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Geology of the NW Indian Himalaya

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Keywords: Himalayan tectonics, nappe tectonics, plate tectonics, continental collision, Himalayan metamorphism

ABSTRACT

This review paper deals with the geology of the NW Indian Himalaya situated in the states of Jammu and Kashmir, Himachal Pradesh and Garhwal. The models and mechanisms discussed, concerning the tectonic and metamorphic history of the Himalayan range, are based on a new compilation of a geological map and cross sections, as well as on paleomagnetic, stratigraphic, petrologic, structural, metamorphic, thermobarometric and radiometric data. The protolith of the Himalayan range, the North Indian flexural passive margin of the Neo-Tethys ocean, consists of a Lower Proterozoic basement, intruded by 1.8–1.9 Ga bimodal magmatites, overlain by a horizontally stratified sequence of Upper Proterozoic to Paleocene sediments, intruded by 470–500 Ma old Ordovician mainly peraluminous s-type granites, Carboniferous tholeiitic to alkaline basalts and intruded and overlain by Permian tholeiitic continental flood basalts. No elements of the Archaen crystalline basement of the South Indian shield have been identified in the Himalayan range. Deformation of the Himalayan accretionary wedge resulted from the continental collision of India and Asia beginning some 65–55 Ma ago, after the NE-directed underthrusting of the Neo-Tethys oceanic crust below Asia and the formation of the Andean-type 103-50 (-41) Ma old Ladakh batholith to the north of the Indus Suture. Cylindrical in geometry, the Himalayan range consists, from NE to SW, from older to younger tectonic elements, of the following zones: 1) The 25 km wide Ladakh batholith and the Asian mantle wedge form the backstop of the growing Himalayan accretionary wedge. 2) The Indus Suture zone is composed of obducted slices of the oceanic crust, island arcs, like the Dras arc, overlain by Late Cretaceous fore arc basin sediments and the mainly Paleocene to Early Eocene and Miocene epi-sutural intra-continental Indus molasse. 3) The Late Paleocene to Eocene North Himalayan nappe stack, up to 40 km thick prior to erosion, consists of Upper Proterozoic to Paleocene rocks, with the eclogitic and coesite bearing Tso Morari gneiss nappe at its base. It includes a branch of the Central Himalayan detachment, the 22-18 Ma old Zanskar Shear zone that is intruded and dated by the 22 Ma Gumburanjun leucogranite; it reactivates the frontal thrusts of the SW-verging North Himalayan nappes. 4) The late Eocene-Miocene SW-directed High Himalayan or "Crystalline" nappe comprises Upper Proterozoic to Mesozoic sediments and Ordovician granites, identical to those of the North Himalayan nappes. The Main Central thrust at its base was created in a zone of Eocene to Early Oligocene anatexis by ductile detachment of the subducted Indian crust, below the pre-existing 25–35 km thick NE-directed Shikar Beh and SW-directed North Himalayan nappe stacks. 5) The late Miocene Lesser Himalayan thrust with the Main Boundary Thrust at its base consists of early Proterozoic to Cambrian rocks intruded by 1.8–1.9 Ga bimodal magmatites. The Subhimalaya is a thrust wedge of Himalayan fore deep basin sediments, composed of the Early Eocene marine Subathu marls and sandstones as well as the up to 8'000 m-thick Miocene to recent Ganga molasse, a coarsening upwards sequence of shales, sandstones and conglomerates. The active frontal thrust is covered by the sediments of the Indus-Ganga plains.

RESUME

Cette synthèse traite de la géologie de l'Himalaya du NW de l'Inde, plus particulièrement des secteurs localisés dans les provinces du Jammu et Cachemire, de l'Himachal Pradesh et du Garhwal. Les modèles et mécanismes qui sont discutés dans ce travail concernent la tectonique et l'histoire métamorphique de la chaîne himalayenne. Ils sont basés sur une nouvelle carte géologique, des coupes géologiques et également sur des données paléo-magnétiques, stratigraphiques, pétrologiques, structurales, métamorphiques, thermo-barométriques et radiométriques.

La marge passive nord indienne de la Néotéthys qui constitue le protolithe de la chaîne himalayenne, est composée d'un socle du Protérozoïque précoce, suivi de sédiments du Protérozoïque tardif à Paléocène, intrudés par des roches magmatiques bimodales de 1.8–1.9 Ga, par des granites peralumineux ordoviciens de 470–500 Ma et par des basaltes tholéiitiques à alcalins du Carbonifère. Des basaltes permiens de plateau continental sont intercalés dans la série. Aucun élément du socle cristallin archaïque formant le vieux craton indien n'a été observé dans la chaîne himalayenne. Le prisme d'accrétion himalayen est issu de la collision continentale de l'Inde et de l'Asie qui a commencé il y a 65–55 Ma, après la subduction vers le N de la croûte océanique de la Néotéthys sous l'Asie et la formation du batholite du Ladakh de type andéen(103–41 Ma), situé au nord de la suture de l'Indus. De géométrie cylindrique, la chaîne de l'Himalaya est constituée d'éléments tectoniques dont l'âge diminue progressivement du NE vers le SW. Le batholite du Ladakh, large de 25 km, forme avec le coin mantellique (mantle wedge), l'appui (butoir) asiatique de la chaîne himalayenne. La zone de suture de l'Indus est composée d'écailles de croûte océanique et d'arcs insulaires, comme l'arc de Dras, obductées et recouvertes par des sédiments d'avant-arc du Crétacé tardif et par la molasse intra-continentale (épi-suturale) du Crétacé supérieur à Eocène précoce et du Miocène. L'empilement des nappes nord-himalayennes daté du Paléocène tardif à l'Eocène atteint une épaisseur de 40 km avant l'érosion. Ces nappes sont composées de roches du Protérozoïque supérieur à Paléocène et comprennent à sa base la nappe de gneiss ordoviciens du Tso Morari contenant des filons basiques éclogitiques à coésite. Une branche du grand détachement en extension de l'Himalaya central, la zone de cisaillement du Zanskar, datée par l'intrusion syncinématique du leucogranite du Gumburanjun de 22–18 Ma, réactive les chevauchements frontaux des nappes nord-himalayennes. La nappe du Haut Himalaya, également appelée la «Nappe Cristalline», chevauchant vers le SW entre l'Eocène tardif et le Miocène, est composée de sédiments du Protérozoïque supérieur au Mésozoïque et de granites ordoviciens. Il s'agit de roches identiques à celles des nappes nord-himalayennes. Le Main Central Thrust, à la base de la nappe du Haut Himalaya, a été créé dans une zone d'anatexie éocène à oligocène précoce formée sous l'empilement de 25-35 km d' épaisseur constitué par les nappes de Shikar Beh de vergence NE et nord-himalayennes de vergence SW. Le Bas Himalaya (Lesser Himalaya), limité à sa base par le Main Boundary Thrust, chevauche au Miocène tardif. Il est composé de sédiments du Protérozoïque inférieur à Cambrien, intrudés par des magmatites bi-modales datées de 1.8–1.9 Ga. Le Subhimalaya consiste en des sédiments de l'avant fosse himalayenne, composés de marnes et grès marins de la série des Subathus à sa base suivis de la molasse miocène de l'Indus et du Gange qui atteint une épaisseur maximale de 9000 m sous le chevauchement frontal du Bas Himalaya. Il s'agit d'une séquence d'argillites, de grés et de conglomérats. La granulométrie augmente vers le haut de la série. Plus au sud, le chevauchement frontal actif est couvert par les dépôts alluviaux du Gange et de l'Indus.

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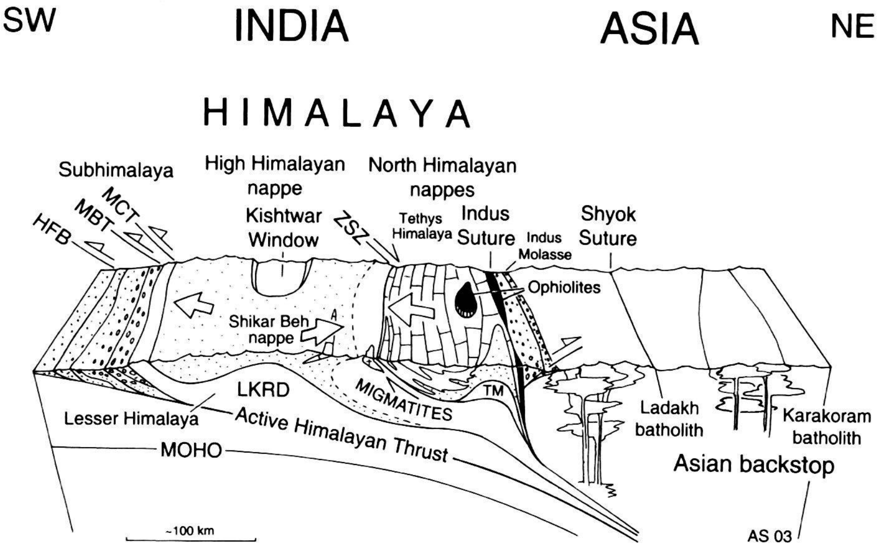

Fig. 1. The main tectonic units of the NW Indian Himalaya
HFB = Himalayan Frontal Boundary
LKRD = Larji-Kullu-Rampur dome
MBT = Main Boundary Thrust
MCT = Main Central Thrust
ZSZ = Zanskar Shear Zone

1. Introduction

In 1964, Gansser based his compilation, in form of a geological map, updated by Fuchs in 1981, at a scale of 1:2'000'000 and his famous monography on the Geology of the Himalaya, on the classical geological work produced by the Geological Survey of India and various international expeditions (Fig. 1). Tourism development during the seventies opened the still poorly known NW Indian Himalayan regions of Himachal Pradesh and Ladakh to new geological studies by American, Austrian, Australian, English, French, Indian, Italian and Swiss researchers.

Since 1979, geologists of the University of Lausanne have carried out a complete and continuous traverse across the Himalayan range between Leh, north of the Indus-Tsangpo suture, and Mandi at the southern border of the frontal thrusts of the High and Lesser Himalayan nappes (Fig. 2), using a multidisciplinary approach combining classical methods, such as geological mapping, stratigraphy, structural geology, petrography, with modern techniques, such as geochemistry, geochronology and thermo-barometry. The new geological data from the NW Indian Himalaya limited by the Ladakh batholith to the N, the Srinagar basin in Kashmir to the W, the Indus-Ganges molasse to the S and Tibet to the E, have been compiled on a geological map at a scale of 1:677'000 (Plate 1). The French Spot satellite images and the 1:250'000 U.S. Army map of India and Pakistan have been used as the topographical base and for toponyms.

The first part of this study describes the stratigraphy and magmatism of the Protolith of the Himalayan range in a chronological order and separately for the main tectonic units (Chapter 2, Fig. 3). The second part deals with the tectonic units, their emplacement and the related metamorphism, magmatism and synorogenic sedimentation. A tectonic model is presented in a concluding chapter.

2. Pre-Himalayan tectonics, sedimentation and magmatism

The Lesser Himalaya

The Lesser Himalayan units, the Lower Crystalline nappe, the Kishtwar, Larji-Kullu-Rampur units and their equivalent, the Chail nappes, the Jaunsar nappe and the Simla-Runkum nappe are composed of a Proterozoic sedimentary sequence deposited on the northern margin of Gondwana and are associated with, or intruded by 1'870-1'900 Ma S-type granites (Frank et al. 1995, Valdiya, 1995). The Ordovician granites, widespread intrusives in the High Himalayan Haimanta or Phe formation, are lacking. In Fig. 3 and 4, the author proposes a synthetic stratigraphic column ranging from early Proterozoic to an Eocambrian age, mainly based on review papers by Brookfield (1993), Frank et al. (1995, 2001), Miller et al. (2000), Srikantia & Bhargava (1998). This sedimentary sequence belongs to the northern part of the North and Central Indian Purana basin. No elements of its basement, the Archean Indian shield, has been observed in the Himalaya (Valdiya, 1995).

The Rampur Formation (Jhingran et al. 1952)
(Proterozoic)

The Rampur Formation is considered to be the basal and probably also oldest formation in the Himalaya. It consists of massive beds of white quartz-arenites alternating with thin layers of sericite and chlorite schists, locally associated with metarhyolites and metabasaltic dikes and lava flows (Rampur basic volcanics, 1'800 ± 13 Ma, U-Pb ages on single zircon, Miller et al. 2000). For Miller et al. (2000), it is not clear whether the Rampur formation quartzites are unconformably overlying the Bandal granitoid complex (Srikantia and Bhargava 1998, Sharma, K.K. & Rashid 2001) or whether the granitoids intrude these Quartzites (Sharma, V.P. 1977, Guntli 1993). The emplacement of silicic volcanites and of S-type per-

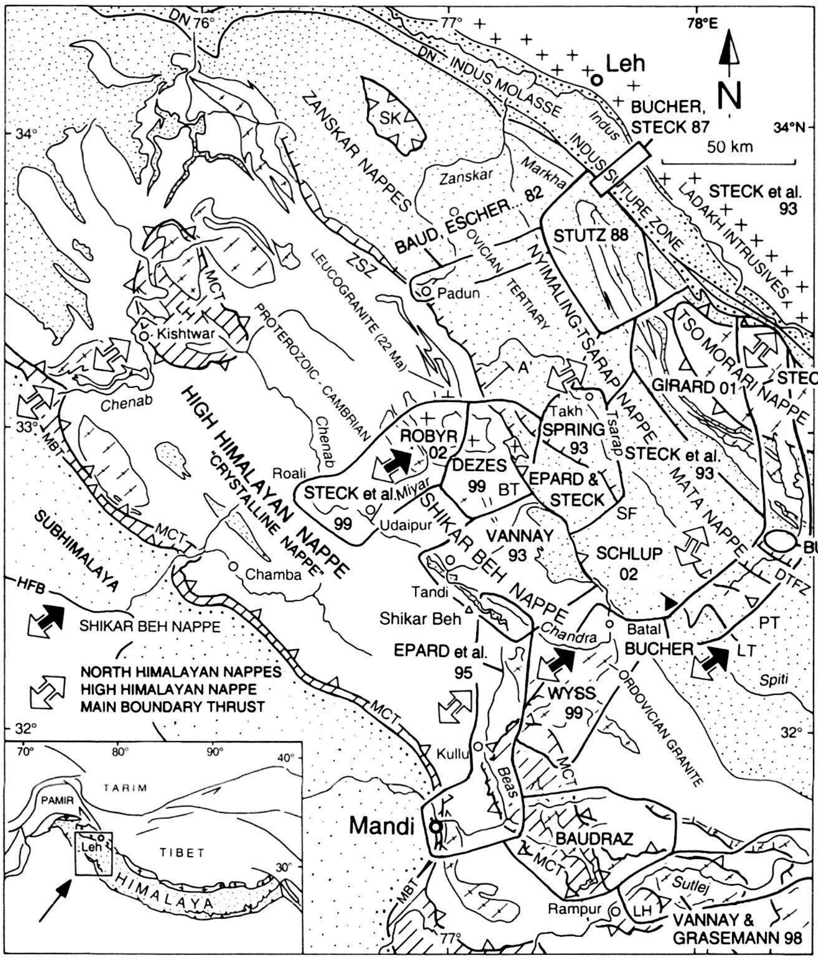

Fig. 2. Geological sketch map of the NW Indian Himalaya based on the area studied by geologists from the University of Lausanne since 1979. The compilation of other regions represented on the geological map (Plate 1) is mainly based on the work of earth scientists from Arizona, Grenoble, India, Milan, Poitiers, Oxford, Turin, Vienna and Zurich.

aluminous granitic rocks occurred around 1.86-1.84 Ga (Miller et al. 2000, Bandal and Sainj granite, Rampur window: 1'800 ± 70 Ma, Rb-Sr whole rock age of 4 samples, metarhyodacite, Rampur window: 1'840 ± 16 Ma, single zircon $^{207}Pb/^{206}Pb$ evaporation technique, Kober 1987, Klötzli 1997; Wangtu granitoid, Lesser Crystalline nappe: 1'866 ± 10 Ma, U-Pb zircon age by Singh et al. 1994). The striking similarities in terms of geological setting, geochemical characteristics and age between Bandal orthogneisses of the Rampur Fm. and the Wangtu Gneisses of the Lower Crystalline nappe suggest that they were formed under similar tectono-magmatic environment during the Paleoproterozoic period. The Rampur Formation represents a northern equivalent of the Aravalli crystalline.

Garsha and Berinag Formations
(Garsha Slates, Thöni 1977, Rampur Formation, Frank et al. 1995)
(Proterozoic)

The Garsha and Berinag formations are composed of quartzites and grey-green phyllites exposed in the Tons and Yamuna valleys and in the Simla klippe and are interpreted as lateral equivalents of the Rampur Formation.

The Shali Formation (Oldham 1888, Pilgrim & West 1928)
(synonyms: Sirban, Deoban, Tejam etc. Oldham 1888; Chail, Pilgrim & West 1928)
(Lower and Middle Riphean)

The Shali Formation, is a massive Mid Proterozoic carbonate platform up to 2000 m thick, composed of calcschists, blue-grey dolomite and limestone with stromatolites: *Kussiella, Collonella* and *Conophyton* of a early to middle Riphean age (Sinha 1972, Valdia, 1989, 1995). It forms the top of the older portion of the Lesser Himalayan sequence (Thöni 1977). In the Shali window, the limestones and quartzites of the Shali Formation are transgressed by the Eocene nummulitic limestones of the Subathu Formation.

The Simla Slates (Medlicott 1864, Pilgrim & West 1928)
(synonyms: Hazara Slates, Dogra slates, Wadia 1934)
(late Riphean-Vendian)

The late Riphean and Vendian Simla and Dogra Slates

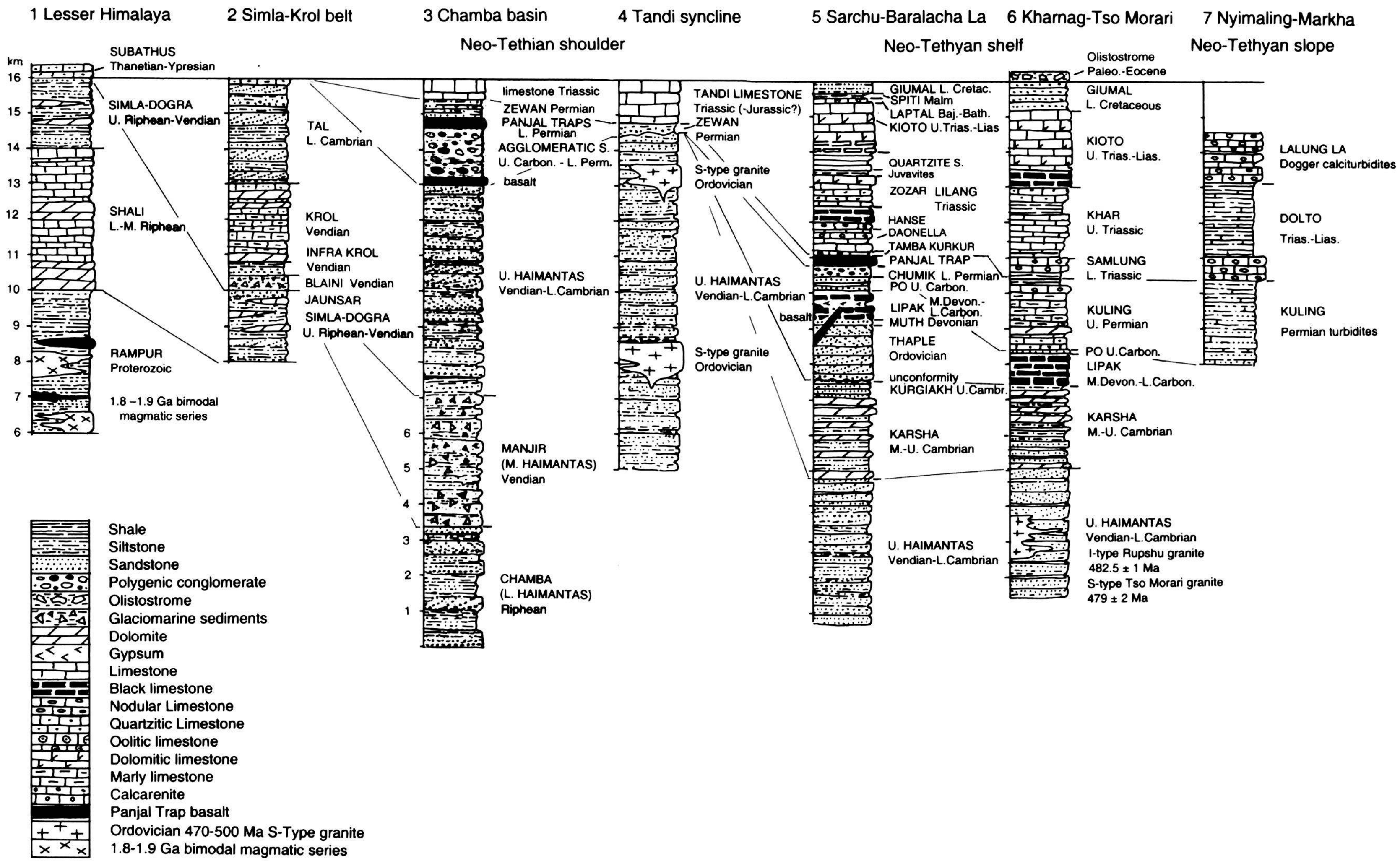

Fig. 3. Compilation of stratigraphic sections of the Lesser and High Himalaya based on stratigraphic sections as well as on biostratigraphic and radiometric ages by the following authors: Auden (1934), Azmi & Pancholi (1983), Bagati (1991), Bhatt et al. (1983), Bhatt & Mathur (1990), Brookfield (1993), Draganits et al. (2002), Frank et al. (1995, 2001), Fuchs (1975), Gaetani et al. (1990), Gansser (1964), Garzanti et al. (1986), Girard & Bussy (1999), Hayden (1904), Hughes & Jell (1999), Kumar et al. (1987), Miller et al. (2000), Pilgrim & West (1928), Ranga Rao et al. (1982), Rattan (1974, 1978, 1985), Spring (1993), Srikantia & Bhargava (1979, 1998), Steck et al. (1993), Stutz (1988), Stutz & Steck (1986), Tripathi et al. (1984), Vannay (1993).
Note that the Late Cretaceous Chikkim limestones are missing in these sections. Erosion relicts of the Late Cretaceous foraminiferal limestone are exposed near Takh on the Tsarap river and on the Chikkim mountain in the upper Spiti valley (Plate 1).

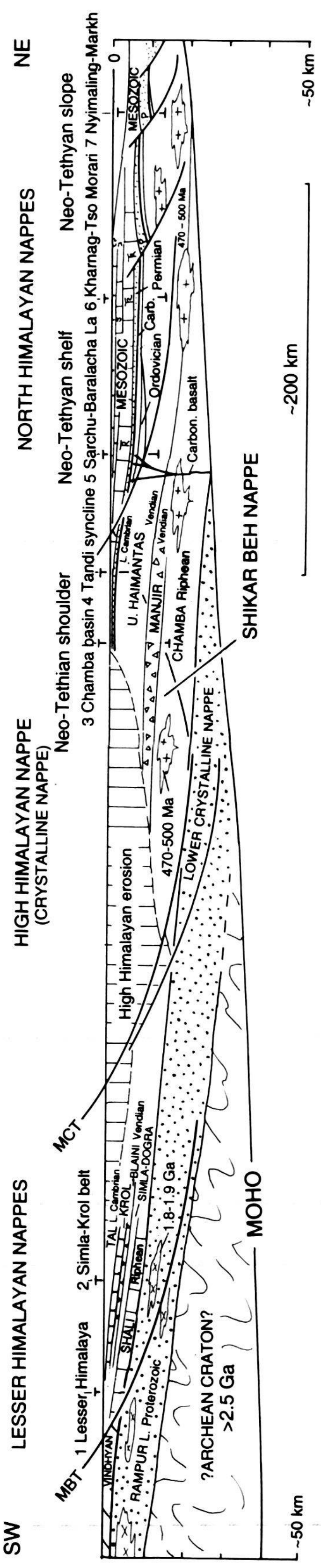

Fig. 4. NE-SW palinspastic section through the NW Indian Himalaya. The palinspastic section was constructed using the stratigraphic sections of Fig. 3 and shortening estimates of ~100 km for the N-Himalayan nappes, of over 300 km for the High Himalayan nappes and of less than 200 km for the Lesser Himalayan nappes. This reconstruction illustrates the geometry of the N Indian flexural passive margin before the continental collision of India and Asia.

consist of over 1'000 m of grey, green, and dark slates, phyllites and siltstones, with locally a thin intercalation of a green limestone. In the Simla-Runkum nappe the nummulitic limestones transgress the Simla Slates.

The Jaunsar Formation (Jaunsar Series, Oldham 1883, Pilgrim & West 1928) (synonyms: Jagas, Nagthat, Auden 1934) (late Riphean-Vendian)

The Jaunsar Formation, called the Jaunsar Series by Pilgrim & West (1928), are composed of about 500 m of quartzites, slates of dark or light green-brown or brown colour, purple phyllites and conglomerates and a few intercalation of carbonaceous phyllites. Pilgrim & West (1928), distinguished in the Simla region the Jaunsar Series forming a thrust sheet between the Simla unit in the foot wall and the Chail unit in the hanging wall, where as Frank (personal communication) considers this series to be a concordant stratigraphic unit in the upper part of the Simla slates. Pant & Shukla (1999) interpret the correlative Nagthat succession as a progradational tide-dominated Proterozoic succession. After Valdia (1995) the paleocurrents in the Nagthat Fm. are mainly NE-directed.

The Blaini Formation (Pilgrim & West 1928) (Vendian)

The Blaini Formation is composed of up to 200 m of diamictite pebbly mudstone alternating with slates of a probable Vendian age (< 700 Ma, Rb-Sr ages on detrital micas, Wolfgang Frank, personal communication). The Blaini Formation rests on the Simla Slates with a clear unconformity (Pilgrim & West, 1928, Frank et al. 2001). These sediments are interpreted as glaciomarine sediments or tillites. Frank (1995) considered the Blaini Formation as a lateral equivalent of the Neo-Proterozoic Manjir Conglomerate of the Haimantas or Phe Formation of the High Himalaya. A thin layer of pink carbonates overlies the pebbly mudstones.

The Krol Formation (Auden 1934) (Vendian)

The Blaini boulder beds are stratigraphically overlain by about 170 m of the black fine-grained sandy slates of the Infra-Krol Formation and up to 1'500 m of limestone and dolomite of the carbonate platform of the Krol belt (Audin 1934, West 1939, Gansser 1964, and Sinha 1987). Stromatolites being the only macrofossils, a Vendian age (Wolfgang Frank, personal communication) is more probable than the Permo-Triassic ("Phanerozoic") age suggested by the above authors. Valdia's conclusion (1995), that the carbonates of the Krol Fm. represent a lateral equivalent of those of the Shali Fm., is not supported by other geologists.

The Tal Formation (Middlemiss 1887) (Early Cambrian)

The sediments of the Tal Fm. are restricted to the Nigali Dhar area. A striking change in deposition took place after the deposition of the calcareous Krol sequence, with the younger beds consisting exclusively of detrital, mostly quartzitic rocks. Auden (1934, Gansser, 1964) distinguished a Lower and Upper Tal, which overlie with an angular unconformity the Krol carbonates. The Lower Tal is up to 1'000 m thick and is characterized by dark, often calcareous graywackes, carbonaceous and micaceous shales and some quartzites. The Upper Tal is up to 2'000 m thick and consists mainly of fine-grained (average grain size of 0.5 mm) arkosic quartzites. The Tal Fm. has yielded Tommotian to Botomian (Early Cambrian) conodonts (Azmi & Pancholi 1983), small shelly fauna (Bhatt et al. 1983, Bhatt & Mathur 1990), brachiopods (Tripathi et al. 1984) and trilobites (Kumar et al. 1987). Early Cambrian trilobites *Xela mathurjoshi* and *Redlichia noetlingi* around the limit of the Lungwangmiaon and Tsanglangpuian, have been identified by Hughes & Jell (1999) in the Lower Tal of the Kumoan Himalaya.

The Lower Crystalline nappe (synonym: Bajaura nappe, Thöni 1977)

The Lower Crystalline nappe is an alpine strongly metamorphosed and deformed unit exposed at the base of the Main Central Thrust in the Kishtwar and Rampur windows. It is composed of mylonitic augengneisses, garnet- and sericite-phyllites, quartzites, graphitic schists and marble layers of the Bajaura Formation and minor concordant sheets of metabasites and Proterozoic granitoids (Wangtu granitoid north of Rampur: 1'866 ± 10 Ma, U-Pb zircon age, Singh et al. 1994). The sediments differ from those of the overlying Haimatas or Phe graywackes. The occurrence of the Proterozoic granites suggests this unit belongs to the Lesser Himalaya rather than to the younger High Himalaya sequence, in which the old granites are missing. The metasediments may be lateral equivalents of the Rampur Formation.

Conclusion on the stratigraphy of the Lesser Himalaya

The units of the Lesser Himalaya, including the Lower Crystalline nappe, are composed of an uniform and typical stratigraphic sequence of a Proterozoic to Early Cambrian age (~2'500–530 Ma) intruded (or underlain?) in its lower part by ~1'850 Ma granitoids and basalts. An early to middle Proterozoic age of the Lesser Himalayan sequence is also suggested by the presence of 1.7–2 Ga detrital zircons (U-Pb-ages, Parrish & Hodges 1996). The sediments of the Lesser Himalaya are transgressed by the Eocene nummulitic Subathu Limestone (Middlemiss 1887, Gansser 1964). During the Tertiary Himalayan thrusting the lesser Himalayan sequence was detached from a same sedimentary basin (Purana) of the North Indian continent (Valdiya 1995).

The High Himalaya

The tectonic units of the High Himalaya (from S to N: the High Himalayan nappe or "Crystalline nappe", the Shikar Beh nappe, and the North Himalayan nappes, including the Tso Morari nappe) are characterized by elements of a single sedimentary sequence of a Late Precambrian to Paleogene age. The sequence may be divided in:

The Gondwanan sediments and related magmatic rocks of Neo-Proterozoic to Early Permian age deposited or emplaced on the northern margin of India which at that time was part of the Gondwanan supracontinent.

The sediments of the Neotethyan shelf and the Neotethyan slope, were deposited on the N Indian margin between the Late Permian opening of the Neotethyan ocean and its closure at the Paleocene India – Asia continental collision some 65-50 Ma ago (Patria & Achache 1984, Jäger et al. 1989, Garzanti et al. 1987, Rowley 1996).

The stratigraphic sections illustrated in Fig. 3 have been compiled from work by Hayden (1904), Gansser (1964), Stutz & Steck (1986), Stutz (1988), Garzanti et al. (1987), Brookfield (1993), Gaetani et al. (1990) Gaetani & Garzanti (1991), Saklani (1993), Spring (1993), Vannay (1993), Epard et al. (1995), Frank et al. (1995) and Wolfgang Frank (personal communication), Steck et al. (1993, 1998, 1999). Gansser (1964) wrote of the Spiti section: "It is this conformable sedimentary sequence of Spiti, from Precambrian onwards, which reflects a calm epeirogenic pre-Alpine history of the Tibetan Himalaya". In addition, Srikantia & Bhargava (1998) recognised the similarity of the stratigraphic columns of the High Himalaya that they grouped together in their Tethys Himalayan Tectogen. This sedimentary and magmatic sequence is crosscut in the Zanskar shear zone by 22 Ma leucogranites (Dèzes et al. 1999).

The Gondwanan Series

Haimantas (Griesbach 1891, Frank et al. 1995)
(Synonyms: Phe Formation (Nanda & Singh 1977), Lolab Fm. (Kumar et al. 1984), Shumahal Fm. (Srikantia & Bhargava 1983), Kunzam La Fm./Debsa Khas Mb. (Kumar et al. 1984), Jutogh Series (Pilgrim & West 1928), Batal Fm. (Srikantia et al. 1980), Vaikrita Series (Griesbach 1891).
(Riphean-Early Cambrian)

Frank et al. (1995) divide the Haimantas into the Lower, Middle and Upper Haimantas. Later, Draganits et al. (1998) proposed the subdivision Chamba Fm., Manjir Fm. and Phe Fm. as equivalents of the three divisions of Frank et al. (1995).

The Lower Haimantas (Chamba Formation, Rattan 1973) (Riphean?)

The Lower Haimantas includes the Blahai and Chamba Fm. and the Pukhri Slates (Rattan 1973). The Lower Haimantas, of a probable Riphean age, represents a more than 5'000 m thick monotonous sequence of meta-pelites, meta-siltstones and meta-graywackes or micaschists with subordinate interbedded carbonaceous layers. The formation shows an overall coarsening-upward trend and sedimentary structures like graded bedding, load casts and flute casts point to a turbitite-type depositional environment (Draganits et al. 1998). The Lower

Haimantas are intruded by the 476 ± 12 Ma old Kaplas, Dalhousie and Mandi granites and the 912 Ma Chur granite (Frank et al. 1995).

The Middle Haimantas (Manjir Conglomerates, Rattan 1973, Manjir Formation, Draganits et al. 1998)
(Vendian?)

The Manjir Conglomerates are up to 3'000 m thick and separate the Lower from the Upper Haimantas. They were designated the Middle Haimantas by Frank et al. (1995). They consist of pebbly diamictites interpreted as glaciomarine sediments, easily recognized by their boulder slate appearance. The matrix supported, polymodal, poorly sorted, angular and rounded clasts, up to 1 m in diameter, are dispersed in a chaotic manner. The clasts consist mainly of quartzites, graywackes, slates and subordinate granites and gneisses. ^{40}Ar-^{39}Ar ages < 700 Ma on detrital white micas in the NW Himalaya (Frank et al. 1995) and 0.8-1.0 Ga U-Pb ages on detrital zircons from the Nepal High Himalaya suggest a Vendian age for the Manjir Formation. Wolfgang Frank (personal communication) considers the glaciomarine Manjir boulder beds of the High Himalaya to be a lateral equivalent of the Blaini Fm. of the Lesser Himalayan sequence, thereby implying that the Lower Haimantas represents a lateral equivalent of the Simla slates. Like the Blaini Formation, the Manjir Formation is overlain by a thin dolomite and limestone layer. SE of the Chamba region, the Manjir boulder slate passes gradually into locally graded meta-graywackes and meta-sandstones intercalated with slates similar to the matrix of the underlying Manjir Formation.

The Upper Haimantas (Draganits et al. 1998)
(Vendian-Early Cambrian)

The Upper Haimantas is up to 9'000 m thick and consists of monotonous graywackes, slates to sandstones with one or two 50m- to 100 m-thick characteristic graphitic quartzite horizons and thin carbonate layers or horizons with carbonate concretions (Frank et al. 1995, Steck et al. 1999). A shallow water detrital marine environment is indicated for the Haimanta sediments by the rhythmic alternation of sandstone and siltstones, the presence of hummocky cross stratification and ichnofossils. Common sedimentary structures, such as graded bedding, load casts, flute casts, chevron marks and groove marks, indicate a turbidite type depositional environment (Draganits et al. 1998). No identifiable fossils have been found in the Haimantas to confirm their generally acknowledged Neo-Proterozoic age. Ordovician granitoids dated at 500 Ma are widespread in the Lower and Upper Haimantas (Le Fort et al. 1986, Pognante et al. 1990, Frank et al. 1995, Girard & Bussi 1999). The 912 Ma-old Chur granite represents a unique age for a granitoid of the Himalaya that intrudes the Lower Haimantas in the Chur peak area (Frank et al. 2001). The Himalayan Cambro-Ordovician magmatism will be discussed in a separate chapter below.

The Karsha Formation (Nanda & Singh 1977)
Synonyms: Parahio Series (Spiti, Hayden 1904), Kunzam La Fm. (Srikantia et al. 1980), Rangamal Fm. (Kashmir, Srikantia & Bhargava 1983)
(Middle and Late Cambrian)

There is a gradual change in the carbonate supply between the Upper Haimantas and the Karsha fms. The lower boundary of the Karsha Formation is arbitrarily defined by the first occurrence of dolomite-bearing beds. In the Kurgiakh valley in southern Zanskar, Garzanti et al. (1986) subdivide the formation into the Mauling (~400 m thick), the Thidsi (200-300 m thick) and the Teta (70-160 m thick) members. A similar sequence is reported by Vannay (1993) from the Baralacha La region.

The arenaceous Mauling member is composed of very fine-grained sandstones and micaceous siltstones similar to those of the Upper Haimantas. Trace fossils characteristic of the Early Cambrian, such as *Rusophycus* isp., *Cruziana* isp., *Arthraria* isp. and *Planolites* isp., as well as synaeris cracks, indicate shallow marine conditions (Stutz 1988, Vannay 1993). The carbonates are ferriferous dolomite with locally hemispheroidal or tabular stromatolites, bioturbation and phosphatic shells.

The carbonaceous Thidsi member comprises thick amalgamated beds of rusty-weathering grey dolomites, locally separated by pyritic greenish shales.

The Teta member is composed of grey nodular stromatolite limestone and marls, sometimes dolomitic, alternating with black schists. The presence of rare trilobites (*Goniagnostus spiniger, Diplagnostus planicauda, Eoshengia sudani*?) indicate the age of the Thidsi member of the Zanskar valley to be Middle Cambrian (Changhian) (Huges & Jell 1999). Like the Haimantas, the entire Karsha sequence was deposited in a shallow water marine environment (Garzanti et al. 1986).

The Kurgiakh Formation (Garzanti et al. 1986)
Synonyms: Rangamal Fm. in Kashmir (Srikantia & Bhargava 1983)
(Late Cambrian)

Garzanti et al. (1986) named a sequence of pelites and sandstones, stratigraphically overlying the Karsha Formation in the Kurgiakh Valley, the Kurgiakh Formation. He subdivided them into a 150 m-thick fossiliferous lower Surichun Member and an upper 150 m-thick silty-sandy Kuru Member. A rich late Middle Cambrian trilobite fauna has been identified in the middle Surichun member (*Goniagnostus aculeatus, Lejopyge armata, Hypagnostus correctus, Linguagnostus* cf. *tricuspis, Clavagnostus* sp., *Agnostid indet., Fuchouia indet., Schmalenseeia amphionura, Damesops sheridanorum, Torifera* sp., *Goniagnostus spiniger, Diplagnostus planicauda, Eospengia? sudani*, Hughes & Jell 1999). The trilobite-bearing beds of the Surichun member consist of black to greenish-grey bioturbated or parallel-laminated pelites. Silty and nodulose dolomites are intercalated in lenticular beds gradually decreasing in thickness upward. Tuffitic layers are present. The lower part of the 300 m-thick Kuru member consists of thin-bedded, greenish-grey siltstones with turbidite sedimentary structures. The middle part comprises thick-bedded (up to 6.5 m) dark greenish-grey sandstones and the upper part consists of a fining-upward megasequence made of amalgamated and multiply graded sandstone beds of gradually decreasing thickness. According to Garzanti et al. (1986), the Karsha carbonate platform passes by gradual deepening into the turbiditic sedimentation of the Kurgiakh Formation, which indicates active tectonic subsidence that largely exceeded the sedimentation rate.

Thaple Formation (Hayden 1904, Nanda & Singh 1977)
Synonyms: Kashmir: Riskobal Fm. (Srikantia & Bhargava 1983), Ladakh-Zanskar: Thango Fm. (Kanwar & Ahluwalia 1979), Spiti: Shian Quartzites, Pin Fm. (Goel & Nair 1977), etc. (Ordovician)

The Thaple Fm. consists of mainly red, sometimes white or green conglomerates and sandstones. The conglomerates overlie equivalents of the Kurgiakh or Karsha Formation with angular unconformity (Pin and Parahio valley, Hayden 1904, Pt. 6, Upper Kamirup Valley, Epard & Steck, in press). The thickness of the Thaple Fm. varies from 250 m to 1'500 m and synsedimentary normal faults suggest a tectonic relief related to extensional tectonism (Spring 1993, Vannay 1993, Steck et al. 1993). Fossils such as corals, crinoids, brachiopods, trilobites indicate an Ordovician to Earliest Silurian age (Hayden 1904, Ranga Rao et al. 1982 and Bagati 1990, 1991). Sedimentary structures such as high-angle crossbeds indicate a fluvial, deltaic depositional environment. Polymictic pebbles of the underlying partially eroded Haimantas, Karsha and Kurgiakh formations occur. The outcrops of the Thaple conglomerates and the overlying white Muth quartzites are restricted to a narrow basin which can be followed from the Kurgiakh valley in Zanskar in the NW, through the Baralacha La, Upper Chandra Valley region and the Spiti valley to the SE. Another correlative sequence is exposed in the Liddar valley of Kashmir, where the Ordovician sediments are formed by pink, purple, green and grey quartz arenites, siltstones and shales (Rishkobal Fm. Srikantia & Bhargava 1983). The whole Gondwanan sedimentary sequence is undifferentiated on the geological map due to the absence of a complete and detailed map of the Kashmir basin.

The Ordovician magmatism

Widespread magmatic activity ranging from and peaking at 500-470 Ma intrudes the Proterozoic and Early Cambrian sedimentary sequence of the High Himalayan units (Le Fort et al. 1986, Mehta 1977, Trivedi et al. 1986). Most of the available radiometric data are Rb-Sr ages that must be used with caution. Precise U-Pb ages on zircon and monazite have been determined in the NW Himalaya by Girard & Bussi (1999: S-type Tso Morari granite, 479 ± 2 Ma, I-type Rupshu granite, 482.5 ± 1 Ma) and Pognante et al. (1990: 472 ± 9 Ma, Temasa granite). No Ordovician granites have been recognized in the Lesser Himalaya. A low-pressure contact metamorphism, with an assemblage of quartz, plagioclase, biotite, white mica, garnet, staurolite, sillimanite, andalusite, has been described by Guntli (1993) in the southern contact aureole of the Ordovician Kaplas granite, Tavi valley, where it has been undisturbed by alpine deformation. A zone of dark and fine-grained hornfels is exposed around the Rupshu granite (Girard & Bussy 1999). Hornfelses with quartz, plagioclase, muscovite, biotite and andalusite in metapelites and diopside, grossularite and wollastonite in metamorphosed Karsha dolomites and limestones, are observed near the Nyimaling granite in the Kanjatse peak area (Baud et al. 1982). The presence of a small quantity of probable contemporaneous basic intrusions, that include rocks at Tso Morari and in the Miyar and Tos Valleys, indicate a bimodal magmatism, but no radiometric ages of the basic rocks are available. Basic dikes with a Tertiary eclogite facies mineral assemblage are widespread in the Tso Morari granite gneiss (Berthelsen 1953). Meta-olivine gabbros are described by Pognante & Lombardo (1989) from the Miyar valley and by Wyss (1999) from the Tos valley, as being a magmatic assemblage of olivine, clinopyroxene, orthopyroxene, spinel, plagioclase (core An 72, rim An 47), biotite, ilmenite and magnetite. The rocks in the Tos Valley are overprinted by a probable Tertiary M1 amphibolite-granulite facies transition assemblage of Cpx, Opx, spinel, garnet, amphibole, ilmenite and magnetite at a pressure of 10 kbar and 700°C and later by an M2 amphibolite facies regional metamorphism (Wyss 1999).

A major, continental-scale tectonic event is required to generate such a large magmatic belt. Girard & Bussy (1999) suggest it is a late stage event in the evolution of the Pan-African orogen. Geotectonic reconstructions traditionally consider that only the extreme south of the Indian plate, situated at a distance of over 3000 km from the Himalaya, was affected by the Pan-African compressional events (e.g. Stern, 1994). As we do not know what kind of crust was situated north of India before the opening of the Paleo- and Neo-Tethyan oceans, the relation of this calc-alkaline magmatism to a zone of subduction can not be excluded. However, the association of the 500 Ma bi-modal calc-alkaline magmatism at a deep tectonic level with syn-sedimentary extensional structures in the Ordovician Thaple sediments now in a high tectonic level of the High Himalayan sedimentary sequence coupled with the fact that we have never observed old pre-Himalayan fold structures in the presently studied NW Himalaya, or an old regional metamorphism testifying of a Pan-African orogeny, as proposed by Srikantia et al. (1977), Mehta (1977, 1978), Garzanti et al. (1986) or Marquer et al. (2000), require another mechanism of magma generation. Wyss (1999) suggests that magmatic underplating in a large zone of extension in the N Indian border of the Gondwana plate is a possible mechanism of this magma generation.

The Muth Formation (Stoliczka 1965)
Synonyms: Wazura and Muth Formations (Srikantia & Bhargava 1983), Kenlung Formation (Nanda & Singh 1977)
(Devonian)

The white Muth quartzites are ~80 m thick and constitute an important marker-horizon, which may be followed from Kashmir to Nepal (Gansser 1964). A littoral foreshore environment is suggested for their deposition by the flat, parallel stratification and wave and current ripples (Elliott 1986), by a bimodal granulometry of the sands (Brookfield 1993) and aeolian dune cross-beds structures (Gaetani et al. 1986). Fossils are extremely rare in this mature quartz-arenite. A Lower Devonian age is generally acknowledged because of its stratigraphic position. Draganits et al. (2001) discovered abundant trace fossils in the Lower Devonian Muth Formation of the Pin Valley in Spiti. The ichnoassemblage consists of *Palmichium antarcticum*, *Diplichnites gouldi*, *Diplodichnus biformis*, *Taenidium barretti*, *Didymaulichnus* cf. *lyelli*, *Didymaulyponomos* cf. *rowei* and *Selenichnites* isp.

The Lipak Formation (Hayden 1904)
Synonyms: Kashmir: Syringothyris Limestones (Middlemiss 1887), Aishmuquam Formation (Srikantia & Bhargava 1983), Zanskar: Tanze Formation/Members A+B (Nanda & Singh 1977)
(Middle Devonian-Early Carboniferous)

The Lipak Formation stratigraphically overlies the Muth quartzite from the Spiti Valley through the Kunzum La, Chandra valley and Baralacha La region. It is essentially composed of a sequence, up to 250 m thick, of calcarenites, limestones and evaporites (gypsum) with conodonts of Middle Devonian (Givetian) to Early Carboniferous (Tournaisian) age (Draganits et al. 2002), and was deposited on a low energy platform with lagoons (Hayden 1904, Vannay 1993). Synsedimentary normal faults intruded by basaltic dikes are common in the Baralacha La region. This Baralacha La dike swarm is related to a major Carboniferous transtensional structure (Vannay 1993).

The Early Carboniferous continental basalts (Vannay & Spring 1993)

The basaltic dikes in the Baralacha La region (Baralcha La dike swarm) have tholeiitic to alkaline compositions and they correspond to a comagmatic suite, which evolved mainly by fractional crystallisation processes. They are not feeder dikes for the tholeiitic Permian Panjal traps, which have a different continental flood basalt composition (Vannay & Spring 1993). It was not possible to date the metabasalts radiometrically, because they have a Tertiary greenschist facies overprint (albite, chlorite, biotite, stilpnomelane, actinolite, epidote, titanite).

The Po Formation (Hayden 1904)
Synonyms: Kashmir: Fenestella Series (Middlemiss 1910), Ganeshpur Formation (Srikantia & Bhargava 1983), Zanskar: Tanze Formation/Member C (Nanda & Singh 1977)
(Late Carboniferous)

Hayden (1904) introduced the term Po Series for a 220 m-thick sequence of sandstones, siltstones and black shales that stratigraphically overlie the Lipak Formation. According to Gaetani et al. (1990), these sediments were deposited in a shallow marine high-energy environment with an important terrigenous component. Bivalves and plant fossils indicate an Early Carboniferous age (Tournesian-Visean?, Hayden, 1904, Gothan & Sahni, 1937, Gaetani et al. 1986, 1990). No diagnostic fossils have been found in the upper black shale member.

The Ganmachidan Formation (Srikantia et al. 1980)
Synonyms: Kashmir: Agglomeratic slates (Middlemiss 1910), Pindobla Formation (Srikantia & Bhargava 1983), Zanskar: Ralakung Formation/Phitsi Member (Nanda & Singh 1977), Chumik Formation (Gaetani et al. 1990), Spiti: Losar Diamictites (Ranga Rao et al. 1982)
(Early Permian)

Quartz arenites, conglomerates and black shales of the Ganmachidam Formation are about 150 m thick and overlie the Po black shales and siltstones with an angular unconformity (Hayden, 1904, Fuchs, 1982, Vannay, 1993). In the correlative Losar diamictites of the Upper Spiti valley, Ranga Rao et al. (1987) identified fossils of Early Permian age and a similar Early Permian (Sakmarian) age has been attributed by Gaetani et al. (1990) to the Chumik Fm. in the Lingti valley, SE Zanskar. The formation is divided into a lower member composed of quartz arenites and conglomerates and an upper member of a polygenic conglomerate, sand- and siltstones and black shales. The Ganmachidam Fm. corresponds to a sequence of coarsening upward sediments corresponding to a distal fluvial deposit or a littoral delta.

The Panjal Traps (Lydekker 1878)
Synonyms: Panjal Volcanics, Panjal Unit, Ralakung Volcanics (Singh et al. 1976), Sankoo Volcanics (Gupta et al. 1983), Phe volcanics (Srikantia et al. 1980)
(Late Carboniferous-Early Permian)

The Panjal Traps represent important continental flood basalts that in Kashmir (Lolab valley, Erin valley and Gulmarg) reach a thickness of 2'500 m deposited under subaerial, marginal marine to terrigenous conditions (Pareek 1983). Their thickness is gradually reduced to the SE and the last Panjal traps are observed SE of the Baralacha La in the Upper Chandra valley (Lydekker 1878, Vannay 1993). The Panjal Traps display relatively primitive tholeiitic and alkaline compositions (Nakazawa & Kapoor 1973, Singh et al. 1976, Bhat and Zainuddin 1979, Honegger et al. 1982, Gupta et al. 1983, Pareek 1983). These rocks originated from an "enriched" P-MORB-type magma, which underwent a limited magmatic evolution by fractional crystallisation with probable crustal contamination (Thompson et al. 1984, Hawkesworth et al. 1990). The volcanic activity began in the Late Carboniferous, producing several hundred meters of intermediate to acid pyroclastics and welded tuffs (Pareek 1983), called the "agglomeratic slate", which overlies Carboniferous shales and shallow water limestones. Stratigraphic relations indicate an age between the Artinskian and Kungurian for the main lava flows (Kapoor 1977, Nakazawa et al. 1975).

The Yunam granite (Spring et al. 1993)
(Early Permian)

A porphyritic granitic dike, 2-15 m thick, crosses the Yunam valley south of Sarchu. The granitic dike has an alkaline to subalkaline composition and yielded an Early Permian U-Pb age on zircon of 284 ± 1 Ma. This alkali granite intrusion may be related to the Early Permian Panjal Trap magmatic event.

The Neotethyan shelf

A strong thermal subsidence of the N-Indian margin started after the eruption of the Panjal Traps concomitant with the Late Permian marine transgression and the opening of the Neo-Tethys ocean (Gaetani & Garzanti 1991, Garzanti 1993, Stampfli et al. 2001).

The Kuling Formation (Stoliczka 1865)
Synonyms: Kashmir: Zewan Beds (Middlemiss 1910), Zewan Formation (Nakazawa et al. 1975 Nakazawa et al. 1981, Srikantia & Bhargava 1983), Lahul: Sarchu Formation (Nanda & Singh 1977), Spiti: Productus Shales (Hayden 1904)
(Late Permian)

Stoliczka (1865) proposed the term Kuling Shales for a sequence of black shale, marls and calcarenites deposited on the Early Permian conglomerates of the Spiti valley. Farther NW these shales overlie the Early Permian Panjal Traps. The Kuling calcarenitic shales were later subdivided in the lower Gechang Member composed of quartz arenites, calcareous bioclastic sandstone, siltites and black shale, and upper Gungri Member composed of siltites and black shales by Srikantia et al. (1980), a nomenclature adopted by many geologists (Fuchs 1982, 1987, Ranga Rao et al., 1982, Baud et al. 1984, Nicora et al. 1984, Gaetani et al. 1986, 1990, Vannay 1993). In the Spiti Valley and the Baralacha La region, the lower member has a thickness of about 65 m and the upper member about 25 m (Vannay 1993). The thickness of the formation rapidly increases to some 100-1000 m in more distal regions of the N Indian shelf with an increase in limestone and dolomite sedimentation (Stutz & Steck 1986, Stutz 1988, Steck et al. 1998). A rich brachiopod fauna with *Lamnimargus himalayansis, Spirifirella rajah* indicate a Early Dzhulfian (middle Tatarian) age for the Gechang Member of the Kuling Formation (Hayden 1904, Nakamua et al. 1985, Gaetani et al. 1990). The correlative Permian Zewan Formation exposed in the Guryul ravine and near Palgham east of Srinagar, was deposited with a stratigraphic gap on the Panjal Traps and has yielded late Midian, Dzhulfian and early Changhsingian ages (Nakazawa et al. 1975, Baud et al. 1996).

The Lilang Group (Hayden, 1904)
(Triassic)

The Spiti valley is a classic area of exposure for the Triassic Lilang Group, about 1'200 m thick, that forms the carbonate platform of the Neo-Tethyan Indian shelf. Most of the stratigraphic horizons are dated by fossils (Hayden 1904, Gansser 1964, Srikantia 1981, Fuchs 1982, Bhargava 1987, Gaetani et al. 1986, Jadoul et al. 1990 and Garzanti et al. 1995). The conodonts of the Triassic sequence of the Spiti valley have been described by Garzanti et al. (1995) whose nomenclature is very similar to the classic stratigraphic subdivisions proposed by Hayden (1904).

The Tamba Kurkur Formation (Srikantia 1981)
(Scythian-Anisian)

The Permian/Triassic boundary is marked by a major break in sedimentation, of up to several My, and is represented by a lateritic horizon in Spiti and an angular unconformity (Hayden 1904, Bhatt et al. 1980, Srikantia 1981, Garzanti et al. 1995). East of Srinagar in Kashmir (Nakazawa et al. 1975, Baud et al. 1996) and Lossar in the Spiti Valley (Hayden 1904, Hugo Bucher personal communication), an alternation, about 14m thick, of centimetric-decimetric black shale and limestone layers ("chocolate limestone") the Scythian and Anisian *Otoceras*, *Ophiceras* and *Hedenstroemia* beds and about 20 m of Nodular limestone overlie the unconformity with the underlying Permian *Productus* Shale (Kuling Formation). They are followed by the competent Niti Limestone (Diener 1908), the Middle Triassic "Muschelkalk" of an Anisian to early Ladinian age that is about 8 m thick.

The Hanse Formation (Srikantia et al. 1980)
Synonyms: Kashmir: Khreu Formation (Nakazawa et al. 1975), Zanskar: Zangla Formation/Middle Member (Nanda & Singh 1977), Garhwal-Kumaun: Kuti Shales (Heim & Gansser 1939)
(late Ladinian-late Carnian)

The Upper Triassic begins with the Hanse Formation that is about 350 m thick and consists of alternating regularly stratified dark limestone, dolomite and marl layers deposited under pelagic to semi pelagic conditions with an important detrital contribution. *Daonella* sp. (Daonella shale and limestone, Hayden, 1904, Bhargava, 1987) indicate a late Ladinian to Carnian age (Hayden, 1904, Srikantia et al. 1980, Gaetani et al. 1986, Garzanti et al. 1995).

The Nimaloksa Formation (Srikantia et. 1980)
Synonyms: Spiti: Tropites Beds (Hayden 1904, Fuchs 1982), Sanglung Formation (Bhargava 1987)
(late Carnian-early Norian)

The Hanse Formation is conformably overlain by 300 m of Carnian to earliest Norian limestone and dolomite with terrigenous intercalations. Its upper massive member, with some megalodontids, also called the Zozar Formation (Gaetani et al. 1986), forms 150 to 350 m high cliffs in the Zanskar area.

The Alaror Group (Garzanti et al. 1995) or Quartzite Formation (Gaetani et al. 1986)
(Norian)

The Norian Quartzite Series (Gaetani et al. 1986, Jadoul et al. 1990) are traditionally subdivided into the 100-200 m-thick Juvavites Beds, the 15–30 m-thick Coral Limestones, the 90-160 m-thick Monotis Shale, and the 30-100 m-thick Quartzite Series, originally defined by Hayden (1904). These classical lithostratigraphic subdivisions, display an upward increase in sandstone with a coral-bearing carbonate band in the middle (Bhargava 1987). Crossbed structures in the quartzite layers are useful polarity markers for structural work.

The brown coloured Norian Juvavites Beds are muddy terrigenous to calcareous shelf sediments with grainstones and oolitic ironstones in its middle part. Bioclasts, commonly concentrated in storm lags, include pelecipods, crinoids, brachiopods, corals, gastropods, ostracods, algae and *Juvavites*. They represent a good marker horizon.

The Coral Limestones form a discontinuous marker unit of grey nodular bioclastic limestones, with interbedded ferruginous siltstones, phosphate nodules and arkoses.

The Monotis Shale is composed of shelf mudrocks, arenites and storm deposited sandstones.

The Quartzite Series consists of fine-grained grey arkoses and green and pinkish quartzarenites with dolomitic to phosphatic clasts, bioclasts and some times Megalodons.

The Kioto Formation (Hayden 1904)
(late Norian-Liassic-Aalenian)

The competent subtidal Kioto limestones and dolomites are 400–900 m thick, contain abundant *Neomegalodon* in some beds, and form spectacular cliffs throughout the NW Himalaya from the Spiti valley to the Zanskar area. They were deposited on a carbonate platform, ranging from late Norian, Liassic to Aalenian in age (Gaetani & Garzanti 1991).

The Laptal Beds (Heim & Gansser 1939)
(early Callowian)

The top of the Kioto platform is truncated by a major unconformity and after a period of at least 10 m.y. was progressively onlapped by the Ferruginous Oolite of the Laptal Formation during the early Callovian global sea level rise (Haq et al. 1987, Gaetani & Garzanti, 1991). The Laptal Beds are characterized by fossiliferous ironstone intervals enclosing a coarsening-upward shale-sandstone sequence deposited on a storm-controlled inner shelf. In the regions where the Laptal Beds are missing, the surface of the Kioto carbonates often shows a zone of red alteration.

The Spiti Shale (Stoliczka 1865)
(Oxfordian-early Berriasian)

After a period of reduced sedimentation and frequent gaps in the late Callovian to early Oxfordian, the Laptal Beds were followed until the early Berriasian time by deposition of offshore pelites, the Spiti shale, on the slowly subsiding N Indian shelf. The thickness of the Spiti Shale varies between 10-30 m in NW Zanskar and 200 m in the Spiti area to the SE (Gaetani & Garzanti 1991).

The Giumal Sandstones (Stoliczka 1865)
(Early Cretaceous)

The offshore Spiti Shale is overlain by the Aptian and Albian Giumal Sandstones, which recorded the multiphase progradation of deltaic clastics on the Zanskar shelf. The Giumal Sandstones form a 200-300 m-thick sequence of grey locally cross or graded-bedded, fine-grained sandstones inter-bedded with black shale. An Early Cretaceous (Valanginian-Albian) volcanic extensional event is indicated by abundant volcanic detritus and local basaltic lavas of alkaline and tholeiitic suites in a large part of the N Indian passive margin, from Zanskar in the NW to Nepal in the SE and correlates with the 115 Ma old Rajmahal Traps of north-eastern India. This magmatism is related to an Early Cretaceous break up episode that preceded the final opening of the Indian ocean (Garzanti et al. 1987, Garzanti 1993, Gaetani & Garzanti 1991).

The Chikkim Formation (Stoliczka 1865)
Synonyms: Zanskar: Fatula Limestone (Gaetani et al. 1985)
(Cenomanian-Campanian)

During the Cenomanian, the outer continental terrace was covered by 100 m light grey to multi-coloured pelagic foraminiferal limestone that formis the white Chikkim cliff in the upper Spiti valley (Stoliczka 1865, Hayden 1904) and the multicoloured Fatula Limestone in Zanskar (Gaetani et al. 1985). This early Late Cretaceous sedimentation marks the onset of the opening of the Indian ocean and the N directed drift of India (anomaly 34, 80 Ma, Campanian, Patriat et al. 1982, Patriat & Achache 1984).

Kangi La Shale (Fuchs, 1977)
(Synonym: Goma Shale, Fuchs, 1977)
(Late Cretaceous)

In Zanskar, the Fatula Limestones are overlain by the Upper Cretaceous Kangi La (Goma) Shales, a 400-600 m-thick sequence of dark grey marls, calcareous siltstones and sandstones in the upper part that was deposited in a bathyal to sublittoral environment (Baud et al. 1984, Garzanti et al. 1987).

Stumpa Quartzites (Gaetani & Garzanti 1991)
(Danian)

The Paleogene succession begins with 10-100 m of white, brown-weathered Stumpa quartzarenites, onlapping an underlying unconformity (Garzanti et al. 1987). These coastal sandstones have a Danian age (Gaetani et al. 1986).

The Shinge La Formation (Garzanti et al. 1987)
(synonyms: Spanboth Limestone, Fuchs 1982, Dibling Limestone, Garzanti et al. 1987)
(Paleocene)

The carbonate sedimentation on the N Indian shelf ends in the Late Paleocene with the 200 m-thick open shelf Dibbling Limestone and oceanward Shinge La pelagic limestone, that is up to 500 m thick. The latter formed on the outer part of the continental terrace and was unconformably overlain by 0-120 m of nummulitic Kesi Limestone. These sediments are capped by an early Eocene hardground (Garzanti et al. 1987).

The Kong Formation (Garzanti et al. 1987)
(synonym: Kong Slates, Fuchs 1982)
(Ilerdian-Ypresian)

The Kong Formation is composed of marine slates and siltstones rich in very fine-grained volcanic detritus with nummulites and assilinas of an Ilerdian and early Eocene (Ypresian) age. These sediments record the erosion of obducted oceanic crust and the collision of the Indian passive margin with Asian active margin. These observations corroborate the paleomagnetic data, which suggest continental collision at about 50 Ma (Patriat & Achache 1984)

Paleogene Olistostrome

To the south of Dat in the Karnagh region, the Cretaceous Giumal Sandstones are overlain by debris-flow conglomerates containing limestone pebbles of Cretaceous to Paleocene age (Fuchs 1986, Garzanti et al. 1987, Steck et al. 1993).

The Neotethyan slope

The Lamayuru Formation (Frank et al. 1977, Fuchs 1977)
Synonym: Lamayuru Flysch (Frank et al. 1977, Fuchs 1977), Markha Unit (Baud et al. 1982).

The sedimentation on the N Indian slope is characterized by a heterogeneous sequence of turbiditic calcarenites and marls with rare fossils, strongly deformed during the Himalayan collision (Stoliczka 1865, Lydekker 1883, Frank et al. 1977a, Fuchs 1977, Bassoulet et al. 1980, Bassoulet et al. 1983, Stutz & Steck 1986, Stutz 1988). Near Lamayuru, the Lamayuru Flysch is composed of a repeated alternation of grey to black siltstone and calcareous grey siltstone with frequent olistolites, olistolite breccias that contains Jurassic microfossils (Bassoulet et al. 1983). The Lamayuru Flysch is exposed all along the Indus suture, from the Dras valley in the NW to the Tso Morari region in the SE (Fuchs & Linner 1996). In the Markha valley, Stutz (1988) distinguished and described different formations that range from Early Permian to Late Cretaceous in age: they include calcareous grey schists of the Luchungtse Formation, over 200 m thick, of a probably Permian age, over 200 m of calcarenites and shales of the Late Triassic to Early Jurassic Dolto Formation and 160 m of calcarenites and calcirudites with phosphorite concretions of Middle Jurassic to Late Cretaceous age. On the Geological map (this paper), the Lamayuru Formation includes sediments of a Permian to Late Cretaceous age of the Neotethyan slope. Georges Mascle (personal communication) has suggested Carboniferous sediments near Karzok Gompa may also belong to these rocks. Virdi et al. (1978) described Permian conodonts from the Taglang La region that may also be included. According to Fuchs & Linner (1996), the Lamayuru Formation of the Shingbuk La region is followed by dated Cretaceous clastic flysch with a few foraminiferal marl beds and nummulitic limestones at the top. Honegger (1983) observed alkaline basaltic sills and volcanites of a Triassic age in the Lamayuru Flysch of the Suru Valley region.

Conclusions on the composition and history of the Pre-Himalayan North Indian continental margin

A palinspastic section of the North Indian continental margin and typical stratigraphic sections are proposed in Figs. 3 and 4. The 2.4-3.2 Ga old rocks of the Archaen craton are exposed in the southern part of the Indian subcontinent (Le Fort et al. 1986). In N India they probably form the basement for the early Proterozoic siliciclastic sediments and bimodal magmatic series of the Aravalli range and the Lesser Himalaya. Archean rocks have never been found in the Himalayan range. The Lesser Himalayan units are mainly composed of Proterozoic sediments and magmatic rocks. In N India, they are overlain by the Middle and late Proterozoic sediments of the Vindhyan Basin a southern equivalent of the Riphean, Vendian to Early Cambrian Shali limestones, Simla-Dogra slates, glaciomarine Blaini boulder slates, Krol limestone and dolomite and Tal sandstone of the Lesser Himalaya and Simla-Krol belt. In the Shali and Simla region they are overlain with an angular unconformity by the Paleocene-Eocene nummulitic Subathu limestone and slates (Pilgrim & West 1928, Auden 1934, Gansser 1964, Brookfield 1993). The tectonic units of the High Himalaya, the Shikar Beh nappe, the High Himalayan nappe or Crystalline nappe and the North Himalayan nappes, are composed of a similar stratigraphic sequence of late Proterozoic to Late Cretaceous age, locally overlain by younger Cenozoic sediments, related to the closure of the Neo-Tethys ocean. The most difficult task is the reconstruction of the geological environment of the early Proterozoic siliciclastic sediments and bimodal magmatic series of the Lesser Himalaya, exposed in the Kishtwar and Larji-Kullu-Rampur windows and in the strongly deformed Lower Crystalline, Chail and Simla-Runkum nappes of alpine age. The siliciclastic sediments were intruded in a lower part by 1.84 Ga granitic and 1.80 Ga mafic magmatic rocks. The Nd depleted mantle model ages on the peraluminous granitic rocks extend to 2.63 Ga, indicating recycling of the older crust and Early Proterozoic to Late Archaen sources (Miller et al. 2000). It is difficult to know, whether this sequence suffered any Proterozoic (Grenville?) orogenic event. Frank et al. (1995) correlate the Vendian Blaini tillite and the overlying Kroll carbonate platform of the Lesser Himalaya with the Manjir tillite and the Upper Haimantas detrital sediments of the High Himalaya. The abundance of fossils in the Phanerozoic makes for very precise stratigraphic interpretation. Fig. 4 proposes a palinspastic cross-section of the Late Triassic

Neotethyan margin of NW India. The principle features of this reconstruction are the following:

1. In contrast to the Alps, where two major angular unconformities cutting through folded and metamorphosed older rocks testify of the Caledonian and Variscan orogenic events, in the Himalaya, no Paleozoic compressional structures are observed. Gansser's conclusion (1964): "It is this conformable sedimentary sequence of Spiti, from Precambrian onwards, which reflects a calm epeirogenic pre-Alpine history of the "Tibetan" Himalaya" that is valid for the whole stratigraphic sequence of the High Himalaya.
2. The early Proterozoic Rampur Formation forms the basement of the Riphean to Early Cambrian sedimentary sequence in the Lesser Himalaya and the Lower Crystalline nappe.
3. The Riphean to Early Cambrian sedimentation of the High Himalayan domain was a long lasting anorogenic detrital accumulation that is over 10 km thick in the Haimanta sequence on the N Indian border (Frank et al. 1995).
4. The glaciomarine pebbly diamictite of the Manjir Formation (Middle Haimantas) represents an important stratigraphic marker horizon of a Vendian age in this monotonous sequence. The Blaini boulder beds in the Lesser Himalayan units are considered to be the lateral equivalent of the Manjir boulder beds in the High Himalayan units (Frank et al. 1995, Draganits et al. 1998).
5. Strong erosion and thermal uplift are indicated by the angular unconformity at the base of the Ordovician Thaple conglomerates. The lower crust is intruded by Ordovician bimodal calc-alcaline but mainly granitic magmatic rocks at 500 Ma (Girard & Bussy 1999, Wyss et al. 1999). At this time, the Cimmerian blocks, like South Tibet and Afghanistan were still attached to the N Indian continental margin. This Ordovician extensional event may be related to the Paleo-Tethys rifting (Stampfli et al. 2001).
6. Early Carboniferous synsedimentary normal faults and a basaltic intraplate magmatism are related to transtensional movements in the N Indian margin suggesting an aborted Carboniferous rift preceding the main Permian Neo-Tethys rifting (Vannay 1993, Vannay & Spring 1993).
7. The emplacement of the Middle Permian Panjal Trap continental flood basalts preceded the opening of the Neo-Tethys ocean. The Cimmerian blocks, Afghanistan and S Tibet, progressively separate from N India. A strong thermal subsidence of the N Indian flexural margin with the formation of the Permian-Cretaceous carbonate platform that began in Middle Permian time (Vannay 1993, Stampfli et al. 2001).
8. The Neo-Tethys margin of N India possesses the geometry of an upper flexural margin (Fig. 4, Wernicke 1985, Voggenreiter et al. 1988). The rift shoulder is exposed in the Pir Panjal range, between the Mesozoic Tandi syncline and the Chamba basin (Gaetani & Garzanti 1991, Steck et al. 1993, Vannay 1993, Stampfli et al. 2001).
9. During the Early Cretaceous, abundant volcanic detritus mixed with quartzose siliciclastics in the Albian shelf sand bodies of the Giumal Sandstone and by a local occurrence of basaltic lava testify to a major volcanic event affecting the northern Indian margin, from Ladakh to Nepal (Bordet et al. 1971, Raina & Bhattacharyya 1977, Sakai 1983, Gaetani et al. 1985, Garzanti et al. 1987). This Early Cretaceous rifting may be a precursor of the NE-directed drift of the Indian plate in the Late Cretaceous. In the Campanian, India began moving away from Africa and synchronously a convergent plate boundary developed along the southern Eurasian margin (anomaly 34, 80 Ma, Senonian; Norton & Sclater 1979, Patriat et al. 1982). The sedimentary record of the period of convergence is exposed in the youngest (Ypresian and early Lutetian) sediments of the North Himalayan nappes of the Zanskar region (Zangla and Lingshed nappes, Garzanti et al. 1987, Gaetani & Garzanti 1991, Rowley 1996). According to Jaeger et al. (1989) and Rage et al. (1995) terrestrial faunas of the Cretaceous/Tertiary boundary age (65.7 ± 2 Ma) in India are similar to coeval faunas of Eurasia and have been inferred to imply collision by 65 Ma. The 55 Ma age of the Tso Morari eclogites suggests that in the NW Himalaya, the subduction of the Indian crust below Asia is of the same age (55 ± 7 Ma Sm-Nd age on grt-gln-whole rock, 55 ± 12 Ma Lu-Hf age on grt-cpx-whole rock and 55 ± <17 Ma U-Pb age on alanite, De Sigoyer et al. 2000). The rapid drop of the convergence velocity from about 15 cm/a to 5 cm/a approximately 50 Ma ago in the early Eocene (Ypresian) (Patriat & Achache 1984) may be another consequence of the continental collision of India with Asia.
10. The time of ophiolite obduction is determined by the first arrival of volcanic and debris from ocean crust from the obducted ophiolites and the accreted Dras arc in the Early Eocene Chulung La Formation of the Zanskar shelf (Garzanti et al. 1987) and the latest Paleocene and Eocene Subathu Formation of the Indus-Ganga fore arc basin (Najman & Garzanti, 2000). Before this time, all detritus deposited on the N Indian shelf was derived from the Indian passive margin to the N (Garzanti et al. 1987). At the time of the Early Cretaceous deposition of volcanic detritus in Giumal sandstones, the N Indian margin was still separated from the active Eurasian margin by the more than 2000 km wide Neo-Tethys ocean. The occurrence of Maastrichtian to Ypresian fossils, in the ophiolitic melange underlying the Spongtang ophiolite klippe, proves that the ophiolite obduction onto the Indian passive margin ended in the late Paleocene (Colchen & Reuber 1987). Thus the classic hypothesis of a Cretaceous continental collision and ophiolite obduction (Searle 1983, 1986) must be revised.

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,

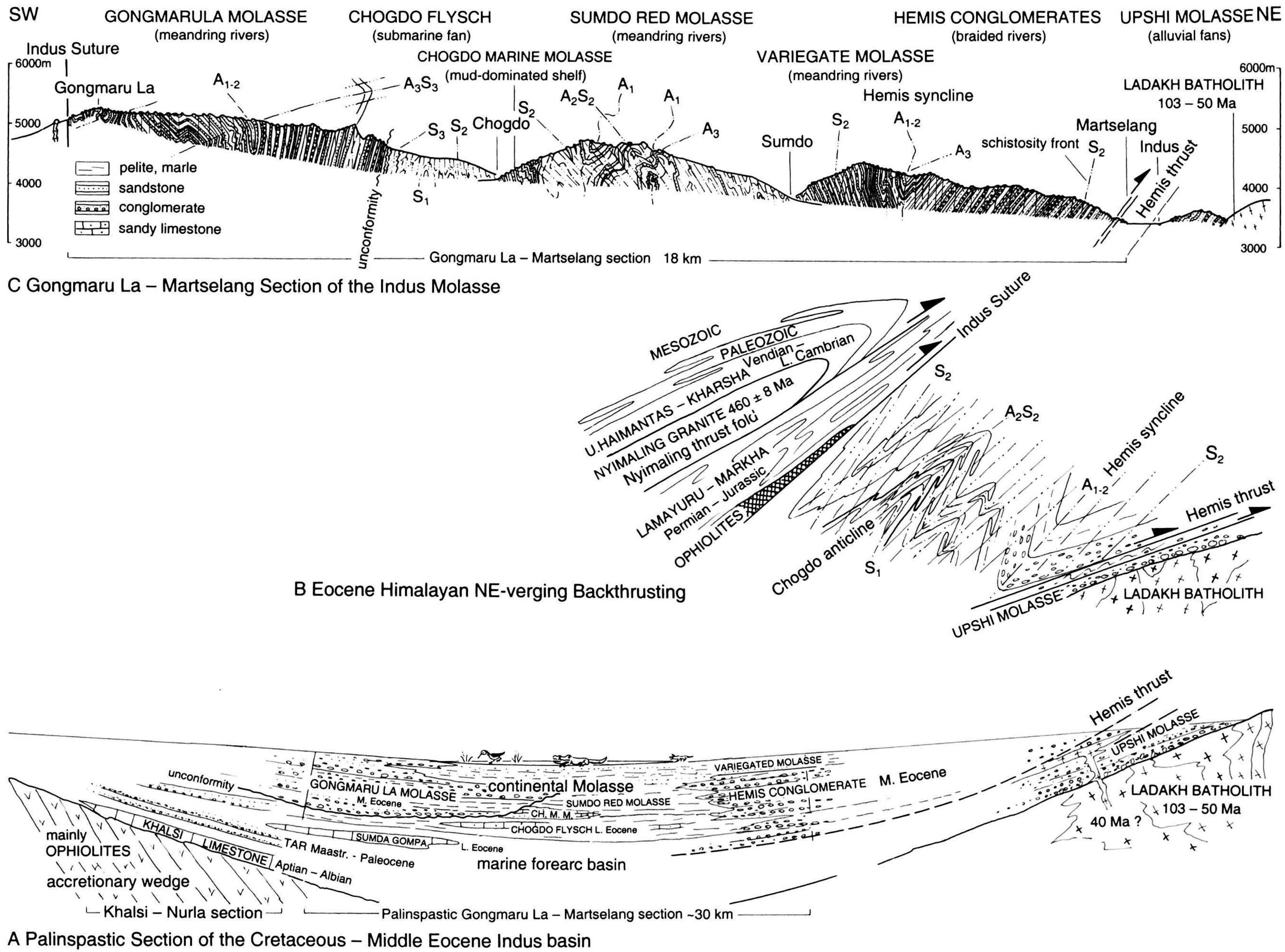

Fig. 5. Geological section of the Gongmarula-Martselang transect through the Indus Molasse (C), the Eocene phase of NE-verging (back) thrusting (B) and the palinspastic section of the Upper Cretaceous – Middle Eocene Indus basin (A) according to Steck et al. (1993).

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

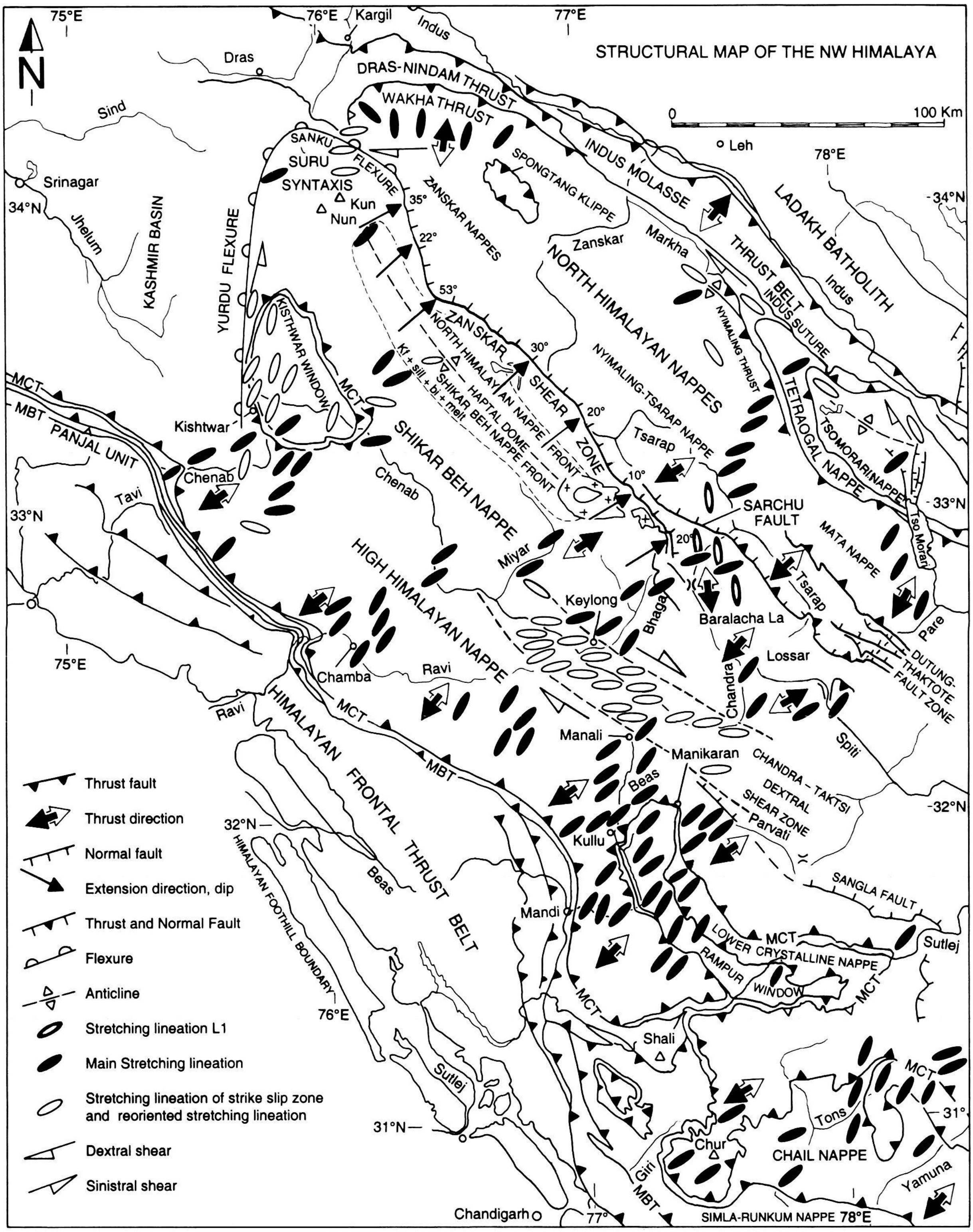

Fig. 6. Structural map of the NW Himalaya, illustrating the main thrusts, extensional faults, dextral and sinistral strike slip zones, stretching lineations and the shear sense deduced from mesoscopic shear criteria. This map represents a compilation of observations by Epard et al. (1995), Epard & Steck (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

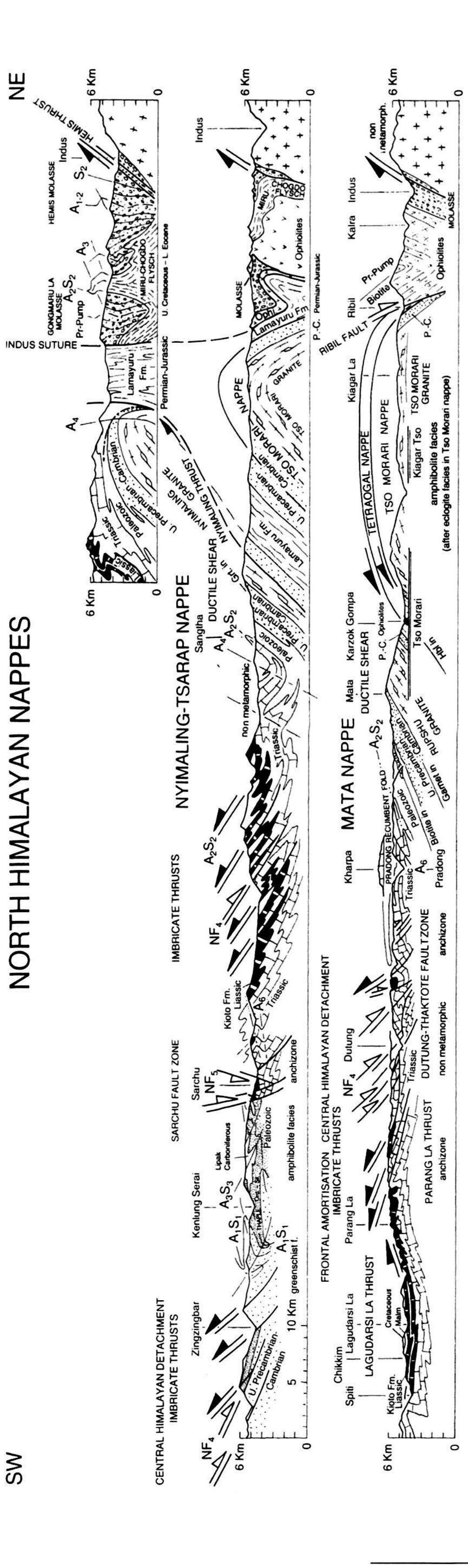

Fig. 7. Three geological sections of the North Himalayan nappes (Nyimaling-Tsarap, Tetraogal, Mata and Tso Morari nappes). The structures of the two S-N oriented and well studied Zingzingbar (Bhaga) – Martselang (Indus) and Chikkim (Spiti) – Tso Morari – Indus transects are projected into a SW-NE plane. 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).

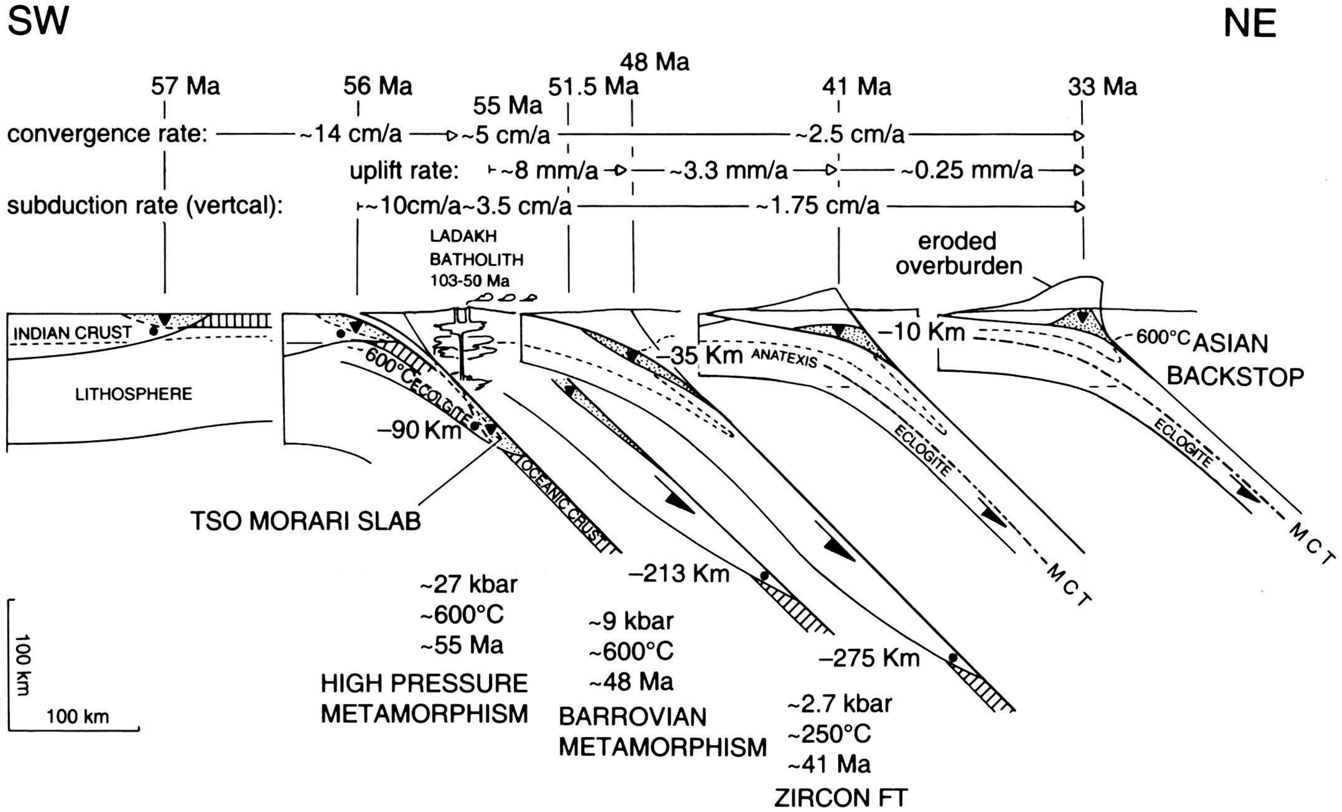

Fig. 8. A kinematic model of the continental collision, the Tso Morari slab subduction and extrusion and the formation of the North Himalayan accretionary wedge (North Himalayan nappes) that occurred between about 56 and 33 Ma during the under thrusting of the North Indian continental margin below Asia. 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.

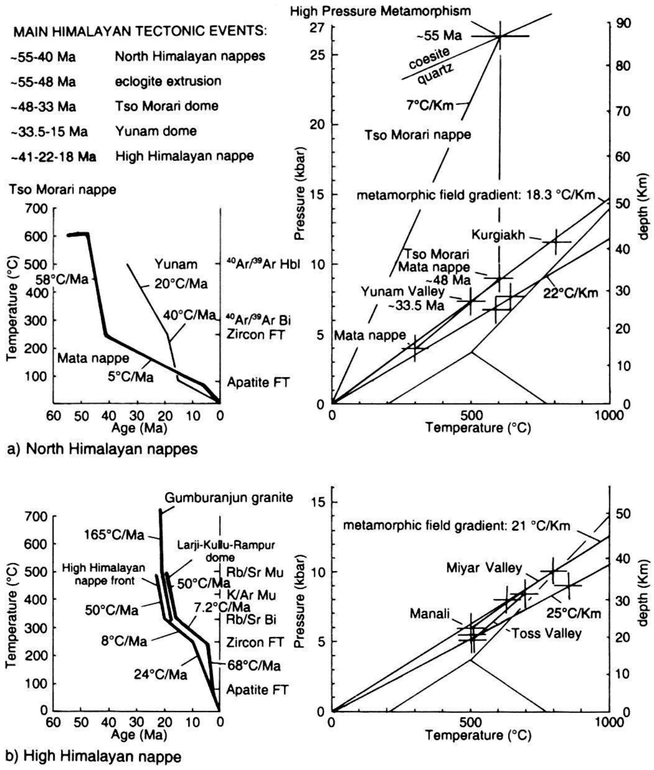

Fig. 9. Temperature-time and pressure–temperature paths of a) the North Himalayan accretionary wedge (North Himalayan nappes) and b) the High Himalayan nappe, based on data from Dèzes et al. (1999), De Sigoyer et al. (2000), Epard et al. (1995), Frank et al. (1977b), Mukherjee & Sachan (2001), Robyr 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

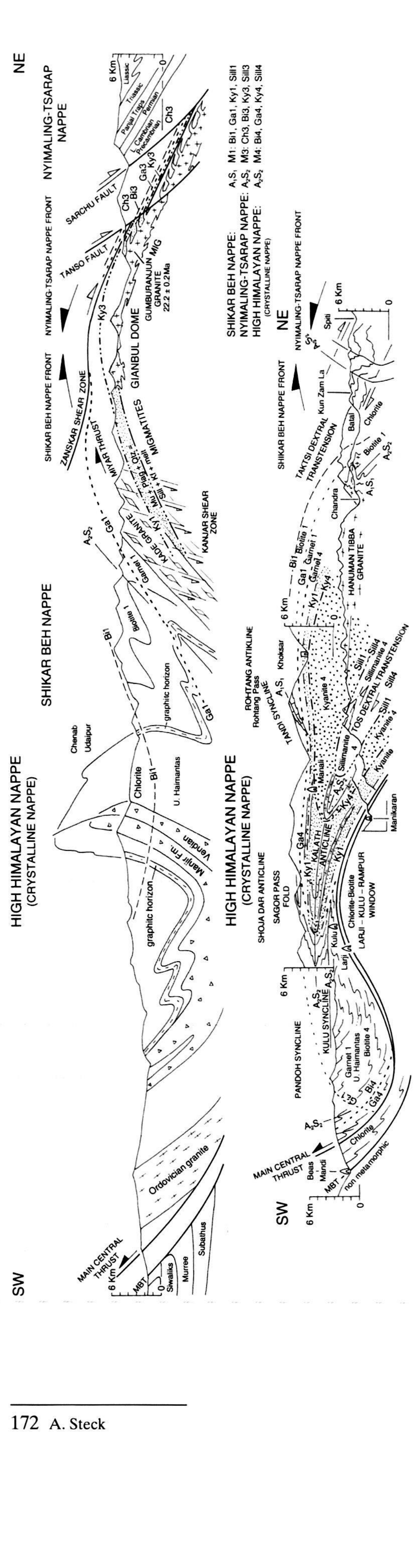

Fig. 10. Geological sections through the High Himalayan nappe (Crystalline nappe). Three phases of Barrovian regional metamorphism may be distinguished in the High Himalaya: the M1 metamorphism of the NE-directed Shikar Beh nappe stack, the M3 metamorphism of the SW-directed Nyimaling-Tsarap nappe stack 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,

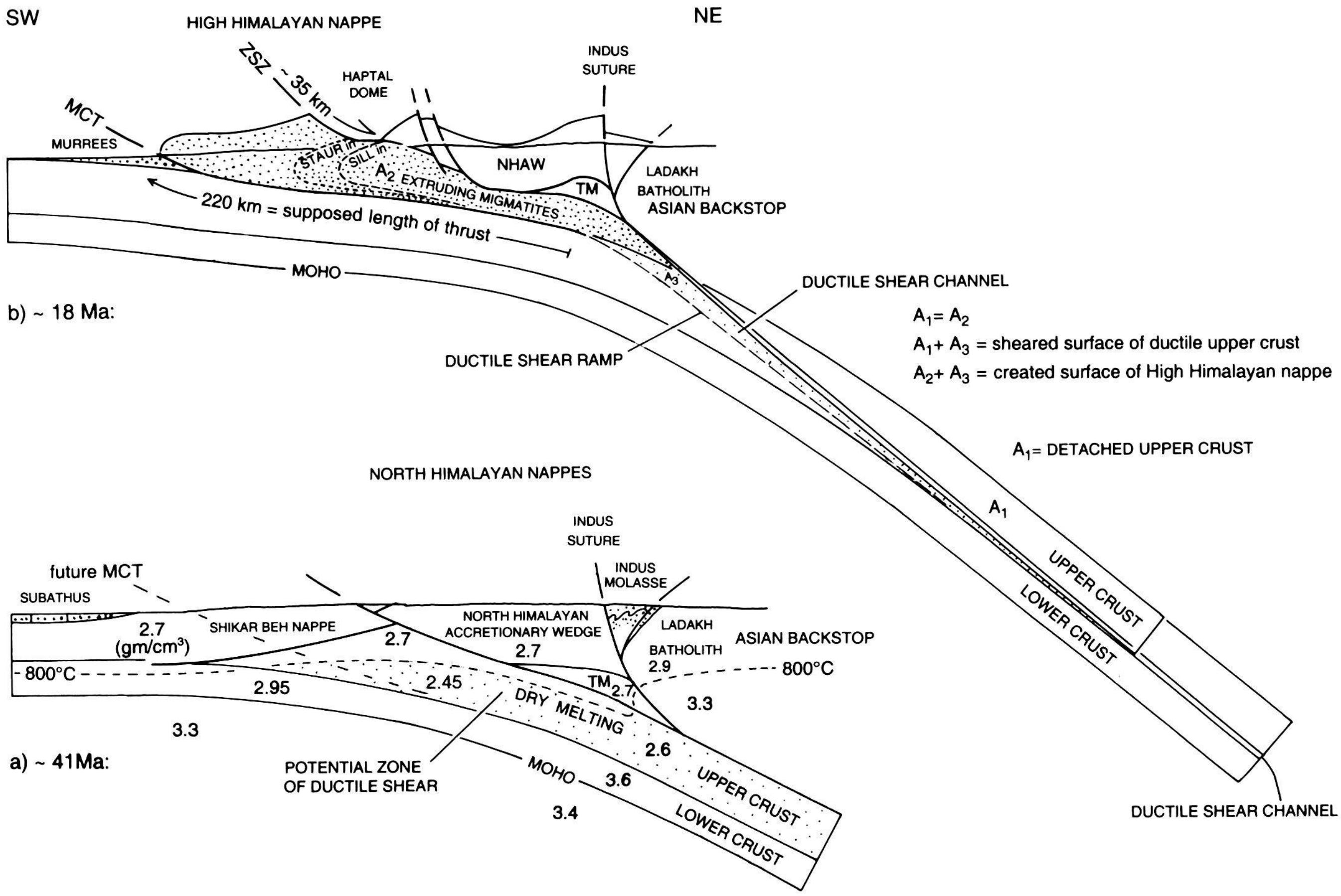

Fig. 11. Empirical model for the formation of the High Himalayan nappe. The lower section illustrates the formation of a dry melting (anatexis) zone in the subducted upper Indian crust below the 55-33 Ma old North Himalayan accretionary wedge and the Shikar Beh nappe stack. The numbers in Fig. 11a are 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-

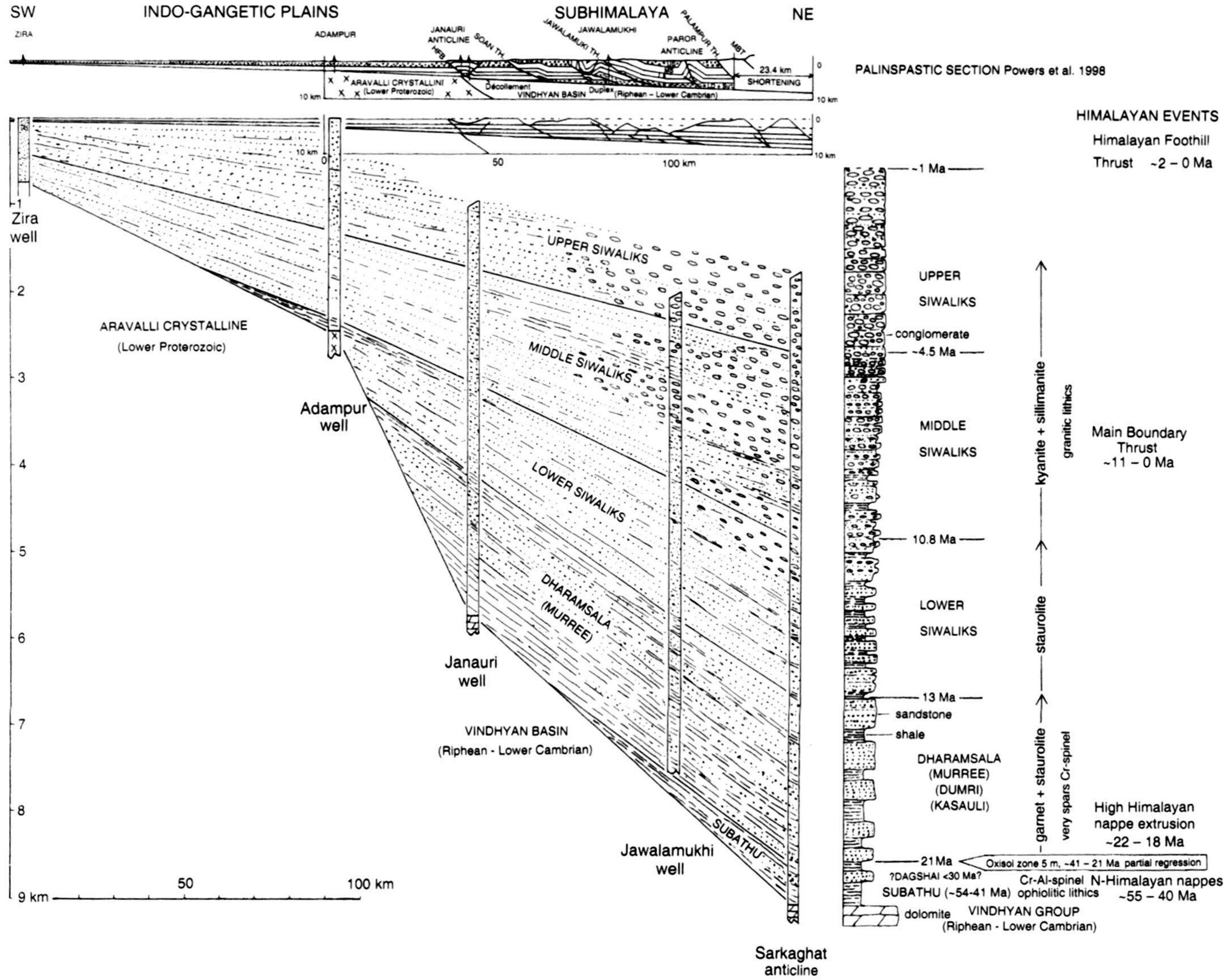

Fig. 12. Palinspastic section of the Subhimalaya molasse basin of the Kangra re-entrant based on surface geology, oil-well seismic reflection data, magnetostratigraphy, sedimentological observations and a balanced and restored cross section by Raiverman et al. (1983, 1993), Lyon-Caen & Molnar, (1985), 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).

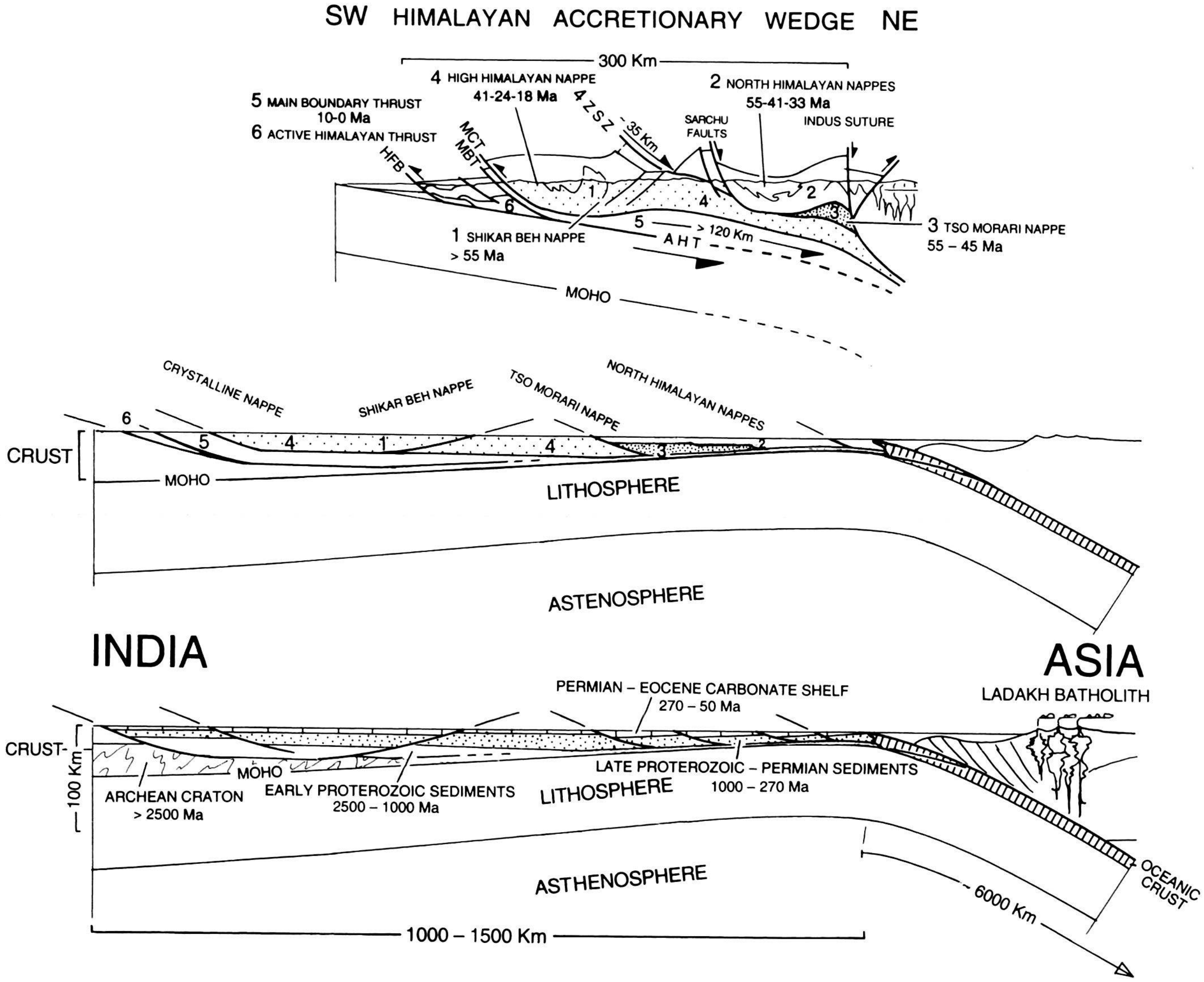

Fig. 13. A geological section of the Himalayan accretionary wedge and palinspastic sections that illustrate the detachment and accretion of successively deeper and older elements of the upper Indian crust during its underthrusting below Asia.

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-

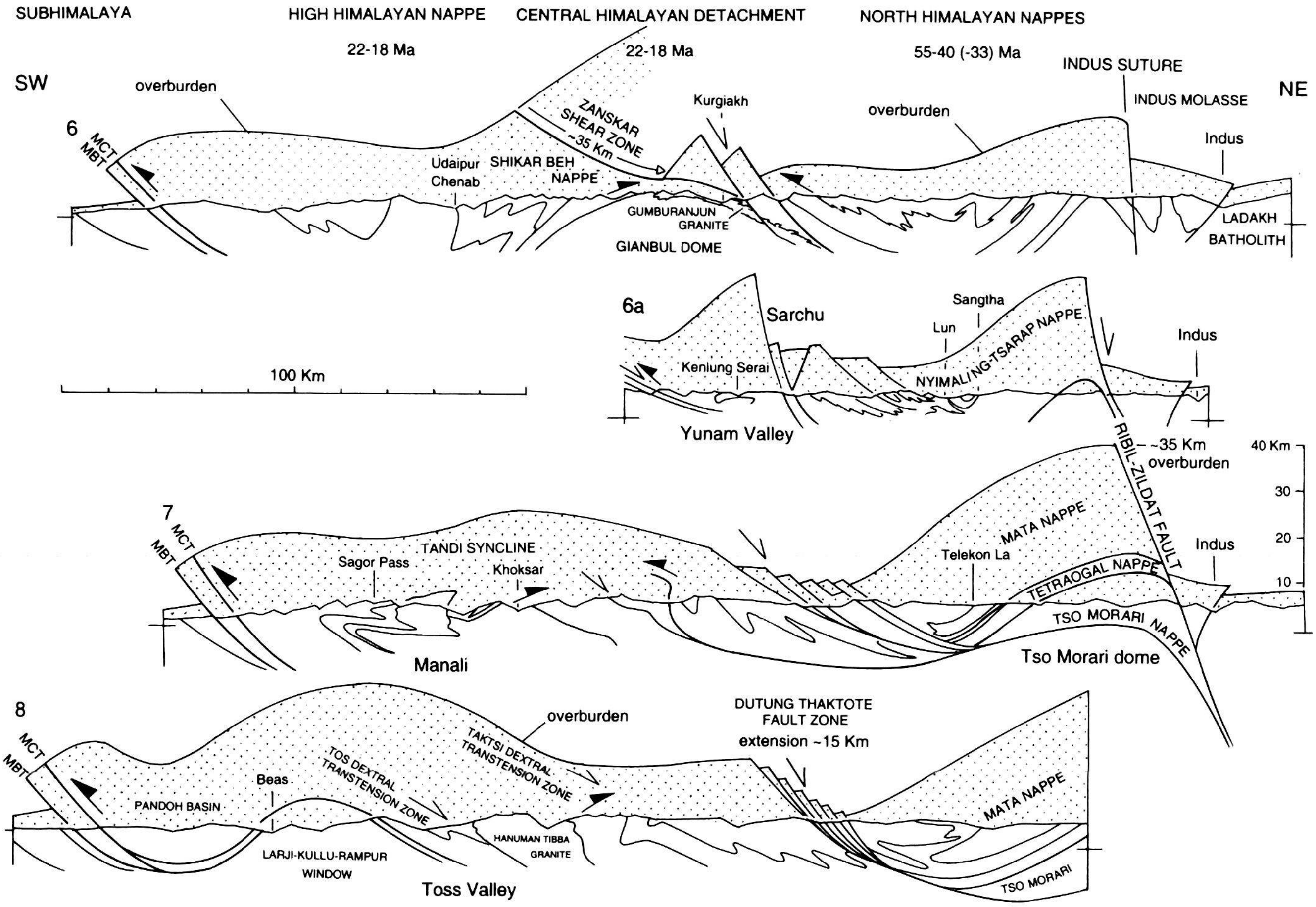

Fig. 14. Estimates of the eroded overburden of the Himalayan accretionary wedge based on thermo-barometric data from Dèzes et al. (1999), De Sigoyer et al. (2000), Epard et al. (1995), Frank et al. (1977b), Girard (2001), Girard et al. (1999, 2001), Robyr et al. (2002), Schlup et al. (2003), Spring et al. (1993a & b), Steck et al. (1993) and Wyss (2000). The numbers 6, 7 and 8 correspond to the cross-sections on Plate 2. 6a follows the Yunam Valley and the Tsara River near Lun and Sangtha. Note the strain partitioning of the Central Himalayan detachment along the strike of the Himalayan range.

graphic sequence ranging from Late Proterozoic to Paleocene age (Hayden 1904, Gansser 1964, Gaetani & Garzanti 1991, Steck et al. 1993, 1998). A pre-Caledonian-granite high grade Barrovian metamorphism in the upper Sutlej valley described by Marquer et al. (2000) is an isolated observation that must be confirmed by new field observations. In the NW Himalaya we studied, we never observed pre-Triassic Variscan or Caledonian crystalline basements. This is a fundamental difference between the Himalaya and the Alps (Steck et al. 2001).

The formation of the Himalayan accretionary wedge

The original position of the main thrust sheets of the Himalayan range are indicated 1 through 6 on the palinspastic section in Fig. 13. The oldest Himalayan thrust sheet is the NE-directed Shikar Beh nappe (1), which is younger than the Liassic limestone of the Tandi syncline and older than the Eocene SW-directed North Himalayan nappe front and the SW-verging structure of the Miocene High Himalayan nappe. The intracontinental Shikar Beh thrust was probably formed by reactivation of an Ordovician, Permian or Mesozoic intracontinental, SW-dipping listric normal fault. The successive detachment of the sediments of the Indian upper crust during underthrusting below the Asian plate is indicated by thrust sheets 2-6. During underthrusting, the sediments were successively detached from the upper Indian crust and accreted on the Asian backstop formed by the Ladakh batholith, starting with the North Himalayan nappes (2) with the external and younger strata and ending in the Lesser Himalayan MBT containing deeper and older sediments of the Indian crust (5). The thrust of the Subhimalayan Ganga molasse (6) and an important seismic activity at a depth of 10-15 km below the MCT (Avouac et al. 2001) and near the Active Himalayan thrust (AHT) demonstrate the ongoing Himalayan orogeny. An estimate of the amount of overburden eroded from the Himalayan accretionary wedge is represented in Fig. 14. The estimate is based

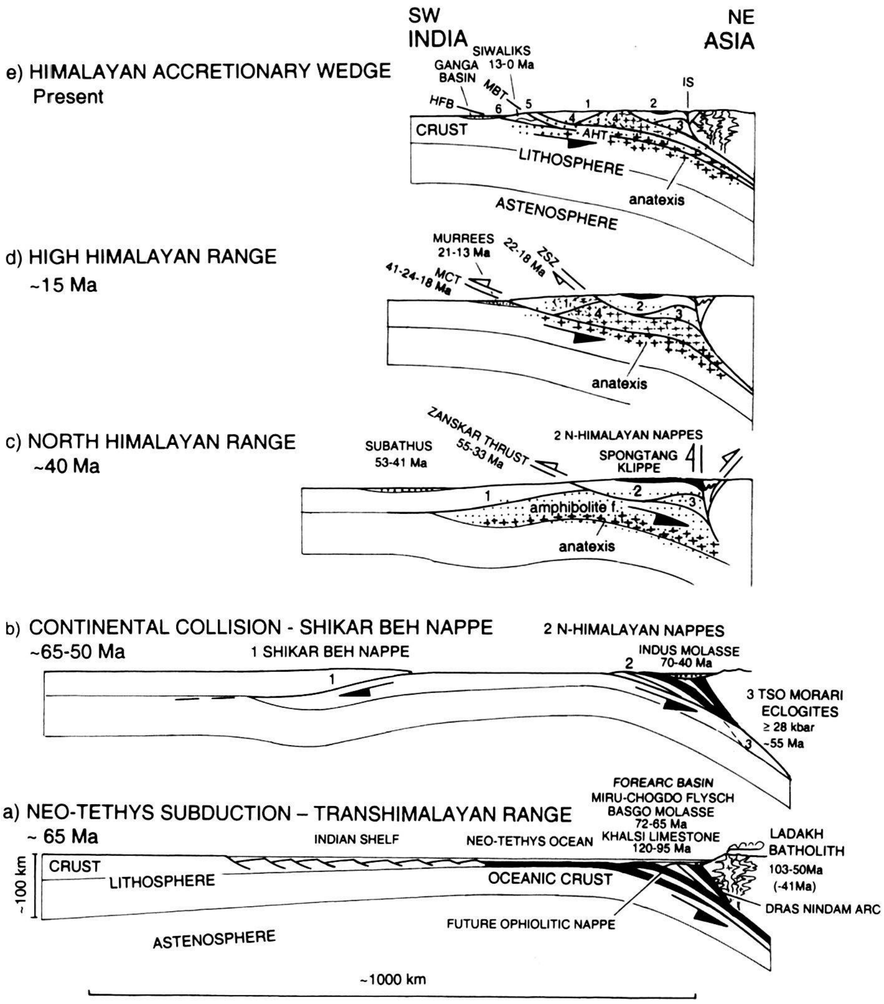

Fig. 15. Model for the kinematic evolution of the NW Himalaya. AHT = Active Himalayan thrust, HFB = Himalayan Frontal boundary, IS = Indus Suture, MBT = Main Boundary thrust, MCT = Main Central thrust, ZSZ = Zanskar shear zone.

on thermo-barometric data (Fig. 9) and the metamorphic map (Plate 4). Schlup et al. (2003) conclude, based on $^{40}Ar/^{39}Ar$ phengite and zircon and apatite FT ages, that the main uplift and erosion of the N Himalayan nappes occurred between 45-40 Ma and was of 2-5 mm/yr. The uplift rate in the High Himalaya was first very slow then increasing since about 21 and especially 11 Ma as recorded by the Indus-Ganga Molasse sediments (Fig. 12). Precise uplift rates for the High and Lesser Himalayan phases are difficult to estimate. The tectonic evolution of the Himalayan range is summarised on Fig. 15 and Tab. 1.

a) Section a) on Fig. 15 illustrates the position of India and Asia before the final disappearance of the Neo-Tethys oceanic lithosphere along a NE-dipping subduction zone. An immature island arc, the future ophiolitic nappe (Spongtang klippe), was accreted to the Asian margin (Reuber et al. 1987, Mahéo et al. 2000), the latter composed of the Dras-Nindam accretionary wedge and the younger 103–50 Ma Ladakh batholith (Transhimalaya batholith) (Weinberg & Dunlap 2000). The Aptian-Vraconian Kalsi limestone was deposited in the forearc basin above the accretionary prism, and was overlain by the marine Maastrichtian Basgo Molasse (Garzanti & Van Haver 1988). The Late Cretaceous-Early Eocene marine Miru-Chogdo flysch is the youngest sediment of the forearc basin (Fuchs & Linner 1995).
b) The intracontinental NE-directed Shikar Beh nappe (1) developed at an early time of continental collision, probably by reactivation of an older SW-dipping listric normal fault in the N-Indian margin. The precise age of this nappe is unknown: it is post Liassic and older than the Eocene frontal thrust of the N-Himalayan nappe stack (Steck et al. 1993, 1998, 1999, Epard et al. 1995, Wyss et al. 1999, Robyr et al. 2002, Robyr, 2003). The units of the N-Himalayan nappe stack (2) were successively detached from the upper part of the underthrust N-Indian crust and accreted to the Asian margin. The obducted ophiolitic Spongtang klippe forms

Tab. 1. Chronology of structural, metamorphic, magmatic and sedimentation phases of the NW Himalaya.

```
Ma:  70      65      60      55      50      45      40      35      30      25      20      15      10      5      0
CRETACEOUS  I  PALEOCENE  I          EOCENE                I  OLIGOCENE  I           MIOCENE           I PLIOC.I

103 Ma +++++++++++++++++++++++           +          Ladakh intrusives             ------- = formation/deposition
?~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~I SW-directed Spongtang ophiolite obduction     ++++ = intrusion
  ?----------------- - - - - - - -? Miru–Chogdo Flysch                             ~~~~ = deformation
   ?-----------?                    Basgo Molasse                                  x x x = metamorphism
Indus Molasse:          ?------------ - - - - - - - - - - - - -x x x x x - - - - - - - - - - - - - - - - ?   M1, M2, M3, M4, M5 =
                        ? ~~~~xx M1xx?       NE-directed Shikar Beh nappe          phases of metamorphism

Tso Morari Nappe:            x M2 x~~x M3 xx
                             ?~~~~~~~~x M3 xx~x~x~x~x~x~x~x~?        SW-directed North Himalayan Nappes
Subathu Formation:            I--------------------I  mid-Eo-Oligocene regression  -I
Tso Morari dome and NE-verging backfolds:        ~~~~ ~~~~~~ ~   ~   ~   ~   ~   ~   ~   ~   ~   ~
SW-directed High Himalayan Nappe and MCT:       ?~x~x~x~x~x~x M4 x~x~x~x~x~   ~   ~  ~  ~   ~~
                              Zanskar Shear Zone and Haptal-Gianbul dome:  I~x M5 x~ ~ ~ ~ ~ ~ ~ ~ ~
                                           Gumburanjun leucogranite:     ++
Murree Formation:                                   --  --  --?  --  --  --  --  I-------------I
Siwalik Formation:                                                                   I------------------------
Main Boundary thrust:                                                                       ?~~~~~~~~~~~
Kishtwar and Larji-Kullu-Rampur domes:                                                          ~~~~~~~~~
Subhimalayan thrust (Active Himalayan thrust):                                                       ?~~

CRETACEOUS  I  PALEOCENE  I          EOCENE                I  OLIGOCENE  I           MIOCENE           I PLIOC.I
Ma:  70      65      60      55      50      45      40      35      30      25      20      15      10      5      0
```

the earliest and highest nappe. At 55 Ma, the Indian continental crust (3) was subducted to a depth of over 90 km and recrystallised under eclogite facies conditions with crystallisation of coesite (De Sigoyer et al. 1997, 2000; ≥ 28 kbar and 700-800°C, Mukherjee & Sachan 2001). The intracontinental or epi-sutural Indus Molasse was deposited between ~70-40 Ma in a longitudinal basin above the Indus Suture and the SE border of the Ladakh batholith (Mascle et al. 1986, Searle, 1986, Garzanti & Van Haver, 1988). Most igneous clasts were derived from the northern Ladakh batholith and radiolarite, basalt pebbles and chrome spinel from the obducted oceanic crust to the S.

c) Before 48–45 Ma, the eclogitic Tso Morari nappe (3), predominantly composed of a low density Ordovician granite gneiss cross-cut by eclogitised basic dykes, was detached from the Indian crust and migrated, by buoyancy forces, along the surface of under thrusting crust to be emplaced at a depth of ~ 30 km below the higher Tetraogal and Mata nappes of the N-Himalayan nappe stack. Extrusion was driven by a mechanism similar to that proposed by Chemenda et al. (1995). Yet the extruding Tso Morari unit was not rigid, but has suffered a strong pure and simple shear deformation. The mylonitic fabric suggests that the Tso Morari slab was progressively squeezed out and moved by buoyancy force as illustrated on Fig. 8. Between 48-45 Ma, the whole N-Himalayan nappe stack, composed of the Tso Morari, Tetraogal and Mata nappes, was overprinted by a Barrovian regional metamorphism, reaching amphibolite facies conditions in the western Tso Morari dome, at a pressure of 9 ± 3 kbar and a temperature of 610 ± 70°C (De Sigoyer et al. 1997, Girard 2001). Between 45-41 Ma and during the formation and up warping of the Tso Morari dome, the amphibolite facies rocks cooled below 300°C to the north of Tso Morari and, at 30 Ma, 30 km farther to the W in the Tso Morari dome (^{40}Ar-^{39}Ar Mica and zircon fission track ages, De Sigoyer et al. 2000, Schlup et al. 2003). During the same period, the frontal thrust of the N-Himalayan nappe stack, exposed in the NW Zanskar crystalline zone, was still active with synkinematic crystallisation of garnets at ~33 Ma (Vance & Mahar 1998, Vance & Harris 1999). The NE-directed backfolds and thrusts of the N-Himalayan range and of the early Eocene Indus Molasse produced a 40-35 Ma old anchizonal metamorphism with prehnite-pumpellyite assemblages in the Indus Molasse (Baud et al. 1982, Van Haver et al. 1986, Steck et al. 1993). The N-verging Wakha and Dras recumbent folds near Kargil are probably formed during this phase (Gilbert & Merle 1987, Gapais et al. 1992). The N-Himalayan nappe was emplaced and exhumed carrying an ophiolitic nappe at its top. This event is recorded by ophiolitic clasts and chrome spinels in the latest Paleocene-Middle Eocene Subathu sandstone and limestone deposits of the southern foredeep basin (or back bulge basin, DeCelles et al. 1998 a, b) of the Eocene N-Himalayan range (DeCelles et al. 1998 a & b, Najman & Garzanti 2000). Only the frontal part of the Subathu foredeep basin is preserved below the Miocene Siwaliks and in some erosion relict on the Lesser Himalayan units. Farther to the NE, there is no evidence for the Subathu foredeep due to non-deposition or erosion of the High Himalayan range (Fig. 4). About 41-33 Ma ago, the formation of the N-Himalayan range was completed (Vance & Harris, 1999, Schlup et al. 2003). It is characterised by its frontal thrust to the SW and the Nyimaling-Tso Morari dome and NE-verging molasse backfolds and thrusts in front of the Asian backstop to the NE. The shortening of the detached upper crustal sediments of Upper Proterozoic to Ypresian age of the N-Himalayan nappes is about 100 km (>89 km, Steck et al. 1993). The average elevation of the N-Himalayan range at this time is difficult to estimate, it was not significant, perhaps of the order of 2000-3000 m elevation.

d) The Subathu marine transgression in the Himalayan foredeep and the Early Eocene epi-suture Indus Molasse sedimentation were followed by a Middle Eocene-Oligocene regression and general up-warping of the Indian crust (Le Fort, 1996, DeCelles et al. 1998). This regression was probably related to an early phase of underthrusting on the Main Central thrust. The late Eocene-Oligocene time interval between the Subathu and Murree-Siwalik sedimentation in the Himalayan foredeep and the Oligocene period of non deposition between the Eocene North Himalayan phase and the late Eocene – Miocene High Himalayan phase coincide with a period of slow convergence rate between the Indian and Asian plates (2.5–4 cm/yr between 36 and 30 Ma, Patriat & Achache 1984). The creation of the High Himalaya nappe started with the late Eocene-Oligocene deep seated intracrustal detachment and underthrusting of the Indian plate below the N Himalayan accretionary wedge (MCT), followed by the Miocene extrusion of the High Himalaya nappe, which is dated at 22–18 Ma (Frank et al. 1977, Hubbard & Harrison 1989, Harrison et al. 1992, Macfarlane 1993, Hodges et al. 1996, Schlup et al. 2003). It is suggested that the position of the intracrustal MCT was determined by a zone of high-grade amphibolite ductile crustal rocks, characterised by the dry partial-melting reaction: muscovite + plagioclase + quartz = K-feldspar + sillimanite + biotite + liquid. This zone was situated at the base of the orogenic lid formed by the pre-existing N-Himalayan and Shikar Beh nappe stacks. This migmatite zone is preserved and exposed in the Zanskar Crystalline, where the temperature increase was buffered by the endogene dry muscovite + plagioclase melting reaction (Honegger et al., 1982, Patiño Douce & Harris 1998, Robyr et al. 2002). The detachment on the deep crustal MCT and the formation of the High Himalayan nappe was predetermined by the existence of a deep seated high-grade amphibolite facies metamorphism and muscovite dry melting metamorphic zone (migmatite zone) in the underthrusted Indian crust below the pre-existing N-Himalayan accretionary wedge, that produced a very ductile crust. In a later phase of SW-directed thrusting of the "Crystalline nappe" on the MCT, a zone of extension, the 22-19 Ma old extensional Zanskar shear zone, developed by reactivation of the frontal thrusts of the N-Himalayan nappes in the roof of the nappe. The extruded ductile High Himalayan nappe is characterised by general shear (pure and simple shear) deformational structures (Vannay & Grasemann 2001). Second generation leucogranitic magmas were formed during this tectonic decompression, also by a dry muscovite + plagioclase melting. One such leucogranite, the Gumburanjun granite, intrudes and dates the Zanskar shear zone (ZSZ) at 22 Ma (monazite U-Pb age, Dèzes et al. 1999, Robyr et al. 2002). In conclusion, it is the thrusting, folding and extrusion of older, still hot metamorphic rocks, that created the inverted metamorphism of the early Miocene High Himalayan nappe (Epard et al. 1995, Steck et al. 1999, Wyss et al. 1999, Robyr et al. 2002). The displacement on the MCT was greater than 100 km, probably some hundreds of kilometres, whereas the extension on the ZSZ was estimated by Dèzes et al. (1999) to be only 35 km. The successive exhumation and erosion of medium- and in a later phase, of high-grade amphibolite facies rocks, is recorded in the heavy minerals deposited in the subsiding Indus-Ganga Molasse foreland basin, with detrital lower amphibolite facies garnets and staurolite in the early Miocene Murree Formation and kyanite and sillimanite in the Late Miocene to Pleistocene Siwalik Formation (Fig. 12; Chaudri 1972, Parkash et al. 1980, DeCelles et al. 1998a & b, Najman & Garzanti, 2000). The SW shore of the Subhimalayan foreland basin molasse migrated southward with the flexuration of the Indian plate below the Himalayan deformational front. The subsidence and rise of the average accumulation rate of the sediments in the Subhimalayan foredeep basin of the Indian Subcontinent and in the surrounding Asian basins increased abruptly at ~21 Ma, with the activation of the MCT and the extrusion of the High Himalayan nappe and the formation of the High Himalayan range (Métivier et al. 1999). The subsidence rate of 0.24 mm/a between 21 –18 Ma increased to 0.83 mm/a between 13-11 Ma.

e) The next structurally lower intracrustal thrust is the Main boundary thrust (MBT), active since about ~11 Ma (Meigs et al. 1995), that is composed of older rocks of early Proterozoic to Cambrian age (5), with an estimated displacement of over 100 km. The increase of sedimentation in the subhimalayan Indus-Ganga basin that started some 11 Ma ago, with a subsidence rate increasing from 0.3–0.6 mm/a (Fig. 12), may be related to the main uplift of the Himalayan range to its actual hight of over 6000 m and the installation of the Monsun rain climate. In situ Th-Pb ion microprobe dating of monazite indicates that the MCT and MBT remain active (Harrison et al. 1998, Catlos et al. 2002). Large recent sedimentary plains of the Beas and Sutlej rivers, between the frontal Himalayan foothills, indicate to an ongoing subsidence of the Foothills, in front of the still active MBT and internal Subhimalayan Molasse thrusts. The present Himalayan accretionary wedge is limited at its base by the Active Himalayan thrust (AHT) and to the N by the Ladakh batholith and Asian mantle wedge backstop. The nappe structures of the Subhimalayan Indus-Ganga Molasse sediments (6), with an active detachment at the base of the Eocene Subathus and the Himalayan Foothill boundary (HFB), the imbricated frontal thrust, with the recent Indus-Ganga alluvial deposits, and earthquake epicentres, at a depth of 10-15 km below the MCT, all indicate an Active Himalayan thrust (AHT) at the base of the present Himalayan accretionary wedge. The cross section through the Subhimalayan belt suggests that this detachment was mainly active in the Quaternary after the last Siwalik deposits since 2.1-1 Ma ago (Fig. 12). The average accumulation rate of the sediments in the basins surrounding the Himalayan range increased exponentially after the Pliocene, testifying to the present day increased tectonic activity, uplift and erosion of the Himalayan range (Métivier et al. 1999).

Estimates of the post-collisional shortening in the Himalaya

Assuming a continental collision at ~50 Ma ago and an average convergence velocity of 5 cm/yr, the shortening between India and Asia at the longitude of this study was about 2500 km (Patriat & Achache 1984). Dewey et al. (1989) suggest 1'800 km of shortening for the western corner of the Himalaya in Pakistan and 2750 km in Assam. According to Tapponnier et al. (1986), about 1'000-1'500 km of shortening might have been absorbed by lateral extrusion in the Asian plate and about the same amount by subduction and thickening of the Himalayan range and the Tibetan plateau. Based on flexural modelling of the elastic Indian lithosphere, Lyon-Caen & Molnar (1985) and Molnar (1990) estimated the convergence rate between India and southern Tibet to be 18 ± 7 km/Ma, thus for a time interval of 50 Ma, 900 ± 350 km of shortening is calculated. Palinspastic reconstructions of some Himalayan structures give only partial information on the crustal shortening. Balanced cross sections indicate about 33 km of shortening for the Subhimalaya (DeCelles et al. 1998, Powers et al. 1998). Steck et al. (1993) determined, from a balanced cross section and a simple shear model, approximately 100 km (>87km) of shortening for the N-Himalayan Nyimaling-Tsarap nappe and >12 km for the Indus Molasse. A similar value of shortening of 85 km for the N Himalayan Zanskar nappes was estimated by Corfield & Searle (2000). In our kinematic model (Fig. 8) during the period of continental collision between ~56 and 41 Ma, an approximate shortening of 450 km between India and the Asian backstop is estimated. Note that our model is based on supposed convergence velocities that are in reality unknown. A minimum displacement of the High Himalayan nappe of 100 km is given by the distance between the internal border of the tectonic Kishtwar and Larji-Kullu-Rampur windows and the frontal MCT thrust outcrop, and the same value is measured for the lower Crystalline nappe (Plate 1, Guntli 1993, Frank et al. 1995). Meigs et al. (1995) estimate about 100 km of displacement on the MBT. In our empirical model for the High Himalayan nappe formation (Fig. 11), a shortening of over 220 km for the period of 41-18 Ma related to the MCT is suggested. The resulting average convergence velocity is ~1 cm/a. This approximation is based on the supposed, but unknown, thrust length of 220 km. Hauck et al. (1998) estimated a minimum displacement of 200 km on the MCT in Eastern Nepal. This result is based on an area equilibrated palinspastic reconstruction. But the authors conclude that a complete restoration of the crust between the Ganga Basin and the Yarlung-Zangbo suture is presently not possible, for two reasons: (1) because the actual displacements along the MCT and STD are unknown and (2) because the degree of internal deformation within the Greater Himalayan allochthon is unknown and probably large. We concur with Hauck et al. (1998) and Hodges (2000) that an estimate of the shortening between India and the Asian backstop during the Himalayan nappe formation is actually not possible.

Doming and NE-verging backfold structures

In the NW Himalaya, doming and backfold structures were developed in response to crustal thickening as NE-verging late conjugate structures of the SW-directed nappes:

1) The 45–40 Ma NE-verging Tso Morari dome of the 55-33 Ma SW-directed N-Himalayan nappe stack, with a distance of ~110 km between the south-western thrust front and the backstop at the Indus suture and the 103-50 Ma old Ladakh batholith,
2) the 22–18 Ma Zanskar crystalline dome and the 24-18 Ma frontal MCT of the Crystalline nappe at a distance of ~115 km and
3) the active uplift of the Kishtwar and Larji-Kullu-Rampur dome between 7 Ma and the present and the Himalayan foothill boundary thrust of the Subhimalaya at a distance of ~120 km.

The similarity of the exposed structures of the three nappe stacks suggests a similar deep crustal geometry of the accretionary wedges as well as a similar mechanism of backfold formation. The mechanism of simultaneous forward and retroshear movements has been studied both in sandbox experiments (Huiqi et al. 1992, Malavielle et al. 1993, Larroque et al. 1995), and with numerical models (Beaumont et al. 1994, 1996). The sandbox experiments, both by Huiqi et al. (1992) and Larroque et al. (1995), show, in the case with low basal friction, that thrusts, folds or thrust faults and conjugate back folds are developed practically simultaneously. The Janauri anticline of the Subhimalayan thin-skinned thrust (Fig. 12 and Plate 2, profile 6 and 7) is an example of this type of structure. In contrast the models with high basal friction show that the conjugate back folds develop after a certain amount of thrusting and crustal thickening. The location of the back folds is forced by the position of the backstop. A good example of this kind of structure is the North Himalayan accretionary wedge with the Nyimaling-Tso Morari dome and the back folds and faults of the Indus Molasse, situated in front of the backstop formed by the Ladakh batholith and the Asian mantle wedge. The situation differs in the case of the High Himalayan nappe with the Haptal-Gianbul dome structure at the frontal edge of the N-Himalayan nappes and is very different in the case of the Himalayan frontal thrusts with the probably synchronous Kishtwar and Larji-Kullu-Rampur domes in the middle of the High Himalaya. The models of Beaumont et al. (1994, 1996) and Escher & Beaumont (1997) suggest that the initiation of up-warping at the backstop may be controlled by reduction in the convergent material that can be accommodated by the subduction channel. During the late orogenic and active Himalayan phase of dextral transpression between India and Asia, the dome structures continue to be uplifted and are often limited by younger normal faults and flexures such as the N-stricking Tso Morari fault and the Tso Kar and Yurdi flexures. The latter is together with the WNW-striking Sanku flexure responsible for the active uplift of the Suru syntaxis (Suru dome).

Conclusion

In conclusion, the Himalayan range was built up during the convergence of the Indian and Asian plates by a typical succession of orogenic phases (Mascle 1985, Le Fort 1996, Hodges 2000), where the preceding phase influences the next younger one. The main phases of the NW Himalaya are enumerated in the following list:

1) The Late Cretaceous and Paleocene ***Transhimalayan batholith phase*** (protohimalayan phase, Hodges 2000), characterised by the 103-50 Ma Andean type Ladakh magmatism, the accretion of the Dras-Nindam arc, the accretion and later obduction of the Spongtang immature island arc forming the southern active border of Asia and forearc sediment deposition. The Transhimalayan batholith, together with the Asian mantle wedge, form the Asian backstop for the Himalayan range.
2) The ***Shikar Beh phase***: the intra continental NE-verging Shikar Beh range of an unknown, probably late Paleocene age,
3) The Eocene ***North Himalayan phase*** (eohimalayan phase, Hodges 2000) creating the SW-directed North Himalayan accretionary wedge.
4) The late Eocene-Miocene ***High Himalayan phase*** (neohimalayan phase, Hodges 2000): The zone of dry intra-crustal melting below the North Himalayan range and the Shikar Beh nappe stack determined the future position of the Main Central thrust at the base of the High Himalayan or "Crystalline" nappe.
5) The late Miocene to present ***Lesser Himalayan phase***, with the formation of the deep-seated intracrustal Main boundary thrust.
6) The active ***Subhimalayan phase*** with the Subhimalayan thrust in front and the Active Himalayan thrust at the base of the present Himalayan accretionary wedge.

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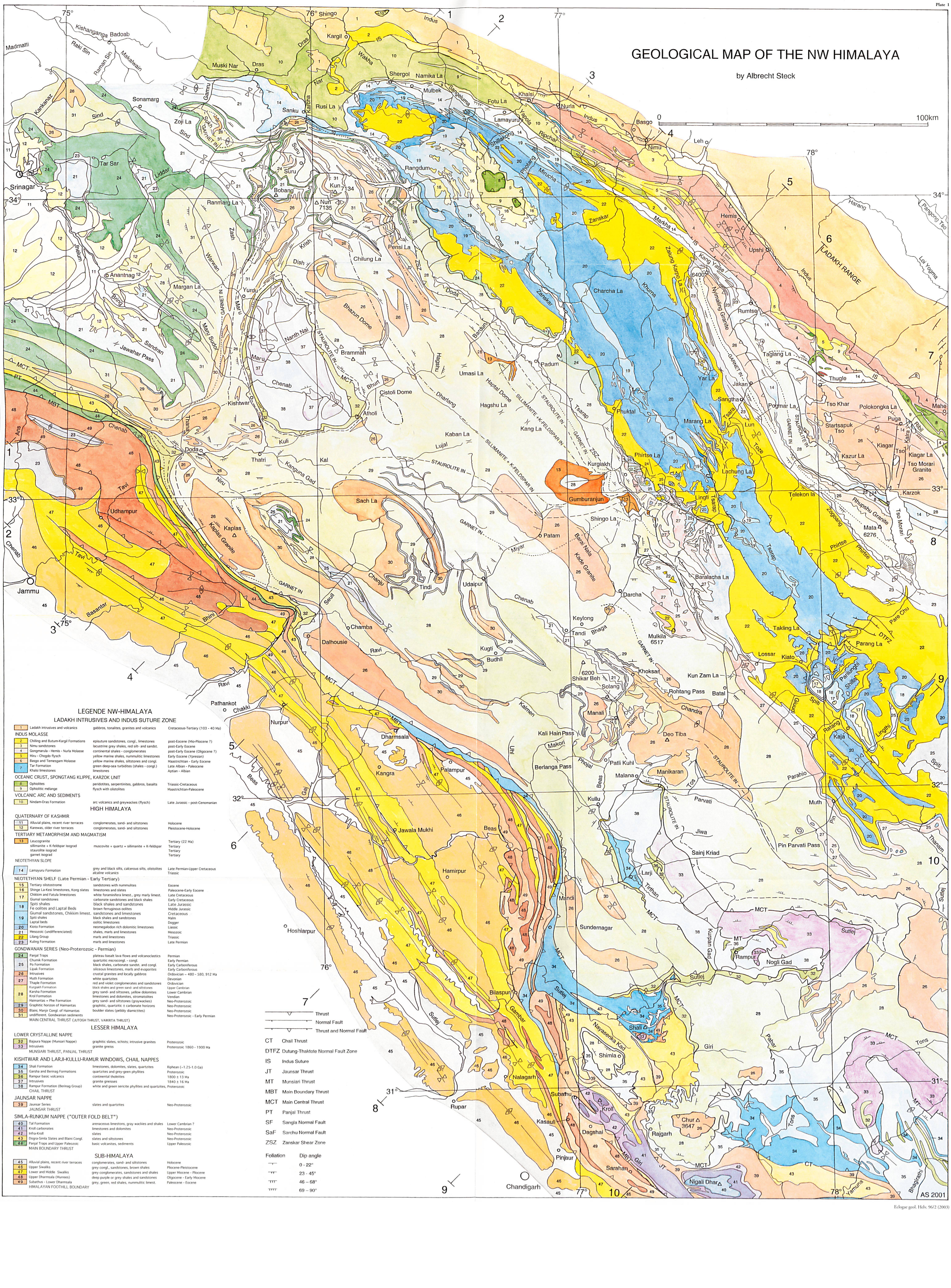
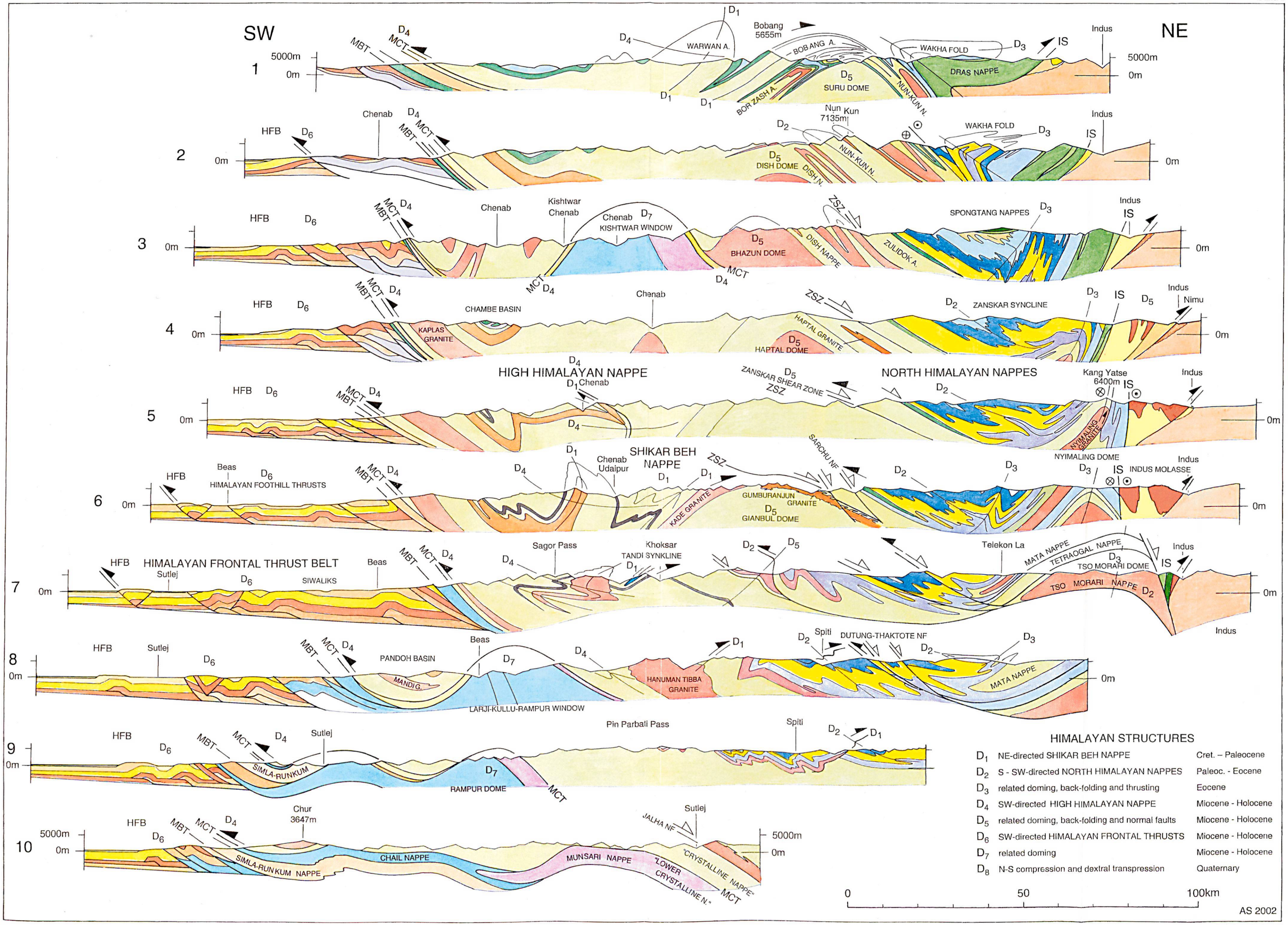
Plate 1: Geological map of the NW Indian Himalaya

Plate 2: Geological sections through the NW Indian Himalaya

Plate 3: Tectonic map of the NW Indian Himalaya

Plate 4: Metamorphic map of the NW Indian Himalaya


GEOLOGICAL MAP OF THE NW HIMALAYA
by Albrecht Steck
0
100km
LEGENDE NW-HIMALAYA
LADAKH INTRUSIVES AND INDUS SUTURE ZONE
1 Ladakh intrusives and volcanics | gabbros, tonalites, granites and volcanics | Cretaceous-Tertiary (103 - 40 Ma)
INDUS MOLASSE
2 Chilling and Butum-Kargil Formations | episuture sandstones, congl., limestones | post-Eocene (Mio-Pliocene ?)
3 Nimu sandstones | lacustrine grey shales, red silt- and sandst. | post-Early Eocene
4 Gongmarula - Hemis - Nurla Molasse | continental shales - conglomerates | post-Early Eocene (Oligocene ?)
5 Miru - Chogdo flysch | yellow marine shales, nummulitic limestones | Early Eocene (Ypresian)
6 Basgo and Temesgam Molasse | yellow marine shales, siltstones and congl. | Maastrichtian - Early Eocene
7 Tar Formation | green deep-sea turbidites (shales - congl.) | Late Albian - Paleocene
Khalsi limestones | limestones | Aptian - Albian
OCEANIC CRUST, SPONGTANG KLIPPE, KARZOK UNIT
8 Ophiolites | peridotites, serpentinites, gabbros, basalts | Triassic-Cretaceous
9 Ophiolitic mélange | flysch with olistolites | Maastrichtian-Paleocene
VOLCANIC ARC AND SEDIMENTS
10 Nindam-Dras Formation | arc volcanics and greywackes (flysch) | Late Jurassic - post-Cenomanian
HIGH HIMALAYA
QUATERNARY OF KASHMIR
11 Alluvial plains, recent river terraces | conglomerates, sand- and siltstones | Holocene
12 Karewas, older river terraces | conglomerates, sand- and siltstones | Pleistocene-Holocene
TERTIARY METAMORPHISM AND MAGMATISM
13 Leucogranite | Tertiary (22 Ma)
sillimanite + K-feldspar isograd | muscovite + quartz = sillimanite + K-feldspar | Tertiary
staurolite isograd | Tertiary
garnet isograd | Tertiary
NEOTETHYAN SLOPE
14 Lamayuru Formation | grey and black silts, calcerous silts, olistolites alcaline volcanics | Late Permian-Upper Cretaceous Triassic
NEOTETHYAN SHELF (Late Permian - Early Tertiary)
15 Tertiary olistostrome | sandstones with nummulites | Eocene
16 Shinge La-Kesi limestones, Kong slates | limestones and slates | Paleocene-Early Eocene
17 Chikkim and Fatula limestones | white foraminifera limest., grey marly limest. | Late Cretaceous
Giumal sandstones | carbonate sandstones and black shales | Early Cretaceous
18 Spiti shales | black shales and sandstones | Late Jurassic
Fe oolites and Laptal Beds | brown ferruginous oolites | Middle Jurassic
19 Giumal sandstones, Chikkim limest. | sandstones and limestones | Cretaceous
Spiti shales | black shales and sandstones | Malm
Laptal beds | oolitic limestones | Dogger
20 Kioto Formation | neomegalodon rich dolomitic limestones | Liassic
21 Mesozoic (undifferenciated) | shales, marls and limestones | Mesozoic
22 Lilang Group | marls and limestones | Triassic
23 Kuling Formation | marls and limestones | Late Permian
GONDWANAN SERIES (Neo-Proterozoic - Permian)
24 Panjal Traps | plateau basalt lava flows and volcanoclastics | Permian
25 Chumik Formation | quartzitic microcongl. - congl. | Early Permian
Po Formation | black shales, carbonate sandst. and congl. | Early Carboniferous
Lipak Formation | siliceous limestones, marls and evaporites | Early Carboniferous
26 Intrusives | crustal granites and locally gabbros | Ordovician ~ 480 - 580, 912 Ma
27 Muth Formation | white quartzites | Devonian
Thaple Formation | red and violet conglomerates and sandstones | Ordovician
28 Kurgiakh Formation | black shales and green sand- and siltstones | Upper Cambrian
Karsha Formation | grey sand- and siltsones, yellow dolomites | Lower Cambrian
Krol Formation | limestones and dolomites, stromatolites | Vendian
Haimantas = Phe Formation | grey sand- and siltstones (graywackes) | Neo-Proterozoic
29 Graphitic horizon of Haimantas | graphitic, quartzitic ± carbonate horizons | Neo-Proterozoic
30 Blaini, Manjir Congl. of Haimantas | boulder slates (pebbly diamictites) | Neo-Proterozoic
31 undifferent. Gondwanian sediments | Neo-Proterozoic - Early Permian
MAIN CENTRAL THRUST (JUTOGH THRUST, VAIKRITA THRUST)
LESSER HIMALAYA
LOWER CRYSTALLINE NAPPE
32 Bajaura Nappe (Munsari Nappe) | graphitic slates, schists; intrusive granites | Proterozoic
33 Intrusives | granite gneiss | Proterozoic 1860 - 1900 Ma
MUNSIARI THRUST, PANJAL THRUST
KISHTWAR AND LARJI-KULLU-RAMUR WINDOWS, CHAIL NAPPES
34 Shali Formation | limestones, dolomites, slates, quartzites | Riphean (~1.25-1.0 Ga)
35 Garsha and Berinag Formations | quartzites and grey-geen phyllites | Proterozoic
36 Rampur basic volcanics | continental tholeiites | 1800 ± 13 Ma
37 Intrusives | granite gneisses | 1840 ± 16 Ma
38 Rampur Formation (Berinag Group) | white and green sericite phyllites and quartzites, | Proterozoic
CHAIL THRUST
JAUNSAR NAPPE
39 Jaunsar Series | slates and quartzites | Neo-Proterozoic
JAUNSAR THRUST
SIMLA-RUNKUM NAPPE ("OUTER FOLD BELT")
40 Tal Formation | arenaceous limestons, gray wackies and shales | Lower Cambrian ?
41 Kroll carbonates | limestones and dolomites | Neo-Proterozoic
42 Infra-Kroll | slates | Neo-Proterozoic
43 Dogra-Simla Slates and Blaini Congl. | slates and siltstones | Neo-Proterozoic
44 Panjal Traps and Upper Paleozoic | basic volcanites, sediments | Upper Paleozoic
MAIN BOUNDARY THRUST
SUB-HIMALAYA
45 Alluvial plains, recent river terraces | conglomerates, sand- and siltstones | Holocene
46 Upper Siwaliks | grey congl., sandstones, brown shales | Pliocene-Pleistocene
47 Lower and Middle Siwaliks | grey conglomerates, sandstones and shales | Upper Miocene - Pliocene
48 Upper Dharmsala (Murrees) | deep purple or grey shales and sandstones | Oligocene - Early Miocene
49 Subathus - Lower Dharmsala | grey, green, red shales, nummulitic limest. | Paleocene - Eocene
HIMALAYAN FOOTHILL BOUNDARY
Thrust
Normal Fault
Thrust and Normal Fault
CT Chail Thrust
DTFZ Dutung-Thaktote Normal Fault Zone
IS Indus Suture
JT Jaunsar Thrust
MT Munsiari Thrust
MBT Main Boundary Thrust
MCT Main Central Thrust
PT Panjal Thrust
SF Sangla Normal Fault
SaF Sarchu Normal Fault
ZSZ Zanskar Shear Zone
Foliation Dip angle
0 - 22°
23 - 45°
46 – 68°
69 – 90°
AS 2001


SW
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Bobang
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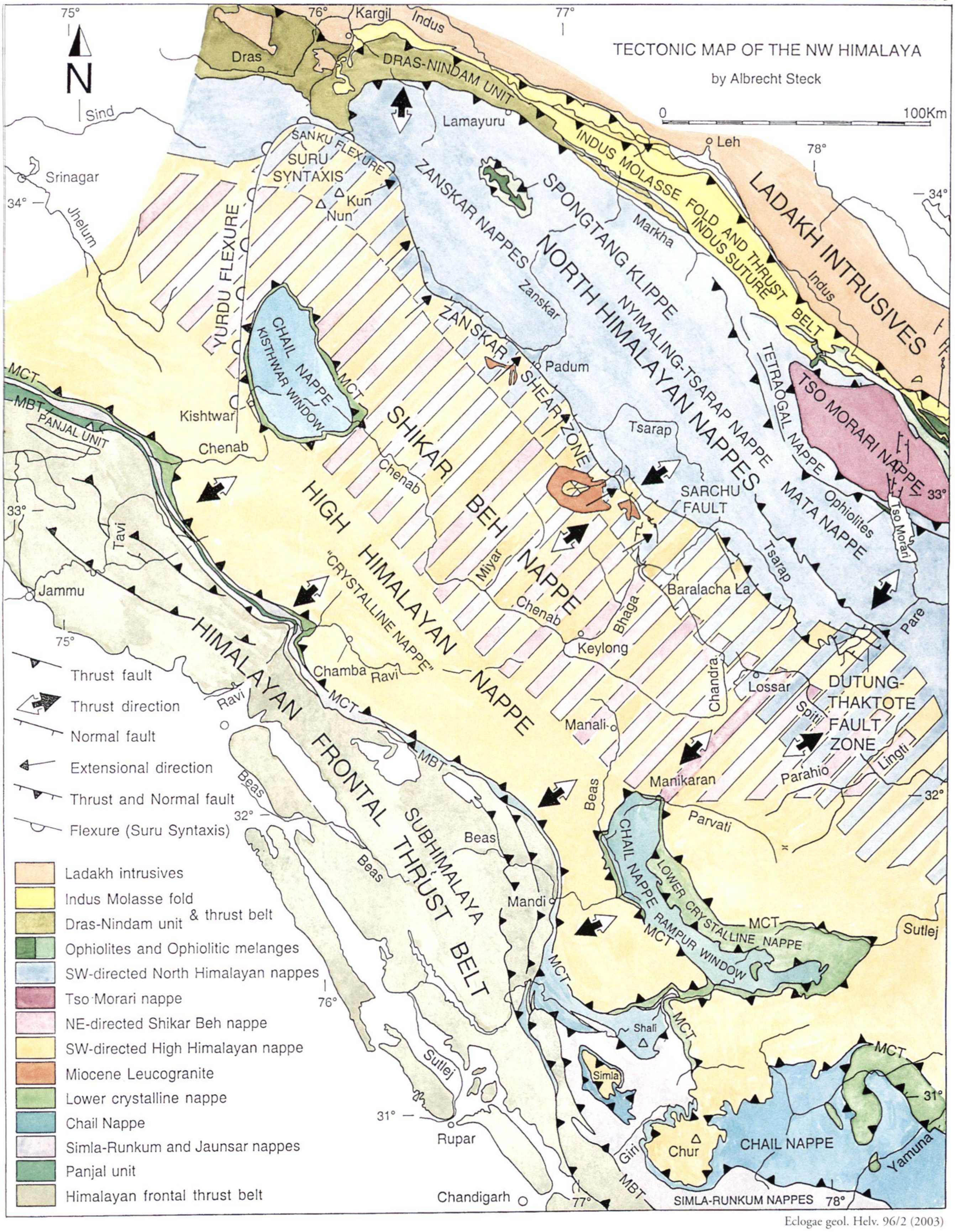
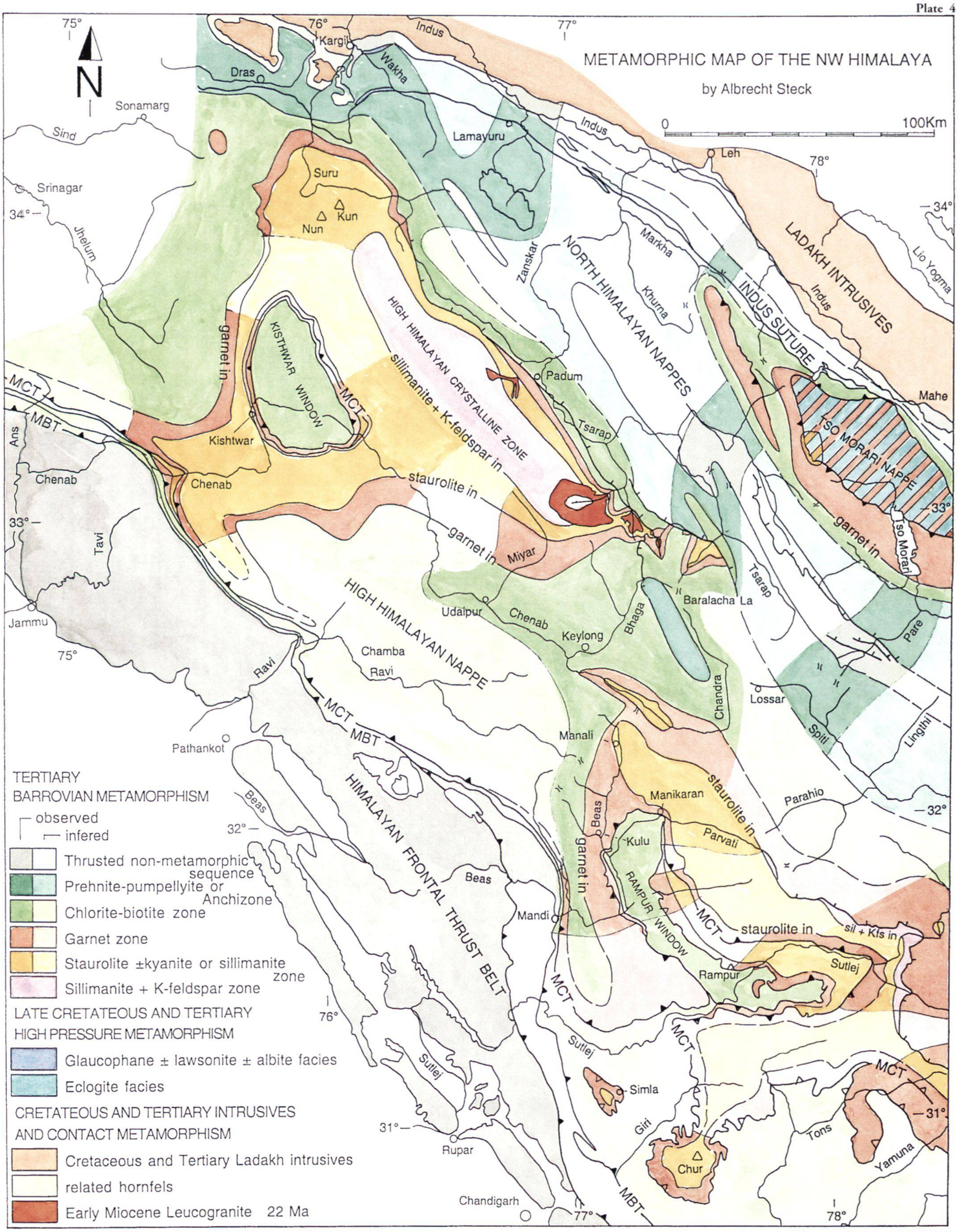
BOBANG A.
WAKHA FOLD
DRAS NAPPE
SURU DOME
BOR ZASH A.
NUN-KUN N.
Nun Kun
7135m
DISH DOME
DISH N.
Kishtwar
Chenab
KISHTWAR WINDOW
BHAZUN DOME
DISH NAPPE
ZULIDOK A.
ZSZ
SPONGTANG NAPPES
CHAMBE BASIN
KAPLAS GRANITE
ZANSKAR SYNCLINE
HAPTAL GRANITE
HAPTAL DOME
Nimu
HIGH HIMALAYAN NAPPE
NORTH HIMALAYAN NAPPES
ZANSKAR SHEAR ZONE
Kang Yatse
6400m
NYIMALING GRANITE
NYIMALING DOME
HIMALAYAN FOOTHILL THRUSTS
Chenab
Udaipur
SHIKAR BEH NAPPE
KADE GRANITE
SARCHU NF
GUMBURANJUN GRANITE
GIANBUL DOME
INDUS MOLASSE
HIMALAYAN FRONTAL THRUST BELT
SIWALIKS
Sagor Pass
Khoksar
TANDI SYNCLINE
Telekon La
MATA NAPPE
TETRAOGAL NAPPE
TSO MORARI DOME
TSO MORARI NAPPE
PANDOH BASIN
MANDI G.
LARJI-KULLU-RAMPUR WINDOW
HANUMAN TIBBA GRANITE
DUTUNG-THAKTOTE NF
Pin Parbati Pass
SIMLA-RUNKUM
RAMPUR DOME
Chur
3647m
JALHA NF
SIMLA-RUNKUM NAPPE
CHAIL NAPPE
MUNSARI NAPPE
"LOWER CRYSTALLINE N."
"CRYSTALLINE NAPPE"
HIMALAYAN STRUCTURES
D_1 NE-directed SHIKAR BEH NAPPE Cret. – Paleocene
D_2 S - SW-directed NORTH HIMALAYAN NAPPES Paleoc. - Eocene
D_3 related doming, back-folding and thrusting Eocene
D_4 SW-directed HIGH HIMALAYAN NAPPE Miocene - Holocene
D_5 related doming, back-folding and normal faults Miocene - Holocene
D_6 SW-directed HIMALAYAN FRONTAL THRUSTS Miocene - Holocene
D_7 related doming Miocene - Holocene
D_8 N-S compression and dextral transpression Quaternary
0
50
100km
AS 2002


TECTONIC MAP OF THE NW HIMALAYA
by Albrecht Steck
0
100Km
N
75°
76°
77°
78°
34°
33°
32°
31°
Kargil
Indus
Dras
DRAS-NINDAM UNIT
Lamayuru
Leh
Sind
Srinagar
Jhelum
SANKU FLEXURE
SURU SYNTAXIS
Kun
Nun
YURDU FLEXURE
ZANSKAR NAPPES
SPONGTANG KLIPPE
INDUS MOLASSE FOLD AND THRUST BELT
INDUS SUTURE
LADAKH INTRUSIVES
Markha
Zanskar
NORTH HIMALAYAN NAPPES
NYIMALING-TSARAP NAPPE
CHAIL NAPPE
KISTHWAR WINDOW
MCT
ZANSKAR SHEAR ZONE
Padum
TETRAOGAL NAPPE
TSO MORARI NAPPE
MBT
PANJAL UNIT
Kishtwar
Chenab
SHIKAR BEH NAPPE
Tsarap
SARCHU FAULT
MATA NAPPE
Ophiolites
Tso Morari
HIGH HIMALAYAN NAPPE
"CRYSTALLINE NAPPE"
Tavi
Jammu
Miyar
Baralacha La
Bhaga
Keylong
Pare
HIMALAYAN FRONTAL THRUST BELT
Chamba
Ravi
Chandra
Lossar
DUTUNG-THAKTOTE FAULT ZONE
Spiti
Lingti
Parahio
Manali
Manikaran
Beas
Parvati
SUBHIMALAYA
Mandi
LOWER CRYSTALLINE NAPPE
RAMPUR WINDOW
Sutlej
Shali
Simla
Rupar
Chur
Giri
Yamuna
Chandigarh
SIMLA-RUNKUM NAPPES
Thrust fault
Thrust direction
Normal fault
Extensional direction
Thrust and Normal fault
Flexure (Suru Syntaxis)
Ladakh intrusives
Indus Molasse fold & thrust belt
Dras-Nindam unit
Ophiolites and Ophiolitic melanges
SW-directed North Himalayan nappes
Tso Morari nappe
NE-directed Shikar Beh nappe
SW-directed High Himalayan nappe
Miocene Leucogranite
Lower crystalline nappe
Chail Nappe
Simla-Runkum and Jaunsar nappes
Panjal unit
Himalayan frontal thrust belt


METAMORPHIC MAP OF THE NW HIMALAYA
by Albrecht Steck
0
100Km
N
75°
76°
77°
78°
34°
33°
32°
31°
Kargil
Dras
Indus
Wakha
Sonamarg
Sind
Srinagar
Jhelum
Lamayuru
Leh
Suru
Nun
Kun
Zanskar
Markha
Khurna
NORTH HIMALAYAN NAPPES
LADAKH INTRUSIVES
Lio Yogma
INDUS SUTURE
HIGH HIMALAYAN CRYSTALLINE ZONE
KISHTWAR WINDOW
garnet in
sillimanite + K-feldspar in
Padum
Tsarap
MCT
MBT
Ans
Kishtwar
Chenab
staurolite in
Mahe
TSO MORARI NAPPE
Tso Morari
Tavi
Miyar
Jammu
Udaipur
Keylong
Bhaga
Baralacha La
HIGH HIMALAYAN NAPPE
Chamba
Ravi
Chandra
Lossar
Pare
Spiti
Lingthi
Pathankot
Manali
Manikaran
Parvati
Parahio
Kulu
RAMPUR WINDOW
HIMALAYAN FRONTAL THRUST BELT
Mandi
sil + Kfs in
Rampur
Sutlej
Simla
Giri
Chur
Tons
Yamuna
Rupar
Chandigarh
TERTIARY
BARROVIAN METAMORPHISM
observed
infered
Thrusted non-metamorphic sequence
Prehnite-pumpellyite or Anchizone
Chlorite-biotite zone
Garnet zone
Staurolite ±kyanite or sillimanite zone
Sillimanite + K-feldspar zone
LATE CRETATEOUS AND TERTIARY
HIGH PRESSURE METAMORPHISM
Glaucophane ± lawsonite ± albite facies
Eclogite facies
CRETATEOUS AND TERTIARY INTRUSIVES
AND CONTACT METAMORPHISM
Cretaceous and Tertiary Ladakh intrusives
related hornfels
Early Miocene Leucogranite 22 Ma