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The maximum Ice Age (LGM) glaciation of the Central- and South Karakorum: an investigation of the heights of its glacier levels and ice thicknesses as well as lowest prehistoric ice margin positions in the Hindukush, Himalaya and in East-Tibet on the Minya Konka-massif

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Abstract

A continuing prehistoric ice stream network between the Karakorum main crest and the Nanga Parbat massive has been evidenced, which, flowing down from the current Baltoro- and Chogolungma glaciers and filling the Shigar valley as well as the Skardu Basin, has flowed together with the Gilgit valley glacier to a joint Indus parent glacier through the Indus gorge. The ice stream network received an influx by a plateau glacier covering the Deosai plateau, which was connected through outlet glaciers to the ice filling of the Skardu Basin and the Astor glacier at the Nanga Parbat, as well as to the lower Indus glacier. The field observations introduced here in part confirm the results as to the Ice Age glacier surface area of Lydekker, Oestreich and Dainelli, but go beyond it. In additon, a reconstruction of the surface level of this ice stream network and its glacier thicknesses up to the highest regions of the present-day Karakorum valley glaciers has been carried out for the first time. In the area under investigation the Karakorum ice stream network showed three ice cupolas, culminating at an altitude of 6200–6400 m. Between the mountain groups towering 1000–2000 m higher up, they communicated with each other over the transfluence passes in a continuous glacier surface without breaks in slope. In the Braldu- and Basna valley ice thicknesses of 2400–2900 m have been reached. In the Skardu Basin, where the glacier thickness had decreased to c. 1500–1000 m, the ELA at an ice level of 3500-3200 m asl had fallen short to the extent that from here on down the Indus glacier a surface moraine cover has to be suggested. However, 80% of the surface of the ice stream network was devoid of debris and had an albedo of 75-90%. The lowest joint glacier terminus of the ice stream network was situated - as has already been published in 1988 – in the lower Indus valley at 850–800 m asl. The reconstructed maximum extension of the ice stream network has been classified as belonging to the LGM in the wider sense (60–18 Ka BP). Four Late Glacial glacier positions (I–IV), with a decreasing ice filling of the valleys, have been differentiated, which can be locally recognized through polish lines and lateral moraine ledges.

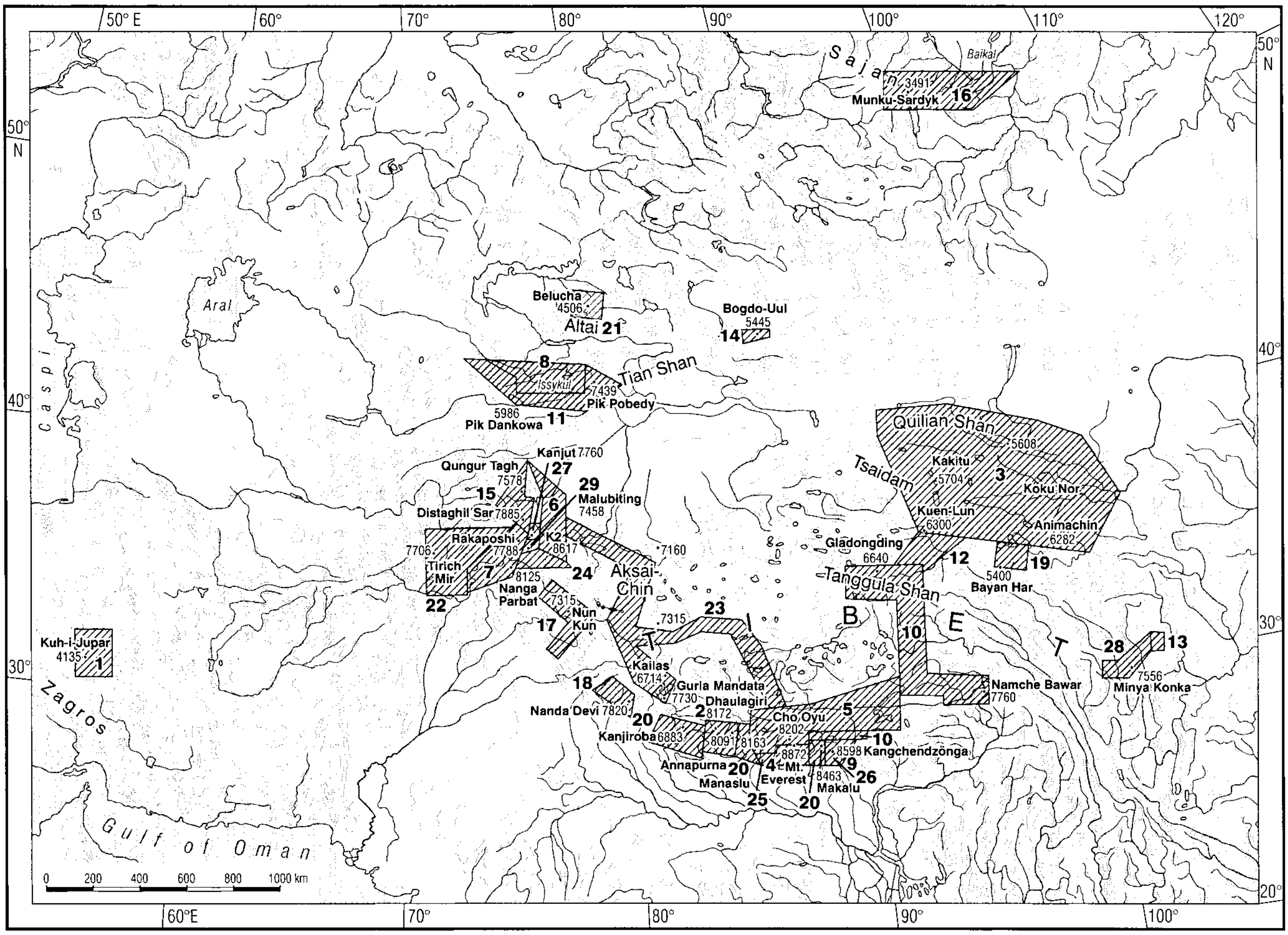
The valley (trough-) flanks with their ground moraine covers, oversteepened by glacier abrasion, have been gravitationally destroyed by crumblings, slides and rock avalanches since the deglaciation, so that an interglacial fluvial-, i.e. V-shaped valley relief has been developed from the in part preserved glacial relief. The contrast of the current morphodynamics with regard to the preserved forms is seen as an indication of the prehistorically completely different - namely glacigenic – valley development and the obvious rapidity of this reshaping at still clearly preserved glacial forms provides evidence of their LGM-age.

In an additional chapter the lowest ice margin positions, so far unpublished, are introduced, which have been reconstructed for the Hindukush, Central Himalaya and on the eastern margin of Tibet.

1. Introduction and method

This treatise is the regional continuation of a detailed and spatially extensive reconstruction of the Ice-Age glaciation in High-Asia. It completes the author's research on the prehistoric extent of ice and glacier thicknesses in High-Asia carried out since 1973 (cf. Figure 1) and published since 1974 (Kuhle 1974–2000) by further investigations in areas which have already been studied earlier or which have not yet been visited (Figure 1, Nos. 2, 4, 7, 17, 22, 24, 25, 26, 27, 28).

The geomorphological and Quaternary-geological methods applied in the field and laboratory are discussed in detail in the papers on empirical Ice Age research and the glaciation history of High-Asia (Kuhle & Wenjing 1988, Kuhle & Daoming 1991, Kuhle 1994a, 1997a, 1999a) published in the GeoJournal series 'Tibet and High-Asia – Results of Investigations into High Mountain Geomorphology, Paleo-



1: *1973, 1974* **2:** *1976, 1977, 1998, 2000* **3:** *1981, 1998* **4:** *1982* **5:** *1984* **6:** *1986* **7:** *1987, 1992, 1995, 2000* **8:** *1988* **9:** *1988/89* **10:** *1989* **11:** *1991*
12: *1991* **13:** *1984, 1991* **14:** *1986, 1992* **15:** *1992* **16:** *1993* **17:** *1993, 1996* **18:** *1993* **19:** *1994* **20:** *1994/95, 2000* **21:** *1995* **22:** *1995* **23:** *1996*
24: *1997* **25:** *1998* **26:** *1999* **27:** *1999, 2000* **28:** *2000* **29:** *2000* Draft: M. Kuhle (2000)

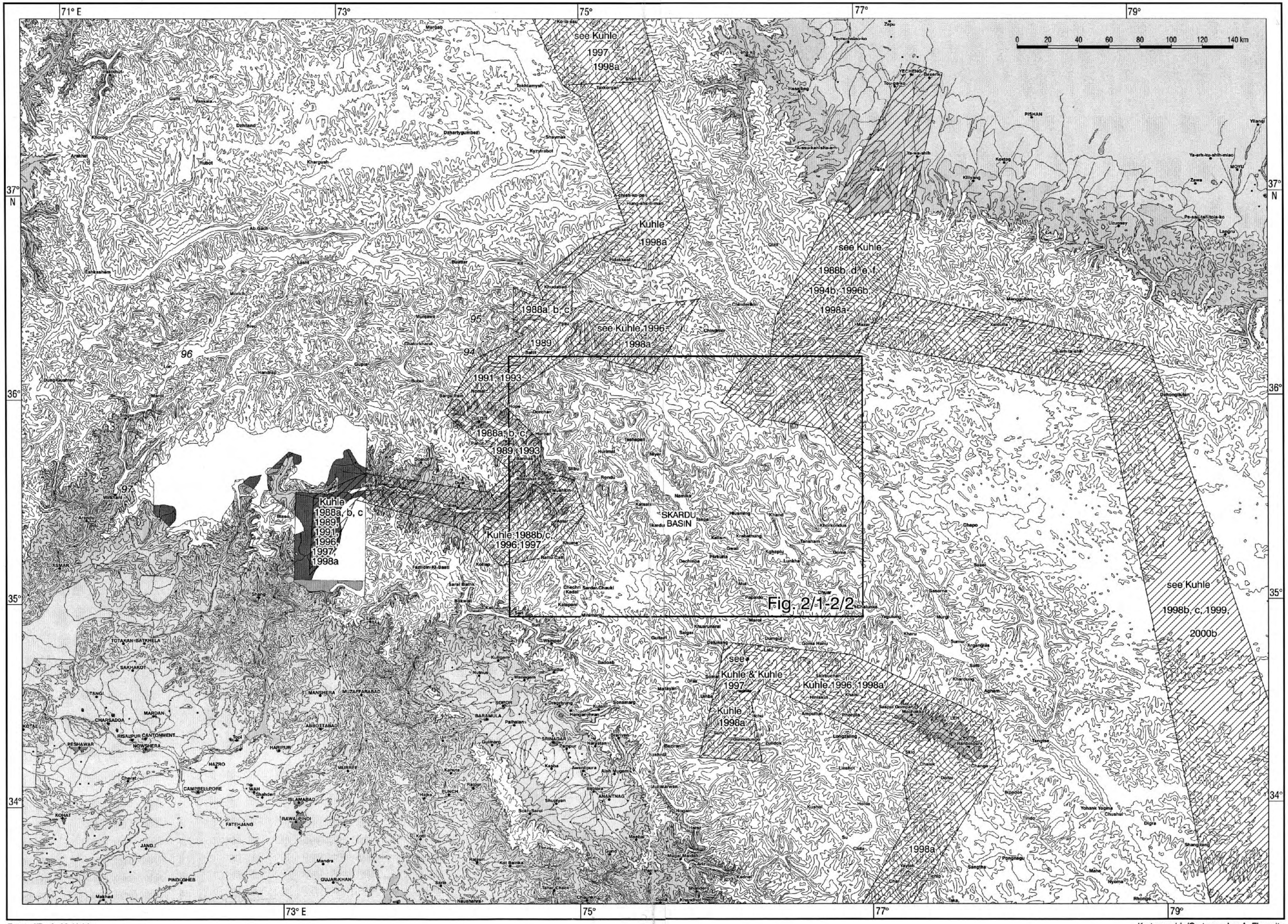
Figure 1. Research areas in Tibet and its surrounding mountains visited by the author. The study presented here introduces new observations on the Ice Age glacier cover from area No. 2, 4, 7, 17, 22, 24, 25, 26, 27, 28.

Glaciology and Climatology of the Pleistocene (Ice Age Research)' volumes I (1988), II (1991), III (1994), IV (1997) and V (1999). In consequence it seems unnecessary to re-introduce these scientifically common methods here.

2. The highest prehistoric glacier levels in the Muztagh Karakorum: K2-S-slope, Broad Peak- and Gasherbrum I (Hidden Peak)-W-slope

In the volume 'Tibet and High-Asia III' the glaciation of the Muztagh-Karakorum N-slope from K2 as far as into the Tarim Basin constituted the subject of the analyses (Kuhle 1994b). Now, the immediate continuation to the S, i.e. the area of the Blaldo valley and present-day Baltoro glacier, has been treated (Figure 1, No.24).

The orographic right-hand source branch of this glacier, the Godwin Austen tributary stream, runs with its orographic left bank along the flank of the Broad Peak (Photo 1 below No.3). Above the present-day glacier level a glacigenic polish band (Photo 1 ▲ on the left below No.3) extends up to a polish line at 6200–6300 m asl (Photo 1 ▬ ▬ on both sides of No.3). The glacigenic rock smoothings in the region of the K2-superstructure are rather sparse. Only the S-spur, which juts out of the S-wall far enough not to be showered by its avalanches, shows glacigenic roundings (Photo 2 ▲ centre). So, too, on the valley flanks, which in the vertical dimension are relatively little extended as e.g. the Nera Peak SE-face, quite large-scale flank polishings are preserved near to the snow-line (Photo 2 ▲ on the left). The enormous detritus masses of the surface moraine in the present-day glacier area (Photos 1–16 □) give an impression of the important down-slope denudation and contemporaneous reshaping during postglacial times. Additionally, the Ice-Age glacier polishing was combined with exarations and detersions, so that the flank surfaces left behind were rough anyway. From the LGM well into the Late Glacial these valley slopes were above the ELA (snow-line), which had been lowered by 1300–700 m. Thus, the meltwater-film on the rock faces needed for the fine-polishing was lacking. Important in this connection is the fact that cold-arid glaciers were concerned, the annual mean temperature of which was about −10 °C at the level of the snow-line. Probably it was even colder. The classic-glacigenic shape of the back-polished mountain



Entwurf/Draft: M. Kuhle

Kartographie/Cartography: A. Flemnitz

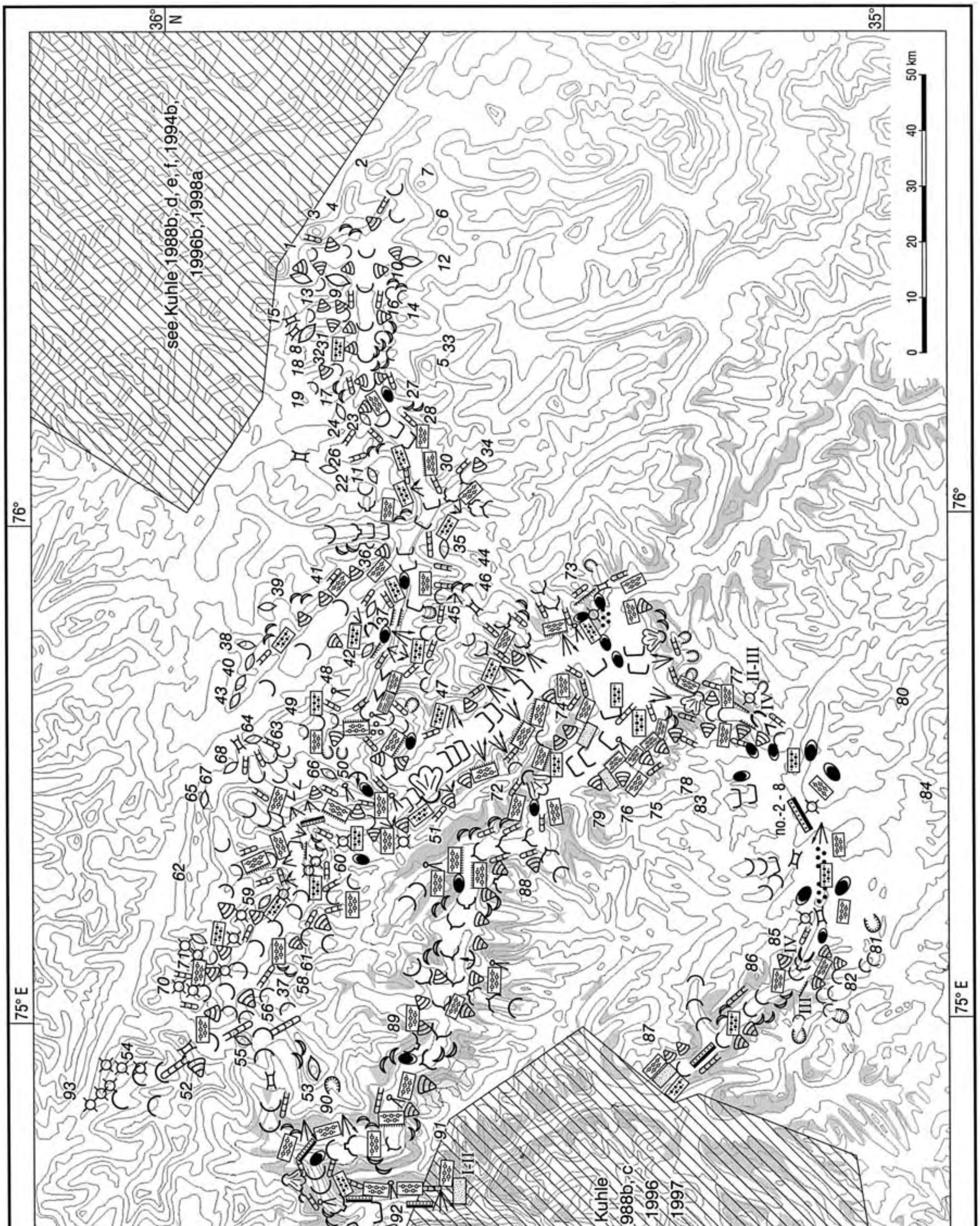


Figure 2/1. Quaternary-geological and glacio-gemorphological map of the Central-Karakorum and Deosai plateau.

LEGENDE / REFERENCE:

15 *Lokalität* / locality

Rundhöcker und ähnliche glaziäre Schlifformen / roches moutonnées and related features of glacial polishing

Sedimentproben, C^{14} Analyse / sampling, C^{14} analysis

Grundmoräne mit erratischen Blöcken / ground moraine with erratic boulders

Gletschertorschotterflur-Terrassen / alluvial terraces in contact with moraines

no.-2 *Gletschertor-Schotterflur-Stadium* / stage of outwash-terraces (explanation in text)

Schwemmschuttfächer, Schotterflurfächer / alluvial fan, outwash fan

Murkegel / mudflow cone

Felshohlkehle durch fluviale Unterschneidung / fluvial undercutting of the valley flank

Kames und subglaziale Schotterablagerung / kames and subglacial gravel deposits

Transfluenzpass / transfluence pass

glaziärer Flankenschliff / glacial flank polishing and abrasion

glaziäre Dreieckshänge / glacially triangular-shaped slopes (truncated spurs)

Kar / cirque

Endmoränen von Talgletschern / terminal moraines of valley glaciers

Ufermoräne, Mittelmoräne, Endmoräne / lateral moraine, middle moraine, terminal moraine (former ice margin)

glazialer Trog ohne und mit Schottersohle / glacial trough without and with gravel-bottom

´schluchtförmiger Trog´ / gorge-like trough

große Blöcke (erratisch und nicht erratisch) / large boulders (erratic or not erratic)

subglaziale Klamm im Trogtalgrund / subglacial ravine cut into the floor of a glacial trough

Gletscherschrammen / glacier striae

V *Kerbtal* / V-shaped valley

glaziales Horn / glacial horn

I-V *Spätglaziale, neoglaziale bis historische Gletscherstände* / Late glacial, Neo-glacial to historical glacier stages (explanation in text)

Podestmoräne, Grundmoränensockel mit Terrassenstufe / pedestal moraine, pedestal ground moraine with escarpment

Grundmoräne mit großen nicht erratischen Blöcken / ground moraine with large non-erratic boulders

Felsnachbrüche an vorzeitlichen Flankenschliffen / rock crumblings on prehistoric flank polishings

Erdpyramiden / earthpyramids

Bergsturz / rock avalanche

Strudeltöpfe / pot-holes

Moränenrutschung / moraine slide

Gletscher / glacier

frühere Untersuchsgebiete / former investigation areas

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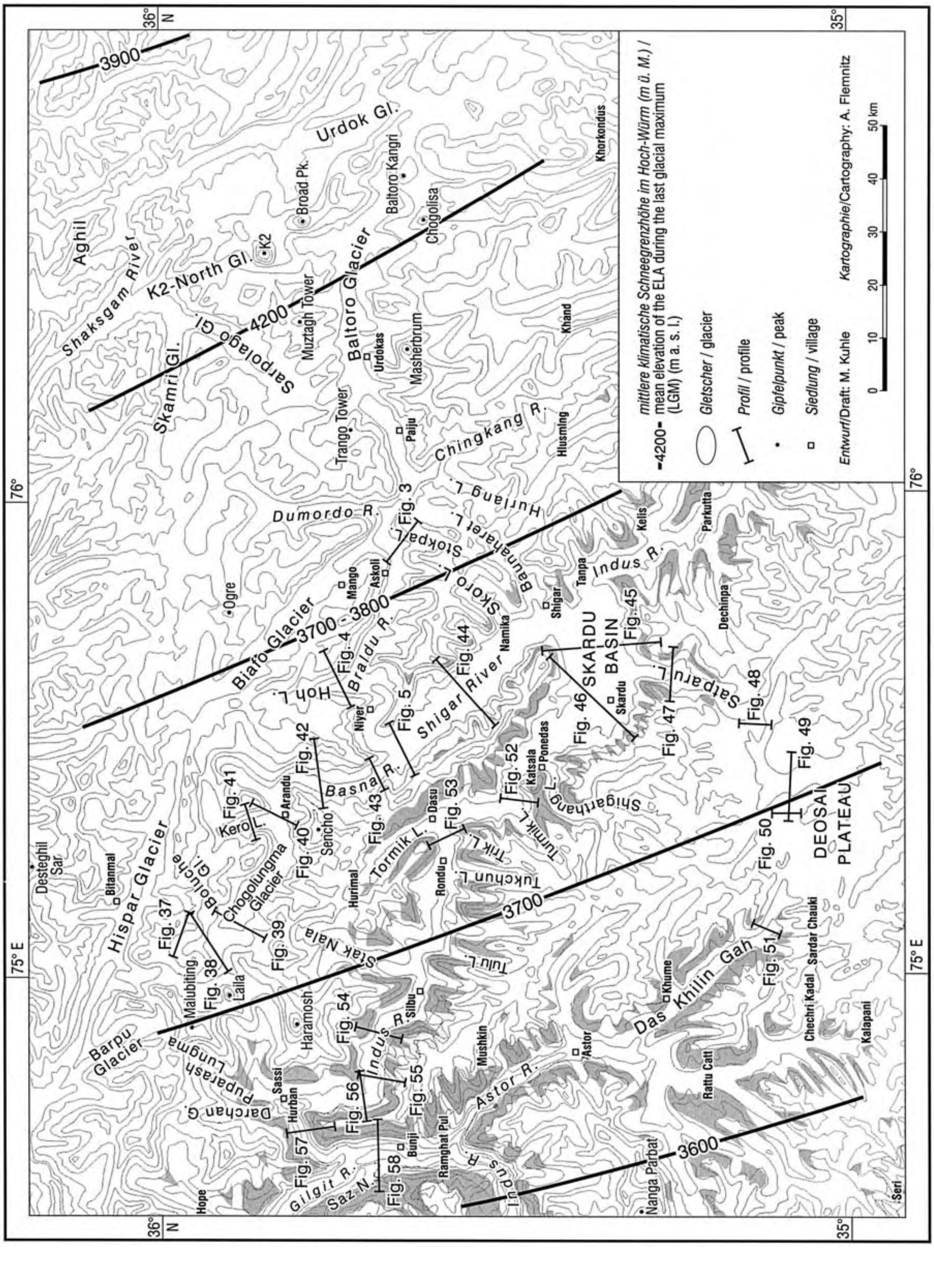
← *Figure 2.* The ice stream network of the Karakorum, West-Himalaya and West-Tibet during the Last Ice Age (LGM) 60–18 Ka BP. Draft: M.Kuhle; Scan: J.Ehlers



2/2 Localities of the glaciogeomorphological and -sedimentological valley cross-profiles during the maximum Ice Age glaciation in the Karakorum S-slope between K2 and Nanga Parbat including the Deosai plateau; see Fig. 3–5; 37–58.

spurs is preserved in the black phyllites of the orographic right valley flank between the inflows of the Savoia-Praqpa- and Khalkhal glaciers (Photo 1 centre between No. 9 and 1; Photo 11 ▲ white). Such glacially triangular-shaped slopes and truncated spurs can be observed in the entire valley system of the Upper Baltoro glacier (Figure 2/1 Nos. 1–4, 9, 10, 12–14). Further classic examples of this form of flank polishing are shown in Photo 3 (◤ between No. 6 and 10), Photo 5 (◥ between No. 3 and 4) and Photo 10 (▲ right, below of No. 2).

A glacio-geomorphologically interesting question as to the determination of the prehistoric height of the glacier level is the one about the subglacial development of sharp rock crests. Two perspectives at right angles to each other show the sharp dividing spur between the upper Baltoro- and Vigne glacier a) in top view (Photo 4 below — —) and b) in profile (Photo 10 right of No. 6; cf. Photo 9 between 7 and 12). Its sharp-jagged profile is a function of the very small confluence-angle of the two glaciers of only 45 °. At the same time it is a geomorphological indication of the gradual lowering of the prehistoric ice level. Overflowing ice makes the forms round, but on this very narrow crest this could not happen without any extractions. At a later stage the ice-free crest was undercut and still more heavily resolved into serrated pinnacles. The result of these considerations is that sharp rock features can have been absolutely overflowed by ice during the LGM, i.e., that their sharpening falls into the Late Glacial.

2.1. On the question of glacigenic or pre-glacial mountain forms in the Baltoro area

Here too, the aforementioned indication points to the nunatakker-problem. As for the K2-pyramid (Photo 2) the term is certainly appropriate because this is still at present fringed by glacier ice and must have been wrapped by ice to a greater extent and higher up during the LGM when the ice level increased by at least 1000 m. However, as nunatakker in the sense of ice-shaped glacial horns only small mountains or summits can be taken in consideration, as, for example, the Marble Peak (Photo 3 No. 9), the satellites of Gasherbrum V (Photo 9 and 10, No. 2; Photo 7 above — —) and probably the Mitre Peak (Photos 3, 5, 9, 10 No. 10). All these summits tower above the rock roundings and flank polishings created by the glacier – as e.g. triangular-shaped slopes and polish bands – by more than 100 metres. Their basal faces are relatively small and their features are the result of basal flank polishing and glacial undercutting up to approximately the highest point (Figure 2/1, No. 9, 10, 13). Compared with the K2 the Muztagh Tower takes up a mid-position (Photo 9, 10 and 12, No. 8). It towers above the geomorphologically demonstrable maximum glacier level by c. 1200 m. However, because these 1200 m are exactly the height of the narrow summit-structure on a much broader and thick-set base, this can be regarded as an indicator for glacigenic sharpening. A further clue is the lance-shaped outline of the summit- structure typical of a flow-form, with which the Muztagh Tower rose above the ice-stream network of the LGM like the conning-tower of a submarine (Photo 12 and Figure 2/1, No. 8). Its sharpening might refer not only to the LGM but also to earlier maximum glaciations in the Pleistocene. Perhaps the huge pyramid of the K2 has also profited from these polyglacial flank polishings. They, too, could have been caused by the conspicuous slope-steepening in the area of the lower 1400–1500 m of this mountain (Photo 2 below — —; Photo 3 and 8 below — — left of No. 1; Photo 5 — — right of No. 1). In contrast, Gasherbrum IV is a pure crest-summit, i.e. even in the prehistoric period it was never entirely surrounded by ice (Photo 7 No. 4). Its pyramid-shape, and especially the steep WSW-face, is due to pre-glacial, probably Tertiary, fluvial backward-erosion. The very massive, completely non-sharpened shape of Broad Peak (Photo 6 and 8, No. 3) shows the characteristics of backward-erosion only within the lower 1800 m of its structure. The three large valley source depressions are separated from each other by two glacigenically-rounded mountain spurs (Photo 6 the two ▲ further up). Above, their steep back-walls are still presently separated from the ungainly double-peak by abrupt slope-flattenings – and will certainly be so even long into the future. These three valleys which have been developed fluvially in the Tertiary, have been modified glacigenically during the Pleistocene as far as into their source depressions (Photo 6 ▽, ▽).

A classic mixed form is the Baltoro Kangri (Photo 10, No. 7). It has not experienced a sharpening like a nunatak or even a glacial horn or Norwegian tindan; it remained massive and ungainly, though it has towered above the level of the ice-stream network (— —) by only just 1000 m, i.e. similar to the Muztagh Tower. However, the flank steepenings within the basal 1400 m of its structure (Photo 10 No. 7 below — —) up to the build-up of the wall, provide evidence of the typical geomorphologic reaction to a strong glacigenic lateral erosion. It even interrupts the Tertiary fluvial source depression – modified to a cirque (hence the second name 'Golden Throne'; Photo 10 No. 7 ○) – by a 600 m-high escarpment (below ○). The different views of the Mitre Peak (Photo 5, 10 and 13, No. 10) prove this mountain to be the purest form of a glacial horn in the sense of a tindan, because it was probably sunk into the glacier up to its summit during the LGM. This is also indicated by the highest prehistoric glacier level running in the farther surroundings about 6100 m. Nevertheless, the ice level is marked at c. 5950 m asl (— —) in the photos, for this altitude alone can be safely evidenced by the glacigenic polish forms along the southern continuation of the crest of the Mitre Peak (Photo 10, the two ▲ left of No. 10). However, smoothings up to the highest point (Photo 5 No. 10) make this still higher glacier level probable. The angular wall-facette, which looks as if it was broken away by undercutting (Photo 13 between No. 10 and — —), could prove the probably already lowered, i.e. Late Glacial marked (— —) ice level. Something corresponding applies to the adjacent granite towers situated somewhat down the Baltoro glacier in the confluence area of Nuating- and Biarchedi glacier (Photo 13 — — on the left and right below No.14). They also reach an altitude of 5800 to 6000 m.

2.2 Continuation of the reconstruction of the glacier level in the Baltoro glacier valley, upper Blaldo valley (Biaho Lungpa) and Muztagh Karakorum S- and W-slope during the LGM

In the extended glacier confluence of Concordia (35°42′ N/ 76°31′30″ E) remnants of classic-glacigenic rock roundings are preserved – so for instance at the exit of the W-Gasherbrum glacier valley (Photo 7 ◥) and the Biarchedi glacier valley (Photo 15 and 13 ▲ below No. 14–11). Their state of preservation depends on the petrography. In contrast to the smooth granite flanks of the Biarchedi valley the lime and lime marl flanks on the right slopes of the upper Baltoro valley chamber are far more resolved and roughened (Photo 11 ◤ black). The remaining comparably large, uniform rock surface (◤ black) due to flank polishings during the LGM and Late Glacial is remarkable. However, it is already flanked by quite young to fresh grooves of falling rocks. On both flanks of the first orographic right-hand side valley of the Baltoro, downwards from Concordia, comparable roughenings caused by the former glacier polishing are evident (Photo 14 ◥ ◤). The uniformity of these flanks forms a classic trough profile as far as the level diagnosed as the minimum height of the prehistoric glacier level (– –).

Such a glacigenic uniformity of the valley flanks can almost be observed in the entire Baltoro glacier valley. This applies especially to the shortened perspective of the view along and upwards the valley (Photo 16). Up-valley, because then the valley flanks from both sides run together. Seen in this way, the Ice Age glacier level at 6000-6250 m asl described above can be recognized very clearly (Photo 16 – –).

2.3 Observations and reflections on the ice thickness in the Baltoro valley

In this longitudinal perspective the upper half of the Baltoro trough valley form (Figure 2/1 between No. 9, 10 and No. 8, 14) becomes perfectly obvious. The lower half is filled by the present-day Baltoro glacier (Photo 16 □). The valley flanks, plunging steeply under the glacier surface, provide indirect evidence of an important ice thickness of the Baltoro glacier. In an extrapolated downward continuation of the two lines of the flank profile (Photo 16 ▲ black and white) under the Baltoro glacier ice, a glacier thickness of over 1000 m becomes probable. This applies in consideration of the glacigenic trough- i.e. U-shaped valley form (Figure 2/1 between No. 9, 10 and No. 8, 14) which must necessarily be assumed. This impression of a substantial ice thickness becomes especially clear in Concordia (Photo 5). Here, the Baltoro glacier attains its most important extension of c. 4.8–5 km in the NE- to SW diagonal between the basal slopes of Broad Peak and Mitre Peak (Photo 5 and 13 between No. 3 and 10). Downward of this confluence the Baltoro parent stream still remains c. 2.8 km wide (Photo 13 below No. 11, middleground). This glacier width continues over a distance of c. 18 km (Photo 17). Its tongue end is still 1 km broad (Photo 48 and 50). An extension like this, as well as the valley bottom height about 3350 m asl at the glacier end, document the above-mentioned glacier thickness of over 1000 m at Concordia – and probably also several kilometres further down (Photo 16). At Concordia, 28 km away from the glacier terminal, the ice level lies at c. 4550 m (Photo 5 □ foreground). The valley bottom declines by 270 m over the same distance of 28 km down-valley of the Baltoro glacier end as far as the inflow of the Biafo glacier at c. 3080 m (locality Korophon). This means a corresponding – now extrapolated up-valley as far as Concordia – height of the valley bottom at c. 3520 m. Then, the valley bottom would lie a good 1000 m below the present-day glacier surface. Though up-valley the regular increase of the gradient curve must be taken in account, this can be considered subglacially as compensated by the increasing thickness of ground moraine and detritus under the present-day Baltoro glacier from Concordia down to its tongue end. Thus, the glacier end at 3350 m asl comes to lie relatively highly against Concordia.

Owing to this, in the area of Concordia the Ice Age glacier level ran at 6000-6250 m during the LGM (Photo 13 – –), c. 2400–2600 m above the valley bottom. By analogy, a mean ice thickness of 2500 m points to a prehistoric ice stream network with the dimensions of the Alpine ice stream network during the LGM (see below).

2.4. The present-day extension of the Baltoro glacier system as a misleading indication of the Ice Age glaciation of this area

In the area around Concordia and 12 km down, as far as the orographic right influx of the Biange ice stream (Photo 17 □ below No. 8) and still further down as far as the orographic left influx of the Mandu glacier (Photo 22 and 25 below No. 5 and No. 27; locality: Figure 2/1 from the cross-profile between No. 9 and No. 10 up to the cross-profile between No. 17 and No. 5) there exists a striking disproportion between the vast kilometer-wide glacier area and the connected small glacier feeding areas above the snow-line. Owing to this, the present-day ice thickness of c. 1000 m and the resulting extended glacier area are the inheritance of the Late Glacial. At any rate, the present-day feeding areas would be unable to build-up the Baltoro glacier system. They are only large enough to maintain it, since the present-day ice level guarantees sufficiently large areas above the ELA. As far as this is concerned the present-day extension of the Baltoro glacier fosters – due to its Late Glacial heritage – the interpretation of a 1500 m-thicker LGM-ice stream network introduced here. Thus, the reason for the extension of the Baltoro glacier system is the inherited ice thickness of a former climate but not the proximity to the present-day climate.

Misleading in the same way, i.e. only seemingly more plausible, is the glacigenic shaping of the flanks above the (lowered) present-day glacier level. The plausibility follows the principle of actualism: where glacier ice is still today attached, it has also been attached higher up in prehistoric times. So, it seems as if the ice were still attached. This applies even if and when important reshapings have taken place.

2.5 Continuation of the reconstruction of the LGM-glacier level in the Baltoro area

Reshapings of this type can be observed between Concordia and the Biange glacier and also several kilometers further down- valley on the orographic right side. Here, on the valley flanks of metamorphic sedimentary rock, rills have been developed crossing the horizontal glacigenic flank abrasion (Photo 17 ◓ between No. 11 and 4) right-angled, i.e. down-slopes, and reshaping it (Photo 17 ⇩; locality: Fig. 2/1 No. 9). The reason for this is the temporary snow meltwater flowing down the slopes. Photo 17 shows a corresponding weather condition with a summery fresh-snow-cover (○) which, by thawing down during a radiation weather condition, causes rill rinsing on the slopes. In a downward direction the depth of the rills decreases, because the glacier has still recently clung to the flanks. In the middle course it is the deepest. Here, the longer glacier-free time and thus rill-development in combination with the most substantial amount of merging water – increasing down-slopes – overlap each other to the summarizing effect of the deepest linear erosion. Upwards, the rills break off in the direction of flat cirque floors. During the LGM the glacier level lay above these cirques (— — above ○). Thus, their shaping took place at a time, when this area was released by the ice because of the dropping of the glacier level, the ELA, however, still ran c. 600–700 m lower than at present. This applies to the late Late Glacial, more exactly, to the Sirkung Stage IV (older than 12,780 YBP after Kuhle 1994b Tab 3 p 260).

The corresponding High Glacial (LGM) ice level, i.e. a corresponding surface level of the glacier between 6000-6200 m, is geomorphologically documented in the Biange glacier valley (Photo 18 — — on the very left and in the right half of the panorama as well as in Photo 19 — —). The structure-dependent reshaping of the prehistoric flank abrasions (◓) which has set in after the Late Glacial (Stage I-IV), i.e. the beginning of which was c. 13,000 to 17,000 years ago, is at an especially advanced stage in the sedimentary rocks and phyllites layered diagonally or vertically in many places. Accordingly, fresh debris cones of crumblings which bury older moraine cores (Photo 18 ▼) occur frequently. They start at the exit of structure-dependent wall gorges. Surface-forming prehistoric moraine deposits are therefore rare. They can be recognized by their fresh cutting forms derived by the snow meltwater (Photo 18 ■ black). Present-day end moraines are built up by a small hanging glacier (■ white) which postglacially polished itself (○) into the orographic left glacigenically abraded flank of the Biange valley (Photo 18 ◓ on the very right). Because its cirque (○) is greater than its moraine volume (■ white), a hanging glacier erosion which has already taken place during the pre-LGM, i.e. in the last interglacial period, is probable. At the same time all features of glacigenic flank abrasion in the Biange valley (Photo 18 ◓) show the characteristics of remnants of mountain spurs polished back to glacigenic triangle-faces (Figure 2/1 No. 8, 9, 13). The Muztagh Tower has been formed to a classic glacial horn by the prehistoric ice stream network (Photo 19). The ice level on its flanks has reached as far as over 6000 m (— —) and its outline, polished lance-like, proves an LGM ice-flow-direction which ought to have followed its longitudinal axis, namely from SE to NW (Figure 2/1 No. 8). This documents that the more important Baltoro ice flowed over to the Sarpo Laggo-Muztagh valley ice. There, it reached the extended Shaksgam parent glacier N of the Karakorum main crest and then flowed rather directly down to the Tarim Basin situated only 140 km away. The Ice Age (LGM) Sarpo Laggo-, Muztagh- and Shaksgam ice stream network has been reconstructed by the author in detail elsewhere (Kuhle 1994b). There, too, are numerous indications of an LGM-ice level running large-scale about 6000 m asl. The ice spill-over took place between the Muztagh Tower and the 6930 m-Peak (No.15, Skilbrum-W-satellite) across the 5500 m-high Moni La and the 5869 m-high Ste Ste Saddle, which made up two transfluence passes (Figure 2/1 between No. 8 and 15) to the Sarpo Laggo tributary stream (Photo 18, LGM-ice level — — between No. 8 and 15). This can be evidenced by the flank abrasion reaching a height of 6200-6300 m on the W-crest of the 6930 m-Peak in the region of the present-day Younghusband glacier (— — below No. 15). On the orographic left flank of the Baltoro glacier valley in the area of the 6344 m-Peak (Biarchedi-N-satellite) (Photo 18 No. 16) flank polishings are rather sparse (Photo 20 ◓). Firstly this is caused by the coarse structure of the bedrock granite. At the places where it is freshly broken away by present-day glacier undercutting (Photo 18 ⇩), remnants of polishing are already completely lacking. Secondly, the N-exposed flank is so heavily glaciated by itself (Photo 20 ▽) that the down-slope glacier denudation and -abrasion has removed the almost horizontal LGM-polishing. Besides the prehistoric polishings and glacigenic abrasions (Photo 17 ◓ on the very left and on the very right) preserved only on rock-ribs and -spurs between the present-day hanging glaciers, more or less reshaped material of ground moraines occurs in these positions which are protected against the present-day wall- and slope erosion (Photo 20 ■ black, 22 ■ white). However, in extreme shadow positions of the glacier flow, where a lot of the prehistoric moraine material has been stripped off by the glacier ground, i.e. ice margin, ground moraine has been preserved, too (Photo 20 ■ white).

On the flanks of the Masherbrum Group situated near the Yermandu tributary glacier which flows into the Baltoro from the S, an extreme case of postglacial denudative and glacigenic reworking of the flanks which partly rise twice as high (see above) above the present-day glacier surface (Photo 20 and 22 No. 5) has been observed. This concerns the E-, NE- and N-flank sections of the Masherbrum main- (7821 m) and Masherbrum-E- (7163 m) summits (Photo 21 No. 5 and No. 33). On the wall spurs (sunny parts of the wall in Photo 21) the denudation by avalanches is extremely effective; the same applies to the locally merging ice run-off in the biconcave wall cauldrons (Photo 21, shaded parts of the wall). Accordingly, on these slopes of the Yermandu glacier valley no clear indicator of the glacier level is preserved as to the prehistoric ice filling of the relief.

If one follows the orographic right glacigenic flank abrasion down the Baltoro valley, its dependence on the rock becomes obvious. Whilst the appearance of the abraded valley slopes with their postglacial rill-features (Photo 17 ▲ between No. 4 and 11) does not change in the course of the 11 km from Concordia to the Lhungka tributary valley, the flanks further down become steeper (Photo 23 ▲). This geomorphological transition takes place along the 3.5 km-distance between the Lhungka valley and down the Muztagh glacier valley (Photo 22 between ■ black and ▲ white). Massive-crystalline bedrock – here granite – occurs already in the area of the inflowing Lhungka valley. It preserves undulating, compact, glacigenic abrasion features (Photo 24 ▲). The rock slope, however, is still flat enough to allow the preservation of prehistoric remnants of ground moraine even in high positions (Figure 2 No. 31; Photo 24 ■; 22 ■ black). This morphological transition region (Photo 25 on the right of No. 18; Photo 28 ▲ below No. 31) passes into the steep-flanked valley section (Photo 25 left of No. 18; Photo 28 left of □ on the very right) at the place where the Baltoro valley suddenly becomes somewhat narrower and at the same time branches off into an orographic left-bend, thus turning from a W- to a SW-direction (Figure 2/1 between No. 17 and 28; Photo 22 on the right of the centre, background). Due to the LGM-ice (Photo 23 – –) rising as far as c. 6000 m, a glacigenically undercut slope has been created here. Accordingly, the c. 2500 m-thick component of the prehistoric ice stream-network, which followed the Baltoro valley up to here in a down-valley W-direction, has been diverted by the right valley flank to the SW. This increase in pressure and thus flank abrasion, resulting from the change of the direction, has caused the steepening (Photo 26 ▲ left of No. 18) into a trough wall. The uniformity and characteristically large-scale rounding (Photo 23 ▲) of this valley flank up to a good 1000 m above the present-day glacier surface (Photo 23 □) reminds one of a Scandinavian fjord-flank.

2.6 Insertion on the glacigenic shaping of granite

Massive-crystalline rock outcrops here as well as in Scandinavia. Sickle-shaped break-offs (Photo 23 ↙) on smoothed rock faces are characteristic of this rock. Accordingly, this type of desquamation takes place approx. concordantly to the surface (Photo 30). Owing to this, a rock-specific structure is probable here, which can only be selectively glacigenically rounded, namely at all places where this granite would also develop roundings by way of any other type of erosion. This means at the same time, that where the granite forms endogenously flat-edged features, the glacier polishing which creates roundings on other rocks, here only develops these flat-edged structures, reworking them in a way. Other variations of break-offs in the granite are determined by the horizontal release joints of the stratification surfaces (Photo 26 ↙).

2.7 Continuation of the reconstruction of the LGM-glacier level in the middle to lower Baltoro glacier area

On the right valley slopes discussed here, the flank polishing of a higher as well as a lower prehistoric glacier level is especially striking. It can clearly be recognized between the inflows of the Biange- and Lhungka glacier valley (Photo 22 the three ▲ on the very right). Here, the younger glacier abrasion is diagnosable by the lightly preserved granite as far as 600 m above the present-day glacier surface (□) (up to at most the level of the second ▲ from the right). This ice level has to be classified as belonging to the youngest Late Glacial, i.e. to the Sirkung Stage IV (older than 12,870 YBP). Together with the corresponding fresh flank abrasion it is also preserved on the orographic left side at the exit of the Muztagh glacier valley (Photo 25 ▲ diagonally right below No. 18).

The rule is: the higher the prehistoric ice level has run, the older it is and the more heavily its geomorphological remains have been reworked and blurred. Accordingly, the indications of the highest level are only preserved in a punctiform manner. They are the basis of an interpolation of that prehistoric ice level between their localities. Such a locality is found in the confluence area between Muztagh- and Chagaran glacier valley on the S-spur of the 6224 m-summit (No. 18). There, the rocks have been abraded by the glacier ice as far as a height of at least 6050 m asl (Photo 27 ▲ white). Indications of high levels exist also in the W-adjacent parallel valleys as e.g. the Biale- and Dunge glacier valley (Photo 28 ▲ white and black right of No. 17; Photo 29 – –). In the Dunge glacier valley the Ice Age-height of the glacier level is documented by the breaks in slope from the flat rock ramps to the steep wall sections of the summit-crests of the Kruksum and the 6544 m-Peak (No. 25 and 19) at the valley exit (Photo 29 – –).

2.8 The development of wall gorges and glacigenically shaped triangular faces in the Baltoro trough valley

The better preservation of the lower-lying polishing faces corresponds with the decreasing size of the wall gorges in a downward direction (Photo 26 ⇩ on the left; 28 ⇩). As far as the geomorphological principle is concerned, the extension of the wall gorges increases – just as every other valley extension – with the increase of the connected catchment area. However, since it is the drop of the ice level which actually renders the downward development of wall gorges possible here, this development is delayed and less advanced. This is the cause of the development of glacigenic triangle-slopes as they are typical of the steep flanks of trough valleys (Photo 23 ▲; 26 ▲ black; 28 ▲ black on the very right and left; Figure 2/1 No. 31, 32, 17, 23, 16, 27).

2.9 The reconstruction of the LGM-glacier level in the area of the lower Baltoro glacier

On the orographic left valley flank on both sides of the influx of the Mandu glacier, the destruction of the prehistoric flank abrasion and the resulting rounding through undercutting of the present-day Baltoro glacier is especially obvious (Photo 25 ⇩). The reason for this is a dependence on the bedding structure of the crystalline rock which on the Urdokas Peak (No. 27) declines coarsely-bedded at c. 40° (40/355) to the N (cf. Photo 32 for details). The lateral erosion undercuts and thus reaches the ac- and bc-jointing which then develops the inevitably vertical to overhanging edges of the break-offs (Figure 2/1 between No. 14 and 27).

Up to the end of the Baltoro glacier the ororographic left flank shows a wealth of indicators of the prehistoric height of the ice level confirming those of the orographic right valley side. A reference to a further Late Glacial, i.e. an already lowered ice level (perhaps during the early-Late Glacial Ghasa Stage I, after Kuhle 1982a and 1994b) provides the polish cavetto of an Urdokas Peak-N-satellite (Photo 32 ⇩) c. 1500 m above the present-day Baltoro glacier surface. Granite mountains, completely rounded by the ice, are also preserved. They must be approached as glacially streamlined hills (Photo 33; Figure 2/1 No. 27). The form of this side of the Baltoro valley is marked by the glacial-interglacial change between a dominant flank abrasion at an ice level (– –) lying at least 1800 m higher than that of today and the cross-running glacial erosion of small tributary glaciers as it takes place at present (Photo 34; 35; 36 – – on the left; 37; 38).

The local influence of the rock structure on the interglacial reshaping is demonstrated by the orographic left flank of the Liligo glacier valley (Photo 38 below No. 28). The E-wall of the Liligo Peak consists of granitic edges of the banking structure, breaking off and eroding back under the control of bc-clefts (Figure 2/1 No. 28). The voluminous debris cones (▼), which are deposited at the wall foot but not tranported down by a hanging ice body, point to a backward erosion after the Late Glacial deglaciation. Ice Age glacigenic flank abrasions and -roundings have been preserved nowhere – not even first signs of them.

In comparison, the flank smoothing of the Baltoro valley further up (Photo 37 ▲ on the left, Figure 2/1 close to No. 27) and down – between the Liligo glacier and the Baltoro glacier end, especially in the area of the locality Liligo – is perfectly preserved (Photo 50 ▲). The prehistoric glacier abrasion has worked nearly concordantly to the surface of the rock banking, which since the deglaciation has at most splintered-off and truncated but has not broken away or deformed (Photo 48 ▲ right). At the base of these rock slopes (Photo 41 ◤ white on the right) even ground moraine, deposited in layers (coarsely-stratified), has been preserved in a thickness of up to 30 m (Photo 41 ■ black; Photo 48 ■ small). Its overlay has kept the rock smooth over millennia (Photo 41 to the right of ■ black; Figure 2/1 next to No. 28). Corresponding conditions can be evidenced on the orographic right side down-valley of the Baltoro glacier tongue at places, where recently fresh-preserved glacigenic rock polishings emerge from its ground moraine cover (Figure 2/1 between No. 11 and 30; Photo 49 ◥ black below ■ white; 50 below ■).

The steepness and relief energy of the glacigenically-shaped orographic right valley flank along the lower 12 km of the Baltoro glacier are remarkable. Between Mt. Biale (No. 17, 6729 or 6422 m) with its S-crest in the E and the Paiju Peak (No. 11, c. 6600 m) in the W, wall-inclines of 60-85° occur at only slightly-staggered vertical distances of 2500 m (Photo 28 left half of the panorama; 45; 46). The relief is striking, though the valley bottom is filled-up by the Baltoro glacier system several hundred meters high. During the LGM, however, this extreme relief was filled with ice and – with the exception of towers and crests – levelled as far as c. 5800-6000 m asl (– –). Evidence of this prehistoric ice level (– –) is provided by glacigenic abrasion forms, as e.g. convex-rounded rock slopes (Photo 28; 36; 39; 40; 42; 44; 45; 46; 49 ▲) and partly well-rounded (Photo 33; 36 below – – on the left; 45 below – – on the right of No. 26) or at least truncated rock ridges and summits (Photo 28 below – – on the right of No. 17; Photo 30 below – –). Additionally, ice-overflowed and structure-dependent 'frayed' summit forms are indicators of the ice-level (Photo 46 below – –).

With all necessary restrictions as to the indicator-value of the glacier abrasion of smooth granite slopes – which are rather structure-dependent and develop quite similarly under different conditions of erosion (see above) – there are numerous possibilities of a clear diagnosis on the orographic right flank under discussion (Photo 36 ▲, right half of the panorama; 39 ▲). Here, rock roundings are combined with glacigenically triangular-shaped slopes and even covered with ground moraine at many places (Figure 2/1 below No. 17 and 23; Photo 39 ■; 42 and 45 ■). Further evidence for the reconstruction of the level of the ice stream network can be provided by the flanks of the side valleys, as e.g. the Trango valley (Photo 39 on the right of No. 26). Here, glacigenic abrasions and flank polishings are obvious over large parts up to over 1000 m above the present-day Trango glacier surface (Figure 2/1 below No. 24 and 26; Photo 40 ▲). Even a small-scale ice transfluence – still during the Late Glacial – to the NE-parallel Dunge glacier between Trango I (No. 23) and the 5753 m-Peak is evident due to unambiguous forms of roches moutonnées (Photo 46 ▲ black). A further transfluence existed between Trango I (No. 23) and Trango Tower (No. 24) (Photo 31 – –; 45 between No. 24 and 23; 46 left of No. 24; 47 on the right of No. 23) during the High Glacial. Thus it is proved, that the Trango Tower was fully enclosed by the ice. The altitude of the glacier level is documented by a lateral erosion undercutting the flanks of the Trango Tower (Photo 31 and 47), so that this summit form can be referred to as being a typical glacial horn (Figure 2/1 No. 24). Especially the lower parts of the shaft of this tower-like horn show roundings still today (Photo 45 No. 24). They go back to the work of the younger but less thick Late Glacial glaciers of the Stages I and II (i.e., Ghasa- and Taglung Stages after Kuhle 1982a) which did not reach so far up. A scarcely less slender but even 278 m-higher glacial horn is the Uli Biaho reaching 6417 m (Figure 2/1 No. 26; Photo 28; 36; 39). Also the Trango Cathedral (No. 23) and the comparably more flattened, but still bold mountain-feature of the Paiju (No. 11) – the highest of all – can be classified as key forms of glacial horns (Figure 2/1; Photo 28, 30, 31, 39, 44, 45, 46). Glacigenic tower-forms of this type can also be observed in the Patagonian Andes, which are still today glaciated and accordingly have been glaciated much more strongly in prehistoric times. Remarkable examples in the area of the Hielo Continental are the granite superstructures of the Fiz Roy and Aiguille Pointcenot. However, they attain only half of the absolute height of these Karakorum Towers.

On the orographic right side of the Baltoro glacier the basal underpolishings are preserved in an exceptional clearness. They have been caused by younger glacier ice the level of which was lowered during the Late- to Neoglacial. On the S-ridge of the 5340 m- spur of the Uli Biaho Group, for instance, four stages of polishing, i.e. abrasion, can be differentiated. 1.) the abrasion as far as beyond the rounded rock summit (Photo 45 ▬ ▬ on the right of No. 26); 2.) the obviously smoother flank polishing below (▲ below ▬ ▬; Photo 44 ▲ white); 3.) the steeper and lighter polish-facet with first breakages, set off against the smooth rock ridge in a downward direction (Photo 44 ↓ on the right; 45 ↓ (short) below ▬ ▬ on the right of No. 26 and ↓ (long); below Figure 2/1 No. 26); and 4.) the also facet-like underpolishing of 3.) owing to the inflowing Uli Biaho tributary glacier (↓ (long) left of (short)) (cf. under different light-conditions: Photo 42).

2.10 Glaciogeomorphology and reconstruction of the LGM glacier- level in the Biaho Lungpa (Blaldo valley) from the terminus of the Baltoro glacier up to the Dumordo (also Panmah-) valley

A 13 km-long valley chamber (Figure 2/1 between No. 11 and 30) follows the Baltoro glacier as far as the orographic left inflow of the Chingkang valley. Its left flank is adjacent to the 5800 m-Peak (No. 30). The orographic right side is made up by the Paiju Peak-S-flank with a maximum height of a good 3000 m (No. 11). The valley bottom, up-valley formed by the glacier (Photo 50 □), here shifts into a glacier mouth gravel floor up to c. 1 km in width, consisting of outwashed and removed moraine material (Photo 50 and 42 ○).

The prehistoric ground moraine cover on the glacier-polished rock of the orographic right flank in the more distant environs of the Paiju locality (Photo 49 and 50 ■ white) has already been introduced (Chapter 2.9). These observations are to be completed by a ground moraine cover with a light matrix on dark bedrock phyllite, containing coarse erratic granite boulders (Photo 48 ■ black; Figure 2/1 between No. 11 and 30). Remnants of ground moraine can be found up to at least 900 m above the valley bottom (○) (□ white below No. 11). The dark colour of the ground moraine surface is the result of a ferromanganese incrustation deriving from the crystallizing-out of ascendent solutions owing to a high potential evaporation. On the basis of an already several millimetres-thick crust-development, a capillary ascent - which was persistent at least during the historical time – becomes the indicator of a longer-term semi-arid climate. Whether this climate already existed in the Holocene is unkown.

Down-valley the two flanks have been glacigenically abraded (the two first ▲ from the left). The orographic right flank shows rock polishings reaching up c. 1000 m from the valley bottom (polished rock-faces glittering in the sun next to the second ◤ from the left; Figure 2/1 between No. 11 and 30). The orographic left flank abrasion on the outcropping edges of the stratum (◥ on the very left) has splintered-off over large parts (Photo 51 ▲). It reaches up to 5400 m asl (▬ ▬), i.e. 2000 m above the gravel floor (○). Large-scale observations of the arrangements of the positions testified to an LGM-glacier level that lay even 200-300 m higher - perhaps 400 m (Photo 48 ▬ ▬). This indicates an LGM ice-thickness of 2200-2400 m, not including the thickness of the gravel under the river bed (Photo 51, 52, 53 ○). However, it cannot be ruled out that the Ice Age Biaho Lungpa (also Blaldu- or Braldu-) glacier flowed on a ground moraine pedestal which was decametres- to several hundred metres-thick. To this points the enormous ground moraine ledge on the orographic left valley slope up-valley of the locality of Bardumal (Photo 51 ■ white; Figure 2/1 above No. 30). But also on the opposite valley slope 60-110 m-thick deposits of ground moraine are preserved at a height of 300-400 m above the valley bottom (Photo 52 ■ on the left; Figure 2/1 on the left below No. 11). On this northern valley side a very important part of the moraines has been redeposited as huge mudflow fans since the deglaciation (Photo 48 ▽ white; 51 and 52 ▽ black on the left; Figure 2/1 on the left below No. 11). At many places a sub-deposit of the present-day river-gravel cover (glacier mouth gravel floor) (Photo 53 ○) by ground moraine is suggested by the plunge of the ground moraines (■) beneath the gravel floor at the foot of the slopes (between ■ on the left and ○). The inflow of the Chingkang valley shows the characteristic glacial-interglacial sequence of forms with glacigenic flank polishing on the upper slope (Photo 53 ▲ black and white; Figure 2/1 No. 30 to the right of No. 35) and interglacial fluvial erosion, undercutting and reshaping the lower slope (below white ▲). This linear cut is so narrow, that it can be considered as being a ravine (Photo 54 ⇩). Probably this ravine has already come into being subglacially during the Late Glacial. At that time a quite important subglacial, i.e. hydrostatically confined, meltwater discharge must have taken place. Owing to its caviation corrasion ability it was especially capable of deep erosion, i.e. the development of ravines. However, the upper edge of the ravine has undoubtedly broken away since the deglaciation, i.e. it has not been polished syngenetically (Photo 53 a little above ▲ white). 500 m up the ravine, ground moraine from the Late Glacial valley bottom has been pressed into it (Photo 54 ■ on the right; Figure 2/1 on the right of No. 35). The steep walls of the ravine in the rock above the ground moraine filling (above ■ on the right) provide evidence of its earlier existence and thus subglacial development (Figure 2/1 left of No. 30). In consequence, its further subglacial forming is very probable (see above).

The LGM-ice levels of the main valley (Biaho Lungpa, also Blaldo or Braldu valley) in the confluence area of the Chingkang valley (Photo 53 and 55 ▬ ▬), which have been reconstructed about 5400-5800 m asl, confirm the complete ice-filling of the Chingkang valley as far as the Double-Peak-massif (No. 34). Though, owing to this, the rock slopes have been glacigenically polished and abraded as far as their culminations, it is almost a V-shaped valley, i.e. a V-shaped gorge. Because the flank abrasion was only capable of polishing a slightly concave bow into the slope profiles, the glacigenic valley type of a 'trough-shaped gorge' or a 'gorge-like trough' has been realized here (Photo 55 the valley cross-profile below No. 34; Figure 2/1 on the left below No. 30) (Kuhle 1982a, 1983a). Quite similar, but formed like a hanging valley, appears the Hurlang Lungma side valley (Photo 55 ⇩). It is also a glacigenically V-shaped valley with a ravine in the area of the talweg (Figure 2/1 on the right of No. 35), which has dissected the c. 180–200 m-high confluence step of the side valley (⇩). The erosion of the ravine profited – and still profits – from the Ice Age moraine material which is transported through. This is the material from which the large mudflow fan has been made up at the valley exit (▽; Figure 2/1 on the left above No. 35).

Down-valley of the Hurlang Lungma influx, glacigenic flank abrasions are obvious on the orographic left-hand side. Only in the upper parts of the slope (Photo 55 on the right and left of ▲) they are dissected by fluvial slope gullies. The reason for this is a prehistorically earlier upper slope, which was already free of ice during the late Late Glacial (Sirkung Stage IV). In contrast to the gullies shown here, those, which in a purely fluvial environment normally develop by backward erosion, are always deeper down-slope than up-slope. The glacigenic tendency towards flank smoothing and the leveling of all unevennesses on the valley flank we are talking about, not only becomes obvious because of the abrasion of protruding rocks (▲), but also through flank niches filled with thick groundmoraine material, which has been pressed into them (■) (Figure 2/1 No. 35).

In the course of the Biaho Lungpa, between the influx of the Hurlang Lungma- and Dumordo valleys, on the orographic right side an at least 20 m-thick ground moraine cover (Photo 56 ■, foreground) is undercut by the gravel floor (○), i.e. the high-water-bed. Near the surface, this has naturally been modified by flushing processes and mudflow activities since the deglaciation. Here, as in many other places, too, the most important ground moraine deposits can be observed in the transition of the trough flanks to the trough ground (Photo 56 below ■ in the background) over a rather large valley section in a longitudinal direction. A representative ground moraine deposit of this type stretches along the foot of the slope around the corner from the Biaho Lungpa into the Dumordo valley (Photo 57 ■ on the left).

During the 27 km-long course of the Biaho Lungpa (valley), from the tongue of the Baltoro glacier up to the confluence with the Dumordo valley and still further down as far as the tongue of the Biafo glacier (locality of Korophon), this main valley is formed like a box-shaped trough valley (Figure 2/1 between No. 11 and 28 up to between No. 36 and 35). Its box-form results from the gravel floor, which for the most part is several hundred metres wide (Photo 56 and 57 ○). The especially significant characteristics of a glacially eroded landscape, as for instance the concave profile lines in the bedrock, can be observed in the phyllites in the Dumordo (Panmah) valley confluence. So, the spur at the exit of the Bakhor Das Lungma has been polished in a classic manner in a concavely-steepened bow up to its culmination (Photo 57 on the right below — — on the left). Fine polishings with glacial striae, however, have splintered out and might only be met at places where they are still protected by a ground moraine cover (Photo 57 ■ on the right). A just as classic example is the upper limit of the glacial erosion which has been in this valley chamber, attaining c. 5650 m (Photo 57 — —). Its convexly abraded rock bulge is set off in a sharp concave bend from the 5810 m-high summit of the Bakhor Das (No. 35), which has the form of a glacial horn (Figure 2/1 No. 35). This polish cavetto (below No. 35) proves a valley glacier body rising 2500 m above the gravel floor (○). An LGM glacier level extending as far as over the summit-level of the Bakhor Das can neither be excluded nor proved. Owing to this, the provable ice thickness attained 2500 m plus the thickness of the gravel floor. At a c. 1.5 km-wide gravel floor – measured in the diagonal of the valley confluence (Photo 57) – and a 35 to 45°-steepness of the valley flanks which plunge under the valley floor, the rock ground might lie 500–700 m lower. An elevation of the subglacial valley ground by ground moraine is probable, but its thickness is unkown. Under consideration of these factors, which can only be very vaguely estimated, an Ice Age glacier thickness of 2600–2900 must be assumed.

Regarding the change of a trough- to a box-shaped trough valley (glacial trough with gravel bottom) the Dumordo valley presents a different picture, because even 10 km downwards of the Panmah glacier tongue end it is still a trough valley (Figure 2/1 on the right above No. 36; Photo 63). Only after these 10 km has the gravel floor, dammed back from the Biaho Lungpa, been established in the Dumordo valley (Photo 60 and 61 ○; Figure 2/1 on the right below No. 36). The reason for this is that the Dumordo valley, being a real side valley, hangs above the Biaho main valley – perhaps even several hundred metres (see above). The bottom of the Choricho W-valley (Photo 62; Photo 61 on the right above ■ white) hangs c. 1000 m above the gravel floor of the Dumordo valley (Photo 61 ○) on the orographic left side. It is a short trough (Figure 2/1 on the left of No. 11). Though this side valley has such a high-hanging bottom and a length, which at 10 km is rather short, its flanks have been glacigenically abraded up to at least 5500 m asl (Photo 62 ▲). The backflow of the ice from the main valley, necessary for this, provides evidence of a complete and at the same time large-scale ice filling of the valley system up to at least 5500 m asl (— —). This ice level is only 150 m lower than that proved on the Bakhor Das, which is situated 15 km away (see above). In this a reciprocal confirmation is to be seen. The development of that confluence-step, covering c. 1000 m (Photo 61 between ■ white and ○), can at least partly be traced back to the more heavily glacigenic abrasion of the ground in a valley of a higher order. Here, the ice was not only 1000 m thicker, but, owing to the diversely ramified and over 100 km-extended catchment area of the Panmah glacier, has moreover flowed much faster. The steep confluence step down to the Dumordo valley has been dissected by two ravines, i.e. gorge-like cuts of a torrent (Photo 60 ⇩;

Figure 2/1 left of No. 11). This – currently backward – linear erosion has been developed during the Late Glacial, when the orographic ELA (snow line) ran about 4200–4400 m asl. At that time it took place subglacially and, accordingly, not yet backwards. Linear meltwater erosion has also occurred in the Dumordo valley. Evidence of this is provided by the narrow cut in the trough ground (Photo 63 below ▽). Here, too, the convexly abraded upper edge of this small cut-profile (△◁) proves the simultaneous existence of the prehistoric glacier in the overlying bed (Figure 2/1 on the right above No. 36). A further strong indicator of a maximum prehistoric ice level at at least 5600–5800 m asl can be observed on the SE-spur of the Bullah (No. 36, 6294 m), where the prehistoric polishing of the outcropping edges of the strata has formed a glacigenically rounded triangular-shaped slope (Photo 56 ▲ below No. 36; Figure 2/1 No. 36). The ground moraine covers reaching up at least 550 m on this abrasion area (Photo 56 ■ black in the background; Figure 2/1 No. 36) have already been referred to. The polishing of the outcropping edges of the strata has been developed on metamorphic sedimentary rocks as e.g. quartzites and hornfels (Photo 58 ▲ in the foreground; Photo 59 ▲; 61 ▲ foreground). Whilst on this orographic right side of the Dumordo- and Biaho valley the postglacial breakages, i.e. those which took place since the deglaciation, occurred relatively sporadically (e.g. Photo 61 below ▲ white on the left; Figure 2/1 on the right above No. 36), on the opposite side of the Dumordo valley they have filled-up postglacial rock gullies with debris (Photo 58 above ▽; Figure 2/1 between No. 11 and 35). A very fresh boulder, for example, which came down the track of this gully and sprang onto the recent gravel floor (Photo 59 ▼), weighs 20 tons. At the place, where the present-day outer bank of the Dumordo river undercuts the abraded rock slopes excavating a cavetto (Photo 64 left of ⇩; Figure 2/1 to the right of No. 36), post-Late-Glacial to recent breakages do also occur. In contrast, huge loose rock, as e.g. decametres-thick ground moraine, reacts to a fluvial undercutting like this through much more large-scale mass movements, as e.g. slides. So, for instance, the reason for the downthrow (Photo 61 ↓ ↓) of the mudflow fans, consisting of moraine material (▽) at the orographic left slope foot of the Dumordo valley, is the undercutting by the high water bed (○).

2.11 The Ice Age (LGM) altitude of the glacier level and the glacier thickness in the upper Baltoro-Biaho Lungpa (Blaldo- or Braldo-) valley system (Muztagh Karakorum-SW-slope) – a summary of Chapter 2. to 2.10

As an essence of the detailed glacier reconstruction, a prehistoric (LGM) ice thickness of 2600–2900 m can be established in the confluence area of the Biaho Lungpa (Blaldo- or Braldo-) and the Dumordo (Panmah-) valley, which is the lowest valley area of the area investigated so far. This means, that an ice thickness of 2400-2600 m (2700–2900 m at maximum; see Photo 5) at Concordia (Chapter 2.3) and a verifiable minimum LGM-ice-level about at least 6200 m, points to an at most 300 m greater ice thickness and a decrease of the ice level as far as at least 5650 m asl. It is evident, that at a lowering of the ice level by 550 m at approximately the same glacier thickness, i.e. an only slight increase (even 300 m are not much in the face of 2500 m ice thicknesses at horizontal distances of 60 km), the ice discharge was equivalent to the incline of the valleys. To put it another way: the valley incline corresponded quite well with the incline of the ice surface in these areas with glacier surfaces lying 1.5 to 2 km above the snow line (ELA).

3. The highest prehistoric glacier level, i.e. the maximum Ice Age (LGM) ice cover in the Biafo- and Chogolungma area with the Braldo- (or Baldo-) and Basna valley as well as in the Shigar valley and the Skardu Basin in the Indus valley

3.1 Former ice levels above the present-day Biafao glacier

The highest summits of the Biafo valley are provided by the Baintha- (or Ogre-) massif with the 7285 m-high Baintha Brakk (Photo 65, No. 38), the 7108 m-high Latok II (No. 39) and the 6422 m-high Uzun Brakk (No. 40). These three mountains show the characteristics of glacigenically sharpened summits, i.e. glacial horns (Figure 2/1 No. 38-40). The most perfect feature is the Latok II (No. 39). Still further up-valley, i.e. upwards of the Biafo glacier, two further glacial horns stand in the confluence area to the Sim Gang glacier (No. 43). The higher one attains an altitude of 5989 m. In contrast to the summits of the Baintha-massif, it was completely covered by the LGM-glacier surface (Photo 69 and Figure 2/1 No. 43). The sharpening lateral abrasion of the ice on these mountains as well as the sporadically preserved glacigenic abrasion on the orographic left rock slopes (Photo 65 and 66 ▲; Figure 2/1 between No. 43 and 36) provides evidence of an LGM-ice level (▬ ▬) between 6300 and 6100 m asl. This prehistoric abrasion has created five classic triangular-shaped slopes between the inflows of the side valleys (Figure 2/1 No. 39, 41, 36). As a rule, the rock abrasions are best preserved halfway up the slopes. On top occurs a more persistent absence of ice and subaerial weathering, whilst right down the development of the wall gorges – corresponding with the most-extended catchment areas connected here – is at the most advanced stage since the deglaciation. In many places, however, their development brought about by backward erosion, has not even reached half the height of the slopes. So, for instance, the flank abrasion on the triangular-shaped slope at the locality of Chaunpisha (Photo 66 ▲ on the left; Figure 2/1 on the left below No. 39) has been preserved very well at its base and - with an increasing height – becomes rougher (Photo 65 below and above ▲). The glacigenic triangle-slope NW of the Dongbar locality has been cut by a wall gorge from top to bottom (Photo 66 ▲ on the left side below No. 41). It is the deepest on top and at the bottom (above ▽ in the middle): on top because of its longer lasting developmental period and below because of an increasing intensity of the process. Apart from the abrasions the ground moraine remnants are the strongest indicators of the glacier level (Photo 65–69 ■).

However, they reach at most as far as 650 m over the present-day Biafo glacier level (Photo 65 ■ on the very left; Photo 66 ■ white on the right side; Figure 2/1 below No. 38 and 41). If at many places even the bedrock breaks down (Photo 66 ⇩; Figure 2/1 No. 38, 39) – how much shorter must then be the preservation of the loose rock of a moraine on the slopes! This means at the same time that the 650 m-moraine level (see above) belongs to the Late Glacial. There remains nothing but the rounded abrasion forms on the rocks for the determination of the LGM-glacier level.

Especially in the area of the locality Mango does the orographic right side of the Biafo glacier valley provide an example of the redeposition of moraine and the reshaping by glacigenic and fluvial dynamics after the deglaciation: the ground moraines on the slopes (Photo 67 the two first ■ from the right) have been damaged and worn down into mudflow fans by rill rinsing and redeposition (▲). The ground moraines have been formed to 'transitional glacial debris accumulations' in the sense of Iturrizaga (1998; 1999). Lateral abrasion by present-day hanging glaciers, as e.g. the Mango glacier (Photo 67 ◊), also causes a reshaping and redeposition of prehistoric moraines (Photo 68 ■). The lowest accumulations resulting from these reshapings are the mudflow- and alluvial fan complexes consisting of several individual forms which extend over kilometres (Photo 67 △; Figure 2/1 above No. 37), which have been kame-like accumulated against the present-day Biafo glacier (□), i.e. its sub-recent lateral moraines (■ on the left, to the right of ▷). The substrate-specific characteristics of moraine material, as e.g. coarse-edged, round-edged, facetted and also well-rounded (○) boulders 'swimming' isolated from each other in a fine, clayey groundmass (Figure 7–36) has been preserved over large parts. This is all the more understandable, because a historic front moraine is also part of the build-up of the kame of Mango (Photo 67 ■ large, diagonally right below No. 37; Figure 2/1 No. 37). A further glacial horn, which has pierced the High Glacial (LGM) glacier level, is the 6282 m-high Gama Sokha Lumbu (Photo 69 and Figure 2/1 No. 42).

3.2 Insertion concerning the problem of rounding and sharpening by glacigenic abrasion

Glaciogemorphologically interesting is the sharpened, c. 5500–5700 m-high crest between Shinlep Brakk (No. 37) and Gama Sokha Lumbu (No. 42). It has been overflowed by a c. 500 m-thick ice stream network (cf. Photo 67 ▬ ▬ to the left and right of No. 37). Today, however, glacigenic roundings no longer occur in some sections (Photo 68, rock crest in the background). Obviously the interglacial hanging glaciers setting in on both sides below the crest (in Photo 67 ◊ and 68 the Mango glacier is visible) have brought about this sharpening by their ground polishing. In the same way, as the abrasive roundings provide evidence of an Ice Age glacier level above the intermediate valley ridges, here, the abrasive sharpening proves the interglacial hanging glaciation subordinated to the relief. In dependence on the progress of the one or the other of the processes, the rounding or the sharpening is more extended. Thus it might be possible that hardly any roundings are preserved at the end of an interglacial period. Owing to this, the fourth dimension – the time – is in the same way decisive for the geomorphological evidence of a prehistoric ice level as the intensity of the prehistoric forming-process itself. For this reason the absence of an indicator can be no counter- evidence but its existence alone can provide evidence. Therefore a chronological differentiation between the LGM- (Stage 0) and Late Glacial- (Stage I, II, III, IV) ice level is possible in some places. The metamorphic sedimentary rocks of the glacial horn – it is a NE-satellite of the Shinlep Brakk (Figure 2/1 No. 37) – on the orographic right side of the Biafo valley near the Brangsa locality, for instance, have been polished as far as a polish cavetto at c. 650–700 m above the present-day glacier level during the late Late Glacial (Stage IV) (Photo 69 ☗ on the very right). At the same time the summit pyramid above has been underpolished. In consequence, the higher abrasion forms are broken away (above ☗ on the very right), because the merely c. 4800 m-high horn was completely covered by the ice surface during the LGM (▬ ▬ on the right).

3.3 The LGM-ice thickness in the Biafo glacier valley

The profile lines of the trough slopes in the Biafo valley with inclines between 32° and 41° dip beneath the present-day glacier surface (Photo 65 ■ black and the shadow-line on the left below No. 40; Photo 69 on the right below No. 36; Photo 70 on both sides of No. 43). In the cross-profile between Mango and Dongbar the Biafo glacier is 1.8 km-wide with a surface about 3750 m asl (Photo 69; Figure 2/1 between No. 37 and 41). Under consideration of a trough-shaped profile line, which becomes flatter towards the trough ground, a trough geometry can be constructed with a rock bottom lying 500 m below the present-day ice level, i.e. at 3250 m asl. 25 km up the Biafo glacier valley, the current ice stream, i.e. Biafo trough, is 3 km-wide at a glacier level at 4450 m asl (Photo 69 cross-profile below No. 43; Figure 2/1 in the area between No. 43 and 42). At the same average inclines of the trough flanks (32°–41°), the trough geometry confirms a height of the trough bottom at 3750 m asl and thus a present-day ice thickness of approximately 700 m. Glacier levels between 6100 and 6300 m asl during the LGM (Photo 65 ▬ ▬, 66 ▬ ▬ on the left and on the very right) yield High Glacial ice thicknesses about 2450–2950 m. Downwards of the Biafo valley the LGM-glacier thickness has increased by at least 200–300 m.

3.4 The prehistoric indicators of the glaciation and ice levels in the confluence area of the Biafo valley into the Braldu (Biahu Lungpa, Blaldu-) main valley

Part of the substantial surface moraine mass of the present-day ice streams, as e.g. the Biafo glacier (Photo 70 □), consists of dislocated prehistoric ground moraine (■). Owing to undercutting of the present-day glacier margin or kame-like mudflow-fans and -cones (▼) – always combined with ice ablation – it has been incorporated into the present-day glacier surface. For this reason the modern interglacial ice streams are much more interspersed with debris than the

Table 1 Glacier stages of the mountains in High Asia, i.e. in and surrounding Tibet (Himalaya, Karakorum, E-Zagros and Hindukush, E-Pamir, Tien Shan with Kirgisen Shan and Bogdo Uul, Quilian Shan, Kuenlun with Animachin, Nganclong Kangri, Tanggula Shan, Bayan Har, Gangdise Shan, Nyainquentanglha, Namche Bawar, Minya Konka), from the pre-Last High Glacial (pre-LGM) to the present-day glacier margins and the pertinent sanders (glaciofluvial gravel fields and gravel field terraces) with their approximate age (after Kuhle 1974-2000).

glacier stadium		gravel field (Sander)	approximated age (YBP)	ELA-depression (m)
-I	= Riß (pre-last High Glacial maximum)	No. 6	150 000 - 120 000	c. 1400
0	= Würm (last High Glacial maximum)	No. 5	60 000 - 18 000	c. 1300
I - IV	= Late Glacial	No. 4 - No. 1	17 000 - 13 000 or 10 000	c. 1100 - 700
I	= Ghasa-stadium	No. 4	17 000 - 15 000	c. 1100
II	= Taglung-stadium	No. 3	15 000 - 14 250	c. 1000
III	= Dhampu-stadium	No. 2	14 250 - 13 500	c. 800 - 900
IV	= Sirkung-stadium	No. 1	13 500 - 13 000 (older than 12 870)	c. 700
V - 'VII	= Neo-Glacial	No. -0 - No. -2	5 500 - 1 700 (older than 1 610)	c. 300 - 80
V	= Nauri-stadium	No. -0	5 500 - 4 000 (4 165)	c. 150 - 300
VI	= older Dhaulagiri-stadium	No. -1	4 000 - 2 000 (2 050)	c. 100 - 200
'VII	= middle Dhaulagiri-stadium	No. -2	2 000 - 1 700 (older than 1 610)	c. 80 - 150
VII - XI	= historical glacier stages	No. -3 - No. -6	1 700 - 0 (= 1950)	c. 80 - 20
VII	= younger Dhaulagiri-stadium	No. -3	1 700 - 400 (440 resp. older than 355)	c. 60 - 80
VIII	= stadium VIII	No. -4	400 - 300 (320)	c. 50
IX	= stadium IX	No. -5	300 - 180 (older than 155)	c. 40
X	= stadium X	No. -6	180 - 30 (before 1950)	c. 30 - 40
XI	= stadium XI	No. -7	30 - 0 (= 1950)	c. 20
XII	= stadium XII = recent resp. present glacier stages	No. -8	+0 - +30 (1950 - 1980)	c. 10 - 20

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High Glacial ones. Naturally, this comparison applies only to Ice Age ablation areas, i.e to the areas below the ELA at the margin of the Karakorum ice stream network (Figure 2). Here, in its centre, the ice surfaces lay 2 km above the snowline, so that no surface moraine could occur anyway. The rounding and facetting of separate boulders (○ on the left) documents the redeposition of prehistoric moraines. The development and preservation of the glacigenic flank abrasion is to a great extent structurally-dependent in its heterogeneity. This becomes obvious on the Bullah-SW-wall (Photo 74 No. 36). Orientated by the edges of the strata and bedding joints, subglacial, large polish faces already break away (Photo 70 ▲ left of No. 35; 74 left of ▲ white below No. 36). Ground moraine fillings in the out-broken wall scores (Photo 70 ■ on the left) are evidence of the subglacial process. There are also only a few undamaged glacigenic abrasion forms on the orographic right side (▲ white and black between No. 44 and 43). In other places, as for instance opposite the Biafo glacier tongue, even easily splintering phyllites still show well-preserved abrasion roundings (Photo 73 ▲; Figure 2/1 above No. 35). This is a clear indication of their minor, i.e. Last Glacial or late Late Glacial age (Sirkung Stage IV).

Correspondingly, the heavier roughenings on the valley flanks at higher altitudes go back to the earlier Late Glacial and LGM (Taglung Stage II, Ghasa Stage I or Stage 0 = LGM). An example is provided by the Shinlep Brakk ESE-spur (Photo 74 ▲ below No. 37; Figure 2/1 on the right of No. 37). Here, as they have been preserved rounded and smoothed on the lower slopes of the trough flanks (Photo 73 ◥ on the left), similarly easy-weathering metamorphic sedimentary rocks have been significantly more strongly roughened at a c. 1000 m-higher level of the valley flanks. A mantling by ground moraine occurs on the lower slope of this roughened abrasion faces (Photo 74 white ■ below No. 37; Figure 2/1 on the right of No. 37). It also covers the rounded mountain ridge further below (Photo 74 ■ black on the right below No. 37). Large erratic granite boulders are embedded into this ground moraine cover; several of them (◊) also overlie glacigenically abraded rock heads (▲ white and black on the right in the fore- and middleground). The ground moraine on the lower slope is covered by a younger, i.e. Late Glacial (Stage I-III) lateral moraine rampart (✓; Figure 2/1 below No. 37). In comparison to the LGM-ice level (— —) (Stage 0 after Kuhle 1980, 1982a, 1998a Tab. 1 p. 82), the Late Glacial glacier margin, which is thus evidenced, has already been lowered by c. 1200-1400 m. Some of the glacigenically abraded rock knolls (Photo 74 (▲ foreground), separated from each other by polish depressions, show classic roches moutonnée-forms (▲ black below No. 41 in the foreground; Photo 76 ▲; Figure 2/1 on the right below No. 37). In the places where a moraine overlay is lacking, they are deeply weathered (▲). The ground moraine cover further down the mountain ridge (■; Figure 2/1 on the right below No. 37) contains large erratics, too (○). It shows periglacial pattern grounds (Photo 76 ■ and ○).

A further orographic left lateral moraine rampart (Photo 77 ■ black; Figure 2/1 below No. 37) in the Braldu valley flank runs at 4000 m, i.e. 1000 m above the valley bottom (○). Between the lateral moraine's outer slope and the valley slope, a classic lateral valley with an alluvial soil (□) has been developed. Upwards of the influx of the Biafo valley, at the same altitude, there are solely ground moraine covers (■ white; Photo 75 ■ white). Corresponding orographic right lateral moraine remnants with lateral moraine depressions, i.e. small lateral moraine valleys, belonging to the Late Glacial Glacier Stages III and IV, occur c. 3 km down-valley (to the W) in the area of the Alpine pasture above the Askoli settlement (Photo 78 between ■ and the left margin). The stone houses of this pasture are situated at 4200 m asl between a large roche moutonnée (Photo 84 ▲ on the left in the middleground; Figure 2/1 on the left below No. 37) and the actual orographic right Braldu valley slope. This glacigenically polished U-shaped saddle (U on the left) lies 1300–1400 m above the main valley floor. Down-valley, on the orographic right side, further Late Glacial lateral moraine ledges (■ black on the right) with lateral moraine depressions (□) and erratic boulders (↓) are preserved (Figure 2/1 on the left of No. 37). On the ESE-spur of the Shinlep Brakk (Figure 2/1 on the right below No. 37) stepped moraine accumulations with erratic granite boulders can be observed down-slope as far as c. 3600 m asl, i.e. up to 500 m above the Biafo glacier tongue (Photo 74 □ on the right). 2.2 km down the Braldu valley, part of this High Glacial (Stage 0 = LGM) ground- and Late Glacial lateral moraine material (Photo 78 □) has been transported down-slope and redeposited into a mudflow fan or -cone (▽ white) since the deglaciation (Figure 2/1 No. 37).

During Stage XI (about 1950) the Biafo glacier tongue has reached the orographic left valley side and basal rock wall of the Bakhor Das (No. 35) over a width of 2.7 km (Photo 74 □ on the very left and right) and thus has crossed the Braldu valley (Photo 73). The Braldu river flowed down beneath the glacier ice. At this time the ice lay at the level of the striking 30 to 50 m-high lateral moraines (Photo 70 ■ black). Also the 'spillway' on the orographic right margin of the glacier tongue (▽) between the roche moutonnée and the trough flank has then still been overflowed by the meltwater (Photo 71 ⇩). About 1950 the old expedition route to the Biafo- and Baltoro glaciers already led through the ravine (Photo 72 ⇩) behind this roche moutonnée (Photo 70 on the left of ▽). During Stage X, i.e. between 1800 and 1920 (Tab 1), at least part of the Biafo meltwater flowed down through this ravine. However, this took place sub-aerially. The 'Younger Dhaulagiri Stage VII' was the last period during which the discharge through this ravine was also still subglacial. One may safely assume that at a snow-line depression of c. 60–80 m, as it is characteristic of this stage (Kuhle 1980; 1998a), the ice has overflowed this roche moutonnée (Photo 70 on the left of ▽) and the ravine behind. For comparison see the detailed investigations on the neoglacial and historic glacier history of the NW-Karakorum 100 km further NW by Meiners 1996: pp 95–183 and 1997: pp 270–301.

To sum up one may say that in the area of the confluence of the Biafo glacier valley into the Braldu valley (Figure 2/1 between No. 37, 36 and 35) the ice level has run between c. 5600 and 6000 m asl at an ice thickness of c. 2600–2900 m during the LGM. The remaining uncertainties of several hundred metres as for the reconstruction of the ice level, partly go back to the inexact summit heights on the 'Karakorum Mountaineering Map 1 : 200,000'.

3.5 The Ice Age glacier filling of the Teste valley in the Mango Range-N-slope (with the Skoro La Gans = glacier) as an example of an orographic left hanging valley above the Braldu valley near Askole

Directly SSE of the Shinlep Brakk (Figure 2/1 No. 37) the Teste trough valley is situated (Photo 84 U white; Figure 2/1 on the right of No. 45) into which one can look from the Shinlep Brakk-S-flank (locality: Photo 79 ▲ on the left). At its valley head the Skora La Gans (glacier) extends below unnamed summits of the Skora massif towering as far as 6000 m asl at maximum (Photo 80 and Figure 2/1 No. 46). This valley head, prehistorically completely filled with the ice of an ice stream network, now undergoes the reshaping into separate Alpine-type glaciers. Owing to this, a process is concerned as it is known of the Alpine High- and interglacial history. Whilst the historic moraine complexes of Stages VIII to XII (Photo 80 ■) – which with ELA-depressions of merely 50 or less metres belong together and are partly still attached to the current glacier margins (□) – show only a small geomorphological distance to the present-day glaciation, the glaciogeomorphological distance to the glaciation during the LGM is far greater (Photo 80 – –; 84 – –; Figure 2/1 below No. 46). The geomorphodynamic mediation is established by the fact, that the oversteepened rock flanks of the Teste trough have broken away since the melting of the ice (Photo 80 and 81 ⇩, Figure 2/1 above No. 46) and thus have overwhelmed the High- to Late Glacial ground moraines on the slope with debris from the rock fall (Photo 80, 81 and 84 ▽ black, Figure 2/1 mudflow cone right of No. 45). Whilst here, at the head of the Teste trough, the late Late Glacial, neoglacial and historical glaciers (Stage IV to c. IX; Tab 1) have strongly undercut and modified the trough flanks, flank polishings are better preserved at the exit of the valley (Photo 81 below – –, Figure 2/1 above No. 46). The reason for this is that the valley bottom was more rapidly cleared of ice since the late Late Glacial, so that then a glacigenic undercutting has no longer taken place. In a corresponding way, the rock shoulder, rounded by glacier abrasion in the area of the glacigenic confluence step into the Braldu valley, is preserved the best (Photo 79 ◢ white; 82). Even Late Glacial ground moraine can be recognized here (Photo 81 ■ white on the left; Figure 2/1 on the right above No. 45). At the valley head of the Teste trough an ice transfluence over the Skoro La has taken place during the LGM (Figure 2/1 between No. 45 and 46, Photo 84 ↷) so that a glacier-connection existed between the Skoro Lungma tributary stream, the Shigar parent glacier and the important ice filling of the Skardu basin.

3.6. The LGM-glacier-filling of the Braldu valley chamber of Askole between the settlements of Ste Ste, Sino and Tonga

The enormous, approximately 3000 m-thick ice filling in the Braldu trough down from the Biafo glacier has been documented in detail (see Chapter 3.4) and depicted by a cross-profile (Figure 3). The glacier filling continuing downwards near the settlement of Ste Ste (Photo 79 and 81 ○) becomes obvious by the photo-perspective from the Teste valley exit (viewpoint ⇩ in Photo 84) on to the Shinlep Brakk (Photo 79 No. 37), which shows the ground moraines and Late Glacial lateral moraines (■) as well as the flank polishings (☁) described in detail (Chapter 3.4). Again down-valley, in the trough cross-profile of Askole on the orographic right side, ground moraines, Late Glacial lateral moraine ledges (Stage II, Tab 1) and roche moutonnée- (Figure 2/1 on the left below No. 37) as well as hanging trough valley forms can be observed about 4100-4200 m asl (Photo 84 ■ white below, ■ II, ☁, ⊔). These indicators lie 1300–1400 m above the Braldu talweg.

3.7 Additional comment on the indicator value of cirques as to the speed with which the Early Glacial ELA must have dropped

On the orographic left valley side below the High Glacial ice surface (— —) are Late Glacial cirques (○) which do not correspond to the height of the snow line, but are oriented by the height of the lowered level of the ice stream network (Figure 2/1 above No. 45). Their development by an individual glacier filling of the cirques had only then become possible, when the appropriate upper valley slopes had been released from the valley glacier. These cirques are no single-phase features of the last Late Glacial, but have been formed during the Early- and Late Glacials of the pre-LGM-glacial periods as well. During the respective Early Glacials they were immediately dependent on the depression of the ELA, but during the Late Glacials they were dependent on the down-thawing of the High Glacial valley glaciers. Comparatively low-lying cirques and nivation niches, as e.g. the hollow mould above the fields of Manjong (Photo 83 ○) at merely 3600–3800 m asl in a N-exposition, have been created in the respective Early Glacials. This happened at a time when the orographic snow-line had already been lowered by over 1000 m, the ice stream network, however, was only built up to a thickness of 1000 m. During the LGM the hollow mould of the cirque above Manjong (see above) has been polished transversely by the Braldu valley glacier, as shown by the preserved roundings of flank abrasion (☁ on the right below No. 45 and 47; Figure 2/1 above No. 45). This genetic sequence is an indication of a very fast drop of the ELA during the Early Glacial. Otherwise the filling of the valley with glacier ice, already exceeding a level of 3600–3800 m at an ELA-depression of merely 700 m (appropriate to the ELA-depression during the Late Glacial Stage IV, see Tab 1) and attaining a level-height about 4000 m (cf. Photo 77), was bound to have already taken place and the cirque glacier, which needs an ELA-depression of over 1000 m for its development, could not have been created.

3.8 Continuation of the comments on the LGM-glacier filling of the Braldu valley chamber of Askole between the settlements of Ste Ste, Sino and Tonga, with geomorphological references to their age

In the valley chamber we are talking about, the numerous deposits of ground moraines, which in many places are preserved very high up the Braldu valley flanks (Photo 83–86 ■; Figure 2/1 on the right above No. 47), are surprising in view of the steep relief and the related erosional speed of loose rock like this. To put it another way: only if these ground moraines are of a very young, i.e. Late Glacial age (17,000 to 13,000 or 10,000 YBP; see Tab 1) they are *not* surprising. Thus, the evidence has to be called to mind once more that the entire glacier filling reconstructed here has to be classified as belonging to the last Quaternary Ice Age, i.e. to the LGM and the subsequent Late Glacial. The dimensions of the seasonal re-deposition of important masses of ground moraine material by mudflows are made clear by the huge, up to over 100 m-thick, mudflow fans and -cones (Photo 83 △). The mudflow fan in the orographic right flank NW of the fields of Askole (Askoli) (Figure 2/1 on the left below No. 37; Photo 83: the two white ▽ in the middle) is especially striking. It still contains decametre-thick ground moraine in situ (second black ■ from the right) at its root.

In some places the ground moraine is better preserved than the actually more solid bedrock. This applies to the areas where the foliation- and bedding planes of the phyllites favour the structural-concordant crumblings (Photo 83 ⇩, 86 ↖).

Beside the Teste trough already described, the Sino valley is a further orographic left trough-shaped side valley (Photo 83 ⊔; Figure 2/1 on the left above No. 45). Through these side valleys and also through the valley from the orographic right side above the settlement of Tonga (Photo 86 ⇩) considerable quantities of gravel are transported by the steeply down-flowing glacier creeks as far downward as the Braldu valley bottom. The alluvial debris- and mudflow fans of these side valley junctions at the settlements of Sino and Tonga (Figure 2/1 between No. 42 and 47) are adjusted to the level of the gravel bottom of the Braldu river (Photo 85 ▽ centre). Its continuous and phased intensified cutting has developed terrace flights in both side valley fans which correspond at comparable levels and steps (Photo 85 ▼ and 86 △). The present-day farmland is situated on these terraces. The fan material consists of displaced moraine, which mudflows have initially transported down into the main valley (Photo 85 □). There it has been more or less washed by the tributary creek (Photo 85 △ centre and 86 △) and then by the Braldu river as well (Photo 85 ○).

The highest preserved roundings and rock smoothings, which must be explained by glacigenic abrasion (Photo 83–

86 ⯊; Figure 2/1 between No. 42 and 47) and which tower above the deposits of ground moraine by approximately 1000 m, attain levels between 5300 and 5600 m (▬ ▬). This is proof of an LGM-ice thickness of c. 2600 m between Cherichor (No. 45), Shinlep Brakk (No. 37) and Koser Gunge (No. 47).

3.9 The traces of the LGM-glaciation in the Braldu valley chamber from the settlement of Tonga as far as the inflow of the Hoh Lungma, and the classic development of a ground moraine pedestal (with a preliminary methodological remark as to the data-density)

Naturally it is not possible that below Tonga the glacier ice filling of the Braldu valley suddenly breaks vertically away by 2600 m and is completely lacking. According to this empirical basis, a greater gap as to the description of the findings might be allowed, without considering the downward continuation of the very thick Braldu glacier as a tributary of a huge ice stream network as being purely hypothetical. On the other side, the methodological strength of a glacial-historic reconstruction is based on as complete as possible a spatial provision of unambiguous field-indicators of a prehistoric ice cover. Because the postglacial reshaping of the glacigenic features with increasing intervals enhances the indicator gaps, the arrangement of the position between the data-localities becomes more and more significant in deciding between the observation 'here was a glacier cover' or 'here was no glacier cover'. However, since the reference to the arrangement of the position of the corresponding field data requires the reader's spatial imagination, but given that only a willing reader is prepared to spare no efforts on that score, the scientific procuring of facts is expected to keep the spatial distance of the field observations as small as possible. Therefore the data are to be continued down-valley in the close-meshed way which has been applied so far. The greater the data-density, the more convincing is the result extracted.

In the area of the settlement of Hoto the orographic left flank of the Braldu trough, covered high up (see above) with ground moraine (Photo 83 ■ in the background, right of the centre), is interrupted by two side valleys which are today still glaciated as far down as c. 3920 m. The steep hanging glaciers flow down from the 6251 (or c. 6400) m-high Koser Gunge (Photo 83 and 84 No. 47) N-slope. Both the side valleys, the northern Pakore valley and the unnamed southern one, show U-shaped forms all over the valleys (i.e., up to the upper margins of the valley flanks) polished out only in the late Late Glacial (Stage IV, Tab 1) (Figure 2/1 above No. 47). Before, the Braldu parent glacier was still so thick, that the tributary glaciers dammed-up by it had no possibility of ground scouring. During the Neoglacial (Stage V-'VII) the steep tributary glaciers were but a few hundred metres-thick and thus not thick enough to develop U-shaped valleys.

A characteristic of the cross-profile of the Braldu valley between the Koser Gunge (No. 47)and the Gama Sokha Lumbu (Photo 91 No. 42) being also over 6200 m-high, is an important moraine filling (Photo 87 ■). This is so substantial that since deglaciation it has not been removed by the Braldu river, here flowing down steeply and torrentially (Photo 87 ▲). A reason for the quantity of loose rocks is the great height of the local catchment area of these mountains, the hanging glaciers of which still into the Holocene were flowing down as far as close to the Braldu talweg. The main reason, however, is to be seen in the gorge-like narrowness of this section of the Braldu valley. As a result, the Ice Age Braldu glacier has developed a huge ground moraine pedestal between the settlements of Hoto and Tosha (Photo 90 and 91 ■ black below; Figure 2/1 between No. 42 and 47). The build-up of a ground moraine pedestal like this is the function of very great ground friction. In order to guarantee the necessary velocity of discharge of the ice supply from up-valley also in a narrow gorge-profile, the glacier has made-up a broad, trough-shaped bed of plastic ground moraine, poor in friction. Obviously the glacial erosion scouring the rock was too slow here, so that the glacier, stripping off its ground moraine, has been pressed up as far as a higher level (cf. Kuhle 1982a: p 49, 50 and 1983a: pp 123–125).

Of an unsurpassed unambigousness are the glacier striations, preserved and visible on the orographic left flank (Figure 2/1 above No. 47; Photos 88 and 89 ⯊). The visibility persists no longer than at most a few decades after the exhumation; then the striations have been wasted away by frost weathering and the glacigenic rock polishings go dull through splintering. Here, the metres-thick ground moraine overlay, isolating from weathering, has only recently slid down and been washed away from the polished rock as a result of heavy rainfalls, which on the water-retaining rock led to soaking. The ground moraine cover still exists in remnants (■) so that these observations were possible. The glacier striations of the granite areas glacigenically treated for the last time during Stage IV (Sirkung Stage, c. 13,000 YBP Tab 1) are almost horizontal (⯊). They thus indicate a glacier mouth terminal at a distance of at least several kilometres.

In many places glacigenic rock polishings and abrasion roundings alternate with ground moraine covers, which in slope depressions are almost 100 m-thick (Photos 90, 91 ⯊ and ■). This thickness can be estimated from earth pyramids of remarkable dimensions which consist of this ground moraine and have been developed on the orographic right side c. 4300 m above the settlement of Gombro (Photo 90 ■ white on the right; Figure 2/1 on the left below No. 48). In the junction area of the Hoh Lungma connected from the N (right), an extended classic glacigenically triangular-shaped slope is preserved (Figure 2/1 on the right of No. 50) overlain by a ground moraine packed up to a thickness of decametres (Photo 90 ■ below the left ○). The valley bottom, which consists of washed ground moraine, shows a height about 2545 m asl in the confluence area. C. 4 km up the Braldu valley, the V-shaped gorge-profile (Figure 2/1 below No. 42; right third of Photo 91) widens to a concavely polished, i.e. trough-shaped profile line (Photo 90 in the fore- and background; Figure 2/1 on the left below No. 48). The connected Hoh Lungma shows a similarly glacigenically-widened V-shaped valley profile with an already trough-like form (Figure 4; Figure 2/1 between No. 50

and 48; Photo 92 below No. 49). Despite the great steepness of 33–40°, important ground moraine deposits are preserved in the depressions and large bowl-shaped concaves of both flanks of the Hoh Lungma (Photo 92 and 93 ■ black above; Figure 4; Figure 2/1 on the right of No. 50 and below 49).

The highest prehistoric glacier traces are presented by abrasion areas still reaching 700 m above the preserved ground moraines. This concerns the classic glacigenic roundings of rock ridges and -heads (Photos 90, 92, 93 ☗). The small cirque forms (Photo 90 ○; Figure 2/1 on the right of No. 50) have been shaped by individual glaciers only since the Late Glacial. By means of these indicators the LGM-glacier surface of this confluence area has been reconstructed at an altitude of at least 4900 m asl (– –; Figure 4).

3.10. The LGM-glaciation of the lower Braldu valley from the inflow of the Hoh Lungma as far as the Shigar valley near the settlement of Mungo

In the further course of the Braldu trough, 6 to 7 km down-valley from the inflow of the Hoh Lungma, the thick ground moraine pedestal, already described further up-valley (Chapter 3.9), is preserved in remnants (Figure 2/1 on the right below No. 50; Photo 92 ■ below). Its surface, which has not yet been reworked, passes with a perfect, concave trough profile bend (above ■ below) into the orographic right rock flank polishings (◥ on the very left). Immediately up- and down-valley, the moraine pedestal has already been removed by the Braldu river. Correspondingly, the remaining remnants, too, are undercut by lateral erosion of the Braldu river – which, with a run-off of several 100 m^3/sec, is extremely energetic – and have been shifted backward in the form of steep crumblings (Photo 93 ■ below) and wasted away. The great density of this ground moraine compacted by the high overburden pressure of an over 2000 m-thick (Figure 3) hanging Braldu parent glacier, is the reason for the development of finely-chiselled earth pyramids on the remnants of this ground moraine pedestal (Photo 92 ↓; Figure 2/1 on the right below No. 50). Trough-profile-like, concave glacier ground scouring is also preserved in the bedrock granite on the valley bottom (Photo 95 ◤ quite below and on the very right). Here, remnants of the ground moraine pedestal (■) and ground scouring on the rock occur immediately adjacent. This may be a reference to the succession of High Glacial ground scouring and the development of a Late Glacial pedestal or it could be an indicator of the – locally separated – simultaneousness of ground scouring here and accumulation of ground moraines on the opposite valley side, as it applies to the outer and inner bank in the fluvial environment. Here, a problem is concerned, which, according to the present-day state of knowledge, must be left open. On the orographic right flank of this valley section flush-bowls and parts of potholes (Photo 94 ○; Figure 2/1 between No. 50 and 47) do occur, created by water flowing under pressure, i.e. subglacial water. They are evidence from a prehistoric glacier, but do not allow a chronological classification – just like the moraine pedestal and the ground scouring. The reason for this is that at this altitude about 2500–2600 m asl, the meltwater features lie 700–900 m below the lowest prehistoric ELA (that one of the LGM = High Glacial = Stage 0; Tab 1). Accordingly, at that time subglacial meltwater has already flowed down and may have developed the pothole sections.

The spatial but not chronological succession of ground scouring and ground moraine sedimentation becomes obvious in the valley chamber of the wasteland of Gohe which follows lower down. Here, the rock spurs, rounded and smoothed by abrasion, protrude from the valley flanks towards the talweg (Photo 96 ☗ large), so that the ice discharge was impeded down-valley. Even a roche moutonnée-like feature has been developed here (☗ white large; Figure 2/1 on the left above No. 47). Further down-valley a protruding glacigenic triangle-shaped face near the settlement of Niyel (☗ below – –; Figure 2/1 on the left above No. 47) documents the increasing abrasion on protrusions. From the valley bottom up to the medial slope heights (slope catena) decametres-thick ground moraines have been deposited (■ black) between these rock protrusions, which means, that they have been stripped off by the glacier ice in front of these projecting rocks and, accordingly, were laid down between them. The increasing moraine thickness toward the lower slope, combined with the development of a ground moraine pedestal, can be observed on the outer banks of the Braldu river (■ black on the right below; Photo 97 ■ black and white below; Figure 2/1 on the left above No. 47).

Down from the valley chamber of the settlements of Niyel and Dasso, the widening box-like valley form of the type 'trough valley' with gravel floor (Photo 97 and 98 ○; Figure 2/1 below No. 50) is obvious. Up to here this glaciofluvial gravel floor was dammed back from the Shigar valley (Photo 99 ○), i.e. the large main valley, 8 km up the Braldu valley. The orographic right flank of the Braldu valley is made up here by the 5778 m-high southern pre-summits (No. 50) of the Ganchen massif (6462 m). Towards the SW, i.e. towards the Shigar valley, being the next-larger main valley, it peters out into a round-polished mountain spur (Photo 98 ☗). This spur (Photo 99 on the left of No. 50) separates the Braldu valley from the W-adjacent Basna valley (in the middle). Its flatly sloping mountain ridge consists of granite and is composed of numerous roches moutonnées (Figure 2/1 below No. 50). Its flanks still show glacier polishings in the form of band polishings of outcropping edges of the stratum (below No. 50; Photo 98 ☗). A persevering, i.e. partly still covering ground moraine overlay (■), has preserved the polishings. On the orographic left valley flank, too, an unambiguous abrasion form occurs in the shape of a glacigenic triangle-slope (Figure 2/1 half-left above No. 47) reaching up to the culmination of the mountain ridge, named Busper, at an altitude of 4564 m. Correspondingly, this mountain ridge was completely overflowed by ice during the Ice Age and the LGM-ice level communicated over the ridge at a continuous level with that one of the Shigar parent glacier. The cross-profile of the Shigar valley (Figure 5) shows the reconstructed Shigar parent glacier flowing with its LGM-level about 4700 m asl (which somewhat further NW has dropped to 4380 m) over this Busper NW-ridge

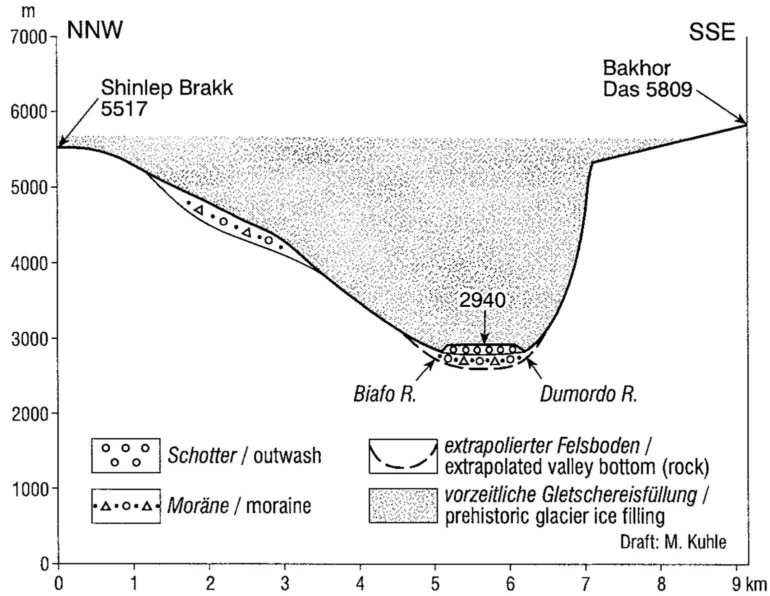


Figure 3. Cross-profile (not exaggerated) across the upper Braldu valley (Biaho Lungpa) from the Shinlep Brakk in the NNW up to the Bakhor Das in the SSE with its minimum glacier ice filling, reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 37 and 35.

(Figure 5 on the right margin) and forming a continuous surface with the tributary stream of the Braldu glacier. The ground moraine overlay preserved on the WSW-side of the 4380 m-ridge on the orographic left flank of the Shigar valley, is metres- to decametres-thick (Figure 5 below the 4380 m-ridge; Photo 99 ■ on the right).

In the section of the Braldu valley we are talking about, the LGM-ice level has been evidenced by a great number of upper limits of polishing and abrasion (Photo 92–97 – –). It lay between c. 4900 m asl in the area of the Hoh Lungma inflow and 4700 m asl (Figure 4) in the area of the Braldu inflow into the Shigar valley (Figure 5). Owing to this, the surface incline of the ice stream ran parallel to the present-day talweg incline down-valley. The maximum ice thickness might have even increased by 100 m from c. 2400–2500 m in the upper cross-profile to 2500–2600 m in the lower one.

3.11. The LGM-glaciation in the upper Chogolungma (Basna) valley from the ESE-slope of the Malubiting- (7458 m) and Spantik massif (7027 m) as far as into the valley chamber of the settlement of Arandu

Kick (1956, 1964 and 1991) has carried out glaciological investigations of the recent Chogolungma glacier and took this opportunity to describe the first observations from the slopes above the present-day ice stream as to the prehistoric glacier. However, his research was not concerned with the reconstruction of prehistoric glacier positions and ice levels.

It is a dendritic tributary glacier system which joins the Chogolungma main glacier from the N with the Moraine glacier (Photo 104 below No. 54; 103 below □ white) and from the S with the Crevasse glacier (Photo 100 between the two ⇩) and the Haramosh glacier (Photo 101 □; 102 above □) (cf. Figure 2/1 No. 52–58, 70–71). At the height of the present-day level of the Chogolungma glacier, the main valley reaches a maximum width of over 2.5 km (Figure 38; Photo 104, 105; Figure 2/1 between No. 71 and 56). Here, the parent glacier is c. 600 m-thick, as can be estimated according to the inter- and extrapolation of the steepness of the valley flanks (Figure 38). By means of the richer grass vegetation on steep scree slopes (Photo 102 ■; 114, 115), which below 4000 m even passes into lush high mountain pastures (Photo 117, foreground), today's greater humidity of the Chogolungma glacier area, compared with that one of the Baltoro, becomes obvious. As to the preservation of the glacigenic flank smoothings, however, no difference can be evidenced due to this differing precipitation in comparison with the Baltoro glacier valley. Here, too, the glacigenic flank abrasion above c. 5000 m asl can only partly be observed in the form of roundings (Photo 101–104 ▲), though the Ice Age trough development is verifiable as far as a polish line at 5900 m (– –; Figure 2/1 No. 55, 56). Between c. 5000 and over 6000 m the abrasions and especially the flank polishings have been roughened by residual snow and frost weathering – which at the black/white lines is in particular active - and destroyed by crumblings (Photo 102 ⇓; Figure 2/1 No. 53, 57); above and at the height of the ELA also by lateral erosion of the current glacier edges (below ▲). Below the ELA the ice margin is no longer attached to the rock, but it is isolated from it by lateral moraine and -small valleys (Photo 113 below ⇩). The undercuttings, which nevertheless do occur, have taken place during the neoglacial (Holocene) glacier levels (Stage V-'VII) at a snow-line that had dropped about 80–300 m against the present-day one (Tab 1) (Photo 113 ⇩). Exemplary crumblings (Photo 101 ⇩), solely deriving from the lateral erosion of a 250–300 m higher glacier level, are to be observed on the north-easternmost rock satellite of the Laila Peak (No. 55). At that time the glacier has still overflowed the rock rounding (▲ on the right). Above 5000 m, current avalanches rework the prehistoric abrasion

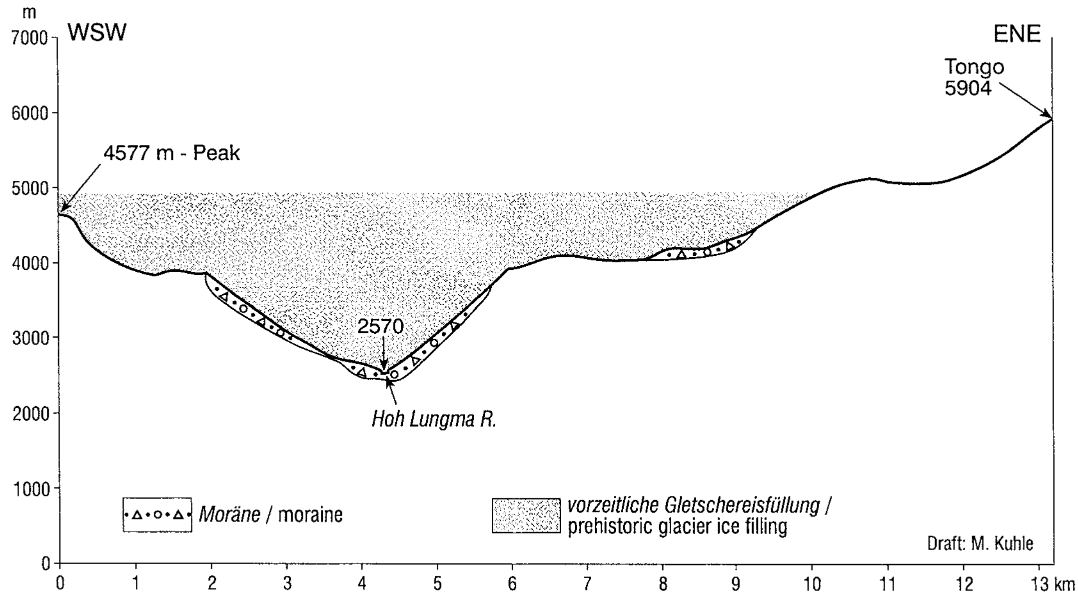


Figure 4. Cross-profile (not exaggerated) across the lower Hoh Lungma near its inflow into the middle Braldu valley (Biaho Lungpa) from the 4577 m-peak in the WSW as far as the Tongo in the ENE with its minimum glacier ice filling reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 50 and 48.

forms downslope (Photo 102 ⇩); naturally, the avalanches profit from the humidity.

The highest prehistoric glacier level can be recognized by the flattened summits of rock towers (Photo 100 ▬ ▬ below No. 52; 102 ▬ ▬ centre and on the right; 103 ▬ ▬), by large, still iced-over flat forms (Photo 113 and 123 ▬ ▬ below No. 54) as well as by flatter ledges, protruding from steep wall faces (Photo 100 ▬ ▬). The latter ones document the fact, that they were protected against the downslope avalanche erosion and -denudation by the level of the ice stream network as the relative erosion base. Naturally, sharpenings of rock spurs and -crests also took place subglacially. This happened at all places, where the crest-parallel scouring was stronger than that which ran transversely to the crest (Photo 101 ▬ ▬ on the right; 113 next to No. 56). Transition features also occur: so, the rock crest E of the 6970 m-high Malubiting E-summit has been somewhat rounded (Photo 100 ▬ ▬ below No. 52; Photo 102 ▬ ▬ on the right). Perfect indicators of prehistoric glacigenic flank abrasions are the triangle-shaped faces, which have been developed from truncated spurs and polished back and abraded further in a facette-like-stretched fashion (Photo 113 at ▬ ▬ on the right below 52; Photo 123 ▬ ▬ on the right of No. 52; Figure 2/1 No. 52). A triangle-shaped face with a classic rounding has been preserved at the exit of the First East-Haramosh glacier valley (Photo 102 and 104 ▲ black; Figure 2/1 No. 56). There are also three glacigenically back-polished triangle faces on the orographic left valley flank (Figure 2/1 below No. 70 and 71; Photo 105 second and third ▲ from the left and on the right below No. 71; see in detail in Photo 110 and 111 ▲). The highest of these indicators at the valley head of the Chogolungma valley attain 6200–6400 m. They prove a connection of the ice stream network with the NW-adjacent Hunza valley glacier over the 5840 m-high Polan La (Photo 113 ▬ ▬ with a bent course on the right below No. 52). At this transfluence pass the thickness of the overflowing ice (Figure 2/1 No. 52) amounted to c. 350–500 m (Photo 123 ▬ ▬ on the right of No. 52). Because the ice flow down the Hunza valley is steeper, i.e. has reached the 2100 m-level over a distance of only 40 km, whilst the distance to this level towards the SE, to the basin of Skardu, is at least 110 km, the LGM-ice-transfluence from SE to NW is probable. A further transfluence pass is the c. 4800 m-high Haramosh La (Figure 2/1 No. 55), leading from the root of the Haramosh valley toward the WSW through the Mani- and Phuparash Gah (valley) to the Indus main glacier. At a distance of only 24 km from this transfluence pass, the Indus gorge runs at a mere 1500 m asl, so that here, too, the ice discharge has taken place from the Chogolungma valley down into the Indus valley. This spillway and ice run-off has brought about the drop of the LGM-ice level from the centre of the upper Chogolungma valley – contrary to the actual conditions of the incline – the Haramosh valley upwards from c. 5900 m down to c. 5400 m asl in the area of the transfluence at the Haramosh La (Photo 101▬ ▬). The level of the Chogolungma parent glacier continued at 5900 m from the inflow of the Haramosh valley, following the main valley down to the SW (Photo 113 and 123 ▬ ▬).

In the uppermost chamber of the Chogolungma valley we are talking about, high-lying ground moraine deposits are only preserved in some places: on the orographic left trough flank in the area of the junction of the Moraine glacier, High- to Late Glacial (Stage 0 to IV) remnants of ground moraines (Photo 103 ■) occur, the highest one of which is situated at 4390 m, 420 m above the current glacier level (Photo 103 ■ white on the right; Figure 2/1 between No. 55 and 70). At the exit of the Sgari Byen Gang Gah further deposits of ground moraine can be noticed on the large glacigenic triangle-shaped face (Photo 107, 109, 111 ■; Figure 2/1 No. 71). The highest ones attain 4600-4700 m (Photo 107 ■

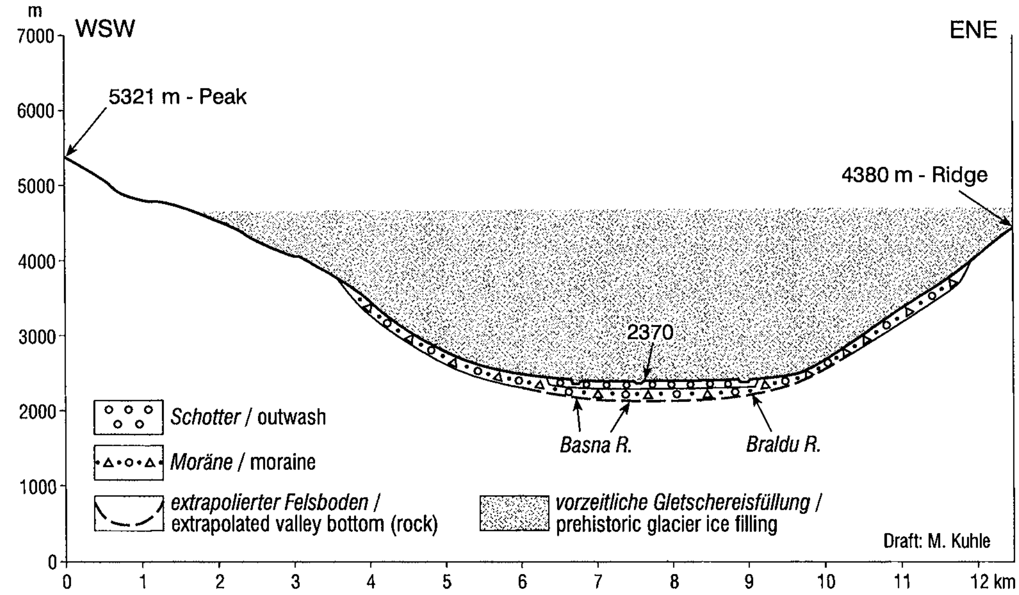


Figure 5. Cross-profile (not exaggerated) across the upper Shigar valley from the 5321 m-peak (NNW of the Munbluk) in the WSW as far as the 4380 m-ridge (NNW of the Busper) in the ENE with its minimum glacier ice filling reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 51 and 47.

white), i.e. slope positions which are situated 800 m above the actual Chogolungma glacier level (3840 m asl). In a similar position the remains of ground moraine covers can be observed at at least 4450 m asl, another c. 2 km down-valley (Photo 105 ■ white on the left; 111 ■ above; Figure 2/1 No. 71). On the next triangle-shaped face down-valley (see above), at the exit of the Bolocho Gah, a ground moraine cover with erratic granite boulders (Photo 105 and 110 ■ on the right; Figure 2/1 on the right below No. 71), transported here from the Malubiting-Spantik massif, reaches up the Aren Cho SW-flank as far as a height of 4380 m (Figure 38). It is diamictic (Figure 9). However, according to Engelhardt (1973: 133; cf. Tucker 1996: 70) the sorting coefficient (So) is 1, which is to be reduced to a good 40% portion of fine sand. There is a second maximum of 5% in the clay, typical of moraines. Thus, the cumulative curve shows the characteristic bimodal course of the cumulative frequency grain size curve (Figure 9) (Dreimanis 1939,1979,1982; Dreimanis & Vagners 1971). In addition to the preforming by the coarse matrix of the bedrock, the sand portions reaching over 80% provide evidence of the short glacigenic distances of transport as they are usual for mountain ice stream networks in the vicinity of ice divides. All SiO2 grains contained are glacially crushed, other grains are partly rounded (Figure 6 sample No. 3).

3.11.1. The orographic left side valleys of the Chogolungma valley: the Sgari Byen Gang- and Bolocho valley and their maximum prehistoric glaciations with remarks on the reworking of glacigenically triangle-shaped slopes and on the development of lateral valleys (Figure 37; Figure 2/1 No. 70/71; Photo 107)

Owing to its then c. 200–300 greater thickness, the Chogolungma main glacier extended by one km with its orographic left marginal area during the historical Stages VII-IX (Tab 1), i.e. it was 1 km wider here than at present and flowed into the Sgari Byen Gang valley as far as c. 4050 m asl (Photo 107 ■ VII–IX). There, the side glacier of this valley flowed towards it (Photo 107 and 106 □ black) and has reached its margin for the last time during Stage IX. Accordingly, the Sgari Byen Gang Gans was a left tributary stream of the main glacier until Stage IX. The matrix of the orographic left historical lateral moraine (VII–IX; Figure 2/1 below No. 70 at the bottom) of the Chogolungma glacier concerned, shows a flat, but broad secondary maximum in the clay up to fine silt of c. 10% (Figure 12) and the characteristic bimodal course with a main peak (primary maximum) in the coarse silt (40%). The overall portion of c. 50% of clay and silt, for instance, corresponds to that of the Norwegian lateral- and end moraines (cf. Haldorsen 1983: 14). The small quartz-portion of the sample is remarkable (Figure 6 No. 6). The minor portion of rounded grains among 99% glacially crushed grains (Figure 6 No. 6) diagnosed in the laboratory, proves the glaciofluvial formative influence of the meltwater and the deposition of moraines far below the snow-line, which has taken place here during the earliest historical Stage VII. At that time the ELA-depression was only c. 80 m compared with today (Tab 1). How small-scale the historic lateral moraine material varies, can be recognized by comparison with the matrix of the orographic left lateral moraine (Photo 106 ■ X on the left; Figure 2/1 below No. 70 above), taken 2.8 km up the Sgari Byen Gang valley. From the previous location to this one, the lime content has decreased by 4%, the humus content by approximately 50% and the granulometric composition with its two peaks has shifted from the clay to the fine silt, i.e. from the coarse silt to the medium sand (Figure 13). Glaciofluvially rounded grains are completely lacking (Figure 6 No. 7). On the floor of the Sgari Byen Gang valley the older ground moraine is overlain by mate-

Probennr. /sample No.	Datum/date	0,2-0,6mm ausgezählte Quarzkörner/ counted quarzgrains	glazigen gebrochen/frisch verwittert;glacially crushed/fresh weathered	äolisch mattiert/fluvial gerundet;dull(aeolian)/ lustrous(fluvially polished)	Anmerkungen / remarks
1	14.07.2000/1	227	227=(100%)	0=(0%)	*gesamte Probe gebrochen* / whole sample crushed
2	16.07.2000/1	210	210=(100%)	0=(0%)	*hoher Quartzanteil, sehr scharfkantig* / high portion of quartz, very sharp crests
3	18.07.2000/1	148	148=(100%)	0=(0%)	*Quartz gebrochen,andere Körner teilw. gerundet* / Quartz crushed, other grains partly rounded
4	18.07.2000/2	45	45=(100%)	0=(0%)	*geringer Quartzanteil, komplett gebrochen* / small portion of quartz, completely crushed
5	19.07.2000/1	76	76=(100%)	0=(0%)	*s. Probe 4* / as sample 4
6	20.07.2000/1	64	63=(98,44%)	1=(1,54%)	*geringer Quartzanteil, 1Korn gerundet* / small portion of quartz, 1 grain rounded
7	22.07.2000/1	226	226=(100%)	0=(0%)	*s. Probe 1* / as sample 1
8	24.07.2000/1	147	147=(100%)	0=(0%)	*gesamte Probe sehr scharfkantig gebrochen* / whole sample crushed with very sharp crests
9	26.07.2000/1	100	100=(100%)	0=(0%)	*s. Probe 3* / as sample 3
10	27.07.2000/1	77	77=(100%)	0=(0%)	*s. Probe 4* / as sample 4
11	28.07.2000/1	51	51=(100%)	0=(0%)	*geringer Quartzanteil, sehr scharfkantig* / small portion of quartz with very sharp crests
12	30.07.2000/1	11	11=(100%)	0=(0%)	*sehr geringer Quartzanteil, andere Körner teilw. gerundet* / very small portion of quartz, other grains partly rounded
13	31.07.2000/1	62	62=(100%)	0=(0%)	*s.Probe 4* / as sample 4
14	31.07.2000/2	-	-	-	*kein Quartz enthalten* / no quartz
15	31.07.2000/3	430	427=(99,3%)	3=(0,7%)	*hoher Quartzanteil, sehr scharfkantig, wenige gerundete* / high quartz portion, very sharp crests, very few rounded grains
16	03.08.2000/1	70	69=(98,57%)	1=(1,43%)	*s.Probe 11* / as sample 11
17	03.08.2000/2	77	77=(100%)	0=(0%)	*homogene Probe, Quartz scharfkantig* / homogenious sample, sharp crests
18	03.08.2000/3	13	13=(100%)	0=(0%)	*sehr wenig Quartz, gebrochen, viele Pflanzenreste* / very small portion of quartz, crushed, many particles of plants
19	06.08.2000/1	155	155=(100%)	0=(0%)	*sehr homogene Probe, scharfkantig* / very homogenious sample, sharp crests
20	09.08.2000/1	61	61=(100%)	0=(0%)	*s. Probe 4* / as sample 4
21	11.08.2000/1	245	237=(96,74%)	8=(3,26%)	*hoher Quartzanteil, wenige kantengerundete* / high portion of quartz, few grains with rounded edges
22	15.08.2000/1	148	148=(100%)	0=(0%)	*gesamte Probe sehr schrarfkantig* / whole sample with very sharp crests
23	17.08.2000/1	15	15=(100%)	0=(0%)	*sehr geringer Quartzanteil, gesamte Probe scharfkantig* / very small portion of quartz, whole sample sharp crests
24	17.08.2000/2	58	56=(96,55%)	2=(3,45%)	*geringer Quartzanteil, wenige gerundete Quartzkörner* / small portion of quartz, few rounded quartzgrains
25	19.08.2000/1	58	58=(100%)	0=(0%)	*s.Probe 4* / as sample 4
26	22.08.2000/1	126	126=(100%)	0=(0%)	*s. Probe 22* / as sample 22
27	23.08.2000/1	355	350=(98,6%)	5=(1,4%)	*sehr hoher Quartzanteil, sehr splitterig, wenige gerundete* / very high portion of quartz, very sharp edges, few rounded grains
28	25.08.2000/1	97	95=(98%)	2=(2%)	*Quartzkörner bis auf wenige scharfkantig gebrochen* / quartzgrains sharp edged, few rounded
29	25.08.2000/2	114	114=(100%)	0=(0%)	*s. Probe 26* / as sample 26
30	28.08.2000/1	609	605=(99,34%)	4=(0,66%)	*s.Probe 27* / as sample 27

Laboranalyse (Mikroskopie) - laboratory analysis (microscopy): N.Schroeder 04.05.2001
Probenentnahme – sampling: M. Kuhle

Figure 6. Morphometric quartz grain analysis of 30 representative samples from Haramosh Muztagh (Chogolungma-Basna valley), Shigar valley, Skardu basin, Deosai plateau, Malubiting-Spantik group (Barpu glacier, Rash lake), Batura-Campire Dior group (Bar valley) (cf. Figures 7–36).

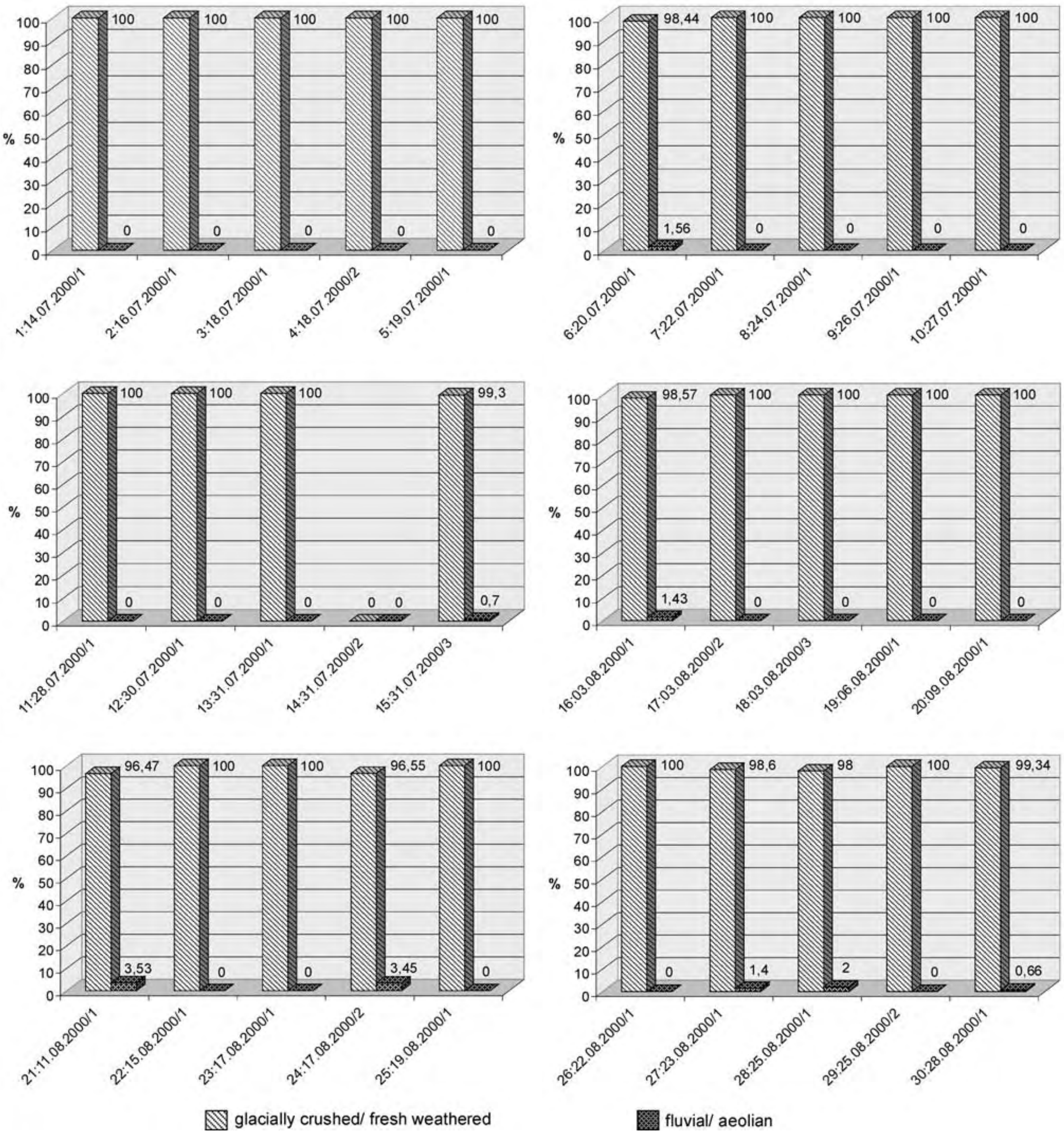


Figure 6. Continued.

rial, which has been redeposited and mixed up many times. This is evidenced by the rounded quartzite boulders, showing perfectly preserved striations, and the edged detritus (Photo 108). Thus, with regard to the reconstruction of the maximum LGM ice infilling of this valley relief, it is futile to record and discuss the sedimentological details which have taken place in the meantime.

The continuous redeposition of high-lying, i.e. High- to Late Glacial ground moraines on the valley slopes as a result of the Holocene to historic and present-day undercutting of the post-Late Glacial glaciers (Photo 106 and 107 ■ black) and their meltwaters (Photo 106 ○) has to be established.

In the Sgari Byen Gang valley these ground moraines stretch as far as a slope height of at least 4800 m asl (Photo 106 ■ black on the right and white; Figure 2/1 No. 71; Figure 37), whilst the glacigenic flank abrasions reach the slope culminations, i.e. the intermediate valley ridges on both valley sides (Photo 106 ☁ on the very left and right; 107 ☁ on the left and right below No. 70; Figure 2/1 No. 70). This local observation proves an LGM ice level reaching beyond the two intermediate valley ridges, and at

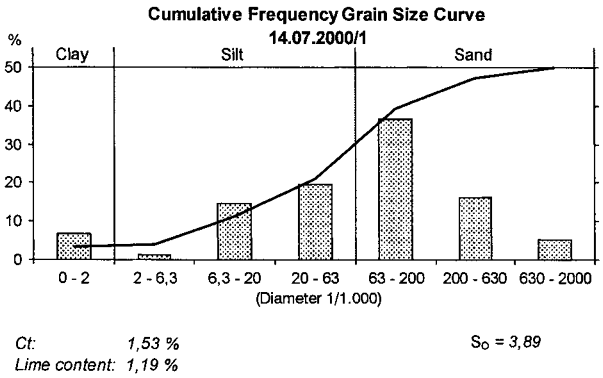


Figure 7. At 3550 m asl on the orographic left flank of the Chogolungma valley, ground moraine matrix (Photo 119 ■ black on the right) taken 250 m above the current glacier from a depth of 0.3 m. The superficial moraine layer has been sedimented during Stage IV (Tab. 1). The primary maximum is relatively coarse-grained, i.e. it lies with 40% in the fine sand; the secondary maximum with 8% clay determines the bimodal course of the cumulative curve; sorting coefficient calculated according to Engelhardt (1973: 133) So = 3.89. Locality: Fig. 2/1 No. 59 (35°53′10″ N/75°13′17″ E); see also Fig. 6 No. 1. Sampling: M.Kuhle.

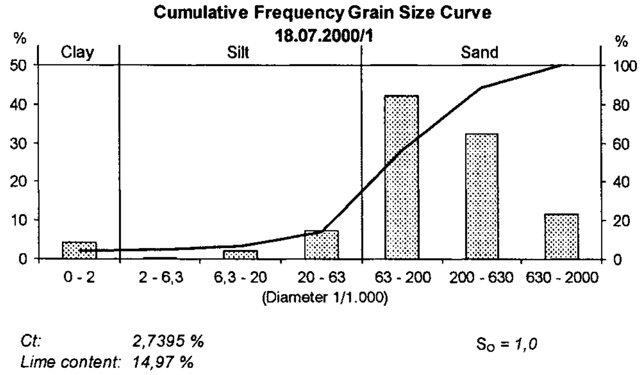


Figure 9. At 3800 m asl (aneroid measurement) ground moraine matrix from the orographic left valley flank above the Chogolungma glacier in the junction area with the Bolocho Gah, taken from 0.3 m below surface. The moraine contains erratic granite boulders and overlies metamorphic sedimentary bedrock (phyllites). The clay portion, which despite the superficial sampling was c. 5%, as well as the bimodal course of the cumulative curve confirm the morainic character; sorting coefficient calculated according to Engelhardt 1973: 133 So = 1. Locality: Fig. 2/1 No. 71, 35°57′55″ N/75°07′35″ E; see also Fig. 6 No. 3. Sampling: M.Kuhle.

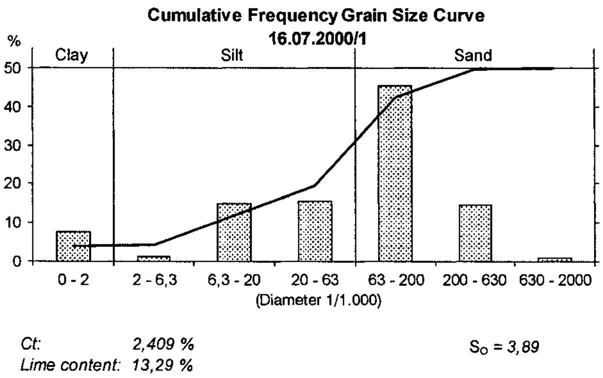


Figure 8. From the orographic right flank of the Kilwari Nala (left side valley of the Chogolungma valley) in the area of its inflow into the main valley, ground moraine matrix taken from 12 m below surface from the exposure wall situated 175 m above the current Chogolungma glacier (Photo 117 ■ black on the right). The moraine is diamictic but slightly stratified; it contains polymict, edged and facetted boulders of partly erratic materials and overlies the granite bedrock. The primary maximum is relatively coarse-grained, i.e. it lies in the fine sand (45%); the secondary one is in the clay (8%). Thus, the bimodal course of the cumulative curve confirms the morainic character. Sorting coefficient calculated according to Engelhardt 1973: 133 So = 3.89. Locality: Fig. 2/1 on the left of No. 59 (35°56′03″ N/75°10′50″ E, 3650 m asl); see also Fig. 6 No. 2. Sampling: M.Kuhle.

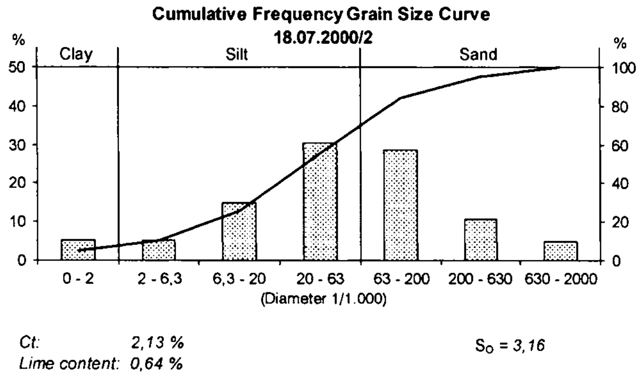


Figure 10. At 3850 m asl, ground moraine matrix from the orographic left flank of the Chogolungma valley, c. 200 m above the current glacier surface up-valley of the locality (yak pasture) Aren Cho (Arinchu) taken from 0.3 m below surface. The moraine matrix contains erratic, facetted and rounded quartzite boulders and overlies thinly stratified bedrock schists (Photo 114 in the foreground). The relatively large clay portion of c. 6% – though the sample was taken close to the surface –, as well as the bimodal course of the curve with a second peak in the coarse silt are evidence of the morainic character. Sorting coefficient So = 3.16. Locality: Fig. 2/1 between No. 71 and 59, 35°57′10″ N/75°08′25″ E; see also Fig. 6 No. 4. Sampling: M.Kuhle.

the same time partly confirms the altitude of the ice level about 6000 m asl (Figure 37) reconstructed here, which has been established by the data of adjacent, over 6000 m-high field positions (see above).

In the parallel, orographic left side valley of the Chogolungma glacier, the Bolocho valley, the glaciogemorphological situation is similar (Figure 37). The current glacier tongue is heavily buried by scree, i.e. covered with surface moraine (Photo 112 □), a large portion of which consists of redeposited Ice Age ground moraine (■ on the left; Figure 2/1 No. 71). The material has been interlaid in neoglacial to historical moraines (■ X and ■ on the right). The matrix of the lateral moraine X of the stage of redeposition show Figures 11 and 6 (No. 5): the twice as high lime content (35.84%) against the Sgari Byen Gang valley indicates a local moraine character dependent on the increasing occurrence of limestone bedrock. Towards the valley exit, i.e. in the direction of the Chogolungma glacier, the lime content already decreases significantly (c. 60%) (cf. Figure 9). As far as 4700 m ground moraine has been preserved on the orographic right valley flank (■ below – – centre); above, up- and down-valley, as well as on the orographic left flank, High- to Late Glacial band polishings and -abrasions of outcropping edges of the strata have been observed (▲; Photo 110 ▲, Figure 2/1 on the right below No. 70). In the upper catchment area and valley head the 'Bolocho Pinnacles' (No. 70), undercut and sharpened by the Late Glacial to postglacial glacier ice, have been totally covered and overflowed by the High Glacial (LGM = Stage 0, Tab 1) ice stream network (Photo 112 – – 0). The same applies to the

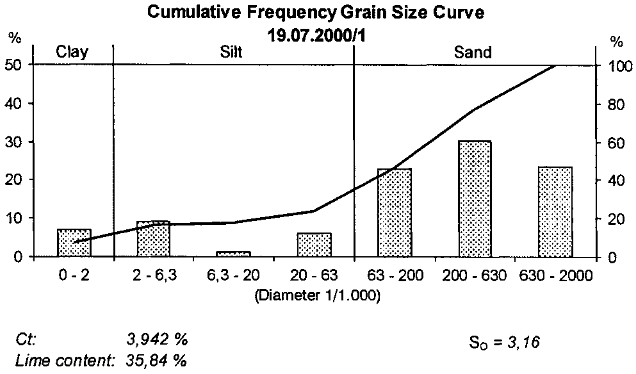


Figure 11. At 4150 m asl (aneroid-measurement) ground- i.e. lateral moraine matrix from the historic, orographic left lateral moraine inner slope (Stage VII–IX) of the Bolocho glacier, taken from 0.4 m below surface. The moraine contains polymict boulders (Photo 112 ○ white) of sedimentary rock and granite mixed with shard-like debris. Locality: Photo 112 ■ black on the right; sorting coefficient So = 3.16. Fig. 2/1 next to No. 71, 35°59′10″ N/75°09′20″ E; see Fig. 6 No. 5. Sampling: M.Kuhle.

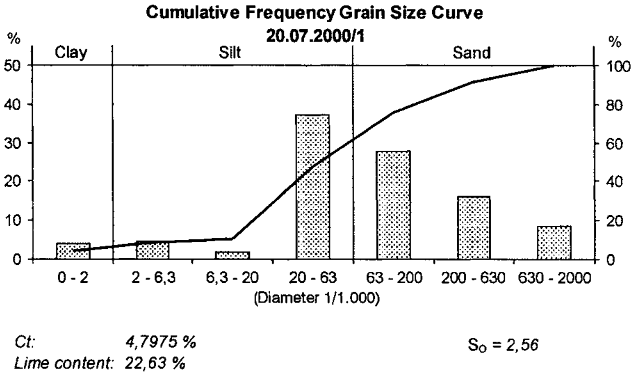


Figure 12. At 3920 m asl (aneroid-measurement), lateral moraine matrix from the historic orographic left lateral moraine (Stage VII–IX) of the Chogolungma glacier, taken from 0.3 m below surface. The moraine contains erratic granite boulders from the Malubiting- and Spantik massif; it overlies thinly stratified sedimentary bedrocks as e.g. schists. Sorting coefficient So = 2.56. Locality: Photo 107 ■ black on the right; Fig. 2/1 below No. 70 bottom, 35°59′03″ N/75°05′35″ E; see Fig. 6 No. 6. Sampling: M.Kuhle.

two valley flank culminations of the Bolocho valley (– – centre; Photo 110 – –; Photo 107 – – on the right), so that here a continuous ice level can be evidenced as far as somewhat above the Aren Cho (No. 71) (Figure 37).

In a similar way as the flanks of the two side valleys under discussion have been reworked and roughened by crumblings with rill-development (Photo 106 and 112 above ◁), the large, glacigenically triangle-shaped face (Photo 105, 3 ▽), situated in the Chogolungma main valley between the two side valley inflows, has also undergone a roughening (Photo 111 ▽, ⇩). Because the observations described immediately make clear, that the LGM-ice-level has considerably towered above the 5310 m-high ridge (Photo 111 – –; Figure 37), the development of the wall gorges and the furrowing during the last 20 Ka must have taken place syngenetically with the thawing-out of the ice stream network. The youngest denudation processes of this sort occurred by means of rock-falls and the discharge of mudflows and avalanches into the historic lateral valley at the foot of the

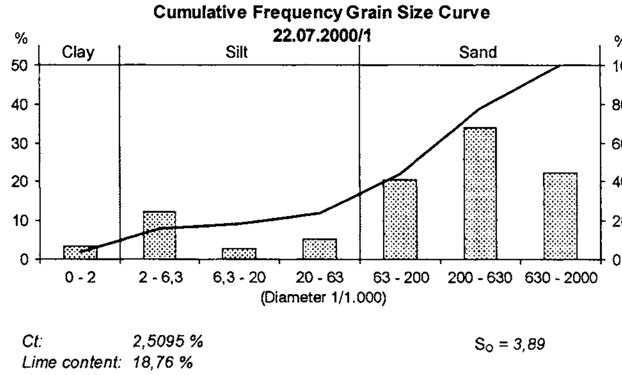


Figure 13. At 4260 m asl (aneroid-measurement), side valley of the Chogolungma. Moraine matrix taken from a depth of 0.3 m from the orographic left, historic lateral moraine of the Sgari Byen Gang Gans (glacier), which has been deposited during the Younger Dhaulagiri Stage (Stage X, Tab. 1) (Photo 106 ■ X on the left). The lateral moraine is composed of autochthonous metamorphic sedimentary rock debris (crystalline schists, phyllites) with portions of quartzite and contains quartzite boulders. The relatively hard, coarse grains of this parent rock determine the very coarse matrix with c. 80% sand and a fine grain peak in the silt. Sorting coefficient So = 3.16. Locality: Fig. 21 below No. 70 top; 35°59′55″ N/75°06′20″ E; see Fig. 6 No. 7. Sampling: M.Kuhle.

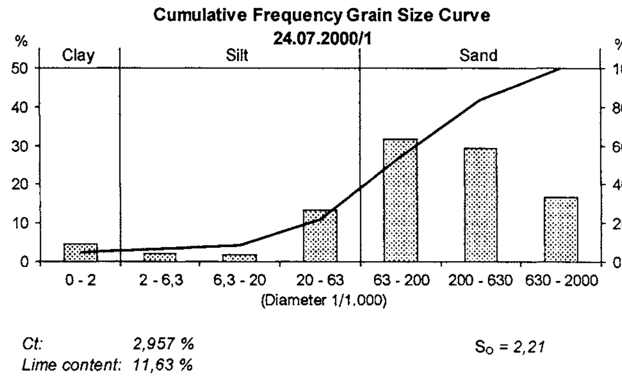


Figure 14. Moraine matrix taken from 0.3 m below surface from the orographic left historic lateral moraine crest of the Chogolungma glacier, which has been deposited during the Younger Dhaulagiri Stage (Stage X, Tab 1) and still clings to the present-day glacier (Photo 111 ■ X). The skeleton portion of the lateral moraine is made up from polymict metamorphic debris of sedimentary rock (crystalline schists, phyllites) with metres-sized boulders of granite and quartzite (Photo 109 ■ on the very right). The relatively hard, large grains of the matrix of this parent rock are the reason for this coarse matrix with c. 75% sand, the fine portion of which makes up the primary maximum. Sorting coefficient So = 2.21. Locality: 3840 m asl; Photo 105 on the lateral moraine ramp ■ X; Fig. 2/1 between No. 71 and 57; 35°58′48″ N/75°05′10″ E; see Fig. 6 No. 8. Sampling: M.Kuhle.

triangle-shaped face (Photo 109). The chronological- geomorphological sequence was as follows: 1. upthrust and deposition of the orographic left lateral moraine of the Chogolungma glacier at an ELA-depression of 30–40 m compared with today during the younger Dhaulagiri Stage X (■ on the very right; Photo 105 and 111 ■ X); 2. glaciolimnical accumulation of the sediments in a lateral lake (Photo 109 □); 3. discharge of the lake and dissection of the lake sediments; 4. build-up of a lateral sander (outwash), i.e. a gravel floor as a lateral valley gravel bottom (○); 5. mudflow- (△) and avalanche cones (◊), adjusted to them, are coming down from the wall gorges (⇩) every year. Accordingly, this 5-phase development has taken place in

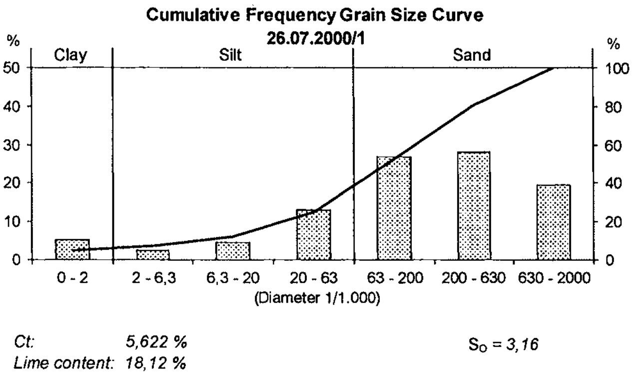


Figure 15. Moraine matrix taken from a depth of 0.3 m from the orographic left historical lateral moraine crest of the Chogolungma glacier, which was deposited during the Younger Dhaulagiri Stage (Stage X, Tab 1) and still clings to the present-day glacier (Photo 114 X). The skeleton portion of the lateral moraine is composed of polymict metamorphic debris of sedimentary rock (crystalline schists, phyllites) with metres-sized boulders of quartzite and granite. The relatively hard, large matrix grains of this parent rock are cause of this coarse matrix with c. 80% sand, the medium sand portion of which forms an only little prominent primary maximum. The sorting coefficient So = 3.16 points to a relatively diamictic matrix. Locality: Photo 114 on the lateral moraine ramp next to X; Fig. 2/1 between No. 59 and 56 (35°56′50″ N/75°08′45″ E; 3690 m asl); see Fig. 6 No. 9. Sampling: M.Kuhle.

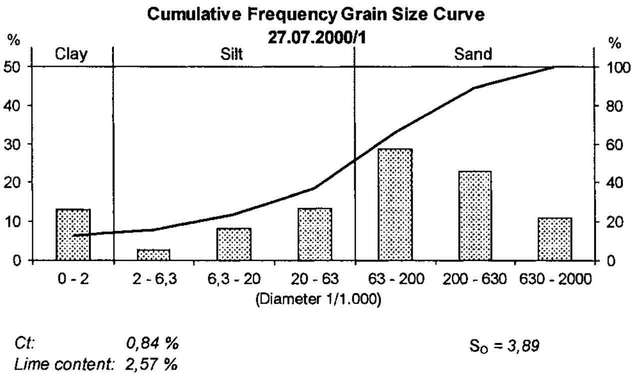


Figure 16. Ground moraine matrix taken from 1.2 m below surface from the orographic right flank of the Chogolungma glacier valley, i.e. Basna valley, 730 m above the present-day surface of the Chogolungma glacier (Photo 126 first white ■ from the right). The skeleton portion of the moraine is composed of polymict rock debris (crystalline schists, phyllites) with metres-sized boulders of quartzite, erratic granite and sedimentary rock. 60% of the matrix grains consist of sand; the fine sand portion makes up a primary maximum which is only little prominent. The sorting coefficient So = 3.89 indicates a relatively diamictic matrix. A secondary maximum lies in the clay (14%) and determines the bimodal grain size distribution. Locality: Fig. 2/1 above No. 60, near the pasture of Guma at 35°51′05″ N/75°18′30″ E, 3630 m asl; see Fig. 6 No. 10. Sampling M.Kuhle.

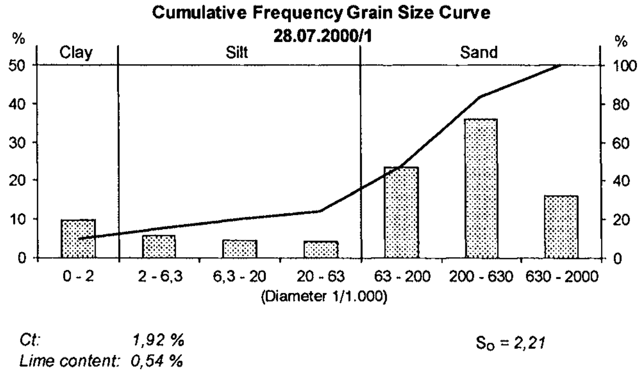


Figure 17. Ground moraine matrix taken from a depth of 2.1 m from the orographic right flank of the Chogolungma glacier valley, i.e. Basna valley, 900 m above the current surface of the Chogolungma glacier (Photo 126 below ▲ from the right; Photo 124 on the right below); the skeleton portion of the moraine is built-up from polymict rock debris (crystalline schist, phyllite) with metres-sized local boulders of quartzite and erratic sedimentary rock. The matrix grains consist of a 75% sand portion; the portion of medium sand shows a clear primary maximum. It is coarse-grained (lean) moraine, transported over a relatively short distance, i.e. local moraine. Evidence of this is also provided by the comparatively low lime content. The sorting coefficient So = 2.21 stands for a relatively homogeneous matrix. A secondary maximum of 10% in the clay causes the typical distribution of the grain sizes with two peaks. Altogether the admixture of residual detritus becomes recognizable. Locality: Fig. 40, Fig. 2/1 above No. 60; above the Guma pasture at 35°51′01″ N/75°18′20″ E; 3800 m asl; see Fig. 6 No. 11. Sampling: M.Kuhle.

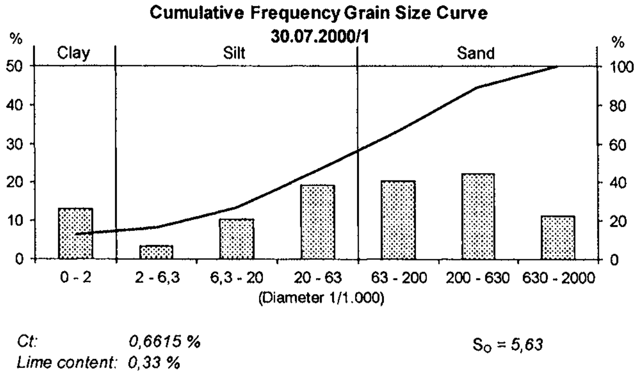
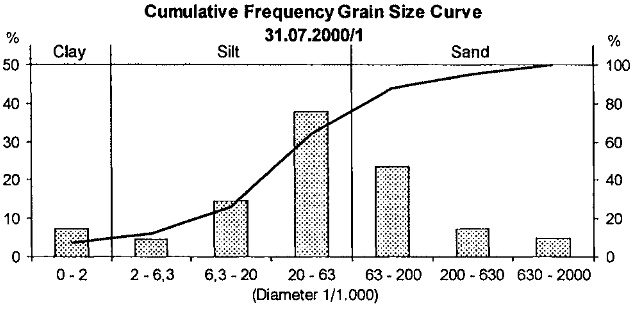


Figure 18. Late Glacial ground moraine matrix of the Sirkung Stage (IV) taken from a depth of 0.2 m from the orographic right flank of the Basna valley, c. 600 m above the talweg of this main valley (sampling at the location depicted in Photo 136). The skeleton portion of the moraine contains metres-sized erratic granite boulders, superimposed upon sedimentary bedrock. The bimodal cumulative curve runs very flat, i.e. a compact primary maximum reaches from the medium sand (22%) via the fine sand (21%) as far as the coarse silt (20%); the clay peak reaches 13%. Correspondingly, the sorting coefficient So = 5.63 indicates a dispersed matrix. Locality: Fig. 42, Fig. 2/1 on the right of No. 60, above the settlement of Tisa Birri at 35°47′20″ N/ 75°22′55″ E, 3230 m asl; see Fig. 6 No. 12. Sampling: M.Kuhle.

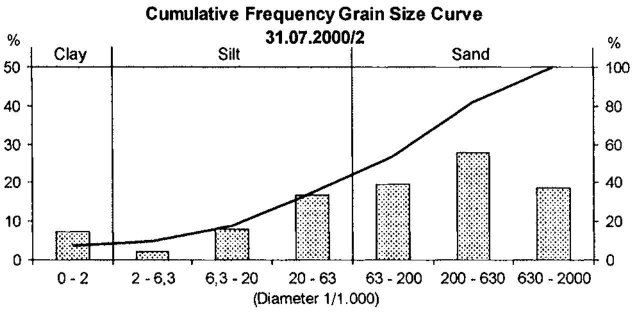
the course of c. 230 years (Tab. 1). There might have been earlier reshapings of this lateral valley – at present no longer verifiable –, but its very first build-up cannot be older than neoglacial, probably Stage 'VII (middle Dhaulagiri Stage c. 2000 YBP). Before, the ELA-depression was over 150 m and the Chogolungma glacier was unable to develop a lateral moraine here.

The lateral moraine of Stage X (Photo 109 ■ on the right; Photo 111 ■ X) clings to the current Chogolungma glacier and, correspondingly, is doubtless of glacigenic origin. In this connection the characteristics of its matrix are especially interesting (Figure 14). The sorting coefficient (So=$\sqrt{Q3/Q1}$: Q1 = grain sizes at the quartile of the grain size cumulative curve at 25%; Q3 = grain sizes of the quartile at 75%) is namely only 2.21 and – without considering the bimodal course of the cumulative curve at a secondary maximum with 5% clay – does not apply to the diamictic character of a matrix typical of moraines. This is an example for the fact that there must be quoted always more than only one criterion. Thus, the morphometric analysis (Figure 6 No. 8) with 100% very sharply-edged crushed SiO2-grains,



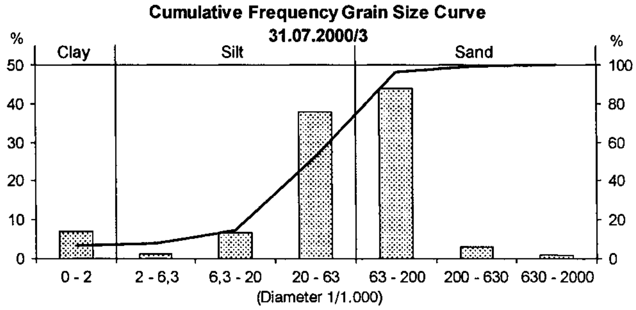
Ct: 5,464 % $S_o = 1,78$
Lime content: 16,99 %

Figure 19. Matrix of Late Glacial ground moraine of the Sirkung Stage (IV) which has been redeposited downslopes several times, taken from an exposure 3.0 m below the slope surface from the orographic right flank of the Basna valley, c. 30 m above the current gravel floor of this main valley (Photo 138 ○); the skeleton portion of the moraine contains polymict boulders, among them metres-sized erratic granite boulders, so that the long-distance transport of the matrix down-valley is also documented. The arrangement of the columnar diagram is bimodal and the cumulative curve relatively steep; this is caused by a prominent primary maximum in the coarse silt (38%). In the comparatively minor secondary clay maximum (8–9%) the secondary denudative redeposition down-slope is objectivized; this is also indicated by the low sorting coefficient So = 1.78. Locality: Fig. 43; Fig. 2/1 above No. 51; E of the settlement of Chutran at 35°41′30″ N/75°25′40″ E, 2410 m asl; see Fig. 6 No. 13. Sampling: M.Kuhle.



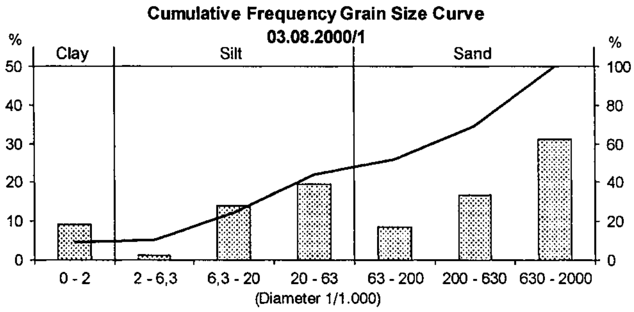
Ct: 0,375 % $S_o = 3,16$
Lime content: 2,43 %

Figure 20. Matrix from rock avalanche material 65 m above the Shigar river on the orographic left valley side, taken from an exposure wall 2.3 m below the slope surface (Photo 141 on the left below No. 50). The skeleton portion of the rock avalanche contains metres-sized, edged boulders, which Hewitt (1999, after Zanettin 1964) describes as tonalites (metamorphic clay rocks). This provides evidence of a cross-valley transport from the orographic right valley flank. The arrangement of the columnar diagram is bimodal because of the substantial clay portion of that source rock. The primary maximum lies in the medium sand (a good 27%), which could also apply to lateral moraines; the sorting coefficient So = 3.16 is also reminiscent of moraine. Locality: Fig. 2/1 between No. 51 and 147; WSW of the Mungo settlement at 35°38′30″ N/75°32′40″ E; 2400 m asl; see Fig. 6 No. 14. Sampling: M.Kuhle.



Ct: 0,62 % $S_o = 2,19$
Lime content: 3,16 %

Figure 21. Late Glacial (Stage IV) ground moraine matrix the surface of which has been slightly glaciofluvially washed, taken from a depth of 0.5 m. It is situated at the lower end of the Shigar valley in the region of the Strongdokmo La, a confluence saddle lying 110 m above the Shigar river between the left Shigar valley flank and the Indus valley (Skardu Basin) (Photo 144 □). The skeleton portion of the moraine contains polymict boulders, among them erratic granite boulders up to metres in size. This provides evidence of the down-valley long-distance transport of the matrix. The arrangement of the columnar diagram is bimodal, the cumulative curve is relatively steep, caused by a very prominent primary maximum in the coarse silt (38%) to fine sand (44%). In the comparatively minor secondary clay maximum (8%) the slight secondary redeposition is objectivized. Sorting coefficient So = 2.19. Locality: Fig. 2/1 on the left of No. 73 (2310 m asl; 35°21′20″ N/75°45′ E), see Fig. 6 No. 15. Sampling: M.Kuhle.



Ct: 1,01 % $S_o = 5,62$
Lime content: 0,69 %

Figure 22. Late Glacial (Stage IV) end moraine matrix in the middle Satpare Lungma (Photo 159 ■ IV) taken from 0.5 m below the surface. The skeleton portion of the decametres-thick accumulation contains granite boulders some metres in size. Due to the thickness of the sediment the genesis of the matrix as a result of in situ weathering of the bedrock can be ruled out, i.e. horizontal transport has taken place. The arrangement of the columnar diagram is even trimodal. This depends on a prominent secondary maximum in the coarse silt (c. 20%) between a primary maximum in the coarse sand (c. 30%, deriving from the granite grit) and a tertiary in the clay (just 10%); correspondingly important is the sorting coefficient So = 5.62. Locality: Fig. 2/1 half-left below No. 77 (35°08′45″ N/75°37′01″ E; 3210 m asl), see Fig. 6 No. 16. Sampling: M.Kuhle.

undoubtedly proves the glacigenic origin, too. The clear similarity of all sedimentological characteristics of this moraine still clinging to the ice, with the ground masses taken in the Bolocho valley (Figure 11 and 6 No. 5; Figure 9 and 6 No. 3) and Sgari Byen Gang valley (Figure 12 and 6 No. 6; Figure 13 and 6 Nr.7) give further proof of its morainic character.

The observations from the uppermost Basna- i.e. Chogolungma valley are to be summarized as follows: In the area of the valley head the LGM-glaciation (Stage 0) of the type ice stream network (Figure 2/1 No. 52-55) has had a maximum ice level at 6400–6200 m asl (Photo 113 and 123 — — right quarter). This ice level has communicated over northern, western and southern transfluence passes with the ice stream networks of the adjacent main valleys: the Hispar valley, the Hunza valley and the Indus valley. As far as the Bolocho side valley and the Second East Haramosh glacier valley (Figure 2/1 No. 56, 57, 70, 71), the verifiable level of the LGM- glacier height decreased to c. 5900–5700 m asl (Figure 37,38; Photo 123 — — centre). Here, maximum

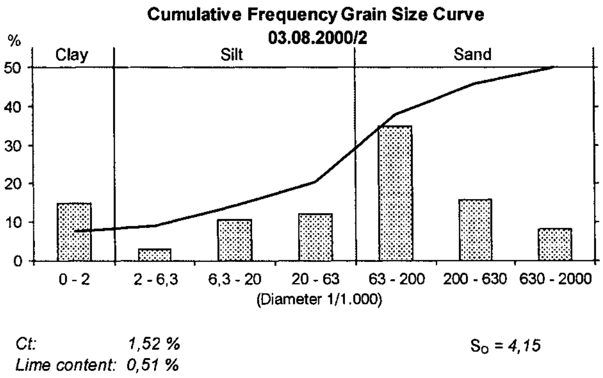


Figure 23. Ground moraine matrix on the central Deosai plateau (on the slope foot in the area depicted in Photo 162, left margin in the background) taken from 0.4 m below surface. The skeleton portion of the decametres-thick accumulation contains polymict boulders up to 0.6 m in length, so that a genesis of the matrix resulting from weathering of the bedrock in situ can be ruled out, i.e. horizontal transport has taken place. The clay- and silt-portions (c. 40%) exclude fluvial transport. The arrangement of the columnar diagram is bimodal, indicated by a primary maximum in the fine sand (c. 33%) and a prominent secondary one in the clay (c. 16%). Considering a moraine, the sorting coefficient So = 4.15 lies in a middle position. Locality: Fig. 2/1 between No. 83 and 84 (35°04'20″ N/75°29'30″ E; 4000 m asl), see Fig. 6 No. 17. Sampling: M.Kuhle.

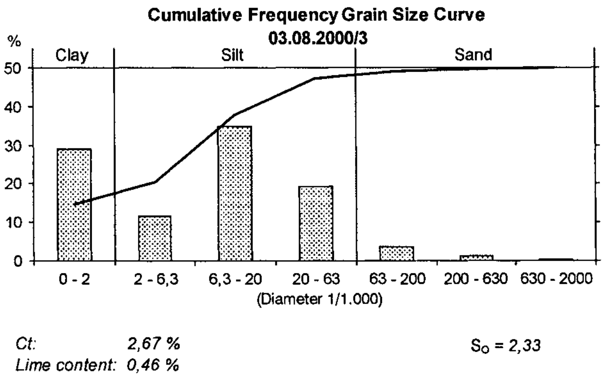


Figure 24. Ground moraine matrix on the western margin of the Deosai plateau c. 10 m above the Scheosar Tso (lake) taken from the slope at the back of the viewpoint of Photo 167 from 0.5 m below surface. The skeleton portion of the c. 2 m-thick accumulation contains polymict boulders up to 0.4 m in length. Among them are gneiss- and granite boulders which, also with regard to their round-edged to rounded forms, cannot be confused with slope debris weathered in situ; secondary solifluidal debris-shifting down-slope seems probable. The clay- and silt portions amount to c. 95% so that fluvial transport has to be ruled out. The arrangement of the columnar diagram is bimodal; this is shown by the primary maximum in the medium silt (c. 35%) and a very marked secondary maximum in the clay (29%). The sorting coefficient So = 2.33 is low, even for a heavily triturated ground moraine; the rest of the parameters argues for a glacial genesis (cf. Fig. 6 No. 18). Locality: Fig. 2/1 above No. 82 (34°59'58″ N/75°14'22″ E; 4180 m asl). Sampling: M.Kuhle.

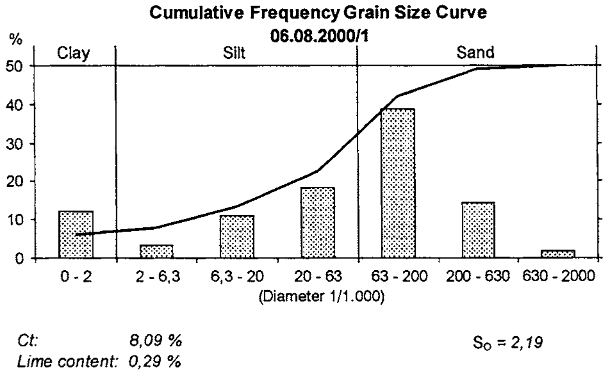


Figure 25. Moraine matrix from the neoglacial to historical (c. Stage V to IX, Tab. 1) orographic right lateral moraine of the present-day Chukutan- or Munrabu hanging glacier on the orographic right above the Barpu glacier, on the N-side of the Malubiting-massif above the Chukutan pasture, taken from 0.5 m below the surface. The skeleton portion of this decametres- to more than 100 m-thick moraine body contains large, to very large, i.e. several metres-long, edged to round-edged boulders. The portions of clay and silt amount to c. 45% and exclude a fluvial transport. The arrangement of the columnar diagram is bimodal, indicated by a primary maximum in the fine sand (c. 39%) and a secondary in the clay (13%). The sorting coefficient, calculated after Engelhardt 1973: 133, is low (So = 2.19) according to a certain sorting by meltwater. The quartz-grain forms, however, reflect only the purely glacial genesis (cf. Fig. 6 No. 19). Locality: Fig. 2/1 below No. 93 on the right (36°09'30″ N/74°54'20″ E, 4050 m asl). Sampling: M.Kuhle.

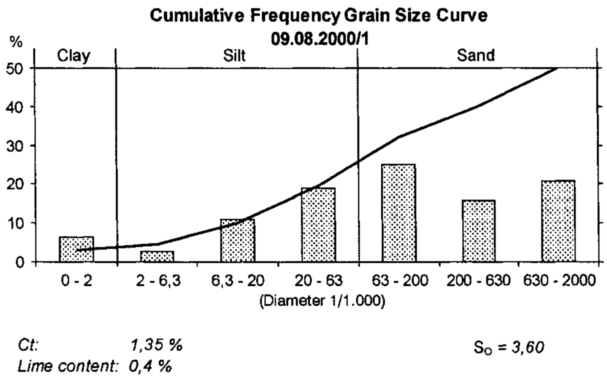
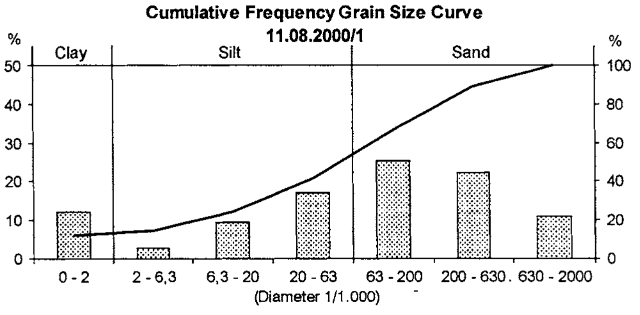


Figure 26. Fine material matrix from a neoglacial to historical (c. Stage V to IX, Tab. 1) orographic right sediment mixture of ground- and lateral moraine up to kame-like bank form of the current middle Barpu glacier, immediately WNW of the Spantik near the pasture-locality of Malenschi, taken from a depth of 0.7 m. The portions of silt and clay amount to c. 40% and exclude fluvial transport; solifluidal transport is possible. The arrangement of the columnar diagram is trimodal. This is due to a primary maximum in the fine sand (c. 24%), a secondary in the coarse sand (c. 20%) and a tertiary in the clay (c. 6.5%). The sorting coefficient So = 3.6, calculated after Engelhardt 1973: 133, is in the middle position; the quartz-grain forms indicate a purely glacial genesis (cf. Figure 6 No. 20). Locality: Fig. 2/1 on the left of No. 54 (36°06′ N/74°54′35″ E; 4200 m asl). Sampling: M.Kuhle.

local ice thicknesses about 2600 m have been attained in the middle of the main valley (Figure 38).

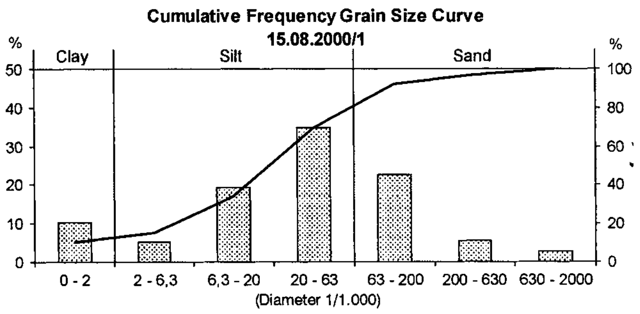
3.11.2. Continuation of the reconstruction of the maximum Ice Age valley filling with glacier ice in the area of the present-day Chogolungma glacier

In the valley chamber connected down the main valley, the cross- profile of which is depicted in Figure 39, kame accumulations of dislocated LGM-ground moraine have been preserved on the orographic left side between c. 3750 and 4000 m (Photo 105 below No. 59 on the right of ■). Kick (1956, 1964) has already described these layered depositions as young, fluvial sediments. However, he has not considered them to be glaciofluvial, i.e. glacigenic embankments, adjusted to a prehistoric glacier margin. They are relatively young and belong at most to the late Late Glacial (Stage IV), probably to the Holocene Neoglacial (Nauri Stage V, Tab 1). Nevertheless, High- to Late Glacial (Stage 0-III) ground moraine, parts of which have been displaced down-



Ct: 1,743 % S_O *= 8,78*
Lime content: 4,92 %

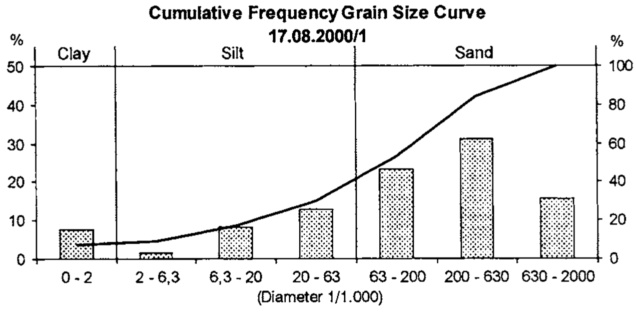
Figure 27. Matrix of earth pyramids from a metres- to decametres- thick Late Glacial to neoglacial (c. Stage IV to V, cf. Tab 1) ground moraine cover on the orographic right of the current lower Barpu glacier beyond the lateral valley SW of the Gutena pasture and the mountain ridge, which separates the Hispar- and the Barpu valley, taken from an exposure wall on the base of the ground moraine, 3 m above the bedrock in the underlying stratum. The portions of clay and silt amount to c. 34%; they exclude fluvial transport. Solifluidal transport is also impossible. Due to a primary maximum in the fine sand (c. 25%) and a secondary in the clay (c.12.5%), the arrangement of the columnar diagram is bimodal. The sorting coefficient (So = 8.78), calculated after Engelhardt 1973: 133, is very large; the quartz-grain forms, too, provide evidence of a purely glacial genesis, though 3.26% of the quartz-grains are fluvially reshaped (cf. Fig. 6 No. 21). Locality: Fig. 2/1 below No. 93 on the left (36°11′10″ N/74°52′ E; 3200 m asl). Sampling M. Kuhle.



Ct: 2,85 % S_O *= 2,78*
Lime content: 0,45 %

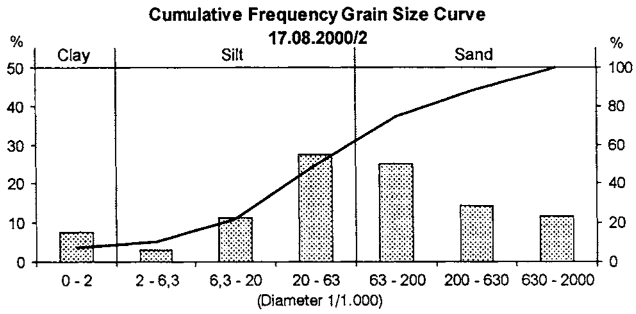
Figure 28. Matrix from an LGM- to Late Glacial (c. Stage 0 to II, cf. Tab 1) ground moraine which is only metres-thick, c. 1900 m above the valley ground on the orographic left flank of the Hispar valley, SE of the Gutena pasture, taken from 15 cm below the surface. The clay- and silt portions amount to c. 70% and exclude fluvial transport; solifluidal dislocation, however, is possible. The arrangement of the columnar diagram is bimodal, caused by a primary maximum in the coarse silt (c. 36%) and a secondary in the clay (c. 10%). The sorting coefficient So = 2.78, calculated after Engelhardt 1973: 133, is rather low. This is due to the homogeneous parent rock. The quartz-grain forms prove a purely glacigenic genesis (cf. Fig. 6 No. 22). Locality: Fig. 2/1 half-right below No. 93 (36°11′35″ N/74°52′40″ E, 4500 m asl). Sampling: M.Kuhle.

slope, has been preserved in the form of extended remnants above and at the margins (Photo 114 ↑). They reach up to an altitude of 4300 m (Figure 2/1 between No. 71 and 59). The matrix (Figure 10) shows a high portion of sand of c. 45%, a primary maximum (30%) in the coarse silt and a weak secondary maximum in the clay. All this is typical of a not far-travelled, and thus little triturated ground moraine as it can solely be taken in consideration for an ice stream network close to a valley head. The quartz grains are completely glacially crushed (Figure 6 No. 4, diagram 18.7.2000/2). The



Ct: 0,65 % S_O *= 3,16*
Lime content: 0,43 %

Figure 29. Matrix from an LGM- to Late Glacial (c. Stage 0 to III, cf. Tab. 1) ground moraine which is only metres-thick, c. 1500 m above the current Barpu glacier surface on the orographic right flank of the Barpu valley WNW of the Rash Phari (5058 m), taken from 20 cm below the surface. The clay- and silt portions amount to only c. 33%, the sand portion reaches c. 67%. This is due to the substantial portions of local moraine from the granite bedrock in the underlying stratum. Solifluidal dislocation is probable. The arrangement of the columnar diagram is bimodal, deriving from a primary maximum in the medium sand (c. 31%) and a secondary in the clay (c. 8%). The sorting coefficient So = 3.16, calculated according to Engelhardt 1973: 133, is usual as to a local moraine in homogeneous granite. The forms of the crushed quartz-grains provide evidence of a frost weathering- or glacial-, i.e. mixed genesis (cf. Fig. 6 No. 23). Locality: Fig. 2/1 on the left above No. 54 (36°00′50″ N/74°53′ E; 4775 m asl). Sampling: M.Kuhle.



Ct: 0,89 % S_O *= 3,16*
Lime content: 0,8 %

Figure 30. Matrix from a Late Glacial (c. III or IV, cf. Tab. 1) decametres-thick lateral moraine ledge on the orographic left, in which the Gutena or Gutens pasture is situated, c. 1400 m above the talweg of the Hispar valley; taken from an exposure, 5 m below the surface. The clay- and silt portions amount to c. 54%, so that fluvial transport can be ruled out. The sand portion reaches c. 46%. This is due to the portions of local moraine from the granite bedrock; solifluidal dislocation is improbable. The arrangement of the columnar diagram is bimodal, deriving from a primary maximum in the coarse silt (c.27%) and a secondary in the clay (c.8%). The sorting coefficient So = 3.16 has been calculated after Engelhardt 1973: 133. It can be explained by the trituration which a far-travelled moraine undergoes during a decakilometres-long transport. The quartz- grain forms testify to a glacial genesis; however, 3.45% are fluvially rounded, providing evidence of the work of meltwater and a position below the level of the ELA (cf. Fig. 6 No. 24). Locality: Fig. 2/1 half-left below No. 93 (36°12′ N/74°52′10″ E; 3900 m asl). Sampling: M.Kuhle.

laboratory analyses are also representative of the orographic left ground moraine covers on the valley slopes in the area and down-valley of the pasture of Arencho (Photo 114, the three ■ below No. 59; 116 ■ on the left and in the middle; 118 ■ on the very left; Figure 39). They attain c. 4400 m asl and are underlain by granite bedrock which shows glacigenic flank abrasion with glacier striae and -polishings (Photo 116

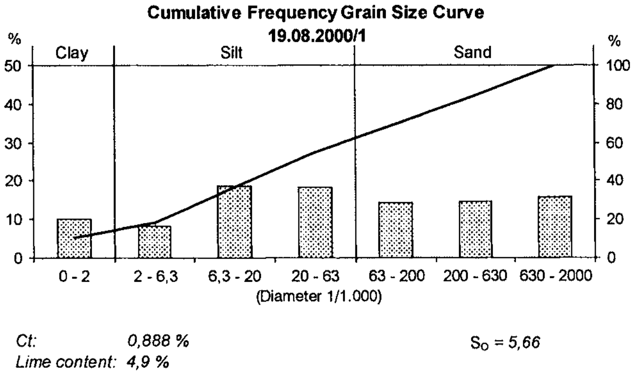


Figure 31. Matrix of a debris flow fan, which has been accumulated up to a thickness of over 100 m from a Late Glacial (c. IV, cf. Tab. 1), orographic right, c. 500 m-high mantling of ground moraine (reaching up to 2700 m asl) in the Bar Nala, in the confluence area of the Daintar Nala, after the deglaciation; the sampling has been taken from a depth of 20 cm. The portions of clay and silt amount to c. 55%, so that fluvial transport can be ruled out. The sand portion reaches c. 45%. This is due to the replacement of local moraine portions from the outcropping phyllites and gneisses. Solifluidal dislocation has to be ruled out at this insignificant sea-level. The arrangement of the columnar diagram is trimodal, deriving from a primary maximum in the medium silt (c. 19%), a very weak secondary in the coarse sand (16%) and a tertiary in the clay (c.10%). The sorting coefficient So = 5.66 has been calculated after Engelhardt 1973: 133. As for a debris flow fan, it is as normal as for a moraine. Due to the debris flow transport, 100% of the sharply-crushed, glacigenic quartz-grain forms have been preserved (cf. Fig. 6 No. 25). Locality: Fig. 2 No. 94 (36°20′50″ N/74°16′30″ E; 2190 m asl). Sampling: M.Kuhle.

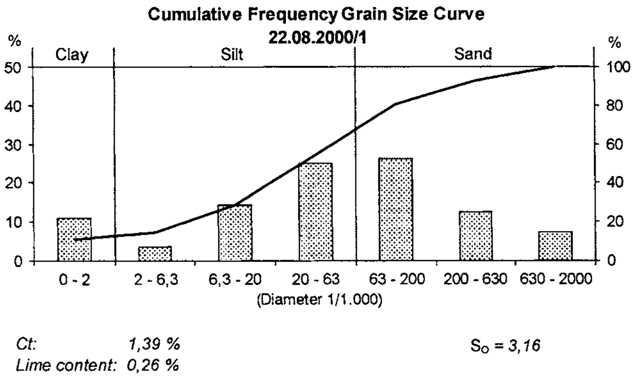


Figure 32. Matrix of the youngest Late Glacial (Stage IV, cf. Tab. 1) orographic right lateral moraine of the Kukuar glacier (Kampire Dior group, S-side), c. 280 m above the current glacier surface, taken from the inner slope from a depth of 30 cm, decametres away from the bedrock. The portions of clay and silt amount to c. 51%; fluvial transport has to be ruled out; an insignificant solifluidal dislocation of the material is possible. Typical of moraines, the arrangement of the columnar diagram is bimodal: an extended primary maximum in the coarse silt to fine sand (together 53%), a secondary in the clay (c. 11%). The sorting coefficient So = 3.16, calculated after Engelhardt 1973: 133, is inconspicuous; typical of moraine are the 100% sharply-crushed quartz-grain forms (cf. Fig. 6 No. 26). Locality: Fig. 2 No. 95 (36°28′50″ N/74°14′ E, 3250 m asl). Cf. Fig. 33. Sampling: M.Kuhle.

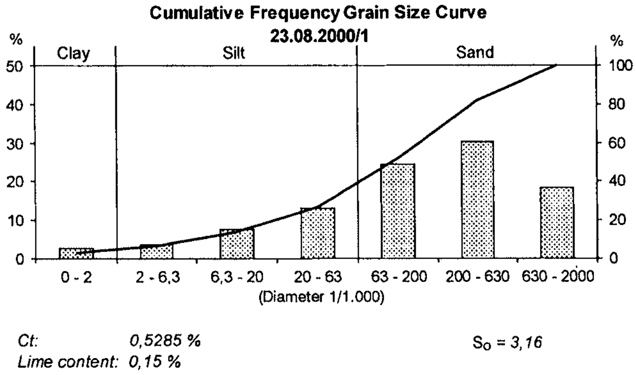


Figure 33. Debris matrix from a slope-furrow with a 50 m-wide floor containing Late Glacial portions of ground moraine (Stage III–IV, cf. Tab. 1) from the orographic right, 35°-steep valley flank slope of the upper Bar Nala, 600 m above the current surface of the Kukuar glacier (Kampire Dior group, S-side), c. 1 km S of the inflow of the Sat Maro glacier; taken from 40 cm below the surface. The portions of clay and silt amount to only c. 27%; solifluidal dislocation of the material is probable. The arrangement of the columnar diagram is monomodal, which is atypical of moraine: there is only one maximum (in the medium sand with c. 30%). The sorting coefficient So = 3.16, calculated after Engelhardt 1973: 133, is typical of moraine, as is also the large portion of sharply-edged crushed quartz-grains (98.6%). However, frost weathering on the bedrock may have played a part. Typical of the participation of fluvial transport on the slope is the 1.4% portion of rounded quartz-grains (5 out of 355 analysed examples) (cf. Fig. 6 No. 27). Cf. Fig. 32. Locality: Fig. 2 No. 95 (36°28′20″ N/74°13′50″ E, 3600 m asl). Sampling: M.Kuhle.

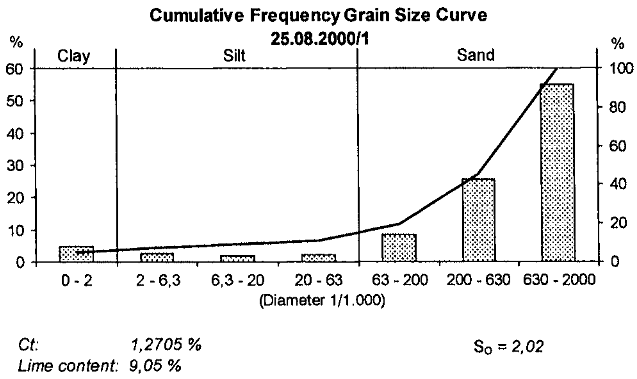


Figure 34. Ground- and lateral moraine matrix of the Late Glacial glacier filling of the Bar Nala (Stage I-III, cf. Tab. 1) in the area of the decametres-, probably over 100 m-thick, inset of the medial moraine of Toltar, at the confluence of the then Baltar- and Kukuar glacier components, 1200 m above the valley bottom of the Nar Nala (Kampire Dior group, S-side). Taken from an earth-pyramid wall, decametres away from the bedrock. The portions of clay and silt amount to only c. 10%; a 90% portion is sandy. As it occurs in the lateral valleys of the glaciers, fluvial transport is probable. The arrangement of the columnar diagram is bimodal and thus typical of moraines: a primary maximum in the coarse sand (c.54%) and a secondary in the clay (c.5%). The sorting coefficient So = 2.02, calculated after Engelhardt 1973: 133, evidences a glaciofluvial washing. There are also fluvially rounded SiO2-grains (2%); typical of moraine are the 98% sharply-edged crushed quartz-grains (cf.Fig. 6 No. 28). Cf.Fig. 35. Locality: Fig. 2 No. 95 (36°29′ N/74°18′20″ E, 3940 m asl). Sampling: M.Kuhle.

→). These polishings and glacigenic abrasions reach as far as beyond the 5178 m-Peak (Photo 114, 116 ▬ ▬ on the left; Figure 2/1 left of No. 59). On the orographic right valley flank of this cross-profile flat, small-scale ground moraine remnants are preserved on schist bedrock up to c. 4350 m asl (Figure 39; Figure 2/1 on the right above No. 58). The flank abrasions reach as far as beyond the current glaciation margins (Photo 114 ▲ below No. 58 and 61). Their upper limit is at an altitude of 5600 m and marks the prehistoric glacier level (Photo 113 ▬ ▬ below No. 58; Photo 114, 118 ▬ ▬ below No. 58 and 61 and on the very left). Including the interpolated ice thickness of c. 500 m of the here 2.3 km wide, present-day Chogolungma glacier, the LGM-parent glacier had a thickness of 2600 m (Figure 39) on this cross-profile. With regard to a glacigenic undercutting of the flanks

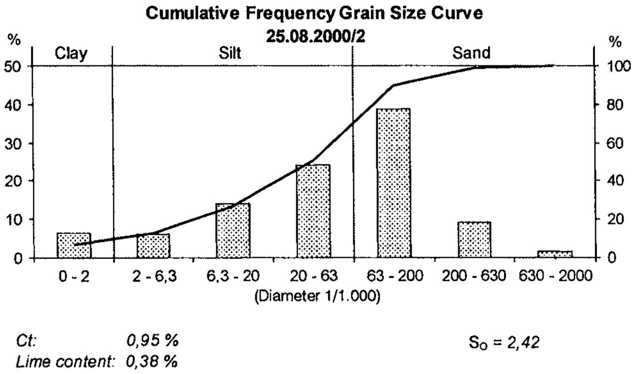


Figure 35. Ground- and lateral moraine matrix at the same locality as in Fig. 34, but 40 m lower, on the moraine slope of the Late Glacial glacier filling of the Bar Nala (Stage I–III, cf. Tab. 1), dispersed by rills into earth pyramids. Here, the portions of clay and silt are much more important; they amount to 52%. 48% of the portions are sandy; coarse- and medium sand together come to only 10%. Fluvial transport, as it occurs in lateral valleys of the glacier, is verifiable. Obviously the sample originates from the core of the lateral moraine. The arrangement of the columnar diagram is bimodal, as is typical of moraines: primary maximum in the fine sand (c.38%), a secondary in the clay (c. 7%). The sorting coefficient So = 2.42 evidences a removal of the pelites by the drainage-water of the melting ice, typical of ablation moraines. Rounded quartz-grains are lacking. Typical of moraine are the 100% quartz-grains, which are crushed sharply-edged (cf. Fig. 6 No. 29). Locality: Fig. 2 No. 95 (36°29′ N/74°18′20″ E, 3900 m asl). Cf. Fig. 34. Sampling M.Kuhle.

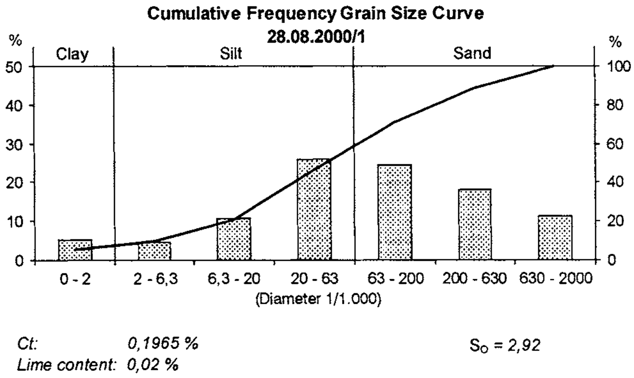


Figure 36. Matrix of the recent orographic right lateral moraine (historical Stage XI–XII, cf. Tab. 1) of the Aldar Kush glacier in the orographic right side-valley of the Bar Nala, SW of the Kutu pasture, 10 m above the glacier surface (Kampire Dior group, S-side). Taken from the lateral moraine ridge, from decametres to over 100 m above the bedrock. Despite a transport distance of a few km at maximum, the portions of clay and silt amount to c. 48%. This is due to the silt-portion of the bedrock in the catchment area. 52% of the portions are mixed-sandy; fluvial transport can be ruled out. The arrangement of the columnar diagram is bimodal, as is typical of moraines: an extended primary maximum in the coarse silt to coarse sand (together c. 50%), a secondary in the clay (c. 5%). The sorting coefficient So = 2.92 evidences a removal of the pelites by the drainage-water of the melting ice, which is typical of ablation moraines. There are 4 (out of 609 analysed) fluvial rounded quartz-grains (0.66%), out of which 99.34% are crushed sharply-edged. (Cf. Fig. 6 No. 30). Locality: Fig. 2 No. 95 (36°25′30″ N/74°18′ E, 3650 m asl). Sampling: M.Kuhle.

due to this ice level and a resulting sharpening of the summits, the Laila Peak, the Haramosh and the 6005 m-Peak are to be considered as glacial horns (Figure 2/1 No. 53, 55, 59; Photos 101, 102, 114, 115, 123). With its altitude of 5571 m, the Aren Cho has pierced the ice level only during the Late Glacial (from Stages I or II on; Tab 1) and towered above the ice stream network. Owing to this, only since that time has it been sharpened by glacigenic side abrasion (Figure 2/1 No. 71, Photo 110, Figure 37, 38).

Near the pasture of Arencho the sub-recent lateral moraine ramp (Stage X), in connection with a small breakthrough of a tributary glacier tongue (Photo 114 ⇦), has been newly buried and straightened by the younger, equally high ice margin, i.e. the left lateral moraine has been freshly remoulded during the short advance in the 20th century (Photo 114 X).

The sample of the matrix of its outer slope is relatively poor in clay (>5%) but rich in sand (approximately 80%). This confirms the minor glacigenic trituration of the matrix during the prehistorically much thicker glaciation of this valley cross- profile (Figure 15). Correspondingly important is the sorting coefficient (So) (Figure 15), i.e. the diamictic character of the matrix. 100% of the quartz grains, however, are glacially crushed (Figure 6 No. 9). The grains of other materials are partly rounded, which reflects the glaciofluvial influence 1100 m below the recent ELA. In addition to the numerous deposits of ground moraine on the orographic left, mapped and marked in detail between the pasture of Arencho and the 6005 m-Peak (Photo 114 ■ below No. 59; Photo 116, 118 ■ on the very left), particular attention ought to be focused on the polished W-spur of the 6005 m-Peak (Photo 116 ▲ black). Here, the Kilwari Nala joins (Photo 117 □ black), so that there exists an intermediate valley ridge, polished down by the prehistoric ice stream network in a glaciogeomorphologically classic way (▲ black, large); it shows a ground moraine overlay (■ white; Photo 118 second ■ from the left). Thus, the mountain spur has been glacigenically abraded and polished into a triangle-shaped face, which has in part already been worn down again postglacially by the neoglacial to historic glacigenic underpolishing (Photo 116 on both sides of ▲ black). In this respect the comparison of this older triangle-shaped face with a younger one in the Kilwari Nala, currently undercut by the confluence of the two glacier branches of the Kilwari Nala, is interesting (Photo 117 ▲ on the very left; Figure 2/1 on the left above and below No. 59). The latter one is younger, but has been roughened and reworked much more heavily by undercutting and obsequent rill rinsing as well as frost weathering in the outcropping edges of the strata of the sedimentary bedrock series.

At the exit of the Kilwara Nala, i.e. the orographic left flank of the Chogolungma valley, samples have been taken from a c. 40 m-thick ground moraine complex (Photo 117 ■ black; Figure 8). The sorting coefficient (So) 3.89 shows the diamictic character as far as into the ground mass. Not only the morphoscopic analysis (Figure 6 No. 2), but also the transport over an average distance of 30–40 km at a very important thickness of the valley glacier, already causes a trituration which is so strong that the growth in matrix has clearly increased compared with the moraines investigated near the valley head, so that its total portion predominates (cf. Ehlers 1994: 43). As to mountain glaciers, naturally the continuous intake of new material is so important that the minor portions of pebbles of only 10% (cf. Nielsen, 1983: 194), as they are typical of far-travelled moraines of inland

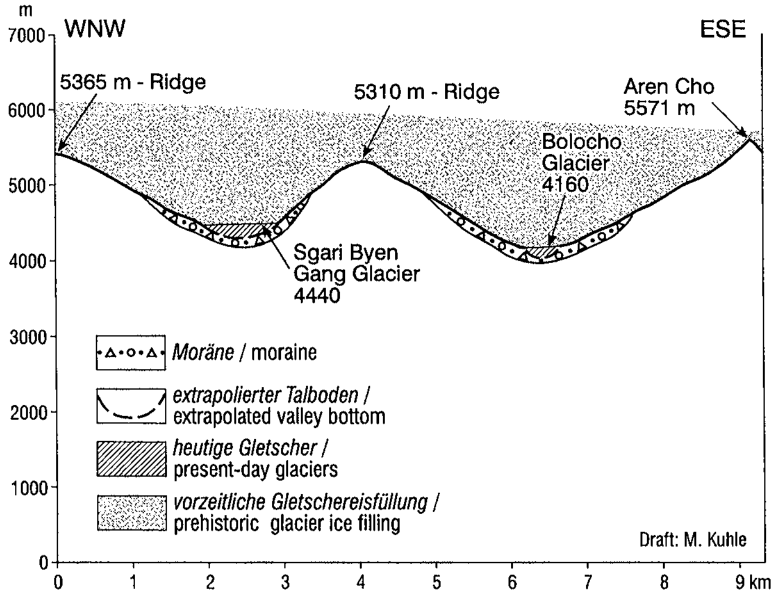


Figure 37. Cross-profile (not exaggerated) across the middle Sgari Byen Gang valley from the 5365 m-mountain ridge in the WNW via the 5310 m-mountain ridge and across the middle Bolocho valley as far as the Aren Cho summit in the ESE with the minimum glacier ice filling of this relief reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 70 and 71.

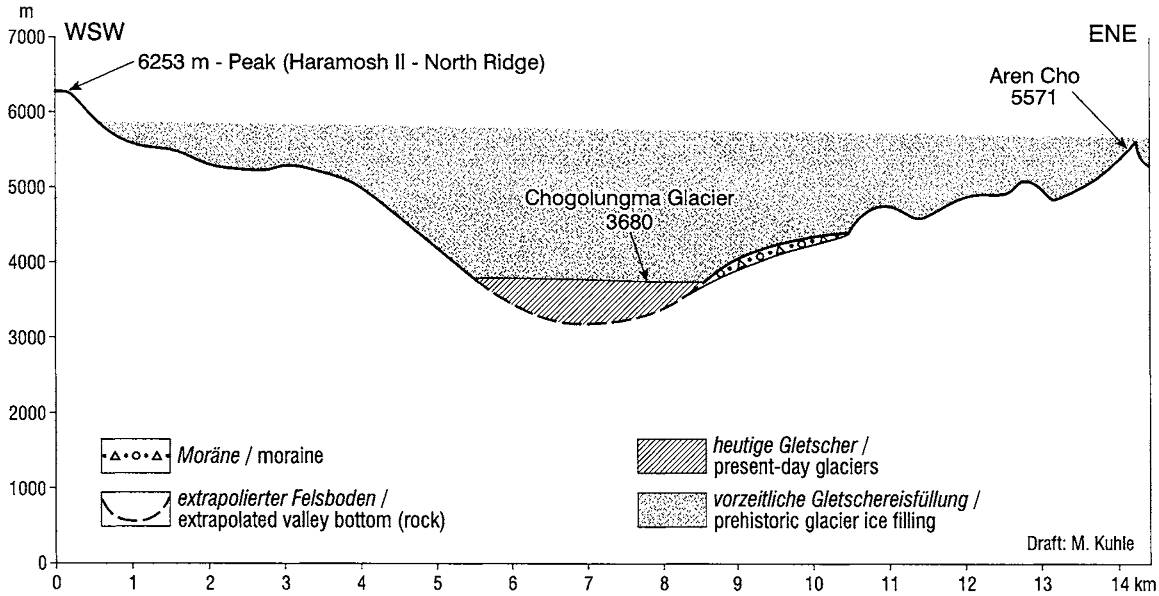


Figure 38. Cross-profile (not exaggerated) across the upper Basna valley with the middle Chogolungma glacier from the 6253 m-peak in the WSW on the N-ridge of the Haramosh II towards ENE as far as the Aren Cho on the orographic left valley side with the minimum glacier ice filling of this main valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 between No. 71 and 56.

ices, could not have been realised in the LGM-Karakorum ice stream network.

The shape of the Ice Age slopes becomes not only blurred by the undercutting of glacigenic flank forms through current glacier margins, but also by a wealth of annual morphodynamics in the lateral valleys. The orographic left lateral valley near the locality of Khurumal (Photo 118), where mudflow cones from ground moraine, dislocated down-slopes (■ on the left of No. 60), have dammed up lateral moraine lakes (↓ ↓) and gravel floors from lateral creeks (lateral sanders) (○) etc. is exemplary. Due to the fact that the backward erosion has not yet reached them, the older ground moraine covers of the upper slopes remain in a better shape than that of the lower slopes (below the two black ■ on the left), which are already heavily decomposed. The base under these ground moraine covers on the slopes are the glacigenically triangle-shaped faces of backward-abraded rock ribs (Figure 2/1 No. 59). In places, the abraded rock has already been denuded from its moraine overlay, but is still unweathered and rounded (next to ■ below No. 59 and ◥ on the left of No. 60). The occurrence of ground moraines and well-preserved flank abrasions indicate a Late Glacial ice level (Stages I to III, cf. Tab 1) about 5000 m asl (▬ ▬ white), whilst clearly roughened older abrasions testify an LGM ice level (Stage 0) between 5500 and 5700 m (▬ ▬ black; Photo 116 ▬ ▬ between No. 59

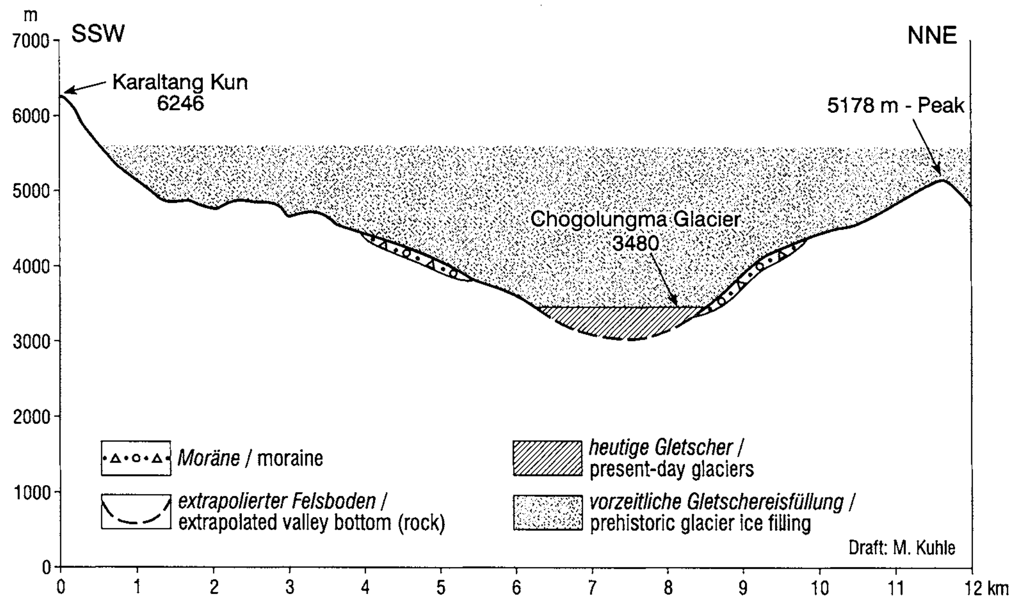


Figure 39. Cross-profile (not exaggerated) across the upper Basna valley with the middle Chogolungma glacier from the Karaltang Kun in the SSW on the Haramosh II N-ridge towards the NNE as far as the 5178 m-peak on the orographic left valley side with the minimum glacier ice filling of this main valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 58 and 59.

and 60). Modified by the metamorphic sedimentary bedrock there, the orographic right flank shows the same indicators of ice levels, i.e. remnants of ground moraines and abrasion forms. They provide evidence of a maximum level at the same altitude as on the left valley side (Photo 118 ▬ ▬ below No. 60–58; Photo 121 ▬ ▬). Old lateral moraine ledges, belonging to the four Late Glacial glacier positions and levels, are lacking, because these levels ran above the corresponding ELA so that no lateral moraine ledges could develop. However, 5 km down-valley of the locality of Khurumal, ground moraine accumulations have been preserved, which, due to their decametres-thickness, show first characteristics of lateral moraine remnants of the youngest Late Glacial Stage, the Sirkung Stage (IV) (Photo 119). Obviously the glacier surface on the glacier cross-profile concerned was already near the ELA, which had a depression of c. 700 m as against today (cf. Tab 1). For this reason, the re-feeding and filling-up of ground moraine mass from the glacier margins is already more important than is the case far above the ELA. The cupola-shaped accumulations are situated c. 200–300 m above the current glacier and contain large, rounded, erratic boulders (○; Figure 2/1 No. 59). The increase of sand portions (>60%, Figure 7) compared with the ground moraine 6 km up the main valley (<60%, Figure 8) proves the progressive intake of fresh debris by the glacier. Owing to this, the trituration, which down-valley increases with the distance of transport, has been over-compensated by fresh debris falling into the marginal cleft, and it opens only in the proximity of the snow-line. Correspondingly important moraine remnants on the orographic left, formed like ground moraine with the first indications of lateral moraines, can be found within 3–4 km further down-valley (Photo 120 ■, 121 second and third ■ black from the right). After the thawing-down of the abutment built-up by the glacier, the important thickness of these moraine complexes has led to large-scale, slow moraine slides. The down-thrown fault is so fresh, that a still active slide has to be suggested (Photo 120 ▼). Apart from a great number of further moraine indicators, the portions of quartz grains, out of which 100% are glacially crushed (Figure 6 No. 1) without indications of frost weathering in situ, undoubtedly preclude a glaciofluvial terrace-genesis. The convex course of the frequency grain size curve in the sand above and towards the right (Figure 7), which might be glaciofluvial, can be observed, too, in connection with coarse primary moraine maxima. In this special case it has been explained by Schreiner (1992: 144) as due to the substantial portions of sand in the source rock.

A large orographic right, i.e. southern, side valley is the Niamur Nala with the Niamur Gans, a dendritic valley glacier, which no longer reaches the Chogolungma parent glacier. This valley is a classic glacigenic trough (Photo 121 ⋃; Figure 2/1 left of No. 60), the basal 600 m of which are mantled by thick ground moraine (■ on the very left). Its glacigenic flank abrasions (▲ on the very left) have attained a good 5000 m asl. Upwards, they are more and more covered by the present-day glaciation. Further down the Chogolungma valley, the Sencho Nala joins from the right (S) (Photo 122). This is a steep hanging valley leading down to the NNW from the Berginsho Church (No. 60), which in its upper half is still glaciated (□ top). It undercuts the High- to Late Glacial glacigenic polish bands of the main valley with their metres- to decametres-thick ground moraine overlays (■). The LGM-glacier level ran about 5500 m asl and thus at such an important height that the right margin of the Chogolungma glacier has reached the summit superstructure of the Berginsho Church (⇩), occupying this side valley and crossing it. For this reason, the side valley has been filled with ground moraine to a depth of at least decametres, stripped from the ground of the parent glacier. In some places, the Sencho Gans tributary glacier has still

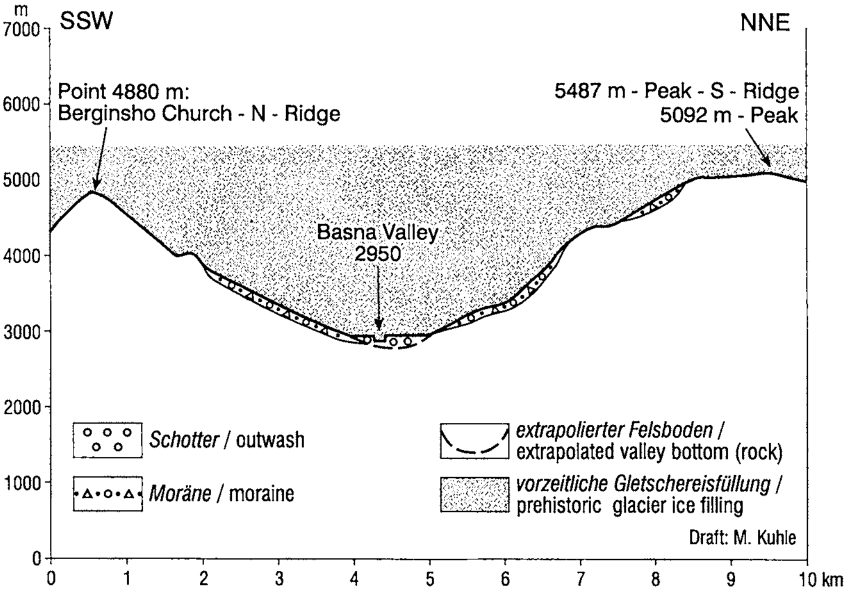


Figure 40. Cross-profile (not exaggerated) across the middle Basna valley down-valley of the current Chogolungma glacier end from the Berginsho Church N-ridge in the SSW towards the NNE as far as the 5487 m-peak S-ridge on the orographic left valley side with the minimum glacier ice filling of this main valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 60 and 1 cm to the right of No. 59.

not been able to remove their remnants since deglaciation (▽ top; Figure 2/1 on the left above No. 60).

Altogether the orographic right flank of the Chogolungma glacier valley discussed here, shows on the one hand a strong division caused by the inflowing side valleys (Photo 123), and on the other glacigenically triangular-shaped slopes (☗; Fig. 2/1 between No. 55 and 60) with ground moraine remnants (■; Fig. 39: km 4–5) which can be observed on the intermediate back-polished mountain spurs. In the erosion shadow of side valleys which provide niches, glacial ground moraine deposits have occasionally also been seen (Photo 122 between ■ and ■ top). The orographic left flank of the valley below the Bukpun-W-(5441 m) and E-(4551 m) summit (Fig. 2/1 half-right below No. 59) has been preserved as a 1.5 to 2 km-high continuous glacigenic polish flank, divided by comparatively small wall gorges (Photo 125, 126 ⇩). These reach as far as the region of perennial snow fields and, accordingly, are active. They carry meltwater and especially in summer they cause big mudflows which undercut the bedrock, but the basal ground moraine covers as well (■), and lay down the dislocated loose material in the form of mudflow fans (Photo 125 △; Fig. 2/1 below No. 59 to the right). In July 2000 quite large mudflows have been observed almost daily. The abrupt increase of the extension of the main gorge above half of the flank height (⇩) derives from deglaciation of the highest flank areas which took place several millennia earlier (see above). In the down-valley continuation of this orographic left flank section of the still glaciated Chogolungma valley, the Kero Lungma joins from the N (Fig. 2/1 from No. 62 in the direction of No. 60; Photo 124 Ʊ; 129 left half). The alluvial- and mudflow fan (◁ white, Fig. 2/1 below No. 59 to the right) at its valley exit has been piled up against the tongue end of the Chogolungma glacier (□). In the Kero Lungma a ground moraine pedestal (Photo 28 ■ below Ʊ Fig. 2/1 to the right of No. 59) has existed into which the talweg has been incised subaerially only after deglaciation. From the material removed during this process, the fan at the valley exit has been built-up. Prior to this and up to the historical Stage X, this material had been continuously eroded by the Chogolungma glacier in the main valley. The lateral moraine of Stage X shows the 100–150 m higher level of the then Chogolungma glacier (Photo 124 and 126 X). Only after the deglaciation following Stage X (Tab 1), i.e. during the last 200 years (Tab. 1), was there enough space for the development of a fan. Only during this time could the Kero Lungma talweg be deeply inset in the way as today, because its level was previously adjusted to the Chogolungma glacier surface. Despite the fact that the Kero Lungma was filled with ice up to beyond its upper edge (Photo 128 ▬ ▬), a classically wide U-shaped valley has not been created in the bedrock (Fig. 41). Only the filling-up by the ground moraine pedestal has significantly strengthened the glacigenic trough profile (Fig. 2/1 to the right of No. 59). As evidenced by flank abrasions, the LGM-ice thickness attained at least a good 2400 m in this valley (Fig. 41). To the orographic right of the Chogolungma glacier end, historical and neoglacial lateral moraines and at least one Late Glacial lateral moraine, that of Stage IV, are verifiable up the valley flank (Photo 124 ■ below; Photo 125–127 ■ IV; Fig. 2/1 above No. 60). This lateral moraine of the Sirkung Stage (IV), on which the pasture of Guma is situated, contains polymictic erratic boulders of crystalline schists and far-travelled granite (○) and is located here 730 m above the valley floor. Above this lateral moraine, older ground moraine deposits have been preserved (Photo 124 to the right of ■ below, large; Photo 126 ■ white above ■ IV). Among others, the morainic character of this polymictic material, rich in matrix, is documented by the bimodal course of the grain size cumulative curve (Fig. 16; Fig. 2/1 above

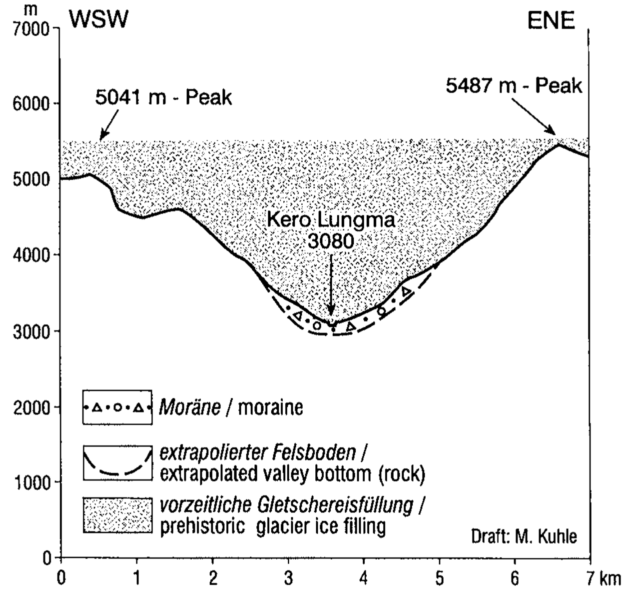


Figure 41. Cross-profile (not exaggerated) across the lower Kero Lungma from the 5041 m-peak in the WSW, towards the ENE as far as the 5487 m-peak on the orographic left valley side with the minimum glacier ice filling of this left side valley of the Basna valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 59 and 1 cm to the right of No. 59.

No. 60) and the 100% crushed quartz grain portions (Fig. 6 No. 10). These ground moraine remnants continue from up-valley towards down-valley along the right valley flank and reach up to an altitude of c. 4000 m asl (Photo 126 the four ■ to the right of No. 63). Additionally, a further sediment sample taken there (second ■ from the left; Fig. 2/1 above No. 60) confirms the ground moraine character of the matrix (Fig. 17 and 6 No. 11). Above, remnants of flank polishings (☗ on the right) are preserved (⇩ on the right) which in many places have crumbled away. The glacigenic sharpening which reaches as far as the rock gendarmes, provides evidence of a maximum glacial ice level of at least 5500 m (Fig. 40) in this valley cross profile.

3.12. Reconstruction of the maximum Ice Age glacier ice filling of the Basna valley down-valley of the present-day Chogolungma glacier terminal and in the connected side valleys

In the valley chamber of the settlement of Arandu the main valley is shaped like a trough (Photo 129; Fig. 2/1 half-right above No. 60). The Basna river, i.e. the meltwater river (○) of the current Chogolungma glacier (□ white) has accumulated a glacier mouth gravel floor to form the valley bottom. To this the tongue end of the Tippur glacier (Tippur Gans) coming from the orographic right side valley of the same name is adjusted (□ black on the right). The settlement of Arandu is situated on the gravel floor of this small glacier (◁; Fig. 2/1 half-right above No. 60), several metres above the gravel floor of the Chogolungma glacier. The Tippur valley, which leads down from the N-flank of the Berginsho Church mountain (No. 60), is a complete valley trough, the flanks of which have been abraded as far as its upper edges in prehistoric times (Photo 130 ☗ white) so that features related to roches moutonnées have been created up there. Currently the tongue of the Tippur glacier advances (▽) and thereby overthrusts the field terraces of Arandu (□). Further down the main valley, the fresh fan of the glacier mouth gravel floor (Photo 131 ○) of the Tippur glacier (△) has undercut neoglacial ground moraine remnants (■ and ■ V black). On both the main valley flanks Late- to High Glacial ground moraine covers are preserved (■ white; Photo 129 ■ black). Situated on a valley shoulder on the orographic right (Photo 130 ■ white on the left; 131 ■ white, centre), they have been erosively removed in the area of tributary talwegs undercutting this valley flank and resedimentated as alluvial-i.e. mudflow-fans (△) at the slope foot. On the orographic left side ground moraine remnants of the Stages III to 0 (=LGM) are preserved up to a height of 4000 m (Photo 133 ■ large), locally they even attain 4500 m (Photo 124 ■ on the very right), that is over 1600 m above the valley bottom (Fig. 2/1 between No. 60 and 65). The moraine remnant mentioned last shows several flat rills of rain- or snow meltwater created since the deglaciation.

In the junction area of the Kushusum Lungpa (Berelter Nala) and the Basna valley the merging ground moraine covers on the slopes of the side valley and the main valley testify to synchronism by their geomorphological interlocking (Photo 134). They suggest that during their last, i.e. late Late Glacial (Stage IV) deposition and reshaping respectively, the bottoms of the main- as well as the side valley were covered by a dendritic valley glacier system and the Kushusum glacier tributary stream flowed into the Chogolungma parent glacier. This left side valley ends slightly V-shaped and further above, as well as its side valleys, shows trough forms (Fig. 2/1 between No. 60 and 67) and bears corresponding traces of an enormous Ice Age (LGM) glacier filling (Photo 133). These are – from below to above – ground moraine remnants (■ small) and evidently glacigenic flank abrasions (the three lowest ☗; Fig. 2/1 between No. 60 and 68) which mediate to an orographic left Late Glacial polish line at 5300 m asl (– – white). The High Glacial flank abrasions reached much higher up (☗ black, top), so that the High Glacial glacier level (– – black) in the uppermost Sokha Nala attained c. 5900 m. During the LGM it crossed the Sokha La, a transfluence pass (Fig. 2/1 between No. 68 and 64), and passed without a threshold into the level of the Biafo ice stream network. The S-wall of the 6066 m-summit (half-right below No. 68; Fig. 2/1 below No. 68) is built up from gneiss, showing classic polish forms up to the very top. Up to every detail they are similar to the polish forms of the Trolltindan wall on the orographic left in the Romsdalen (Norway), which also reach as far as the summit and thus provide evidence of an ice thickness of over 1800 m. In the confluence area of the Basna valley concerned here, an LGM-ice thickness of at least 2700 m is verifiable (Fig. 40 and 41). Obviously, the work of the glacigenic flank abrasion carried out by this High- to Late Glacial ice stream network, has sharpened all summits which are over 5400 m-high into glacial horns (Photo 132 and 133; Fig. 2/1 No. 64, 65, 68).

The two tributary valleys – the Aralter- and the Dungus valley (Photo 127) – also connected to this confluence area

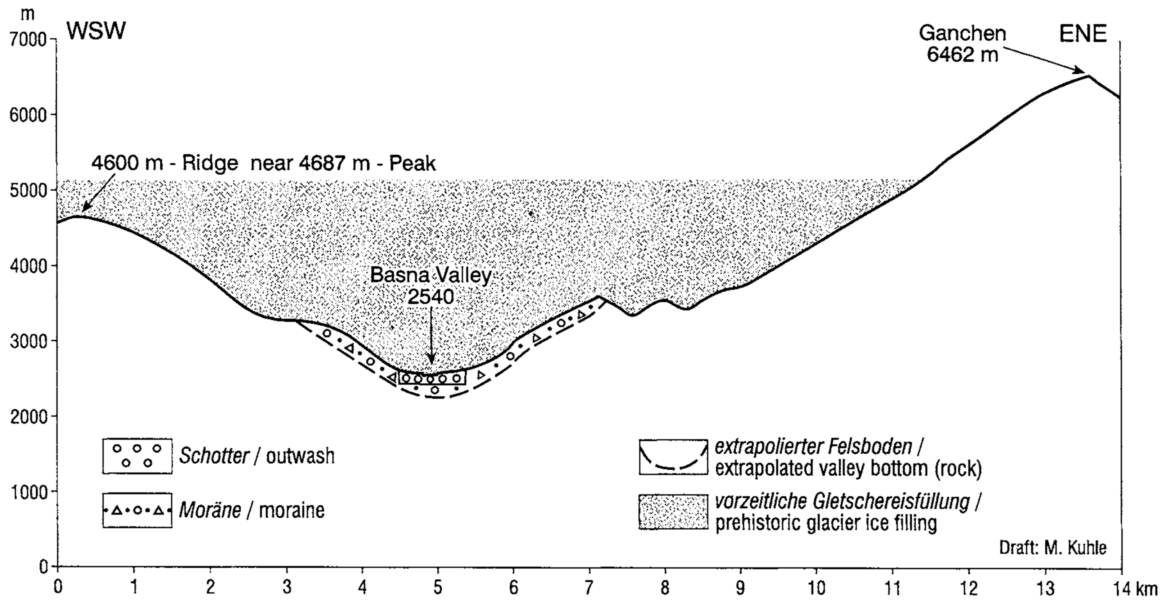


Figure 42. Cross-profile (not exaggerated) across the middle Basna valley from the WSW, from the 4600 m-crest on the orographic right, towards the ENE as far as the Ganchen on the orographic left valley side, with the minimum glacier ice filling of this main valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 60 and 66.

of the Basna valley, are stepped hanging valleys, the originally glacigenically rounded junction thresholds of which have in the meantime been transformed by the meltwater creeks of the current valley glaciers into a V-shaped valley stretch (Fig. 2/1 between No. 63/64 and 60). Evidence of this are the alluvial fans (Fig. 2/1 to the left of No. 66) discharged from these V- shaped valley stretches (Fig. 2/1 left of No. 66) since deglaciation, as e.g. the alluvial fan of the settlement of Bisil (△ white). Above, the valleys are wider and have – at least in sections – trough valley cross-profiles (Ʊ; Fig. 2/1 left of No. 63 and 64). Ground moraine remnants on the valley flanks (Photo 132 ■ white) as well as glacigenic flank abrasions on the bedrock (◥ ◤) indicate a prehistoric level of the ice stream network about 5400 to 5500 m (Photo 127 and 132 ‒ ‒). The pinnacle features of the 5563 m-summit (No. 64) and the adjacent course of the crest (Fig. 2/1 No. 64) prove a glacigenic undercutting by this glacier level (‒ ‒), but it cannot be ruled out that this level (‒ ‒) represents just the oldest Late Glacial (Stage I). However,this level – which perhaps was not a maximum – also ran 2600 m above the subrecent to present-day gravel floor of the valley chamber of Gon (Photo 127 □).

The Basna valley bottom with its gravel floor is the current glacier mouth gravel floor of the Chogolungma- and Tippur glacier (Photo 126, 129, 131, 134 ○). It is divided into a sequence of small glacier gravel floor terraces with a height of several metres at maximum (Fig. 2/1 left of No. 66). With their help one can recognize the retreat of the 50 km-long Chogolungma glacier during the last decades.

3.13. Reconstruction of the maximum Ice Age valley filling with glacier ice in the Basna valley from the locality of Bulcho down-valley as far as its confluence with the Braldu valley

On the orographic left valley flank, which at the same time forms the W-flanks of the mountains Hikmul (No. 63, c. 6300 m), Ganchen (No. 66, 6462 m) and the 5046 to 5778 m-high southern fore-summit (No. 50) of the Ganchen massif (Photo 136), ground moraines up to maximum heights of 3900 to 4100 m are preserved (Photo ■ 135 white; 136 the two ■ white on the left; Fig. 2/1 between No. 60 and 50; Fig. 42). They lie c. 1300 to 1500 m above the talweg and have still been reached by the main glacier during the Dhampu Stage (III, Tab 1). The ground moraine cover of the Sirkung Stage (IV) is continuously preserved over a slope-parallel distance of approximately 2 km (Photo 136 ■ IV on the right) up to a height of 3600 m. Even the morphology of primary exaration traces and pressure edges of the attached glacier margin (❙) can be observed. In some places this ground moraine is decametres- to over one hundred metres thick (Photo 135 ■ black). As usual, its thickness increases towards the valley bottom, so that there existed a ground moraine pedestal built-up in the course of the Late Glacial, which during the deglaciation has been removed along the talweg of the Basna river (Photo 136 □). The accumulations on the valley ground modified to mudflow fans (▲ white; Photo 135 △) which make up the current fields, contain remains of this ground moraine pedestal. The High Glacial glacigenic flank abrasion (Fig. 2/1 between No. 60 and 50) reaches up from 4800 to 5100 m (Photo 135 ☗; 136 the two left ☗), thus indicating the minimum height of the LGM-glacier level (0‒ ‒; Fig. 42). Sections of rock crests, situated below this ice level during the LGM, have then been undercut and afterwards roughened by the ice level which during the Late Glacial had dropped below their level (Photo 136 ‒ ‒ between No. 66 and 50). The south-

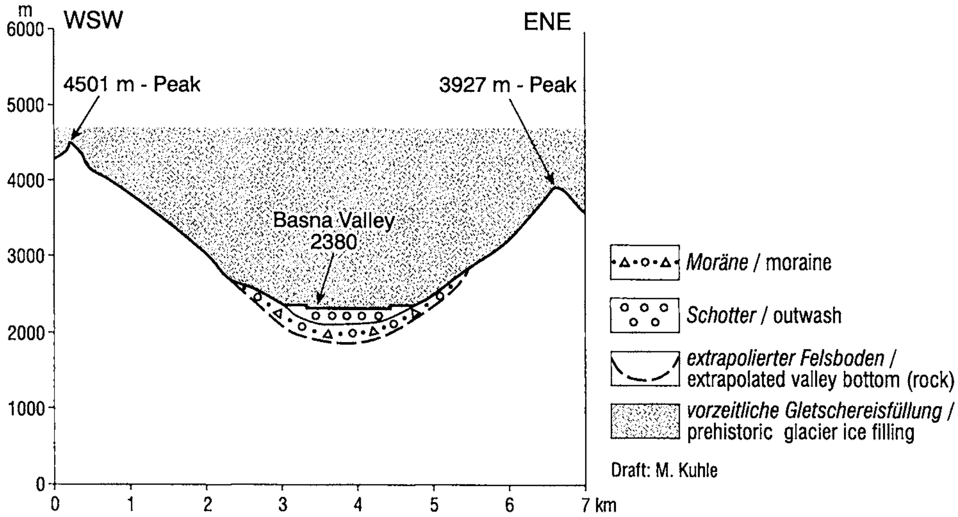


Figure 43. Cross-profile (not exaggerated) across the lower Basna valley, 4.3 km above its inflow into the Shigar valley, from the WSW from the 4501 m-peak on the orographic right towards the ENE as far as the 3927 m-peak on the orographic left valley side with the minimum glacier ice filling of this valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 approximately between No. 50 and No. 51.

ern Ganchen fore-summit has been undercut by the High- to Late Glacial glacier level and sharpened into a glacial horn (Fig. 2/1 No. 50).

The ground moraine cover of the orographic right flank of this valley section has corresponding extensions, because here, too, late Late Glacial ground moraines are preserved up to a height of 3600 to 3700 m (■ on the very left, right below and below No. 51 as well as on the very right; ■ IV on the left; Fig. 2/1 on the right of No. 60; Fig. 42, however, this cross-profile does not include the highest orographic right ground moraines, but only those up to 3250 m asl). Their substantial thickness reaching decametres, can be recognized here as well as on the opposite valley side by typical downthrows of large rotational slides (↓), by which the ground moraine packing gives way when the valley glacier has melted down and the abutment of the ice is absent. There are outcropping metamorphic rocks in the underground, but in the ground moraine cover 'swim' erratic granite boulders (○) up to several metres in length, which in the longitudinal direction of the valley have been transported down over a long distance. The fine material matrix has a very weak maximum in the medium sand. However, it is nearly reached by the coarse silt and fine sand, whereby with a good 13% clay a high sorting coefficient (So = 5.63) is connected (Fig. 18). The SiO2-grains, contained in an only small quantity, are without exception crushed. The rounded grains of soft rocks reveal an involvement of water, as is to be expected c. 1000 m below the corresponding late Late Glacial snow line (ELA of Stage IV = 4300 m) (Fig. 6 No. 12). The highest orographic right mountain ridges, fringing the Basna valley trough above the Tisa Birri settlement, reach 4687 m. They have been rounded by glacier abrasion (Photo 136 ▲ white), so that the level of the LGM-ice stream network must have been situated above (=0– – on the right; Fig. 42). As for the mantling of the slopes with ground moraine and the glacigenic flank polishings reaching higher up, the glaciogemorphological characteristics of this valley section continue down-valley without interruption as far as the valley cross-profile 4.3 km above the junction with the Shigar valley (Fig. 43) (Fig. 2/1 from the cross-profile between No. 60 and 50 downwards up to above No. 51). In this lower course, the Basna valley is also a classic whole-valley trough, filled with glacier ice up to the upper margins of the flanks (Photo 136 – – below No. 51 half left and half right).

4 km down from the Tisa Birri settlement, the orographic right main valley flank of the Mutuntoro Klas (valley) is interrupted (Photo 137). Its orographic left flank, which joins the orographic right flank of the Basna valley, is extensively covered with ground moraine up to a height about 4000 m asl (■ white; Fig. 2/1 centre, between No. 60 and 51). These ground moraine deposits and also the abrasions extending even higher up, provide evidence of a level of an Ice Age tributary valley (– –) corresponding to that of the main valley. They thus confirm the overall picture of an LGM-ice stream network glaciation with communicating main- and tributary glacier levels. Since deglaciation the ground moraine pedestal of this tributary stream (■ black) has been fluvially undercut and partly dislocated into the mudflow fan at the valley exit (○; Fig. 2/1 centre, between No. 60 and 51). C. 6 km down-valley of the inflow of the Mutuntoro Klas (valley), the geomorphological trough valley profile of the Basna valley E of the settlement of Chutran was mapped between the 4501- and 3927 m-peak (Fig. 43). On the orographic right side the ground moraine matrix was analysed (Fig. 19). The rock slope on which the ground moraine lies, is a 45°-steep outer bank, undercut by the Basna river (in Fig. 43 WSW of the valley floor), so that since deglaciation the ground moraine has been displaced down-slopes and the characteristics of its grain sizes have undergone a loss in significance. However, the morphoscopic characteristics of the quartz grains have been completely preserved (Fig. 6 No. 13).

In the entire course of the 23 km-long Basna valley, which is the subject of this chapter, numerous glacigenically triangle-shaped slopes have been formed (Fig. 2/1 between No. 66, 60 and 51; Photo 135 ▲ on the right below No. 66;

136 ⯊ on the right below No. 50; 139 ⯊ on the very left). These polyglacial abrasion features, polished for the last time during the LGM to Late Glacial, alternate in the longitudinal profile of the valley with slope sections showing decametres- to over 100 m-thick ground moraine deposits. An example of this is the orographic left valley slope above the settlement of Thurgu in the WSW-flank of the 3927 m-peak (Photo 138 ⯊; Fig. 43 ENE of the valley bottom). In the area where the outline of the valley flank runs convexly, at the same time forming an inner bank in the current river bed, the spur-like protruding bedrock has been polished back by the glacier (⯊). Where the course of the valley flanks is indented concavely, there was a flow shadow behind the up-valley spur (⯊ on the left) into which ground moraine (■ black) has been pushed and then – in front of its down-valley rock projection (⯊ on the right) – stripped off by the left glacier edge up to a significant thickness. After deglaciation these moraine deposits (■ black), now lacking an abutment, are especially susceptible to the formation of rills with earth pyramids (Fig. 2/1 on the right above No. 51) and also slides. The accumulation on which the settlement of Thurgu is situated (△ white on the right) contains primary remnants of ground- and lateral moraine of the late Late Glacial (Sirkum Stage IV), however, it is covered by a mudflow fan of dislocated ground moraine. This has been transported here from the enclosed depression formed in the meantime (to the right of ■ black). In the adjacent region, where the ground moraine was primarily thin, it has even been preserved up to the upper slope, i.e. up to approximately the mountain ridge (■ white). As evidenced by the glacigenic abrasion traces on the highest elevations of both the valley flanks (Photo 138 ▬ ▬; 99▬ ▬ on the right and centre; 136 ▬ ▬ on the left below No. 51; 139 half left below No. 50), in this valley cross-profile (Fig. 43) the level of the LGM-minimum thickness of the ice must have been above their culminations, running at 4700 m asl. A good 4 km down from this cross-profile, the lower Basna valley joins the Braldu valley (Chapter 3.10); they both are glacigenic troughs with a gravel bottom (Fig. 2/1 on the right above No. 51). Here, the orographic left valley flank, we are talking about, peters out and dips as a mountain spur (Photo 139 ⯊ black, centre) under the Shigar valley bottom. Its perfectly glacigenic rounding (⯊ black, centre; Fig. 2/1 between No. 50 and 51) and ground moraine cover (■ on the left) once more provides evidence (see above; Photo 136 on the right of No. 50; Photo 138; Fig. 43) of the joint glacier surface of the Basna- and Braldu ice streams which already existed at a distance of over 7 km upwards of their confluence during the LGM (Stage 0).

3.14. Summary of Chapter 3.11–3.13: The LGM-glaciation in the Chogolungma-Basna valley from the upper valley head in the Malubiting- (7458 m) and Spantik (7027 m) massif as far as its lower valley exit and the confluence with the Braldu valley, i.e. its inflow into the Shigar valley (Fig. 2/1 from the profile between No. 52 and 53 up to the profile between No. 50 and 51)

As indicated by the seven exemplary, glaciogeomorphological valley cross-profiles of the Chogolungma-Basna valley and several of its side valleys (Fig. 37–43), investigated in detail, the valley system of this main valley was filled with ice in a glacier thickness of 1800 m (Fig. 37) up to c. 2900 m (Fig. 42) during the LGM. Thus glacier thicknesses comparable to those of the Braldu valley system have been reached (cf. Fig. 3; Chapter 2.–3.10). Just as in this valley system, the most important ice thickness was several kilometres down-valley of the current glacier terminals. In the entire area concerned, the LGM-glacier surface lay above the ELA (snow line). This is documented by the fact, that no LGM-lateral moraines have been found, but only ground moraines which upwards of the valley flanks have without exception turned into glacigenic abrasion- and polish features, i.e. pure erosion forms. Towards the source area of the main valley and as far as into its tributary valleys, the LGM-ice thickness decreased to its smallest amounts (Fig. 37). At the lower valley exit, in the confluence area with the Braldu (Blaldu-) valley, i.e. at the place where the two valleys merge to form the Shigar valley, the glacier thickness including the ground moraine which probably moved as far down as the rock ground, has still come to approximately 2500 m (Fig. 5). Because rock bottom heights of the trough ground are concerned, extrapolated from the dipping trough flanks (see above), the real LGM-ice thickness might differ from the data of the cross- profiles (Fig. 3–5 and 37–42) by a few hundred metres at maximum. In this case the maximum ice thickness of 2900 m (see above), for instance, would have to be corrected to at most 2600 m (Fig. 42) which in principle is no differing result. Neither the dimensions of the infilling of the relief with glacier ice nor the resulting prehistoric glaciation type are affected by this inexactness. Its was definitely a maximum glaciation which exeeded the thickness of the Ice Age Alpine glaciers, reaching at most 1800 to 2000 m, by 20 to 30%. The type of glaciation was that of an ice stream network communicating with the glacier systems of the Hunza valley in the W and the Hispar-Biafo-Baltoro system in the N up to the NE across the highest passes. 'Communicating' means to be connected with an approximately unified, uninterrupted ice surface levelled out by the transfluences. Corresponding to that of the simultaneous upper Braldu glacier (Photo 5 and 65–67), the highest LGM-ice levels reached an altitude of 5800-6400 m (Photo 113).

3.15. The maximum Ice Age glaciation of the Shigar valley

At the head of the Shigar valley, in the confluence of the Braldu- and Basna (or Basha) valley, Oestreich (1906: Fig. 28) has already approached the accumulations of edged coarse boulders (Photo 99 □ and 139 ○) as 'rock-slide debris'. Dainelli (1922, plate LIV) classifies them as terminal moraine of his '4th Advance' and Norin (1925) has described them as 'low morainic ridges . . . protruding through the gravel formation in the middle of the valley . . . ', deposited by 'a composite glacier' of the Braldu- and Basna valley glaciation (cited after Hewitt 1999: 231). Hewitt (ibid.) provides a detailed analysis of at least three 'rock avalanches' the 'conspicuous breakout scars (of which lie) 600 m-wide and 700 m-high' in the orographic right flank of the Shigar valley. He refers to them as Ghoro Choh I-III rock avalanches (ibid. Fig. 10). According to his findings of mega-clasts in the eastern valley flank above the Haidarbad settlement 'Bollah', 7 km away from its breakout scar, the rock avalanche has surged up the orographic left slope by even 150 m. He suggests that the rock avalanches discharged into a Shigar valley, which was free of ice during the Holocene.

We are interested in the traces of the preceding maximum glacier filling of this valley section, the flank abrasions of which might have led to the slope steepenings which have created the conditions for important postglacial rock avalanches like these. Authors like Norin (1925) and Dainelli (1922) – the latter during his '4th Advance' which most likely corresponds with the Late Glacial (Stage I–IV after Kuhle 1982a) - have assumed that in former times the glaciers from the Basna- and Braldu valley have reached up to here. As we have seen in the chapters above, the ice thickness probably has markedly exceeded the ice fillings reconstructed by these authors, even during the Late Glacial (Stage I–IV after Tab 1). In the course of this chapter this will become still clearer. First indications as to the height of the break-off are provided by the observations and data of Hewitt (1999) – for us this is at the same time a glacigenically prepared height of the crumbling away. It lay 700 m above the valley bottom (see above), that is at approximately 3100 m asl. The ground moraines found on this valley flank reach c. 700 m higher up (Fig. 5, WSW valley side). The glacigenic flank abrasions on the orographic right even reach as far as 4600–4700 m asl, so that an LGM-glacier level (Photo 99 ‒ ‒ on the very left) is indicated a good 2200 m above the valley bottom.

In his main opus Lydekker (1883) describes indicators of a prehistoric glaciation of the Braldu valley as far as its inflow into the Shigar valley. In the 'Geology of part of Dardistan, Baltistan and neighbouring districts' (1881: 47f) he reconstructs a 500 m-thick glacier with the help of glacial traces up to the confluence with the Basna valley. With the exception of the U-shaped valley form and erratic boulders near the Strongdokmo La, he has found no indications of a glaciation further down-valley, but he assumes that a Shigar glacier has reached the Basin of Skardu. Oestreich (1906: 80) confirms this suggestion. These observations of the two researchers concern a former glacier thickness of c. a quarter of that of the author, who has found traces of a maximum ice thickness of a good 2200 m (see above) in the Shigar valley. The ice thickness of 500 m reconstructed, i.e. assumed by these researchers, most likely corresponds to the late Late Glacial ice thickness during Stage IV (=Sirkung Stage, Tab 1), which 17 km up the Basna valley still came to c. 1000 m (see Photo 136 ■ IV).

The only accumulative slope ledge on the orographic left which might be a lateral moraine, is the rock fall next to the Mungo settlement – the Ghoro Choh I–III rock avalanches (Fig. 2/1 on the right of No. 51; Fig. 20; Fig. 6 No. 14). But despite the great similarity of the granulometric composition to that of a moraine, the complete absence of quartz grains is an indication against glacigenic transport over a far distance and the corresponding mixing through. Apart from that, the glacier ice has left behind a completely evacuated valley. This argues in favour of the suggestion that even the last late Late Glacial glacier filling, that of the Sirkung Stage IV, has flowed through the entire Shigar valley as far as the Indus valley. The orographic right accumulations are decametres-thick remnants of ground moraine (Photo 99 ■ white; Fig. 2/1 on the right above No. 51) undercut by the gravel bed of the Holocene Shigar river (glacial trough with gravel-bottom between No. 51 and 47; Fig. 5). These ground moraine remnants have partly been interrupted and modified by mudflows fans from the side valleys (Photo 141 ▽ small; Fig. 2/1 from right below No. 51 to right below No. 72). Mudflow fans of this type are also on the orographic left (Photo 140 and 141 ▽ in the foreground; Fig. 2/1 between No. 72 and 46); for the most part they consist of displaced ground moraine material from the valley flanks (Photo 140 ■ on the right; 141 ■ white; Fig. 2/1 on the left of and below No. 47) and side valleys (Photo 140 ■ on the left and in the middle). Ground moraines of the 'pedestal moraine' type can be diagnosed in the very thick accumulations formed like terraces on the orographic right side (Photo 141 ■ black between No. 74 and 51). They have been heaped up by the tributary glaciers against the Shigar parent glacier (Fig. 2/1 to the right of No. 72–74). For the reconstruction of minimum ice thicknesses which clearly exceed the 500 m observed in the Shigar valley (see above) the ground moraines, which as remnants are preserved on the slopes, are sufficient. Three exemplarily high ground moraine remnants have remained on three glacigenically triangular-shaped slopes on the orographic right (Fig. 2/1 above No. 72) at 3600–3800 m asl, 1200–1500 m above the gravel floor (Photo 141 ■ white on the right and left below No. 51, on the right below No. 72). They are so thick that they have been dissected into several metres-deep flush rills and on the steep parts of the slopes are fissured into earth-pyramid-like pipes (between ■ white and black on the right below No. 72). C. 18 km further down-valley an orographic right moraine remnant, too, reaches quite far, namely as far as c. 3600 (■ below No. 74), up a glacigenically triangular-shaped slope (Fig. 2/1 No. 74). The glacigenic flank abrasions are highest preserved where the density of the side valleys is only small (⇩ and ↓) and, accordingly, the undercutting by their slopes is relatively ineffective. The highest one reaches 4650 m (▲ below No. 51); the almost freshly preserved flank polishing on the orographic right (▲ on the left below No. 74) trends from the valley bottom up to a good 3400 m. On the orographic left side an adequate glacigenic picture of the valley flank is presented (Photo 140; Fig. 2/1 on the left and below No. 47). The highest covers of ground moraine reach c. 3550 m (■ on the right) up to 3800 m asl (Photo 143 ■ background on the right). Here too, the glacigenic flank abrasions have left behind a series of triangle-shaped slopes situated between the inflows of the side valleys (▲ on the

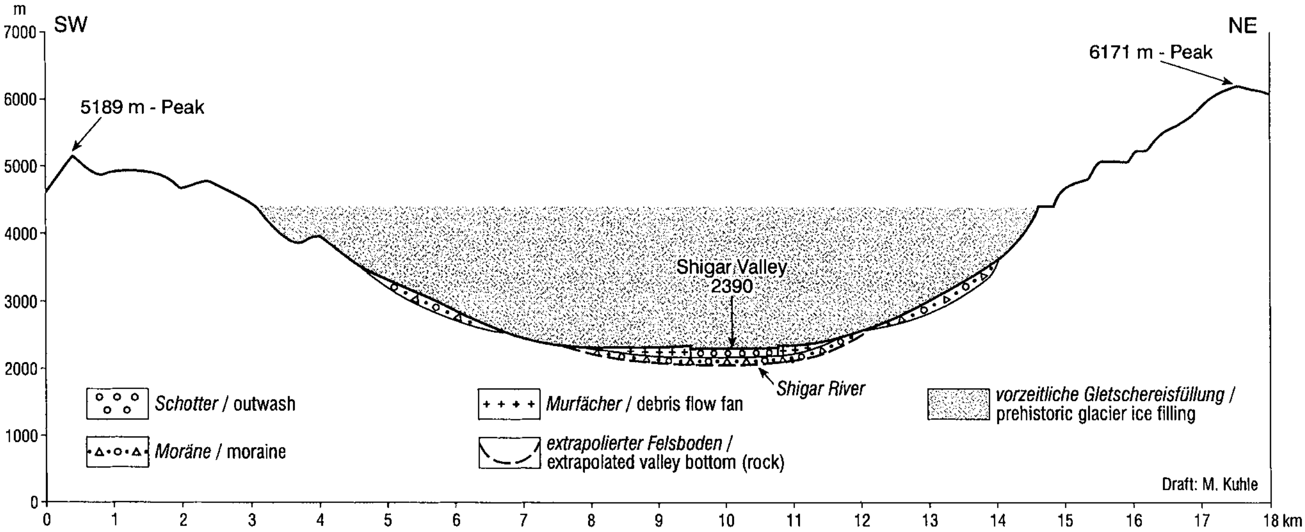


Figure 44. 17 km-wide cross-profile (not exaggerated) through the middle Shigar valley, half-way between the confluence of the Basna- and Braldu valley and 30 km above its inflow into the Indus valley (Skardu Basin) from SW, from the 5189 m-peak situated on the orographic right in the Haramosh Range, a SSE- satellite of the 5770 (or 5691)m-peaks (Fig. 2/1 No. 72), to the NE as far as the 6171 m-peak, the southern summit of the 6251 (or 6400) m-high Koser Gunge in the Mango Range (Fig. 2/1 No. 47) on the orographic left valley side with the glacier ice filling of this classic trough valley reconstructed for the LGM. Locality: Fig. 2/2; Fig. 2/1 between and below No. 72 and 47.

right; Fig. 2/1 No. 47), which partly have been smoothed by band polishing of the outcropping edges of the stratum (Photo 140 ▲). The polish line runs down from c. 4650 m as far as 3500 m (— —; Photo 143 — — on the right of No. 47 up to 0— —). Evidence of this LGM-glacier-level is provided by the corresponding orographic right polish lines at the same altitude (Photo 141 — — white on the left of No. 51 as far as — — black on the left of No. 74). Owing to this reconstruction of the thickness of the Shigar parent glacier, the ground moraine findings in the hanging side valleys at a height of 4400 to 4600 m become understandable (Photo 140 ■ below ⇩; 141 on the right of ◡), because the side glaciers have communicated with the parent glacier. Accordingly, the central cross-profile of the Shigar valley (Fig. 44) shows a trough with a postglacial gravel bottom (glaciofluvial gravel floor) to which syngenetic mudflow fans are adjusted. This trough has been flowed through by a main valley glacier at a thickness of c. 2100 m during the LGM and polished out during several Pleistocene glaciations.

Two orographic left side valleys of the valley chamber of the Shigar valley discussed here, exemplify traces of its glaciation: the Skoro Lungma and the Baumaharet Lungpa (Photo 42). Dainelli (1922, plate V and LIV) has identified moraines at the valley exit of the first, which he classifies as belonging to his 3rd Glacier-expansion. However, not only longish ice margin moraines are concerned as he has mapped them, but even ground moraines covering both slopes of the Skoro Lungma over large parts and reaching as far as c. 3400 m asl (Photo 143 ■ on the right of No. 47, Fig. 2/1 on the left of No. 46). Hewitt (1999, Fig. 2), too, does not contradict their glacial genesis, i.e. he has not diagnosed them as being rock avalanche debris. Dainelli (ibid.) has also found moraines in the NW-parallel Daltombore Lungma (Fig. 2/1 on the left below No. 45) which leads down from the Cherichor massif (No. 45, 6030 m). Without doubt, over 1000 m-thick tributary glaciers have still joined the Shigar parent glacier here, even during the Late Glacial (Stage I–IV, Tab 1). The Skoro Lungma tributary stream had a short-path connection over the Skoro La (Fig. 2/1 transfluence pass below No. 45) to the upper Braldu glacier in the valley chamber of the Askole settlement (cf. chap. 3.5.). It has polished out a relatively wide valley, which in the upper part of its cross-profile is thoroughly trough-like-concave. Only near the talweg has the subglacial Late Glacial meltwater erosion set gorge- and V-shaped cross-profiles into the trough ground (Fig. 2/1 on the left below No. 46). In comparison, the Baumaharet Lungpa (Photo 142) is a gorge-shaped trough valley with very steep flanks (Fig. 2/1 above No. 73 on the left) the glacigenic flank abrasions of which (above No. 73) have been roughened by crumblings, but can still be recognized high up (▲). In this valley only basal ground moraine remnants with earth pyramids have survived (■, Fig. 2/1 on the left of No. 73). Oestreich (1906: 79/80) has already identified them as being remnants of pedestal moraines (Fig. 2/1 on the left above No. 73), accumulated by the tributary glacier at its confluence with the main glacier. He has applied them to provide evidence of the simultaneous Shigar parent glacier. In this connection the pedestal moraines, i.e. the ground moraine pedestal at the exits of the orographic right Shigar side valleys must be referred to (Photo 41 ■ black between No. 74 and 51). During the last Late Glacial Stage (Sirkung Stage IV; Tab 1), when the ELA ran c. 700 m lower than at present (current ELA c. 5000 m asl), i.e. at 4300 m, the side glaciers had probably still reached the Shigar parent glacier. From the LGM up to this stage the ground moraines of the glacier tongue-ends of the side valleys, tending to melt down, have been progressively accumulated against the lowering edge of the main glacier to form pedestals of ground moraine. At the same time the main glacier became narrower, so that the moraine pedestals have been built-up and project into the main valley.

Up to the present their remnants (■ black between No. 74 and 51) have been preserved in the dead angle of the sub- and postglacial meltwater erosion (⇩). This backward linear erosion, which has first removed the ground- and pedestal moraines at the valley exits, has today already reached the high valleys, dissecting their trough grounds (Fig. 2/1 between No. 72 and 74) of bedrock (▢) as well. The LGM- to Late Glacial heights of the snow-line (ELA) between 3700 and 4300 m suggest the subglacial meltwater erosion of the lower 1300-1900 altitude metres as far down as the main valley bottom below 2400 m asl (see Tab 1). Above 4300 m, the bottoms of the side valleys have been hardly dissected so far.

A further glaciogeomorphological key locality is the orographic left transfluence pass from the Shigar- into the Indus valley, the Strongdokmo La (Fig. 2/1 on the left of No. 37), which has already been discussed. Here, a typical glacial polish depression- and polish threshold landscape has been preserved, the depressions of which are filled with loose material (Photo 144 and 145 □) pierced by roches moutonnées (Photo 144 ▲ white and black below No. 47; Photo 145 ▲ black on the right; Photo 146 ▲ white on the right). Part of the loose material can be evidenced as being ground moraine (Fig. 21 and 6 No. 15). 430 quartz grains of this ground moraine have been microscopically examined: 427 were glacigenically crushed, the remaining three were rounded, i.e. glaciofluvially reshaped. The ice overburden has pressed this obviously metres- to decametres-thick ground moraine (Photo 144 ■ black, middle) to the side of the roche moutonnée rocks (▲ white) and then it has been forced upwards along them. It has to be stressed that this is no weathered material of roches moutonnées, accumulated as a small debris cone from a rock rill, but that ground moraine is concerned here! The polymict boulders (○) – among them erratic granite boulders – (○ black), lying 110 m above the present-day gravel floor of the Shigar river, are fluvially speaking not understandable. The roche moutonnée forms (▲ black below No. 47 and white; Fig. 2/1 on the left of No. 73) point to a S-direction of the ice flow, out of the Shigar- and into the Indus valley (Skardu basin). Without doubt this flow direction concerns the glaciers of the Late Glacial (Stage I–IV), which were the last with a forming effect. Before (Stage 0–I), the ice filling of the Indus valley – being the main valley with an overriding importance – might have been too thick to permit a substantial inflow of tributary glaciers. In contrast to a High Glacial (LGM = Stage 0) minimum glacier level at 3400 m asl (– –; Photo 141– – left margin; 147– – on the left below No. 73) confirmed by flank abrasions (Photo 141 ▲ left margin, Photo 144 ▲ black on the right) and their upper limit, the Late Glacial Shigar glacier was no longer 1200 m thick (as the LGM glacier), but only 400 m. This is evidenced by the orographic left lateral moraine above the confluence saddle (■ I–IV; Fig. 2/1 on the left of No. 73). The relatively important thickness of the LGM-Shigar glacier in the area of its confluence into the Indus valley is also documented on the orographic right side, where – opposite to the confluence saddle (Strongdokmo La) –, flank abrasions (Photo 143 ▲ black on the left) reach up to a height of at least 3400 m (– – on the right of No. 79). Ground moraine is preserved on the prominent pillar, polished-out concavely (▲ white on the right; Photo 141 ▲ third from the left), as well as on the rock ledges next to it (Photo 143 the three ■ on the left of No. 47; Photo 144 ■ on the left; Fig. 2/1 on the right of No. 74). The local height of the LGM-ice-level at the exit of the Shigar valley is large-scale and confirmed by the simultaneous ice level in the Skardu basin, which is even verifiable by high-lying ground moraines (Photo 148 ■ on the left). This ground moraine- and polish line locality (Fig. 2/1 below No. 74) is situated immediately behind the polished rock spur at the exit of the Shigar valley (see above, Photo 143 ▲ on the right of No. 79), on the side of Peak No. 74 which faces Skardu (Photo 149 – – white and black on the right below No. 74). It is not possible that the Shigar glacier has polished it, but only the Indus glacier. The 160°-angle formed by the flanks of the Shigar- and Indus valley makes clear that a dead angle has existed here, so that the Shigar glacier was unable to polish.

Owing to these large-scale arrangements of the positions of the glaciogeomorphological indicators of an LGM-ice-level, the glacigenic shaping of the southern section of the Strongdokma La becomes understandable and trivial (Photo 145 ▲ and ■). The partly rounded rocks and round-polished hills rise up to a height of a good 2900 m, i.e. 700 m above the present-day level of the Shigar river. In parts they are abraded in such a perfectly streamlined fashion (▲ on the right) that this additionally supports a local thickness of the overflowing ice of c. 500 m (LGM-ice level c. 3400 m asl, see above). Here, too, very large, light erratic granite boulders have been met in the moraine formation as well as on the dark phyllite bedrock (↓ ↓). The fresh state of preservation of the glacigenic forms in a continental mountainous environment signified by an insolation- and frost weathering like this, confirms the assumption of their LGM- to late Late Glacial age (Stage 0–IV, Tab 1).

3.15.1. Summary of the observations on the maximum Ice Age glaciation of the Shigar valley (Fig. 2/1 from between No. 50 and 51 to between 74 and 73)

Norin (1925, Tavl. 5) who has obviously taken into consideration neither the approach of Lydekker (1883) nor Oestreich (1906) or Dainelli (1922), thought the Shigar valley to have been free of glaciers during the Ice Age, whereas Lydekker and Oestreich have indirectly come to the conclusion that a Shigar glacier has reached the Indus valley. The thickness of this Ice Age glacier, however, was left open. Our reconstruction proves that the upper Shigar glacier must have been c. 2400 m thick (Fig. 5), the middle c. 2150 m (Fig. 44) and the lower in the area of its confluence into an Indus parent glacier still 1200 m – plus an estimated thickness of the gravel body of c. 100 m on the valley bottom under the current Shigar river (Photo 143 □). The surface of the ice stream was c. 10.7 (Fig. 5) to 11.5 km-wide (Fig. 44). The Braldu- and Basna glacier were its source branches, and 5- to 24 km-long (Baumaharet Lungpa) tributary glaciers have joined from the SW and NE side valleys.

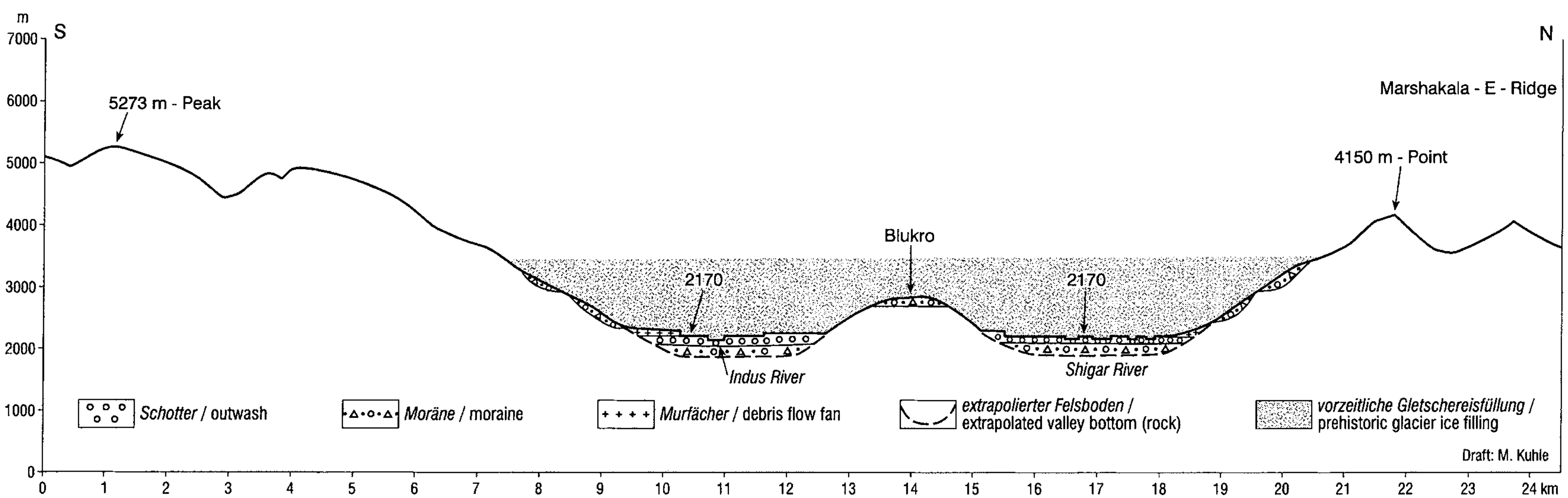


Figure 45. C. 21 km-wide cross-profile (not exaggerated) of the middle Indus valley, here of the over 36 km-long Skardu Basin between the 5273 m-peak (a NW-satellite of the 5321 m-peak, No. 77) and the Marshakala-E-ridge (on the right of No. 74), 12.3 km down-valley of the upper, i.e. eastern margin of the basin. The cross-profile lies in the confluence area of the Shigar- and Indus river between the Deosai plateau in the S on the orographic left and the Haramosh Range in the N on the orographic right valley side. In the lower 1500 altitude-metres of this profile, situated S and N of the ground-moraine-covered Blukro, the glacier ice-filling reconstructed for the LGM has left behind two classic trough-cross-profiles, the bedrock of which is also mantled with ground moraine in some places. Locality: Fig. 2/2; Fig. 2/1 between No. 77 and 74.

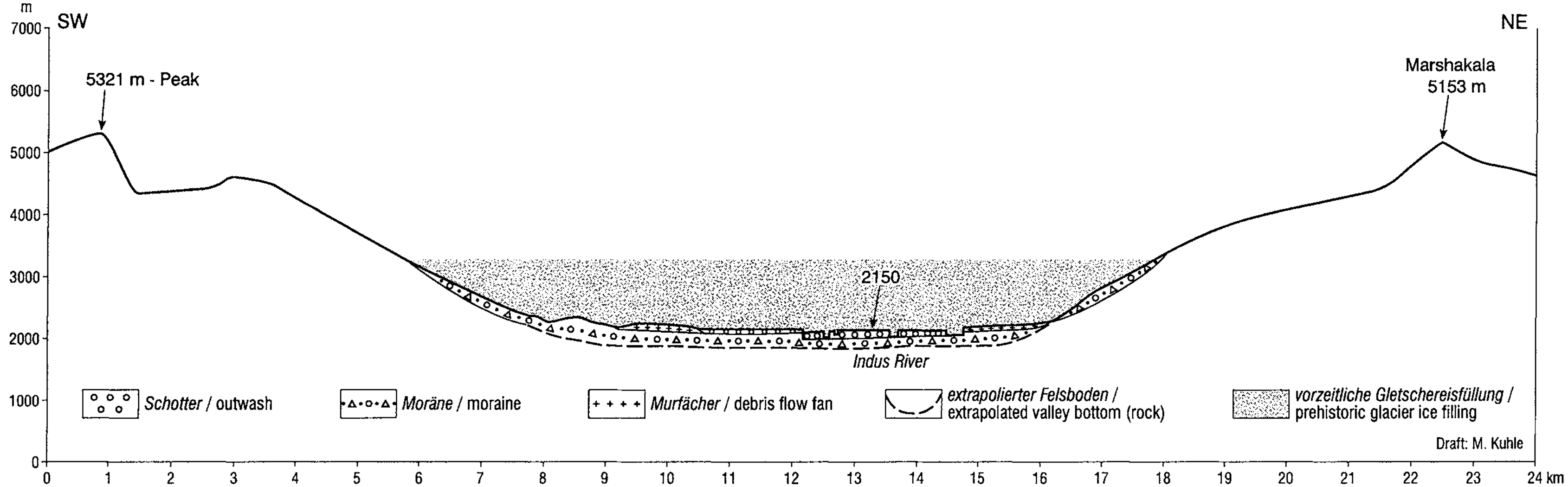


Figure 46. 22 km-wide cross-profile (not exaggerated) of the middle Indus valley, here of the good 36 km-long Skardu Basin between the 5321 m-peak (No. 75) and the Marshakala (on the right of No. 74), 16 km above its lower NW-exit leading into a further stretch of the Indus Gorge. The basin lies between the Deosai plateau in the SW on the orographic left and the Haramosh Range in the NE on the orographic right valley side. In the lower 1400 altitude-metres of this profile the glacier ice-filling reconstructed for the LGM has left behind a classic trough cross-profile the rock of which is covered with ground moraine. Locality: Fig. 2/2; Fig. 2/1 between No. 75 and 74.

3.16. The highest continuous prehistoric glacier level, i.e. the maximum Ice Age (LGM) ice cover in the Basin of Skardu in the middle Indus valley (Fig. 2/1 between No. 72–74 and 75–79)

With regard to the confluence area of the Shigar- and Indus valley (Photo 146) the question arises concerning the existence of an Indus main valley glacier. Oestreich (1906: 72 Fig. 25), who has already diagnosed the glacigenic roche-moutonnée- landscape of the Strongdokmo saddle in the confluence area of the Shigar valley and was confirmed by the author (Chapter 3.15), also points to a roche moutonnée next to the Narh settlement at the exit of the Skardu Basin situated to the east (ibid.: 61 Fig. 19). Accordingly, an Indus main valley glacier has reached the Skardu Basin. This finding, too, has been confirmed by our observations which contribute further data. So, for instance, metres-thick remnants of ground moraine have been found up to a height of c. 3200 m asl on the orographic right side of the Indus valley far above the Narh settlement, 5–7 km up the Indus, in the dead angle of the 3981 m-spur in the area of the inflow of the Shigar valley (Photo 147 ■ small on the right below No. 73; Fig. 2/1 below No. 73). Orographic right deposits of ground moraines of an Indus parent glacier like this reach up to the flank of that 3981 m-spur (Photo 147, the two small ■ on the left below No. 73). A ground moraine complex, channelized by rills and dispersed into earth-pyramid-like pillars, even reaches a thickness of several decametres at a height of 3300 m, i.e. 1100 m above the Indus valley bottom (↑). It has been preserved on a steeply sloping, glacigenically abraded rock face. The highest glacigenic abrasion areas here, form a continuous polish line about 3400 m (– – fine, below and half right below No. 73; Fig. 2/1 below No. 73). On the orographic left side the flank polishing is preserved on a glacigenically triangular-shaped face (Fig. 2/1 below No. 73) even up to 3700 m (– – fine on the very right). On this valley side ice influx has come from a 4200-4600 m-high remnant of an old face with self-glaciation (0– –). During the LGM the ELA (snow-line) did not run higher than 3800 m asl. Remnants of flank abrasions with corresponding rock roundings (Fig. 2/1 on the right above No. 77), but also remnants of ground moraines, have been preserved on the orographic left side as far as the inflow of the Satpare Lungma (◤ on the very right; ■ medium-large and large on the right of No. 73). These glacigenically erosive and accumulative remnants are interrupted by postglacial crumblings, which develop rock falls here, and, somewhat up the Indus valley, in part attain the dimensions of rock avalanches (▽; Photo 146 ⇩ and △; Fig. 2/1 on the right above No. 77) (cf. Hewitt 1999: 222 Fig. 2 No. 4–7). These rock spallings, which naturally also carry along deposited moraine, ought to be understood as a result of the slope-steepening glacigenic flank polishing. Owing to this, they are described here – as well as in the entire preceding glaciogeomorphological text – as glacigenically-prepared 'crumblings', which have been developed after deglaciation. According to this genetic classification, those crumblings are indirect glacial indicators.

The hose cirque in the Shinkar summit-wall is a convincing indicator of the ELA (snow-line altitude) (Photo 147 ○; Fig. 2/1 No. 73). It reaches as far down as 3400 m; the medium height of its catchment area is c. 5100 m, so that the orographic ELA can be calculated according to the formula: mean altitude of frame-ridge (m asl) − altitude of tongue end (m asl)/2 + altitude of tongue end (m asl) = 5100 − 3400/2 + 3400 = 4250 m asl. The present-day ELA in a SW-exposition runs here at c. 5100 m; this is evidenced by the couloirs filled with perennial firn and firn-ice, which are situated above the cirque bottom (above ○). Correspondingly, the snow-line of the prehistoric cirque glacier, which has formed this hose cirque, ran 850 m lower than the current ELA. An ELA-depression by 850 m is characteristic of the Dhampu- Stage III (cf. Tab 1). Accordingly, at that time the glacier tongue last reached the lowest preserved area of the bottom of that hose cirque at 3400 m asl. During the LGM, when an ELA-depression of 1200–1300 m existed to the north as well as to the south (Kuhle 1994b: 266ff; Fig. 138; 1997: 121, 153ff; Figs. 39–45), this hanging glacier tongue has reached the reconstructed (see above) Indus glacier level at 3400 m (– – below No. 73). During the early-Late Glacial Stages I and II the already lowered Indus glacier level has perhaps even just been reached – or even just not been reached – by this cirque glacier. This exemplary calculation has been carried out for a SW-exposition which is unfavourable for glaciation. It has been undertaken to make clear that during the LGM the hanging glaciers, flowing steeply down from the over 5100 m-high surrounding mountains (Fig. 2/1 No. 51 and 72–79), have reached the Indus glacier surface in the Skardu Basin, situated between c. 100 and 600 m below the snow-line. The two lowest cirques (Photo 146 ○ white; Fig. 2/1 on the right of No. 77) indicate a significantly lower orographic snow-line in the shady N-exposition.

In correspondence to the reconstructed Ice Age glacier level about and over 3400 m asl, the roches moutonnées rising up to 2925 or even 2984 m (Fig. 2/1 on the right below No. 74 and on the left below No. 73) and the rock hills in the Skardu Basin (Photo 145 ▲ black and white on the right; 147 ▲ black on the left) lay under a glacier cover of 400 m at minimum during the LGM. Their glacigenic abrasion forms, analysed from different perspectives (cf. Photos 145–150), confirm this more (Photo 145 ▲ on the very left and right; 146 ▲ on the very right; 147 ▲ on the left and right below No. 73) or less clearly (Photo 145 ▲ black and white below No. 76; 146 ▲ black and white on the right of No. 76; 147 ▲ on the left and 148; 149 ▲ white and 150). Thus, the arrangement of the positions of abrasion forms on isolated 'riegels' (barrier mountains) in the Skardu Basin, prove a minimum height of the LGM-ice level of over 3000 m (apart from the flank abrasions and deposits of ground moraines on the valley flanks of this Indus valley-section which have already been introduced). This observation is confirmed by erratic boulders and ground moraines preserved on several rock hills. The erratic boulders on the roches moutonnées of the Strongdokomo La have already been discussed (Chap. 3.15). The two rock hills north of Skardu, isolated from the valley flanks and flowed round by the Indus (Photo 147 and 149), show overlays of ground moraines and erratics (the two ■ left of No. 73; Photo 148 ■ on the right and ↓ ↓; 147 the two ■ on the left; 145 ■ below No. 76; 150 ■ and ↓; 146 ■ below No. 76; Fig. 2/1 above No. 77). The author agrees with Lydekker (1881: 48) and Oestreich (1906: 70–73 and Fig. 20) who approach these accumulations as moraines; the latter has also mapped their position in an approximately correct way. However, the ground moraine overlay on the eastern 'riegel' (barrier mountain) is not only limited to its western margin (Photo 148 ■ on the right), as described by Oestreich (ibid.), but it covers the entire rock 'riegel' as far as its NE-side (Photo 145 ■ below No. 76). In addition, the accumulation reaches not only up to 300 m above the Indus, but up to 480 to 690 m on the eastern 'riegel' (Photo 148) and 240 to 440 m on the Karpochi, i.e. the western 'riegel'. The two researchers suggest that – probably because of the minor altitude of that moraine overlay – the Ice Age 'Bascha-Braldu-glacier' has come to an end here in the Skardu Basin. The author, by contrast, reconstructs an ice filling in the Skardu Basin which was at least four-times thicker, and an Indus parent glacier which has completely flowed through this basin and continued NW of the Ponedas settlement down into the Indus gorge (Fig. 2/1 on the left of No. 72) (see below Chap. 5).

On the SSW-slope of the Karpochi the loose rock, approached as moraine, is exposed. It is a substrate rich in matrix with a high density, into which polymict boulders up to metres in size are embedded in isolation from each other. Erratic granite boulders are among them (cf. Photo 150 ↓ ↓); edged and rounded boulders are assorted. These deposits (see also Photo 149 ■ black; 150 ■) as well as the accumulations on the Blukro (Fig. 45), the eastern of the two central 'riegels' (Photo 147 the two ■ on the left; 148 ■ on the right; 149 the two ■ white below No. 73), give the impression of a mantling with ground moraine of both rock hills by an Indus glacier overflowing them from E to W. The small-scale combination of these accumulations with the well-preserved glacigenic rock abrasions and -polishings points in the same direction (Fig. 2/1 above No. 77; Photo 145 ▲ below No. 76; 146 ▲ black; 147 ▲ on the left; 149 ▲ black, large; 150 ▲ white). According to the presently active wasting away of these abrasion roundings in the form of fresh crumblings (Photo 147 ⇩; 148 ⇩), their convergent emergence as weathering forms has to be ruled out, because they must be prehistoric features the development of which has no equivalent here today. The freshness of these crumblings is made obvious by the fact that they interrupt (⇩) continuous rock roundings (Photo 147 ▲ on the left) as well as by the debris slopes, which they have developed below (Photo 148 ▽) and which are adjusted (○) to the current level of the Indus-riverbed (□). The central, completely separated position of the Karpochi and Blukro (Fig. 45; Photo 145, 146 below Nr.76; 149 on the left margin and on the left below No. 73; 147; also the Photos 153 and 155 show the Karpochi's 4 km-distance to the S-flank of the basin), 4 to 5 km away from the marginal slopes of the basin, and their relatively important height above the basin-bottom on which this loose material is situated, namely 240 to 440 m, i.e. 480 to 690 m, preclude its genesis by the deposits of rock avalanches as far as possible.

The easternmost section of the orographic right flank of the Skardu Basin has already been introduced as to its remnants of ground moraine and the glacigenic flank abrasions preserved (Photo 147 below No. 73). So, too, glaciogeomorphological indicators like this continue downwards of the junction with the Shigar valley to the W and then NW (Fig. 2/1 below No. 74). There, flank abrasions (Photo 149 ■ black on the right below No. 74) reach up to c. 3450 m asl (— — white); (Fig. 45 on the left below the 4150 m-point). At the place, where the Shigar valley flank turns into the Indus valley flank (— — fine on the left), the glacigenic polish band is covered with ground moraine up to approximately this altitude (Photo 148 ■ below — —; Fig. 45 on the left below the 4150 m-point). It concerns covers several metres in thickness into which postglacial flush rills are set in. Ground moraine overlays over relatively large expanses of the smoothed and rounded abrasion faces (Photo 151 ■ and ▲) have also been preserved on the left (to the E) and right (to the W) of the junction with the Marshakala or Strandokmo Lungma, as well as on the flanks of the tributary valley. If even the rock breaks away at many places since the deglaciation (⇩), the ground moraine must have been eroded the more intensively. Accordingly, its covers have been furrowed or – below erosional channels in the rock – removed by the short but heavy discharge of the rainwater as far down as the bedrock (on the right and left of ■ on the very left). The highest finding of moraine is preserved here at c. 1270 m above the Indus (↑). Judging by the glacigenic form of the side valley, an Ice Age Strandokmo tributary glacier has reached as far as the Indus glacier surface – despite its S-exposition. For this, an orographic ELA at c. 4200 m asl was necessary, i.e. an ELA depression of c. 1000 m.

On the opposite, southern flank of the Skardu Basin (Photo 152), ground moraines have been preserved in niches up to a height of 3350 m, but also in remnants of decametres in thickness at the foot of the slope (■; Fig. 2/1 above No. 77; Fig. 45 on the right of the 5273 m-peak). Flank abrasions as roundings of banking edges are only verifiable at a small scale (▲). At a large scale these remnants of abrasions have merged to a polish band with a polish line at c. 3400 m (— —; Fig. 45 on the right of the 5273 m-peak). The process of thawing-down of the glacier margin here since the LGM, i.e. the highest glaciogemorphologically verifiable glacier level (see Fig. 45) becomes clear through a pedestal moraine, which 'grows out of' the valley flank (■ white, large). It concerns the pedestal moraine of a steep hanging glacier flowing down from a cirque (○) (Fig. 2/1 above No. 77). This Late Glacial pedestal moraine has been accumulated step by step from dislocated ground moraine of the main glacier (Indus glacier). At the same time it shows the characteristics of a ground moraine ramp over which the cirque glacier has flowed into the main glacier. During a last stage such a ground moraine ramp or overthrust ground moraine ramp has been pushed against the main glacier as only an end moraine and then, kame-like, built-up against it. Owing to the vast catchment area of the Indus glacier, showing branches of valley glaciers of many hundred kilometres in length, this has still occupied the Skardu Basin at a time when the tongue of the cirque glacier had already retreated from its surface and come to an end high up in the valley flank. NW of the exit of the Satpara Lungma, for which Hewitt (1999, Figs.7,8) has evidenced a large rock avalanche (Photo 153 □; Fig. 2/1 above No. 77), the orographic left flank of the Skardu Basin (Indus valley) discussed here, continues. At the mountain spur between the exit of the Satpara Lungma and the western parallel valley, the 'Burji Lungma', four areas of rock break-out (⇩), separated from each other and controlled by bc-clefts, are situated. These rock avalanches have been approached as postglacial crumblings (Fig. 2/1 on the left above No. 77) of glacigenic flank abrasions. Hewitt (ibid. Fig. 2, Tab 1) has marked this area as '2.' and indicated the accompanying rock avalanche deposits as 'Astama'. The correspondingly precipitous rock faces and -edges of these breakages (Photo 153 ⇩) contrast clearly with the rounded rock areas (▲) of the main valley flank which continues down the Indus. This is the glacigenic flank abrasion reaching up to an altitude about 3400 m (– –).

This height of the polish line, which provides evidence of the Ice Age glacier level, increases by 100 m at minimum in the orographic left side valley, the 'Burji Lungma' (– – below No. 78). Accordingly, the flank polishings (▲; Fig. 2/1 on the right of No. 78) there testify to a tributary glacier, the level of which was adjusted to the Indus parent glacier, which it has joined. The same applies to the Satpara tributary glacier stream, the polish line of which rises continuously from the Skardu Basin as far as the Deosai Plateau (– – below No. 77; Fig. 2/1 on the left of No. 77; cf. Photo 157-160). Obviously there existed a simultaneous outlet glacier flowing down from the Deosai Plateau and joining the Indus glacier (see Chap. 4). The orographic left flanks of the 'Burji Lungma' merge in a rock spur, which in a classic fashion has been shaped glacigenically. It has been sharpened from both sides by the merging ice streams and also – in accordance to the brittle medium provided by the glacier ice – rounded by abrasion (second ▲ from the left). Downwards from c. 3200 m, metres- to decametres-thick ground moraine has been deposited on this spur (■ below the second ▲ from the left). In the course of 14 km from the inflow of the 'Burji Lungma' to the NW, along the left flank of the Skardu Basin, the LGM glacier surface reconstructed by means of polish lines and the highest remnants of ground moraine (all ■ on the right of No. 78) has dropped by c. 150-200 m (Fig. 2/1 on the right of No. 78-79).

Corresponding indicators are the characteristically triangular- shaped faces themselves, forming the basin flank (Fig. 2/1 on the right of No. 78–79). These are truncated spurs polished-back between the inflows of the side valleys, as can also be observed in the currently glaciated areas of the Baltoro-, Biafo- and Chogolungma glacier (cf. Photo 28, 66, 121). There, they are still partly in the process of development and show the rock- specific variations of this key form.

Out of the four species of ground moraines described of the orographic left flank of the Skardu Basin (see text of Photo 153), the highest preserved, less-thick remnants are most important for the reconstruction of a maximum ice level and glacier thickness. On the triangle-shaped face SW above the Tindschus settlement, they reach up to 3300 m (the two ■ black on the left above ■ white; Fig. 2/1 on the right above No. 75). The cross profile through the middle Skardu Basin depicts this overlay of loose material, too (Fig. 46 on the right below the 5321 m-peak). The probability and even the possibility that loose material is concerned here which differs from that of ground moraine, is low, i.e. it can be ruled out. A rock avalanche would only have a narrow catchment area here; a break-out scar is lacking. A gravitational mass-movement like this, once released, would not have been deposited on this solid rock slope, poor in friction, with an approximately steady steepness of c. 32°, but would have fallen at least as far as the slope foot without slowing down. There is also no roughness or a hilly toma-(debris-)landscape on the surface, as is typical of the debris of rock avalanches (cf. Photo 153 on the right of □). The geomorphological opposite is true, because there exists a cover of loose material smoothed on the surface (■ black above ■ white), which has not even been subsequently dissected by rills. A glacier, however, like a huge bricklayer's trowel, spreads its ground moraine evenly and without friction and thereby smoothly over the rock slope areas. For this reason and because there is no alternative, the loose rock can be diagnosed sedimentologically and geomorphologically as being ground moraine.

The fact, that the ground moraine covers have been deposited as far up as the glacier level, gives proof of a glacier surface at 3300 m that lay below the simultaneous snow-line (ELA).

The ground moraine ramps on the lower slope, bent like segments of trough profiles (■ below No. 75 and 79), have been developed simultaneously with those high ground moraine covers. As far as their surface is preserved, it is smooth, too. However, the very thick ground moraine remnants discharged from the side valleys, as e.g. the 'Bainsah Lungma' (■ on the left below ▲ on the left of No. 75) and the steep side valleys SW of the Tindschus settlement (■ white), are younger. They have been thrown up by the side glaciers of these side valleys, which are still partly glaciated, against the Indus main glacier in the form of pedestal moraines or ground moraine ramps (Fig. 2/1 on the right of No. 75 and 76) and accumulated to an important thickness. This has taken place during the Late Glacial (Stage I–IV) when the main glacier had already melted down and receded at least from the higher parts of the valley flanks. Norin (1925 Tavl.5) has also described part of these accumulations, namely those at the exit of the 'Bainsah Lungma', as moraines, but as end moraines. In his map they fringe two small basins of the tongues of side valley glaciers at the exit of the side valley. However, he assumed the Skardu Basin (Indus valley), to have been glacier-free during the Ice Age.

In the downward-continuation of the first cross-profile through the Skardu Basin (Fig. 45) and also down-valley the Karpochi and the junction with the Marshakala Lungma, the ground moraine mantle on the orographic right valley flank is obvious (Fig. 2/1 No. 74; Fig. 46 on the left below the Marshakala). So, for instance, these debris slopes of ground moraine have been observed at the junction of the gorge of Kuardu, where Godwin-Austen (1864: 23 and 26) has already found 'lacrustine deposits' and 'conglomerates with huge blocks' about 4000 feet, i.e. 1200 m above the Indus, which he considered to have been transported here by the river. Norin (1925, Tavl.5), too, has mapped his so-called 'Indus conglomerates' at approximately this locality, but less far up the slope. Whilst these two authors assumed a glacier-free Skardu Basin, Oestreich (1906: 73) has interpreted these accumulations – at least up to 300 m above the current Indus level – as being moraines. According to the author's opinion this is correct. Lydekker (1883: 36f) agrees insofar as he comes to the result that the Shigar glacier has reached the Skardu Basin, and at the same time overflowed the hills of the Blukro and Karpochi. On the other hand he agrees with Godwin-Austen with regard to a lacrustine, i.e. fluvial deposit at Kuardu, which he confirms up to a relative height at 300 m (Lydekker 1883: 67).

The very thick, locally many times reworked orographic right ground- to lateral moraine deposit at Kuardu and further down-valley (Photo 154 second ■ from the left) inevitably shows a Late Glacial character as to its great variety of the features preserved. There are indications of pedestal moraine ramps (Fig. 2/1 between No. 72 and 74) and at some places of kames with interposed glaciofluvial sediments. Its most highly situated remnants (see above) provide evidence of an LGM-glacier level at 3300 m asl (▬ ▬ on the left; Fig. 46 on the left below the Marshakala). The corresponding orographic left moraine deposit (Fig. 46 on the right below the 5321 m-peak) basally forms the very thick moraine pedestal in the area of the Schagari and Tindschus settlements (Photo 153 ■ white). As a tapering-out ground moraine cover, it reaches as far as a relative height of 1200 m (3300 m asl) (■ black below No. 76–79). 12 km downwards of this cross-profile (Fig. 46), the jagged, almost 3100 m-high rock head named Katsala (N-satellite of the Phara, No. 79) is situated above the Koneore settlement (Tsok) at the north-western end of the Skardu Basin. The ground moraine remnant preserved there (Fig. 154 ■ white; Fig. 2/1 above No. 79) documents that it has been mantled and overflowed by the Indus glacier (▬ ▬). Glacier traces in the form of ground moraines (the left ■ black) have also been preserved on the orographic right flank of the Skardu Basin above the Shot Kore and Kandore settlements. In the meantime, several of the glacigenically rounded flank abrasions are half-free (▲) of moraine (Fig. 2/1 below No. 72). Here, towards the lower end and exit of the Skardu Basin, the LGM-glacier level has been lowered to c. 3100 m asl. This decrease in thickness of the Indus glacier to c. 1000 m is a result of the abrupt increase of the gradient at the exit of the basin, deriving from the gorge-stretch of the Indus valley which sets in between the Poneda and Gulcho settlements. However, this decrease in thickness in consequence of the increasing velocity of flow, has been counteracted by the reduction of the width of the valley base from c. 7 km (Fig. 46) in the basin to c. 1.3 km at the head of the gorge near Gulcho. It is not necessary to discuss once more the interpretation of Godwin-Austen (1864), Dainelli (1922), Norin (1925: Tavl.5), Owen (1988b) and Haserodt (1989: 196/197), who considered the valley-obstructing, coarse-blocky accumulation of Katzarah (Kachura) (Photo 154 ▽) to be an Ice Age end moraine. This has already been disproved by Hewitt (1999), and the author agrees with him in every detail.

The Indus glacier, i.e. the orographic left side glacier from the Shigarthang Lungma, which just reaches the bottom of the Indus valley, did not come to an end here during the maximum extension of the ice – as these authors have assumed – but, according to the author's reconstruction, the Indus glacier had a thickness of c. 1000 m here during the LGM and flowed down the gorge stretch of the Indus. It cannot be ruled out that Late Glacial end- and lateral moraine material of the Stages II–IV (cf.Tab 1) is buried under the edged, coarse-blocky debris of the Katzarah rock avalanche. The ice of the Late Glacial Stage I (Ghasa-Stage) which belonged to an ELA-depression of 1100 m, and probably also that of Stage II (Taglung-Stage: ELA-depression 1000 m, see Tab 1) has completely covered the bottom of the Skardu Basin and left the basin through the gorge, continuing downwards.

3.16.1. Summary of the maximum Ice Age glaciation of the Skardu Basin (Indus valley) and its ice level (Fig. 2/1 between No. 79, 74, 73 and 77)

Whilst Lydekker (1883: 36f), Oestreich (1906: 73) and Dainelli (1922: Plates V and LIV) have reconstructed a Shigar glacier – the latter also an Indus main glacier – which has reached the Skardu Basin and was somewhat more than 300 m thick during the LGM (cf. also Haserodt 1989: 199), the author, who also refers to an Indus parent glacier, into which the Shigar glacier has flowed, reconstructs an ice thickness of c. 1000 to 1500 m (Fig. 45 and 46) in the Skardu Basin – 1500 m at its upper start in the E and 1000 m at the lower end in the WNW. The thickness of the bottom of loose rock in the Skardu Basin and thus the exact height of the rock-bottom underneath, can only roughly be extrapolated from the steepness of the plunging outcropping valley flanks about 1900 m asl (Fig. 45 and 46). At the then ice level, the Skardu Basin was c. 34 km-long and up to 13 km-wide. At its upper start in the E, above the inflow of the Shigar valley, its glacier-ice-filling reached up to 3500–3700 m asl, in the middle up to 3300–3400 m asl (Fig. 45 and 46) and at its exit in the WNW up to 3100 m. At the upper start of the basin this High Glacial glacier surface only just reached the level of the snow-line (ELA), but in the middle and lower basin area it lay 200 to 600 m below, so that several strings of surface moraine could have existed. In the confluence area of the Shigar- and Indus glacier, in the continuation of the Shinkar WSW-crest (Fig. 2/1 diagonally left below No. 73), the most important medial moraine string might have been developed from the merging lateral moraines of these two large tributary streams above the Strongdokmo La. The position of the glacier surface below the snow-line is evidenced

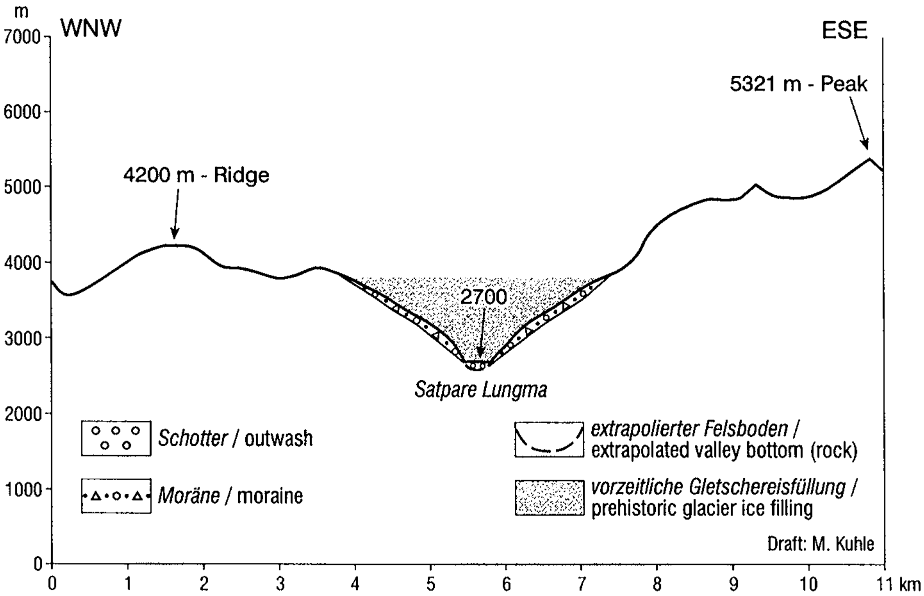


Figure 47. Cross-profile (not exaggerated) of the lower Satpare Lungma 4.8 km above its exit, W of the 5321 m-high summit (No. 77). The ground moraine covers on the slopes provide evidence of a Satpare glacier being a NNE-outlet-glacier of the Deosai plateau with a minimum thickness of 1100 m during the LGM. On the immediately connected valley slopes the then glacier feeding areas above 3800 m were too small to develop this thick Satpare Lungma glacier, so that the mass balance points to an outlet glacier. Locality: Fig. 2/2; Photo 157 before ᗜ; Fig. 2/1 on the left of No. 77.

due to the fact, that the ground moraine on the valley flanks reaches up to the polish line and thus up to the ice level (see Fig. 46).

4. The Ice Age glaciation (LGM) of the Deosai plateau (Fig. 2/1 between No. 80-84 and 75-78)

4.1. The Satpare (Satpara, Satparu, Satpura) Lungma (valley) (Fig. 2/1 on the left of No. 77)

Oestreich (1906) and then also Dainelli (1922), Norin (1925) and Schroder et al. (1993) have approached the Satpara Tso (lake) as being dammed-up by a terminal moraine. Dainelli and Norin have classified this moraine as belonging to an ice margin position of an Ice Age outlet glacier of the Deosai plateau. The first has suggested, too, that it belonged to his '4th Advance'. After Dainelli, the '3rd Advance' of the outlet glacier has left behind its deposits in the area of the Skardu settlement. The massively disturbed gravel fan- and lacrustine sediments there have been approached by Bürgisser et al. (1982), Owen (1988a, unpublished) and Cronin (1989) as 'glacitectonic or disturbed till', so that the conclusion of Dainelli (1922) was confirmed. This contradicts Hewitt (1999: 228–231) who not only queries the interpretation as 'glacitectonic', but also considers the accumulation damming-up the Satpara lake as being a 'rock slide-rock avalanche'.

The author nearly completely agrees with Hewitt's opinion. 'Nearly', because besides the correctly described rock avalanche deposits (Fig. 2/1 on the left above No. 77; Photo 155 □) and river terraces (○) etc. (cf. Hewitt 1999: Fig.7), accumulations of ground moraine do also exist in different altitudinal positions and states of preservation above those rock avalanche deposits on the orographic left side, 2 to 3 km up-valley of the Satpara Tso (Photo 155 ■ white on the left and the two first black from the left). With regard to the arrangement of the positions, the glacigenic abrasion forms and rock roundings are also connected with them (☗; Fig. 2/1 half-left above No. 77). They provide evidence of an LGM-ice level (▬ ▬ white) about 3400 m asl, so that an ice thickness of c. 1000 m has been reconstructed. This suggests at the same time that the Satpara Tso rock avalanche(s) ('Satpara Lake-Skardu rock avalanche(s)') has (must have) taken place in the Late Glacial up to the postglacial period, i.e., more exactly, that it (or they) occurred in this area after the deglaciation of the Satpare Lungma glacier during the Late Glacial Stage III (Dhampu Stage, cf. Tab 1). Accordingly, the observation of Oestreich (1906: 77), who has described 'considerably rounded gneiss boulders below a rock avalanche talus' here, can be integrated, too, as being correct. To be precise, high-lying ground moraine material is concerned here (Photo 156 ■), which has come down together with the rock avalanche(s) and survived relatively undamaged.

Thus, a tributary glacier, the Satpare Lungma glacier, has joined the Indus glacier (the ice filling of the Skardu Basin, see above) and still had a surface height about 3500 m (▬ ▬) in this junction. Therefore, no 'narrow gorge of the Satpua Valley' (Hewitt 1999: 228) exists here at the exit of this valley, but rather a classic trough valley (Photo 157 ᗜ; Fig. 2/1 on the left of No. 77), the upper slopes of which have been oversteepened by glacier flank erosion and, owing to this, have been prepared for the rock slide-rock avalanche(s).

4 to 5 km up-valley, ground moraine covers (■; Fig. 2/1 on the left of No. 77) testify to an increase of the LGM-glacier level as far as an altitude about 3800 m (▬ ▬ right half of the panorama), so that the ice-thickness of the Satpare Lungma glacier has reached a good 1100 m here (Fig. 47).

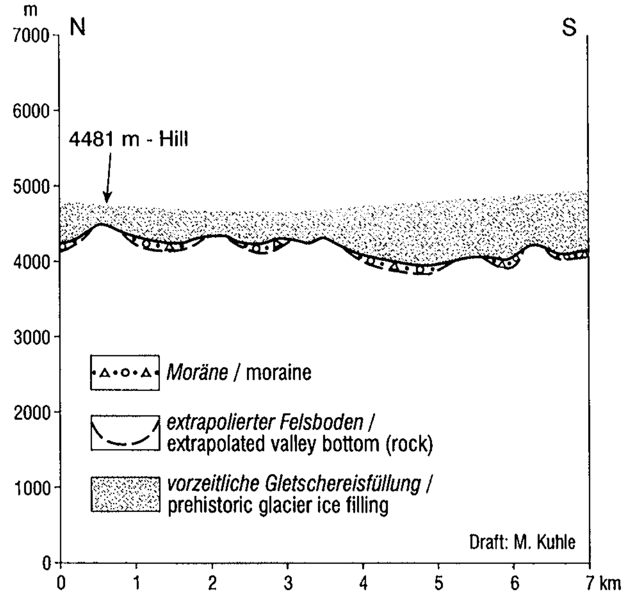


Figure 48. One of three exemplary profiles (not exaggerated) in the area of the Deosai plateau (see also Fig. 49 and 50); in order to leave the signature legible, the ground moraine, however, has been represented as thicker than it is. The hilly to highland-like surface, with a difference in height of 700 m on this NS-profile, shows the area N of the 4481 m-hill, depicted in Photo 161. The LGM-glacier-surface has been reconstructed according to the suggestion that it must have run above the hills with and without a ground moraine cover, which have been abraded by the glacigenic ground polishing. This is hypothetical insofar, as the exact ice thickness is no longer verifiable. Due to the fact that the ELA lay below the height of the plateau, the ice thickness is realistic. The depression of the glacier surface is a result of the ice which increasingly flowed down to the near Satpare Lungma on the NNE margin of the plateau. Locality: Fig. 2/2; Fig. 2/1 in the middle between No. 77,80 and 84.

The tributary glaciers joining from both sides (▬ ▬ white and ▬ ▬ black on the very right; Photo 158 ▬ ▬ on the very right) were impounded against the main glacier up to the same level of the glacier surface. This, too, is proved by ground moraines (Photo 157, the two ■ on the left) and glacigenic rock abrasions on the slopes (☗ on the very left and right; Photo 158 ☗ on the very right). Accordingly, the moraine remnants on the orographic right lower slope of the main valley (Photo 157 ■ II–III) are classified as belonging to the Late Glacial Stages II and III (cf.Tab 1). At that time, when the ELA- (snow-line) depression had decreased to a mere 1000 to 800 m and the ELA ran here, on the NNE margin of the Deosai plateau, at c. 4000 m, the Satpare Lungma outlet glacier has only just reached the valley exit, i.e. the Skardu Basin. The valley bottom of the Satpare Lungma has been made up from the late Late Glacial sander (Stage IV, gravel field No. 1) and last from the current gravel floor (□ black) as well as from alluvial fans (▽ large and white), so that the ground moraine is completely covered here (Photo 158 □).

In the valley chamber connected up-valley, remnants of ground moraine pedestals are preserved (■ white and below ■ black, small, on the right) which point out that before the accumulation of the gravel floor on the valley bottom (□) has taken place, the ground moraine was dissected and partly removed by the Satpare river. Whilst the ground moraines have survived mainly on the lower slope sections, i.e. reach at most half of the slope height (e.g. ■ right of the right (⇩, 500 to 700 m above the valley bottom), abrasion roundings (☗) and upper polish lines (▬ ▬ white on the left) indicate a slightly increasing LGM glacier surface in an up-valley direction (0▬ ▬) and an ice thickness about 1000 m. On the orographic right the Late Glacial glacier level, already described 3.7 km down-valley, can again be observed in the lower third of the slope (■ II–III; Fig. 2/1 on the left of No. 77).

Another 3 km further up-valley, an end moraine is situated which is especially developed on the orographic right side and can be approached as a Late Glacial ice margin position of the Sirkung Stage (Photo 159 ■ IV; Fig. 2/1 half-left below No. 77) at 3200 m asl. This height is consistent with an ELA about 4200 m at that time. The material of the end moraine has been analysed sedimentologically (Fig. 22) and morphometrically (Fig. 6 No. 16) and confirmed, so that a rock avalanche or a mudflow accumulation as an alternative can be ruled out. On the orographic left valley side ground- and lateral moraine remnants of the next-older Dhampu Stage (■ III) are preserved. Besides sporadic remnants of ground moraines, the older, higher-lying Late Glacial (Stage II–I) glacier traces as well as those of the valley glacier with the verifiable maximum altitude of the glacier level (▬ ▬0), have left behind flank polishings and -abrasions (■ ☗) as well as glacigenically triangle-shaped slopes (Fig. 2/1 half-left below No. 77).

A valley section of c. 5 km in length follows, which, glaciogemorphologically speaking, is comparatively sterile. Due to debris tali from surficially crumblings, deriving from easily-weathering mica-schists which outcrop above, it shows a V-shaped valley cross-profile. Nevertheless, high-lying glacigenic flank abrasions are obvious (Fig. 2/1 on the left below No. 77). The two glacigenic side valleys connected to the orographic right, and a cirque floor, also situated to the E, are evident (Fig. 2/1 below No. 77).

In the upper reaches of the Satpare valley, i.e. where at 4100 m the Satpare Lungma comes down from the Deosai plateau (Fig. 48) leading to the NNE, a typical start of a glacier valley has been preserved (Photo 160). It is as characteristic as those in the former Norwegian fjell- and high plateau areas when they were completely glaciated by an inland ice sheet. The upper 5 km of the valley show a trough valley cross-profile (Fig. 2/1 on the left below No. 77). Whilst the first 2 km are rather flat (left half), the following 3 km down-valley are already similar to an Alpine U-shaped valley with basal debris tali (▽) and upper slopes which are roughened by crumblings in the bedrock (on the right and left above ▽). As for this valley head, the roche-moutonnée-like glacigenic abrasion knobs on the culminations of the valley flanks (☗ white and black in the middle and half-right; Fig. 2/1 on the left below No. 77) are evidence of an LGM-ice surface (▬ ▬) which had priority over the relief. This points to a corresponding snow-line at a height markedly below 4000 m, probably about 3700 to 3800 m, because only several 100 m above the ELA, the glacier ice on the bottom of the valley head – which, as happens here, leads down rather steeply –, is able to rise above the valley flanks.

4.2. The prehistoric glaciation of the Deosai plateau (Deusi high plateau)

Where the geomorphologic high plateau character sets in, a large-scale to completely covering ground moraine landform can be observed (Photo 160 second ■ from the left; 161 ■). Especially on the up to km-wide high valley bottoms (white) and in the smaller depressions between the roches moutonnées and abraded glacially streamlined hills (☗; Fig. 2/1 between No. 77, 80, 84), the ground moraine creates a complete cover. Towards the rock knobs the moraine decreases (■ black) and in places the polished rock is devoid of it (☗ on the very left; Fig. 48).

Here, the pioneering feat of Oestreich (1906: 86–87, 91) has to be emphasized. He has reached the Deusi plateau, as he called it, not through the Satpare Lungma (see above) – as we did – but over the Burje La, following the Burje creek (in the Burji Tschu) from the Skardu Basin up to the N margin of the high plateau. Crossing the Deosai plateau from here, he describes a prehistoric glacial landscape with these characteristic words: 'Alle Formen sind sanft gerundet, wellig und kuppig, alles ist von dem einstmals darüber hinweg streichenden Eise geschliffen worden.' (ibid.: 86). However, his assessment as to the expansion of the glaciers and the ice thickness connected to it is contradictory, when he writes: 'Die Deusi waren zur Eiszeit ein gewaltiger Eisbehälter, und es scheint, dass sie nicht etwa nur ein Firnbecken waren, sondern vielmehr eine wirkliche Gletscherausbreitung, nach Art der Vorlandgletscher.' (ibid.: 86). Accordingly, a piedmont glacier of this type would only have flowed down from the mountains fringing the plateau to the plateau area and thus into the mountain foreland. Later, however, he says '. . . , zu Zeiten mögen die Gletscher individualisiert gewesen sein, zu Zeiten verschmolzen zu einer allgemeinen Gletscherbedeckung. Die Deusi stellten alsdann ein grosses Eisreservoir dar, und das Eis strömte nicht nur in der natürlichen Abzugsrinne, im Schigartal, hinab, sondern, wie ich vermute, auch über die Erniedrigungen der Umrandung.' (ebd.: 87). The latter he proves with the help of erratics. According to our findings, the second of the two glaciation variations was true for the LGM, because a continuous ice cover had been built up at a thickness of c. 300 to 700 m above the mountain ridges or even 900 m above the high valley bottoms (Fig. 48–50).

Evidence of ground moraine (e.g. on the profile-line in Fig. 48) is provided by metres-sized (Photo 161 □), far-travelled boulders – because they are facetted to rounded – in a polymict composition (○) in or upon a clay- and silt-containing matrix (■ white). Besides the non-sorted composition of the material the undulating surface with its non-uniform, medium to flat incline also contradicts a fluvial accumulation. Mudflow dynamics (debris- or mudflows) on the plateau have to be ruled out, because catchment areas providing the necessary quantities of debris- and water at a sufficient difference in height and gradient of the ground, can also be excluded. In addition to gneiss boulders which may originate from the underground and which, together with conglomerate blocks, Oestreich (ibid.) has already suggested to be morainic, sedimentary rocks as e.g. quartzite (Photo 161) and granites also occur in the boulders.

In the centre of the Deosai plateau (Fig. 2/1 in the middle between No. 83 and 84), where the creeks join to form the Shigar river (Photo 162), the valley bottom has been fluvially remoulded (□, ●). On the slopes glacigenic abrasion roundings can be evidenced (e.g. ☗ on the very left) as well as sedimentologically classic ground moraines (■ e.g. on the very left; Fig. 23) with portions of quartz grains from which up to 100% are glacially crushed (Fig. 6 No. 17). On the valley bottom below, i.e. in the talweg area, the accumulation of a glaciofluvial terrace landscape (Photo 162) has inset since the Late Glacial deglaciation, i.e. after the Taglung Stage (II; see Tab. 1). Here, glaciofluvially washed and displaced ground moraine material is concerned, sedimentated in the form of two Late Glacial glacier mouth gravel floor terraces (□ 2 and ● 1). Between these oldest and highest terraces the gravel floor bodies No. −0 to −2 and last the historic to recent gravel fields No. −3 to −8 have been deposited (□ −0−−8) (Tab 1) (Fig. 2/1 between No. 83 and 84) during the neoglacial period. The postglacial reshaping on the Deosai plateau, however, is only restricted to loose rocks like these. Also in this respect it is similar to the 500 to 1000 m higher Tibet (Kuhle 1988g, 1991a, 1999). The glacially streamlined hills, as well as the roche-moutonnée-like rock knobs (☗) set upon them, and the completely preserved trough valleys and transfluence passes with trough-shaped cross-profiles (⋃) have survived without any modification (Fig. 2/1 between No. 83, 82, 84). Thus, the roundings of the highest mountain ridges provide evidence of a covering LGM-glaciation of the centre (‒ ‒; Fig. 49). This is indicated, too, by a further exemplary locality with a ground moraine overlay and a surficial scatter of relatively large, far-travelled boulders (Photo 163, 164, Fig. 2/1 on the right above No. 81). The insignificant degree of weathering of the boulder surfaces – despite the freeze-and-thaw climate on the highland – points to a Last Glacial, more exactly, Late Glacial sedimentation during the Taglung Stage (II) c. 15,000 years ago.

3 km W of this W/E-profile studied in detail (Fig. 49), beyond a hill ridge, a further segment of the landscape (Photo 165) has been observed, which cannot be explained fluvially or periglacially. The valley slanting to the S, is flowed through anastomisingly by the orographic right tributary river of the Schigar river. It has created a valley floor of boulders by washing-out the ground moraine (□). At the same time the valley slopes have been undercut by lateral erosion, so that loose material has also been displaced from there. It is metres-thick material (■) consisting of polymict, partly erratic, rounded boulders (○) 'swimming' in a clayey-silty matrix (■ black on the very left). Owing to this, weathering in situ has to be ruled out. Sedimentologically speaking these diamictites bear all the characteristics of far-travelled material, which at the same time has not been fluvially displaced. They are considered to be moraines and – geomorphologically – ground moraines. The two terrace forms (■ black below ‒ ‒ middle and ■ below No. 78) are also made up from loose rock and can most easily be interpreted as being remnants of a ground moraine pedestal. In

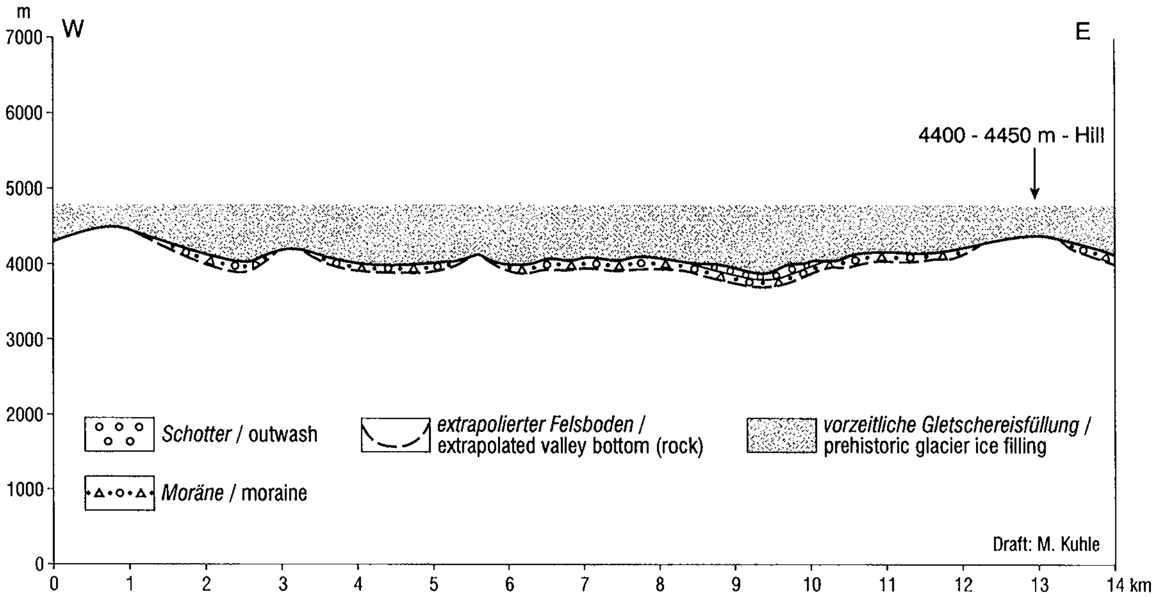


Figure 49. One of three exemplary profiles (not exaggerated) in the area of the Deosai plateau (cf. Fig. 48 and 50). The gravel cover and the ground moraine are represented as thicker than they are so to leave the signature legible. The hilly to highland-like surface with a difference in height of 600 m on this WE-profile shows the area depicted in the Photos 163 to 166. The LGM-glacier-surface indicated has been reconstructed according to the suggestion that it must have been situated above the hills with or without a ground moraine cover, which have been abraded by the glacigenic ground scouring. This is hypothetical insofar, as the exact thickness of the ice is no longer verifiable. Due to the fact that the ELA ran c. 400 m below the height of the plateau, the ice thickness of 200–900 m at a compact outline of the high plain of 24 to 28 km on average is realistic. Locality: Fig. 2/2; Photo 163–166; Fig. 2/1 on the right below No. 85 and on the right above No. 81, both at the same distance.

this case, the ground moraine pedestal would have been cut by the river down to a depth of about 15 to 20 m and removed over the entire width of the river flood land (□). At the same time, the fine material would have been flushed out so that the coarse blocks would have been displaced solely vertically downward, remaining consolidated in the river flood land (□). A different interpretation of these terraces is that they might have been lateral moraines on both sides of a late Late Glacial valley glacier during Stage IV (Tab 1), which have lastly been reshaped to form lateral kames. In this case the terrace would have been created by the accumulation of slope material against the valley glacier, which had melted-down to a flatter, narrower shape, across the original lateral moraine.

The abraded fjell-landscape (☗) makes an LGM-glacier surface (= Stage 0 = Würm glaciation) about 4800 m probable (Fig. 49); this also becomes obvious 3 km more to the W down-valley (Photo 166 – –; Fig. 50). Here, the ground moraine landscape can immediately be diagnosed by the large-scale scatter of boulders transported by the glacier ice (▲; Fig. 2/1 half-right above No. 81) as well as by the typical later fluvial reshaping on the slopes (▽); this applies also to the forming of the glacially streamlined hills and the polish thresholds (☗; Fig. 2/1 half-right above No. 81). The flushing rills (▽ white) are of a minor age and they testify to a complete geomorphological change of the regime compared with the rounding and accumulating processes during the Ice Age. It is impossible that these have been periglacial, because currently, i.e. at the time of the rill development, solifluction also takes place here. This becomes evident through drop-shaped solifluction garlands (Photo 167 below ☗ black on the right) and – together with

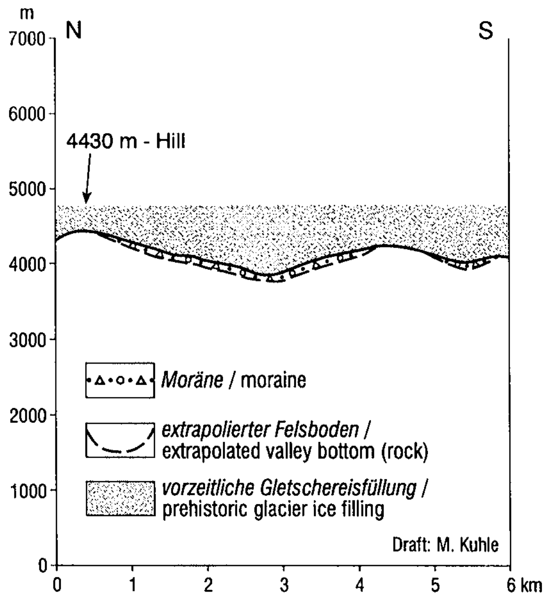


Figure 50. One of three exemplary profiles in the area of the Deosai plateau (see also Fig. 48 and 49); the ground moraine has been represented as thicker than it is so to leave the signature legible. From the high valley bottoms as far as the upper slopes, the ground moraine becomes thinner and toward the hill-tops it breaks off completely. This is a typical effect of ice covers which have priority over the relief. The ice overlay concerns the LGM. Locality: Fig. 2/2; Photo 166; Fig. 2/1 on the right below No. 85 and on the right above No. 81, both at the same distance.

the roughening of the small-scale exposed bedrock (above ☗) by frost weathering – brings about an only slightly reshaping of the glacial landscape. Nevertheless, the morainic character of the loose-rock overlay has been exemplarily investigated on representative hill slopes (Fig. 24; 6 No. 18; Fig. 2/1 on the right below No. 85).

The Scheosar Tso (lake; Photo 167) shows all features of a glacigenic pass lake as they are typical of transfluence passes formed by glacier scouring (well-known Alpine examples: Toten- lake on the Grimsel-pass, Upper-Engadine-lakes on the Maloja-pass etc.). Its over-deepened basin has been created by ground scouring of the overflowing ice, i.e. as a polish depression between the polish thresholds (⌓). The speed of the ice discharge forced by the western plateau margin might explain the locality of the lake.

Whilst on the eastern N-margin of the Deosai plateau, where the Satpare Lungma sets in (Photo 160 left half) and the ice discharge into the Satpare Lungma outlet glacier has taken place in a soft gradient curve of the underground (Chapt. 4.1), developing a large funnel which has sucked away the ice, i.e. a large-scale depression in the glacier surface of the plateau (Fig. 48) – here, on the western margin of the plateau, the Sartschur La (pass), a rock threshold (Photo 167 ⌃; Fig. 2/1 on the right above No. 81) does exist, over which the ice has flowed down abruptly and steeply into the Das Khilin Gah (valley). However, here too the LGM-ice-level must have dropped from c. 4800 m on the plateau (Fig. 50) to c. 4450 m asl (– – on the very right). This means that the ice above the roche moutonnée of the marginal threshold (⌓ white) would have been only 130 m thick. This is a minimum thickness for the development of distinct roche moutonnée forms like these. Because this plateau area lies c. 600 m above the LGM snow-line altitude, a greater ice thickness could be suggested without any problem.

4.3 The prehistoric glaciation of the upper Das Khilin Gah deriving from an LGM-outlet-glacier, which flowed down from the Deosai plateau to the WNW (Fig. 2/1 below No. 85)

The upper 13 km of the Das Khilin Gah and the connected side valleys, as well as source depressions of a still minor order, have been glacigenically abraded and rounded as far as above the valley flanks (Photo 168 ⌓). There rounded rock steps (second ⌓ white from the left) and roches moutonnées occur (the first ⌓ black and fourth ⌓ white from the left) as well as truncated spurs with glacigenically triangle-shaped faces (third ⌓ white from the left and first and third ⌓ black from the right) and preserved trough cross-profiles (U) (Fig. 2/1 below No. 85). Further evidences of accumulation are provided by the ground moraine covers, observed from the talweg up to at least 400 m up the valley slopes (■; Fig. 2/1 above No. 82). Thus, a valley glaciation is proved empirically, which also continues on this side of the transfluence pass (⌃) leading from the Deosai plateau to the W (Sartschur La). The Das Khilin Gah glacier is to be understood – as just the synchronous Satpare Lungma glacier (see above) – as an outlet glacier of the Deosai plateau.

Besides the glacigenic abrasion forms discussed, typical 'tors', which have survived the Late Glacial glacier scouring, are also preserved (▲). The development of this 'tor'-example (▲) has been prepared by subglacial meltwater erosion below the Late Glacial snow-line. During the LGM the ELA ran at c. 3600 m (in the Nanga-Parbat-massif 50 to 60 km to the W at 3400 to 3600 m asl, Kuhle 1997: 123) and during the Late Glacial Stages II to III about 300 to 500 m higher, i.e. 100 to 300 m higher than in the area of the 'tor' (▲). Therefore at 3850 m asl a subglacial discharge had already been developed here at that time, dissecting the trough valley bottom (between below ⇩ and on the right of ▲) (root of a subglacial ravine: Fig. 2/1 below No. 85). So, the steep-walled 'tor' might have been made up in this way, i.e. by fluvial erosion below the glacier. In this glaciogeomorphologic and paleoclimatic connection, given by the glacier indicators and the ELA, the Late Glacial lateral- and end moraine at 3900 to 3600 m (Photo 168 ■ IV; the second ■ IV from the left is a corresponding ground moraine remnant or kame) is appropriate. It has to be classified as belonging to the Sirkung Stage (Tab 1) (Fig. 2/1 IV below No. 85).

4.4. Summary of the prehistoric glaciation on the Deosai plateau

The detailed geomorphological findings in the field, as well as the sedimentological laboratory analyses, testify to a complete glacier cover of the hilly relief of the high plateau up to a level at 4700–4900 m asl. This is the surface height of the maximum Ice Age glacier level (Fig. 48–50). Thus, the nearly round glacier face of the Deosai plateau amounted to c. 24 km in diametre. The glaciation, which, according to the freshness of its forms and sediments, belongs to the last Ice Age, provides evidence of a snow-line-depression as far as below 4000 m asl. This is the medium hill- and high valley bottom surface of the plateau. Calculated with the help of current small mountain glaciers and firn patches (Photo 161,162,166 No. 80–82 and 84) the present-day climatic snow-line runs about 4800 m. This means that the plateau was glaciated up to Stage III (the Late Glacial Dhampu Stage according to Tab 1), because at a large-scale the ELA-depression amounted to 800–900 m. During the LGM (Stage 0, Tab 1) the snow-line has even run c. 400 m below the height of the plateau. In a semi-arid climate this means a difference in temperature of nearly 3°C (2.8°) on the glacier ground. This points to a cold-based ice lying on the Deosai plateau during the LGM, and to a warm-based during the Late Glacial. The glacigenic abrasion forms preserved are rather smoothly-polished and softly-shaped (Photo 161, 164, 166, 167 ⌓), whilst cold-based ice, which from time to time is frozen to the underground, produces rough exaration- and detraction forms. Accordingly, the final Late Glacial ice cover might have been the cause of their development.

The outlet-glaciers, which during the LGM (= Würm = Stage 0) were decakilometres-long and flowed down radially from the central Deosai-plateau-glacier into the valleys leading down from its margin, as e.g. the Satpare Lungma or the Das Khilin Gah, reached as far as 2500 and 2400 m and then joined the next larger main glaciers, the Indus- and the Astor glacier respectively. During the Late Glacial Stages II or III (cf. Tab 1) the terminus of the Satpare outlet-glacier has even just reached the Skardu Basin (Indus valley), whilst the

Satpare Lungma glacier came to an end at 3200 m asl during Stage IV without having flowed through half the length of the Satpare Lungma (Fig. 2/1 IV at No. 77). During this final last Late Glacial Stage IV, when the ELA ran about 4200 m asl, i.e. already 200 m above the Deosai plateau, the Satpare Lungma glacier probably was no longer an outlet-glacier but a local valley glacier. Now it was mainly fed by the local, shortly-connected, steep hanging- and side glaciers on the orographic right, flowing down from the comparatively high 5032 m- and 5058 m-peaks (Fig. 2/1 below No. 77). Probably, the Deosai ice had already melted down over large parts, so that the Satpare Lungma valley head was nearly free of ice. This does not correspond to the ice margin position of Stage IV in the Das Khilin Gah (Fig. 2/1 IV below No. 85). Here, the high summits are lacking and the highest Deosai plateau area, a good 4200 m-high, is situated only 4 km away.

Generally, the minor reshaping of the glacial landform of the Deosai plateau since the deglaciation which actually only took place in the loose material, has to be stressed. Exemplary for this are the Late Glacial to postglacial glaciofluvial terraces, i.e. gravel plains on the high valley bottoms (Photo 162), the insignificant development of mudflow- and alluvial debris fans (Photo 164 ◁ Fig. 2/1 on the left above No. 84) or the glaciofluvial flushing of ground moraine pedestals and the corresponding compaction of the boulders contained (Photo 165 □). The cirque development observed in the marginal mountain-chains (Photo 162 ○; Fig. 2/1 No. 81) occurred repeatedly during the Pleistocene, early- as well as late-glacially. It last took place in the last Late Glacial, i.e. during Stage IV (Sirkung Stage, cf. Tab 1) when the ELA ran about 4200–4300 m.

4.5. The maximum Ice Age glaciation (LGM) of the middle and lower Das Khilin Gah as the large right side valley of the Astor (Astar) valley next to Nanga Parbat (Fig. 2/1 on the left of No. 85–87)

The middle Das Khilin Gah presented here is a rather well-preserved glacial valley with a characteristic trough cross-profile (Photo 168 ᗜ on the right; Fig. 2/1 No. 85). Characteristic abrasion forms of glacier flank polishing are also the triangular-shaped slopes on the back-polished mountain spurs between two side valley junctions (▲ on the right of No. 82; Fig. 2/1 No. 85). Additionally, mantlings of ground moraine are verifiable on the slopes (■ small, on the very right; Fig. 2/1 No. 85,86). The highest, unambigously reconstructable valley glacier surface (– – on the left below No. 86) has run in the valley chamber of the Jidim settlement at c. 3900 m, i.e. about 600 m above the valley bottom (Photo 169 – – on the left). The LGM-snow-line at c. 3600 m asl (see Chapt. 4.3) definitely lay below the glacier surface, so that a glacier feeding area the type of a firn stream glacier has existed as far as Jidim and still further down-valley. Accordingly, glacier ice- and -firn lay also on the higher valley flanks (Photo 168 – – below No. 82 and 86–85).

From the orographic left three glacigenic trough valleys (Photo 169 below No. 81, ᗜ and left of it; Fig. 2/1 No. 81–82

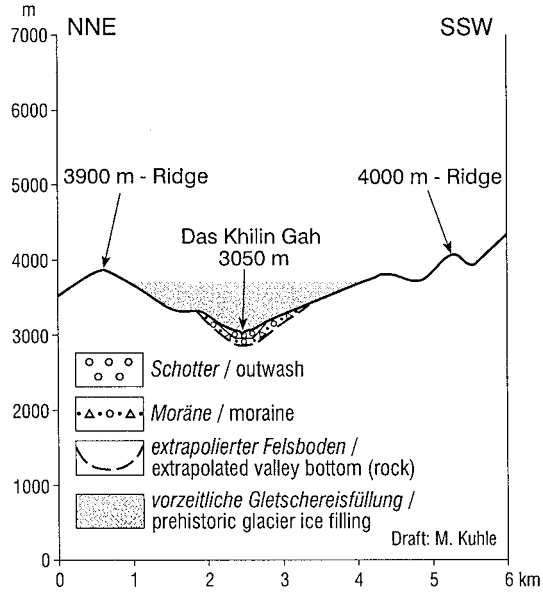


Figure 51. Valley cross-profile (not exaggerated) through the middle Das Khilin Gah seen facing NNE (orographic right flank) to SSW (left flank) up-valley; to leave the signature legible, the gravel cover and the ground moraine have been represented as thicker than they are. The lower half of the LGM glacier- filling of the valley is documented by moraine deposits. On the valley bottom they are covered with the youngest Late Glacial gravel floor (No. 1; cf.Tab. 1). The upper half of the glacier ice-filling as far as 3700 m is proved by traces of glacigenic abrasion. Locality: Fig. 2/2; Photo 169; Fig. 2/1 half-left below No. 86.

and above 82) join the valley chamber investigated in detail, which leads from the Jidim settlement up to 3.2 km further down. So, the Das Khilin Gah becomes a main valley of a next higher order. Here, the orographic left flank, i.e. the mountain spur on the first intermediate valley ridge, has been abraded as far as beyond its culmination (▲ white; Fig. 2/1 on the right of No. 82). This also applies to the abraded slope as far as the next-lower orographic left valley junction (▲ black on the left), which has therefore also to be approached as a glacigenically triangular-shaped face (Fig. 2/1 above No. 82). On the orographic left a third polished mountain spur (▲ black on the right) confirms a highest LGM-ice level, which, following the valley incline, has dropped from 3900 m to 3700 m asl (– – and – –0; Fig. 51) in this valley section. On the lower slopes Late Glacial ground moraine remnants and the remains of lateral moraine ledges have been observed (the three ■ black and white above; Fig. 2/1 III on the left below No. 86), which, considering the arrangement of their position to the end moraine of Stage IV (Photo 168 ■ IV, Fig. 2/1 IV below No. 85), belong to the Dhampu Stage (III, cf. Tab 1). On the steeper trough valley flanks late Late Glacial (Stage IV) to postglacial rock crumblings (Fig. 2/1 on the left below No. 86) have heaped up debris tali over ground moraine cores since the deglaciation (Photo 169 △).

From this locality and the corresponding trough-valley-like cross-profiles (Fig. 51) 15 to 16 km down the Das Khilin Gah, the Taramdas settlement is situated. In this valley section ground moraine remnants have been preserved on both valley slopes (35°10'25″ N/74°58'40″ E; 2700 m asl) which are at least 30–35 m-thick. They reach as far as c. 100 m-high up the slopes – on the orographic right even somewhat higher (Fig. 2/1 on the left of No. 86). As far as this valley cross-profile and still further, up to the inflow into the Astor valley, sporadically preserved flank abrasions are evident (Fig. 2/1 on the left of No. 86 up to left of No. 87). The trough valley flanks are interrupted by the junctions of side valleys. As a result, glacigenically triangular-shaped slopes have been formed, created by back-polished mountain spurs of intermediate valley ridges. The valley bottom of the Das Khilin Gah is made up from a gravel accumulation with terrace steps, interpreted as being a Late Glacial gravel floor (glacier mouth gravel floor or sander) (Fig. 2/1 on the left below No. 87). In some places it might cover deposits of ground moraine. The large orographic right side valley leading from the Khume settlement ('Khume Lungma or Tschu') up the main valley, is also a typical trough valley (Fig. 2/1 between No. 86 and 87). The massifs of the 4848 m-peak (No. 85), the 4843 m-peak (No. 86) and the 5718 m-peak (No. 87) belong to its upper catchment area as well as northwestern marginal areas of the Deosai plateau (Fig. 2/1 between No. 86 and 83), so that at a comparable Ice Age climatic snow-line (ELA) about 3600 m asl and the same exposition, a glacier ice filling similar to that of the main valley, the Das Khilin Gah, has to be suggested anyway.

About 2.5 km above the inflow of the Das Khilin Gah into the Astor valley in the valley chamber of the Pakora settlement, (35°15'10″ N/74°53'20″ E; height of the valley bottom c. 2600 m asl) a c. 120 m-high kame-terrace is situated on the orographic right, accumulated from an orographic right side valley against the Late Glacial Das Khilin Gah glacier arm (Fig. 2/1 on the left somewhat below No. 87). It consists of displaced moraine material. A younger postglacial alluvial fan reaches through the kame-terrace, which in the meantime has been dissected by the creek of this right side valley. This kame terrace provides evidence of a Late Glacial glacier thickness of c. 130 m and might be classified as belonging to the Taglung (II)- or Ghasa Stage (I; cf.Tab 1). A maximum thickness of the Das Khilin Gah glacier, joining the Astor glacier, is proved here by the mantlings of ground moraine (Fig. 2/1 on the left of No. 87) above the kame level up to an altitude of c. 2850 m and the glacigenically triangle-shaped face (Fig. 2/1 on the left below No. 87; cf. Kuhle 1996: 155, Fig. 2 on the right above No. 31 and 32) reaching c. 100 m higher up, as well as by the adjacent glacigenic rock-abrasions. The ice was at least 350 m-thick, but it might, however, have been far thicker (see below). Directly opposite, on the NE-exposed, orographic left valley flank, classic ground moraine in the shape of a ramp formed like a trough-profile segment, confirms this finding of an Ice Age Das Khilin Gah glacier (Fig. 2/1 on the left below No. 87, on the left above Khume).

Here, however, the question has to be asked, whether the glacier ice has come from the much more important Astor valley, i.e. has flowed into this side valley from below, or whether it has followed the Das Khilin Gah from above as far down as the valley exit (see below).

But first reference is made to further observations in the field: c. 2-3 km down-valley of those orographic right kame terrace (see above), at the lower valley exit, the bridge is situated, over which the route switches to the orographic left valley flank (35°15'30″ N/74°51′ E; 2450 m asl). Here, undisturbed ground moraine deposits, which have been heavily compacted by the ice pressure, are preserved on both sides of the Khilin Gah (here also named Das Khirim Gah) river in a primary position. They contain facetted to polymict boulders in size up to several metres, which, isolated from each other, 'swim' in a fine matrix (Fig. 2/1 on the left somewhat below No. 87). Again further down-valley, the Astor valley with its valley chamber of Gurikot continues in a NW-direction. Here, the Astor river has dissected the gray-coloured ground moraine reshaped by mudflow- and alluvial fans as well as by fluvial furrowing since the deglaciation, to a depth of c. 100 m (see Kuhle 1997: 205 Photo 86 ■ on the right below; 131/132 Fig. 28 on the right of No. 35). In the context of the glacier history of the entire Nanga Parbat massif, the author has investigated in detail this area of the Basin of Gurikot in the Astor valley in the summer of 1987 (Kuhle 1988b: 588; 1988c: 11; 1989: 271–273; 1991: 299; 1993: 108–110; 1996; 1997; 1998a: 90–94. The glaciogeomorphological and sedimentological findings provide evidence of a maximum, probably LGM-ice-level in the Astor valley in the junction area of the Das Khilin Gah at c. 3700 m (Kuhle 1996: 138 Photo 1 ▬ ▬ below No. 5, 142/143, 154 Fig. 2 No. 32-36; 1997: 102/103, 131/132 Fig. 28 No. 32–36); this means c. 1250 m-high above those side valley bottom which is still partly covered with ground moraine (see above, Fig. 2/1 on the left, somewhat below No. 87). During the Late Glacial Stage I the Astor glacier thickness has still amounted to 750 m (Kuhle 1997: 203–205, Photo 85 ▬ ▬ I–II) so that the Astor glacier surface lay c. 300 m above that side valley bottom. Owing to this, we must imagine an Astor glacier which flowed into the Das Khilin Gah. The significantly higher catchment area of the Nanga Parbat-massif, which accordingly – compared with that of the Das Khilin Gah and the Deosai plateau connected to it – is at present still more strongly glaciated (climatic snow-lines (ELA): currently 4600–4700 m, during the LGM 3400-3500 m asl), has built-up an ice flow in the Astor glacier, which has probably pressed into this side valley over many kilometres before it could be compensated by the Das Khilin Gah glacier-flow down-valley toward the NE. The fact that the Ice Age (LGM) Astor glacier has met the side valley glacier up to a level about 3200–3500 m asl, may have been the cause of an ice connection between the Deosai plateau and the Nanga Parbat ice stream network. In the perspective of the mass balance, the medium height of the catchment area at an ELA about 3600 m asl (see above) was too low to allow the Das Khilin Gah glacier to flow much deeper down than a few kilometres down-valley of the Jidim settlement (i.e. up to approximately the cross-profile of Fig. 51).

From the 5718 m-peak (Fig. 2/1 No. 87) as a relatively very high, local catchment area which even today shows high valley glaciers (see Kuhle 1997: 205 Photo 86 ● below

No. 1), a steep valley glacier in continuation of the current cirque glacier flowed down the 'Pakora Lungma' toward the SW to the Das Khilin Gah valley bottom (ibid.: 131/132 Fig. 28 on the right above No. 31) during the LGM. Because of that, a glacier infilling of the lower Das Khilin Gah has taken place in addition to the ice injection of the Astor valley. The Ice Age 'Khume Lungma' glacier (Fig. 2/1 between No. 86 and 87) with its 4843 m (No. 86) to 5718 m-high (No. 87) catchment areas has also contributed to this.

Finally it has to be pointed out that Dainelli (1922), who agrees with Oestreich (1906) as to the glaciation of the Deosai plateau (Dainelli 1922: Tav.LXXVII.Fig. 2), in his Tav.CLXXVII. has already sketched out a continuous valley glacier from the Deosai plateau through the Das Khilin Gah as far as the Astor glacier, which he claims to belong to his 'Terza Espansione Glaciale'. However, he has not visited this area and has thus obviously deduced the Das Khilin Gah glacier on theoretical grounds.

5. The Ice Age glaciation (LGM) of the middle Indus valley between the Skardu Basin (Ponedas settlement) and the inflow of the Gilgit river (Fig. 2/1 between above No. 79 and above No. 92)

Dainelli (1922 Tav.V. and Tav.CLXXVII. as well as Tav.XLI Fig. 2; Tav.XLII Fig. 1; Tav.XLIII. Fig. 1 and 3) assumed for his 'Terza Espansione Glaciale', which corresponds to the Würm Glaciation or LGM, an ice-free Skardu Basin. At the same time, however, he suggested a glaciation of the lower exit of the Skardu Basin – which is discussed here – deriving from the orographic left side valley. Here he has reconstructed a Shigarthang Lungma glacier, considering the 'Kazarah rock avalanche' – as the accumulation of coarse boulders at Kazarh (or Cazzura, i.e. Kachura) has been approached by Hewitt (1999) – to be its end moraine (see above Chap. 3.16). The author, however, who also interprets this accumulation as a postglacial rock avalanche, has come to the conclusion that the LGM Indus glacier has not only reached the Skardu Basin, but has even filled it (Fig. 45 and 46) to the extent that, leaving the basin and flowing into the downward valley stretch of the Indus gorge, its ice was c. 1000 m-thick (see Chap. 3.16.1). The corresponding glacier level lay at c. 3200 m (Photo 154 ▁ ▁).

Going on from these data of the Skardu Basin, we follow only the glacier traces from the basin into the Indus gorge. Here, at the start of the gorge, i.e. W of the 'Kazarah rock avalanche', glacigenic abrasion forms are evident on both gorge flanks from the talweg (Indus) at 2145 m asl upward (Photo 170 ▲ white and black on the right; Fig. 2/1 below No. 72). The roche moutonnée, half of which has been linear-erosively worn down by a torrent (▲ white on the right), evidences once more that the smoothings in the fresh and completely unweathered rock originate from prehistoric times and that they are not to be confused with forms of weathering. Due to the state of roughening since the deglaciation, two ice levels can be differentiated: an older, higher one up to which the polishings reach, which have been more strongly roughened by weathering and which has to be classified as being from the LGM (Stage 0; ▁ ▁0); and a second, running where the fresh abrasions come to an end, so that it can accordingly be referred to as being from the oldest Late Glacial Ghasa-Stage (I; cf.Tab 1) (▁ ▁I). Moraines in situ (■), but also replaced moraines (▽), occur in the zones near to the talweg. It has not been proved that the large polymict boulders contained (◊ and ○) are partly erratic (Fig. 2/1 below No. 72). As to the findings in this valley section it has to be taken into consideration that the orographic right flank of the Indus gorge rises steeply from the valley bottom as far as beyond the current snow-line. The hanging glacier flowing down from the 5220 m-peak (a SE-satellite of the 5770 m-peak, Fig. 2/1 No. 72) in a SW-exposition, has joined the Indus parent glacier providing local moraine. Here, Dainelli (1922 Tav.CLXXVII.) has correctly mapped an orographic right side glacier for the LGM, i.e. his 'Terza Espansione Glaciale', and also an orographic left one, which has left the Turmik or Meyto Lungma (Fig. 2/1 on the right below No. 88; 1.5 km down the Indus from the cross-profile Fig. 52) and reached the Indus main valley, because here, too, a catchment area of up to 5405 m is connected. Dainelli points out that, 2 km down the Indus from this inflow of the side glacier, his Indus glacier filling, which he supposes to be only 14 km-long from the Skardu Basin up to here, has totally ended (ibid.). The author, however, assumes a c. 1200 m-thick Indus main valley glacier, verifiable by ground moraines and, further above, flank abrasions (Fig. 52), i.e. an ice stream, which, due to the inflow of side glaciers, has become 200 m thicker than in the lower Skardu Basin. In the valley chamber of the Dasu settlement (Fig. 53), the Indus glacier was even c. 1300 m-thick during the LGM: whilst the talweg declined, the glacier remained at approximately the same level. This has to be related to the substantial ice inflow from the orographic right Tormik Lungma (Fig. 2/1 on the left of No. 51) and the left Trik Lungma (Fig. 2/1 above No. 88) (Photo 171 ▽).

Here, both flanks of the Indus valley and these side valleys have risen to c. 1400–1600 m above the ELA during the LGM, so that the side glaciers, also without the Indus main glacier, would have reached the main valley bottom at about 2140–1900 m asl. Naturally, they have in any case reached and nourished the surface of the Indus glacier about 3100–3200 m. In many places this applies especially to the orographic right side of the Indus gorge with its side valleys before it has reached the Gilgit river.

Flank abrasions and glacigenically triangle-shaped faces occur on the back-polished mountain spurs between the wall gullies (Fig. 2/1 between above No. 88 and below No. 72). In several of the steepest gorge sections they are the only glacier traces indicating the glacier level by their upper limits (Fig. 52). The gorge flanks, concavely polished out by glacier abrasion are remarkable, so that a typical 'gorge-like trough' (Kuhle 1982a, 1983a) has been developed (Fig. 2/1 between above No. 88 and below No. 72). In the wider Indus valley section near the Dasu (or Smurdo) settlement (Photo 171), large portions of morainic material have been preserved on the valley bottom (■ large); ground moraine has survived on the flanks, and especially in the position

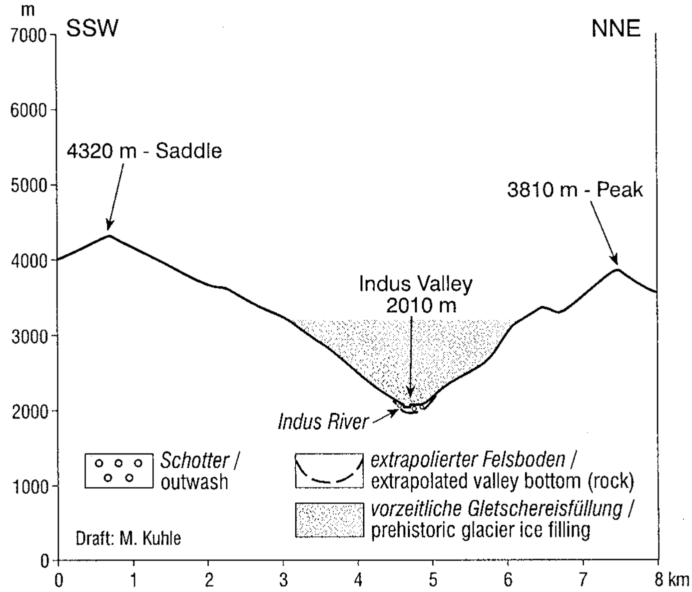


Figure 52. Valley cross-profile (not exaggerated) across the Indus Gorge from facing SSW (orographic left flank) to NNE (right flank) looking down-valley. In order to leave the signatures legible, the gravel cover has been represented as thicker than it is. Ground moraine deposits on the valley bottom have been covered or replaced by Late Glacial to recent gravel floors (No. 2 to −7; cf. Tab. 1) since the deglaciation after the Taglung Stage (II). The glacier ice filling up to c. 3200 m is evidenced by glacigenic traces of abrasion. Locality: Fig. 2/2; 7 km down the Indus from the viewpoint of Photo 170; Fig. 2/1 on the right of No. 88.

of the spur unfavourable to removal, up to 3200 m asl (■ small; Fig. 53 on the right flank: Fig. 2/1 on the left of No. 51). The moraine mass on the valley bottom profits from the pedestal moraines at the exits of the Tormik- and Trik Lungma (Fig. 2/1 between No. 88 and 51). Though the valley bottom was situated at least 1700 m below the ELA during the LGM, and despite its subglacial dissection by hydrostatically confined meltwater, developing a subglacial ravine (Fig. 2/1 half-left below No. 51), roches moutonnées have been preserved (☗ on the right; Fig. 2/1 above No. 88). In the course of the 40 km-long Indus gorge stretch continuing down-valley up to the next exemplary cross-profile (Fig. 54), the 'gorge-shaped trough- or trough-shaped gorge character' (ibid.) does not undergo a change worth mentioning (Fig. 2/1 between above No. 91 and above No. 88), i.e. it naturally changes only in the confluences of the side valleys. Two examples are provided by the junctions of the 'Tulu Lungma' (Photo 172) after 24 km and the '5090 m-peak S-valley' (Fig. 173) after 38 km. Here, glacigenically triangle-shaped slopes, developed from truncated spurs, are the rule (Fig. 2/1 below and on the right below No. 89). Owing to a lesser decline of the slopes, ground moraine remnants have survived on them (Photo 172 and 173 ■; Fig. 2/1 below No. 89). The bottoms of the hanging side valleys have been regularly dissected by subglacial ravines, developed here during the LGM up to the deglaciation (below and half-right below No. 89) (Photo 172 ↓; 173 below ▽). Between the exemplarily narrow, trough-shaped gorge profiles (Fig. 54 and 55), 3 km down-valley from Fig. 54, the Ice Age Bulache Gah glacier has joined the Indus glacier, flowing down from the currently still glaciated 5569 (or 5559; Fig. 2/1 No. 91) m-massif, which forms the dividing crest to the Astor valley, adjacent to the SW (cf. Kuhle 1997: 131/132 Fig. 28 on the right of No. 46) (Photo 174). Here, too, a larger ground moraine complex (■; Fig. 2/1 on the right above No. 91), similar to that at the exit of the 'Tulu Lungma' (see above), has been preserved on a truncated spur. This topographic situation is favourable to moraine accumulation, because the side glacier has edged out the parent glacier and thus created space for its pedestal moraine.

On the bottom of the Indus gorge roches moutonnées or remnants of them have been preserved sporadically (Photo 173 ☗ on the very right; Fig. 2/1 on the left below No. 89). The characteristic forms of dispersion, such as e.g. rills, roots of earth pyramids, moraine slides or crumblings of abraded rock, are only cursorily pointed out here; they occur frequently and regularly (Photo 171–174; Fig. 2/1 between No. 72–91).

It is nearly impossible to interpret the loose rocks on the valley slopes in the area of the next lower Indus valley cross-profile (Fig. 55) as being not glacigenic, i.e. as something different from moraines (Photo 175). A large rock avalanche has to be ruled out, because the loose rock has been laid down in several complexes which are separated from each other (■); several small rock avalanches are impossible, because, falling down the mountain flank, they must have stopped on their own in order to come to a standstill. Due to the steepness of the slopes this is unthinkable (see e.g. ■ black, small). Accordingly, a valley filling up to their height, namely about 800 m above the talweg, would be necessary to explain those accumulations as rock avalanches. However, for a huge rock avalanche such as this, the necessary break-out scar is absent. Small rock falls can be observed in many places (⇩; Fig. 2/1 above No. 91); but they have taken place since the deglaciation, because they are adjusted to the moraines (◁), i.e. have even occurred in their material afterwards. A catchment area for the development of small local hanging glaciers on the valley slopes is also

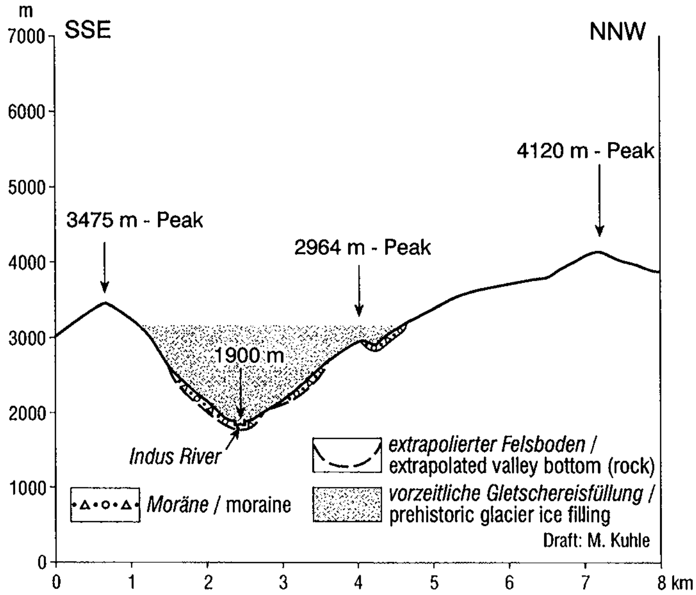


Figure 53. Valley cross-profile (not exaggerated) across the Indus Gorge from facing SSE (orographic left flank) to NNW (right flank) looking down-valley. The sketched-out LGM-glacier-filling of the valley is evidenced by deposits of ground moraine, which, even in the talweg, have not yet been replaced by a gravel cover at a thickness worth mentioning. The valley cross-profile has been widened, i.e. polished out to a trough-shape (cf. Photo 171 at). On the orographic left side, the glacier ice-filling has been evidenced by glacigenic abrasion traces up to an altitude about nearly 3200 m asl. Locality: Fig. 2/2; 3.5 km down the Indus, i.e. the talweg, from the viewpoint in Photo 171; Fig. 2/1 above No. 88.

lacking: in this valley chamber the slopes are neither high enough to reach a catchment area (cf. Fig. 55) nor do side valley forms exist, which, at an ELA about 3800-3700 m, could have reached down to the Indus valley talweg. This proves that ground moraines (■; Fig. 2/1 above No. 91) of an Indus parent glacier are concerned here. Only the feeding area of the huge connected ice stream network had an area beyond 3800 m sufficiently large for the ice to flow down to (here) 1550 m. The abrasion forms testify to an ice thickness of approximately 1600 m here. Corresponding are the conditions in the area of the example in the next lower Indus valley cross-profile, 8–10 km down-valley (Fig. 2/2 No. 56; Fig. 56; Photo 176). Hewitt (1998: 4, Fig. 1 No. 96) has mapped a rock avalanche in this valley chamber. But here, too, a break-out scar large enough to have the material surged up to over 1000 m-high is lacking. In many places (△ above) minor rock falls (⇩) have come down onto the ground moraines (■; Fig. 2/1 half-left above No. 91) on the orographic right slope.

From this cross-profile on (Fig. 56) the valley develops from a 'gorge-shaped trough' to a trough with a V-shaped profile near to the talweg (Fig. 2/1 on the left above No. 91 up to left below No. 90). Despite all structural-geological conditions which are unknown in this valley, the V-shaped profile has to be deduced from the subglacial meltwater stream, and the widening of the valley from the decreasing decline of the talweg towards the Gilgit river (see also Fig. 57 and 58).

The Puparash or Haramosh Lungma, newly reinvestigated by Meiners (2001) as to its prehistoric glaciation, is connected to the Indus valley with an exit mantled high-up with moraines (Photo 177 ■) on the orographic right (Fig. 2/1 on the left somewhat above No. 53). The LGM-level of this tributary stream, flowing down from the highest feeding areas of the Karakorum at over 6000 and 7000 m, was adjusted to the Indus parent glacier about 3100 m asl (— —0). This high level of the side valley glacier only becomes understandable by the back-damming ice filling of the main valley, particularly because the Puparash- or Haramosh Lungma is a very short and steep valley and so would be unable to contain a thick glacier without an abutment. At 1400 m, 15 km away from the confluence with the Gilgit valley, the Puparash-Haramosh Lungma glacier has once more provided the Indus glacier with an important ice mass.

Here, the main valley already bears the characteristics of a normal glacigenic trough valley (Fig. 2/1 on the left of No. 90; Photo 178). The end of the mountain spur on the orographic left had been reshaped to a 'riegel' (bar mountain) and this to a roche moutonnée (▲ on the left of large; Fig. 2/1 on the left of No. 90). During the Late Glacial (Stage IV) the space between the roche moutonnée and the spur (large) has still been reached by the tongue of the Puparash-Haramosh Lungma glacier and filled with ground moraine and moraine material. In the main valley the Indus parent glacier had already melted down. This succession of High Glacial ground polishing with abrasion forms (▲) and Late Glacial mantling with ground moraine (■) can be observed in this valley chamber in many places (Fig. 2/1 on the left of No. 90). After the deglaciation the replacement of the moraine cover (▲ and △) followed, which locally has taken place several times and in great variety, as well as the youngest fluvial aggradation (Fig. 2/1 on the left of No. 90) and its dissection (○ white). The current, i.e. interglacial glacigenic shaping, is limited to cirques (○ black; Fig. 2/1 No. 90) at an altitude beyond 4800 m asl.

7 km down the Indus valley from the junction of the Puparash-Haramosh Lungma, in the area of a glacigenically polished-out valley-cross-profile which at the level of the

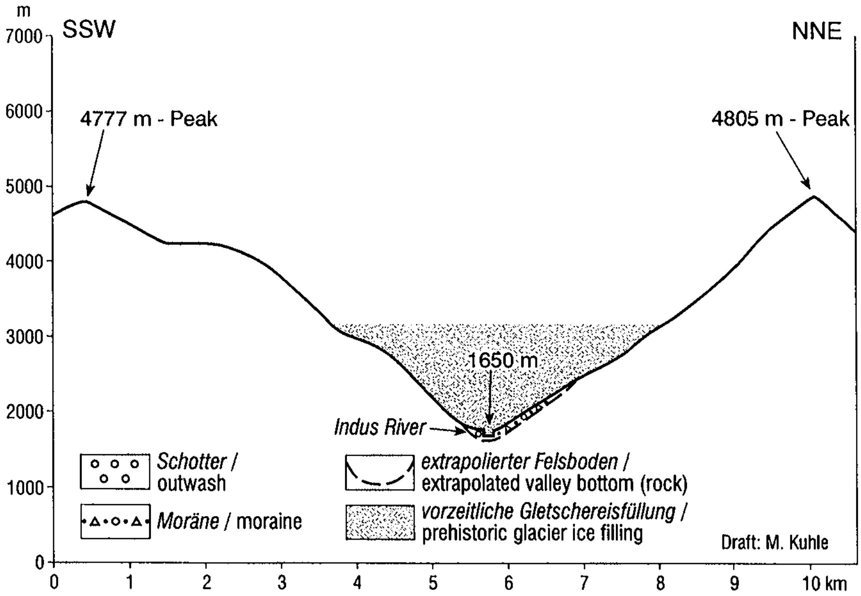


Figure 54. Valley cross-profile (not exaggerated) across the Indus gorge from facing SSW (orographic left flank) to NNE (right flank) looking down-valley. The sketched-out LGM-glacier-filling of the valley is documented by deposits of ground moraine on the orographic right as far as nearly half the height of its filling. They have been replaced by a gravel cover in the talweg since the deglaciation after the Taglung-Stage (IV; see Tab. 1) at the latest – perhaps already subglacially. The valley cross-profile is nearly V-shaped and has only been slightly concavely widened by glacigenic flank polishing over at most a few 100 m in the line of slope. On the orographic left the glacier ice-filling is completely evidenced by glacigenic abrasion forms up to a good 3100 m asl; on the orographic right it becomes obvious as to the upper 850 m. Locality: Fig. 2/2; 1.7 km down the Indus, i.e. the talweg, from the viewpoint in Photo 173; Fig. 2/1 between No. 89 and 91.

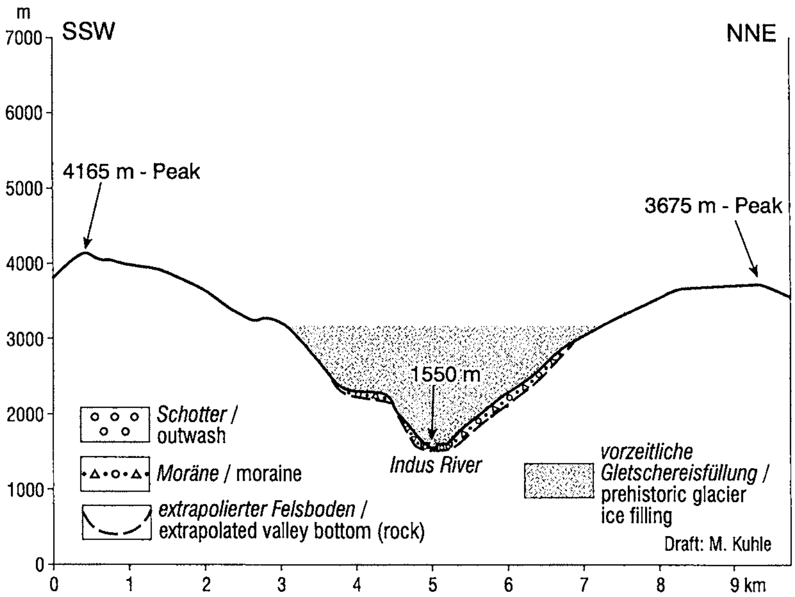


Figure 55. Valley cross-profile (not exaggerated) across the Indus Gorge from facing SSW (orographic left flank) to NNE (right flank) looking down-valley. On the orographic right the indicated LGM glacier-filling of the valley is evidenced by ground moraine covers as far as 2900 m; in the talweg they have been replaced by a gravel cover or they have already been syngenetically interrupted subglacially through a gravel floor between the left and right valley slope. The lower area of the valley cross-profile is nearly V-shaped; the upper one, above the rock shoulder on the orographic left, has been widened by glacigenic flank polishing (Photo 175). However, due to the ground moraine overlay, the true position of the surface of the rock flank can only roughly be interpolated. The glacier ice-filling up to an altitude of a good 3100 m asl is evidenced as for the upper 750 m by glacigenic abrasion forms on the orographic left. Locality: Fig. 2/2; 2 km up the Indus, i.e. the talweg, from the viewpoint in Photo 175; Fig. 2/1 immediately above No. 91.

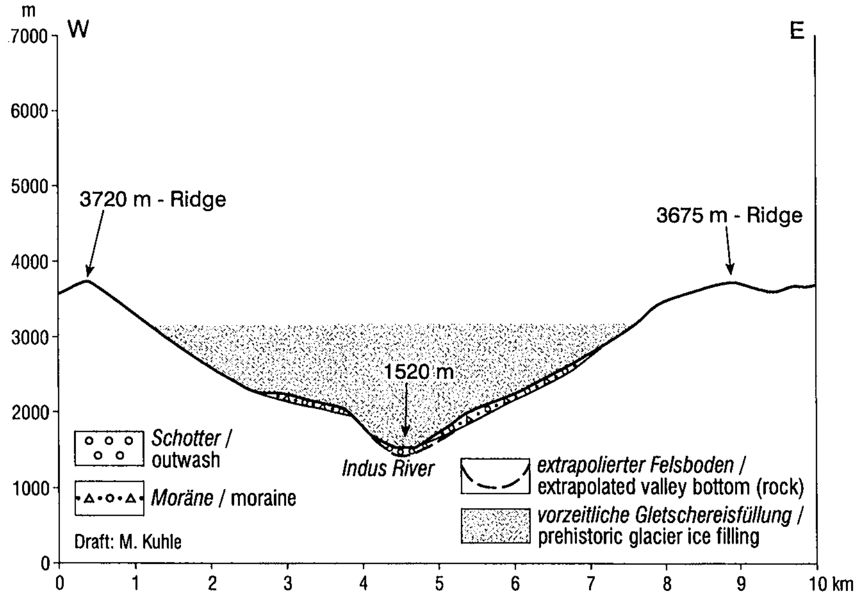


Figure 56. Valley cross-profile (not exaggerated) across the Indus Gorge above the inflow of the Gilgit river, facing W (orographic left flank) to E (right flank) looking down-valley. On the orographic right the indicated LGM glacier-filling of the valley is evidenced up to an altitude at 2650 m and on the orographic left up to 2200 m. On the orographic left side these moraine covers on the slopes are interrupted by steep rock slopes; on the valley bottom, near the talweg, they have been replaced by the deposit of gravels. Gravel layers of washed-out ground moraine like these (Photo 176 □ and ○) might be syngenetical and so could have already been developed beneath the High- to Late Glacial glacier. The lower 550 m of the valley cross-profile, however, are already trough-shaped but narrow; the upper segment of the cross-profile as far as 3200 m asl is also trough-shaped and, in addition, very wide; the lower segment has partly been formed by the syngenetical, subglacial meltwater erosion. As for the upper 500 m the glacier ice-filling up to a height of a good 3100 m asl has been evidenced by glacigenic abrasion roundings in the bedrock. Locality: Fig. 2/2; c. 2 km up the Indus, i.e. the talweg, from the viewpoint in Photo 176; Fig. 2/1 somewhat to the left above No. 91.

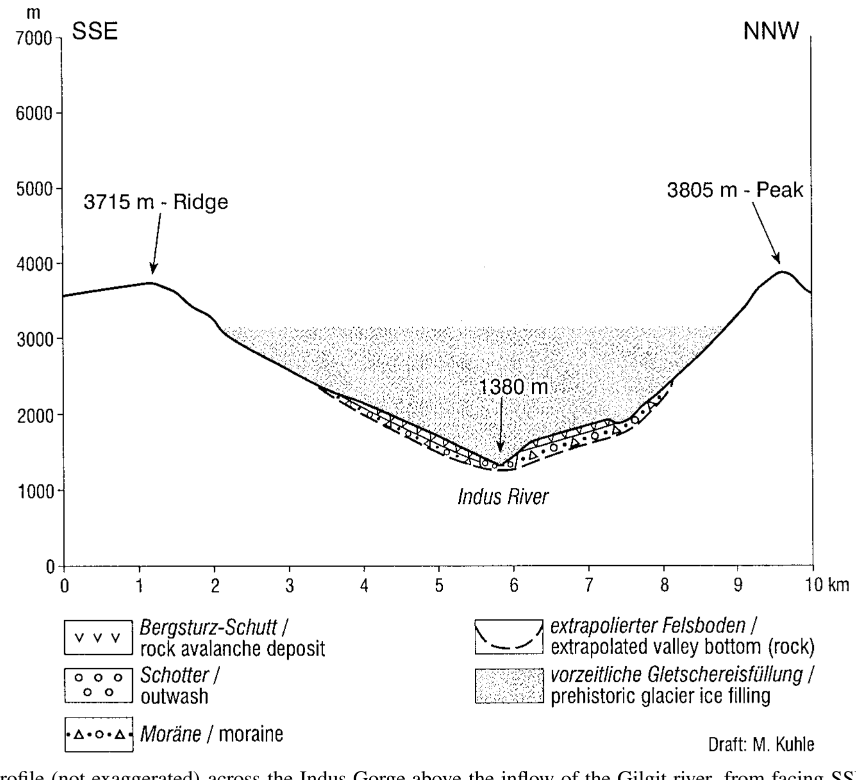


Figure 57. Valley cross-profile (not exaggerated) across the Indus Gorge above the inflow of the Gilgit river, from facing SSE (orographic left flank) to NNW (right flank) looking down-valley. The indicated LGM glacier-filling of the relatively wide trough (Photo 179 ᑌ) is evidenced by ground moraines as far as a height of c. 2400 m (orographic right). The ground moraine layers are covered by deposits of rock avalanches on both slopes; on the valley bottom, near to the talweg, they have been replaced by gravel deposits. Gravel layers of washed-out ground moraine like these (Photo 178 and 179 ○) might be syngenetical and so could have already been created beneath the High- to Late Glacial glacier. The glacier ice-filling up to a good 3100 m is proved by glacigenic abrasion roundings in the bedrock occurring on the upper c. 750 m. The trough form which, compared with the up-valley cross-profiles (Fig. 52–55), is wide, can be explained by the decreasing incline of the valley bottom (this applies also to Fig. 58). Locality: Fig. 2/2; 1.7 km up the Indus, i.e. the talweg, from the viewpoint in Photo 179; Fig. 2/1 on the left of No. 90.

LGM-ice is 7 km-wide (Fig. 57; Fig. 2/2), very freshly preserved forms of glacier polishing and -smoothing have survived (Photo 180 and 179 ▲ large). Naturally, these traces of the ice are freshest where they have still recently been covered with ground moraine (■) (Fig. 2/1 on the left of No. 90 and half-right above No. 92). On the glacier bottom, a good 2 km below the ELA and at an LGM-ice-thickness of c. 1700 m, a waterfilm has existed between ice and rock, developed on thermal grounds as well as because of the exceeding of the pressure-melting-point. Thus, the gleaming fine-polish, i.e. the shining rock-polish becomes understandable. The sporadic flush-bowls (Photo 180 ▽) document the simultaneous presence of hydrostatically confined water with a forming effect. However, exactly these polish-forms (Photo 180) belong to the Late Glacial, i.e. Dhampu- or Sirkung Stage (III or IV, Tab 1), i.e. to the time of the last prehistoric glacier cover. Naturally, the ice pressure had finally decreased, but the portions of water between the ice and the rock, necessary for the fine-polish, continued to exist. The separate flush-bowls (▽) prove, that all other forms are not fluvial and thus cannot be explained by the Indus at a higher river-level.

Here, too, the stratigraphically younger cover of ground moraine has again been worn down in many places, i.e. replaced, during a longer period (Photo 179 △); there, the polishes have already splintered-off (▲ small). The different colour of the ground moraine covers is striking: the dark one on the dark rock (■ white) mainly consists of local moraine and is covered with slope debris: the light one (■ black) is denser and erratic (Fig. 2/1 on the left of No. 90 and half-right above No. 92); replaced lake sediments are also contained. The last process of sedimentation, which has still not come to an end, is that of the rock avalanches (□; Fig. 57) and smaller rock crumblings (△ on the very left; Photo 180 ⇩) taking place since the deglaciation. It has been prepared by the glacigenic oversteepening of the valley flanks, which, as a result of the glacier erosion, have developed a U-profile (Fig. 57). Thus, the rock avalanches are the last stage of the glaciogemorphological cycle and no alternative.

In the continuing, c. 8 km-long, section of the trough valley (Photo 181 U), the Indus has again cut into the rock of the trough ground; here, too, the development of this dissection might have been caused by subglacially confined water, which, due to cavitation corrasion, had an especially erosive effect (Photo 181 ↓; Fig. 2/1 above No. 92). During the LGM the Indus glacier ground lay c. 2200 m below the snow-line. Even its surface lay 500 m below it. From the heavily torn surface of this gorge glacier, meltwater has run through all crevasses into – and finally also beneath – the glacier, and then – accelerated by enormous hydrostatic pressures – has been pressed with extremely high velocities through subglacial ice tunnels. After this linear erosion, the glacier has continued its work, polishing the ground of the valley bottom. This is proved by the abrasion roundings (▲ in the middle, below). So, the basal flank steepening in the orographic left flank of the Indus trough, down-valley of the inflow of the Gilgit river (Photo 182 ▼), can either be approached as a postglacial, fluvial undercutting by a moving outer bank of the Indus river, or as a subglacial fluvial undercutting below the joint Gilgit tributary stream and Indus parent glacier.

The four-stepped terrace landscape filled into the glacigenically shaped valley relief (from □ to ○; Fig. 2/1 above No. 92; cf. Photo 178 ○ white, 179, 181 ○) is young. It is younger than the roches moutonnées (Photo 181 ▲ in the middle below), which have naturally been released from the ice only since the deglaciation and are covered by its terrace gravels. These 50 to 210 m-high gravel bodies are composed of Holocene to historic river gravels of the Indus, dammed-up by rock avalanches and debris flow cones or -fans from the Indus valley flanks and side valleys. They have been dissected syngenetically with the cutting of overspills into these barriers of loose material by the Indus. An example of this process is provided by the debris flow fan from the Bundschi Gah (Fig. 2/1 below No. 92; Photo 182 ▽). It is built-up from dislocated moraine material (■ I–II; Fig. 2/1 on the left of No. 91) and has dammed-up the Indus several times, so that at least a few of the terrace levels concerned have been developed in this way (○ black; Photo 181 ○).

Down-valley of the orographic left mountain spur in the junction area of the Indus- into the Gilgit valley, the joint left valley flank of the Indus-Gilgit valley starts, named Indus valley from here on (right half of the panorama; Fig. 2/2 between the cross-profiles Fig. 57 and 58). This slope in the transition to that main valley of a higher order shows glacigenic flank abrasions (three ▲ on the right), just reaching 3100 m, which mark the prehistoric glacier level (– – black on the right and white fine). It corresponds to that one up-valley, marked by a clear polish cavetto (– – white on the left). Approximately half of its abrasion faces is still covered with remnants of ground moraines (three ■ white on the right; Fig. 2/1 half-right above No. 92). Currently these debris bodies are removed through rill cuttings and at the same time locally reshaped into earth pyramids (Fig. 2/1 on the right of No. 92), i.e. they are definitely prehistoric. The only different prehistoric possibility as to their development is glaciation, so that the alternative development as debris tali has to be ruled out.

According to the data extracted by the author in 1987 (Kuhle 1988b: 588), the Gilgit- and the Indus glacier have flowed together to form a joint lower Indus parent glacier (Fig. 2) in the area of this valley cross-profile. Dainelli (1922 Tav.CLXXVII.) suggested for this confluence neither an Indus- nor a Gilgit- and lower Hunza valley glacier. The author reconstructed a Hunza glacier, which, as the Manu- and Bagrot glacier, has flowed into the also glaciated Gilgit valley, so that their ice merged into the Indus glacier (cf. also Kuhle 1988a,c; 1989, 1991, 1993). The communicating glacier levels of these tributary streams have just reached a height of 3100 m (Fig. 58; Fig. 2/2). The joint glacier surface is evidenced by further abrasions and deposits of ground moraines (Photo 182 ▲ and ■; Fig. 2/1 on the right and half-right below No. 92) in the orographic left valley flank downwards of the confluence concerned. Here, this large ice stream, which was the most important outlet glacier of

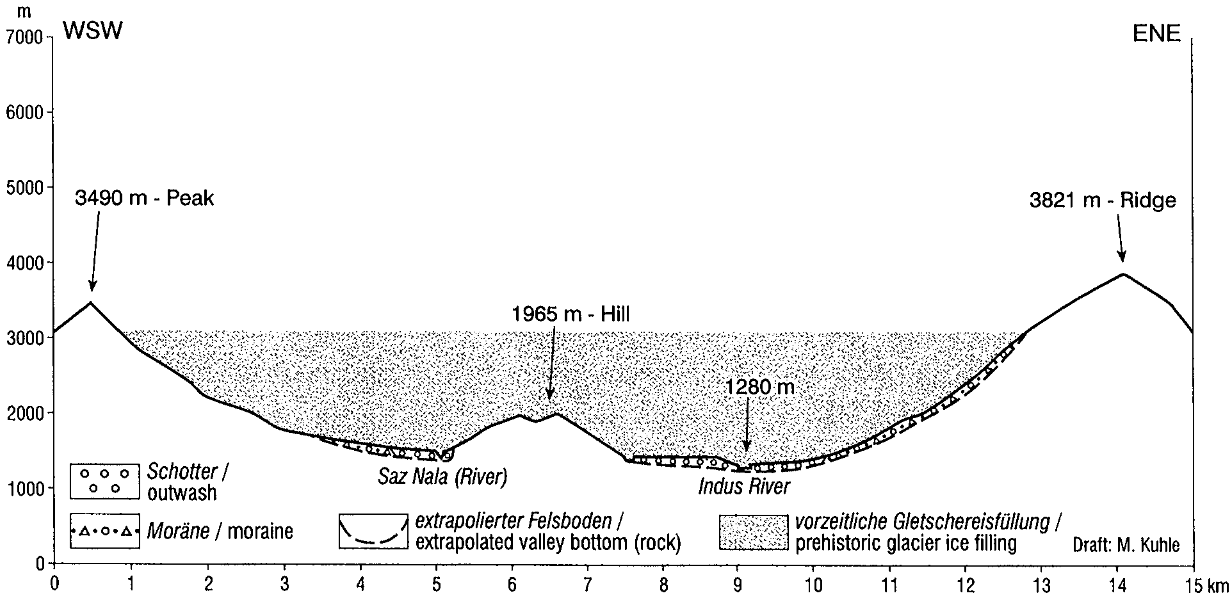


Figure 58. Valley cross-profile (not exaggerated) across the Indus valley, 5.5 km down from the inflow of the Gilgit river, in the junction area of the Saz Nala (river), facing WSW (orographic right flank) to ENE (left flank) looking up-valley. The indicated LGM glacier-filling of the c. 13 km-wide trough (Photo 182 U) was about 1800 m-thick and can be evidenced by a ground moraine cover up to an altitude of c. 3000 m (on the orographic left). This glacier ice-filling is also confirmed by glacigenic abrasion roundings in the bedrock which are sporadically exposed. The valley bottom is made up of terraced gravel deposits up to 2.5 km in width; part of these gravel layers might be syngenetical with the ground moraines and so has already been developed beneath the High- to Late Glacial (Stage 0 to I or II, see Tab. 1) glacier and fluvially reworked after the deglaciation. Locality: Fig. 2/2; 5.6 km down the Indus valley, i.e. from the viewpoint in Photo 182 directly to the S; Fig. 2/1 half-left above No. 91.

the Karakorum-S-slope (Fig. 2), had a width of somewhat over 11 km and was c. 1700 m-thick. It has left behind an - according to its size, harmonic – trough valley cross-profile (Fig. 58), which, since the deglaciation, has been filled to a box-shaped glacial trough valley with a gravel bottom (U; Fig. 2/1 above No. 92). The slope form of the 1965 m-hill (No. 92), which in contrast is stretched and precipitous, becomes understandable by the fluvial undercutting of the Indus (□).

5.1. Summary of the maximum glaciation (LGM) of the Indus valley in the course of its gorge between the Skardu Basin and the Gilgit river (Fig. 2/1; 2/2; Fig. 52-58)

This 135 km-long, gorge-like narrow and, as for the incline of the talweg, steep valley course, has been continuously flowed through by an ice stream, which was 1000 to 1700 m-thick (in the confluence with the Gilgit valley). This is confirmed by ground moraine covers on the slopes and glacigenic flank abrasions up to a height of at least 3200 m at the head of the gorge (= exit of the Skardu Basin) and somewhat below 3100 m asl at its end.

The upper four valley cross-profiles (Fig. 52–56) show a V-shaped gorge character, in parts widened to a 'trough-shaped gorge' or a 'gorge-shaped trough' by flank polishing. At a steep valley incline, glacigenically V-shaped valleys of this type are not unusual, but in the Karakorum they are rather rare, because the valley incline is lacking. However, they are the rule for all prehistorically glaciated Himalaya cross-valleys (Kuhle 1982a: 59/60, 1983a: 154/155) and are to be deduced from the ground polishing, which, against that of the flank polishing, reduced by the tractive forces of the glacier, has increased. This idea goes back to Visser (1938 vol.2: 139), who in this connection has pointed out that the trough profiles in the glaciated Karakorum always occur in the areas, where the insignificant valley incline leads to an increase of the pressure forces in the glacier, and thus also to an increased polishing of the flanks.

A further factor which has increased the ground polish was the subglacial meltwater erosion beneath the Indus glacier. Above the gorge stretch, i.e. throughout the large Skardu Basin over at least 60 km the glacier surface has already lain below the LGM snow-line at 3800–3600m asl. The meltwater beneath the 1000–1700 m-thick ice (here already 1500–2300 m below the ELA) as hydrostatically confined water, had a very high velocity of flow and thus erosional force. This led to the subglacial exaration of a narrow, ravine-like cross-profile below the trough cross-profile. The overlying glacier worked over the ground by polishing it afterwards, so that the ravine-profile has been syngenetically widened to a V-shape (cf. Photo 173 ▲ black = trough bottom, below ▲ white, middle and on the right = V-shaped profile; 175 U = trough and below ▼ = V-shape; Fig. 55 and 56) (as for details see Kuhle 1982a: 60; 1983a: 158–160). Based on this interpretation, it is difficult to understand the development of a ground moraine pedestal or pedestal moraine, verifiable in the Indus gorge stretch (Photo 75 ■ large), as being syngenetical. Where bedrock has been eroded, the synchronous accumulation of ground moraine is impossible. Synchronism is at most understandable by means of spatial differentiation, i.e. in such a way, that ground moraine has been accumulated to a rather important thickness in the flow shadow behind rock ribs. But perhaps a chronological separation applies: in this case, ground moraine pedestals have been built-up during the deglaciation, in the time of the last Late Glacial Stages III to IV (cf.Tab 1).

Due to the Rakaposhi- and Haramosh chain, which immediately to the N reach up to over 6000 and even 7000 m, the 135 km-long section of the Indus gorge has received an enormous influx of ice from the N through its orographic right side valleys, from the Ice Age Tormik glacier, Stak (or Sak) Nala glacier with the Kohtia Lungma tributary stream from the Haramosh-E-flank (7397 or 7409 m; Fig. 2/1 No. 53), Boroluma glacier, Ishakapal glacier, Puparash-Haramosh Lungma glacier (cf. Meiners 2001) and Darchan Gah glacier, to quote only the most important of them. But also from the left side valleys the glacier termini have still reached the 3200–3100 m-high level of the Indus parent glacier, because the orographic snow-line had run there in a N-exposition at only 3600 m during the LGM.

In this shady gorge stretch the very productive side glaciers on the orographic right might still have fed the Indus parent glacier, when in the Late Glacial sections of the upper Indus valley had already been cleared of ice.

The Ice Age glacier influx from the Gilgit valley connected to the orographic right, has already been investigated (Kuhle 1988a-c,1989, 1991, 1993) (cf. Fig. 2) as well as the orographic left glacier inflow into the lower Indus parent glacier by the Astor glacier and the rest of the side glaciers from the Nanga Parbat massif (Kuhle 1996, 1997). The Photos 111 and 112 of the last mentioned study (ibid.: 225, 226) show the area of the Nanga Parbat (Rakhiot valley and current glaciers) with the High Glacial glacier level of the Indus parent glacier down- valley of the junction with the Gilgit valley, introduced in detail here.

6. Summary of the field- and laboratory data as to the maximum Ice Age (LGM) glaciation of the Central- and South Karakorum in the Braldu-, Basna-, Shigar- and Indus valley system as well as on the Deosai plateau between the Skardu Basin and the Astor valley concerning the heights of their glacier levels and their ice thicknesses (Fig. 2/1 and 2/2)

6.1. The glaciogemorphologically reconstructed prehistoric glacier extension in the light of the state of knowledge so far

As for the results of the field investigations mainly carried out in 1997 and 2000 introduced here in detail, it is fundamentally new that a continuous ice stream network has existed for the entire research area (Fig. 2/1) and that no LGM-glacier terminus could be reconstructed. Only during the Late Glacial did this ice stream network disperse into separate glaciers. The High Glacial ice stream network merged into the Indus glacier and thus had only one lowest glacier terminus, namely that of the Indus parent glacier. In the course of his previous fieldwork done 115 km down-valley from Nanga Parbat (113 km down from the junction of the Rakhiot valley) between the Sazin settlement and 20 km down the Indus (down from 35°32′ N/73°18′ E), the author has reconstructed the lowest LGM-tongue end of the Indus glacier at c. 850–800 m asl (Kuhle 1988a-c, 1989, 1991, 1993, 1996, 1997, 1998a) (cf. Fig. 2). Schroder (1989: 144), who has listened to the author's presentation in Leicester, March 1988, approximately follows this reconstruction, suggesting a glacier terminus 25–100 km away from Nanga Parbat, down the Indus. Dainelli (1922) has not visited this section of the Indus valley down-valley of the Skardu Basin, but has assumed an 18 km-long glaciation of the Indus valley for the LGM, deriving from the Astor valley and the valleys of the Nanga Parbat massif exposed to the WNW (ibid. Tav. CLXXVII). In contrast to the reconstruction of the author, based on field investigations, Dainelli's ice has only occupied the foot of the Nanga Parbat and has come to an end c. 95 km up-valley of the Indus glacier terminus indicated by the author. A further difference is, that in Dainelli's map no Indus valley glacier is marked between this influx of the Astor valley glacier and the Skardu Basin, and hence over an Indus stretch of 150 km. According to the assumption of Dainelli (see Chapt. 5) an inflow of a valley glacier has not taken place from the Gilgit- and lower Hunza valley either. The author, however, has evidenced a continuous Hunza-Gilgit ice stream for the LGM, which he has made probable with the help of fresh glacier striae (Kuhle 1988a,b,c, 1989, 1991,1993) (see Fig. 2 and 2/2). The Astor glacier, which Dainelli (ibid.) considered as having reached the Indus, has been confirmed by the author's field investigations (1987) (Kuhle 1988b,1996,1997) (see Fig. 2/2). According to their fieldwork (1934) Finsterwalder (1938: 174) and Troll (1938a,b) contradict this opinion. They pointed out that the Astor glacier has extended only somewhat farther than up to the Astor settlement, and that the Astor valley, continuing in a downstream direction over 25-30 km, remained non-glaciated. In connection with the Das Khilin Gah glacier (see Fig. 2/2), which has joined the Astor glacier, the author has newly referred in detail to his reconstruction of an Astor glacier being c. 1500-thick in the LGM (see Chapt.4.5.) (Kuhle 1996,1997). Haserodt's interpretation (1989) has already been discussed (Kuhle 1996 and 1997). He describes the Indus valley below the Skardu Basin as not being flowed through by an Ice Age valley glacier, but as having only been reached by the tongue ends of side valley glaciers.

Dainelli (ibid.) has not provided field observations of the Indus valley from a few kilometres away from the Skardu Basin in a downstream direction, because he has not visited this area. However, for his Mindel- and Riß-glaciation (1st and 2nd glaciation) he has postulated an Indus glacier which he supposed to have filled the valley below the Skardu Basin. A certain inductively obtained anticipation of the author's results may be seen in Dainelli's observation (1922) that the Skardu Basin, which he had investigated in detail, had been filled and polished out by the glacier ice during these two older (earlier than LGM) glaciations. This assumption, however, is not based on a 1400 m-thick ice-filling of the basin, as has been reconstructed by the author (Fig. 45, 46; Fig. 2/2), but on a valley bottom, which during his 1st Quaternary glaciation was still higher and – until the end of his 2nd glaciation (Riß- or pre- LGM) – has been polished down to the level of the present-day rock bottom below the lake sediments and gravels. He describes and maps (ibid. Tav.5) moraine-like deposits on the two 'riegels' Blukro and Karpochi as well as on the basin flanks c. 500–1000 m above the valley bottom and classifies them as belonging to the 2nd glaciation, without inferring from them the corresponding glacier thickness at that time. This may be explained by the fact that Dainelli considers the height of the trough bottom as having been much higher at the start of the 2nd glaciation than at its end, whilst the highest moraine findings could be dated as being from its start (from the 2nd glaciation). A further observation of Dainelli (1922 Tav.LIV.), which is important from the perspective of the findings introduced by the author, are the deposits mapped as 'Morene della 2. espansione' at greater heights on both slopes of the Shigar valley. These are sediments confirmed by the author as moraines, i.e. ground moraines (Fig. 5 and 44; Fig. 2/2), but which he has classified as belonging to the LGM glacier filling. Summing up, the following has to be established: whilst the author recognizes the preserved glacier traces as indicators of the LGM, which have completely reshaped an older glacier history (during a previous stage of uplift of the Karakorum at a different sea-level), Dainelli goes so far back into the past that – due to the then inevitably very important influence of the tectonic history of uplift (Schneider 1956, 1957), which for the most part is unknown – his findings allow neither a reconstruction of the ice levels or glacier thicknesses nor a paleoclimatic approach by calculations as to the snow-line and glacier surfaces. Compared with this, Dainelli's glacier reconstructions for his 3rd glaciation (1922 Tav.CLXXVII) are realistic. Chronologically it belongs to the LGM, but, according to the author's observations in the field, it corresponds to the glacier extension during the High Glacial (Stage 0 = LGM) only in some areas – so e.g. on the Deosai plateau –, whilst for the most part it applies to that during the Late Glacial (in part to that of the Stages I, II, III or IV; cf.Tab. 1). After having discussed the observations of Dainelli, which were new at that time, Lydekker (1883) and Oestreich (1906), being the definitely classical authors of the prehistoric glaciation of the Shigar valley as far as down to the Skardu Basin, are to be acknowledged again (see Chap. 3.15) in this summary. Both authors have also introduced personal field observations on the Ice Age glaciation of the Braldu- and Basna valley, merging into the Shigar valley (ibid.). Norin (1925), however, who has worked there later and confirms the results of Lydekker (1883), Oestreich (1906) and Dainelli (1922) only in part, has supposed the Braldu- and Basna valley to have been prehistorically glaciated, but neither the Shigar valley nor the Skardu Basin. Nevertheless, according to his investigations small valley glaciers have reached the basin margin from much lower catchment areas from the S and SW. Oestreich (1906) was obviously the first researcher who has recognized the complete glaciation of the Deosai plateau (see Chap. 4.2.).

In part very critical comments on the field data and investigation results of the classic researchers quoted here – with which the author does not agree –, but also a useful compilation of the older Quaternary-geological expedition literature as to the prehistoric glaciation of the Karakorum we owe to von Wissmann (1959). As for the Indus valley section which continues 90 km SSE of the Skardu Basin from the Kargil Basin upward, as well as for the connected Zanskar Himalaya (Fig. 2) is to be referred to the author's further field observations on the maximum Ice Age glaciation (Kuhle 1998a: 88/89; Kuhle & Kuhle 1997); also for the area 50 km NW of the Biafo glacier (Fig. 2) in the Shimshal region (Kuhle 1996a, 1998a) and on the Karakorum N-slope (Fig. 2) from the K2 up to the Tarim Basin (Kuhle 1988b,d,e,f; 1994b; 1996b; 1998g.

Probably the Late Glacial end moraines (Stage I-IV, Tab 1) in the Indus valley section with its large side valleys Braldu-, Basna- and Shigar valley, connected at Skardu, have in part been glaciofluvially and -limnically buried in the Skardu Basin (Fig. 2/2 in the area of Profile 45 and 46). Downstream of the Skardu Basin, in the narrow Indus gorge up to the junction of the Gilgit river (Stage IV and III) (Fig. 2/2 between Profile 52 and 57), but also further downvalley in the Nanga Parbat area (Stage II and I) (Fig. 2/2 Profile 58 and below) they have been removed by the Indus river bearing plenty of water, and finally their remnants have been covered by the river gravels. Late Glacial end moraines (Stage III and IV) from the outlet glaciers of the Deosai plateau glacier have survived – at least in remnants – in the Satpare Lungma and Das Khilin Gah (Fig. 2/2 S of Profile 47; at Profile 51) (see Chapt.4.1 and 4.5). As for the rest, the Late Glacial glacier levels of the Stages I to IV (Tab 1) and their ice thicknesses have been reconstructed with the help of sporadically preserved lateral moraine ledges with and without erratics.

With regard to the differentiation of the Late Glacial ice margins from the Holocene, neoglacial to historic ice margins of the Stages V to XII (cf Tab.2) in the Karakorum is to be referred here to the detailed observations of Meiners (1996, 1997, 1998, 2001). The postglacial reworking and destruction, i.e. reshaping of ground- and lateral moraines, as well as the development of kames and related glacigenic lateral forms of those High (LGM) to Late Glacial (I–IV) Karakorum glaciers since the deglaciation have been investigated and typified by Iturrizaga (1997, 1998, 1999, 2001). The comprehensive work of Kalvoda (1992) on the geomorphology of the Himalaya and Karakorum is completely indispensable with regard to the postglacial high mountain geomorphology in our regional connection as well as to the general understanding. As for the hardrock geology the standard work of Searle (1991)is to be referred to.

6.2. The glaciogeomorphologically reconstructed heights of the ice levels and glacier thicknesses of the prehistoric ice stream network

In addition to the reconstruction of the maximum extension of Ice Age glacier cover, also studied by the researchers mentioned above, the author's field observations over several months and their interpretation were focused on the glaciogemorphological evidence of the corresponding heights of the ice level and glacier thicknesses at that time. Only in this way is the extension of the glacier surface of the prehistoric Karakorum ice stream network recognizable. Merely the height of the glacier surface provides indications as to the extension of the feeding area and the actual glacier form as well as the dome-shaped geometry of the surface of this ice stream network. Its dammed-back, accumulative elevation and the autonomous-increase of its catchment areas above the snow-line (ELA) become only understandable in this way. Additionally, the ice thicknesses point to the flow dynamics of the glacier and the glacio-isostatic load of the crust. From the extension of the Karakorum ice stream network (Fig. 2) on, the reconstructed glacier face is not only an indicator of the Ice Age (LGM) thermal and hygrical climate change in the subtropics, but, due to its surface albedo, also an effective cooling-amplifier (Kuhle 1985b, 1986e, 1987f, 1988b, 1998a, 2002 etc.).

Distributed over the reserach area (Fig. 2/2), the 25 representative glaciogeomorphologic valley- and glacier cross-profiles (Fig. 3–5, 37–58) show prehistoric glacier levels from 6200 (Fig. 37) down to 3100 m asl (Fig. 58): the first is situated in the currently still glaciated regions of the highest summits in the Central Karakorum (Muztagh-Karakorum) and the latter in the area of the Indus outlet glacier below the inflow of the Ice Age Gilgit glacier (Fig. 2/2). In the area of the present-day upper Baltoro glacier between K2, Broad Peak, Gasherbrum IV, Baltoro Kangri and Chogolisa the ice level has even reached 6300 m asl (extreme value: 6400 m) (Chap. 2–2.5). So too in the area of the current Biafo glacier, the LGM-glacier surface has reached similar sea levels (Chap. 3.3). Correspondingly, in the areas of the present-day Chogolungma-, Biafo- and Baltoro glacier three dome-like ice culminations have existed in the large-scale cross-profile of the LGM ice stream network (Fig. 2/2). These very flat ice cupolas cannot just been explained by the reconstructed glacier cover, height of the ice level and glacier thickness in the research area discussed here, but only by the entire Karakorum ice stream network, extending over c. 125,000 km^2, depicted in Figure 2, and the mountain relief lying beneath. The ice-cupolas of these domes have communicated with each other over the transfluence passes (Fig. 2/1), that means they have formed an approximately continuous, i.e. unbroken, level of the ice surface without today's breaks of the incline. This adjustment of the level has taken place across the transfluence passes. The ice overlay of the Deosai plateau has made up a fourth dome, from which outlet glaciers also flowed down the nearly actiniformly-arranged valleys (Chap. 4.4). All these ice levels with their large-scale surface inclines have met at the lowest point of their joint ice discharge exactly at the place where, N of Nanga Parbat, the Astor glacier has merged into the Indus glacier (Fig. 2/1 below No. 92; Fig. 2/2 below No. (Fig).58). Here, too, the talwegs join, so that the LGM-ice stream network has flowed down subordinated to the relief, i.e. following the valley incline.

In the middle Braldo valley (or Braldu-,Biaho-,Blaldo Lungpa) an ice thickness of about 2600–2900 m has been reached (Fig. 3; Fig. 2/2). In the Baltoro glacier valley the ice thickness amounted to c. 2400-2900 m (Chap. 2.11.), in the Biafo glacier valley to c. 2450–2950 m (Chap. 3.3), in the Chogolungma glacier valley and the continuing Basna valley to c. 1800–2900 m (Chap. 3.14; Fig. 37–42; Fig. 2/2). In the valley exits the Braldu glacier was still 2500 (Fig. 4) and the Basna glacier c. 2600 m-thick (Fig. 43; Fig. 2/2). All these most important valley glacier thicknesses of the Ice Age Karakorum lay above the snow-line (cf. Fig. 3–44 with the climatic snow-line altitudes in Fig. 2/2). They have still exceeded the maximum LGM-glacier thicknesses of the Alpine ice stream network reaching c. 1800–2000 m by 20–30%, and may have had glacioisostatic effects up to a depression about 50–100 m. In the very wide Shigar valley the ice thickness has been reduced from c. 2400 m (Fig. 5) via 2150 m (Fig. 44) to c. 1300 m at the valley exit (Chap. 3.15.1). This corresponds approximately to the then ice thickness in the connected Skardu Basin; there it has reached c. 1500 m (Fig. 45 and 46; Fig. 2/2) and decreased to c. 1000 m towards the NW, i.e. the exit of the basin (Chap. 3.16.1). In the Indus gorge below the Skardu Basin the level of the glacier surface has only minimally dropped by 100 to 150 m, but the ice thickness has again increased up to 1700 m as far as the Gilgit glacier (Fig. 52-57; Fig. 2/2; Chap. 5.1). The reason for this is the deeply dissected and steep course of the gorge, showing numerous narrow bends. In this gorge the ice was nearly blocked, i.e due to the resistance by friction it flowed down very slowly. At the same time the gorge glacier has been dammed-up by the c. 1800 m-thick Gilgit glacier as far as its suface level at c. 3100 m asl. The Indus- and the Gilgit glacier have both formed a parent glacier with a width up to 11 km (Fig. 58), which came in contact with the Astor glacier and the Nanga Parbat ice stream network and mediated to an ice margin at only just 800–850 m (Chap. 6.1).

The ice level of the Deosai plateau glacier which on average is 24 km in size, lay approximately between 4700 and 4900 m asl, so that ice thicknesses of c. 200 m above the hill cupolas, and at most c. 900 m above the high valley bottoms, could be extrapolated as being probable (Fig. 48–50; Fig. 2/2; Chap. 4.4). The Deosai outlet glaciers investigated, the Satpare Lungma (Fig. 47) and Das Khilin Gah (Fig. 51; Fig. 2/2), have produced glacier connections to the Skardu Basin, i.e. Astor valley, and thus to the ice stream network of the Karakorum S-slope, and at the same time to the Indus ice stream network. In these terms, the Deosai glacier can be indicated as being one part of this entirety.

The altitudes of the climatic glacier snow-lines (ELA) in the research area, summarized in Figure 2/2, in connection with the 25 cross-profiles of the valley glaciers, lead to the deduction that the communicating surface of the most productive glacier branches of the ice stream network has remained below this snow-line from the lower Shigar valley on as far as the Skardu Basin between Profile Fig. 44 and Profile Fig. 45. Here, the ELA has run between about 3700 and 3800 m asl during the LGM. From this it follows that in this area the surface moraine cover sets in and, due to the growing ablation process down the Indus, may have increased rapidly. A condition for this is the feeding of the glaciers, also participating during the glacial periods, from the steep faces of the highest mountains as e.g. the K2, Broad

Peak, Ogre, Spantik, Laila etc., which at that time still towered 1000–2000 m above the highest glacier surfaces. Even more debris has been produced by the participation of the numerous ground- and lateral moraines, which in the narrow valley network were very massive, and which in the confluences have first met to form internal- and then medial moraines. In the middle to lower Skardu Basin (Fig. 2/2 from Profile (Fig.) 46 on), this surface moraine developed from avalanche-rock-debris merging with medial moraines of the Central Karakorum (Muztagh), might have completely covered the Indus parent glacier. Naturally, the participation of the covering glaciation of the Deosai plateau in this debris supply was only subordinated. Despite this substantial, albedo-decreasing production of surface moraine, at least 80% of the surface of the ice stream network were situated outside, i.e. at the same time above the surface moraine areas, so that they have reflected c. 75-90% of the global radiation on snow, firn and ice (Kuhle & Jacobsen 1988). The result of this was a cooling and autonomous amplification as to the build-up and preservation of the Karakorum ice stream network (Kuhle 1988e).

In the course of several glacial periods, the pre-glacial valley relief with its high plateau remnants (Deosai plateau) lying in between, the development of which had started in the pre-Pleistocene and which was created by backward river erosion, has been glacigenically polished out, i.e. abraded and – below the ELA – at the same time subglacially dissected (Chapt. 5.1). During the interglacial periods the fluvial relief has been rejuvenated. At present the destruction of the glacial relief is again taking place.

6.3. Postglacial rock avalanches as a confirmation of the LGM-glaciation

The valuable reconstruction of rock avalanches in the Karakorum, mainly and most recently carried out by Hewitt (1998 and 1999), is rightly an objection to the glaciation concept of Dainelli (1922) and the epigonal literature, which postglacial rock avalanches has misinterpreted as LGM- and Late Glacial end- moraines.

Hewitt (1999) introduces three examples of misinterpretation (see above) from the Basin of Skardu and the Shigar valley, i.e. our investigation area. As has already been described, the author completely agrees as to these isolated cases. At the same time, however, he does not recognize a contradiction to the glaciation history of the Karakorum, discussed here, but, in contrast, a confirmation. According to the author's opinion, which contradicts that of Dainelli, especially in the Central Karakorum, no LGM- (3rd espansione) end moraines are verifiable, because the ice stream network had only one joint ice margin position at c. 850–800 m asl in the lower Indus valley (Chapt. 6.1). In addition the Late Glacial end moraines have not been preserved, but have been completely removed by the abundant and very erosion-effective meltwater in the forefield of the melting-back ice stream network, especially in the narrow valley receptacles (cf. Kuhle 1994b: 184–186 Fig. 47 and 48). Where this cannot be applied, as e.g. in the 10 km-wide Skardu Basin, they have been buried by damming-up of sediments caused by rock avalanches. Owing to this, the author considers the concept of the search for end moraines in the Karakorum – pursued by many researchers – as methodologically wrong. Thus he has concentrated his work on the sedimentological and geomorphological analysis of the valley flanks and the search for ground- and lateral moraines there, as well as on the reconstruction of flank polishes, for it is mainly the reconstruction of the ice levels which enables a prehistoric glaciation in the high mountains to be estimated. The analysis of end-moraines, however, is naturally due mainly to the mountain forelands and lowlands (Kuhle 1991b). The glaciogeomorphological analysis of valley cross-profiles and valley flanks introduced here, which has led to the reconstruction of the c. 125,000 km^2 extended and continuous Indus-Karakorum ice stream network (Fig. 2), also confirms through the realization of glacigenically over-steepened trough valley cross-profiles the destabilisation of the valley flanks (cf. Kuhle, Meiners & Iturrizaga 1998). Accordingly, rock avalanches as have been described by Hewitt (1998 Fig. 1) do not contradict, but rather confirm the very large, 1000 to 2900 m-thick glaciation of the type of ice stream network reconstructed by the author as the actual condition for those prolific postglacial crumblings.

Anyone who through glaciogeomorphological data collection has worked on the still extended Karakorum glaciers as e.g. the Baltoro-, Biafo-, Chogolungma glacier, will know about the youngest, prehistoric surfaces of the trough walls, which have been – and still are – heavily roughened by current crumblings every day – and even of those to which valley glacier ice is still today attached on the lower slopes. Because of this relief-specific cause of gravitational mass movement, occurring not only in the bedrock, but naturally rather in the loose rock on the steep slopes, the attempts of so-called absolute datings on the background of the previous physical- technical possibilities in the high mountains are methodically wrong. The continuing primary, secondary, tertiary, quaternary etc. replacement of the loose material as well as of the datable boulder surfaces, their uncovering by flushing, and also their burying which takes place again and again and, in the Karakorum, even increases to an extreme extent because of the insignificant plant cover, disqualifies this attempt at establishing ages in this steep relief nearly to the category of a pseudo-scientific method. Dating attempts such as these, as well as the investigation of end moraines (see above), are rather due to the lowland than to the Karakorum Himalaya. The cold aridity with its poverty of plants is unfavourable even for a C14-analysis.

However, the desire for an absolute dating of those prehistoric glaciations of the ice stream network can be fulfilled geomorphologically. The author has classified it as belonging to the LGM, because the interglacial destruction of the trough forms – as has been evidenced geomorphodynamically by the current flank roughening and as has been confirmed by the numerous postglacial crumblings observed by Hewitt (ibid.) – takes place so fast that the preservation of the forms of a pre-LGM glaciation can be ruled out. Additionally, the history of the up-lift of the Karakorum

(Schneider 1956, 1957 and others) points to the time mark LGM as to the most extended glaciation by the fact, that the height of the high mountains was noticeable lower during the pre-LGM. Thus, for the development of this extended ice stream network a snow-line depression of several hundred metres more than in the LGM would have been necessary. However, this is a climatically very improbable and therefore inadmissible additional consideration.

According to the hierarchy of the scientific question, it has been primarily worked on the geomorphological reconstruction of the prehistoric glacier cover. Inevitably the exact age of the phenomenon which has to be evidenced is secondary. A wrong dating is not able to relativize the existence of a prehistoric ice stream network.

7. Lowest prehistoric ice margin positions of several valley glacier systems in the Hindukush, Himalaya and in East-Tibet near the Minya Konka-massif (Fig. 1)

7.1. The lowest preserved, probably LGM-ice margin position in Chitral, Hindukush (Fig. 1 No. 22)

Haserodt (1989) has suggested that the Ice Age Chitral glacier terminated c. 8 km up-valley of the Buni settlement, so that the Mastuj- or Chitral main valley should have been free of ice. Kamp (1999), in contrast, assumes that the prehistoric Chitral glacier has reached at most 100 km further down-valley, as far as the Drohse settlement at 1300 m asl. According to the author's reconstruction the LGM-Chitral glacier has even flowed down 15 km further (Photo 183; Fig. 2 on the left above No. 97), i.e. as far as the valley chamber of Mirkhani at 1050–1100 m asl (35°28'40″ N/71°46'30″ E; Fig. 2 No. 97). This ice stream was not only thicker than 500 m, as has been supposed by Kamp (ibid.: 189), but three-times thicker. Whilst Kamp (ibid.) does not assume a transfluence of the Tirich Mir glacier over the Zani pass, the author's findings of granite erratics above the depression of the pass argue in favour of this transfluence (Photo 184; Fig. 2 No. 96), thus providing evidence of a more important ice thickness.

7.2. The Ice glacier filling of the Lahaul valley (SE-Zanskar Himal) and two ice margin positions (probably LGM) in the SE Pir Panjal Range SW-slope, NE of Dharamsala (Fig. 1 No. 17)

The 3980 m-high Rothang (Jot) pass (32°21'45″ N/ 77°14'50″ E) was a transfluence pass over which the Lahaul glacier has flowed into the Kullu valley adjacent to the S. This is evidenced by glacigenic abrasion forms (Photo 185 ▲ large) as well as its ground moraine cover (■). Due to this transfluence the level of the Lahaul glacier has run at a height just about 4300–4400 m (▬ ▬). Its ice thickness has amounted here to c. 1100–1200 m (see also Kuhle 1998a).

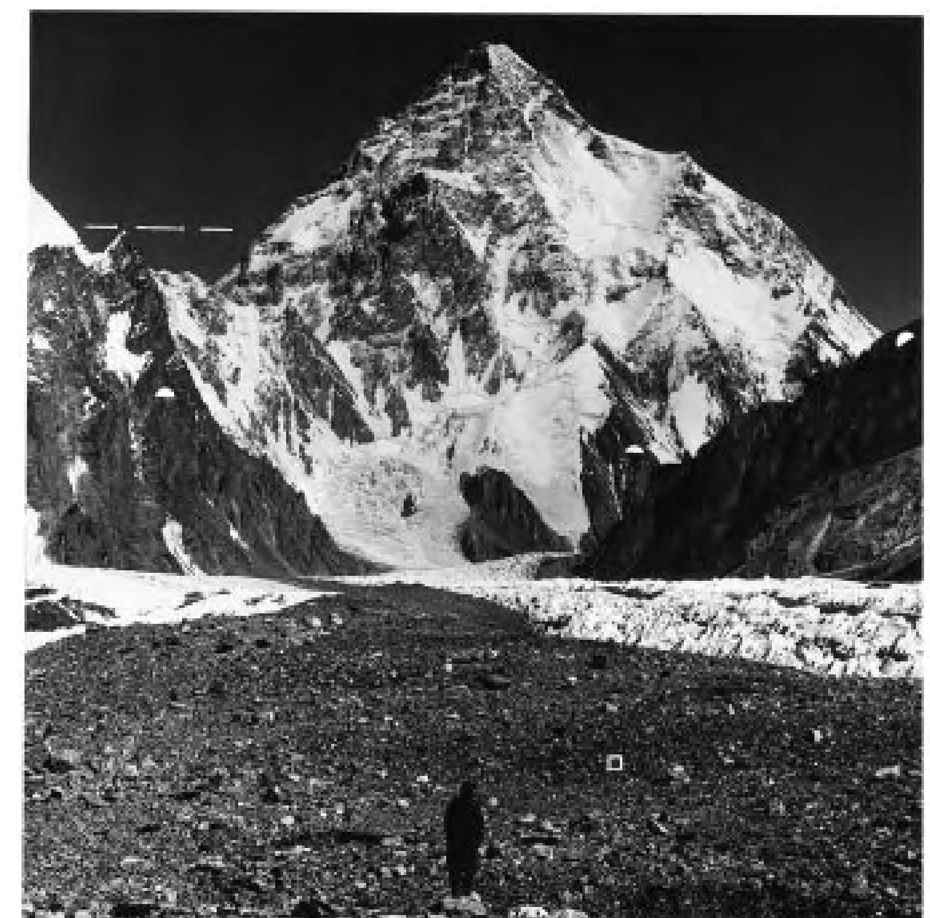
The two parallel valleys which from the 4971 m-high SE Pir Panjal massif NE of Dharamsala peter out into the Himalaya foreland to the SSW, have been glaciated as far down as the foreland (Photo 186, 187 ▬ ▬). Their lowest ice margin positions, recognizable by well-preserved lateral- and end moraines, can be considered as belonging to the LGM. They are situated below the exits of the Tori valley at 1250 m (32°12'30″ N/76°22'30″ E) and the Triund valley at 1200 m asl (32°12'50″ N/76°21'10″ E).

7.3. Supplements to the lowest LGM-ice margin positions of the Annapurna Himalaya (cf. Kuhle 1982a) (Fig. 1 No. 2)

In his map on the LGM glacier cover, sketched out during the expeditions 1976 and 1977 (ibid. Bd.II Abb.8), the author has put the end of the Ice Age Modi Khola glacier above the confluence with the synchronous Chomrong Khola glacier. However, a question mark has been added to the glacier terminal (ibid. Abb.8?, on the left below D). In the meantime further interpretations of the data and additional findings in 1995 and 1998 yielded information that its tongue received an influx from the Chomrong Khola- and Kyumnu Khola glacier during the LGM (Stage 0) and flowed down as far as c. 800 m asl up to the Dobila locality at the junction of the Jare Khola (28°13'50″ N/83°43′ E) (Photo 188 ᗜ; 189 ↓). The trough valley cross-profile of the lower Modi Khola (ᗜ) reaches up to there, as does an orographic right kame-terrace (Photo 188 ■). Down-valley the glacier mouth gravel floor terraces of Chuwa (Kusma) set in (ibid. Bd.I: 71/72, Bd.II Abb.7 and 105–108; Kuhle 1983a: 193–196). Among others, the following observations testify to this prehistoric glacier extension: ground moraines up to at least 2400 m asl at the spur between the Chomrong- and Modi Khola (■ black on the right) and on the orographic right side at the exit of the junction area of the Kyumnu Khola up to 2000 m (near Udi) (■ black on the left); prehistoric subglacial potholes on the orographic left between Landrung and Tolka at a height of 300–500 m above the current talweg (△) and an orographic right ground moraine cover (■ white) at the inflow of the Bhurungdi Khola at c. 1450 m asl. Accordingly, the Modi Khola glacier near Birethanti (cf. Kuhle 1982a Bd.II Abb.8,d) still had a thickness of approximately 400 m (■ white).

The LGM-Seti Khola glacier has also been mapped as being uncertain (ibid.: Abb.8? above K). With the help of two up to 3 m-long erratic gneiss boulders (Photo 190 ■) on a rock head of schist bedrock W of Ghachok at 1500-1540 m asl, 350-390 m above the current talweg (Photo 190 ↓), its LGM-glacier tongue end (position of the glacier mouth) can be extrapolated at c. 1000 m asl (⇩). At an ice thickness of c. 300-400 m and a glacier width of 2.5–3 km at Ghachok, the High Glacial Seti Khola glacier has in all probability reached the junction of the Seti- and Yamdi Khola situated 6 km further down-valley (28°16'20″ N/83°57'30″ E) (⇩). The erratic boulders derive from the Himalaya main crest (Annapurna III–IV). According to the arrangement of their positions they can neither be explained by mudflows (or related phenomena) (Fort 1986: 118) nor by a 'Late Glacial glacial extent' (ibid.: 108). Fort, however, does not come to a decision as to these two possible interpretations of the accumulations in the Seti Khola cross-profile of Ghachok. Both of them can be ruled out by these erratics.

↑ *Photo 1*. 360°-panorama of the Upper Baltoro glacier with its northern tributary stream, the Godwin Austen glacier, taken from 4760 m asl (35°47′ N/76°31′ 20″ E) (Fig.2 No.1,3,6,9,10). At the end of this tributary stream, right in the north, the 8617 m-high K2 (Tschogori), second-highest mountain on earth (No.1), is situated. In front of it the Savoia glacier flows from the left (orographic right) side into the Godwin Austen glacier. The Chogolisa (No.6, 7665 m) is in the S with Concordia in front at an altitude of 4500-4600 m. Here, the Godwin Austen- and Upper Baltoro glacier are joining and the Baltoro glacier flows down to the W, i.e. to the right. To the orographic left is the WSW-flank of the 8047 m-high Broad Peak (No.3, Palchan Kangri). To the orographic right the Marble Peak (No.9, 6238 m) and Mitre Peak (No.10, 6010 m) can be seen. Right of the Marble Peak is a spur peak (— —) polished glacigenically up to its sharpened summit (◤ on the right below No.9); it has got a glacially triangular-shaped slope (Fig.2 on the right above No.9). (□) mark the present-day strings of surface moraine. They are interrupted by detritus-free bands of glacier ice. The subtropically intense insolation creates white ice pyramids on them. (▼) signify present-day debris cones, deposited against the margins of the valley glaciers. They consist of older ground moraine material, dislocated from higher valley flank positions, as well as from the crumbling rock flanks of prehistoric glacigenic polish areas. In places where the denudation and the development of rills has only slightly advanced, glacigenic flank polishings are still preserved (◤). (— —) marks the Ice Age (LGM) glacier level at c. 6000-6250 m asl. It is evidenced by flank polishings breaking off in an upward direction (e.g. Fig.2 on the right below No.1) as well as by polish cavettos. This maximum glacier level ran 1200-1400 m above the present one. Photo M.Kuhle, 15.9.1997.



← *Photo 2.* Taken at c. 4800 m from the surface moraine (□) of the Godwin Austen glacier looking towards the N to the S-face of the K2 (8617 m) with the Fillipi glacier. The surface moraine is made up from polymict boulders of sedimentary rocks as e.g. crystalline schist, quartzite and gneiss. Only some of them are rounded at the edges. The rounded rock heads (◤ on the right) at the foot of the Broad Peak have been polished by the Godwin Austen glacier, the ice of which attained a much higher altitude during the High- (LGM) to Late Glacial (Fig.2 on the right below No.1). Flank polishings are also preserved on the opposite valley flank (◓ left). The glacier has polished back the confluence spur in the inset between Godwin Austen- (right) and the Savoia glacier (left) (◓ on the left). This glacially triangular-shaped slope and 'truncated spur' (Fig.2 left of No.1) reaches up to the 6394 m-high satellite of the Nera Peak sharpened by the ice stream network of the glacier. Up to this altitude, i.e. c. 1400 m above the modern level, the LGM-glacier level can be evidenced (– –). The upper part of the glacially triangular-shaped slope has already been roughened by meltwater, ravines and rock falls (above ◓ on the left) since the deglaciation. Such denudation and roughening takes place within only a few centuries or millennia. This is documented by the again roughened rock island in the Fillipi glacier (left of ◓ centre). Photo M.Kuhle, 15.9.1997.

→ *Photo 4.* At 4650 m asl, looking down the Godwin Austen glacier which over large parts is covered with surface moraine (□), via Concordia (Baltoro glacier) situated c. 4 km away, facing S to the 7665 m-high Chogolisa (No.6). (◥ ◤) mark prehistoric flank polishings on metamorphic sedimentary bedrocks. The flattening of the profile line of the Chogolisa-N-spur provides evidence of the linking up of the two polish bands and flank polishings of the joining Vigne- (right) and Upper Baltoro glacier (left) at a far higher Ice Age glacier level. (– –) is the corresponding glacier polish limit, i.e. the maximum prehistoric ice level (LGM) at c. 6300 m asl. (■) signifies remnants of ground moraine, reshaped by avalanches and meltwater which regularly show forms of debris cones. See Fig.2 No.6. Photo M.Kuhle, 15.9.1997.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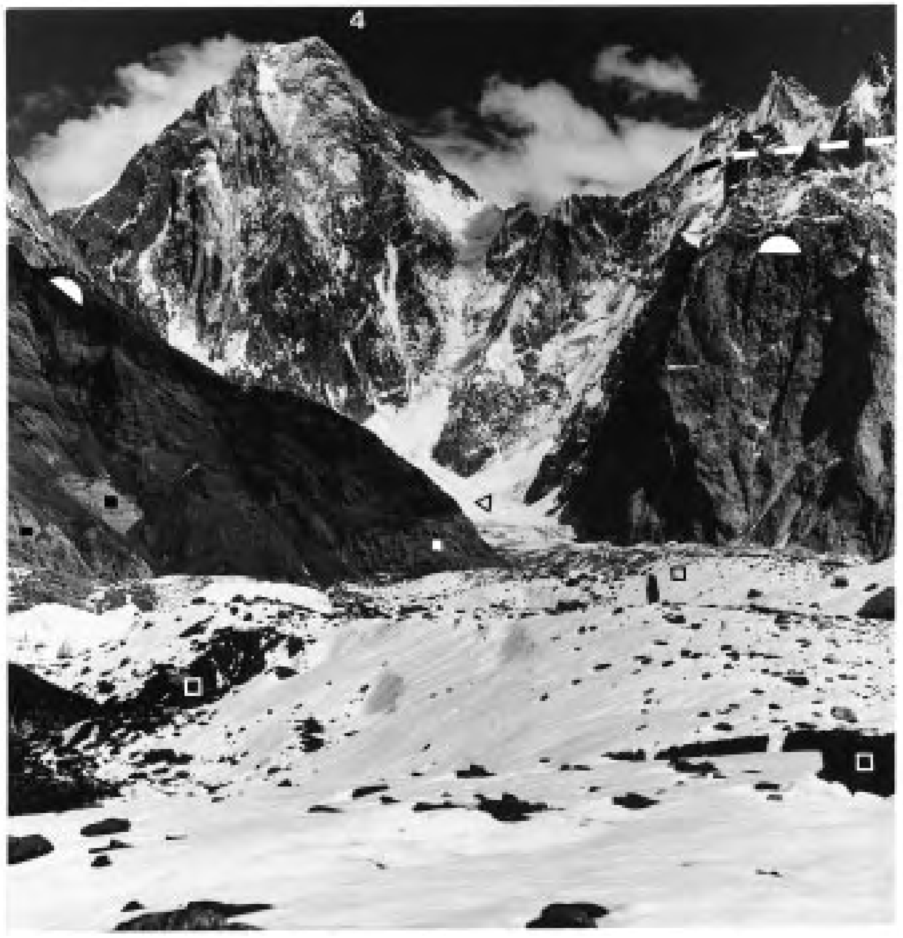
← *Photo 3.* 200°-panorama taken from the large central medial moraine of the Godwin Austen glacier (□) at 4700 m asl (35°46′ N/76°31′ 45″ E) facing S with the Chogolisa (No.6, 7665 m) and the glacier confluence Concordia in the middleground, the drainage of the Baltoro parent glacier to the E with the Paiju Peak (No.11, 6600 m) and the Godwin Austen glacier with the K2 (No.1, 8617 m) in the N. Furthermore the Mitre Peak (No.10, 6010 m), the Masherbrum (No.5, 7821 m) and the Marble Peak (No.9, 6238 m) are visible. (– –) marks the glacio-gemorphologically recorded position of the Ice Age (LGM) glacier surface. It ran between 5950 m asl at Concordia (– – middleground of the panorama-centre) and 6400 m asl in the area of the Upper Baltoro glacier (– – on the very left) and upper Godwin Austen glacier (– – right third). The valley flanks, which extend up to the prehistoric glacier level (– –), have been polished glacigenically (◤). Since the deglaciation, i.e. since the lowering of the glacier level up to its present-day position (□), these glacigenic smoothings and roundings are broken away and at the same time have been roughened. The debris of the breakages, also containing crumbled and dislocated ground moraine, have formed fresh debris cones and -slopes in new slope-gullies (▼). The dark sedimentary bedrocks are especially susceptible to this postglacial reshaping (in the areas of ◤). The massif-crystalline gneisses create steeper, very coarsely structured valley flanks (below – – between No.5 and 11 background). The Marble Peak has towered above the glacier surface by c. 150 m and due to the great stability of the marble has formed a nunatak with steep faces, i.e. a classic glacial horn. (See Fig.2 No.1,9,10,6). Photo M.Kuhle, 15.9.1997.

↑ *Photo 6.* At 4550 m, 1 km E from Concordia, i.e. from the centre of the glacier confluence, looking towards the NNE across the Baltoro- and lower Godwin Austen glaciers to the Broad Peak (No.3, 8047 m). (□) mark the Baltoro glacier covered by petrographically different surface moraine. The two components of the Broad Peak glacier flowing together in the junction area are visible (▽). (◓) shows a few of the glacigenic rock polishings which have only been preserved in parts. They mediate to the prehistoric glacier level (– –) evidenced by their occurrence. The undercutting of the rock faces through the present-day glacier margins of the Baltoro glacier (□) and of the Broad Peak glacier components (▽) seems to have a steepening effect and at the same time to intensify the breakages (Fig.2 No.4) that due to the absent contiguity of the ice take place anyway. Photo M.Kuhle, 16.9.1997.

← *Photo 5.* 300°-panorama taken from Concordia at 4550 m asl, i.e. from the centre of the confluence area of Godwin Austen- (on the very left) and Upper Baltoro glacier (panorama-centre) from the glacier surface moraine covered with freshly fallen snow (□) (35°33′ 30″ N/76°44′ 30″ E). The camp (for scale) is put up between large boulders of the surface moraine, 1.8-3 km away from the valley flanks. Correspondingly, the present-day ice thickness is to be interpolated with the help of the dip of the valley flanks to c. 1300-1500 m. The K2 (No.1) is situated in the N, the Broad Peak (No.3, 8047 m) in the NE, the Gasherbrum IV (No.4, 7980 m) approx. in the E, the Baltoro Kangri (Golden Throne, No.7, 7312 m) in the SE, the Mitre Peak (No.10, 6010 m) in the S and the Paiju Peak (No.11, 6600 m) approx. in the W. (◓) marks the Ice Age (LGM) to Late Glacial glacier flank polishings on glacially triangular-shaped slopes and back-polished mountain spurs between the exits of side valleys on metamorphic sedimentary bedrocks (phyllites). (– –) is the High Glacial (LGM) ice level related to these flank smoothings running between an altitude of 5950 m at the Mitre Peak (No.10) and 6400 m between K2 and Broad Peak (No.1 and 3). Thus, the ice must have attained the very substantial thickness of 2700-2900 m here at Concordia. The sharp glacial horns of the Mitre Peak and the adjacent summits to the right (right quarter of the panorama) suggest that they have towered above the LGM-ice surface just a little. See Fig.2 No.1,3,4,6,7,10. Photo M.Kuhle, 3.9.1997.



← *Photo 7.* Looking from the large Concordia glacier confluence area at 4550 m asl towards the S to the 7980 (or 7925)m-high Gasherbrum IV (No.4). Viewpoint on the surface moraine (□) of the Baltoro glacier, which joins the Gasherbrum glacier coming from the E (▽). The highest visible summit further to the right is Gasherbrum V (7321 m). It is also part of the feeding area of the Gasherbrum glacier, supplying it with avalanches. (– –) shows the minimum altitude of the Ice Age glacier level, (◓) are remnants of polish bands and flank polishings in the bedrock below, as well as postglacial crumblings verifiable according to their glacigenic roundings (Fig.2 No.4). At some other points of these rock slopes decametres-thick High- to Late Glacial remnants of ground moraine occur which were pressed into the glacier flow shadow of rock niches and rills and stripped off by the glacier bottom (■). Photo M.Kuhle, 3.9.1997.

→ *Photo 8.* View from the Upper Baltoro glacier (□ centre) at 4660 m asl (aneroid-measurement) facing NNW (No.9; Marble Peak 6238 m) via N across the Godwin Austen glacier tributary stream, the K2 (No.1) and the Broad Peak (No.3) with the junction of the Broad Peak glacier as far as NNE to the junction of the Gasherbrum glacier (□ right). The Ice Age (LGM) glacier level (– –) deduced and interpolated from the partly preserved glacier polishings (◓) and triangular-shaped mountain spurs on the intermediate valley ridges (◓ on the very left and second and fourth from the left) (Fig.2 No.9,3,4) runs approx. 1400-1500 m above this confluence centre of the Baltoro glacier. The reconstruction suggests the sharpening of the Marble Peak SW- crest to have been created by the undercutting lateral erosion of the High Glacial ice stream network (– – below No.9). (▼) shows debris cones consisting of material from crumblings on the surface and of High- to Late Glacial ground moraine in the core. Photo M.Kuhle, 16.9.1997.

↓ *Photo 9.* 360°-panorama at 4680 m asl (aneroid-measurement) from the Upper Baltoro glacier in the confluence area with the Vigne glacier (below No.12), 3.9 km SE of Concordia. A frozen supra-glacial meltwater lake is in the foreground. It has been melted into the glacier ice, rich in debris (□). The Baltoro Kangri (No.7, 7312 m) is situated in the SE, in the confluence area of the Abruzzi glacier (from the left) and the northern Chogolisa glacier. The Trinity Peak (No.12, 6614 m), which is part of the Masherbrum Range, is in the SSW. Between the upper Baltoro- and Vigne glacier lies the N-crest of the Chogolisa-massif glacigenically sharpened especially during the LGM (– – between 7 and 12). Between Mitre Peak (No.10) and Marble Peak (No.9) the Muztagh Tower (No.8, 7273 m) becomes visible in the NW. The 7360 m-high Skilbrum (No.13) is one of the highest mountains of the orographic right side valley system of the Godwin Austen glacier (-valley). It is situated beyond the large glacier confluence of Concordia in the NNW. Behind the white glacier string made up of glacier ice poor in moraines with pyramid forms (left half of the panorama and on the very right) short, steep side valley glaciers coming from the Gasherbrum group flow into the Upper Baltoro glacier (▽). (■) is a Late Glacial to Holocene orographic right lateral moraine remnant. (▼ on the left) signifies a Holocene debris cone the surface of which is made up from the debris of crumblings. It has a High- (LGM) to Late Glacial ground moraine core. The snow of avalanches from above causes small linear mudflows to flow down over the cone surface (between ■ and ▼). (▼ on the very right) shows ground moraine material reshaped by slope flushing. (◓) are glacigenic flank polishings which up-slope are coming to an end. Thereby they document the Ice Age glacier surface (– –). These polishings have formed the mountain spurs of the intermediate valley ridges by backward denudation to typical triangular slopes. Their classic shaping becomes especially clear between No.12 and 10 (Fig.2 right of No.10). Photo M.Kuhle, 16.9.1997.

↓ *Photo 11.* Long-distance exposure of the Skilbrum massif at 4725 m asl facing NW. From this viewpoint the 7360 m-high main peak of the Upper Baltoro glacier (surface moraine ☐) can be seen (somewhat to the right below No.13). The seemingly higher summits left and right of No.13 are the 7103 m-high Savoia III and the 7263 m-high Savoia Kangri. The marked prehistoric glacier surface (– –) runs between c. 6400 (left) and 6000 m asl (right). (◓) show the glacigenic major forms of flank polishings on back-polished mountain spurs roughened by postglacial breakages in its different petrography: in metamorphic sedimentary rock (◓) and in massif-crystalline rock (◤). (▲) is a postglacial debris slope made up from dislocated ground moraine and crumblings of the bedrock, deposited kame-like against the present-day glacier margin. Photo M.Kuhle, 16.9.1997.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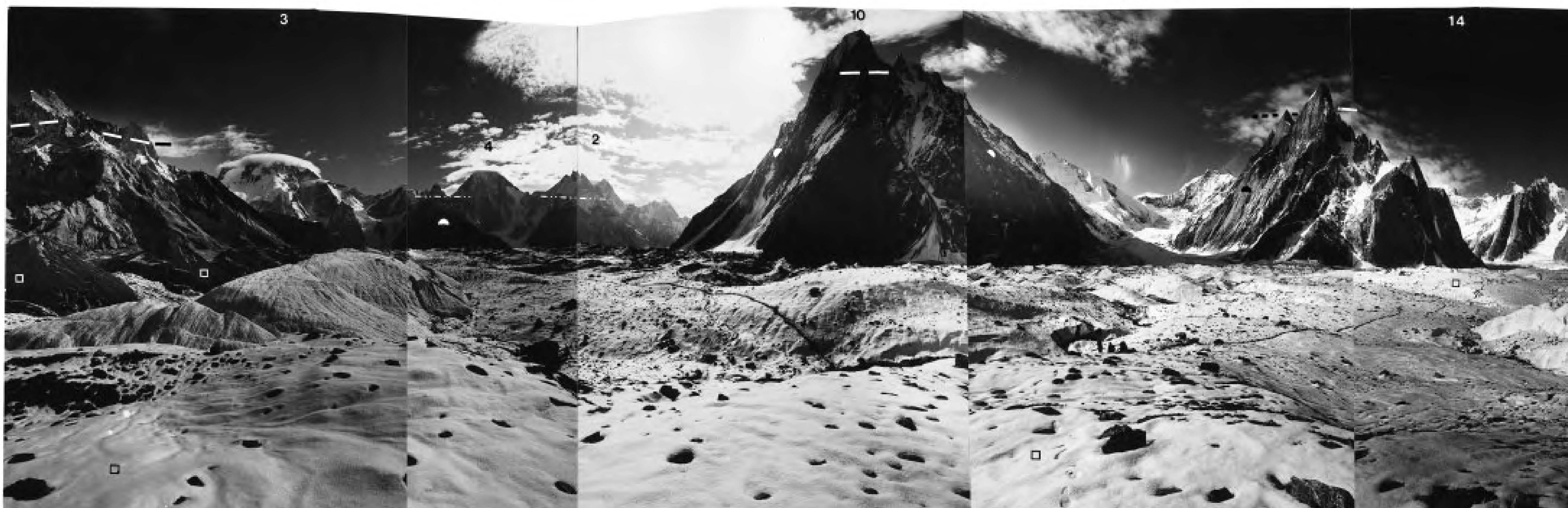
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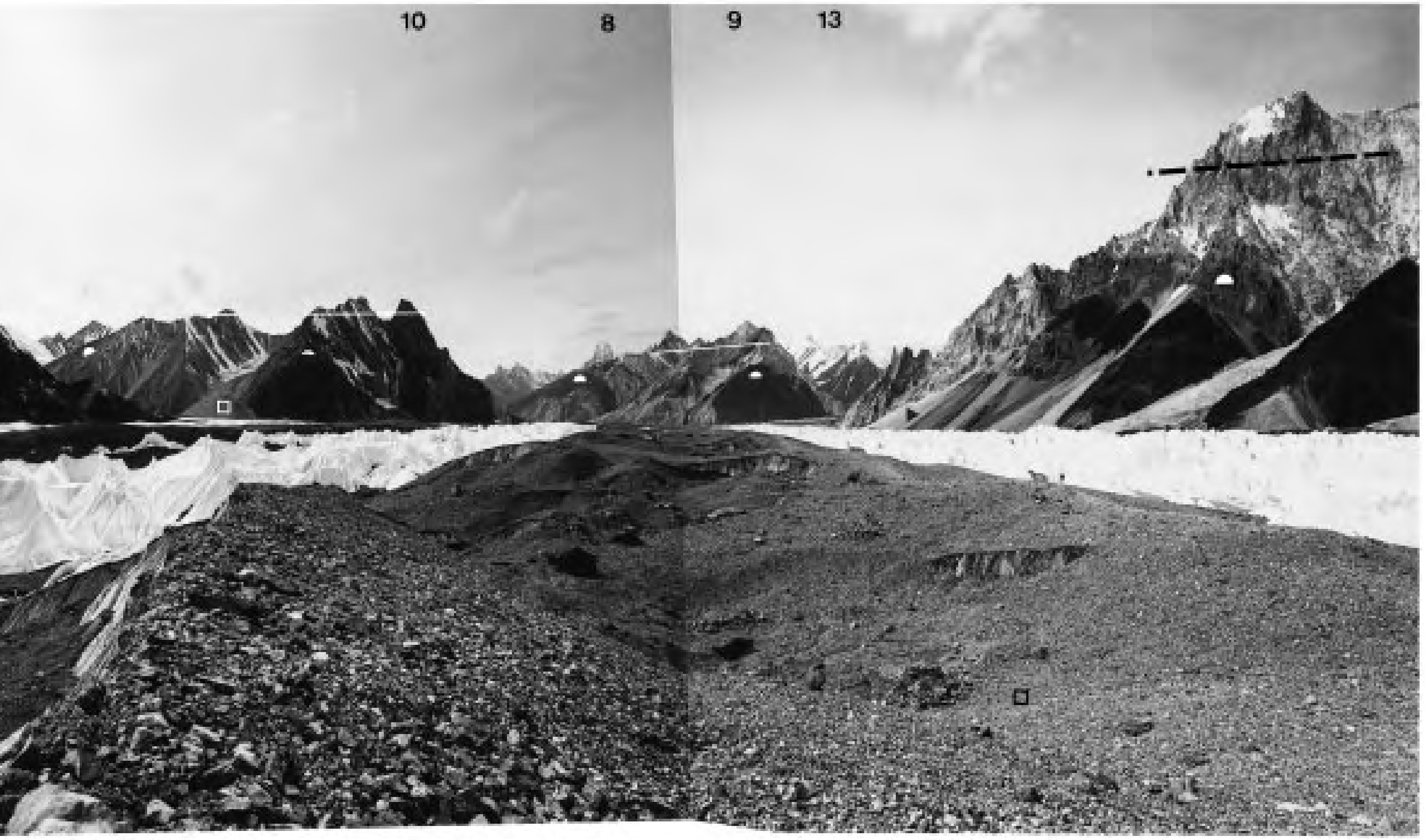
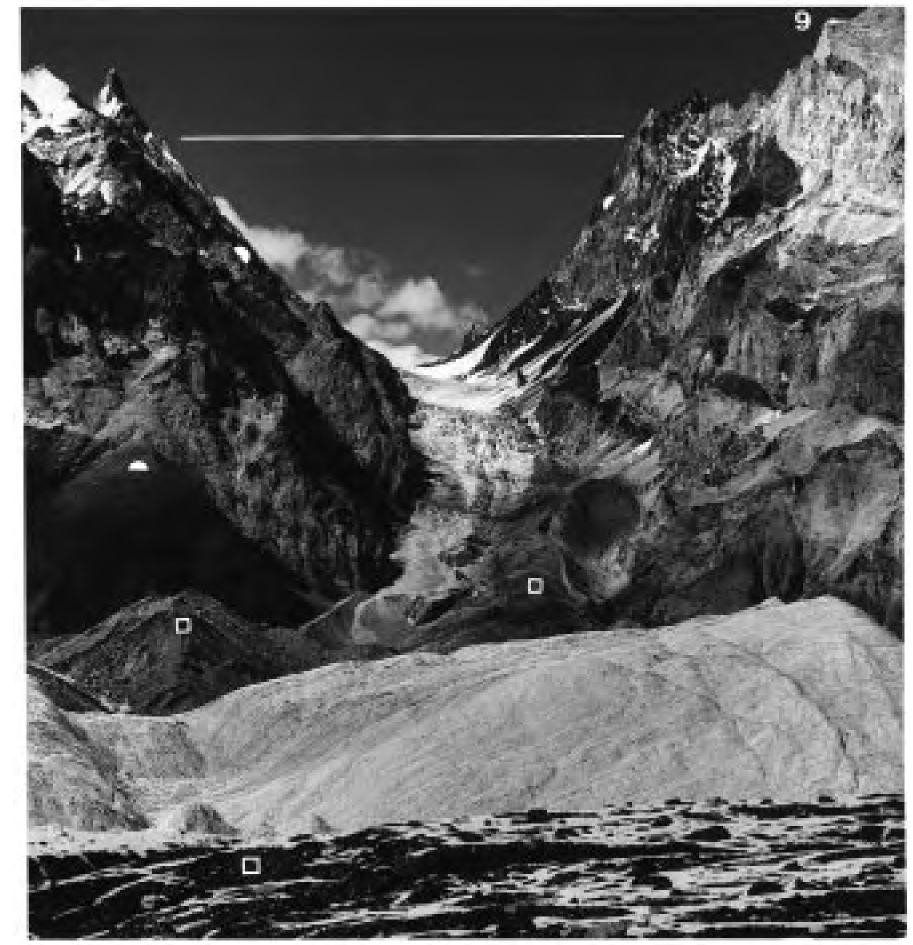


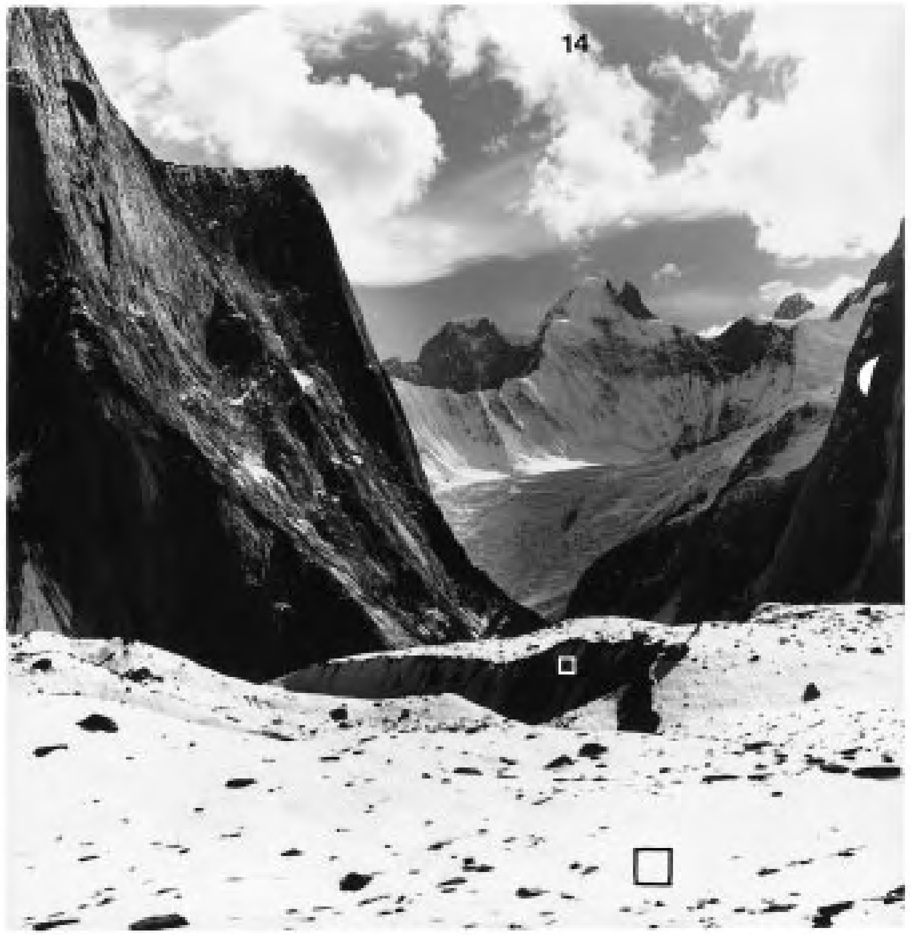
← *Photo 10.* This 360°-panorama was taken at 4725 m asl from the surface moraine (third □ from the left) of the Upper Baltoro glacier. No.2 is the position of the Gasherbrum-group with the satellites of Gasherbrum V (7321 m) and Gasherbrum VI (7003 m) in the ENE. Its short valley glaciers (first and second □ from the right) are the orographic right catchment area of the Baltoro parent glacier. No.7 is the 7312 m-high Baltoro Kangri (Golden Throne), No.6 the Chogolisa (7665 m) situated up-glacier to the SE and S, respectively. Down-glacier facing NW, the Mitre Peak (No.10, 6010 m) can be seen; on the right side of the joining Vigne glacier the Muztagh Tower (No.8, 7273 m, for details see Photo 12), the Marble Peak (No.9, 6238 m) and the Skilbrum (No.13, 7360 m; for details see Photo 11) - the last three mountains beyond Concordia - are visible. In many places remnants of glacigenic flank polishings are evident (⯊). Some of them have developed major forms, i.e. glacially triangular-shaped slopes in the shape of classic back-polished mountain spurs (e.g. ⯊ large) (cf. also Fig.2 between No.9,2,10). The reconstructed maximum High Glacial (LGM) glacier level (– –) ran above these roundings (⯊) , i.e. below the sharpened crests with their acute rock towers. In the case of the important, acute-angled ice confluence of the prehistoric Upper Baltoro glacier and Vigne tributary stream the N-crest of the Chogolisa group has also been sharpened subglacially (– – on the right below No.6). To enable one to imagine the important dimensions of the present-day subtropical Baltoro valley glacier system with glacier extensions of 2-3 km and over 100 m-wide strings of surface moraine, our camp with its large kitchen-tent situated below No.8-9 provides a rough scale. Photo M.Kuhle, 16.9.1997.

← *Photo 13.* 360°-panorama taken at 4540 m asl, 3.2 km down the glacier, i.e. W of Concordia, from the middle of the Baltoro glacier. The surface moraine (□) is covered with fresh snow which fell a few days before. On the left as well as on the right margin of the panorama, in the N, the 6238 m-high Marble Peak (No.9) is depicted. No.3 is the 8047 m-high Broad Peak approx. in the NE, wearing a cap of clouds; No.4 the 7980 m-high Gasherbrum IV in the E, also with a cloud cover. On the right the 7321 m-high Gasherbrum V with its satellites is situated and somewhat behind on the right Gasherbrum I (Hidden Peak, No.2), the 8068 m-high main summit of the Gasherbrum group. The G I is part of the catchment area of the Abruzzi glacier joining the Upper Baltoro glacier. No.10 = Mitre Peak (6010 m-high); on its right side, directly to the S, the Nuating- or Mitre glacier. In this direction, on the path in the foreground tramped into the snow by expedition-porters, three persons provide a scale. No.14 is the 6781 m-high Biarchedi, belonging to the catchment area of the orographic left Biarchedi tributary glacier, which also merges into the Baltoro glacier from the S. No.11 marks the Paiju Peak (c. 6600 m-high) lying c. 30 km W of the viewpoint down the Baltoro glacier. (⯊) indicate rock smoothings derived from the prehistoric glacigenic flank polishing which, since the drop of the ice level, have still not been completely remoulded and blurred. In places, these polish bands have created the characteristic glacigenic triangular-shaped slopes (e.g. ⯊ on the very left and the two first from the right) (Fig.2 No.4,9,10). They have been developed in massive-crystalline rock (⯊ right and left below No.10 and 14) as well as in sedimentary rock (⯊ on the very left and the two on the right). Their occurrence in the latter has a greater indicator value, because in the massive-crystalline rock they largely tally with the rock structure anyway and have even partly been preformed by it. (– –) signifies the LGM-glacier level about 6000 m asl, i.e. c. 1400-1500 m above the present-day glacier surface. Photo M.Kuhle, 3.9.1997.

← *Photo 12.* Long-distance exposure of the summit of the Muztagh Tower (No.8, 7273 m) from the Upper Baltoro glacier at 4725 m asl facing WNW, looking across a far lower rock crest in front which shows patches of firn ice. The LGM-glacier level (▬ ▬) lies at ca. 6000 m. The lateral undercutting of this mountain by the High Glacial ice stream network created a classic glacial horn. It is the highest glacial horn on earth. Its longish, lance-shaped outline from ESE to WNW (Fig.2 No.8) testifies a superficial ice drainage in this direction axis. We are looking to the ESE short side of the horn, i.e. to its ESE crest. Photo M.Kuhle, 16.9.1997.

→ *Photo 14.* Looking from 4580 m asl to the N facing a hanging valley glacier in the orographic right flank of the Baltoro valley. It is covered with dark surface moraine (☐ right) and flows into the Baltoro glacier (the two ☐ on the left). (——) indicates the glacier level during the LGM reconstructed according to the trough- and U-shaped valley cross-profile (Fig.2 No.9). Up to this height the preservation of the glacigenic flank polishings is complete and continuous showing the profile lines characteristic of trough valleys (◥ ◤). An even considerably higher prehistoric ice level is not to be ruled out. However, the heavy decomposition of the valley flank into ribs, pillars and sharp satellites and towers does not allow an unambigious evidence. The orographic left trough valley flank is the Marble Peak (No.9) -WSW face. Photo M.Kuhle, 3.9.1997.





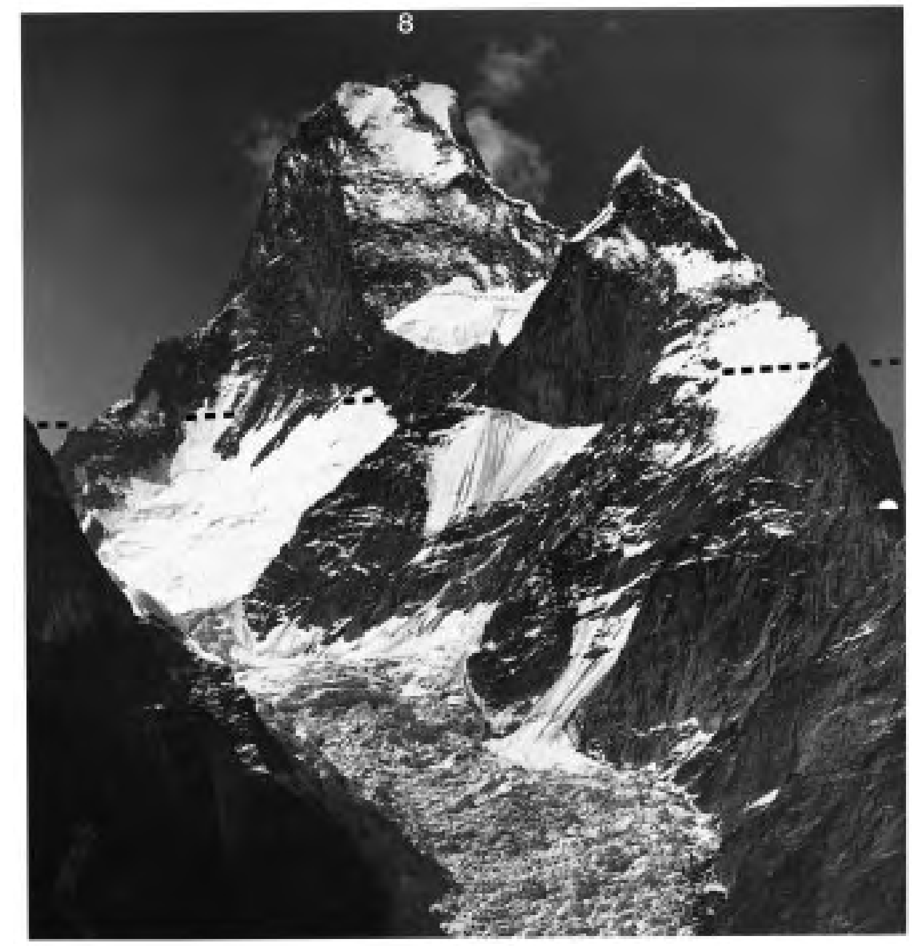
← *Photo 15.* Taken at 4465 m asl from the surface moraine (□) of the Baltoro glacier covered with fresh snow. Viewpoint a few 100 metres up-glacier from the junction with the orographic left side valley glacier, the Biarchedi glacier, facing SW to the Biarchedi (6781 m, No.14) and the Upper Biarchedi glacier with its large ice fall. (◥) marks the glacigenically polished trough- i.e. U-shaped valley flanks (Fig.2 No.14). The orographic left flank shows a polish face on an outcropping rock pillar (◤ white) rounded in the horizontal- and stretched in the vertical profile. A bi-concave polish profile, i.e. a rock face hollowed out horizontally and vertically by polishing, can be observed on the orographic right flank (◥ black). The granite bedrock of these trough valley walls has got a fibrous structure and roughening caused by numerous rock breakages since the deglaciation (Fig.2 No.10). Their slabby appearance is characteristic of massive-crystalline rocks and can also be recognized on the flanks of fluvially shaped valleys. Thus, these trough valley flanks do not show any obviously glacial indicators of flank polishing. This means that at this location a fluvial phenomenon of convergence going back to the rock structure can only be ruled out by relating the positions to the numerous unambiguous forms of glacier polishing in the nearest vicinity. Photo M.Kuhle 3.9.1997.

→ *Photo 16.* Viewpoint at c. 3950 m asl approx. 1.7 km away from Urdukas down the Baltoro glacier on the surface moraine (□) of the main glacier (Baltoro) between huge granite boulders (□ in the foreground). Expedition porters with their loads in the foreground (for scale). At a distance of 31 km towards the E, still beyond Concordia, the 7980 m-high Gasherbrum IV (No.4) is situated; left of it (to the N) the 8047 m-high Broad Peak (No.3). The shortened perspective shows the Baltoro trough almost half-filled by the present-day glacier. Glacigenic flank polishings on both sides (◓) provide evidence of its ice filling during the LGM. (– –) mark the ice level at Concordia (in the background further right) to have been about 6000 m (Fig.2 No.9) and in the middle valley section (middleground on the left) about 5600 m asl at that time. Photo M.Kuhle 18.9.1997.

← *Photo 17.* 350°-panorama taken at c. 4330 m asl from the c. 25 m-high, striking ice ridge covered with surface moraine (the left and right □, next to three persons) in the middle of the Baltoro glacier (35°44′ 40″ N/76°25′ 40″ E) in the area of the Gore Camp. Upward of the Baltoro glacier, W of the viewpoint, Gasherbrum IV (No.4) and Mitre Peak (No.10) are situated and downward, in the E, the c. 6600 m-high Paiju Peak (No.11). From the orographic right the Biange glacier joins the Baltoro glacier (□ centre). It emerges from the tributary streams Dre glacier and Younghusband glacier, flowing together below the Muztagh Tower (No.8, 7273 m) which lies NE from here. The Ste Ste Saddle, the 5869 m-high source saddle of the Younghusband glacier, becomes visible on the right side of the Muztagh Tower. No.5 marks the 7821 m-high Masherbrum, SSE of this viewpoint. In front of this mountain, i.e. below its NE-face, the Yermandu glacier joins the Baltoro. This confluence is below No.27, but 4.2 km before the somewhat over 6000 m-high Urdokas Peak (No.27.) (■) indicate end moraines in hanging- and side valley mouths the glaciers of which no longer reach the Baltoro parent glacier. Their material consists of older (LGM- to Late Glacial) ground moraine which has been pushed together. It is mixed up with the edged debris of boulders of postglacial crumblings. Over large parts the valley flanks have been rounded by the work of the prehistoric glacier. However, glacigenic rock polishings occur only in places (◓). Their state of development and preservation varies according to the petrovariance of the bedrocks. As for the granite it takes the form of large slabs (the two right ◓) but as for the sedimentary rocks it is finely splintered (the three ◓ between No.8 and 4). (– –) shows the highest geomorphologically demonstrable glacier level. The polish line (– –) runs almost continuously about 6150-6000 m asl. (⇩) signifies erosion rills which have been set - and still are - into glacigenic trough flanks by the meltwater of small Late Glacial hanging glaciers and Holocene to historic snow fields at right angles to the flank polishing. Photo M.Kuhle 2.9.1997.

↑ *Photo 18.* From the central surface moraine (□) of the Biange glacier composed of the side moraines of the Dre glacier (left of No.8) and the Younghusband glacier (between No.8 and 15), a 360°-panorama was taken at 4480 m asl (35°46′ 25″ N/76°23′ 30″ E). The surface moraine is recruited from the present-day flank polishing which undercuts the rock slopes of the Muztagh Tower (No.8) The Ice Age ice fillings of the upper valley chambers being then a good 1000 m thicker (– – on both sides of No.8) lent the Muztagh Tower the shape of a glacial horn. No. 15 is the 6930 m-high W-satellite of the Skilbrum. Between No.8 and 15 the Ste Ste Saddle (5869 m) can be seen. Downwards to the SSW, the Biange glacier converges with the Baltoro parent glacier; the 6344 m-high N-satellite (No.16) of the Biarchedi stands behind. It consists of granite. (⏶) marks the glacigenic flank roundings and polish remnants. In places, rills and wall gorges are filled with superficially reshaped remnants of prehistoric moraine (■ black) or with debris cones, which are to a great extent made up from crumblings (▼). The present-day glacigenic lateral erosion causes even steeper lower slopes in some places (⇩). (□ white) is a glaciofluvially dissected end moraine which was accumulated by an interglacial hanging glacier - and still is by that one of today. (– –) is the High Glacial (LGM) glacier level. Photo M.Kuhle, 5.9.1997.

→ *Photo 22.* At 4220 m asl from the middle of the Baltoro glacier, right N of the 6344 m-peak (No.16) and S of the locality Biange Goro (35°44′ N/76°23′ E) a 190°-panorama, taken from one of the 8 to 14 m-high ice hills covered by surface moraine (□). The left margin of the panorama lies towards the SSE; the Masherbrum (No.5, 7821 m) stands in the SW beyond the inflow of the Yermanendu glacier (on the left below ◇). The Baltoro glacier, still 24 km-long from here, flows down to the W in a slight S-bend (right of the panorama centre). As far as approx. 1800 m above the present-day ice surface the orographic right side of the glacier is flanked by valley slopes polished glacigenically during the prehistoric period (⏶). On these slopes (■ black) - and also on the orographic left side (■ white) - Ice Age to Late Glacial ground moraines have been preserved. They are situated up to 700 m above today's glacier level (■ black). (▽) are deposits of detritus at the exits of gullies which are adjusted to the glacier surface and provide the lateral moraines with detritus. (◇) indicate fresh rock crumblings due to the modern lateral erosion of the Yermanendu glacier. (– –) is the highest demonstrable prehistoric glacier level running about 6000-6150 m asl. Photo M.Kuhle, 6.9.1997.

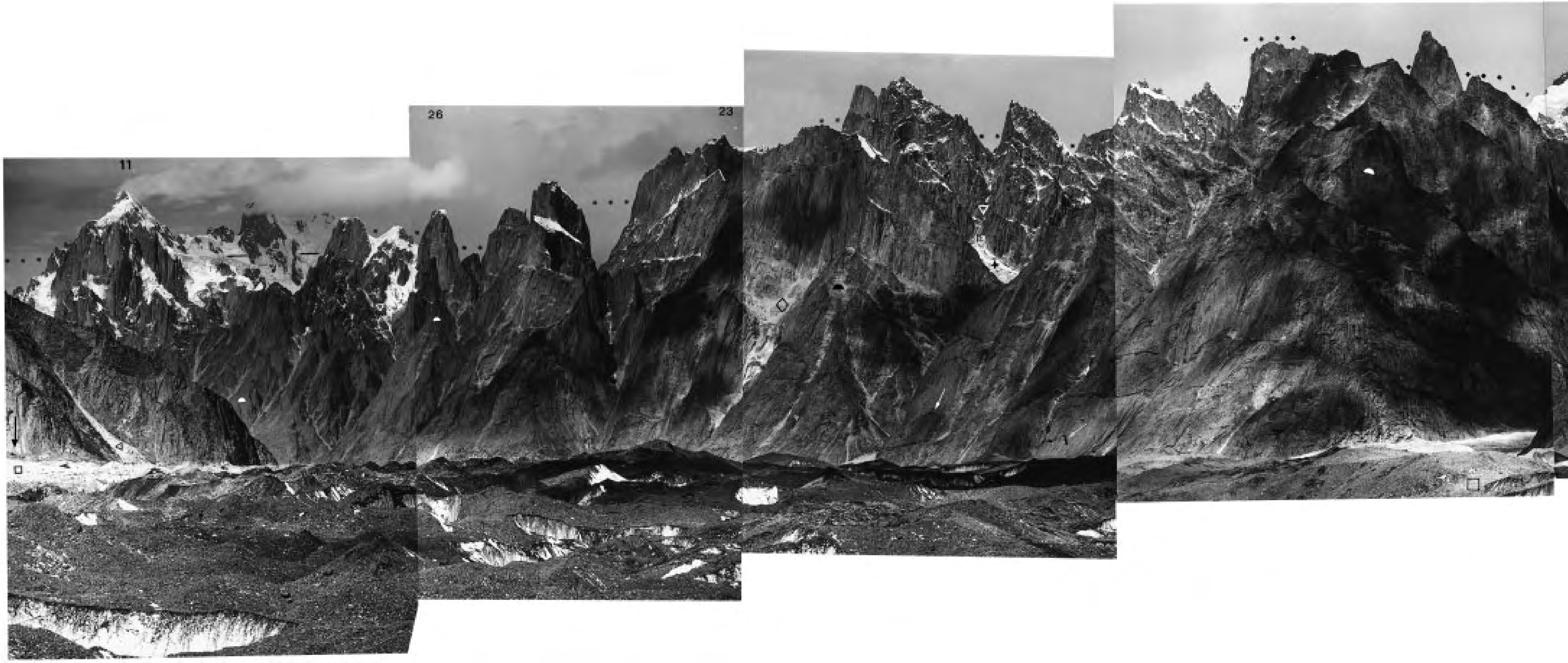


↑ *Photo 19.* This picture shows a section of the panorama-photos 17 and 18 in a slightly changed view. The S-face of the Muztagh Tower (No.8, 7273 m) and its E-satellite, the 6719 m-high Black Tooth, falls away to the Dre glacier by maximally 1600 m. Despite the homogeneous petrography it is divided into two parts, i.e. at the point where the up to 80°-steep upper wall passes into the flatter lower wall with an inclination of 60-45°, a bend goes through. This indicates the Ice Age glacier level (— —). At places where - dependent on the spur position - the denudation by avalanches is lacking, a glacigenic rock rounding has been preserved (◓). The Muztagh Tower and the Black Tooth are situated on a ENE/WSW-trending crest (cf. Photo 12). They form a basally connected, glacial horn with two summits (Fig.2 No.8). Photo M.Kuhle, 2.9.1997.

← *Photo 21.* From the confluence area of the Yermanendu- (middleground) and the Baltoro (foreground) glacier looking SSW (c. 200°) to the Masherbrum group. (□) is one of the largest boulders from the surface moraine of the Baltoro glacier. At the time when the picture was taken, the moraine - lying 1000 altitude-metres below the present-day glacier snow-line (ELA) - was covered by a three decimetres-thick layer of fresh snow. The 7821 m-high Masherbrum (No.5), the c. 3000 m-high flank of which falls away to the Yermanendu glacier, consists of reddish granite. The true shady, bi-concave section of the flank's NE-wall rises compactly over a distance of 2000 m and with an incline of 50-80° from two wall foot glaciers up to the firn-covered summit (the shaded wall). Towards the W there stands the 7163 m-high Masherbrum E (No.33). Its flanks are mantled by decametres-thick ice with numerous ice balconies. Photo M.Kuhle, 1.9.1997.

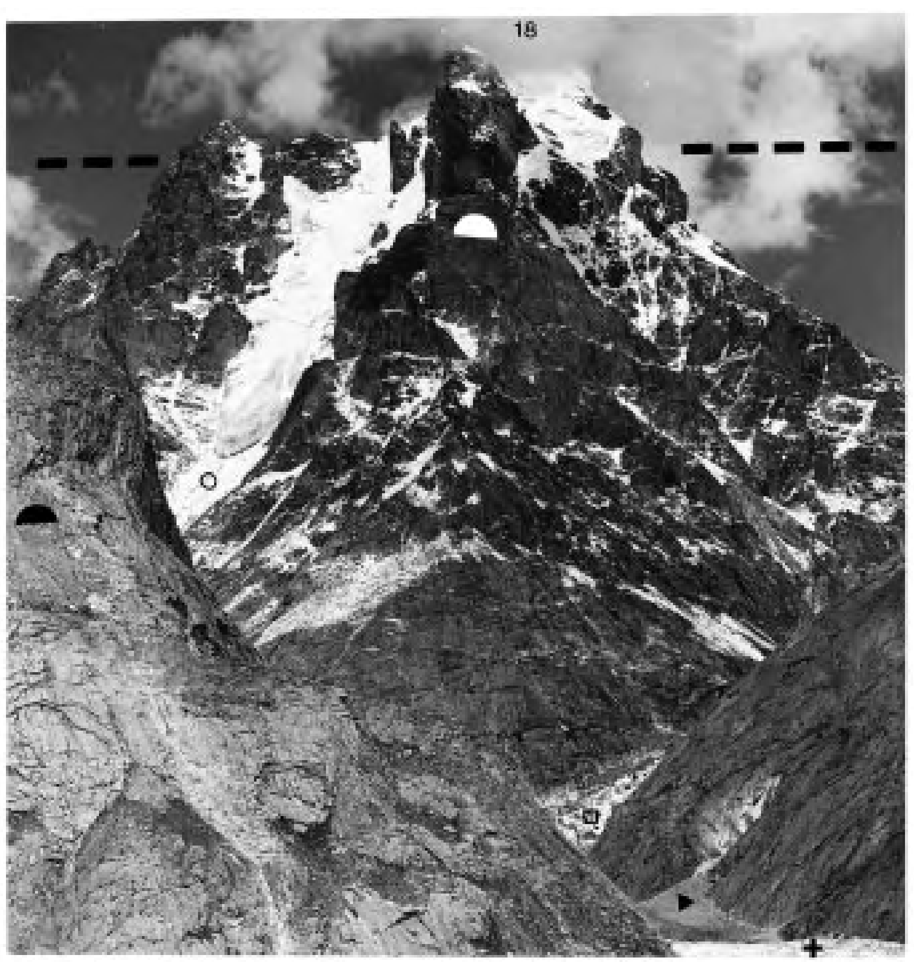
↑ *Photo 20.* Panorama at 4310 m asl from the middle of the freshly snow-covered Baltoro glacier (□) W of the Gore Camp across the orographic left valley flank taken from S to W. The glacigenically rounded granites of the left half of the panorama (◓) belong to the N-slopes of the Biarchedi group. No.5 is the 7821 m-high Masherbrum, the steep NE-flank of which falls away towards the Yermanendu glacier (above □). No.27 is the 6368 m-high Urdokas group beyond the Mandu glacier, a further right tributary glacier of the Baltoro glacier from the Masherbrum group. (■) are morainic accumulations of different ages: (■ white) is a prehistoric ground moraine core covered by avalanches, which is preserved in a valley inlet; (■ black on the left) shows prehistoric moraine material with a permafrost core causing a rock glacier-like self-movement of the moraine. (■ black on the right) is a prehistoric, probably Late Glacial ground moraine remnant c. 200 m above the Baltoro glacier. (▽) are hanging- and wall-foot glaciers remoulding and erasing the Ice Age horizontal glacigenic flank polishing (Fig.2 above No.14) down-slopes, i.e. at a right angle. (– –) marks the Ice Age glacier surface reconstructed with the help of partly preserved indicators of glacier polishing. Its level runs about 6000 m. Photo M.Kuhle, 1.9.1997.

↑ *Photo 24.* At 4200 m asl, c. 1.8 km away from Camp Gore I, down the glacier, is a glacier mill on the surface in the middle of the Baltoro glacier (35°44′ N/76°19′ 58″ E) showing especially large dimensions. (↑) points to a porter with a load (for comparison of the size) on the upper edge of the glacier mill. Approximately 20 m below the glacier surface covered by moraine (□ white) the supra- and intra-glacial water joins creating a 150 m^2 extended meltwater reservoir. (▽) marks the steep tongue-front of the advancing (1997) Lhungka glacier. Two tributary streams are part of its catchment area flowing down on the S-flank of the 6307 m-high Lhungka summit (above ▽). There is a further tributary glacier in the western parallel valley, which today no longer reaches the Baltoro glacier (above □ black). Both the short-trough-like tributary valleys dissect the orographic right Baltoro valley slope, which has been polished by the High Glacial ice (◓). (■) marks a prehistoric remnant of ground moraine. (– –) indicates the LGM ice level. Photo M.Kuhle, 27.8.1997.



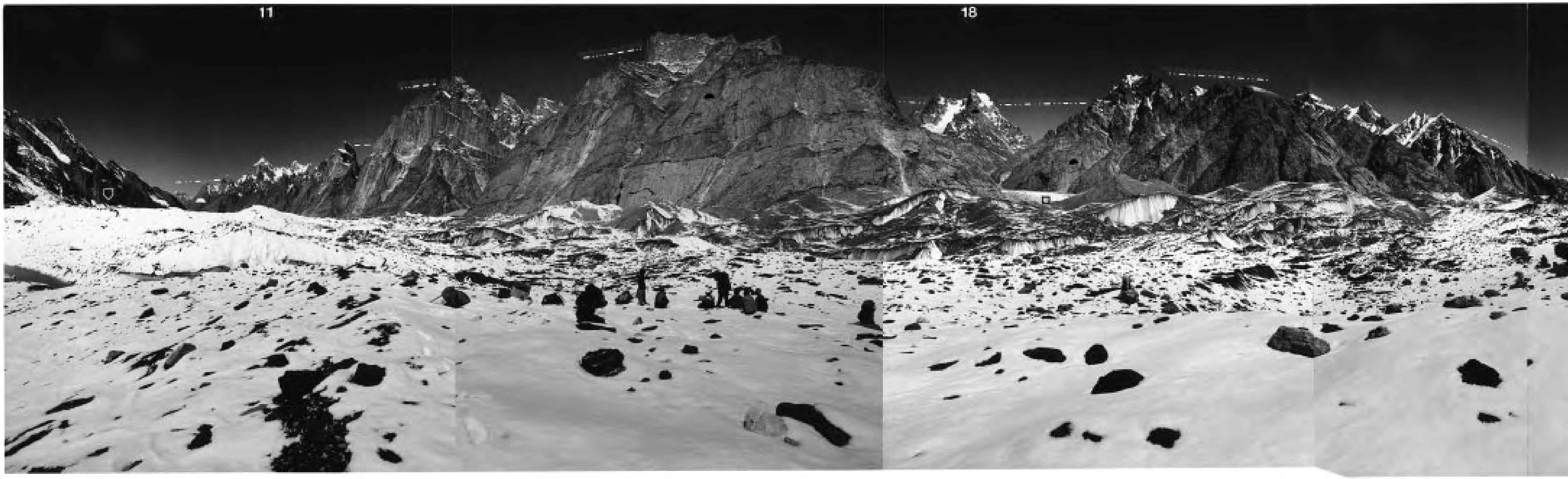
↑ *Photo 23.* Detail of the orographic right Baltoro trough valley flank at 4220 m, looking from the middle of the glacier downwards in an approx. W direction. The glacial horn of the Uli Biaho I (No.26, 6417 m) stands exactly in the W; to its left, 3 km down the glacier, is the c. 6600 m-high Paiju Peak (No.11). The main summit of the Trango group (No.23) is the 6286 m-high Trango Cathedral (Trango I). No.17 indicates the 6729 m-high Mount Biale. In front of it, i.e. E of this mountain, the Muztagh glacier joins the Baltoro glacier from NNW (above ☐ on the right). The Mt. Biale, the N-flank of which is turned away in this perspective, belongs to the catchment area of the Muztagh glacier. (◓) marks the glacigenic rock roundings which are often combined with glacigenically triangular-shaped slopes on back-polished mountain spurs. (◇) is a rock face destroyed on a relatively large scale by falling rocks. Above (◇) the bedrock granite develops column-like structures of breakages. (✓ white) signifies crescent-shaped crumblings. Avalanche tracks (▽ white) adjusted to the hanging glaciers (+) in the steep walls, destroy the rock, too. (▽ black) is a gully, progressively filled up by moraine and also debris of fresh crumblings in a downward direction. (↓ black) shows a fresh break-away which results from the undercutting by the orographic left glacier margin. Through this debris and the material of crumblings the increasing thickness of the surface moraine (☐) towards the glacier margins can be explained. (– –) is the evident maximum prehistoric ice level at altitudes about 6000 m. Photo M.Kuhle, 6.9.1997.

→ *Photo 27.* No.18 is the 6224 m-peak in the confluence-triangle between Chagaran- (☐) and Muztagh glacier (+) (35°48′ 40″ N/ 76°17′ 40″ E). The wall gorge glacier in its SW-flank, hanging steeply down, terminates as a cold to tempered glacier with a steep tongue-front and a meltwater-string only in the middle of summer (○). (◓) are glacigenic rock smoothings which occur up to an altitude at 6050 m asl. They provide evidence of a High Glacial (LGM) glacier level (– –) running at least 1600 m above the present-day level of the Muztagh glacier (+). (▼) is an a few centuries-old debris cone below a fresh crumbling which has been (and still is) sedimented on the surface of the Muztagh glacier. Proximally its mantle is continuously supplemented by the debris of the crumblings whilst distally it is eroded by the glacier flow. Photo M.Kuhle, 1.9.1997.

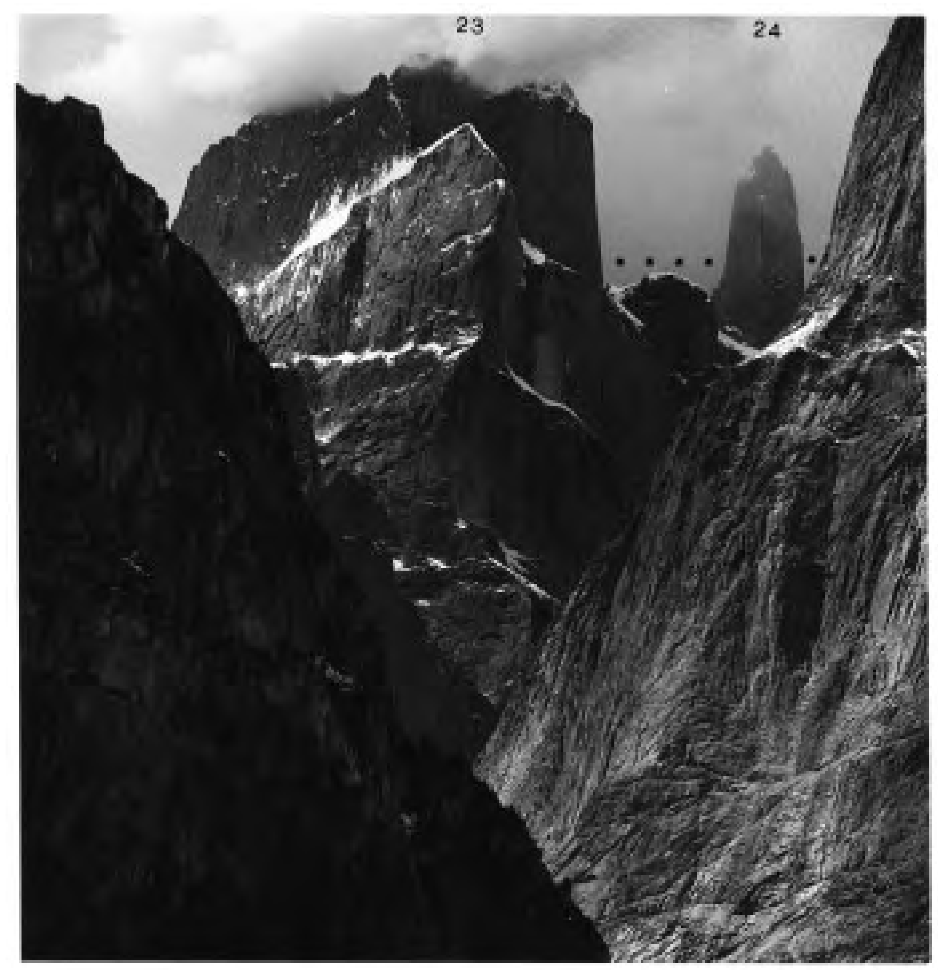


→ *Photo 29:* From the surface moraine of the Baltoro glacier (35°43′ 40″ N/76°14′ 25″ E) at 3920 m asl, looking up the Dunge glacier (□) facing NNW (335°) towards the 6544 m-summit (No.25), the western satellite of the Kruksum (No.19, 6600 m). It is separated by an orographic left tributary stream of the Dunge glacier (in the blind spot of the rock crest, below No.19 middleground) from the Mt. Biale W-face (also not visible). On the orographic right - seen from the Dunge glacier as well as from the Baltoro glacier - the granite-crest of the Trango Cathedral rises 2000 m, i.e. up to over 6000 m. There, the glacigenic flank smoothings are preserved the best (◥). (– –) indicates the LGM glacier level. At the valley head of the Dunge valley it has attained a height of c. 6200 m. The glacier snow-line (ELA) runs at c. 4800 m asl, that is in the upper third of the ice fall, torn by transverse crevasses. Photo M.Kuhle, 18.9.1997.

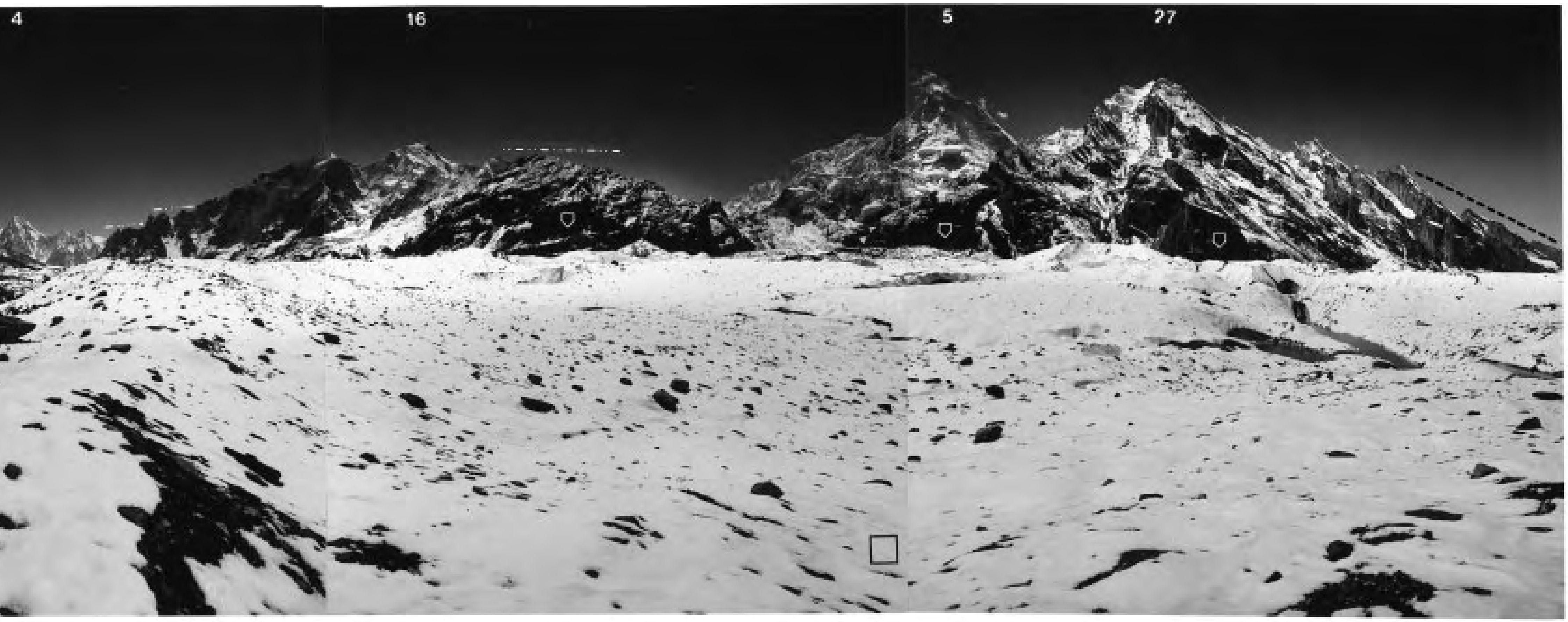
→ *Photo 30.* ESE-flank of the Trango Cathedral (Trango I, No.23, 6286 m); the picture was taken from the confluence area of the Dunge glacier and the Baltoro parent glacier after a sudden drop in temperature with heavy snowfall lasting two days. The snow cover indicates the flatter parts of the granite face, whilst at places where the snow is lacking the face is steeper. The former have been rounded glacigenically. However, the rounded rock forms do not occur up to the Ice Age glacier level (– –) everywhere. On the rounded, as well as on the edged, faces of the steep walls sharp breakages have been formed. Their slabby break-aways seem to preserve the round forms rounded and the straight ones edged and straight. Photo M.Kuhle, 28.8.1997.



↑ *Photo 25.* 360°-panorama from the orographic right, i.e. northern large medial moraine string of the Baltoro glacier (☐ large; 35°44′ 20″ N/76°18′ E) at c. 4090 m asl. The porter group with its load (between No.11 and 18) having a rest, gives an impression of the dimensions of the surface moraine boulders thawing out of the fresh snow. Down the glacier towards the W the c. 6600 m-high Paiju (No.11); in the N a 6224 m-high granite peak is situated above the inflow of the Chagaran glacier into the Muztagh glacier (No.18), which is one of the large orographic right tributary glaciers of the Baltoro glacier (☐ below No.18). Gasherbrum IV (No.4, 7980 m) stands up-glacier in the E. No.16 is the 6344 m-high Serac Peak on the orographic left of the Baltoro glacier. At its base the Yermandu glacier joins the Baltoro glacier. The commanding orographic left summit is the Masherbrum (No.5, 7821 m). Its N-flank provides the Mandu glacier, situated below, with ice. Between the second and third (⇩) from the right this glacier reaches the Baltoro parent glacier. No.27 are the up to 6368 m-high Urdokas Peaks on the orographic left side of the Baltoro glacier. Due to the moraine debris, the glacier margin is particularly capable of polishing and undercuts the valley flanks by lateral erosion. This becomes especially clear in the steepening of the lower slopes (⇩). On the orographic right valley flanks consisting of granite, the prehistoric glacier polishings are the best preserved (◓). Glacigenic rock roundings occur on the spur-shaped flank of the 6224 m-summit up to approx. 6000 m asl (◓ below No.18) (see Photo 27). According to these traces of the polish line, the Ice Age (LGM) segments of the glacier level (– –) can be reconstructed at an altitude of 5950 to 6100 m asl. Photo M.Kuhle, 1.9.1997.



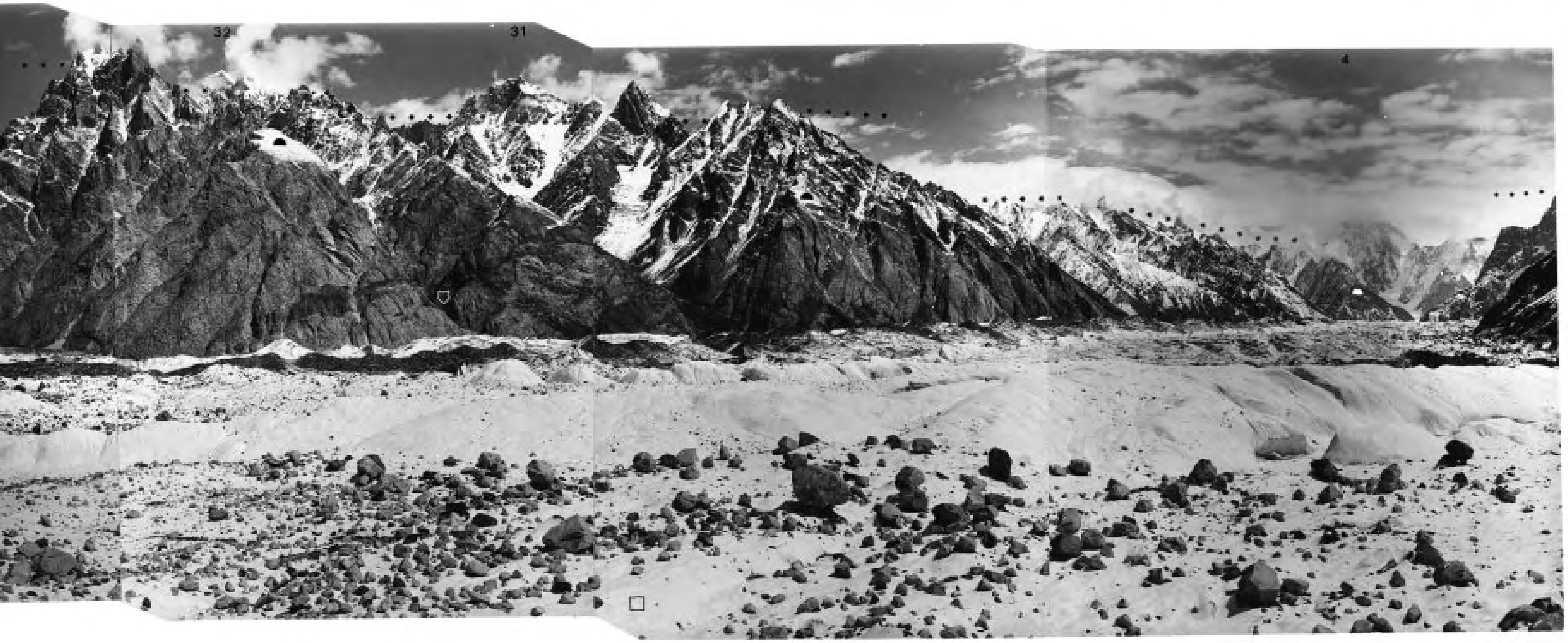
→ *Photo 31.* The ESE- (left of No.23) and ENE-flanks (right of No.23) of the Trango Cathedral (Trango I, No.30, 6286 m). View at 4150 m from the middle of the Mandu glacier not far from its inflow into the Baltoro parent glacier. Nearby is the 6239 m-high Trango Tower (No.24) the summit form of which is typical of the petrography of granite, but which at the same time can also be explained by glacial erosion. (– –) is the highest prehistoric glacier level at 6050 m asl to be evidenced here. Besides the minor decline of its talweg, the Ice Age glacier drainage of this valley system, fringed by over 2500 m-high trough flanks, has been heavily impeded by friction forces. They have found a glaciogemorphological expression in the tub-like steepness of the valley walls and the extraordinarily uniform flank abrasion. Photo M.Kuhle, 31.8.1997.



← *Photo 32*. On its orographic right the Baltoro glacier is flanked by the locality of Urdokas - a neoglacial to historic lateral moraine area. Here stands the Urdokas Peak, the northernmost satellite of which (No.27, c. 5600 m, 35°43′ 40″ N/76°14′ 20″ E) is seen from the E. Up to the highest point of the summit the flanks are glacigenically abraded (◥) and sharpened. This documents that they were completely overflowed by the LGM-glacier. During the Late Glacial, when the glacier level was lowered, a polish cavetto (⬠) has been developed in the rock face of the summit. This points to a longer-surviving height of the glacier margin at this level. The torn glacier ice in the fore- and middleground is part of the Urdokas glacier. This present-day hanging glacier marginally underpolishes the summit of the Urdokas satellite. Fresh snow had fallen during the previous two days. Photo M.Kuhle, 29.8.1997.

↑ *Photo 26.* This 180°-panorama was taken from the middle of the Mandu glacier (□ half-right in the foreground), c. 1.2 km upwards (to the S) from its confluence with the Baltoro glacier (35°43′ 10″ N/76°17′ 55″ E, 4150 m asl), looking across the Baltoro glacier and its orographic right valley flank. The panorama's left margin is in the W, the 6224 m-peak (No.18) is in the N and Concordia with the Gasherbrum IV-Peak (hidden by clouds), forming the valley exit, lies in the E. No.23 is the 6286 m-high Trango Cathedral, the main-summit of the Trango Group. No.24 is the Trango Tower (6239 m) built up of granite. The highest summit in the middle section of the right Baltoro valley flank is the Mount Biale (No.17, 6729 m); coming out of its SSE-flank, the Biale tributary stream reaches the Baltoro parent glacier from the N (□ middleground on the left). (□ in the middleground on the right) is the inflow of the ramifying Muztagh glacier, also from the N, into the parent glacier. (■) shows end moraine material in the forefield of the present-day Lhungka glacier flowing down from the Lhungka Peak (No.32, 6307 m asl) and Biange Peak (No.31, 6427 m asl). The orographic right, large-scale flank polishings of the Baltoro valley (◓) have only been slightly geomorphologically modified by concordant crumblings. They start at the horizontal stratification surfaces (↓). These are classic release joints as they are characteristic of granite. (◇) marks still fresher, but different crumblings the form of which is controlled by a vertical, pillar-like structure of the granite bedrock. (⇩) show wall gorges which become narrower in a downward direction. From time to time, i.e. during the Pleistocene periods, the embedment of a valley glacier has more and more put a stop to their development. (– –) indicates the LGM glacier level. Despite its position about 1000 m below the present-day ELA (snow-line), the surface moraine of the Baltoro glacier wears a 30 cm-thick cover of freshly fallen snow (□ foreground). Photo M.Kuhle, 31.8.1997.

→ *Photo 33.* View from the orographic left, moraine-covered margin (□) of the Baltoro glacier in the area of the inflow of the western Urdokas- (35°43′ 20″ N/76°15′ E, 3990 m asl) into the Baltoro glacier, facing SE. The valley flank of granite bedrock is perfectly glacigenically rounded (◓). Its large-scale vertical pattern of dark water stripes, indicating the areas of the down-flowing water, makes clear, that - with the exception of only one gully (above ▼) - up to the present no fluvial gully-system could develop. At many places the almost horizontal glacier striae and lunar-like fractures are overlain by water stripes parallel to the line of dip without having been reshaped or disintegrated. (▼) is a fresh small cone of coarse granite-debris which has been broken off from the young, steep and narrow erosion rill and eroded during the postglacial period. (– –) shows the minimum altitude of the prehistoric glacier level derived from the flank which is polished round up to the culmination. Photo M.Kuhle, 8.9.1997.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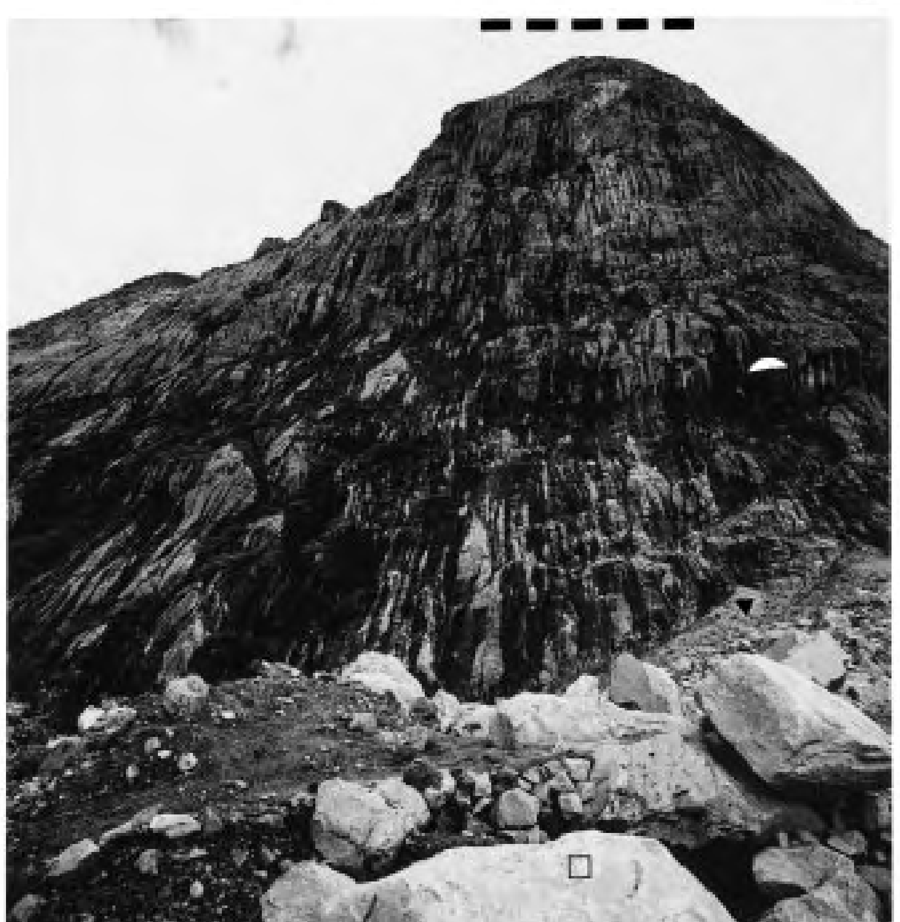

↑ *Photo 28.* 180°-panorama from Urdokas, i.e. from the orographic left neoglacial lateral moraine of the Baltoro glacier (35°43′ 50″ N/76°16′ E) at 4120 m asl: on the very left the Paiju Peak (No.11, c. 6600 m-high) in the W; the Uli Biaho (No.26, 6417 m), the Trango Cathedral (No.24, 6286 m), next to it the Trango Tower (No.23, 6239 m) and Mt. Biale (No.17, 6729 m) in the NNW; the Biange Peak (No.31, 6427 m) in the NE and the Broad Peak (No.3, 8047 m) in the ENE. Correspondingly, the inflows of the following orographic right tributary glaciers into the Baltoro parent glacier are to be recorded as being from WNW via N up to NE: Trango glacier (☐ on the very left), Dunge glacier (☐ half left), Biale glacier (☐ middle) and Muztagh glacier (☐ right). Here, the Baltoro ice stream, completely covered by surface moraine, is exactly 2 km-wide (fore- to middleground). (▲) is fresh end moraine debris of the Urdokas hanging glacier, which comes to an end with a steep tongue breaking c. 200 m above the Baltoro glacier. This glacier hangs in the N-flank of the c. 6000 m-high Urdokas Peak N-satellite. (▽) is an example of a small flank valley running down steeply. Its debris bottom is dislocated by mudflows. In places several metres-thick ground moraine remnants occur under the debris cover. (⇩) indicates an especially large wall gorge in the granite of the orographic right Baltoro valley flank, deepened today and also during the Pleistocene interglacials. In the High Glacial periods it was polished back by an almost 2 km thicker Baltoro ice stream. (◓) are glacigenic flank abrasions preserved up to an altitude at 6050 m asl. They document an LGM-ice- thickness up to at least this height (– –). Photo M.Kuhle, 22.8.1997.


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← *Photo 34.* Viewpoint at 3990 m asl on the Baltoro glacier which is covered with a very coarse surface moraine of edged granite boulders (□, for scale: porters with their loads crossing the boulders). We are in the confluence area of the western Urdokas glacier (35°43′ 20″ N/76°14′ 90″ E), looking SSE to the superstructure of the Urdokas Peak (No.27, altitude of the major summit: 5988 m). (– –) is the highest demonstrable prehistoric glacier level; (◓) mark glacigenic roundings of the bedrock granite from that time. (↓) indicate glacier valley flanks undercut through lateral erosion by the present-day glacier, which has melted back to only minor dimensions. The rock slopes, shaped by the LGM-glaciation, are thus steepened from their lower slopes upwards. This forces rock crumblings, so that the present-day glaciation remoulds the prehistoric glacigenic forms. Photo M.Kuhle, 22.8.1997.

← *Photo 35.* Taken at 3920 m asl from the Baltoro glacier in the confluence area of the western-most Urdokas Peak N-glacier (35°43′ 30″ N/76°13′ 50″ E) facing S, looking up this orographic left tributary glacier to the 6368 m-summit (No.29, Urdokas Group). (– –) is the LGM glacier level of the prehistoric ice stream network. (◓) mark rock roundings of the bedrock granite belonging to this stage of the maximum glaciation. Undercut by the present-day glacier, the rock flanks show numerous fresh crumblings (↓), producing the surface moraine boulders (□). Some of them form glacier tables on the sheer ice. In the foreground three porters for scale. Photo M.Kuhle, 22.8.1997.

→ *Photo 36.* Panorama taken from the orographic left lateral moraine (■) of the present-day Baltoro glacier, here completely covered by a decimetres- to metres-thick polymict surface moraine (□ on the right side), seen from a viewpoint (35°43′ N/76°13′ 45″ E) at c. 3860 m asl at the locality Robutze: in the WSW a northern satellite of the 6251 m-high Liligo Peak (No.28); below are detritus cones (▶) mainly consisting of displaced and remoulded morainic detritus; in front lies the Liligo glacier (□ on the left) only just reaching the Baltoro glacier; in the W stand Paiju Peak (No.11, c. 6600 m-high) and the 6756 m-high Choricho (No.22); the c. 6600 m-high Kruksum (No.19) lies towards the NNE, on the right side of the glacigenically rounded granite walls of the Trango group (the two ◓ on the right). (– –) indicates the LGM-glacier level at 5900 and 6100 m asl. The highest flank polishings reach as far as this altitude (◓). The lateral moraine (■) between the large, polymict, massive-crystalline boulders (○) and the sedimentary rocks shows noticeably more matrix and a finer one than the surface moraine (□ right). Photo M.Kuhle, 22.8.1997.

→ *Photo 37.* Panorama taken from the orographic left margin of the Baltoro glacier (□ black) at the exit of the Liligo valley (35°42′ 45″ N/76°12′ 20″ E; 3850 m asl). The left side of the panorama shows the orographic left margin of the Baltoro glacier, looking upwards. It runs in an ENE-direction. In the centre of the photo, i.e. in the S, the heavily advancing tongue of the Liligo glacier (□ white), blackened by the surface moraine, can be seen. Half-right it comes into contact with the Baltoro glacier margin (□ large). No.28 is the summit superstructure of the 6251 m-high Liligo Peak. The right margin of the photo lies in a WSW-direction looking down the Baltoro valley. (◓) are glacigenic flank smoothings. Depending on the rock they occur differently: the polishing on a thinly stratified rock (◓ on the left) is preserved more smoothly and that one on the coarsely stratified granite more roughly (◓ right). The postglacial fluvial rock gullies (⇓) provide the substratum for the youngest small debris bodies in the form of fans or cones (▼). (○) is the modern, i.e. only a few years old, gravel floor (sander) of the Liligo glacier, being in the process of build-up. (□ on the left) marks a debris slope consisting of surface moraine (person on the right of the left □ for scale). Behind the debris slope is visible the dark-grey ramp of sheer ice on which surface moraine slides down (▽). (□ large) are metres-thick fluvial sands lying on the glacier ice. Photo M.Kuhle, 9.9.1997.