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ginning 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-

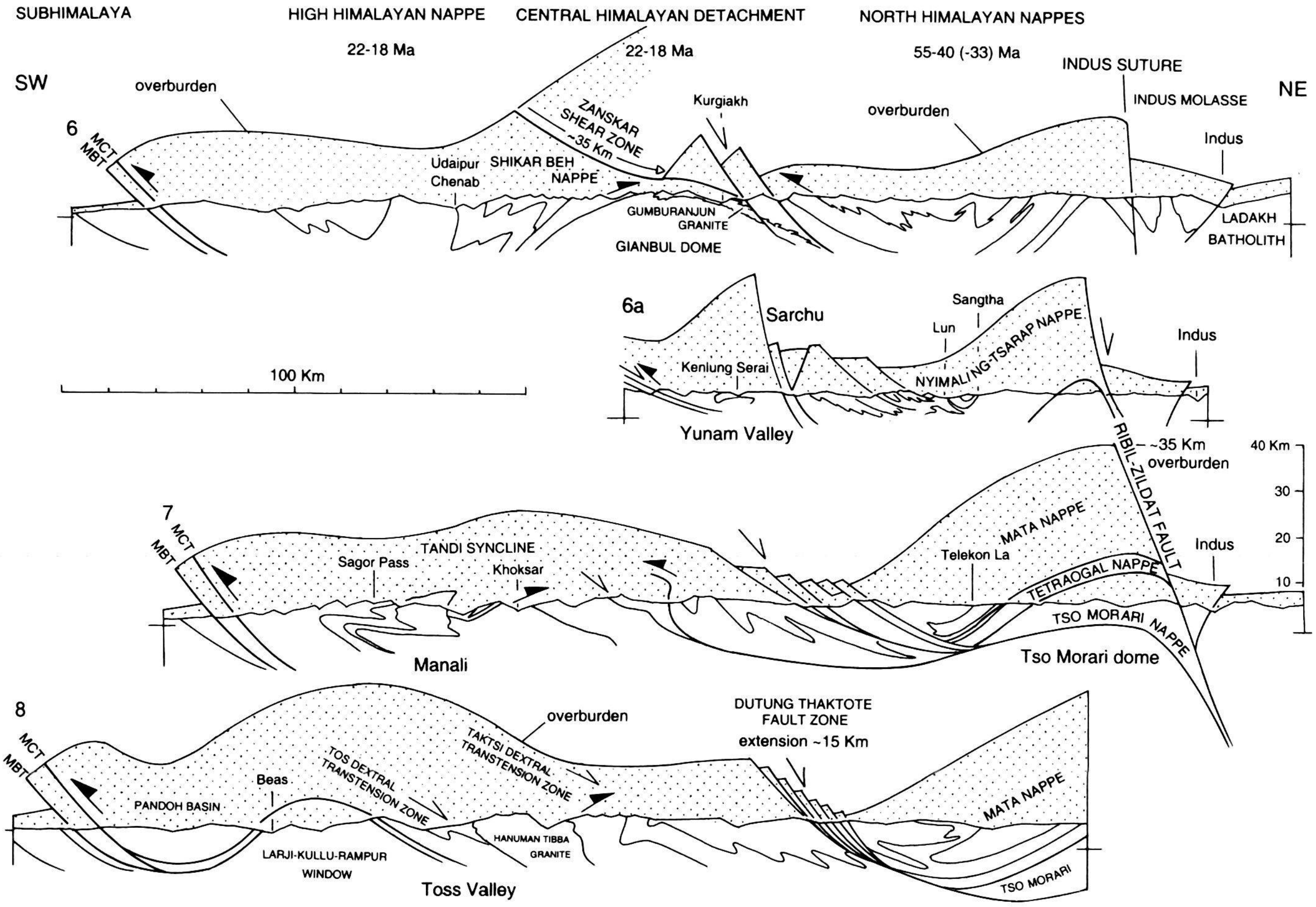

Fig. 14. Estimates of the eroded overburden of the Himalayan accretionary wedge based on thermo-barometric data from Dèzes et al. (1999), De Sigoyer et al. (2000), Epard et al. (1995), Frank et al. (1977b), Girard (2001), Girard et al. (1999, 2001), Robyr et al. (2002), Schlup et al. (2003), Spring et 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

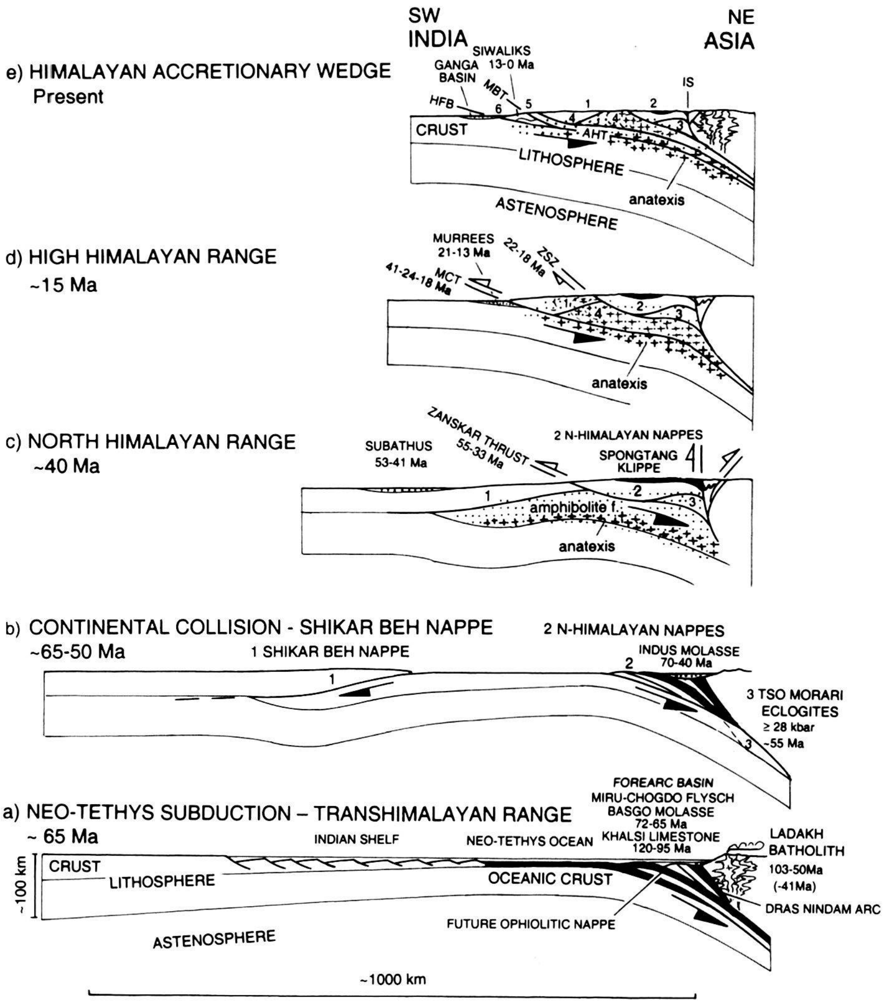

Fig. 15. Model for the kinematic evolution of the NW Himalaya. AHT = Active Himalayan thrust, HFB = Himalayan Frontal boundary, IS = Indus Suture, MBT = Main Boundary thrust, MCT = Main Central thrust, ZSZ = Zanskar shear zone.

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.

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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
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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 retro-shear 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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