# Landform evolution of the Makalu – Barun region in the East Nepal Himalaya

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## Abstract:

Geomorphological analysis of landform patterns in the Makalu – Barun region of the Nepal Himalaya, illustrates their relation to morphotectonic and climate-morphogenetic features of relief-building processes. High-mountain landforms in the Himalaya are the result of morphotectonic processes, as well as the denudation and erosional efficiency in different palaeoclimatic conditions during the late Cainozoic. Recent climate-driven morphogenetic processes in the extremely dissected relief with an elevational gradient of over 7 000 m are described in the framework of extraglacial and glacial zones, the periglacial zone and the seasonally cold/warm humid zone. The geomorphological observations on a decadal scale suggest that the frequency and magnitude of recent landform changes in the Makalu – Barun region are increasing from a very cold and dry extraglacial zone across a large periglacial area up to subtropical landscape with humid climatic conditions. Observations in the East Nepal Himalaya suggest significant feedbacks between the rate of tectonic exhumation of deep crystalline rocks and the intensity of climate-morphogenetic processes. Cooling history of rocks investigated by low-temperature thermochronology has confirmed rapid crustal uplift and erosion in the Makalu – Barun region during the late Cainozoic. The very high rate of valley incision also stimulates isostatic compensation, which is one of the factors influencing the uplift of the Makalu, Mount Everest and other extremely high massifs in the East Nepal Himalaya during the Quaternary. The observed recent landform changes confirm the high intensity of climate-driven morphogenetic processes, especially with very effective erosion and transport of weathered material in the periglacial and seasonally warm humid (monsoon) mountain zones. Dynamic changes of landscape pattern are important evidence of the present-day severe natural hazards.

**Key words:** landform evolution, glacial and periglacial processes, East Nepal Himalaya

## 1. Introduction

The Himalaya is regarded as the most perfectly developed collision orogene with distinctive features of the active lithospheric plate tectonics. Research on the geological structure and the landforms of the Himalaya has identified the extreme intensity of the geodynamic processes which, especially during the Quaternary, remodelled these mountains to their present-day shape (Iswata 1987; Kalvoda 1988, 1992; Fielding 1996 and others). The Makalu – Barun region is situated between the Khumbu Himal (Mount Everest 8 848 m a.s.l.) and the Arun valley (1 350 m), in the morphotectonically conspicuous zone of the High Himalaya nappes (Bordet 1961, Jaroš and Kalvoda 1978a, b). Observations of landform patterns of peculiar relief types in the Makalu – Barun region suggest extremely high rates of denudation, sediment transfer and deposition. The vertical hierarchy of variable high-mountain reliefs is striking, and ranges from the extremely cold arête ridges of the Makalu Massif (8 475 m), through the heavily glaciated and periglacial areas, to the seasonally cold/warm humid Lower Barun and Arun valleys (Figure 1). The aims of the presented paper focused on the landform evolution of the Makalu – Barun region in the East Nepal Himalaya are in: a) analysis of landform patterns and climato-morphogenetic processes, b) evaluations of exhumation and erosion of rocks related to morphotectonic processes during the late Cenozoic.

Geomorphological analysis of landform patterns in the Chomolongma and Makalu Massifs and the Barun Khola valley regions has been undertaken by Kalvoda (1978, 1979a, b, 1982, 1984a, 1992, 2003, 2007). The largely glaciated part of the East Nepal Himalaya is an area exceeding 400 km$^2$ comprising complicated systems of hanging, slope and valley glaciers which flow from the extremely high crests (Figure 2). The relief between the Chomolongma Massif and the wider surroundings of the Barun Khola valley forms a natural section across the sequence of rocks and the Himalayan landscape. The observation of landforms provides evidence of very dynamic landscape evolution in the Quaternary. The nature of the Makalu – Barun region has been studied since the middle of the 20$^{th}$ century mainly focused on a) geology and geophysics (e.g. Bordet 1961; Jaroš and Kalvoda 1976, 1978a, b; Palivcová et al. 1982; Schärer 1984; Brunel and Kienast 1986; Hubbard 1989; Hubbard and Harrison 1989; Hubbard et al. 1996; Yagi and Minaki 1991, Lombardo et al. 1993, Visona and Lombardo 2002; Svojtka et al. 2003), b) geomorphology (Kalvoda 1978, 1979a, b, 1982, 1984a, 1992, 2003, 2007; Kalvoda and Valenta 1997; Kalvoda et al. 2004; Kuhle 2004, 2005), c) biology and geoecology (Smolíková and Kalvoda 1981; Daniel et al. 1985; Byers 1996; Carpenter and Zomer 1996; Zomer et al. 2001, 2002). The challenge remains of determining the landform development in the Makalu – Barun region in which the geomorphological record of active orogeny is integrated with the investigation of landscape patterns and recent climate-morphogenetic changes of the Himalayan relief.

The dynamics of geomorphological processes in the vertical climato-morphogenetic zones of the Makalu – Barun region shows that glacial and periglacial processes are very effective at destroying the rock massif uplifted during collision orogeny. Rapid unroofing and exhumation of deeper parts of the rock massifs is reliant upon vigorous transport agencies, such as transgression of glaciers and intensive activity of wind in extraglacial and glacial zones, and/or rapid action of water in periglacial and seasonally humid cold/warm zones. The key factor is the long-term interaction between the intensity of morphostructural processes, including the extent of the tectonic exhumation of deep-crust rocks, and the changeable rates of denudation which also involve the outward flux of eroded material (Burbank et al. 1996, 2003; Zeitler et al. 2001; Bishop 2007; Kalvoda 2007; Kalvoda and Goudie 2007). The extremely high rate of valley incision also stimulates isostatic compensation (Wager 1937), which is one of the factors influencing the orogenic uplift of the Nepal Himalaya throughout the Quaternary.

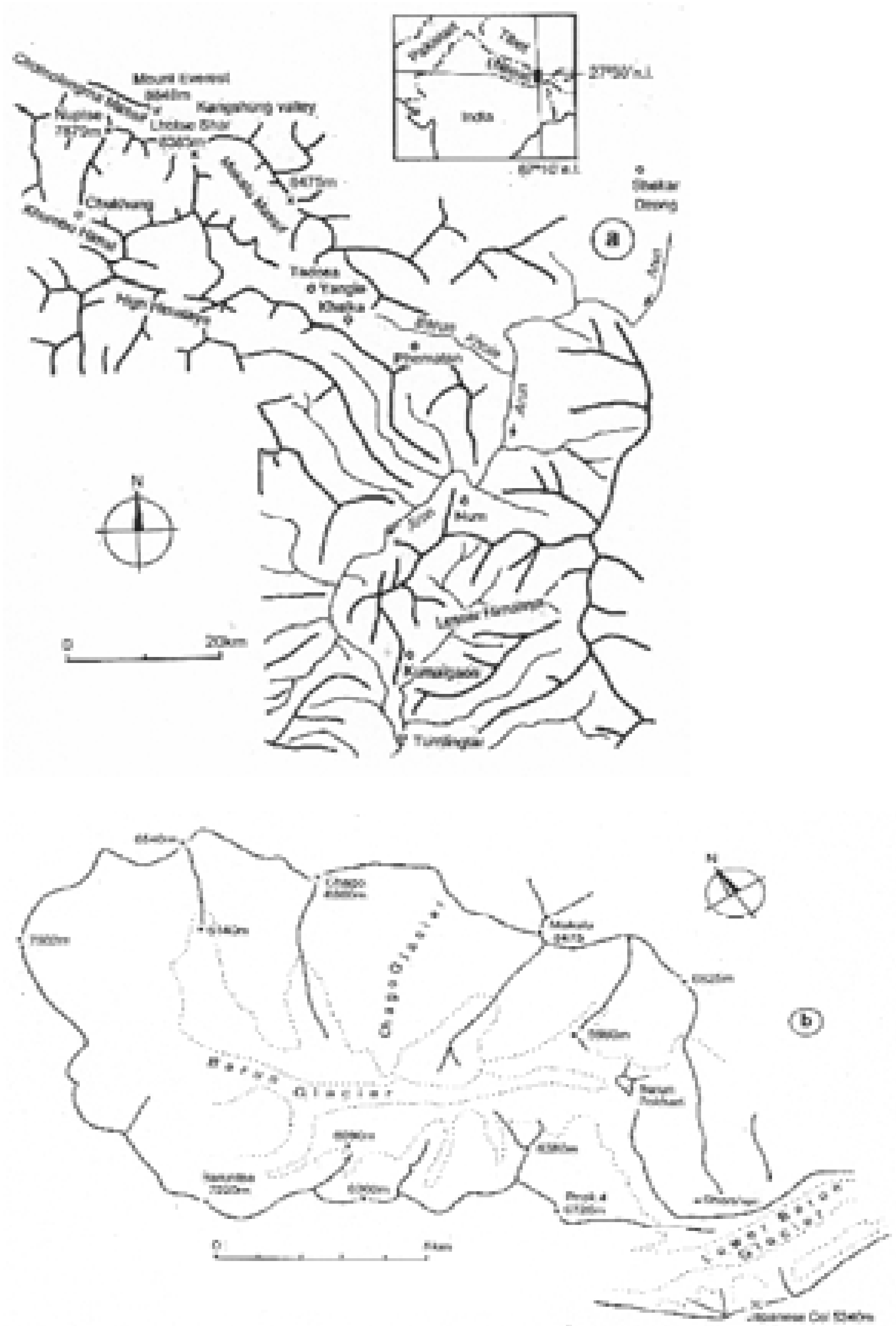



Figure 1: Geographical position of the Makalu – Barun region (a) and described geomorphological localities (b) in the East Nepal Himalaya.

## 2. Exhumation and erosion of rocks related to morphotectonic processes

Geomorphological analysis of landform patterns in the Makalu – Barun region of the East Nepal Himalaya provides evidence of the dynamics of exhumation and erosion of rocks during ongoing collision orogeny in the late Cenozoic. The relief of the uppermost cliffs of this part of the East Nepal Himalaya represents a thick rock succession comprising three main lithostratigraphical units (Jaroš and Kalvoda 1976, 1978a, b); the Chomolongma Group (Figure 3), the Makalu Formation and the Barun Group. In the Makalu – Barun region, the Chomolongma Group occurs only in the vicinity of the unnamed peak of 7 502 m elevation. Its lower part consists of pelitic, slightly metamorphosed Palaeozoic rocks. The Makalu Formation consists of a number of Miocene granite bodies associated with extensive contact injection zones intruded into the lower part of the Chomolongma Group and the upper part of the Barun Group. The later consists mostly of biotite gneisses, originally of Precambrian age (Bordet 1961). The complex of sillimanite-, kyanite- and cordierite-bearing gneisses, calc-silicates and amphibolites (Bordet 1961**;** Lombardo et al. 1993) is intruded by Namtche migmatitic orthogneisses (Brunel and Kienast 1986**;** Pognante and Benna 1993) and overlain by biotite-sillimanite gneisses ("black gneisses" of the Rongbuk Formation). At lower elevations, below the Shershon site (Figure 1) towards the Main Central Thrust, the Barun Group consists of an up to 5 000 m thick formation of banded biotite gneisses with garnet, kyanite and sillimanite with interlayered garnetiferous amphibolites, pyroxenite lenses and granulites.

The biotite-sillimanite gneisses in the upper part of the High Himalayan nappe were intruded by tourmaline leucogranites of the Makalu Massif during the late Oligocene and early Miocene (24.4 – 21.7 Ma; Schärer 1984). The bedrock topography on granites is developed especially on the ridges and the rock faces of the Massifs of Peak 4, Baruntse, Chago and Makalu (Figure 2). Yellowish-white granites form strikingly steep rock faces, often polished by wind and avalanches, occasionally ice-affected, which display extensive desquamation planes. Compact bedrock topography is developed on black paragneisses along the lower parts of the eastern cliffs of Peak 4, below elevation points 5 860 and 6 260 m and in the foreland of the southwestern wall of the Makalu.

An outstanding problem in the orogenic evolution of the High Himalaya is the timing of the thermal metamorphic peak and the widespread melting event that resulted in production of voluminous masses of leucogranitic magma. The emplacement of leucogranites into the nappes of the High Himalayan rocks is constrained to 24 – 17 Ma (Searle et al. 1999 and references therein) but the thermal peak of metamorphism has been dated with a large spread of ages. Prograde metamorphic ages reported from the eastern Himalaya overlap the emplacement ages of the leucogranites but there is > 10 Ma age delay between the metamorphic thermal peak and the onset of the leucogranitic magmatism reported from the central and western parts of the Himalayan orogen (Harrison et al. 1997a, b, 1999).

The metamorphism of the High Himalayan nappe in the Makalu and adjacent Mount Everest regions (Figure 1) is related to crustal thickening during the collision (T = 550 – 680°C, P = 8 – 10 kbar) followed by decompression and crustal melting at P = 2 – 7 kbar (Brunel and Kienast 1986; Hubbard 1989; Hubbard and Harrison 1989; Lombardo et al. 1993; Pognante and Benna 1993). The thermal peak of metamorphism was close in time to the reported emplacement age of the leucogranites (24 – 17 Ma; Searle et al. 2003). It was also suggested that no external heat source was required to explain the formation of leucogranitic magma in the Mount Everest and Makalu Massifs (Visona and Lombardo 2002). The peak of metamorphism in the study region is constrained between 42 – 28 Ma (Lombardo et al. 1993; Pognante and Benna 1993). This time span corresponds well with the evidence reported from the Mount Everest region (36 – 28 Ma; Simpson et al. 2000) for the timing of prograde

Figure 2: The northwestern wall of the Makalu Massif (8 475 m) consists of the Miocene leucocratic granite body and its injection zones into paragneisses. The extraglacial semiarid climate-morphogenetic zone above Barun glaciers displays conspicuous landform patterns of intensive glacigenic, nival, cryogenous and eolian processes. (Photo Jan Kalvoda.)

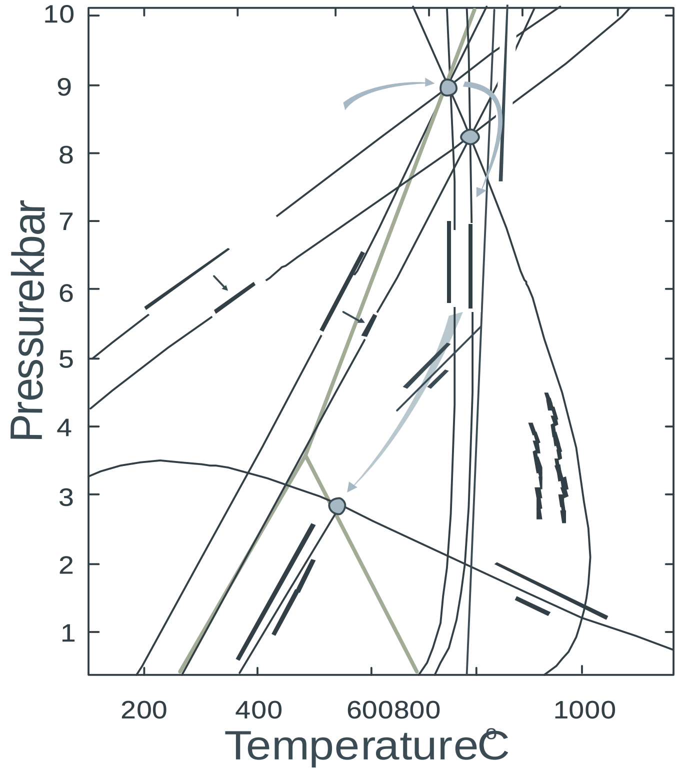
Figure 3: The uppermost part of the very steep southern walls of the Nuptse (7 879 m) and Lhotse Shar (8 383 m) with evident features of cryogenic and eolian modellation, consisting of leucocratic granites of the Miocene age and their upper injection zones intruded into the biotite gneisses of the Chomolongma group. (Photo Jan Kalvoda.)

metamorphism. Given the age of crustal melting constrained by the U–Pb systematics in zircons extracted from the Makalu leucogranites (24.4 – 21.7 Ma; Schärer 1984), there is only a short period of time remaining between the peak of metamorphism and the emplacement of leucogranites into the black gneisses of the Rongbuk Formation. This time delay can be explained by thermal re-equilibration of the buried crust before the onset of the large-scale crustal melting.

There are several constraints on the metamorphic evolution of rocks in the studied section of the High Himalayan slab. These are based on observations made on samples used in this study as well on other samples collected along the studied section (Kalvoda 1984a, 1992; Jaroš and Kalvoda 1978a, b; Palivcová et al. 1982). The studied section of the High Himalayan slab in the Makalu region is well exposed along the Barun Khola river (compare Kalvoda 1992; Svojtka et al. 2003; Kalvoda et al. 2004). The base of the section is marked by a series of thrusts of the Main Central Thrust zone that separate the High Himalayan slab from the Lesser Himalayan sequence of weakly metamorphosed sediments and granitic intrusions, such as the Num orthogneiss. Buriánková and Košler (2001) reported U–Th–Pb crystallisation ages of monazite from high-grade Barun gneisses that form the upper part of the High Himalayan slab in eastern Nepal. Monazite is useful accessory mineral that can be dated to constrain the timing of metamorphic events in amphibolite and higher grade facies rocks. Monazite in metamorphic rocks commonly develops at temperatures below the closure of the U–Th–Pb isotopic system. Its ages are interpreted as dating conditions close to the thermal peak of metamorphism.

Typical samples of cordierite-biotite-sillimanite-garnet gneiss (e.g. Hi – 262, compare Figure 3) as a part of the Namtche migmatite orthogneiss unit were collected from outcrops in the lower part of the eastern wall of Peak 4 at an elevation of 5 200 m. It also contains K-feldspar, plagioclase, quartz and a small amount of muscovite; apatite, monazite and zircon are present as accessories. Monazite is present as up to 150 µm near-euhedral grains in biotite and sillimanite aggregates. Sample Hi – 270 is a banded migmatitic cordierite-biotite-sillimanite-garnet gneiss collected from a rock pillar in the eastern wall of Peak 4, 1 km NE of icefall of the Lower Barun glacier at the elevation of 5 200 m. The mineral assemblage is similar to sample Hi – 262, except for K-feldspar and muscovite that are not present. Monazite occurs as near-euhedral grains in sillimanite and biotite aggregates. Sample Hi – 282b is cordierite-sillimanite-biotite gneiss with garnet collected from a pillar near the lower margin of a large ice field on the SW wall of Makalu, elevation 6 350 m. Monazite in this sample is present as 50 – 200 µm irregular grains enclosed in K-feldspar and biotite. Sample Hi – 292 was collected from a rock pillar gorge in the lower part of SW wall of Makalu at an elevation of 5 500 m (Figure 2). It is a migmatitic gneiss with mineral assemblage comprising quartz, K-feldspar, plagioclase cordierite, biotite and sillimanite. Apatite, rutile, zircon and monazite are present as accessories; monazite forms 50 – 100 µm near-euhedral grains in biotite and K-feldspar. Sample Hi – 293 is garnet-biotite-sillimanite gneiss collected from outcrops adjacent to the sheet-like intrusions of leucogranite in the central part of the SW wall of Makalu at an elevation of 6 800 m. The mineral assemblage is similar to sample Hi – 292 but sample Hi – 293 also contains an accessory amount of corundum. Monazite is present as up to 150 µm near-euhedral inclusions in biotite, K-feldspar and sillimanite.

Syn- and post-kinematic garnet porphyroblasts in the migmatitic gneisses of the Makalu area are compositionally zoned and often contain inclusions of alumosilicates, plagioclase, micas, monazite and allanite. An early high-pressure event is documented by the presence of kyanite relics in the garnet-biotite-sillimanite gneisses (Buriánková and Košler 2001). The high-pressure event was followed by decompression and an increase in temperature that resulted in the kyanite to sillimanite transformation, and cordierite formation. Buriánková and



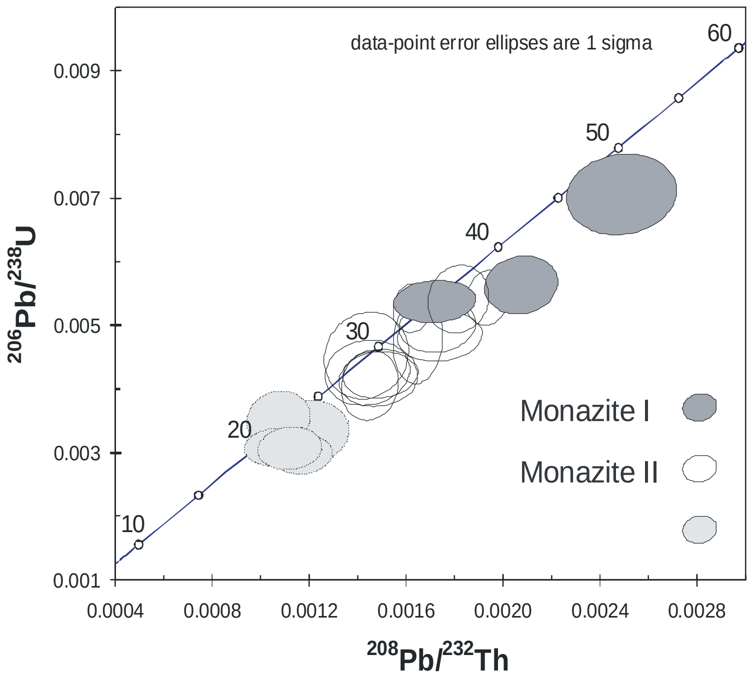




Figure 4: Metamorphic features determined on typical samples of Barun gneisses from the Makalu area: Figure 4A – pressure–temperature grid of gneisses showing their clockwise metamorphic path; Figure 4B – total U–Th–Pb concordia diagram for monazites.

Košler (2001) utilised conventional geological thermometers based on the Fe–Mg exchange reactions to derive temperature conditions of metamorphism for the studied rocks. The summary of pressure – temperature (P–T) estimates and the resulting P–T path are given in Figure 4A. Presented data are consistent with a clockwise P–T evolution through the kyanite stability field at maximum pressures of at least 9 kb and temperatures close to the biotite dehydration melting at ca 800 °C in the sillimanite stability field.

Kyanite has been found in garnet-bearing quartz-feldspathic gneisses elsewhere in the Makalu region (Buriánková and Košler 2001; Košler et al. 2001; Svojtka et al. 2003). Sillimanite was present either as fine fibrous aggregates replacing kyanite grains or in elongate fibrous aggregates resulting from the reverse biotite dehydration melting reaction. Sillimanite was also formed during cordierite breakdown, forming fibrous aggregates surrounding the pinitized cordierite grains. The presence of kyanite and relics of kyanite grains within fibrous sillimanite in some samples along the studied section of the Makalu – Barun area are indicative of kyanite → silllimanite reaction. The presence of K-feldspar and garnet in leucosome suggests a biotite dehydration melting reaction during the decompression. Reverse reaction and formation of cordierite (melt → Crd + Bt + Kfs + $H_2O$) took place during cooling following the decompression. Further cooling and increased fluid activity resulted in formation of chlorite at the expense of biotite and partial break-up of cordierite.

Košler et al. (2001) utilized a VG PlasmaQuad 2 S+ quadrupole ICPMS instrument coupled to a NdYAG laser to measure Pb/Th and Pb/U isotopic ratios in monazites in standard petrographic polished thin sections. The results of U-Th-Pb dating are plotted in a conventional concordia diagram in Figure 4B. Monazite is present in the studied samples in several textural relationships with other minerals. The texturally oldest monazite (I) is represented by subhedral grains (30 – 200 µm in size) that are included in K-feldspar and biotite (sample Hi – 282b). This monazite also overgrows the sillimanite aggregates that formed as a result of kyanite breakdown and several monazite crystals were found to contain inclusions of sillimanite that have similar orientation as the sillimanite fibres in the adjacent matrix. Back-scattered electron imaging revealed the presence of compositional inhomogeneities in grains of monazite (I) grains that could be interpreted as older inherited domains. Laser ablation ICPMS dating of this monazite yielded a wide range of ages between ca. 50 – 35 Ma. Given the textural relation of monazite (I) and sillimanite, this obtained age should constrain the phase of decompression following the peak metamorphic conditions.

Another type of monazite (II) forms subhedral to euhedral crystals in averaging 50 µm in size at triple grain boundaries between feldspar, biotite and garnet (or cordierite) and at the contacts between apatite and biotite in samples Hi – 262, Hi – 292 and Hi – 293. This textural relation suggests that there was a phase of monazite growth during the retrograde phase of metamorphism, possibly associated with the formation of Crd, Grt, Bt and Kfs. Laser ablation ICPMS dating of this monazite yielded a wide range of ages between ca. 40 – 28 Ma. The crystallization of monazite (II) constrains a maximum age limit for the thermal peak of metamorphism, characterized by formation of sillimanite, biotite and cordierite ± garnet.

The third phase of monazite growth (Figure 4) is constrained to domains adjacent to the intrusions of leucogranites. This monazite is euhedral and up to 500 µm in size (samples Hi – 270 and Hi – 292). Monazite (III) grains are homogeneous in back-scattered electron images. The ages of the third type found in granite-like domains ranges between 24 – 19 Ma, i.e. overlapping in time with the intrusion age of the Makalu leucogranite (24.4 – 21.7 Ma; Schärer 1984).

In situ laser ablation dating of monazite inclusions in biotite, sillimanite, feldspar and quartz from the migmatitic gneisses of the Makalu area record a polyphase metamorphic evolution between 35 – 21 Ma (Buriánková and Košler 2001; Svojtka et al. 2003). The older monazites constrain a maximum age limit of ca 35 Ma for the sillimanite grade metamorphism, while the younger monazite ages (ca 21 Ma) reflect a thermal event associated with the intrusion of the Makalu granite at 24.4 – 21.7 Ma (Schärer 1984). Pressure and temperature conditions of the later event, recorded by garnet composition and Fe–Mg exchange between garnet and biotite, were 3.2 – 5.0 kbar and 600 ± $25^0$ C, respectively. Maximum age of the metamorphism recorded by Th–U–Pb isotopic systems in monazite from the Makalu area corresponds to the timing of prograde metamorphism reported from the Mount Everest region (36 – 28 Ma; Simpson et al. 2000).

The presence of kyanite in the mineral assemblage of the Barun gneiss, and its breakdown to sillimanite, are indicative of decompression that started before or shortly after the rocks reached temperatures close to the biotite dehydration melting (Svojtka et al. 2003). This succession of metamorphic reactions is typical of collision orogens where burial of rocks and an increase in pressure precedes the heating, followed by decompression and exhumation of rocks towards the surface (Simpson et al. 2000**;** Beaumont et al. 2001). The total amount of denudation today in the rugged high-mountain relief of the Himalaya can be estimated at approximately 6 000 m per Ma (Hubbard et al. 1996**;** Searle et al. 2003).

The major exhumation events related to denudation and uplifting processes in the Makalu area were recorded by Svojtka et al. (2003) and Kalvoda et al. (2004). The low-temperature thermochronometry of the crystalline units in the upper part of the Barun valley above the Main Central Thrust zone was studied at 4 600 – 5 000 m elevation. Zircon and apatite grains from the biotite-sillimanite paragneiss, migmatitic orthogneiss and glacifluvial sediments were analysed. The fission-track zircon cooling ages for the migmatitic orthogneiss and paragneiss were 7.1 ± 1.0 Ma and 12.2 ± 1.0 Ma, respectively, interpreted as resulting from a steady slow cooling through the zircon partial annealing zone between 310 – 230° C. Fission-track ages of zircons from glacifluvial sediments (9.0 ± 0.7 Ma and 9.2 ± 1.0 Ma) represent a mixture of the fission-track cooling ages of rocks exposed in the upper part of the Barun valley (Svojtka et al. 2003). Cooling ages for migmatitic orthogneiss and biotitic-sillimanite paragneiss were determined as 3.2 ± 0.2 Ma and 6.6 ± 0.6 Ma, and the glacifluvial sediments yielded an age of 3.7 ± 0.5 Ma and 4.0 ± 0.5 Ma.

The estimate of an average cooling rate is based on the ratio of closure temperature and obtained ages (Figure 5). The apatite and zircon fission-track data suggest (Svojtka et al. 2003) that the exhumation / denudation processes are characterized by initially high cooling rates of ~ 46° C/Ma (7.1 Ma – 3.2 Ma) for the migmatitic orthogneiss and ~ 32° C/Ma (12.2 Ma – 6.6 Ma) for the paragneiss. The cooling rate for the migmatitic orthogneiss and the paragneiss from the Pliocene / late Miocene to the present time decreased to ~ 22° C/Ma and ~ 11° C/Ma, respectively. Modelling of the thermal evolution of apatites by AFT Solve software (Ketcham et al. 2000) has shown that the migmatitic orthogneisses cooled from the apatite partial annealing zone (60 – 120° C) to 20º C since ca. 3.0 Ma (Figure 5). The temperature has not significantly changed since approximately 1 Ma ago (Svojtka et al. 2003; Kalvoda et al. 2004). The apatite fission-track data documenting a rapid decrease of temperature from 120º C to 20º C in the gneisses between 3.0 – 2.0 Ma suggest the existence of dissected mountain relief at that time which probably developed at a substantially lower elevation during the Pliocene (Svojtka et al. 2003; Kalvoda et al. 2004). The rapid decrease of temperature of the exhumed crystalline rocks is perhaps evidence for an episode of intensive erosion and denudation of the palaeorelief of the High Himalayan nappe.

High temperature and pressure geochemistry of the main rock units relates to the evidence of igneous emplacement and metamorphic episodes connected with collision orogeny in the Cenozoic (e.g. Beamont et al. 2001; Searle et al. 2003). Described igneous and metamorphic events are related to the plate tectonic setting and in this period in the evolution of the East Nepal Himalaya the uplift was caused primarily by tectonic processes. Low-temperature thermochronometry indicates an extent of exhumation and erosion of crystalline rock massifs as well as their position with respect to the evolving landscape elements during the long-term and very intensive processes of mountain building (Bishop 2007).

Presented thermochronological results from the Makalu – Barun area are in correspondence with relatively extensive studies in other Himalayan regions (e.g. Burbank et al. 1996, 2003; Beaumont et al. 2001; Zeitler et al. 2001; Simpson et al. 2000; Searle et al. 2003). However, test of our preliminary regional interpretation using a larger set of samples of crystalline rocks will be valuable. Rock samples for thermochronometry can be collected on steep walls of peaks or in deeply incised valleys where the outcrops can be found. We can ask, what would a measurement be on similar samples obtained still deeper below the surface under a ridge line between incised valleys. It might tell different results which might mean the thermochronometry measurements are biased by its location in the extremely dissected high-mountain relief. Low-temperature thermochronology used in complex studies of long-term landscape evolution of tectonic active orogens has variable interpretations (e.g. Burbank et al. 1996, 2003; Bishop 2007) due to the configuration and age of paleosurfaces as well as different patterns of crustal flow, rock uplift and erosion.

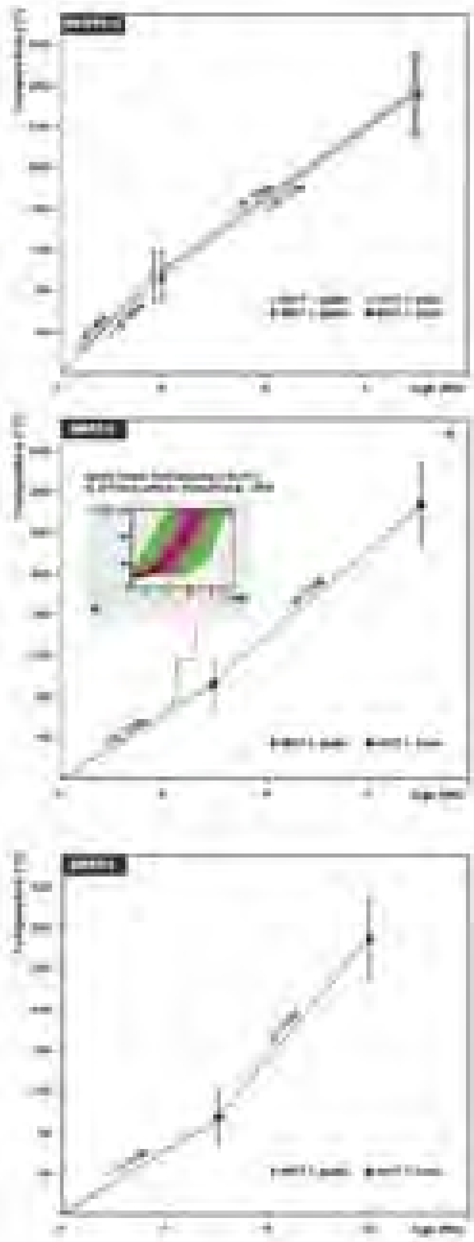


Figure 5: A closure temperature vs. cooling age plot for the studied Makalu – Barun area samples. Key: MKFT1 and MKFT2 – gneisses in glacifluvial sediments, MKFT4 – migmatitic orthogneiss, MKFT5 – biotitic-sillimanite paragneiss.

## 3. Landform patterns and climate-morphogenetic processes

The rugged high-mountain landscape of the Makalu – Barun region displays an elevational gradient of over 7 000 m with very varied climatic and biogeographical zones. Recent climate-driven morphogenetic processes in the extremely dissected relief can be described in the framework of extraglacial and glacial zones, the periglacial zone and the seasonally cold/warm humid zone.

The extraglacial high-mountain zone with a rock-cut landscape of alpine-type ridges displays a dynamic integration of deep weathering with major glacial and nival morphogenetic processes in a very cold and semi-arid environment (Figures 2 and 3). The intensity and duration of temperatures below freezing point led to deep rock disintegration and macrogelivation (Kalvoda and Valenta 1997; Kalvoda 2003). Avalanches and rockfalls are frequent, and aeolian erosion and stagnation of the volume of ice and snow masses is very conspicuous. Platforms in the shape of small altiplanos and large glacial valleys serve as an accumulation space for snow and glacier masses of the High Himalayan Range. The present equilibrium-line altitude in the Barun and Khumbu Himal areas lies between 5 600 – 5 700 m.

The erosional landforms of alpine-type relief are developed in two structural denudational levels. The lower surface occupies the roof part of the Barun nappe at 4 900 – 5 200 m a.s.l., the second occurs in the upper parts of the Makalu Massif (Figure 2) and it is situated more than 1 000 – 1 200 m higher than the first. The Makalu (8 475 m) is a huge peak, partly isolated tectonically, as are the Baruntse and the unnamed peak of 7 502 m elevation. Peak 4 is an extensively glaciated mass sharply bounded by rock faces. The summits of other peaks are either the terminations of lateral crests or the eroded relicts of connecting ridges between the main mountain massifs. The denudational slopes dipping up to 48$^{o}$, mostly covered with a 20 – 120 cm thick stratum of blocky and sandy-gravelly detritus, extend over large areas, especially between the confluence of the Chago and Barun glaciers and the ridge of elevation point 6 140 m, below the peak of the 6 090 m elevation, and in the surroundings of the south-southwestern peak of 6 260 m elevation.

The upper end of the Barun glacier valley cuts into the main ridge of the High Himalaya between the Chomolongma and Makalu Massifs. This glaciated area passes with a wide transfluence of ice masses into the Kangshung valley to the east of Mount Everest. The nunataks are developed in two types: (1) massive rocky crests, the heights of which exceed the maximum thicknesses of the fossil glaciers (e.g. ridges with debris eluvium in the environs of the 6 140 m elevation point, and those in the Chago valley) and (2) steep exposed relics of pinnacles between slope and hanging glaciers, e.g. northwest of the 7 057 m elevation point peak, in the lower parts of the northwestern and southwestern cliffs of the Makalu and west of the 6 825 m elevation point. The glacial polish on the groups of *roches moutonées*, exposed 1.5 km from the northern peak of elevation point 6 260 m in the vicinity of the 6 540 m peak below the northwestern face of the Makalu and above the Barun Pokhari lake towards the Japanese Col (Figure 1), provide evidence of the considerable extent of the earlier stages of glaciation.

Range and volume of Himalayan glaciation has been very changeable during the Quaternary which depends on the variable paleogeographical history of the large region between the Indian Ocean and the Tibetan Highland. The complexity of landform evolution and dimensions over which interactions of both exogenic and endogenic processes operate in the Himalaya and the Tibetan Highland is studied by Fielding (1996), Owen et al. (1998, 2002), Burbank et al. (1996, 2003) and Kuhle (1999, 2004). In these papers presented conceptual models of the chronodynamics and the range of orogenic and isostatic uplifts of the Tibetan Highlands and neighbouring regions are supported by recent geomorphological and geological observations as well as geophysical measurements. Integration and feedbacks of very dynamic orogenetic and climate-morphogenetic processes as main cause of large-scale glaciations and their changes during the Quaternary are accepted (Fort 1966, 2004; Kalvoda 1978, 1992, 2007; Owen et al. (1998, 2002; Kuhle 1999, 2004, 2005; Benn and Owen 2002; Owen 2004). Historical-genetical models of long-term paleogeographical changes of the natural environment in Tibetan Highland and its neighbouring regions are in these papers considerably different. However, field exploration of outlying areas, research of genesis and the age of landforms, including dating of late Cenozoic sediments, necessary for correlations and testing of these conceptual models, are quite progressively evolving.

It is not possible to verify satisfactorily the hypothesis about a large-scale inland glaciation of the Tibetan and Himalayan region in the Pleistocene (e.g. Kuhle 1999, 2005) only by our local observations in studied extraglacial, glacial and periglacial zones of the Makalu – Barun area. The methodological reason is that a set of landforms of mountain massifs and valleys is very young due to extreme intensity of climate-morphogenetic processes. Geomorphological analysis deals only with existed relief types and landforms. Therefore, its local results can be interpreted for large-scale geographical contexts only approximately and in different manner. However, an architectonic configuration of morphostructural and glacigenous landforms give evidence (e.g. Figures 2 and 6) about existence of ancient and large glaciations of the Makalu – Barun region in the Middle and Late Pleistocene, which were considerably older that preserved relics of glacigenous and glacifluvial sediments.

Figure 6: The upper Barun valley area with a step-like climate-morphogenetic zoning and conspicuous morphostructural patterns of the region. Dejection planes of rockfalls on extremely steep and high walls of crests built by crystalline rocks are displayed as well as glaciofluvial and lacustrine sediments or groups of moraines which can be correlated with equivalent landforms in the Mount Everest region. In the Solo Khumbu area Richards et al. (2000) have dated these groups of moraines as the result of three advances at 18–25 ka, 10 ka and 1–2 ka. (Photo Jan Kalvoda.)

Intensive glacigenic, nival and cryogenic destruction of the mountain slopes is indicated by fresh tension cracks in scree slopes or by the sharp lines of mostly glaciated crests (Figure 3). Gravitational landforms include active talus fans developed at the foot of huge rock faces, as well as talus fans of great areal extent, and accumulation piles resulting from rockfalls and landslides. A cover of sliding detritus, mostly consisting of boulders, is formed on slopes. The lateral moraines of glaciers are frequently overlain by accumulations of slope debris ranging in size from coarse sand to boulders.

The glacial zone of the Chomolongma and the Makalu – Barun regions displays a recent regression of glaciers and a rapid decrease in their volumes (comp. Odell 1925; Heuberger 1956; Fushimi 1978; Kalvoda 1978, 1979a, b; Kuhle 1984, 1997, 1999; Heuberger and Weingartner 1985; Zheng Benxing 1988; Kőnig 2001). The spreading of the periglacial zone to the detriment of lower areas of the very cold glacial zone is striking (Figure 6). The valleys and ridges are fully filled with glacier masses at high altitudes above ca 6 000 m a.s.l. Large ice source areas often contrast with very narrow canyon-like lower parts of valleys. A striking phenomenon is the occurrence of the relics of sediments in accumulation landforms of Upper Pleistocene age and younger than 50 x $10^3$ years (Bordet 1961; Kalvoda 1984a, b, 1992, 2007; Iswata 1987; Yagi and Minaki 1991; Kuhle 1999, 2004, 2005; Richards et al. 2000). The southwestern side of the Makalu Massif (8 475 m) has its foot at altitudes of 4 900 to 5 000 m. Observations from the years 1971, 1973, 1976 and 2002 have shown (Kalvoda 2003, 2007) that conspicuous recent changes in the rock slope patterns and, especially, in the volume of ice masses accompanied by a recession of frontal parts of hanging glaciers, are only found in the lower parts of the walls.

The oldest accumulation landforms are represented by fossil moraines of the Barun glacier. In presented paper are used the names of the moraines after equivalent glacial accumulations in the Khumbu Himal region (compare Iswata 1976a, b; Kalvoda 1978; Richards et al. 2000). The moraines of the Dusa type from younger period of the Late Pleistocene (Kalvoda 1978, 1984a; Late Glacial Stages I – IV by Kuhle 1999, 2005), have been preserved between the foot of the eastern rock faces of the peak of elevation point 6 380 m in the Peak 4 Massif and the Barun Pokhari lake, as well as north of the Shershon site. These moraines adjoin the valley slopes, attaining thicknesses of 120 – 160 m and above Shershon at least 60 – 80 m. The surface of the moraines of the Dusa type is slightly furrowed due to periglacial processes and is consolidated by alpine steppe vegetation.

Holocene moraines of the Changri type lie at the southern foot of the Makalu Massif. They are the relics of glacier advances (related to Neo-glacial Stages V – VII after Kuhle 2005) during which ice masses filled the floor of the lower parts of the Barun valley up to the Shershon site. The Changri-type lateral moraines are flat and 15 – 20 m high and the frontal moraines, up to 30 m thick, have their arched ridges chaotically distributed. The valley glaciers in the Makalu – Barun region are bounded by the conspicuously asymmetrical walls of the lateral and frontal moraines of the Khumbu glaciation stage. Surfaces of these Late Holocene moraines (Historical Stages VIII – X, Kuhle 2005) with relative heights of 12 – 40 m are practically unconsolidated. The moraines consist of coarse sandy unsorted material, with a pronounced predominance of light-coloured granitic material prevailing over dark gneisses and varied coloured migmatites.

The moraine of the recent Lingten oscillation has been preserved on the western side of the main Barun glacier tongue, at altitudes of 5 000 – 5 050 m, and in front of the slope glaciers of the western and northern cliffs of Peak 4 (Stage XI after Kuhle 2005). The moraine occurs as 5 to 15 m high, broadly-based ridges, zig-zag or arched in shape, with a chaotic granular texture. The entire Barun glacier tongue and its lateral ice flows from the valley of the eastern part of the Baruntse Massif and from the Chago valley below 5 500 m a.s.l. are covered with an almost continuous surface moraine (Stage XII, Kuhle 2005).

Of particular note is a time correlation of the Lower Barun glacier icefall situated at the end of the narrow hanging valley north of the Chukhung Massif. The shape of the icefall and its pattern of crevasses has not changed since the year 1973 (Kalvoda 2003, 2007), but the whole lower lying flat tongue of the glacier is close to extinction in the near future (compare Figures 7 and 8). Recent decrease of ice masses is also evident from position of small glacier tongues in the lateral valleys related to their moraines of the Late Holocene age. The shape of moraines has not changed in the last 30 years and their ridges are only more rounded.

In the foreland of the present-day glaciers a system of glaciofluvial and lacustrine depositional landforms is developed. The Late Holocene to recent terraces and cones of outwash sediments represent the earlier of two generations of glaciofluvial landforms which are situated in the depressions between individual oscillation ridges of moraines. The presence of present-day and former lakes of glacial origin on valley floors (Figures 6 and 8) is indicated by terraces of lacustrine sediments up to 6 m thick surrounded throughout by fossil moraines. Of the periglacial landforms, paved and polygonal soils of limited areal extent have been found (Kalvoda 1979b; Kalvoda and Smolíková 1981), although always at altitudes of below 5 200 m on fossil moraines.

Young aeolian landforms consist of fine sandy to silty accumulations especially originating from unconsolidated morainic, lacustrine and slope sediments. The terraces, composed of up to 3 m thick wind-blown sand, rest in the lee of fossil morainic ridges on the flanks of the slope-glacier tongue of the eastern face of Peak 4, and to the northwest of the Shershon site. Irregular small active dunes up to 50 cm high border, for example, the bottoms of the depressions between the Holocene and recent moraines of the Barun glacier tongue, the environs of the Barun Pokhari lake and its southern foreland (Figure 6). The aeolian sands are strikingly light-coloured, white to yellow, with angular grains and a high proportion of quartz and micas.

Accelerated rates of erosion have been observed in the periglacial environment located outside the present glaciers (Iswata 1976a, b; Kalvoda 1979a, b, 2007). The recent rapid retreat of the glaciers is accompanied by a distinctive increasing of the active periglacial zone. This increases the volume of transported products of denudation and the level of geomorphological hazards, including frequent rapid events of mass movements triggered by earthquakes, avalanches, flash floods and landslides. Present decreases in the distribution of permafrost has implications for landscape stability (Kalvoda and Smolíková 1981; Smolíková and Kalvoda 1981), which is mirrored in solifluction, rock-glacier movements and sediment release into streams and rivers.

From the icefall south of the Peak 4 Massif to the Tadoza site, the valley floor is practically filled with chaotic ridges of fossil oscillation moraines, rockfalls, talus fans and the Lower Barun glacier tongue (Figure 7). In the Barun Khola river valley, glacial and periglacial landforms, which reflect the influence of a former cirque and slope glaciers, occur especially to the east of the Lower Barun glacier. They are located above the Yangle Khalka site and the Arun river canyon-like valley (Figure 1) in the foreland of structurally controlled denudational slopes which are developed along the front of the Main Central Thrust.

The Barun Khola canyon-like valley in the area between Tadosa and Phematan sites has considerably U-shaped pattern (Kalvoda 1979a, b, 1984a, 1992). It is primarily of glacial origin, similarly as the Iswa Khola valley in the west, with main stages of its relief evolution during the Middle and Late Pleistocene. This period of glacigenic evolution of the Barun Khola valley probably includes Glacial Stages – 1 and 0 (especially the Last Ice Age) described in the Himalayan region by Kuhle (1999, 2005). The glacial landforms also occur in the lateral, mostly hanging, valleys and on the crests between them.

Figure 7: The main Lower Barun glacier tongue with a system of icefalls is conspicuously changed and the decrease of its volume is accompanied by formation of a new lake dammed by a not very high recession moraines (Figure 7 – 1973, Figure 8 – 2002). These accumulation landforms are situated below slope glaciers and icefalls at the bottom of an earlier broad glacial valley. (Photo Jan Kalvoda.)

Figure 8: Present-day lake originated in the last three decades in frontal part of the Lower Barun glacier is dammed by low recessional moraines and recent landslides. (Photo Jan Kalvoda.)

The periglacial weathering features of the marginal ridges of the Barun Khola valley disappear even before this valley enters the rocky cliffs of the High Himalayan nappe in the evergreen monsoon mountain forest vegetation zone. The Barun Khola valley floor, up to 2 800 m a.s.l., is covered by thick (up to ten metres) deposits of glaciofluvial and slope sediments cut by vertical erosion from the Phematan locality to as low as paragneisses and granulites of the lower part of the Barun nappe.

The lower part of the Barun valley is constantly being reshaped by huge and frequent slope movements and simultaneous rapid erosion of glaciofluvial and slope sediments deposited in accumulation landforms of Holocene age. The Barun and Arun river regions are areas of frequent natural disasters with high risks involved to all types of human activities. A large amount of new rockfall accumulations has been found in the lower part of the Barun Khola valley, which is part of a subalpine conifer zone with Abies and Juniperus. Steep erosion-denudational slopes of the Arun river valley, with very frequent slope movements, are covered by a monsoonal evergreen forest, which is being intensively reshaped by burning and clearing (Daniel et al. 1985; Zomer et al. 2001, 2002). The catastrophic course of landscape changes stimulated by human activities (Bayers 1996; Carpenter and Zomer 1996) is detected even in national parks of the Nepal Himalaya.

The dividing ridges between the Tsang Po and Arun rivers lie 120 km north of the Chomolongma Massif in the Shekar Dzong area, and reach altitudes of 6 500 m. In contrast to the Tibetan Himalaya, the Arun valley is of a very moderate gradient, in the High Himalaya it suddenly becomes steeper in its deeply incised canyon. Kuhle (2005) suggested that the Arun glacier (in length approx. 80 km and/or Barun–Arun glacier 67 km) has flowed down to ca 500 m a.s.l. in the maximum of the Last Ice Age glaciation (60 – 18 Ka BP, Stage 0).
At the confluence with the Barun Khola gorge, the Arun canyon is incised as deep as 1 050 m a.s.l., and after a further 12 km, near the village of Num, it is at 850 m a.s.l. During our repeated field works in the relief section from Hatia site to Num and Sedoa villages and Tumlingtar area (e.g. Kalvoda and Smolíková 1981; Kalvoda 1984a; Kalvoda and Valenta 1997; Kalvoda 2007), variable relics of accumulation landforms with Holocene and recent series of fluvial, slope and glacifluvial sediments were documented. The Nepalese part of the Arun valley, up to the northern margin of the Tumlingtar intermontane basin in the Lesser Himalaya, has a steep irregular gradient. The ridges of the Lesser Himalayan relief north of Kumalgaon village distinctly plunge under the frontal parts of the crystalline rocks of the High Himalayan nappe. Wager (1937) suggested that isostatic response to Arun valley incision may contribute to Himalayan range uplift.

The observed recent landform changes confirm the high intensity of climate-driven morphogenetic processes (compare Figures 7 and 8), especially with very effective erosion and transport of weathered material in the periglacial and seasonally warm humid (monsoon) mountain zones. This is in striking contrast to the relatively small range of denudation and transport of weathered material in the northern cold and semi-arid climatic zones of the Himalaya in Tibet (Fielding 1996; Kuhle 1999, 2004). The palaeogeographical consequence of these long-term differences is the very deep penetration of erosion and denudation of rock massifs in regions of steep windward Tibetan-foreland transitions with the influence of humid air masses (Beaumont et al. 2001) leading to development of the deeply-entrenched high-mountain relief of the Himalaya. Extreme exhumation of deep crystalline rocks in the Himalaya during the late Cenozoic is the result of morphotectonic processes as well as the effective tuning of palaeogeographical changes to the extension of the main climate-morphogenetic zones.

## 4. Conclusions

Geomorphological analysis of landform patterns in the Makalu – Barun region has shown that high-mountain landforms of the East Nepal Himalaya are the result of morphotectonic processes integrated with denudation and erosional efficiency in different paleoclimatic conditions during the late Cenozoic. There are significant feedbacks between the rate of tectonic exhumation of deep crystalline rocks and climate-morphogenetic processes. High rates of denudation and valley incision in the Himalaya during the mature stage of collision orogeny has also stimulated isostatic component of uplift in the Quaternary.

Cooling history of rocks investigated by low-temperature thermochronology has confirmed rapid crustal uplift in the Makalu – Barun region during the late Cenozoic. Tectonic transfer of crystalline rocks was accompanied by long-term denudation and selective erosion of paleosurfaces. Coupled rapid orogenetic and climate-morphogenetic processes created extremely dissected topography today in the East Nepal Himalaya. The fundamental role of surface processes connected with the origin of the Himalayan landscape has been in removing an extremely large volume of regolith and/or Quaternary sediments out of the mountain belt.

The style of glaciation in the East Nepal Himalaya and neighbouring areas changed in successive advances during the Quaternary as a result of the evolution of valley system being created by the earlier events (e.g. Owen et al. 1997; Kuhle 2004, 2005). This implies a shift from ice caps to lengthy valley glaciers. Erosion is more localised along the valley alignments and in a succession of glacial advances driven by global climate changes, is a cumulative process. The products of weathering and erosion are shifted discontinuously through the landscape in the process of paraglacial sediment storage and their erosion (Benn and Owen 2002; Bishop 2007). Tectonic uplift is needed to trigger the incision that evacuates the sediment fills which is also connected with isostatic response.

Observations of landscape changes in the Makalu – Barun region on a decadal scale suggest that the frequency and magnitude of recent landform processes are increasing from very cold and dry extraglacial and glacial zones across a large periglacial zone up to seasonally cold/warm and subtropical humid zones. Extreme changes in Himalayan landscapes are driven and/or accompanied by present-day severe natural hazards such as earthquakes, rapid slope movements, flash floods, deforestation and soil erosion.

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