, 41–57.

Zeitler, P. K. 1985 *Tectonics*, **4**, 127–151.

Zhang, Z. F., Zhu, M. X. & Liu, S. B. 1981 *Geol. Ecol. Stud. of Qinghai-Xizang Plateau, Beijing, May 1980*, vol. 1, pp. 895–904. Beijing: Science Press.

Discussion

V. S. Cronin (*Department of Geology, and Center for Tectonophysics, Texas A&M University, U.S.A.*). The exposures of the Transhimalayan batholith along the Indus–Tsangpo Suture Zone (ITSZ) seem to cluster in roughly half a dozen large masses that are separated from one another by valleys (see Gansser 1981, figure 2). Analysis of satellite imagery suggests that the valleys may be fault bounded. It is tempting to speculate that the large granitic masses are separated from one another by normal faulting and/or that they are separated along northwest-trending, right-lateral strike-slip faults. Both strike-slip and normal faults can accommodate east–west extension along the ITSZ at the southern edge of Tibet. Hirn *et al.* (1984) have inferred a major strike-slip fault along the ITSZ, based on deep seismic profiling. Armijo *et al.* (1986) describe right-lateral strike-slip faulting along the Karakoram–Jiali Fault Zone, *ca.* 200 km north of the ITSZ.

Does Dr Le Fort have any field data that might indicate whether the Transhimalayan batholith has been segmented by normal or strike-slip faulting? In a similar vein, does he know of any field data to support the hypothesis of strike-slip faulting along the ITSZ?

References

Armijo, R., Tapponnier, P., Mercier, J. L. & Han T. L. 1986 *J. Geophys. Res.* **91**, 13803–13872.

Gansser, A. 1981 In *Zagros–Hindu Kush–Himalaya geodynamic evolution* (ed. H. K. Gupta & F. M. Delany) *Am. Geophys. Union, Geodyn. Ser.* vol 3, pp. 111–121. Washington, D.C.: American Geophysics Union.

Hirn, A., Nercessian, A., Sapin, M., Jobert, G., Xu Zhong Xin, Gao, En Yuan, Lu De Yuan & Teng Ji Wen 1984 *Nature, Lond.* **307**, 25–27.

P. Le Fort. I thank Dr Cronin for stressing the segmented appearance of the Transhimalaya batholith (THB). I am also tempted to interpret this as a result of faulting. However, if we know a number of NS grabens such as the Thakkhola–Mustang one (Colchen *et al.* 1986; Tapponnier *et al.* 1986) that cut through the Indus–Tsangpo Suture Zone (ITSZ) and THB, they do not correspond to the major interruptions of the THB otherwise oriented NE–SW. Right-lateral strike-slip movement would better fit with these major gaps. However, we have very few indications that may support this interpretation and the relative movement should be rather limited at the level of the THB.

From the interpretative simplified maps of Tibet by Taponnier *et al.* (1986, figure 2) and Armijo *et al.* (1986, figure 33), there is a band of right-lateral decoupling along the Karakoram–Jiali Fault Zone (KJFZ), that presently separates northern Tibet from southern Tibet and India. It is likely that this zone was previously active some 150 km more to the south along the Transhimalaya zone when it was in the position of the KJFZ.

As for the ITSZ, strike-slip movement along it has been documented by Burg (1983) in the ophiolitic formations including the late Liuqu conglomerates in the region S and SW of Lhasa. Inferences of older right-lateral strike-slip movements have been put forward for the generation of the small episutural Eocene to Miocene, continental basins of the Ladakh region by Mascle *et al.* (1986). It should also be noted that after suturing, relative movement was transferred towrds the south where strike-slip has been documented in the High Himalaya crystallines (see Burg 1983; Pêcher *et al.* 1984; Gapais *et al.* 1984), this zone becoming the actual limit between India and Tibet during the Himalayan building.

References

Mascle, G., Hérail, G., Van Haver, T. & Delcaillau, B. 1986 *Bull. Cent. Rech. Explor.–Prod. Elf–Aquitaine,* Pau, **10**, 181–203.

Pêcher, A., Bouchez, J. L., Cuney, M., Deniel, C., France-Lanord, C. & Le Fort, P. 1984 In *10éme Réunion sci. Terre, Bordeaux,* p. 436. Soc. géol. France.

Tapponnier, P., Peltzer, G. & Armijo, R. 1986 In *Collision tectonics* (ed. M. P. Coward & A. C. Ries), Geol. Soc. London, spec. publ. no. **19**, 115–157.