The Himalayan tectonics, metamorphism, magmatism and sedimentation

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3. Elements of the Neo-Tethys ocean crust, islands ascs and the Ladakh bathlith of the Asian margin

Oceanic crust, Spongtang klippe, Karzok unit and blueschists of the Indus suture zone.

Obducted oceanic lithosphere on top of the North Himalayan nappes crops out in the Kiogar ophiolite nappe in the Kumaon (Heim & Gansser 1939, Gansser 1964) and the Spongtang klippe in Zanskar (Reuber et al. 1987). In the NW Indian Himalaya, ophiolitic melanges with serpentinites and peridotites, gabbros, pillow basalts of a MORB composition and Late Jurassic radiolarian cherts and limestones occur in the two major ophiolitic melange zones north and south of the Dras-Nindam unit. Blueschists occur in tectonic melanges with Late Cretaceous foraminiferal limestones (Honegger et al. 1982, 1989). In the Spongtang klippe, Reuber et al. (1987) distinguish an upper thrust sheet consisting of tectonized harzburgites and a lower one composed of an ophiolitic melange, Dras volcanics and flysch with Maastrichtian to Paleocene fossils (Garzanti et al. 1987). The Nidar ophiolites (Fuchs & Linner 1996, Mahéo et al. 2000), comprising pillow lavas, gabbros and slightly serpentinized lherzolitic peridotites, are exposed in the Indus suture zone, north of Ribil. The Karzok ophiolite, consisting of a chromite and metagabbro lens, situated on the thrust plane between the Tetraogal and Mata nappes, crops out on the W shore of the Tso Morari (Steck et al. 1998). The mafic rocks of the Spongtang klippe, the Nidar and Karzok ophiolites are considered to be cogenetic. REE patterns and radiogenic Nd isotopic compositions are diagnostic of an N-Morb like depleted mantle source (Mahéo et al. 2000). The authors consider the Spongtang, Nidar and Karzog ophiolites as slices of a same immature intra-oceanic arc. Blueschists occur along the Indus suture in Ladakh in the form of tectonic thrust slices and as isolated blocks within mélange units and in tectonic mélanges associated with Upper Cretaceous foraminiferal limestone (Honegger et al. 1982, 1989). Near Shergol in the Wakha valley, they lie in a mélange zone between the Dras-Nindam zone in the north and the Lamayuru zone in the south. The mineral assemblage of the metabasic volcanoclastic rock sequence is characterised by lawsonite-glaucophane/crossite-Na-pyroxene-chlorite-phengite-titanite±albite±stilpnomelane. P-T estimates indicate temperatures of 350 to 420°C and pressures around 9-11 kbar. Geochemistry indicates a primary alkaline character of the blueschists, which suggests an oceanic island or a transition MORB type primary geotectonic setting. K/Ar isotopic investigations yield middle Cretaceous ages (Honegger et al. 1989).

The Dras-Nindam Formation (Bassoulet et al. 1978a & b, 1983, Honegger et al. 1982, Dietrich et al. 1983) (Callovian-Cenomanian)

The Dras volcanics and their lateral equivalent the shallow-water volcano-sedimentary Dras-Nindam flysch unit constitute a major geological zone, which follows the Indus Suture for more than 400 km between the Karakorum and Leh (Frank et al. 1977a, Bassoulet et al. 1978 a et b, 1983, 1984, Honegger et al. 1982, Dietrich et al. 1983). A 15 km-thick pile of thrust sheets, comprising volcanics, pyroclastics, volcanoclastic sediments and radiolarian cherts, are exposed near Dras. Island arc tholeiitic basalts alternate with dacitic rocks of the calc-alkaline series. Bulk-chemistry, REE-patterns and relictic primary minerals, such as magnesiochromite, clinopyroxene, hastingsite hornblende and Ti-magnetite, suggest that the volcanics belong to island arc tholeiite and calc-alkaline rock series, typical of present island arcs in the Caribbean and Pacific (Honegger et al. 1982, Dietrich et al. 1983). The associated flysch sediments yield Callovian to Cenomanian ages (that include inclusions of Albian to Cenomanian Orbitolina limestones, Wadia, 1937). According to Cannat & Mascle (1990), the volcano-detrital Nindam flysch deposits have an Aptian to Early Eocene age and represent the product of the erosion of a volcanic arc.

The Ladakh intrusives of the active Asian margin

Subsequent to the formation of the Upper Jurassic and Lower Cretaceous Dras volcanic arc of the NW Himalaya, the further subduction of oceanic crust produced large volumes of magmatic rocks of the Ladakh batholith calc-alkaline plutons, which intruded the Dras volcanics to the west of Kargil and the southern border of Asia from the Cenomanian to the Lutetian (Honegger et al. 1982, Cannat & Mascle 1990). To the north of the Indus-Tsangpo suture zone, the Eurasian plate was bordered over a distance of 2500 km by the 30 to 50 km-wide Andean type Transhimalayan (Gangdese) plutonic belt. The Ladakh intrusives represent a north-western segment of this belt. In the Ladakh range, the magmatic activity occurred between 103 and 50 Ma, i.e. between a Cenomanian gabbronorite of 103 ± 3 Ma (concordant U-Pb zircon age) and an Eocene quartz-diorite of 49.8 ± 0.8 Ma (U-Pb zircon age, Weinberg & Dunlap 2000). These ages are similar to the 94.2 ± 1 Ma to 41.1 ± 0.4 Ma U-Pb ages on zircon from samples of the Lhasa-Xigaze region in Tibet (Schärer et al. 1984). A 103 Ma Ladakh biotite-granodiorite intruded the Dras volcanics near Kargil, producing an aureole of contact metamorphism (Frank et al. 1977a). This cordillera type calc-alkaline suite testify to the partial melting of the Asian mantle above the NW-directed underthrust Neo-Tethys ocean floor. The Ladakh intrusives are generally non-metamorphic. Prehnite mineral assemblages in a shear zone within a diorite near of the village of Chumatang to the west of Mahe in the Indus Valley are interpreted as a product of a post-magmatic hydrothermal activity, rather than of a regional metamorphic overprint (Schlup et al. 2003).

4. The Himalayan tectonics, metamorphism, magmatism and sedimentation

The Indus Group (Garzanti & Van Haver 1988)
(synonyms: Indus Molasse, Tewari 1964, Fuchs 1982; Indus Clastics, Garzanti & Van Haver 1988)
(Late Cretaceous-Neogene)

Late Cretaceous to Neogene sediments were successively deposited in a fore arc basin of the Ladakh batholith and in an intermontane molasse basin. They are exposed over a distance of 2000 km along the Indus–Tsangpo suture zone, between the Nanga Parbat in Pakistan and the Namche Barwa syntaxis in Assam (Mascle et al. 1986). Sinclair & Jaffey (2001) restricted the use of the term Indus Group to the molasse sediments that have been deposited during the Tertiary in an intermontane basin. A simplified synthetic stratigraphic column based on observations by Dainelli (1933), Tewari (1964), Fuchs (1977, 1979, 1982), Frank et al. (1977a), Bassoulet et al. (1978b, 1982), Baud et al. (1982), Brookfield & Andrews-Speed (1984); Van Haver (1984); Bucher & Steck (1987); Searle et al. (1990), Sinclair & Jaffey (2001) and Steck et al. (1993) is illustrated in the palinspastic section of Fig. 5.

The Khalsi Limestone (Bassoulet et al. 1978 a and b, 1984) and Thar Formation (Van Haver 1984)
(Aptian-Albian)

The oldest sediments in the southern part of the Indus Group sedimentary sequence are the Aptian-Albian Khalsi Limestones, up to 50 m thick, exposed S of the Indus river between Saspul and Nurla that form a tectonic contact with the ophiolitic melange zone and Nindam Flysch of the Indus suture. The platformal carbonates contain an abundant fauna of orbitolinas, rudists and nerineas. They are interpreted by Garzanti and Van Haver (1988) as a limestone platform forming an atoll or deposits of a fore-arc basin perched on the Nindam accretionary wedge in front of the Ladakh batholith. The Khalsi Limestones are unconformably transgressed by 200 m of Tar Formation, Late Cretaceous green turbidites, rich in belemnites with two coarse conglomerate bodies interpreted as deep-sea fan turbidites of a Maastrichtian and probable Paleocene age, overlain by shallow-water carbonates with gastropods, bivalves and nummulites of the Paleocene Sumdha Gompa Formation and finally capped by sandstones and pelites deposited in a deltaic environment and paleosols (Garzanti & Van Haver 1988).

The Miru-Chogdo Flysch (Miru Flysch, Fuchs & Linner 1996; Chogdo Flysch (Baud et al. 1982)
Synonyms: Jurutze Flysch, Jurutze Marls, Brookfield and Adrews-Speed (1984), Chogdo Flysch and Chogdo marine Molasse (Steck et al. 1993), Nummulitic Limestones (including Sumdha Gompa Formation, Parcha Kanri Conglomerate and Urucha Marl (Garzanti & Van Haver 1988), Jurutse, Sumda and Chogdo Formation (Sinclair & Jaffey 2001)
(Cenomanian-early Eocene)

The transition from the open sea fore-arc basin sedimentation in front of the Ladakh arc to the continued marine sedimentation of the intra continental molasse basin is gradual and coincides with the deposition of the yellow fine-grained Miru-Chodgo Flysch with nummulitic limestone lenses of a supposed early Eocene age. The Miru-Chogdo Flysch overlies with a stratigraphic continuation the Maastrichtian-Paleocene Tar and Early Eocene Sumdha Gompa Formations and is succeeded by the continental red shales and red and green conglomerates of the Gongmarula -Sumdo-Hemis-Nurla Molasse. New observations by Fuchs & Linner (1996) in the Kiari section, N of the Tso Morari dome, revealed microfossils such as *Rotalipora cushmani* and *Dicarinella algeriana* of a <middle to late Cenomanian age and *Hedbergella planispira* and *Globigerinelloides bollii* of an age not older than the Turonian in the Miru Flysch. These new observations document that the Chogdo-Miru Flysch facies may also be a lateral equivalent of the Tar and Sumda Gompa Formations.

The Gongmarula-Hemis-Nurla Molasse
(early Eocene?)

The term Gongmarula-Hemis-Nurla Molasse has been chosen to designate all the different types of meandering and braided river sediments and deltaic prograding river fans composed of red and/or green shales, sandstones and conglomerates deposited in the intra-continental Indus Molasse basin, within the Paleo-Indus valley (Fig. 5). The Gongmarula Molasse corresponds to red siltstones, which alternates with red or green sandstones and conglomerates; the Hemis Molasse is mainly conglomeratic; where as the Nurla Molasse and Sumdo red Molasse are dominated by shales and sandstones. Many other formation names have been proposed for the molasse sediments and there are also major differences between the stratigraphic columns proposed by different geologists working in the Indus Molasse (Baud et al. 1982, Brookfield & Andrews-Speed 1984; Van Haver 1984; Bucher & Steck 1987; Searle et al. 1990, Sinclair & Jaffey 2001 and Steck et al. 1993). There is a major difference between the stratigraphic columns proposed by Garzanti & Van Haver (1988), with the Hemis Conglomerates at the base and the Nurla Formation on top, and the succession of Sinclair & Jaffey (2001) with the Nurla Formation at the base and the Hemis Conglomerates on top. A palinspastic reconstruction based on a detailed structural and stratigraphic study of the Martselang–Gongmaru La geological transect suggests that the coarse Hemis, variegated and Stok (Gongmaru La) conglomerates and the fine-grained red Sumdo and Gongmaru La shales and siltstones represent interfingered lateral equivalents within a 4000 m thick pile of meandering and braided river deposits and prograding deltas with local lakes and marshes (bird foot prints, rain drop marks, etc. in the Gongmaru La region) (Fig. 5 and Figs. 4–10 in Steck et al. 1993). Our reconstruction can explain the contradiction between the stratigraphic columns of Garzanti & Van Haver (1988) and Sinclair & Jaffey (2001) as well as propositions by other authors. The detrital sediments contain granitoid pebbles of the Ladakh rang, red radiolarites, basalts and chromite from the obducted Neo-Tethys ocean crust and the Dras island arc and limestone and marble pebbles from the Himalayan range (Van Haver 1984, Bucher & Steck 1987). The Gongmaru La–Hemis–Nurla Molasse is not dated with fossils. The molasse sediments, overprinted by a Late Eocene regional metamorphism, is thus of an Early and Middle Eocene age (Van Haver et al. 1986). These synorogenic sediments were deposited in an intramontane Indus basin, over 30 km wide,

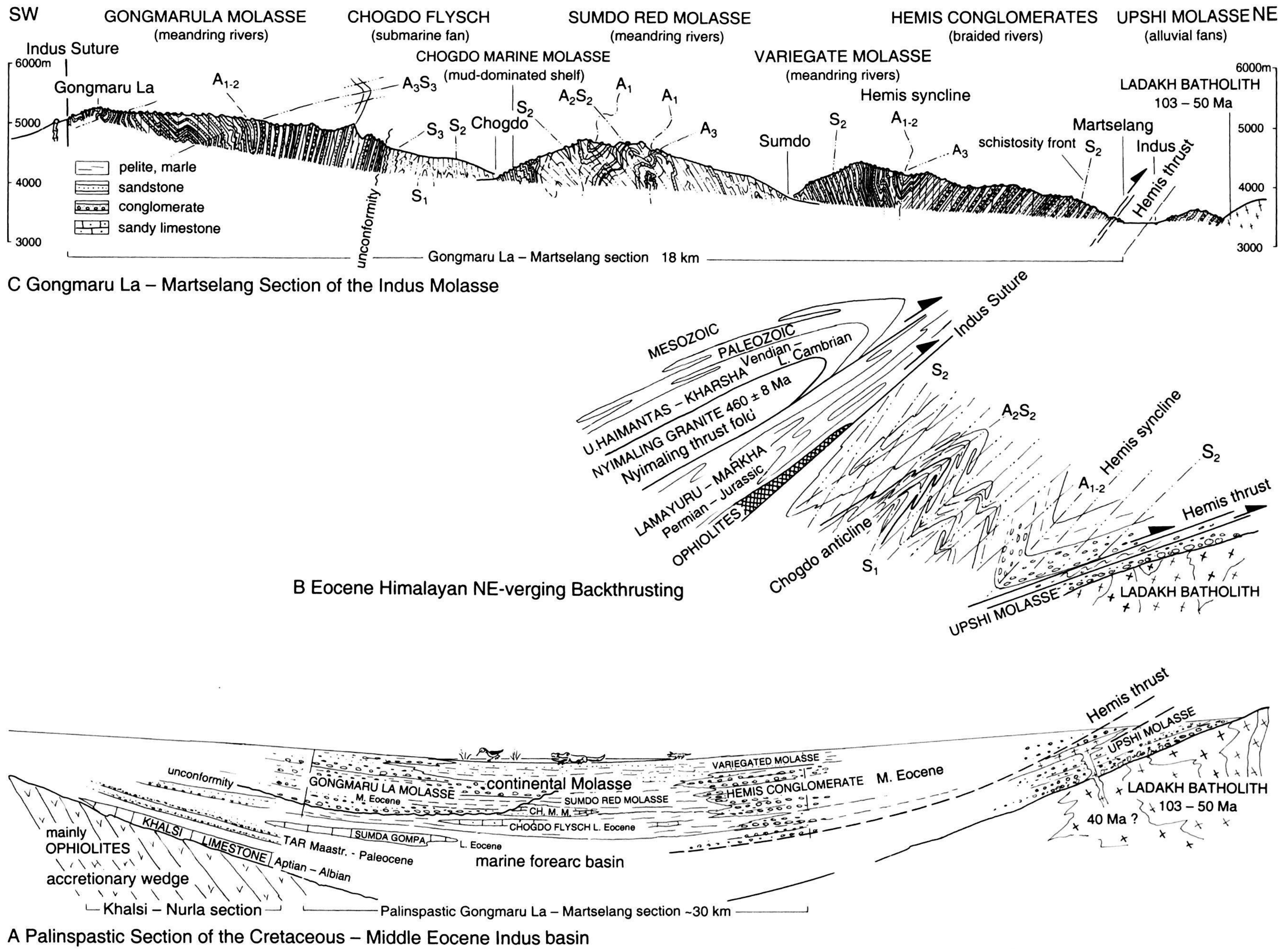

Fig. 5. Geological section of the Gongmarula-Martselang transect through the Indus Molasse (C), the Eocene phase of NE-verging (back) thrusting (B) and the palinspastic section of the Upper Cretaceous – Middle Eocene Indus basin (A) according to Steck et al. (1993).

after the Early Eocene. Between Leh and Upshi, the Gongmarula-Hemis-Nurla Molasse is overthrusted by the Hemis imbricate thrust on the autochthonous molasse conglomerates (Upshi Molasse (Fig. 5); Frank et al. 1977a, Steck et al. 1993), the SE structural equivalent of the Basgo Formation.

The Basgo Formation (Basgo Unit, Baud et al. 1982)
(Maastrichtian)

On the northern border of the Indus basin, the Ladakh granitoids are unconformably overlain by alluvial and deltaic deposits, starting with the 10-200 m of coarse conglomerates of the Basgo Formation (Basgo-Skinning Formation or Molasse) changing laterally and locally into lacustrine limestones and marls with a Maastrichtian ostracod fauna (Van Haver 1984). It is not clear if the undated autochthonous Upshi Molasse (Frank et al. 1977a) represents a lateral equivalent of the autochthonous Basgo Molasse or of the younger Nimu Sandstones, we have decided to show on our map.

The Nimu sandstones
(synonyms: Nimu grits Brookfield & Andrews-Speed 1984, Sinclair & Jaffey 2001; Nimu Formation, Van Haver 1984, Garzanti & Van Haver 1988)
(Paleocene-early Eocene)

The Nimu Sandstones and Temesgang Formation stratigraphically lie on the Basgo Formation. Garzanti & Van Haver (1988) consider the two undated units to have different ages, the Nimu Formation as the younger and Temesgam Formation as the older sediments of the molasse stratigraphic column. These two similar formations are represented by the same colour on our map. The undated Nimu Sandstones, up to 600 m thick, consist of an alternation of mudstones and sandstones, are preserved in the central part of the Indus basin between Saspul and Upshi. Near Upshi, they stratigraphically overlie the Ladakh granitoids (Frank et al. 1977a). Freshwater pelecypods, slumps and turbiditic sedimentation indicate a lacustrine environment. The Indus valley was occupied by a large lake (Van Haver 1984, Garzanti & Van Haver 1988, Searle et al. 1990). The contact with the overlying Gongmarula-Hemis-Nurla Molasse is tectonic (Hemis thrust Frank et al. 1977a, Steck et al. 1993).

The Temesgam Formation (Van Haver 1984)
(Paleocene-early Eocene?)

During the Paleocene and early Eocene, more than 1000 m alternating fine-grained yellowish pelites and reddish sandstones of the Temesgam Formation was deposited in stratigraphic continuation on the Basgo deltaic sediments. The sedimentary features and occurrence of freshwater pelecypods suggest that these fine-grained sediments are lateral equivalents of the Nimu Formation.

The Chilling and Butum-Kargil Formations (Honegger 1983)
(Mio-Pliocene?)

The Chilling and Butum-Kargil Formations are composed of coarse continental conglomerates and sandstones, probably deposited during the Mio-Pliocene time on the folded older Indus Group deposits and the vertical units of the Indus Suture zone. (Wanlah Conglomerate, Mascle et al. 1986, Cannat & Mascle 1990).

The tectonics of the sediments of the Indus Group

A major unconformity and erosion gap in the Indus Molasse profile of the Martselang-Gongmarula transect is interpreted as evidence for important tectonic movements during the Molasse deposition (Fig. 5C, Steck et al. 1993). Furthermore, some younger epi-sutural conglomerates unconformably overly folded molasse deposits in the Markha valley (Christan Talon, personal communication). Major differences in the description and interpretation of the tectonics of the Indus Group have led, as mentioned above, to very different and contradictory stratigraphic columns (Baud et al. 1982, Brookfield & Andrews-Speed 1984, Cannat & Mascle 1990, Van Haver 1984; Bucher & Steck 1987; Garzanti & Van Haver 1988; Searle et al. 1990, Sinclair & Jaffey 2001 and Steck et al. 1993). A synthesis based on these divergent proposals is impossible, as the differences are often based on incompatible observations, interpretations and structural concepts. A synthetic stratigraphic column and a palinspastic model of the Martselang–Gongmarula valley transect (after Steck et al. 1993) are illustrated in Figs. 5A and C. This interpretation is fundamentally different to the structural descriptions and interpretations of the same profile by Baud in Baud et al. (1982), Van Haver (1984), Garzanti & Van Haver (1988) and Sinclair & Jaffey (2001). The main difference is due to the fact that some strongly deformed stratigraphic contacts between very ductile shale and competent conglomerate or sandstone layers have been erroneously interpreted as faults by Baud. The geometry of the Indus Molasse, at a regional scale, is characterised by a succession of open anticlines and synclines of greater than 1 km wave length, such as the spectacular NE-striking syncline of Hemis Gompa to the south of Leh (Fig. 9 in Frank et al. 1977a), a molasse fold style that was already documented by Dainelli (1933, Fig. 39 in Gansser 1964). From detailed structural analyses up to 3 foldgenerations and two schistosities related to the NE-verging late Himalayan back folding and thrusting to the north of the Indus Suture can be recognized. The Hemis syncline is a second phase A_2 fold with a weak axial surface schistosity S_2. Early A_1 folds and axial surface schistosity S_1 were developed in the more ductile shales and siltstones of the Chogdo Flysch and locally in the younger red Sumdo Molasse shale and siltstone. Second folds A_2 with their axial surface schistosity S_2 were then back-rotated into their present vertical position concomitant with a late updoming of the Ladakh batholith to the N and the rotation of the molasse transgressional surface to the present 60° SW-dipping position. Local third generation NE-verging open folds A_3 were developed after the preceding tilting of the molasse A_2 folds and S_2 schistosity. Detailed geological mapping of the Indus Suture zone of Eastern Ladakh between the Taglang La, Ribil and Mahe by Fuchs & Linner (1996) revealed a complicate structure for the Indus Suture zone. They suggest an early phase of SW-directed imbricate thrusts of the ophiolites with the overlying autochthonous Miru-Chogdo Flysch and Indus Molasse. Two such thrust sheets are distinguished in the middle section of Fig. 7. These early imbricates were then overprinted by the younger NE-verging folds and thrusts of the Indus Suture zone.

The metamorphism of the Indus Group

The sediments of the Indus Group are affected by a very low grade metamorphism: high diagenesis to the NE at the contact with the Ladakh intrusives which grades up to the anchizone or prehnite-pumpellyite facies to the SW in the Gongmarula Molasse (Fig. 5C & Plate 4). This zonation is based on illite crystallinity and prehnite-pumpellyite mineral assemblages (Plate 4; Van Haver et al. 1986, Steck et al. 1993). The metamorphism is related to the NE-directed back thrusts of the N Himalayan nappes and the NE-verging folds, thrusts and schistosity in the Indus Molasse and the Nyimaling thrust (Fig.7) in the Nyimaling-Tso Morari dome (Van Haver et al. 1986, Steck et al. 1993). Based on K-Ar dating of separated micas, Van Haver et al. (1986) estimated the age of this tectono-metamorphic event to be 35-40 Ma. These data are in agreement with the age of cooling below 300°C of the regional Barrovian metamorphism of the North Himalayan nappe stack exposed in the Tso Morari region (De Sigoyer et al. 2000, Schlup et al. 2003). These ages indicate also that the sediments of the Indus Group, with exception of the Chilling and Butum-Kargil formations of probable Mio-Pliocene age, are not younger than the Late Eocene. Likewise, their deposition in an epi-sutural basin is related to the erosion of the Ladakh batholith to the N and of the accretionary wedge composed by the Late Paleocene – Eocene North Himalayan nappe stack to the S.

The Himalayan nappes

Six phases of nappe emplacement have been distinguished in the Himalayan range based on new studies of the Leh–Mandi transect of the NW Indian Himalaya (Plate 3). From oldest to youngest, the phases are characterized by the implacement of the following nappes:

1 The NE-directed Shikar Beh nappe.
The SW-directed North Himalayan nappes composed of
2 The eclogitic Tso Morari nappe and
3 The higher (non eclogitic) units of the N Himalayan nappe stack
4 The High Himalayan nappe, "Crystalline nappe" or Main Central thrust (MCT).
5 The Lesser Himalayan nappes or Main Boundary thrust (MBT.)
6 The Himalayan frontal thrust belt or Active Himalayan thrust (AHT).

Each nappe stack is responsible for crustal thickening and a perturbation of the geothermal gradient, due to the underthrusting of cold below hotter crustal slices and followed by thermal relaxation. So each nappe stack has induced its own orogenic metamorphism that will be described separately. Plate 4 shows the metamorphic zonation of the NW Indian Himalaya. In the Tso Morari dome only, the high-pressure metamorphism restricted to the Tso Morari nappe has been distinguished from the zones of the younger Barrovian metamorphism related to the entire North Himalayan nappe stack. The distinction of the other phases of similar Barrovian type metamorphism is difficult and hence not represented on the metamorphic map. The contact metamorphic aureoles around the Ordovician granites are likewise not illustrated.

The Shikar Beh nappe

The oldest nappe structure of the Himalayan range is the NE-directed Shikar Beh nappe (Steck et al. 1993, 1998, 1999, Vannay & Steck 1995, Epard et al. 1995, Wyss et al. 1999, Robyr et al. 2002). Its most spectacular structure is the NE-verging Tandi syncline of Permian and Mesozoic sediments, exposed in the Pir Panjal range, south of Tandi, on the left bank of the Chandra-Chenab river. Early NE-directed folds and thrusts of the Shikar Beh nappe have been documented over a distance of 200 km between the Miyar valley and the Lagudarsi La east of Kiato in the upper Spiti valley (Steck et al. 1999). A medium pressure regional metamorphism is related to this early Himalayan nappe stack. Kyanite and staurolite assemblages in metapelites appear between Kalath in the upper Beas valley and Khoksar in the Chandra valley. The estimated overburden is about 20 km (Epard et al. 1995). From this area, the degree of the regional metamorphism decreases gradually NE-ward down to the anchizone with pumpellyite mineral assemblages in metabasites south of the Baralacha La. The latter mineral assemblage crystallised below an overburden of less than 10 km (Frank et al. 1977b, Vannay 1993, Steck et al. 1993). The metamorphic zonation is explained by a nappe stack thrusted to the NE (Plate 4, Steck et al. 1993, Vannay, 1993, Epard et al. 1995, Vannay & Steck, 1995, Wyss, 2000). A penetrative SW-dipping syn-metamorphic stretching lineation and micro- and mesoscopic scale top-to-the NE shear sense criteria reveal the NE-directed thrust direction (Fig. 6, Steck et al. 1999, Robyr et al. 2002, Robyr 2003). The NE-verging folds and thrust of the Shikar Beh nappe front were in a later stage overprinted by the SW-directed folds of the younger N Himalayan nappe front, such as near Batal in the Chandra valley (Fig. 10; Steck et al. 1993, Wyss et al. 1999) and at the Lagudarsi La in the Spiti valley (Fig. 7; Steck et al. 1998). The NE-verging Shikar Beh nappe structures are overprinted by the younger SW-directed folds and schistosities of the High Himalayan nappe (Fig. 10). The Shikar Beh thrust is older than the about 22 Ma Early Miocene Main Central thrust at the base of the High Himalayan nappe (Frank et al. 1977b, Hubbard & Harrison 1989). Zircon FT data of Schlup et al. (2003) from the Tso

Morari region show that the main emplacement of the North Himalayan nappes occurred before 40 Ma and Vance and Harris (1999) suggest that synkinematic garnets of the nappe front crystallizeed ~30 Ma ago. In conclusion, the NE-directed Shikar Beh nappe represents the oldest Tertiary intracontinental nappe structure of the north Indian crust accreted in the Himalayan range. This compressional structure has probably been formed by reactivation of a pre-existing SW-dipping listric normal fault of the North Indian continental margin. Le Fort (1997) compares the early NE-directed Shikar Beh thrusts and folds of the NW Himalaya with the late backfolds of the Alps, an assumption that is in contradiction with the structural history.

The N–verging Warwan, Bobang, Bor Zash and Wakha recumbent folds and N-directed Dras nappe

The tectonic map of the NW Himalaya and geological sections of Honegger (1983) show the N-verging Bor Zash, Bobang and Warwan recumbent folds to the NW of the Kisthwar window and in the Nun-Kun (Suru dome or Suru syntaxis) region (Plate 2, section 1). Honegger (1983, p. 89 and geological sections on its Fig. 47) suggests the following mechanism:

"Als Hypothese kann man sich für die zusammengefaltete und gestauchte Struktur dieses Gebietes folgende Entwicklung vorstellen:

1 Aufrechtstehender Faltenbau
2 Entwicklung der Decke (Kristallindecke) im Untergrund, bestehende Faltenstrukturen erhalten NE-Vergenz um NW - SE laufende Achsen
3 Einsetzen der Falten mit SW – Vergenz
4 Abdrehen der Warwan Antiklinalen, der Bobang – und Bor Zash Einheit in N-S Richtung mit Ausbildung des Suru Domes.

Translation:

One hypothesis would be to imagine the following development for the superimposed structures of this region:

1 Vertical folding.
2 Development of the (Crystalline) nappe at a deeper level, pre-existing fold structures became NE-vergent with NW-SE striking fold axes.
3 Starting of SW-verging folds.
4 Rotation of the Warwan anticlines, the Bobang – and Bor Zash units in a N-S direction with formation of the Suru dome.

In other words, Honegger (1983) suggests that the huge Bor Zash, Bobang and Warwan recumbent folds are early Himalayan folds that have been rotated in their final N-verging position during SW-directed thrusting of the deep-seated "Crystalline nappe".

A different interpretation is proposed for the NE-directed young Wakka and Dras thrusts and folds. Gilbert (1986) studied the N-verging Wakha recumbent fold of Triassic and Liassic sediments and the underlying N-directed Dras nappe. According to Gilbert (1986), Gilbert & Merle (1987) and Gapais et al. (1992), the Wakha fold nappe represents a late Himalayan structure, postdating the piling up of the units above the MCT, due to synconvergent spreading of the ductile Himalayan metamorphic rocks. They relate the spreading with the Early Miocene Zanskar extensional shear zone.

In conclusion, there are strong arguments for the existence of an early Himalayan NE-directed Shikar Beh nappe stack. It is also reasonable to propose the same interpretation for the N-verging Bor Zash, Bobang and Warwan recumbent folds of the High Himalaya in NW Zanskar. On the other hand, the interpretation of the N-directed Wakha recumbent fold and the Dras thrust as structures related to the Shikar Beh nappe system is improbable. A convincing suggestion is that these thrusts may be related to a late orogenic synconvergent spreading to the N of the Early Miocene Zanskar detachment (Gapais et al. 1992). This type of NE-verging folds are often developed in the roof of north-dipping faults of the Central Himalayan extensional detachment system (Pêcher 1991, Steck et al. 1998, Epard & Steck in press). It is proposed that the N-directed Wakha and Dras thrusts are of the same late phase as the NE-verging backfolds and thrusts of the Indus Molasse south of Leh. A major unanswered question remains the exact age of the different NE-verging structures of the western Zanskar region. More radiometric dating and structural work is needed.

The Barrovian metamorphism (M1) of the Shikar Beh nappe stack

The amphibolite, greenschist and prehnite-pumpellyite facies metamorphic zones of the Rohtang Pass - Chandra valley - Baralacha La transect are related to the stack of the NE-directed Shikar Beh nappe (Plate 4, Steck et al. 1993, Vannay, 1993). The change in the thickness of the actually eroded overburden, from over 20 km in the Rohtang La region to less than 10 km in the Baralacha La region, is explained by an overthrust in a NE direction, one that is corroborated by the NE-verging Tandi Syncline of Mesozoic rocks. As well as in the Miyar valley transect farther to the NW, the regional metamorphism is related to the NE-directed structures of the Miyar thrust zone, a local expression of the Shikar Beh nappe (Steck et al. 1999, Robyr et al. 2002, Robyr, 2003). In the Kullu valley, between the Rohtang pass and Mandi, Epard et al. (1995) distinguished two phases of Barrovian recrystallisation, relicts of an early metamorphism (M1) and mineral assemblages (M4) that crystallised syn- and postkinematically with respect to the High Himalayan nappe emplacement. The metamorphic map (Plate 4) shows for the Kullu valley the position of the early isograds of the relictic Shikar Beh nappe M1 metamorphism. It was not possible to find unaltered mineral assemblages of the M1 metamorphism for a thermo-barometric microprobe study because of the overprint by the younger M4 metamorphism of the Crystalline nappe. Radiometric Ar-Ar dating of hornblende of the Shikar Beh M1 metamorphic event was not successful, due to Argon overpressure.

The North Himalayan nappes

A stack of N-Himalayan nappes in the Kumaun Himalaya transect was first described by Heim & Gansser (1939) and, after the opening of Ladakh to western geologists, in the Zanskar region by Bassoulet et al. (1980) and Baud et al. (1982). Today, the nappe stack model for the so-called Tethys or Tibetan zone of the Himalaya is generally accepted (Burg & Cheng 1984, Stutz & Steck 1986, Searle 1986, Searle et al. 1987, Steck et al. 1993, 1998, Ratschbacher et al. 1994). Fuchs (1982, 1985, 1986, 1989, Fuchs & Linner 1995, 1996) still believes in a folded but autochthonous Tethys Himalaya sedimentary sequence.

The eclogitic Tso Morari nappe

The deepest unit of the North Himalayan nappe stack is the Tso Morari nappe, exposed in a tectonic window below the higher Tetraogal and Mata nappes of the Tso Morari region (Figs 6, 7 and Plate 3, Steck et al. 1998). The Tso Morari nappe is composed of over 95 vol. % by the 479 ± 2 Ma Ordovician Tso Morari granite (zircon U-Pb age, Girard & Bussy 1999), associated with Upper Proterozoic - Cambrian sediments and cross-cut by mafic dikes. The strongly mylonitic schistosity, with at least two schistosities and isoclinal fold generations, and a strong NE-SW directed stretching lineation are parallel to its upper contact with the overlying Tetraogal and Mata nappes (Figs 6 & 7). These structures were formed during its subduction and later extrusion parallel to the plane of underthrusting of the Indian crust (Steck et al. 1998). The mechanism of nappe emplacement will be discussed later. A younger NW-SE oriented streching lineation has been developed during late dextral strike-slip movements south of the Indus Suture (Fig. 6).

The high-pressure metamorphism (M2) of the Tso Morari nappe

The rocks of the Tso Morari unit are strongly mylonitised and have successively been overprinted by an eclogitic high-pressure metamorphism M2 discovered by Berthelsen (1953) and a younger Barrovian metamorphism M3 (Plate 4, Fig. 7; De Sigoyer et al. 1997, Girard 2001). De Sigoyer et al. (1997) observe the following high-pressure mineral assemblages:

Metabasites: garnet-omphacite-quartz-glaucophane-phengite (Si 3.56)-paragonite-zoisite-rutile.

Fe-rich metapelites: garnet-jadeite-glaucophane-chloritoid-paragonite-phengite (Si 3.58-3.24)-zoisite-chlorite±biotite,

Mg-rich metapelites contain kyanite – Mg-chlorite,

Intermediate metapelites contain staurolite-kyanite-biotite-chlorite.

P-T conditions of ≥20 ± 3 kbar and 550 ± 50 °C have been identified for the high-pressure metamorphic peak (De Sigoyer et al. 1997) and 9 ± 3 kbar and 610 ± 70 °C for the Barrovian metamorphism (De Sigoyer et al. 1997, Girard 2001, Schlup et al. 2003). Mukherjee & Sachan (2001) identified with Raman spectrometry coesite. This very high pressure SiO_2 polymorph occurs optically in close association with garnet and omphacite as inclusions in radial fractured garnets. This indicates a minimum pressure of 27 Kbars at ~600 °C, pointing to a depth of subduction of over 90 km (Massone, 1995). The mylonitic Tso Morari granite is composed of the non-critical mineral assemblage: quartz - K-feldspar - albite - oligoclase - garnet (pyr 4-20 – gross 15-35 – Alm+Spess 60-80) – clinozoisite – phengite (Si 3.1-3.28) – biotite – titanite – ilmenite. For an unknown reason the sodic plagioclase was not transformed into jadeite. Also the muscovite dehydration dry melting reaction was not attained (Girard 2001). The predominance of this low-density gneiss (>95 vol. %) in the Tso Morari nappe may explain its extrusion driven by buoyancy forces. The isograds of the Barrovian regional metamorphism cross-cut and seal the contact of the lower Tso Morari and the higher Tetraogal and Mata nappes after their emplacement (Plate 4, Fig. 7; Girard 2001). The high-pressure mineral assemblages have been dated at 55 ± 7 Ma on garnet-glaucophane-whole rock with the Sm-Nd method, at 55 ± 12 Ma on garnet clinopyroxene-whole rock with the Lu-Hf method and 55 ± 17 Ma on alanite with the U-Pb method (De Sigoyer et al. 2000). The kinematics of the exhumation and emplacement of the eclogitic Tso Morari nappe will be discussed below together with higher units of the N Himalayan nappe stack or accretionary wedge.

The higher (non eclogitic) units of the North Himalayan nappe stack

The North Himalayan nappes occupy the northern internal part of the Himalayan range (Plate 3 and Figs 6, 7). They are arched down in a vertical root along the Indus-Tsangpo suture to the NE. Towards the SW, their frontal thrusts coincide with the crest line of the High Himalaya (Haptaldome, Plate 1), the Zanskar Crystalline zone between the Nun-Kun to the NW and the Gianbul dome with the Tertiary Gumburanjun leucogranite intrusion to the SE. Farther to the SE, the frontal trusts and folds cross the Baralacha La, the upper Chandra valley, the Kunzum La and follow the high summits south of the Spiti valley. A great number of thrust units have been defined in the Tibetan or Tethys zone of the Himalaya by different authors since the discovery of the Zanskar nappes by Bassoulet et al. (1980). The geometry of the North Himalayan nappe stack varies along strike and a lateral correlation of the defined units is difficult because of the numerous gaps in field observations. Thus, only a few major thrusts of the North Himalayan nappe stack are documented on the geological map and sections of this paper (Plate 1, 2 and 3). Between the Baralacha La and the Pare valley, the frontal thrusts are not continuous or cylindrical but succeed in an oblique en-échelon array

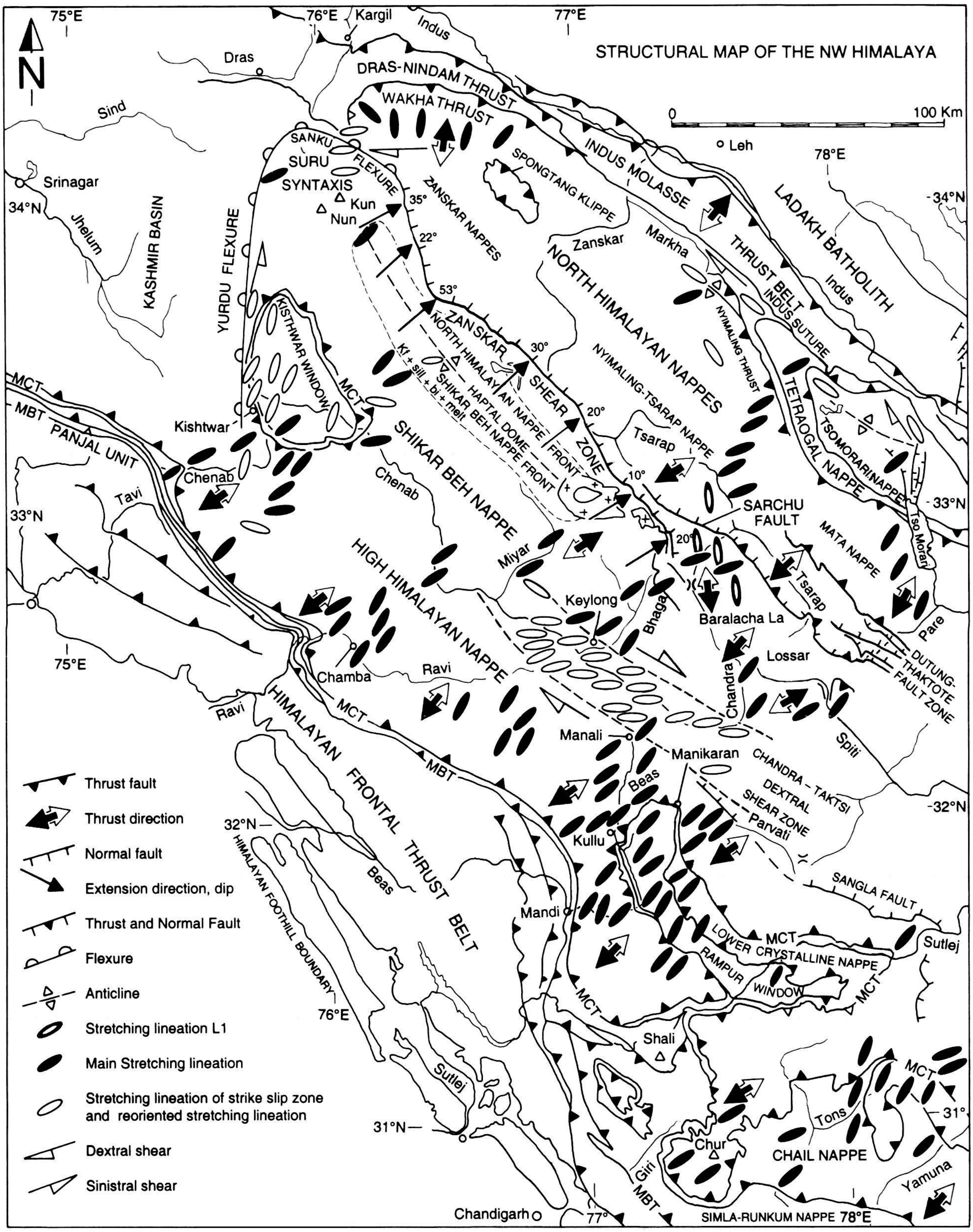

Fig. 6. Structural map of the NW Himalaya, illustrating the main thrusts, extensional faults, dextral and sinistral strike slip zones, stretching lineations and the shear sense deduced from mesoscopic shear criteria. This map represents a compilation of observations by Epard et al. (1995), Epard & Steck (in press), Gapais et al. (1992), Gilbert (1986), Gilbert & Merle (1987), Guntli (1993), Herren (1987), Honegger, (1983), Kündig (1989), Patel et al. (1993), Robyr (2002), Robyr et al. (2002), Spring (1993), Stäubli (1989), Steck et al. (1993, 1998, 1999), Stutz & Steck (1986), Stutz (1988), Vannay (1993), Vannay & Steck (1995) and Wyss et al. (1999).

(Plate 1, 3 and Fig. 6). A geometric feature that may be explained by the anti-clockwise rotation of India relative to Asia during collision, that resulted in a component of dextral transpression. In connection to the Nyimaling-Tsarap nappe of the Leh - Baralacha La transect and the Mata nappe of the Tso Morari-Spiti section, it has been demonstrated that the southwestern frontal imbricate thrust and fold structure, of a high (brittle) tectonic level, gradually pass to the NE and, with increasing tectonic depth, to a ductile shear zone characterised by recumbent folds (Fig. 7; Steck et al. 1993). It is important to note, that the Zanskar nappe stack has a thickness of up to 40 km (Fig. 14) and does not represent the thin-skinned fold and thrust belt with a single basal detachment and pop-up structures, proposed by Corfield & Searle (2000). The magnitude of the displacement on the SW-verging folds and thrusts of the Nyimaling-Tsarap nappe becomes gradually less from NE to SW, ending in absorption of the displacement at the SW front of the structure. The mean NE-direction of under thrusting of the Indian border below Asia is expressed by the generally SW-verging folds F_2 and F_3, NE-dipping axial surface schistosities S_2 and S_3, with a NE-dipping stretching lineation L_2 and L_3, and related top-to-the SW shear sense criteria (Fig. 6). Detailed structural work in the Baralacha La – Lingti valley region revealed an older generation of E-verging isoclinal folds F_1 with N-plunging fold axis parallel to an intense N-plunging stretching lineation L_1 (Fig. 6). Top-to-the S shear criteria in deformed conglomerates and the porphyritic Permian Yunam granite dyke to the S of Sarchu indicate a early N-directed underthrusting of India below Asia. A spectacular isoclinal F_1-fold hinge of Carboniferous Lipak marls surrounded by the white Muth quartzite and the red Thaple conglomerate is exposed N of Kenlung Serai on the Manali-Leh road (Fig. 7; Epard & Steck in press, sketch drawing by Maurizio Gaetani, Fig.7 in Baud et al. 1984). This F_1 recumbent fold is exposed on both sides of the Yunam valley S of Sarchu (Plate 1). In older publications, this structure has wrongly be attributed to a younger F_3 fold phase (Spring & Crespo, 1992, Spring, 1993, Steck et al. 1993); so structures F_1, F_2, S_1, S_2 and L_1, L_2 in Steck et al. (1993) become F_2, F_3, S_2, S_3 and L_2, L_3, and F_3, S_3 and L_3 become F_1, S_1 and L_1 in this paper (Epard & Steck, in press). The succession of the early N-directed, followed by the main NE-directed, underthrusting of India below Asia indicate an anticlockwise rotation of India relative to Asia during an early stage of continental collision. The SW-front of the SW-directed shear of the North Himalayan nappes is documented by rotated garnets in the Zanskar Crystalline zone between the Suru valley to the NW (Gilbert 1986, Vance & Harris 1999) and in the Zanskar and Kurgiakh valleys to the SE (Patel et al. 1993, Dèzes et al. 1999). High level tectonic structures of the N-Himalayan nappe front are exposed farther to the SE, in the Baralacha La axial depression of the Himalayan range. In this region, frontal imbricate thrusts pass gradually to a frontal fold belt with a frontal absorption of the SW-directed movements (Plate 3, Fig. 7). The highest unit of the North Himalayan nappes is the Spongtang ophiolitic klippe. Ophiolitic rocks

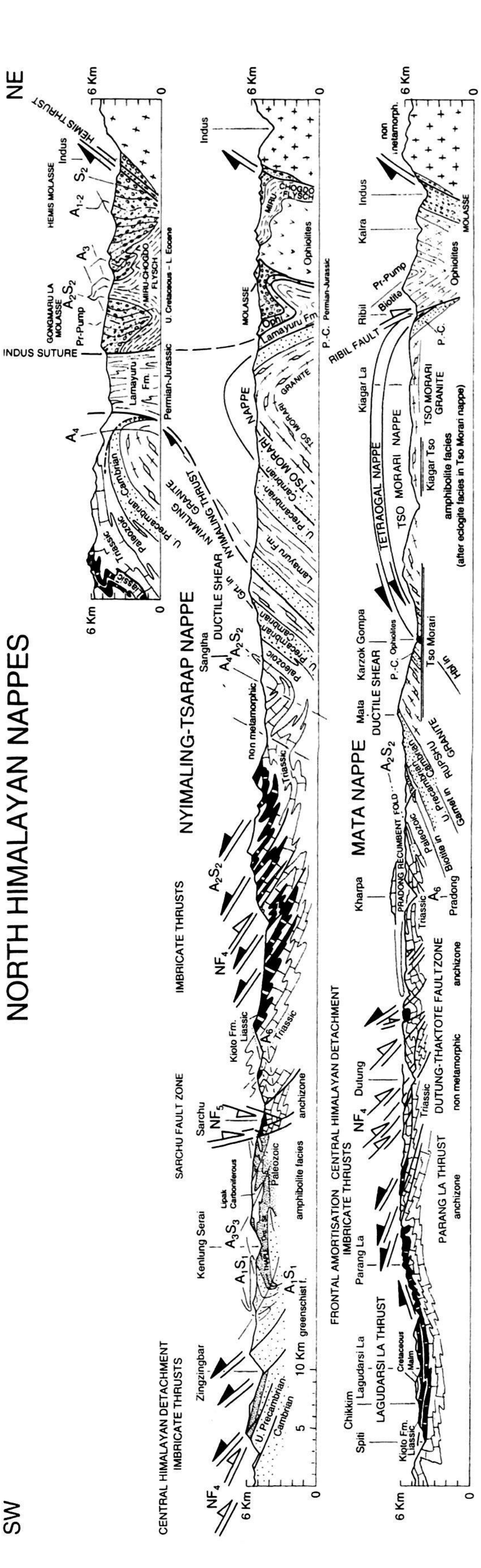

Fig. 7. Three geological sections of the North Himalayan nappes (Nyimaling-Tsarap, Tetraogal, Mata and Tso Morari nappes). The structures of the two S-N oriented and well studied Zingzingbar (Bhaga) – Martselang (Indus) and Chikkim (Spiti) – Tso Morari – Indus transects are projected into a SW-NE plane. Two phases of regional metamorphism are developed in the Tso Morari transect: a 55 Ma eclogite facies metamorphism with coesite restricted to the Tso Morari nappe, and a Middle and Late Eocene Barrovian metamorphism affecting the whole nappe stack (De Sigoyer et al. 1997, 2000, Epard & Steck, in press, Girard, 2001, Girard et al.1999, Mukherjee & Sachan, 2001, Spring et al. 1993, Steck et al. 1993, 1998).

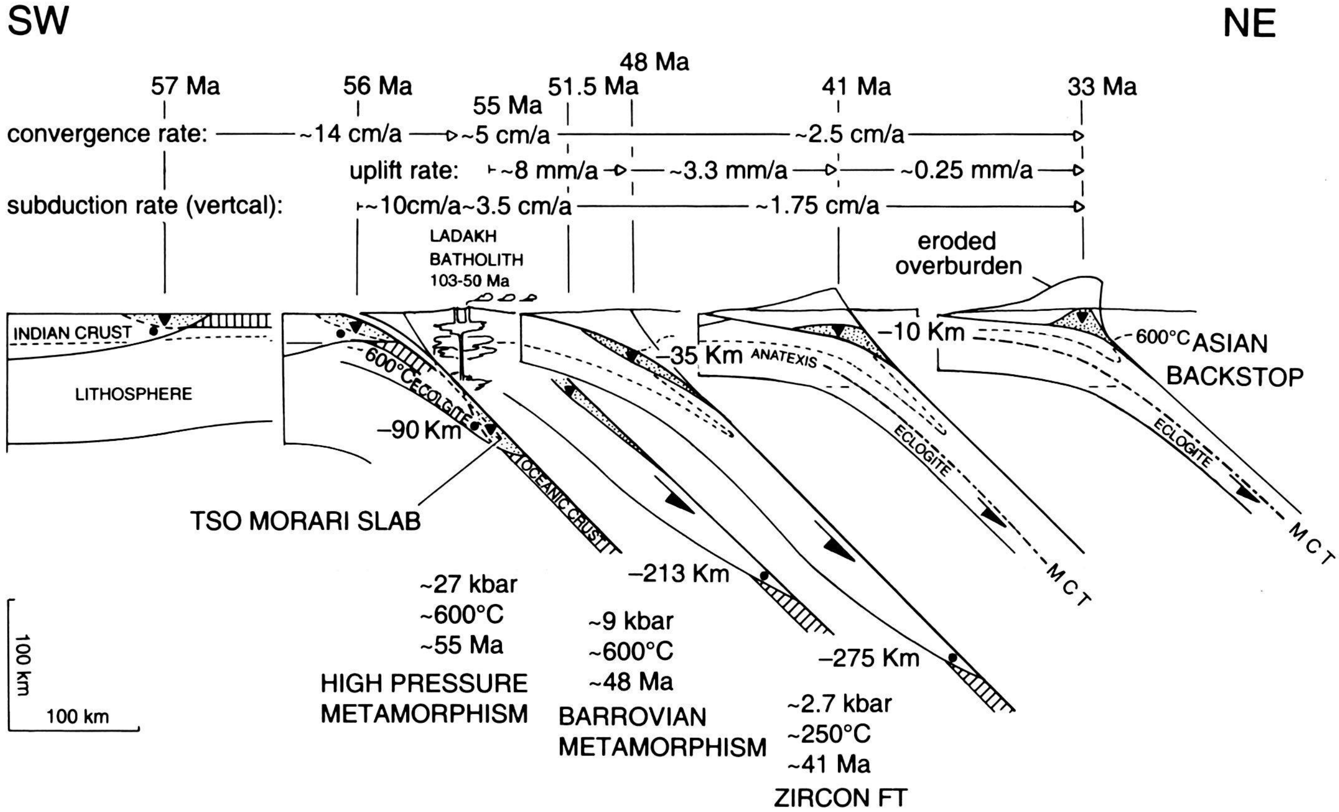

Fig. 8. A kinematic model of the continental collision, the Tso Morari slab subduction and extrusion and the formation of the North Himalayan accretionary wedge (North Himalayan nappes) that occurred between about 56 and 33 Ma during the under thrusting of the North Indian continental margin below Asia. The thermo-barometric and radiometric data are from De Sigoyer et al. (2000), Mukherjee & Sachan (2001), Schlup et al. (2003) and Weinberg & Dunlap (2000). Approximate convergence rates between India and the Indus suture are based on the studies of Patriat & Achache (1984) and the model of Tapponnier et al. (1986). A shortening of about 450 km between India and the Asian backstop is estimated for the period between 56 and 41 Ma.

from Ladakh were already mentioned at the end of the 19th and the Early 20th century (Lyddeker 1883, Mac Mahon 1901). However, modern studies of the Spongtang ophiolitic klippe started only in the nineteen-seventies with the opening of Ladakh to western geologists (Fuchs 1977, 1982, Bassoulet et al. 1980, 1983, Baud et al. 1984, Searle 1986, Reuber 1986, Reuber et al. 1987, Garzanti et al. 1987, Cannat & Mascle 1990, Searle et al. 1997, Corfield & Searle 2000). In the Spongtang region, the geological section 3 on Plate 2, was inspired mainly by an unpublished geological section (Michel Colchen, personal communication), which was partially reproduced and discussed in Cannat & Mascle (1990), and by observations from Corfield & Searle (2000). Michel Colchen's geological section appears, for a structural geologist, to be the most convincing interpretation when compared to the numerous propositions by the other above mentioned geologists working in the area. The Spongtang ophiolite, composed of mantle harzburgite, tectonically overlies a complex ophiolitic and sedimentary mélange with an Upper Campanian to Maastrichtian matrix, in turn overlying younger Turonian-Ypresian foraminiferal limestones of the Fatula, Kangi La, Dibling and Kesi formations and Lutetian Kong Slates of the Zanskar nappes (Reuber et al. 1987, Colchen & Reuber 1987, Corfield & Searle 2000). According to Corfield & Searle (2000) and Mahéo et al. (2000), this nappe stack results from an early phase of Late Cretaceous obduction onto the flysch sediments of the outer passive Indian margin followed by the final phase of Early Eocene emplacement on the Indian carbonate shelf. Mahéo et al. (2000) explain the obduction of the oceanic lithosphere exposed in the Spongtang klippe, the Nidar and Karzok ophiolites, with their similar geochemical signature of an immature arc, by a second zone of underthrusting in the Neo Tethys oceanic crust, situated to the south of the main subduction related to the Dras Arc and Ladakh batholith. About 100 km (>87 km) of crustal shortening in the detached N-Himalayan nappe stack was estimated by Steck et al. (1993) using a simple shear model. A similar shortening was estimated by Corfield & Searle (2000) for the Western Zanskar nappes. Not that this value is very different to the over estimations proposed by Mc Elroy et al. (1990) based on the wrong model of a thrust fan that is linked to a single basal detachment.

The Barrovian metamorphism (M3) of the North Himalayan nappe stack

A regional metamorphism ranging from non-metamorphic grade (diagenesis) to amphibolite facies grade, with staurolite-kyanite-sillimanite mineral assemblages, overprints the North Himalayan nappe stack. The highest grade rocks are exposed in the footwall of the Zanskar shear zone (Honegger et al. 1982, Gilbert 1986, Herren 1987a & b, Dezes et al. 1999, Searle et al. 1992, 1999, Vance & Harris 1999), S and W of Sarchu in the Yunam and Kamirup valleys (Srikantia & Bhargava 1982, Spring 1993, Epard & Steck in press) and in the Tso Morari dome (Thakur 1983a, Guillot et al. 1997, De Sigoyer et al. 2000, Girard 2001). The dry-melting muscovite+plagioclase+quartz = sillimanite+K-feldspar+melt isograde is reached only in the Zanskar Crystalline dome (Honegger et al. 1982, Patiño Douce & Harris 1998). A maximum P and T of about 11.8 kbar and 820°C were estimated in the Zanskar dome in the Gianbul valley (Dèzes et al. 1999, Robyr et al. 2002), and 9-10 kbar and 650-700°C in the western Tso Morari dome (De Sigoyer et al. 2000, Girard 2001). In the latter area, the isogrades cut obliquely through the tectonic contacts of the Tso Morari, Tetraogal and Mata nappes (Plate 4; Girard 2001). In the Tso Morari region, the medium pressure Barrovian overprint was dated between 48 and 45 Ma, 48 ± 2 Ma ^{40}Ar-^{39}Ar age on phengite, 47 ± 11 Ma Sm-Nd age on garnet-hornblende - whole rock and 45 ± 4 Ma Rb-Sr age on phengite-apatite - whole rock (De Sigoyer et al. 2000). Similar $^{40}Ar/^{39}Ar$ ages of ~51-47 Ma from white micas that grew along the foliation and stretching lineation in Triassic phyllites, are recorded from below the main N-Himalayan thrust in Tibet (Ratschbacher et al. 1994). Cooling below 300 °C, by uplift and erosion of the Tso Morari dome, started in the Tso Morari-Kiagar Tso region at ~40 Ma (zircon fission track ages, Schlup et al. 2003). It was followed 10 Ma later, 30 km farther to the W, in the highest grade sillimanite zone rocks of the Tso Morari dome (31.1 ± 0.3 Ma ^{40}Ar-^{39}Ar age on phengite, 29.3 + 0.3 Ma and 29.0 ± 0.4 Ma age on biotite, De Sigoyer et al. 2000). The data of De Sigoyer et al. (2000), Mukherjee & Sachan (2001) and Schlup et al. (2003) show that the subduction to a depth of over 90 km of the Indian granitic crust and exhumation occurred between 55 and 40 Ma (Fig. 8 & 9). These ages also constrain the period of emplacement of the N-Himalayan nappes. Gilbert (1986) was the first to recognise that the Miocene extensional Zanskar shear zone reactivated the older SW-directed frontal thrusts and deformed the high-grade metamorphic rocks below the Zanskar nappe stack. Recently, these observations were generally confirmed (Patel et al. 1993, Dèzes et al. 1999, Walker et al. 1999, Wyss et al. 1999). Using the Sm-Nd method, Vance & Harris (1999) dated the prograde growth of garnets in the kyanite zone rocks from the Suru region at 33-28 Ma. They relate garnet growth to the post-collisional thrusting south of the suture zone. The data of Vance & Harris (1999) suggest that the North Himalayan nappe front was still active at depth, in the stability field of garnet, some 33-28 Ma ago.

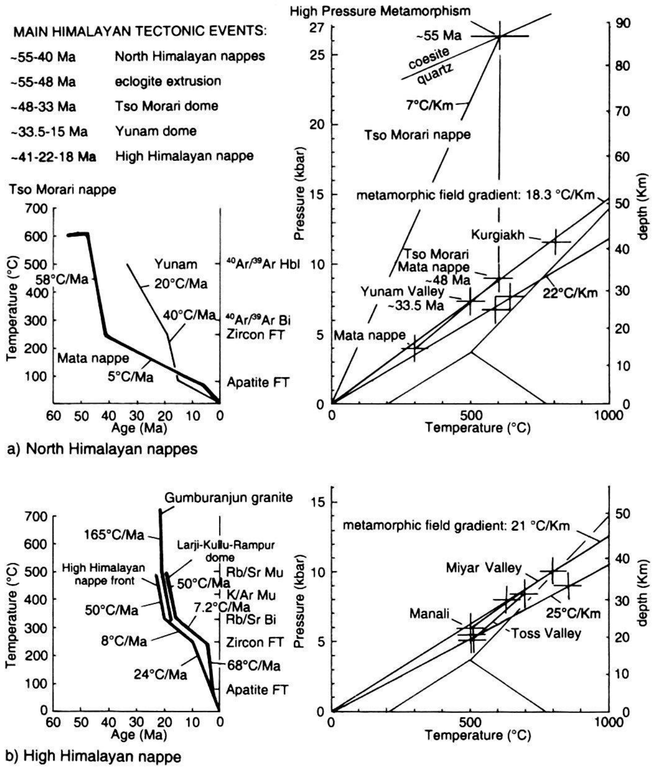

Fig. 9. Temperature-time and pressure–temperature paths of a) the North Himalayan accretionary wedge (North Himalayan nappes) and b) the High Himalayan nappe, based on data from Dèzes et al. (1999), De Sigoyer et al. (2000), Epard et al. (1995), Frank et al. (1977b), Mukherjee & Sachan (2001), Robyr et al. (2002), Schlup et al. (2003) and Spring et al. (1993a & b). Note that the cooling path of the Yunam dome, in the frontal part of the North Himalayan Nyimaling-Tsarap nappe (Fig. 9a), is related to the cooling path of the Zanskar crystalline doming (Gumburanjun granite cooling path on Fig. 9b). The cooling path from the Kullu Valley transect (Fig 9b) is similar to the data of Jain et al. (2000) from the same region and to the Umasi La Atholi transect in the Zanskar crystalline zone (Sorkhabi et al. 1997).

During the same period, the internal part of the North Himalayan nappe in the Tso Morari dome had already been uplifted by doming and had cooled below 300°C (Schlup et al. 2003). Prince et al. (2001) dated deformed Early Miocene leucogranites, derived from fluid-enhanced melting, of the High Himalaya in Garhwal. Sm-Nd garnet dating indicates a crystallisation age of 39 ± 3 Ma. We suggest that this metamorphism belongs to the M3 phase related to the N-Himalayan nappe stack.

Kinematics of the North Himalayan nappes

A kinematic model of the continental collision between India and Asia, as well as the formation of the North Himalayan accretionary wedge between 56 and 33 Ma, is proposed in Fig. 8. We suggest that the continental collision and subduction of the North Indian continental margin began ~ 56 Ma ago. Following the subduction of the Indian crust to a depth of ~ 90 km, at

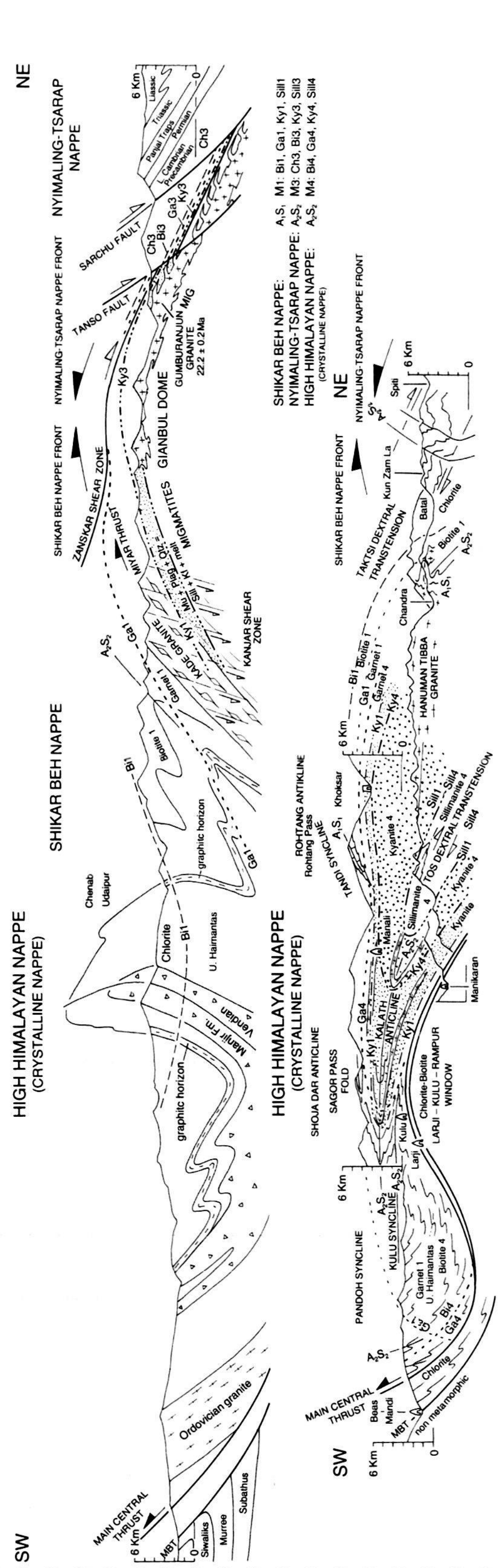

Fig. 10. Geological sections through the High Himalayan nappe (Crystalline nappe). Three phases of Barrovian regional metamorphism may be distinguished in the High Himalaya: the M1 metamorphism of the NE-directed Shikar Beh nappe stack, the M3 metamorphism of the SW-directed Nyimaling-Tsarap nappe stack both of which were transported and folded by the SW-directed High Himalayan nappe under high temperature conditions. The folded isograds of the M4 metamorphism of the High Himalayan nappe (Crystalline nappe) indicate a stage of crystallisation of the SW-extruding High Himalayan or Crystalline nappe. The upper profile corresponds to section 6 on Plate 2 (Steck et al. 1999, Fig. 2), Robyr et al. 2002). The lower profile, situated near section 7 of Plate 2 was constructed from the geological Kullu Valley transect (Epard et al. 1995, Fig. 2) and the Manikaran – Tos Valley – Kun Zam La – Spiti transect (Wyss et al. 1999, Plate 1 and Wyss 2000, Plate 1).

a subduction rate of ~10 cm/a, the granitic Tso Morari slab, with its eclogitic dikes and coesite (Mukherjee & Sachan 2001), was detached by ductile shear, squeezed out by tectonic compression between the North Indian plate and the Asian mantle wedge and moved by buoyancy forces (Chemenda et al. 1995) along the plane of subduction. The uplift from a depth of 90 km to 35 km in 7 Ma denotes an uplift rate of ~8 mm/a. Metamorphic fluids, derived from dehydration of antigorite and amphiboles of the subducted oceanic crust and mantle (Ulmer & Trommsdorff 1995), may have also facilitated the wet deformation and acted as lubricant for the extrusion of the Tso Morari slab. Guillot et al. (2000) suggest that serpentinites were formed on the border of the Asian mantle wedge above the subduction zone. These serpentinites may acted as a lubricant in the hangingwall of the Tso Morari slab. Some 48 Ma ago, the Tso Morari slab reached the base of the accretionary wedge, where it recrystallised along with the higher North Himalayan nappes, under amphibolite facies conditions (De Sigoyer et al. 2000, Girard 2000). This was followed between 48-33 Ma, by a phase of doming and NE-vergent backfolding in front of the Asian backstop formed by the Ladakh batholith (Schlup et al. 2003). The temperature-time and pressure-temperature path of the North Himalayan nappes is shown in Fig. 9a. The late Paleocene and Early Eocene age of the North Himalayan nappes is indicated by the Tso Morari and Mata nappe cooling curve; whereas the temperature-age path of the Yunam valley amphibolite facies rocks (between Kenlung Serai and Sarchu on Fig. 7) does not record the age of the North Himalayan nappes, but the later Oligocene to Miocene up-doming, related to the younger extrusion of the High Himalayan nappe.

The High Himalayan nappe or "Crystalline nappe"

The most spectacular structure of the Himalayan range is the "Crystalline nappe", named herein the High Himalayan nappe, which overthrusts the Lower Crystalline nappe and the Lesser Himalaya along the Main Central thrust (MCT). The name "Crystalline nappe" was introduced because in the central and eastern part of the Himalaya, this nappe is composed of high-grade amphibolite facies metamorphic rocks (Von Lóczy 1907). In the NW Himalaya, the degree of the Tertiary regional metamorphism varies between the prehnite-pumpellyite facies or anchizone and migmatites of the sillimanite + K-feldspar zone. The low grade metamorphic rocks at a high tectonic level are preserved in a central Chamba-Baralacha La E-W striking axial depression. The high grade rocks are exposed due to the two deeply eroded axial culminations of the Zanskar crystalline dome, with the Kishtwar tectonic window, and the dome structure around the Rampur window (or Larji-Kullu-Rampur window). The stratigraphic column of the High Himalayan nappe comprises sediments of a Late Proterozoic to Upper Cretaceous ages. The main metamorphism is of a regional Barrovian type, with a first phase M1 related to the early Shikar Beh nappe stack,

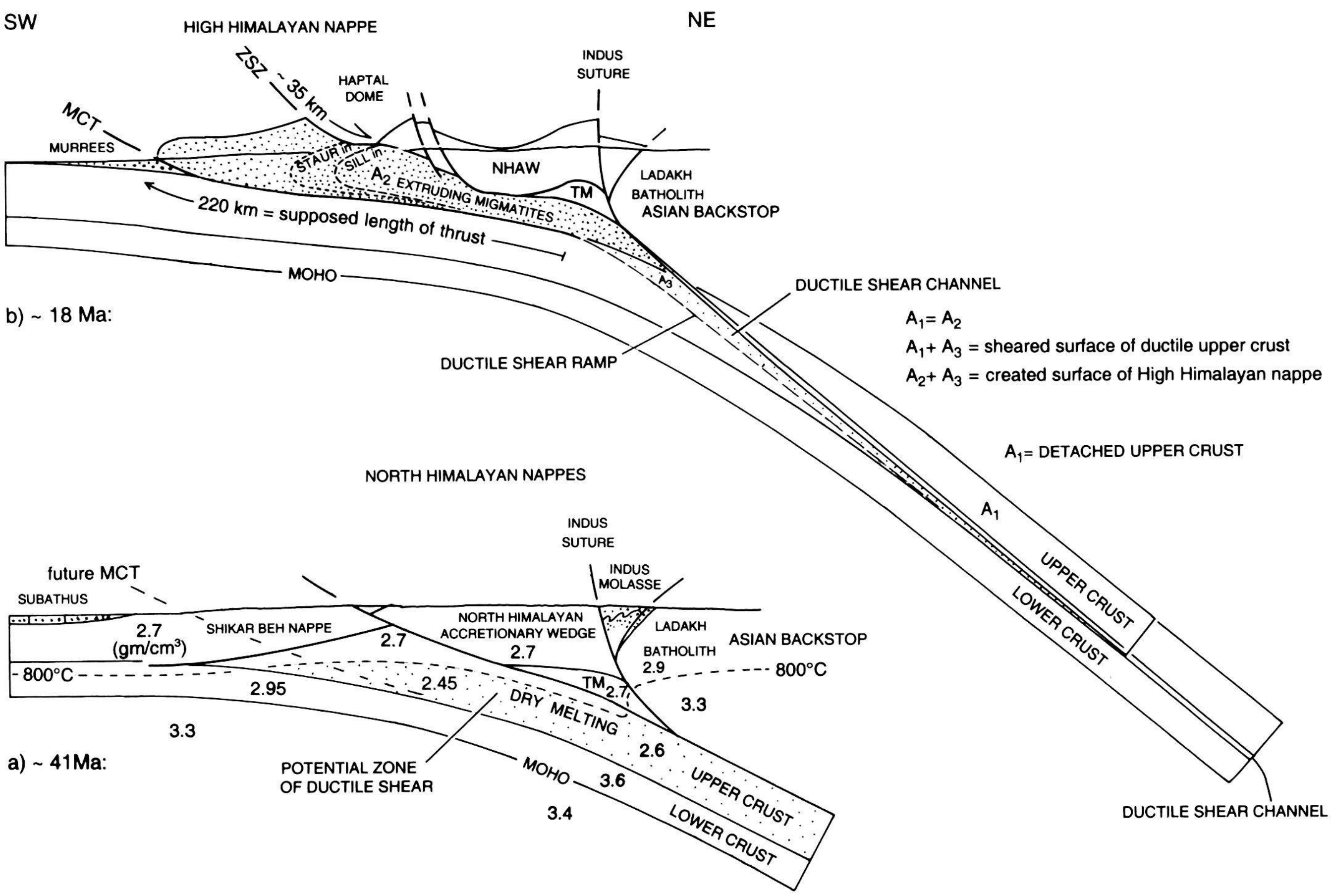

Fig. 11. Empirical model for the formation of the High Himalayan nappe. The lower section illustrates the formation of a dry melting (anatexis) zone in the subducted upper Indian crust below the 55-33 Ma old North Himalayan accretionary wedge and the Shikar Beh nappe stack. The numbers in Fig. 11a are rock and migmatite densities in gr/cm^3 after Bousquet et al. (1997) and calculated after Lange & Carmichael (1990). During further underthrusting, a ductile shear zone develops and the upper crust is detached by pure and simple shear while a part of the ductile crust is squeezed out along the base of the Main Central thrust (MCT) and along the Zanskar shear zone (ZSZ) on top of the High Himalayan nappe. The cooling of the extruding nappe is dated at 22-19 Ma (Dèzes et al. 1999). It is not possible, at present, to quantify the crustal shortening related to the formation of the High Himalayan nappe for lack of information on the displacement on the MCT and the degree of internal deformation within the nappe (Hauck et al. 1998). The supposed crustal shortening is of ~220 km.

a second phase M3 related to the North Himalayan nappe stack and a third phase M4 of crystallisation related to the emplacement of the High Himalayan nappe. Zones of low-pressure contact metamorphism are developed around Ordovician granite intrusions. SW-verging folds, a generally NE-plunging stretching lineation, and top-to-the SW shear structures indicate that the High Himalayan nappe, with the Main Central Thrust at its base, was formed by ductile detachment of the upper sedimentary cover of the Indian plate during underthrusting below Asia (Plate 3 & Fig. 6, 10, 11). The magnetude of displacement on the MCT is at least 100 km, which corresponds to the distance between the frontal MCT and the internal border of the Rampur window, but more probably it is in the range of some hundred kilometres. An exact estimation is not possible (Fig. 11; Hauck et al. 1998, Hodges, 2000).

The tectonic units of the Lesser Himalaya

The tectonic units of the Lesser Himalaya (Lesser Himalayan Tectogen of Srikantia & Bhargava, 1998) are characterised by a similar Proterozoic - Lower Cambrian sedimentary sequence. In the Shali-Deoban and Simla domains, these rocks are unconformably overlain or transgressed by the Thanetian-early Ypresian Kakara Formation and the Ypresian-Lutetian Subathu Formation (Auden 1934, Gansser 1964, Srikantia & Bhargava 1998). The Chail nappe exposed in the Larji-Kullu-Rampur Window consists of Proterozoic sediments, intruded by or overlying the 1'800 ± 13 Ma Rampur basalt and an 1'840 ± 16 Ma metarhyodacite (Miller et al. 2000). No igneous rocks of this Proterozoic age have been observed in the High Himalaya domain. The nappes of the Lesser Himalaya are overthrusted along the Main Boundary thrust (MBT) onto the

Pliocene to Pleistocene Upper Siwaliks, implying that the MBT at the base of the Lesser Himalaya is an active thrust (Tab. 1). Meigs et al. (1995) suggest an initial displacement along >100 km of the Main Boundary thrust prior to 10 Ma. They argued that sediment-accumulation curves of the Himalayan foreland basin show an accelerated accumulation rate and inferred that subsidence began at ~11 Ma and that apatite fission-track ages and track-length measurements indicate rapid cooling below ~105°C between 8 and 10 Ma.

The Lower Crystalline nappe (Frank & Fuchs 1970)

The name Lower Crystalline nappe was introduced by Frank & Fuchs (1970) for a sheet of metamorphic rocks at the base of the MCT in Nepal. Guntli (1993) and Frank et al. (1995) used the same name for a tectonic sheet at the base of the High Himalayan or Crystalline nappe overlying the Chail nappe of the Lesser Himalaya exposed in the Kishtwar and Rampur windows and in front of the High Himalayan nappe. The Lower Crystalline nappe is a strongly deformed tectonic unit, with basal mylonites similar to the High Himalayan nappe above the MCT, but within rocks with Lesser Himalayan stratigraphic affinity. Thöni (1977) named this unit, comprising metasediments and metagranites exposed in the Larji-Kullu-Rampur window, the Bajaura nappe. According to these authors, the Lower Crystalline nappe, or Bajaura nappe, is a basal slice of the Crystalline nappe. However, the rocks of the Bajaura Fm. are quite different from the overlying sediments and Ordovician granites of the High Himalayan nappe and they overlie or are intruded by Proterozoic 1.8-1.9 Ga intrusives (Miller et al. 2000). The lithology of the Lower Crystalline nappe resembles rocks in the Chail nappe. For this reason the Lower Crystalline nappe is here considered as a unit of the Lesser Himalaya (Fig. 4). Like the High Himalayan nappe, the Lower Crystalline nappe represents a sheet of Tertiary metamorphic rocks that has been thrust over low grade metamorphic units producing a reverse metamorphism like in the High Himalayan nappe. This suggests that this highest unit of the Lesser Himalaya was metamorphosed in an internal position before it was thrusted over a distance of more than 100 km in a SW-direction above other Lesser Himalayan units.

The Barrovian metamorphism (M4) of the High and Lesser Himalayan nappes

In contrast to the North Himalayan nappe stack, where deeper tectonic units are generally characterised by high grade metamorphism, in the High Himalayan nappe as well as in Lower Crystalline thrusts, in the Kishtwar window, in the lower Chenab valley to the W of Doda (Guntli 1993) and in the Rampur window (Vannay & Grasemann 1998, 2001, Vannay et al. 1999), high grade metamorphic rocks have been over trusted above lower grade metamorphic rocks (Plate 4). The fact that topographically higher regions expose rocks with a higher grade of metamorphism was discovered in 1878 by von Lóczy (1907, Gansser 1964). This reverse or inverted metamorphism in the High Himalayan nappe overlying low grade metamorphic rocks lead to the introduction of the name "Crystalline nappe". In the Himalaya, the high-grade metamorphism has a Tertiary age and overprints, with variable grade, sedimentary rocks of the Tethyan stratigraphic sequence of Late Proterozoic to Early Eocene age. Hence, the Tertiary crystalline rocks cannot be described as the basement for the sediments of the Tethys domain. The name "Crystalline nappe" is equally misleading, because in the Chamba basin of the NW Himalaya, this tectonic unit is composed of very low-grade sediments passing gradually to the high-grade amphibolite facies and migmatite zone rocks of the Zanskar crystalline zone to the N (Plate 4). Considering of these relations indicate that the neutral term High Himalayan nappe should be introduced as an alternative name to the more than 100 year old traditional name Crystalline nappe (observed in 1878 by von Lóczy 1907, Gansser 1964). The P-T conditions and the age of the Barrovian regional metamorphism of the High Himalayan nappe have been studied by numerous scientists; they are in the NW Himalaya Frank et al. (1973, 1977), Guntli (1993), Honegger (1983), Honegger et al. (1982), Kündig (1989), Patel et al. (1993), Pognante & Lombardo (1989), Searle (1986), Searle et al. (1988, 1992, 1999), Stäubli (1989), Epard et al. (1995), Stephenson et al. (2000, 2001), Thöni (1977), Vance & Harris (1999), Vannay & Grasemann (2001), Vannay et al. (1999), Wyss (1999, 2000), Wyss et al. (1999), Prince et al. (2001). On our metamorphic map (Plate 4), it was not possible to distinguish the metamorphism M4 of the High Himalayan nappe from the M3 metamorphism of the North Himalayan nappe front in the Zanskar region and the older M1 metamorphism of the NE-directed Shikar Beh nappe stack. Steck et al. (1999), Robyr et al. (2002) and Robyr (2003) demonstrated that the Barrovian regional metamorphism M1 of the southern part of the Zanskar metamorphic dome was generated by the Shikar Beh nappe stack. Patel et al (1993), Spring et al. (1993), Steck et al. (1993), Dezes et al. (1999), Vance & Harris (1999) and Walker et al. (1999) showed that the Barrovian metamorphism M3 of the northern flank of the Zanskar Crystalline dome was related to the North Himalayan nappes. So this Zanskar High grade metamorphic zone was formed during older orogenic events, extruded and transported during the emplacement of the younger High Himalayan nappe (Fig. 11). In the Beas (Kullu) valley transect, Epard et al. (1995) demonstrated that in this region the older transported M1 metamorphism of the Shikar Beh nappe stack was over printed by a younger M4 Barrovian metamorphism of the High Himalayan nappe. Likewise, the younger isograds of the High Himalayan nappe metamorphism are passively folded and overthrusted to the SW (Fig. 10, 11). In the deeper high-grade M4 metamorphic zone of the Toss valley transect east of Manikaran (Parvati valley), only relicts of the M1 metamorphism are recognizable (Wyss 2000). All these observations indicate that the tectono-metamorphic processes in the High Himalayan nappe are complex and that the main metamorphism of the High Himalayan or "Crystalline" nappe corresponds to a pre-existing transported and folded zone of metamorphic rocks (Fig. 10, 11). The age of cooling by extrusion and erosion of the High Himalayan nappe is constrained by mica ^{40}Ar-^{39}Ar cooling ages of 24 to less than 18 Ma (Fig. 9b; Frank et al. 1977b, Hubbard & Harrison 1989, Harrison et al. 1992, Hodges et al. 1996, Schlup et al. 2003). The deeper units of the Lesser Himalaya (Chail nappe, Simla-Runkun nappes, Panjal unit) are overprinted only by a low-grade greenschist facies metamorphism M4 with chlorite and sometimes biotite and stilpnomelane assemblages, resulting from the overthrusting and subsequent heating by the High Himalayan and Lower Crystalline nappe.

The Zanskar shear zone, the Miocene leucogranites, the Gianbul dome, the Kanjar shear zone and related retrograde metamorphism (M5)

A zone of NE-directed low-angle normal fault extension has been reported by several geologists in the central part of the Himalayan range (Burg & Cheng 1984, Burchfiel & Royden 1985, Burchfiel et al. 1992, Herren 1985, 1987a & b, Gilbert 1986, Searle 1986, Pêcher 1991, Patel et al. 1993, Dèzes et al. 1999). It was named the South Tibetan detachment by Burchfiel & Royden (1985), the North Himalayan detachment by Pêcher et al. (1991) and the Zanskar shear zone by Herren (1985, 1987a, b) in the NW Indian Himalaya. Due to its central position in the Himalayan range, Epard & Steck (in press) named it the Central Himalayan detachment. The Zanskar shear zone reactivated the older thrust faults of the frontal imbricate structure of the North Himalayan nappes (Plate 3, Fig. 6, 7; Gilbert 1986, Patel et al. 1993, Dèzes et al. 1999, Robyr et al. 2002, Epard & Steck in press). In the Zanskar region, the extension is dated by the synkinematic intrusion of the Gumburanjun leucogranite intrusion at 22.2 ± 0.2 Ma (U-Pb monazite age), which cooled below 300°C ~19.8 ± 0.1 Ma (^{40}Ar-^{39}Ar mica age) ago (Fig. 8b; Dèzes et al, 1999). These original data were later confirmed by Walker et al. (1999). These ages corroborate the data from the Nepal Himalaya (Burchfiel et al. 1992, Hodges et al. 1996). The NE-directed extension on the Zanskar shear zone occurred in the same period as the SW-directed extrusion of the High Himalayan nappe on the MCT (Frank et al. 1977b, Hubbard & Harrison 1989). Based on a combined structural and thermo-barometric study, Dèzes et al. (1999) estimated a normal displacement on the supposed 20° NE-dipping Zanskar shear zone of 35 km. Dèzes (1999) also observed a retrograde M5 metamorphic evolution in the Zanskar shear zone, starting from the staurolite-kyanite zone of the pre-existing Barrovian M2 metamorphism of the North Himalayan nappes, with the successive crystallisation of sillimanite, cordierite, andalusite and margarite. Robyr et al. (2002) and Robyr (2003) observed a similar retrograde metamorphic evolution in the Khanjar shear zone, on the southern limb of the uplifted and eroded Gianbul dome: the staurolite-kyanite-mica assemblage of the M1 Shikar Beh nappe metamorphism is successively replaced by sillimanite, cordierite and andalusite.

In the Zanskar area, the ductile Zanskar shear zone, about 2 km wide, forms the boundary between the high-grade rocks of the Zanskar crystalline zone to the south and the very low-grade rocks of the Zanskar nappes (North Himalayan nappes) to the north. From the Kurgiakh valley and the Gumburanjun leucogranite intrusion to the southeast, the regional metamorphic grade decreases rapidly towards the Chamba-Baralacha La axial depression in the Himalayan range (Plate 1 and 4). At a high tectonic level, the Zanskar shear zone gradually fans out forming the more discrete low-angle extensional faults of the Tapachan fault zone (Epard & Steck, in press). Further east, with the en échelon position of the frontal thrusts of the N-Himalayan nappes, the extensional low-angle normal faults occur again in a similar en échelon position (Plate 3, Fig. 5; Steck et al. 1998). Girard et al. (1999, 2001) showed that the Lachalung La and Dutung-Thaktote fault zones form the contact between anchi- and non-metamorphic rocks. An extensional displacement of more than 15 km was estimated for the Dutung-Thaktote normal fault fan (Fig. 6 and 14).

The history and origin of the metamorphism of the High Himalayan or "Crystalline" nappe

The famous reverse metamorphism in the High Himalayan or "Crystalline" nappe was discovered in 1878 by von Lóczy (1907), Gansser (1964). Most new observations in the NW Himalaya have arrived at a conclusion that the high grade metamorphic rocks of the so-called "Crystalline nappe" were extruded with the MCT at its base and the South Tibetan detachment at its roof (Burchfiel & Royden 1985, Searle & Rex 1989, Pêcher et al. 1991, Hodges et al. 1992, Chemenda et al. 1995, Grujic et al. 1996). Based on observations in the Sutlej river transect, Grasemann et al. (1999) and Vannay & Grasemann (2001) demonstrated that the emplacement of the Crystalline nappe occurred through a general shear extrusion, a combination of pure and simple shear. All these kinematic models do not answer the fundamental question concerning the origin of the metamorphism. An answer to the question of the origin of the transported metamorphism of the Crystalline nappe of the NW Himalaya is found in the work of Epard et al. (1995), Patel et al. (1993), Vance & Harris (1999), Dezes et al. (1999) and Prince et al. (2001). Epard et al. (1995) demonstrated that the metamorphism in the Kullu valley was created by heating at deep levels in the early Shikar Beh nappe stack and that these still hot metamorphic rocks (M1) where then overthrusted on the MCT during the extrusion of the Crystalline nappe. This observation is confirmed for the Miyar valley in the southern Zanskar Crystalline zone, where the main regional Barrovian metamorphism is related to the NE-directed structures of the Shikar Beh nappe (Steck et al. 1999, Robyr et al. 2002, Robyr 2003). Patel et al. (1993), Vance & Harris (1999) and the Dèzes et al. (1999) demonstrated that the metamorphism of the northern Zanskar crystalline zone is synkinematic to the frontal thrust of the N-Himalayan nappes and due to the emplacement of this nappe.

Vance & Harris (1999) dated the synkinematic crystallisation of garnet from the Suru region at 33 Ma by the Sm-Nd method. Prince et al. (2001) dated the crystallisation of early Miocene leucogranites of the High Himalaya at 39 ± 3 Ma by the Sm-Nd garnet method. We suggest that this anatexis belongs to the M3 metamorphism of the N-Himalayan nappe stack. These ages are 10-15 Ma older then the extrusion of the High Himalayan nappe. In conclusion, there exists a major difference between the Barrovian metamorphism M4 of the High Himalayan nappe and the Barrovian metamorphism M3 of the N-Himalayan nappes. In the North Himalayan accretionary wedge, the Barrovian regional metamorphisms M3 resulted from the crustal thickening and subsequent heating of the nappe stack. Whereas in the High Himalayan nappe, rocks heated and derived from below the North Himalayan nappe stack (M3 metamorphism), in the southern Zanskar Crystalline rocks in the Kullu transect, as well as from the southern Shikar Beh nappe stack (M1 metamorphism), were thrusted over the cold rocks of the Lesser Himalaya producing a new nappe stack of hot rocks with a second Barrovian Metamorphism M4 (Fig. 10).

The progressive emplacement of the High Himalayan nappe may be divided in three successive phases (Fig. 11):

1. The first phase corresponds to the underthrusting of the Indian plate below the N-Indian accretionary wedge with the ductile shear detachment of the upper Indian crust along a pre-existing zone of dry melting below the N-Himalayan and Shikar Beh nappe stacks. This early phase followed the formation of the N-Himalayan accretionary wedge at least after 41 Ma.
2. In a second phase, the underthrusting was progressively accompanied by the extrusion of the Crystalline nappe with its detachment in the hanging wall from the N-Himalayan accretionary wedge, by reactivation of its frontal imbricate thrusts (Dèzes et al. 1999, Vance & Harris 1999, Vannay & Grasemann 2001). The extruding migmatites of the High Himalayan nappe sheet may be considered as a ductile material with a physical behaviour similar to a highly viscous Newtonian fluid. The driving force acting on the ductile shear channel, responsible for the channel flow type extrusion (Turcotte & Schubert 1982) may be of two types: a buoyant force and a compressional one between the more rigid Asian backstop, comprising the Ladakh batholith and the Asian upper mantle wedge, and the subducting elastic Indian lithosphere. Beaumont et al. (2001) and Grujic et al. (2002) have suggested that the gravity difference (buoyant force) between the high temperature migmatites and the colder crustal rocks of the orogenic lid and the subducting Indian plate, represents the driving force. The main difference between our model (Fig. 11) and the Beaumont et al. (2001) and Grujic et al. (2002) model lies in the localisation of the migmatite zone below the North Himalayan accretionary wedge and not below Tibet. This second phase, responsible for significant uplift and erosion of the High Himalaya, is dated at 22-18 Ma (Dèzes et al. 1999).
3. In a third phase, the formation of the Gianbul and Haptal domes of the Zanskar crystalline zone developes as a zone of up warping at the frontal edge of the North Himalayan nappes and above a reduced section of the subducting channel (Robyr et al. 2002). Strong denudation of the growing High Himalayan range may also have assisted the extrusion of the High Himalayan nappe.

The Subhimalaya

The geological map of the Subhimalaya, or the Himalayan frontal thrust belt (Plates 1 and 3), composed of Tertiary molasse type sediments, was compiled from data in Raiverman et al. (1983). The Tertiary sediments reach their maximal thickness of 7'000-8'000 m at the inner border of the Indus-Ganga basin, below the Lesser Himalayan Main Boundary thrust front (MBT) (Plate 2 and Fig. 12; Raiverman et al. 1983, Burbank et al. 1996, De Celles et al. 1998a, Powers et al. 1998). The thick sediment accumulation is caused by the flexural loading of the Indian margin since the latest Paleocene, the time of the India-Asia continental collision and the beginning of the detachment and accretion of the N-Himalayan nappes from the upper part of the under thrusted Indian crust. According to DeCelles et al. (1998a), the stratigraphic sections of the Subhimalaya of Pakistan, Northern India and Nepal are very similar and will be described together.

The Subathu Formation (Valdiya 1980)
(Thanetian-Lutetian)
(synonyms: Subathu Formation in N India, Valdiya 1980; Kohat or Balakot Formation in Pakistan, Bossart & Ottiger 1989; Bhainskati Formation in Nepal, DeCelles et al. 1998a)

Srikantia & Bhargava (1998) note, "During the period, after the Lower Cambrian Tal sea until the advent of Paleogene transgression, the Lesser Himalaya Tectogen of Himachal Pradesh experienced a phase of non-deposition and remained a positive area; though in some sectors of this tectogen there was minor incursion of Permian sea as in the "Autochthonous fold belt" bordering the Paleogene Subathu-Murree zone of Jamu and Kashmir, Darjeeling foothills and Arunachal sub-Himalaya and also in Salt Range of Pakistan". During latest Paleocene (Thanetian) to Middle Eocene (Lutetian) time, sandstones, siltstones, mudstones and nummulitic limestones, up to 1'500 m thick, of the Subathu Formation were deposited in a shallow marine environment on the subsiding Precambrian Indian crust south of the Neo-Tethys carbonate shelf, as shown by a palynoflora and a marine fauna of large foraminifera (Raiverman et al. 1983, Karunakaran & Ranga Rao 1976, Mathur 1978, Bossart & Ottiger 1989). Subathu equivalents were deposited also on the Riphean Shali Fm. of the Lesser Himalaya, but are missing, probably due to erosion, in the High Himalaya (Gansser, 1964). Bossart and Ottiger (1989) described the Kohat Formation of Pakistan as a relatively monotonous succession of fining upward cyclothems, with marls intercalations that contain nummulites and assi-

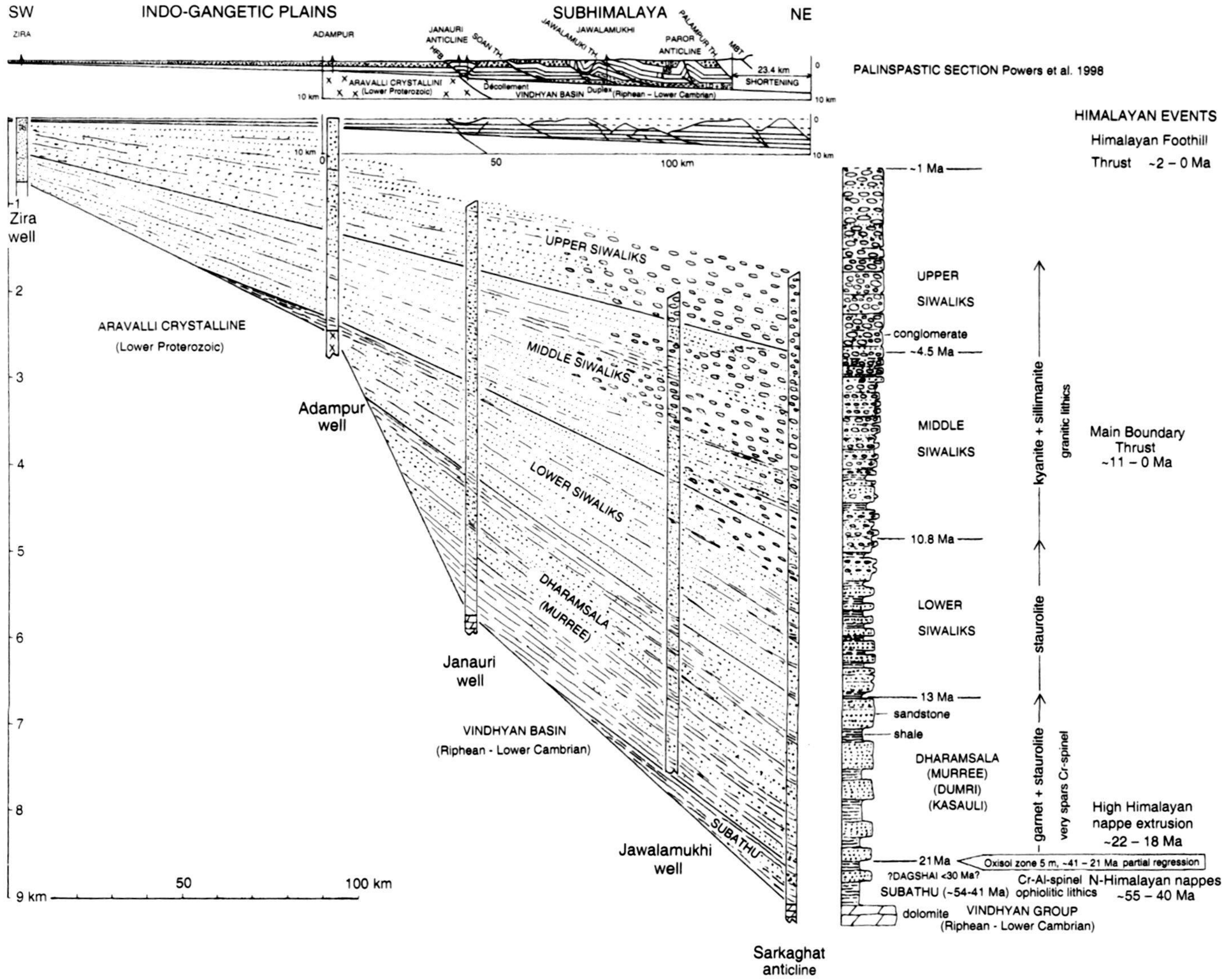

Fig. 12. Palinspastic section of the Subhimalaya molasse basin of the Kangra re-entrant based on surface geology, oil-well seismic reflection data, magnetostratigraphy, sedimentological observations and a balanced and restored cross section by Raiverman et al. (1983, 1993), Lyon-Caen & Molnar, (1985), Powers et al. (1998), completed by sedimentological and magnetostratigraphic data from the Nepal Himalaya by DeCelles et al. (1998a & b) and Najman & Garzanti (2000). The illustrated stratigraphic section of the N Indian Subhimalaya is very similar to the East Pakistan and Nepal sections (DeCelles et al. 1998a).

lines. Modal analysis of the Subathu sandstones shows that these sediments were derived from a mixture of sedimentary, volcanic, and ophiolitic protoliths. Rock fragments of felsitic volcanites, chert, serpentine and rocks with high-Al to high-Cr chromian spinel indicate a northern ophiolitic source and do not favour any contribution from the Deccan Traps (Fig. 12). The Subathu shallow marine sediments testify to the erosion of the obducted oceanic crust and immature arc, from the Spongtang klippe in Zanskar (Reuber et al. 1992, Mahéo et al. 2000) and the ophiolitic klippen of the Amlang La - Kiogar region of the Kumaon Himalaya (Heim & Gansser 1939, Gansser 1964) that overlies the Eocene North Himalayan nappe stack. This detrital sedimentation is contemporaneous with the deposition of the marine fluvio-deltaic redbeds of the Early Eocene (Ypresian and Lutetian) Chulung La and Kong formations that overlie the Late Cretaceous to Paleocene nummulitic limestones of the continental shelf exposed in Zanskar (Garzanti et al. 1987). These Paleocene sediments of the N Indian shelf record emplacement of ophiolitic rocks in the North Himalaya (Garzanti et al. 1987).

The Murree Formation (Wynne 1877, Shah 1977, Bossart & Ottiger 1989)
(Synonyms: Dagshai and Kasauli Formations, Valdiya 1980, Najman et al. 1993, and Lower and Upper Dharamsala Formations in N India (White et al. 2002); Dumri Formation in

Nepal, Sakai 1983, DeCelles et al. 1998a; Murree Formation s.s. in Pakistan, Shah 1977, Bossart & Ottiger 1989) (Oligocene? -early Miocene)

The name Murree Formation was originally used for molassic red and green siltstones and shales of latest Paleocene to Miocene age, that crop out through out Pakistan (Wynne 1877). In the Murree Hill station and in the Hazara-Kashmir syntaxis of Pakistan, these sediments are Eocene. These shallow marine and tidal deposits are the lateral equivalents of the Kohat and Subathu marls and limestones (Bossart & Ottiger, 1989). The Stratigraphical Committee of Pakistan decided that only the early Miocene continental shales and siltstones of the Kohat-Potwar Province of Pakistan should be asigned to the Murree Formation s.s. (Shah 1977, Bossart & Ottiger 1989), the definition used in this paper. However, it is quite possible that some Murree sediments on our map (Plate 1), drawn after Raiverman (1983), belong to the Murree group, as defined by Wynne (1877).

The top of the Bhainscati Formation in Nepal is marked by a distinctive, mappable 15 m-thick zone of oolitic ironstone beds and hematitic siltstone passing gradually upward into a ~5 m-thick red-and-white mottled paleosol (oxisol zone on Fig. 12; Sakai 1983). This zone, documented through Nepal and Northern India, is related to an Oligocene regression in N India lasting 15 m.y. (Valdiya 1980, Najman et al. 1993, DeCelles et al. 1998a, Najman & Garzanti 2000). The basal sandstones of the Dumri Formation overlie, with a sharp and irregular contact, the older paleosol and ironstones. The Murree Formation is composed of clastic, molasse-type sediments, mainly red and green shales, graywackes and subordinate conglomerates of an alluvial facies that is more than 2500 m thick (Shah 1977, Bosshart & Ottiger 1989, Powers et al. 1998, Najman & Garzanti 2000). The age of the continental Murree Formation and its equivalents is constrained by magnetostratigraphy and detrital mica Ar-Ar ages (White et al. 2002). Bossart & Ottiger (1989) and DeCelles et al. (1998) proposed an early Miocene age for these rocks. The Oligocene period of non-deposition was questioned by Bhatia & Bhargava (2002) through records of palinofossils in the Dharamsala sediments of a late Eocene-Oligocene age. Their observations may suggest that lake sediments were deposited in local basins during the period of Oligocene regression and were later reworked. The Oligocene time span of non-deposition in the Indian fore land basin corresponds to the a similar time interval between the emplacement of the North Himalayan nappes between 55 and about 40 (–33?) Ma (Vance & Harris 1999, Schlup et al. 2003) and the late thrusting and extrusion of the High Himalayan nappe or "Crystalline nappe", between about 22 and 18 Ma (Frank et al. 1977, Hubbard & Harrison 1989, Dèzes et al. 1999, White et al. 2002, Schlup et al. 2003). The deposition of the Dagshai Formation, corresponding to the lower part of the Murree Formation between <30 and 21-17 Ma, indicates erosion of the rapidly exhumed metamorphic slab of the High Himalayan nappes on the Main Central thrust between 22-18 Ma. Detrital modes show that the Dagshai Formation was predominantly derived from very low-grade metamorphic rocks, followed by predominant increasing grade metapelitic lithic grains, such as garnet and staurolite, and rare to negligible volcanic and ophiolitic detritus in the main Dagshai Formation (Najman & Garzanti 2000). The Kausali Formation is up to 1300 m thick and composed of greenish grey sandstone with minor claystone dated by early Middle Miocene plant fossils (Fiestmantel 1882). These fluvial sandstones are characterised by abundant metamorphic lithic fragments, detrital micas and zoned amphibolite facies grade garnets and staurolite. The increase in metamorphic lithic fragments illustrates the strong erosion of the rapidly exhumed High Himalayan metamorphic rocks (Najman & Garzanti 2000). For the periode of 21-13 Ma, about 2000 m of shales and sandstones of the Murree Formation were deposited with a mean sedimentation rate of 0.24 mm/a. In the Kangra reentrant of NW India, the southern shore of the Dharamsala (Murree) basin is formed by the over 1500 m high Aravalli crystalline range (Fig. 12; Lyon-Caen & Molnar 1985, Powers et al. 1998).

The Siwaliks or Siwalik Group (Srikantia & Bhargava 1998, Tandon 1991
(synonym: Ganga Molasse)
(Miocene-Present)

The Siwalik Group is over 5'600 m thick and consitsts of a coarsening upwards sequence. It is traditionally subdivided in three members, the 13-10.8 Ma old Lower Siwalik Member, up to 1'800 m thick, composed of an alternation of sand- and claystone with minor pebble horizons, the 11–4.5 Ma old Middle Siwaliks, up to 2'100 m thick, composed of sandstone with minor claystone and conglomerates closer to the MBT and the 4.5-1 Ma old Upper Siwalik Member, over 2'300 m thick, consisting of conglomerates and an increasing proportion of sandstone, away from the MBT (Fig. 12; Raiverman et al. 1983, DeCelles et al. 1998a & b, Harrison et al. 1993, Powers et al. 1998). The heavy minerals studied in the Nepal Himalayan foreland are characterised by a distinctly higher grade suite of metamorphic minerals, like kyanite and sillimanite and granitic lithics derived from unroofing of medium- to high-grade rocks of the High Himalaya metamorphic zones (Chaudri 1972, Parkash et al. 1980, DeCelles et al. 1998, Najman & Garzanti, 2000).

The 13-1 Ma, late Miocene to Pleistocene age of the Siwalik Formation is constrained by magnetostratigraphy based on the time scale of Cande & Kent (1992, Johnson et al. 1983, Rangao et al. 1988, Appel et al. 1991, Harrison et al. 1993, Meigs et al. 1995, Powers et al. 1998), volcanic ashes of 10 Ma, 3 Ma, 2 Ma and 1.6 Ma (Burbank & Johnson 1983) and vertebrate fossils in Nepal (West et al. 1991). The following very high sedimentation rates may be calculated for the Siwaliks: Lower Siwaliks 0.83 mm/a, Middle Siwaliks 0.3 mm/a and Higher Siwaliks 0.6 mm/a).

The Himalayan Frontal thrust belt

The Himalayan Frontal thrust belt is bounded by the Himalayan Foothill boundary (HFB) and covered by recent fluvial deposits of the Indus and Ganga fluvial systems to the SW and the Main Boundary thrust (MBT) to the NE. It is composed of non-metamorphic Cenozoic sediments of the Ganga foredeep basin (Raiverman et al. 1983, 1993). Balanced and restored structural cross sections of the thrust belt, illustrated in Plate 1, 2 and 3 and Fig. 12, are well constrained by seismic reflection profiles and exploration wells (Raiverman et al. 1983, 1993, Lyon-Caen & Molnar 1985, Tandon 1991, Burbank et al. 1996, DeCelles et al. 1998 a,b, Powers et al. 1998, Mukhopadhyay & Mishra 1999). For the Kangra reentrant (section 7 on Plate 2 and Fig. 12), the surface of detachment (the active Himalayan thrust AHT) is situated on the base of the Subathu Formation and dips at an angle of ~9 ° to the NE (Powers et al. 1998). The depth of the pre-Tertiary basement is about 8000 m below the Main Boundary thrust (MBT). The Dharamsala Formation and Lower Siwaliks were deposited on Riphean-early Vendian dolomites of the Vindhyan basin in a fore deep basin limited to the south by the Aravalli crystalline range, more than 1000 m high, composed of Lower Proterozoic rocks (Fig. 12). The step in the southern Subhimalayan molasse basin determined the position of the Himalayan Foothill boundary (Janauri anticline) and the front of the decollement of the Subhimalaya on the Himalayan Foothill thrust. The wedge-shaped cross sectional shape of the Neogene Ganga basin is controlled by the deflection of the Indian plate overlying an inviscid fluid upper mantle below the weight of the growing Himalayan nappe stack (Lyon-Caen & Molnar, 1985). The youngest Siwalik strata have an age of ~4 Ma in the Kangra reentrant (Sarkaghat anticline section on Fig. 12) and of 1 Ma in Kashmir (magnetostratigraphy and fission track ages, Burbank et al. 1996). These periode of 4–1 m.y. of non sedimentation or erosion may indicate the time interval of the earliest detachment of the Ganga Molasse thin-skinned belt. Powers et al. (1998) estimate a minimum of shortening of 23 km in the Siwalik basin, SW of the Palampur thrust, that occurred since 1.9-1.5 Ma, yielding a shortening rate of 14 ± 2 mm/yr. As the displacement of the Palampur thrust is greater than 10 km, the shortening of the whole Cenozoic Ganga basin is greater than 33 km. This conclusion is based on the assumption that all movements occurred in the plane of the NE-SW oriented geological section and it does not consider possible dextral displacements related to the probable oblique collision between India and Asia.

The Neogene and Quaternary Himalayan accretionary wedge and active Himalayan structures

Many field observations and geophysical data support active deformation in the Himalayan mountain range. According to Patriat & Achache (1984), the India-Asian convergence is at the present about 5 cm/yr and about 1/3 – 1/2 of the convergence may be absorbed by shortening in the Himalayan range. The location of earthquake hypocenters at a depth of about 10-15 km below the Main Central thrust and the related thrust-type fault plane solutions indicate an Active Himalayan thrust (AHT) at the base of the Himalayan accretionary wedge (Fig. 13 & 15; Molnar 1990, Avouac et al. 2001, Qin et al. 2001). A NE-dipping thrust, at the base of the latest Paleocene to present day Ganga molasse sediments, is documented by seismic reflection lines (Fig. 12; Raiverman et al. 1983, Burbank et al. 1996, DeCelles et al. 1998, Powers et al. 1998). No seismic reflection survey has been shot in the NW Himalaya between the MBT and the Indus-Tsangpo suture. An extrapolation of the seismic reflection data to this area from Nepal and Buthan Himalaya (Hirn et al. 1984, Alsdorf et al. 1998 and Hauck et al. 1998) is questionable. Nevertheless, the INDEPTH TIB-1 data (Hauck et al. 1998) were extrapolated and used for the construction of a geological profile (Fig. 13). Hauck et al. (1998) interpret a strong 9°NE-dipping reflection at a depth of about 30 km below the STDS of Nepal to be the active main Himalayan thrust and a second parallel reflection at a depth of 62.5 km to be the Moho discontinuity. At about 100 km south of the Yarlung Zangbo suture, both the AHT and Moho reflections disappear. The surface profile of Fig. 13 corresponds to the geological section 6 of Plate 2 and was combined with the deep structures of the INDEPTH TIB 1 data from Buthan. The study of the Spot satellite images suggests that late anticlines in the Himalayan range continue to be uplifted relative to adjacent synclines. This is evident in the Tso Morari region where the lowered upper Pfirtse River basin is filled with huge gravel plains, whereas farther to the N the Mata range is uplifted and strongly eroded. The up warping of this range is also controlled by the uplift of the footwall of the E-dipping Tso Morari normal fault. The fault continues to the north in the conjugate faults of the N-striking active Kiagar Tso graben, and continues en-échelon in E-dipping and N-striking normal faults through the Ladakh batholith to the north of Mahe. Many late Himalayan structures are compatible with active N-S (or NNE-WSW) compression, W-E (or WNW-ESE) extension and dextral shear along the NW-SE striking Indus-suture. For example:

1. The huge W-E striking Chamba basin – Baralacha La structural depression between the Kishtwar, Zanskar and Tso Morari domes to the N and the Rampur – Sutley high to the south and the NNE-dipping Sanku flexure at the northern border of the Suru syntaxis. These structures also form type 1 interference patterns (Ramsay 1967) between older NW-striking domes and backfolds and younger E-striking folds (Plate 4).
2. The NNE-SSW striking and WNW-dipping Yurdu flexure between the Kashmir basin to the W and the Zanskar crystalline – Kishtwar dome to the E (interpreted as a W-dipping normal fault by Fuchs (1975), (Plate 3 & Fig. 6).
3. The E-dipping Tso Morari normal fault, the NNE-SSW striking conjugate Kiagar Tso normal faults and the ESE-dipping normal faults in the Ladakh batholith to the N of Mahe (Plate 1, 3 and Fig. 6; Steck et al. 1998).
4. The E-dipping flexure at the western border of the Tso Kar (interpreted as a normal fault by Fuchs & Linner, 1996).

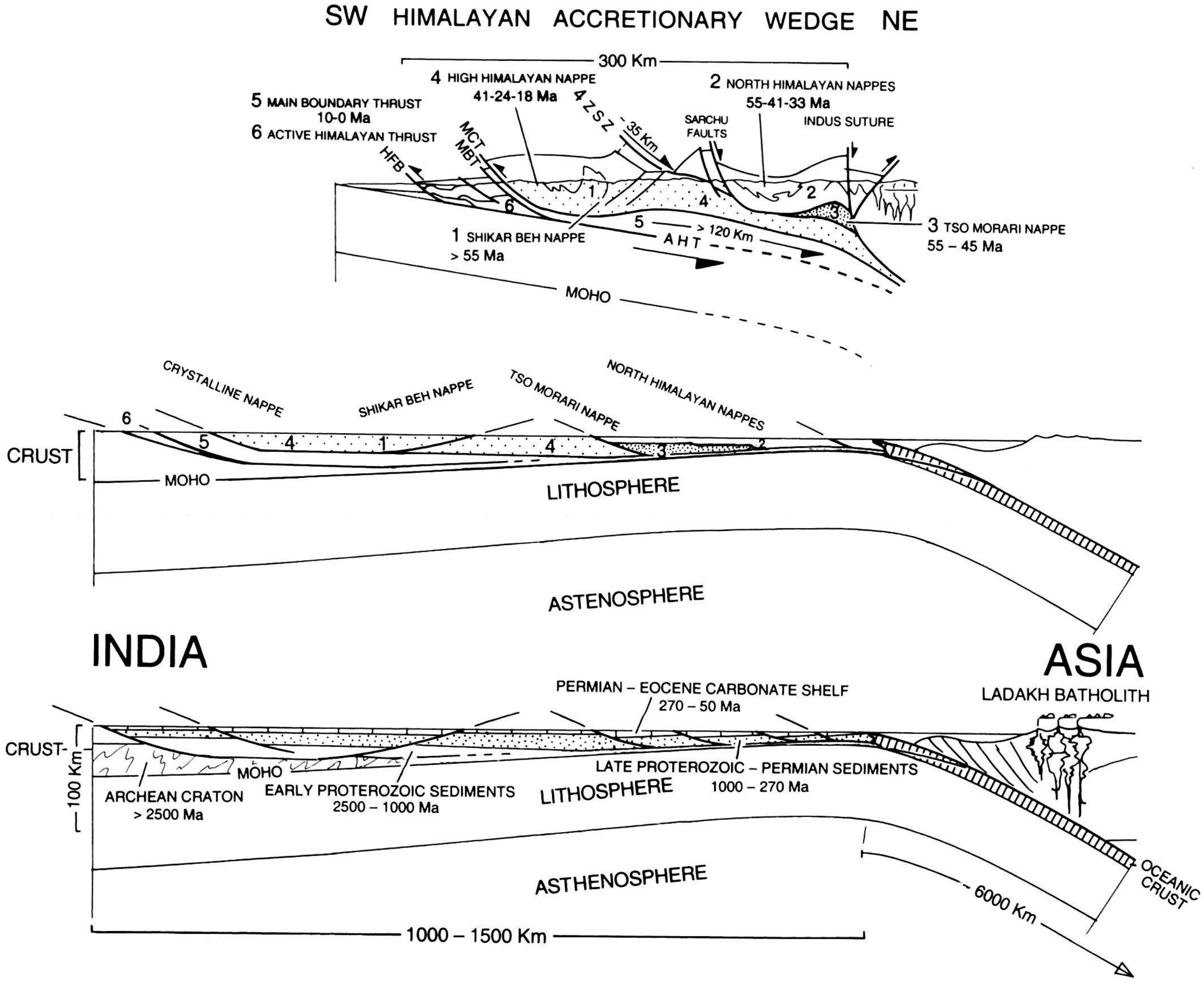

Fig. 13. A geological section of the Himalayan accretionary wedge and palinspastic sections that illustrate the detachment and accretion of successively deeper and older elements of the upper Indian crust during its underthrusting below Asia.

5. Late NW-striking strike slip structures along the Indus suture (Stutz & Steck 1986, Steck et al. 1993).
6. The Suru syntaxis is the result of the NNE-striking Yurdu flexure with a sinistral component and the ESE-striking dextral Sanku flexure (Gilbert et al. 1983).
7. The NNE-striking sinistral Ravi valley reentrant and a similar indentation in the Kangra reentrant to the south of Mandi (Mandi flexure) are interpreted as ESE dipping late flexure zones.

The structures enumerated above belong to a system of similar structures of a Pliocene and Pleistocene age, such as the Thakkhola-Mustang graben in North Nepal and southern Tibet described by Colchen (1999) and others. Late Pliocene to Holocene pre-glacial, glacial and alluvial sediments of Karewa series of Kashmir basin are up to 2'000 m thick and testify to subsidence in Kashmir basin to the W of the Yurdu flexure (Plates 1 and 3). We conclude that the Suru dome or Suru syntaxis, between the Yurdu and Sanku flexures are of the same Late Pliocene to Holocene age (Fig. 6). Seeber & Gornitz (1983) used river profiles along the Himalayan arc as indicators of active tectonics and demonstrated that the MCT remains an active structure at the present. The alluvial plains of the Gaj, Beas and Sutlej rivers are 10-20 km wide and lie between the frontal Siwalik anticlines. They suggest important subsidence of the Ganga foredeep basin in front of the active Himalayan range. Métivier et al. (1999) showed that the curve for overall average accumulation rates for Asian sedimentary basins, since the beginning of the Paloegene, shows an exponential evolution from slow accumulation rates up to the beginning of Oligocene time, after which, the rates quickly increase exponentially. They suggested that extrusion and crustal shortening are complementary processes. Molnar & England (1990) suggested that the late Cenozoic global climate change towards lower temperatures, which increased alpine glaciations, made the climate stormier and caused perturbations in the air humidity, the vegetation, precipitation and erosion, are a possible cause for late Cenozoic uplift of the Himalayan range. For Whipple et al. (1999) neither fluvial nor glacial erosion is likly to induce significant peak uplift.

5. Discussion and conclusions on the structural and metamorphic evolution of the Himalayan range and its sedimentary record

The age of the India/Asia continental collision

The formation of the Himalayan range started with the continental collision of India and Asia. The collision is a complex event, which is constrained by a number of independent data. The continental collision is preceded by the period of subduction of the Neo-Tethys oceanic crust below the Asian plate, documented by the main emplacement of the Ladakh batholith (Transhimalayan batholith) intrusions between 103-50 Ma (Honegger et al. 1982, Weinberg & Dunlap 2000) followed by later intrusions at 41 Ma (Schärer et al. 1984). Since 65-50 Ma, the Transhimalayan batholith and the Asian mantle wedge formed the backstop of the growing N-Himalayan range. Geological evidence shows that the continental collision of India and Asia started in the region of the NW Himalayan Nanga Parbat syntaxis about 65 Ma ago, as indicated by a terrestrial fauna exchange between India and Asia in Pakistan at the Cretaceous/Tertiary boundary (Jaeger et al. 1989). Patriat & Achache (1984) correlate the reduction in convergence velocity between India and Asia from ~15 cm/Ma to the present day velocity of 5 cm/Ma and the beginning of an anticlockwise rotation of India ~52 Ma ago (anomaly 23) with the continental collision. This anticlockwise rotation of India during the continuation of collision is corroborated by other paleomagnetic studies (Besse et al. 1984, Klootwijk et al. 1985, Appel et al. 1995, Schill et al. 2001) and by structural data (Pêcher 1991, Steck et al. 1993, Wyss et al. 1999, Epard & Steck in press). Evidence indicating tectonic uplift in the Zanskar shelf at 57 and 54 Ma ago, corresponding to stratigraphic unconformities at the base of the Early Eocene Kesi Limestone and the Chulung La and Kong formations, are also explained by the continental collision (Van Hinte 1978, Garzanti et al. 1987). The change from marine to continental sedimentation in the epi-suture Indus Formation during the Ypresian results also of the continental collision (Garzanti & Van Haver 1988). The 55 ± 15 Ma radiometric age of the Tso Morari eclogites (De Sigoyer et al. 2000), which crystallised at a depth of over 90 km as indicated by the crystallisation of coesite (Mukherjee & Sachan 2001), suggests a beginning of subduction of the Indian continental crust at this time (Fig. 8, 9). If the convergence rate between India and Asia was of 14 cm/a, as shown on Fig. 8 (Patriat & Achache 1984), the time necessary to subduct the Tso Morari gneiss to a depth of 90 km was ~1 Ma. All these data testify to an Early Eocene age for the continental collision of India and Asia, ranging from late Ypresian in the west to the Lutetian in the east (Rowley 1996).

The geometry of the N-Indian margin before continental collision

Stratigraphic sections and a palinspastic reconstruction of the N Indian margin before continental collision are illustrated in Fig. 3 and 4. Before collision, North India represented a flexural upper plate margin (Wernicke 1985, Steck et al. 1993, Stampfli et al. 2001), characterised by a regular stratified sequence of Lower Proterozoic to Tertiary rocks, crosscut by listric normal faults related to the Ordovician, Carboniferous and Late Permian to Early Cretaceous continental extensions. The Indian crust south of the margin is composed of the >2'500 Ma old Archean craton. Elements of this old craton were never found in the Himalayan range. The Archaen craton is overlain by early Proterozoic siliciclastic sediments of the Rampur Formation that are crosscut by, or overlie 1'800-1'866 Ma bimodal basic and granitic intrusions (Kober 1987, Klötzli 1997, Singh et al. 1994, Miller et al. 2000). Similar rocks are exposed in the Lesser Himalaya. The Rampur Formation represents a northern equivalent of the Aravalli crystalline basement. The next younger strata, ~1'000-590 Ma, are composed to the south, in the future Lesser Himalayan units, of an alternation of carbonate and siliciclastic sediments that belong to, the Lower and Middle Riphean stromatolitic Shali Limestone, the Upper Riphean-Vendian Simla-Dogra Slates, the Vendian Blaini glaciomarine boulder slates, the Vendian stromatolitic Krol limestone and dolomite and the Lower Cambrian siliciclastic Tal Formation. Outcrops of Permian marine sediments below the Thanetian-Lutetian Subathus of the Lower Chenab and Ans valleys testify to a local Permian back shoulder basin of the Indian continent. The external part of the Indian crust, the future North and High Himalayan nappes, are mainly composed of siliciclastic sediments that range from 1000-270 Ma and consist of the Riphean Chamba Fm. (or Lower Haimantas), the Vendian glaciomarine Manjir boulder slate (or Middle Haimantas) a northern lateral equivalent of the Blaini sediments (or Middle Haimantas), the Vendian – Late Cambrian Upper Haimantas and locally a fore-bulge basin of the Paleo-Tethys Indian continent with Ordovician, Silurian, Devonian, Carboniferous and Lower Permian siliciclastics and carbonates. The Late Proterozoic and Cambrian graywackes were intruded during Ordovician extension by granite sheets and basic dykes of a calc-alkaline suite (Frank et al. 1995, Girard & Bussy 1999). A carbonate platform, up to 3000 m thick, was formed on the N Indian shelf during the opening of the Neo-Tethyan ocean, from the Late Permian to the Ypresian (270-50 Ma)(Gaetani & Garzanti, 1991). It is important to note, that the units of the High Himalaya are composed of a single strati-