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The Pleistocene Glaciation in the Karakoram-Mountains: Reconstruction of Past Glacier Extensions and Ice Thicknesses

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Abstract: Geomorphological and Quaternary-geological field- and laboratory data (Fig.1) are introduced and interpreted with regard to the maximum Ice Age (LGM) glaciation of the Central- and South Karakoram in the Braldu-, Basna-, Shigar- and Indus valley system as well as on the Deosai plateau between the Skardu Basin and the Astor valley (Fig.2). These data result from two research expeditions in the years 1997 and 2000. They show that between c. 60 and 20 Ka the Central Karakorum and its south slope were covered by a continuous c. 125,000 km² sized ice stream network. This ice stream network flowed together to a joint parent glacier, the Indus glacier. The tongue end of the Indus glacier reached down to 850 ~ 800 m a.s.l. In its centre the surface of this Indus ice stream network reached a height of a good 6000 m. Its most important ice thicknesses amounted to c. 2400 ~ 2900 m.

Keywords: Karakoram; paleoclimate; glacial geomorphology; last glacial period; coalescing ice stream network; glacier thickness

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1 Introduction: Aim of the Study and Characteristics of the Investigation Area

Before the author visited the area of this study (Fig.1 and 2), he had investigated neighboring topographically and climatically representative key areas from 1986 to 1996 and in 1999. His investigations were focused on the question if these areas were glaciated or not and if so, how extended and thick the past ice covers were, situated in the western Himalaya including Zansgar Himal and Nanga Parbat (Fig.1 hatched) massif, in the western Kuenlun (Fig.1 hatched) and Tibet up to the Ladakh Range and in the northern (Fig.1 hatched) and northwestern Karakoram. In all these areas he has found indicators providing evidence of an extended valley-ice-stream-network and a plateau glaciation. In 1997 and 2000 the two geomorphologic expeditions took place shown in Figures 1 and 2. They led to the Central Karakoram (Muztagh- and Haramosh group) and across the Deosai Plateau, both belonging to the catchment area of the middle Indus. Aim of the fieldwork in these new areas was to investigate if they, despite their current semiaridity, were also glaciated and

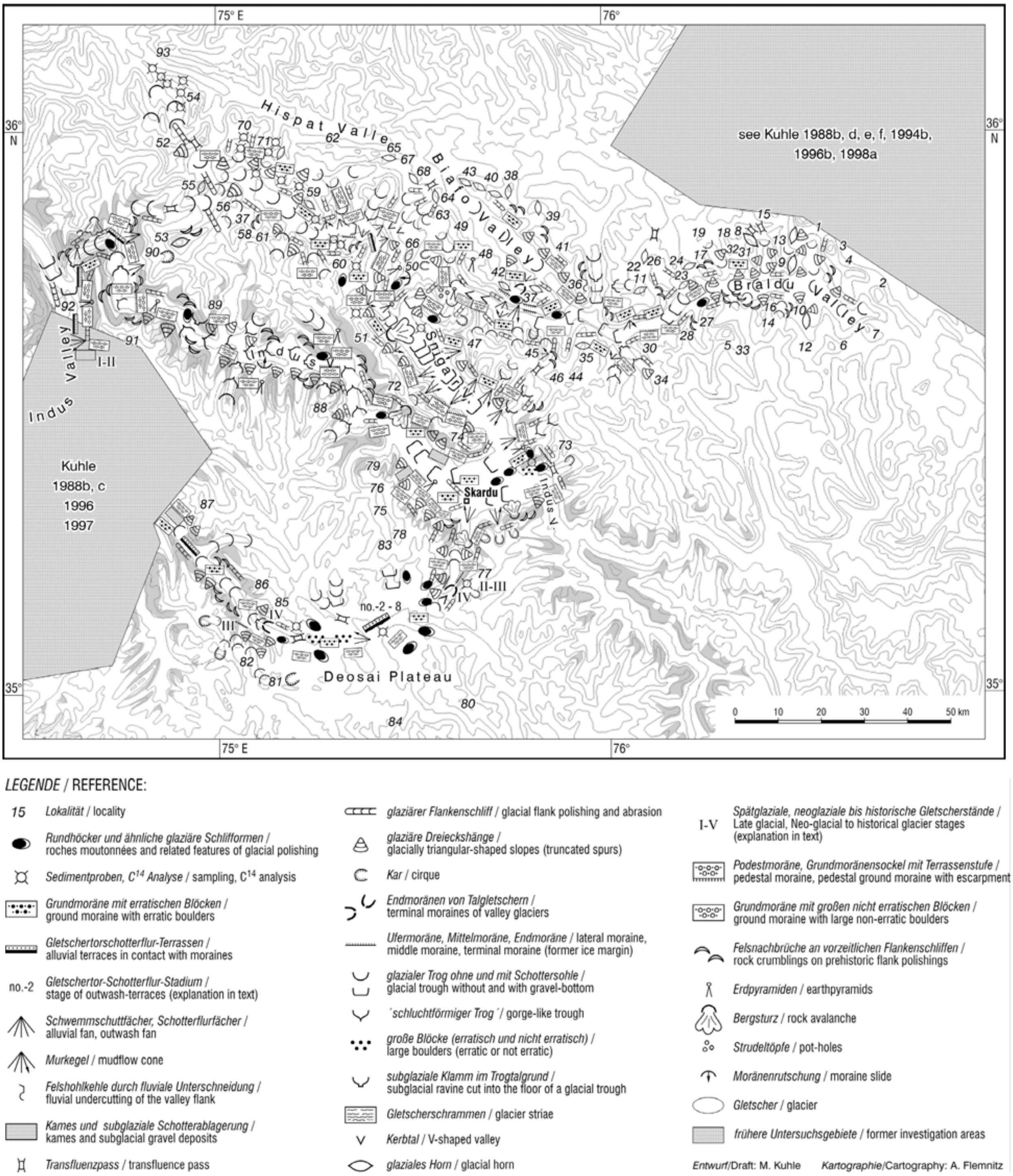

Fig.1 Quaternary-geological and glacio-gemorphological map of the Central-Karakorum and Deosai plateau between K2 and Nanga Parbat.

to find geo-morphological and sedimentological indicators of a past glaciation and its extension during the last glacial period.

The current climate in the southern slope of the Karakoram is subtropic-semiarid. From 1951 to 1989 the annual precipitation in the Skardu Basin

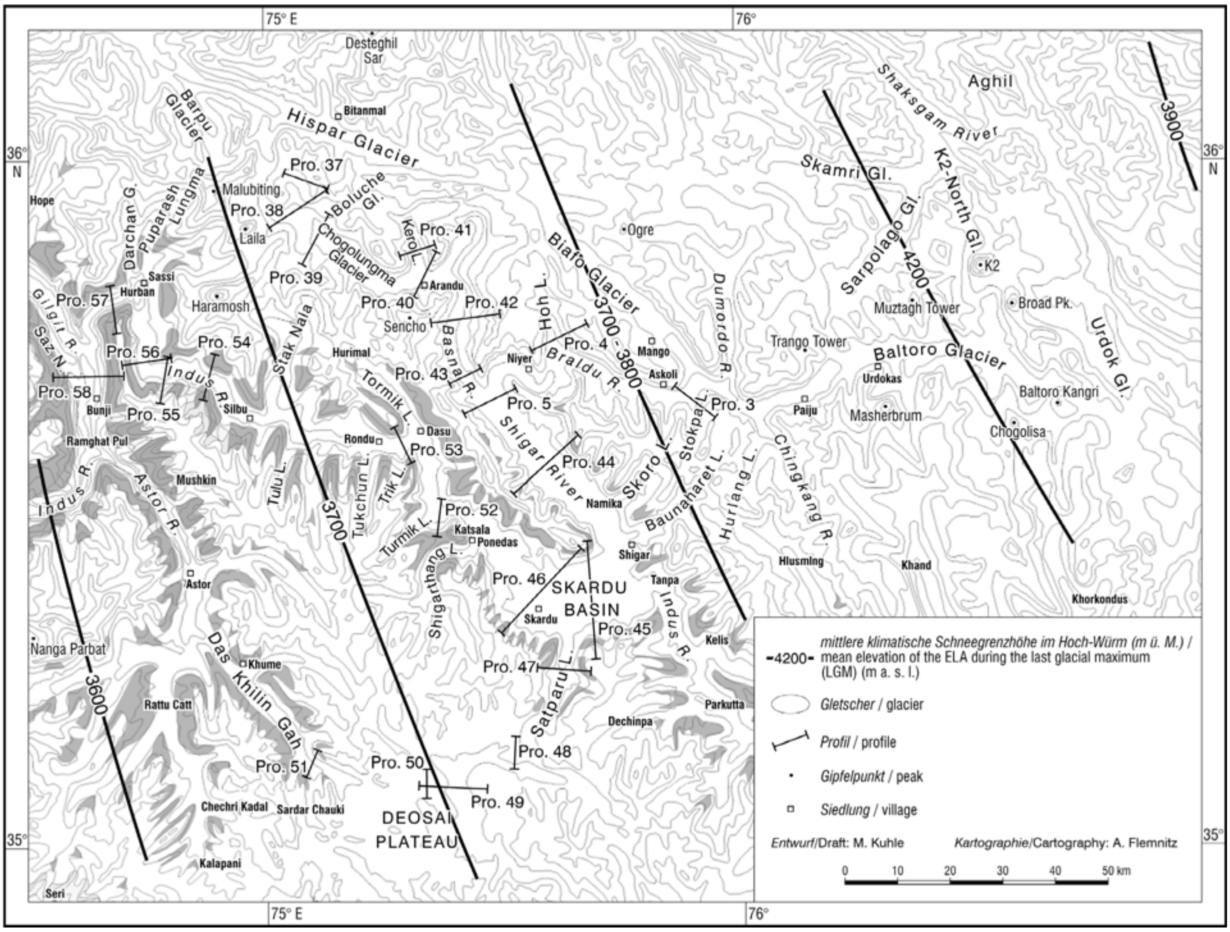

Fig. 2 The ice stream network during the maximum glaciation in Last Ice Age (LGM) c. 60-18 Ka BP and the localities of the glaciogeomorphological and -sedimentological valley cross-profiles in the Karakorum S-slope between K2 and Nanga Parbat including the Deosai plateau; see Fig.3.

on the bottom of the Indus valley at 2181 m a.s.l (Fig.2 climate station 1) reached only 208 mm/yr (Miehe *et al.* 2000). Here, in the centre of the study area, the mean annual temperature is 11.5 °C (ibid.). The lowest recorded temperature reached -21.0 °C (ibid.). On the glaciers above the current glacier equilibrium line altitude (ELA = glacier snow line) about 5000 ~ 5400 m a.s.l, precipitation increases exponentially to markedly above 1000 mm/yr. This has been shown by measurements of glacier evaporation and glacier runoff in the Karakoram. The ablation of the K2 north-glacier is 1200 ~ 1300 mm/yr, of the Skamri glacier 1300 ~ 1400 mm/yr and of the Sarpolago glacier (Fig.2) at least 1500 mm/yr (Ding 1987: 25). According to the gradient of 0.6 to 0.7 °C/100 m measured in 1986 between 3960 and 5330 m with four climate stations on the valley bottom (2 stations) and on the K2 north-glacier (2 stations), the annual mean temperature at the ELA (5300 m a.s.l) is -10.1 to -12.3 °C (Fig.2 climate stations 2~5). This corresponds to the glacier ice temperature plus 1 °C at an ice depth over 10 m and indicates semiarid, cold glaciers (Kuhle 1988a: 414). Xie (1987), however, measured –6 °C for ice at 8 m depth at the same locality (6. 10. 86). This is thus only an approximate confirmation of the author's calculation and points to a somewhat greater humidity.

Today the area is still heavily glaciated. Despite its subtropical position about 35°N, the largest extra-arctic and extra-subarctic glaciers do occur there. The Baltoro glacier is 60 km long (Fig.2).

Fig. 3

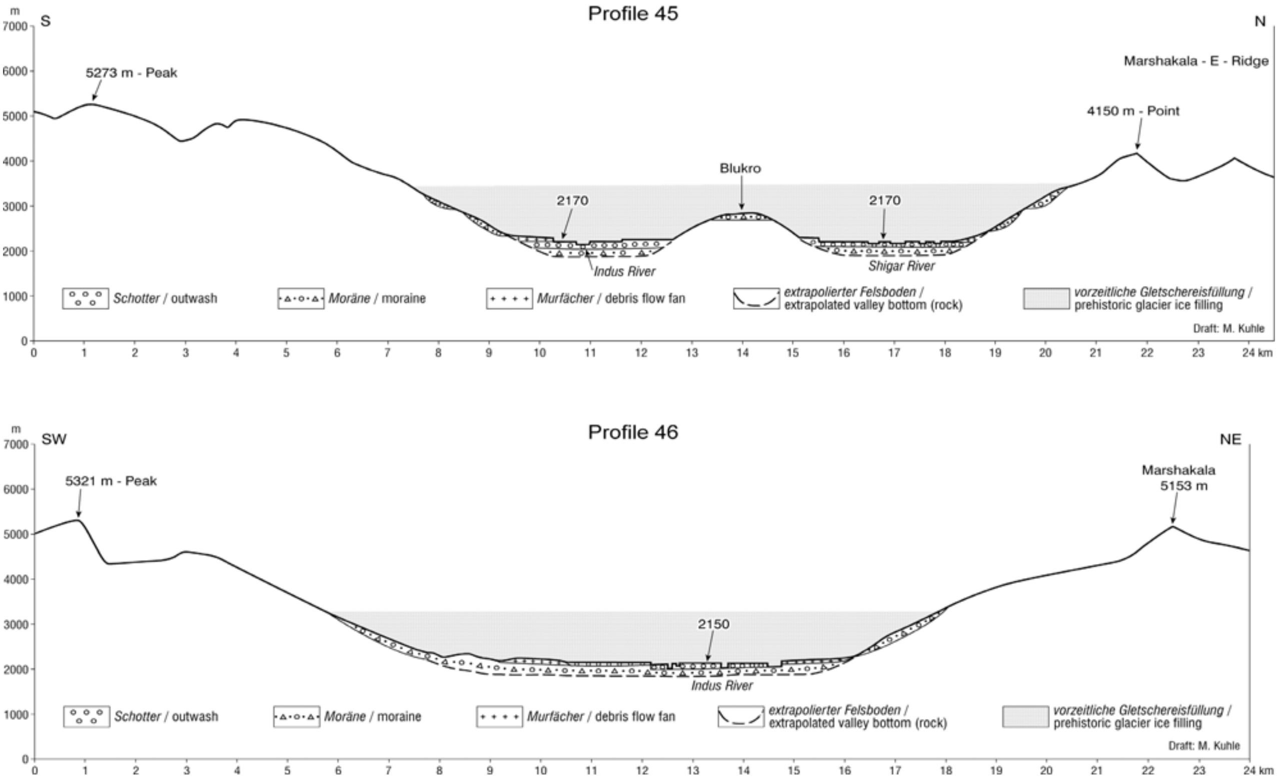

Fig. 3 Two representative glaciogeomorphological and -sedimentological valley cross-profiles through the Indus Valley (Skardu Basin):

Profile 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, Fig.1 No.77) and the Marshakala-E-ridge (Fig.1 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 Profile 45; Fig.1 between No.77 and 74.

Profile 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 (Fig.1 No.75) and the Marshakala (Fig.1 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 Profile 46; Fig.1 between No.75 and 74.

As a connected system the Biafo-Hispar glacier extends (Fig.1 No.67, Fig.2) over 74 km on the Hispar pass. The 38 km long Chogolungma glacier reaches down farthest (Fig.2). It comes to an end in the Basna valley at 2900 m a.s.l (Fig.1 between No.59 and 60). For the Aletsch glacier in the Alps the accumulation area ratio (AAR) - which indicates the ratio of the glacier feeding - to the ablation area - is 0.66, i.e. the feeding area is twice the size of the ablation area. In the Karakoram, however, this ratio is reverse. Here, the AARs are c. 0.4 ~ 0.3. The Baltoro glacier, for instance, with an

AAR of 0.3, maintains a stable position of the tongue end despite this poor situation of nourishing. But it is impossible that it could come into being under these bad conditions. Accordingly, especially the largest valley glaciers of the investigation area seem to have got their ice masses from significantly cooler periods of the ending Late Glacial. Supporting for this stability of preservation are the large-scale, metres-thick surface moraines which are a protection against ablation over decakilometres.

2 Methods

Glaciogeomorphologic observations in the research area have been mapped in detail. Locations of typologically unambiguous individual phenomena, i.e., glacier indicators, have been recorded with the help of 31 signatures (Fig.1). In this study the significance of “glacial striae” (Fig.1), “roches moutonnées and related features of glacial polishings” like glacially streamlined hills or “glacial flank polishings and abrasions” lies in the fact that they can be interpreted as obviously glacigenic erosional forms. Corresponding to the height of their position in the relief, i.e. up the valley flanks or even on mountain ridges or “transfluence passes” between two valleys (Fig.1), they provide evidence of a glacier filling up to the corresponding glacier level. This applies also to “glacially triangular-shaped slopes (truncated spurs)” (Fig.1). Here, segments of valley flanks between inflowing side valleys are concerned which have been polished back glacigenically. At the same time these are elements of the larger form “glacial trough” (Fig.1). According to roughnesses above, as e.g. wall gorges and gullies and a “polish or abrasion line” which separates the smooth and the rough rock face, the glacier trimline is recognizable. Corresponding observations can be made with the help of the trough cross profile. Where its concave course comes to an end in an upward direction, the minimum altitude of the glacial trimline is reached. “Glacial horns” (Fig.1) belong to the same glacigenic forms of erosion as troughs. These are summits between two trough valleys which have received their significant steepness by glacigenic undercutting and back-polishing on two sides. They indicate a mountain relief filled by glacier ice very far upward and thus are evidence of past ice stream networks. In many places “rock crumblings on past flank polishings” as well as “rock avalanches” (Fig.1) suggest past glacigenic forms, because they are gravitational mass movements on flanks which have been oversteepened by glacier polishing, i.e. “trough valley flanks”. The typically postglacial reshaping of “glacial troughs” by “fluvial undercutting of the valley flanks” (Fig.1) is a further factor which prepares “rock crumblings on past flank polishings” and “rock avalanches”. The wide spread occurrence of these secondary features of erosion proves the development from a glacigenic relief of a trough valley of the last glacial period to an interglacial V-shaped valley relief. Owing to this it is a secondary characteristic of the past glacial relief. “Gorge-like troughs” (Fig.1) are modifications of “glacial troughs” at a steep incline of the valley bottom. Here, the tractive forces within the glacier are predominant so that the ground scouring predominates the flank polishing. “Subglacial ravines cut into the floor of a glacial trough” (Fig.1) are indicators of the course of trough valleys below the snow-line, because they have been additionally shaped by subglacial meltwater erosion. Along the deepest valley courses they have been developed by High Glacial activities and - when the snow-line increased - by Late Glacial activities along higher valley courses. In the Karakoram the latter applies also to “cirques” (Fig.1), because during the maximum of the last glacial period the snow-line ran too low there, so that cirques could not be formed. Instead, an ice stream network developed, composed of valley glaciers (Fig.1, 2). In positions high above fluvial talwegs or on rock slopes without a present-day fluvial catchment area “pot-holes” (Fig.1) can be a hard indicator of a past ice filling of the relief supra- or subglacially canalizing the water needed for their development.

Accumulative glacier indicators which may variformly overlie the glacial-erosive features have been observed as “ground moraine with erratic boulders” (Fig.1, 3) and “ground moraine with large non-erratic boulders” (Fig.1, 3). The higher their position was above the valley bottom, the greater was the former ice thickness (Fig.3 profile 45 and 46). The same applies to “large boulders (erratic or non-erratic)” (Fig.1). In addition, their edged, facetted or rounded form has been taken in consideration as an indicator of the way and the distance of their transport. Typical of the Karakoram and its wealth of debris is the preservation of decametres- to over 100 m thick "pedestal moraines, pedestal ground moraines with escarpment" (Fig.1). They are the result of many thick layers of ground moraine, one on top of the other, which have been deposited step by step during the melting process. "Earthpyramids" (Fig.1) are characteristic residual forms of lodgement till. They are considered evidence of recent fluvial dissection of former glacigenic accumulations. At the same time they make understandable that the preservation of these past forms must be incomplete. This applies also to "moraine slide", "debris flow cone" and "alluvial fan, outwash fan" (Fig.1, 3, profile 45 and 46), because they concern redeposited moraine material which originally lay higher up the valley flanks. "Kames and subglacial gravel deposits" (Fig.1) are accumulations which are intercalated into lateral moraines. In the investigation area they have developed during the post-high glacial decay of ice. "Alluvial terraces in contact with moraines" are forms of glacifluvial dissection of older sanders caused by further regression of the glacier mouth. These sander faces, i.e. outwash accumulations, form the postglacial trough valley bottoms in the investigation area ("glacial trough with gravel bottom") (Fig.1) so that box-profiles result (Fig.3). The postglacial "rock avalanches" (see above) (Fig.1) deposit their accumulations on the gravel floor of valleys as well as on moraine accumulations.

Representative samples have been taken for laboratory analyses (Fig.1) in order to provide a detailed diagnosis and to show the actual occurrence of lodgement till in those topographical positions which suggest former glacier trimlines at high altitudes. Data of 30 moraine samples were analysed. The sediment analyses: C/N - determination (Elementar Analyser Leco CHN 1000), lime content determination, grain size analysis, determination of the sorting coefficient in the matrix spectrum (Engelhardt 1973) and morphoscopic quartz grain analysis (Mahaney 1991) support and complete the proof of a huge former glacigenic landscape. The morphoscopic quartz grain analysis doesn't make it immediately possible to recognize the material as "glacially crushed or freshly weathered". But the petrographical analysis in the field, i.e. the content of erratic material and partly also the lime content of the debris cover, indicates that the material is glacially crushed and not freshly weathered in situ. The sorting coefficient S_o (= $\sqrt{Q_3 / Q_1}$) considers the ratio of the grain sizes of the first quarter Q_1 of the grain size distribution curve to that of the third quarter Q_3. It provides further evidence as to the differentiation of fluvial and morainic accumulations. If only one grain size exists in the sediment, then $S_o = 1$. The greater the coefficient is, the more intensively mixed are the different grain sizes. This is the characteristic of moraine matrix. Accordingly, the minor C-portion, the bi/trimodal grain size distribution and the lack in sorting as well as the very high percentage of glacially crushed quartz grains provide evidence of lodgement till even up to very high positions in this steep valley relief. Thus, these analyses are further accumulation indicators of the former ice cover and glacier thickness and - in some places - even of the minimum altitudes of the trimline.

"Terminal moraines of valley glaciers" and "lateral moraine, middle moraine, terminal moraine (former ice margin) " are accumulative indicators of the ablation areas. In the investigation area (Fig.1, 2), left by the joint glacier terminal of the ice stream network had left during the maximum glaciation of the last glacial period, they are mainly evidence of Late Glacial glacier margins far below the highest trimlines (Fig.1) which had already decreased. Therefore they are marked by the numbers I ~ IV as being of late- to neoglacial age (Fig.1). This is not the place to discuss the late- and neoglacial periods in detail. However, the existence of these younger indicators is important, because they render the differentiation of seven late- to neoglacial and at least four historical glacier positions possible. This number is approximately corresponding to that of glacier positions of the post-last-glacial period diagnosed worldwide and, owing to this, may be an important indication as to our correct dating.

The catalogue of signatures applied has been developed for the map 1 : 1 million (ONC G-7, 1974). The locations of the sediment samples are marked in Figure 1. All 92 type localities supporting the results of this study are numbered. They concern areas where the arrangement of the position of indicators provides unambiguous evidence of a glacier cover. Within the scope of this paper it is impossible to provide a detailed description of the indicators in the particular topographical connection of these 92 type localities. Owing to this, only the most significant glaciogeomorphological phenomena are shown, which due to their position in the valley relief are especially hard indicators. In accordance with the scale of Figure 1 they are strongly generalized. References to them are given in the text, the figures and the table. In addition to this large-scale map and the recording of type localities - which do not only occur on the valley floors but partly also on remote slopes and mountain flanks - 25 geomorphologic profiles, mainly valley cross-sections, distributed over the entire investigation area, have been recorded (Fig.2). The two profiles selected to provide evidence of the large-scale Ice Age glaciation (Fig.3) are shown in Figure 2. Combined with the two-dimensional, vertical representations in Figure 1 and 3 they give an impression of the three-dimensional arrangement of the indicators which leads to the reconstructed glaciated area on the one hand and the local ice thickness on the other. The aim of this is a spatial proof-system based on the arrangement of the positions, so that the profiles complete the glaciogeomorphologic map especially with regard to the indication of past glacier thicknesses. Apart from that, the proved ice thickness provides local references as to the horizontal extension of the glacier cover. All indicators marked in the maps and profiles have been documented on the spot by photos in a medium-sized format. In this study four of these photos are shown as exemplary proofs.

3 The Glaciogeomorphologically Reconstructed Trimlines and Glacier Thicknesses of the Past Glacier Network

In addition to the reconstruction of the maximum extent of the glacier cover of the last glacial period (= Last Ice Age = Stage 0, after Kuhle 1998: 82 Tab.1), field investigations and laboratory analyses were focused on the determination of trimlines and glacier thicknesses.

A continuous past glacier network (Fig.2) as well as the extent of the Karakoram ice cover was reconstructed (Fig.1). Some representative local trimlines are recognizable in cross sections of Figure 3.

With regard to the arrangement of their positions, the erosive and accumulative indicators of the past glacier activities and their partly postglacial reshaping by rock crumblings, rock avalanches, debris flow cones and alluvial fans are compiled in a geomorphologic map (Fig.1).

In a central section of the investigation area (Fig.1 No.35) was a maximum ice thickness of the past ice stream network of 2700 m. This is proved by covers of lodgement till on the rock slopes and further above by flank abrasions up to an abrasion or polish line, i.e. past trimline at 5700 m a.s.l. The rock flank has also been polished by the glacier. The very resistant bedrock quartzite has only been slightly splintered and roughened by postglacial weathering for c. 13,000 years. For the last time the valley was filled with glacier ice during the last late glacial stage (Sirkung Stage IV = older than 12,780 YBP, after Kuhle 1994: 142/143 Tab.2). Merely the fine polish is lacking, whilst the abrasive rounding has been preserved.

In place of numerous topographically similar localities in the investigation area Figure 1 (for example No.37) shows the high glacial filling of the ice stream network of the deeply incised Karakoram relief up to a trimline at 5700 ~ 5900 m a.s.l and also the late glacial lining of the valley flanks by ground- and lateral moraines. Accordingly, the lateral moraines belong to the late glacial stages Dhampu (III) and Sirkung Stage (IV) (Kuhle 1998: 82 Tab.1). The later lateral moraine which belongs to the last (youngest) late glacial glacier position (IV) is c. 13,500 to 1,300 years of age (older than 12,780 YBP, after Kuhle 1994: 142/143 Tab.2) and underwent an ELA-depression of c. 700 m against the current course of the snow-line (cf. Fig.1 No.77).

In many places the transition from a basal mantling of ground moraine up the slope and glacigenic flank abrasions up to a polish line, i.e. glacier trimline at 5600 m can be observed. With regard to a more western section of the investigation area it occurs e.g. in the locality of Figure 1 No.59. This transition is characteristic, because the highest trimline which has been preserved by flank polishing, has developed far above the corresponding snow-line and therefore must have been formed by glacigenic abrasion without a ground moraine accumulation worth mentioning - which should have been preserved up to these days. The lower 430 altitude-metres consisted of late glacial lodgement till and lateral moraine of the Sirkung Stage IV, after Kuhle (1998:82 Tab.1).

An example of the different type of relief and its former completely covering glaciation is the Deosai Plateau (Fig. 1 between No.83 and 84). The plateau shows the typical remnants of an ice cap: wide-spread ground moraines with partly large erratic boulders far away from each other, roches moutonnées and glacially streamlined hills providing evidence of a minimum altitude of the trimline at 4800 m.

Distributed over the research area, the 25 glaciogeomorphologic valley- and glacier cross-sections taken in the field (Fig.2 Profiles 3~5, 37~58; Fig.3 Profiles 45 and 46) indicate past trimlines from 6200 m a.s.l (at the locality in Fig.2 Profile 37) down to 3100 m a.s.l (Fig.2 Profile 58). The first trimline is situated in the currently still glaciated regions of the highest summits in the Central Karakoram (Muztagh Karakoram): K2 up to Ogre (Fig.2) and the latter, lower trimline in the area of the Indus outlet glacier below the inflow of the Ice Age Gilgit glacier (Fig.2 Gilgit river). In the area of the present-day upper Baltoro glacier (Fig.2) between K2, Broad Peak, Gasherbrum IV, Baltoro Kangri and Chogolisa, the ice has even reached 6300 m a.s.l (maximum value: 6400 m a.s.l). Thus, also in the area of the current Biafo glacier (Fig.2) the glacier surface of the last glacial period has come to similar altitudes.

Correspondingly, in the areas of the present-day Chogolungma-, Biafo- and Baltoro glacier three dome-like ice culminations have existed in the glacier network of the last glacial period. These very flat ice domes can be demonstrated by the reconstructed glacier cover, height of the trimline and glacier thickness as well as by the entire Karakoram glacier network, at that time extending over c. 125,000 km^2 (Kuhle 1988b, 1998), i.e. by a basal face which was much larger than that of our study area. The ice thickness depends on the extent of an ice dome and also on the mountain relief lying underneath, because this is decisive for the ground friction by ice outflow and the degree of ice accumulation. Accordingly, the extremely high mountain relief of the Karakoram with its enormous valley depths was the cause of the ice thicknesses reconstructed. The ice domes have communicated with each other via transfluence passes, i.e. they have formed an approximately continuous ice surface without today's breaks in slope. The levelling has taken place across the past transfluence passes, as indicated by smoothly-convex, rounded longitudinal profiles and trough-shaped cross-profiles with lodgement till, partly with erratics on the bottom. The ice cover of the Deosai Plateau (Fig.1 and 2) has made up a fourth dome, from which outlet glaciers came down the valleys nearly symmetrically and in all directions (actiniformly) (Fig.2). All these trimlines with their - seen on a large scale - minor surface inclines, which are equidirectional, 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, below Profile 58). Here, too, the talwegs join, so that the glaciers of the last glacial period flowed down-valley controlled by the relief.

In the middle Braldo valley (or Braldu-, Biafo-, Blaldo Lungpa), the ice thickness was about 2600 ~ 2900 m (Fig.1 and 2, Profile 3). In the Baltoro glacier valley it was c. 2400 ~ 2900 m, in the Biafo glacier valley 2450 ~ 2950 m, in the Chogolungma glacier valley and the continuing Basna valley c. 1800 ~ 2900 m (Fig.2 Profile 37-42).

In the valley mouths the Braldu glacier was still c. 2500 m thick (Fig.2 Profile 4 and 5 immediately down the Braldu valley) and the Basna glacier c. 2600 m (Fig.2 Profile 43). All these glacier thicknesses of the main valleys of the Ice Age Karakoram lay above, i.e. up-valley of the snow-line (cf. Fig.2 cross-profile positions 3 ~ 44 with the climatic snow-line altitudes). They have still exceeded the maximum glacier thicknesses of the last glacial period of the Alpine glacier network - reaching c. 1800 ~ 2000 m - by 20 ~ 30%, and may have had a glacioisostatic effect as far as a depression about 50 ~ 100 m. In the very wide Shigar valley the ice thickness has been reduced from c. 2400 m (Fig.2 Profile 5) via 2150 m (Profile 44) to c. 1300 m at the valley mouth. This corresponds approximately to the ice thickness in the connected Skardu Basin (Fig.2, Profiles 45, 46), where it was c. 1500 m (Fig.3 Profiles 45; 46) and decreased to c. 1000 m towards the NW, i.e. the mouth of the basin. 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 increased again up to 1700 m as far as the Gilgit glacier (Fig.2 Profiles 52 ~ 57). 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 surface level at c. 3100 m a.s.l. The Indus- and the Gilgit glacier have both formed a parent glacier with a width up to 11 km (Fig.2 Profile 58), which came in contact with the Astor glacier and the Nanga Parbat glacier network (Kuhle 1997) and mediated to an ice margin at only just 800 ~ 850 m a.s.l (Kuhle 1988b).

The trimline of the Deosai plateau glacier which was c. 24 km in diametre (Fig.1 and 2), lay approximately between 4700 and 4900 m a.s.l (Fig.2 Profiles 48 ~ 50), so that ice thicknesses of c. 200 m above hill cupolas, and at most c. 900 m above the high valley bottoms, could extrapolated as being probable. The Deosai outlet glaciers through the Satparu Lungma (Fig.2 Profile 47) and Das Khilin Gah (Profile 51) have produced glacier connections to the Skardu Basin (Fig.2, 3 Profiles 45 and 46), i.e. Astor valley (Fig.2), and thus to the glacier network of the Karakoram S-slope, and at the same time to the Indus glacier network. In these terms, the Deosai glacier can be indicated as being a part of this entirety.

The altitudes of the climatic glacier snow-lines (ELA) (Fig. 2), in connection with the 25 cross-profiles of the valley glaciers, suggest that the communicating surface of the most productive glacier branches of this glacier network remain below the snow-line from the lower Shigar valley on as far as the Skardu Basin between Profile 44 and Profile 45 (cf. Fig. 2 and 3). Here, the ELA has run between c. 3700 and 3800 m a.s.l during the last glacial period. From this one can draw the conclusion that in this area the surface moraine cover has started and, due to the growing ablation process down the Indus, may have increased rapidly. A condition for this is the feeding of the glaciers by avalanches from the steep faces of the highest mountains such as K2 (8616 m), Broad Peak (8047 m), Gasherbrum IV (7980 m), Masherbrum (7821 m), Ogre (Baintha Brakk, 7285 m), Malubiting (7458 m), Haramosh (7409 m) etc., which has also taken place during the glacial periods. At that time these mountains 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 in the confluences have first met to form internal- and then medial moraines. In the middle to lower Skardu Basin (Fig.2, from Profile 46 on) these surface moraines which were made up of avalanche rock-debris and merged with medial moraines of the Central Karakoram (Muztagh Karakoram with K2, Masherbrum, Muztagh Tower and others Fig. 2), might have completely covered the Indus parent glacier. 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 moraines, at least 80% of the surface of the glacier network was 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. The result of this was a cooling and autonomous amplification resulting in the build-up and preservation of the Karakoram glacier network (Kuhle 1988c).

The current snow-line altitudes of the glaciers (equilibrium altitude = ELA) have been determined after the method of V. Klebelsberg (1948: 31) according to the condition of the glacier surface at places, where the fresh lateral moraines begin to thaw out. In the Karakoram the ELA-depressions could be appraised most immediately by arithmetical averaging of lowest altitudes of current and past snow-line positions, following V. Höfer (1879). We know from experience that this is an integral method of approximation, which is only usable for paleoclimatic estimations (cf. Kuhle 1988c, 1994). The lowest past tongue ends of the Karakoram glacier network came down to c. 800 m a.s.l (Indus parent glacier), i.e. on average c. 2600 m lower than the current valley glacier tongues. This evidences an ELA-depression of c. 1300 m.

The empirically found out ELA- (snowline-) depression of c. 1300 m suggests a drop in the average annual temperature by 7.8 to 9.1 °C during the last glacial period. Whether the present-day precipitation, which decreases to 210 ~ 140 mm/yr in the valleys, but amounts to 1200 ~ 1500 mm/yr at altitudes above 5000 m a.s.l (see above), would have been sufficient to create such an ice build-up at a higher relative humidity, or if the conditions were more hygric, cannot be estimated reliably. However, the reduced evaporation has led to the development of lakes in the currently arid Tarim Basin, Gobi Desert, Mongolia and in South-Siberia (Wünnemann *et al.* 1998). Perhaps the glacial ice build-up of the Karakoram system has also profited from their evaporation.

4 The New Results Concerning the Past Glacier Reconstruction in Relation to the Observations of Other Researchers

The author's field investigations in 1997 and 2000 suggest that a continuous glacier network existed over the entire research area (Fig.1, 2). No glacier terminus of the last glacial period could be reconstructed. Only during the late glacial period did this glacier network disperse into dendritic valley glacier complexes of still large dimensions. The High Glacial glacier network merged into the Indus Glacier and had only one lowest glacier terminus, namely that of the Indus parent glacier (Fig.1 hatched on the left below the research area concerned, Fig.2 left margin). In the course of previous fieldwork done 115 km down-valley from the 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 this tongue end of the last glacial period of the Indus glacier at c. 850 ~ 800 m a.s.l (Kuhle 1988 a~ c, 1989, 1991, 1997, 1998, a.o.).

Shroder (1989: 144) approximately follows this reconstruction, suggesting a glacier terminus 25 ~ 100 km away from the Nanga Parbat, down the Indus.

Haserodt (1989), however, strongly opposed to it. He considered the lower section of the Indus valley to have been entirely free of a valley glacier and interpreted several of the moraines as deposits from local side valley glaciers.

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 Last Glacial Maximum (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 which is 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 suggested 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 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 last glacial period indicated by the occurrence of fresh glacier striae (Kuhle 1988b, 1989, 1991) (see Fig.2 left margin). Within a few centuries or millennia this fresh striae, glacigenic rock polishes and -abrasions as well as the rock roundings become reworked beyond recognition by weathering and denudation of walls and slopes. This process is accelerated by the steep relief of the Karakoram. Accordingly, it is impossible to consider the forms as being older than 60 Ka or even to classify them as belonging to the pre-last glacial period before c. 120 Ka. Due to the very heavy denudation which takes place in the steep relief of these mountains since deglaciation, the preservation of the forms under the moraine cover since the pre-last glacial period has to be ruled out, too. The Astor glacier, which Dainelli (ibid.) suggested has reached the Indus, has been confirmed by the author's field investigations (1987) (Kuhle 1988b, 1997) (see Fig.2). According to their fieldwork (1934) Finsterwalder (1938: 174) and Troll (1938a, b) contradict this opinion. They pointed out that the Astor Glacier has come 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 this regionally greater context the reconstruction of the glaciation history of the Hindukush has to be mentioned, which is situated northwest-adjacent to the investigation area.

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 was ree of ice. Kamp (1999), in contrast, assumes that the past Chitral glacier has come at most 100 km further down-valley, up to the Drohse settlement at 1300 m a.s.l. According to the author's reconstruction, in the Last Glacial period the Chitral glacier has even flowed 15 km further down, i.e. up to the valley chamber of Mirkhani at 1050 ~ 1100 m a.s.l (35°28'40"N/ 71°46'30"E). This ice stream was not only thicker than 500 m, as supposed by KAMP (ibid.: 189), but three-times thicker, that mean 1500 m. Whilst Kamp (ibid.) does not assume a transfluence of the Tirich Mir glacier over the Zani pass, the author's findings of granitic erratics above the depression of the pass argue in favour of this transfluence. This gives evidence of a more important ice thickness.

As for a small part of the investigation area of this study, namely the Skardu Basin, glaciogeomorphological observations have been recorded in five papers. Besides Godwin Austen (1864), Dainelli (1922) and Norin (1925: Tavl. 5), as well as Owen (1988a, Fig. 2) and finally Haserodt (1989: 196/197) have misinterpreted the rock avalanche of Katzarah (Kachura) (Fig. 1, below No.72) at the exit of the Skardu Basin as a moraine. It is not necessary to discuss once more the interpretation of these authors, who considered the valley-obstructing, coarse-blocky accumulation of Katzarah to be an Ice Age end moraine. This has already been disproved by Hewitt (1999).

Haserodt's (1989) interpretation, too, has already been discussed in detail (Kuhle 1997). He described that the lower Indus valley (see above) and also the Indus valley below the Skardu Basin were not flowed through by an Ice Age valley glacier, but that they have only been reached by the tongue ends of side valley glaciers.

Oestreich (1906), Dainelli (1922), Norin (1925) and Shroder *et al.* (1993) has approached the Satpara Tso (lake) as being dammed-up by a terminal moraine. Dainelli and Norin classified this moraine as belonging to an ice margin position of an Ice Age outlet glacier of the Deosai plateau. Additionally, the first suggested that it belonged to his "4th Advance". After Dainelli the deposits of the "3rd Advance" of the outlet glacier were left behind in the area of the Skardu settlement. Burgisser *et al.* (1982), Owen (1988b) and Cronin (1989) interpreted the massively disturbed gravel fan and lacrustine sediments as "glaciotectonic or disturbed till", so that the conclusion of Dainelli (1922) was confirmed. This contradicts Hewitt (1999: 228 ~ 231) who not only queried the interpretation as "glaciotectonic", but also considered the accumulation damming-up the Satpara lake as being a "rock slide or rock avalanche" (Fig. 1, on the left above No. 77).

The author nearly completely agrees with Hewitt's opinion. "Nearly", because besides the correctly described rock avalanche deposits (Fig. 1, on the left above No. 77) and river terraces etc. (cf. Hewitt 1999: Fig. 7), accumulations of lodgement till 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. With regard to the arrangement of the positions, the glacigenic abrasion forms and rock roundings are also connected with them (Fig. 1, half-left above No. 77). They provide evidence of an Ice Age-ice level about 3400 m a.s.l, 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 (after Kuhle 1998: 82 Table 1 c. 14,250 ~ 13,500 YBP). Accordingly, the observation of Oestreich (1906: 77), who has described here "considerably rounded gneiss boulders below a rock avalanche talus", can also be integrated as being correct. To be precise, high-lying material of lodgement till is concerned, and has come down together with the rock avalanche(s) and survived relatively undamaged. Thus, a tributary glacier, the Satpare Lungma glacier (Fig. 2 Profile 47), has joined the Indus glacier (the ice filling of the Skardu Basin, see above Fig. 1 ~ 3, Profiles 45, 46). It still has a surface height about 3500 m in this junction. Correspondingly, there exists no "narrow gorge of the Satpua Valley" (Hewitt 1999: 228) at the exit of this valley, but rather a classic trough valley (Fig. 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(s) or rock avalanche(s).

Dainelli (1922) did not provid field observations of the Indus valley from a few kilometres away from the Skardu Basin in a downstream direction, because he did not visit this area. However, for his Mindel- and Riß-glaciation (1st and 2nd glaciation) he 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 might be seen in Dainelli's observation that the Skardu Basin had been filled and polished out by the glacier ice during these two older (earlier than last glacial period) glaciations. This assumption, however, was not based on a 1400 m thick ice-filling of the basin, as reconstructed by the author (Fig. 2 and 3, Profiles 45, 46), but on a valley bottom, which during Dainelli's 1st Quaternary glaciation was still higher and - until the end of his 2nd glaciation (pre-last glacial period or Riss glaciation) - has been polished down to the level of the present-day rock bottom below the lake sediments and gravels. He described and mapped moraine-like deposits on the two "riegels" (bar mountains) Blukro (cf. Fig.3 Profile 45) and Karpochi (Fig. 1 and 2 Skardu Basin) as well as on the basin flanks c. 500 ~ 1000 m above the valley bottom and classified them as belonging to the 2nd glaciation, without inferring from them the corresponding glacier thickness at that time. This might be explained by the fact that Dainelli considered the height of the trough bottom as having been much higher at the start of the 2nd glaciation than at its end, whilst he dated back the highest moraines to the start of the 2nd glaciation. A further observation of Dainelli, which was 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 were sediments confirmed by the author as lodgement tills (Fig. 1 from No. 50 via 47 up to 73, Fig. 2 Profiles 5 and 44), but which he classified as belonging to the last glacial period (Würmian) -glacier-filling (c. 60 ~ 18 Ka). The chronological classification is based on the geomorphological, i.e. relative dating method and thus on an abundance of indicators (Fig. 1) as e.g. the preservation of moraines on steep slopes (e.g. Fig.3) and the arrangement of their positions in relation to nearly unweathered glacier striae (e.g. Fig. 1, in the middle between No. 47 and 48) and related indicators of polishing as glacial flank polishing and abrasion (Fig. 1), which naturally are rare in this morphodynamically active steep relief - as well as on absolute datings (^{14}C and TL) of peaty layers in valley-blocking positions and limnic sediments on lateral- and end moraines (Kuhle 1994: 142/143, Table 2: 237/264; 1997: 100, Table 2). ELA-calculations recording the snow-line depressions according to the last glacial period and late glacial glaciation supported the chronological classification of the glacier positions (Kuhle 1988b: 588, 1994: 260, Table 3: 266/267, 1997: 123).

Summing up, the author recognizes the preserved glacier traces as indicators (Fig. 1) of the last glacial period (c.60 ~ 18 Ka), which have completely reshaped an older glacier history (during a previous stage of uplift of the Karakoram at a different sea-level). Dainelli went so far back into the past that, due to the then inevitably very important influence of the tectonic history of uplift, which for the most part is unknown - his findings allow neither a reconstruction of the trimlines 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 (ibid.) is realistic. Chronologically it belongs to the last glacial period, but, according to the author's observations in the field, it corresponds to the glacier extension during the High Glacial Stage 0 (after Kuhle 1998. 82, Table 1, equivalents to Last Ice Age, equivalents to last glacial period Isotope Stage 2 ~ 4. all these periods are of an identical duration, that is c. 60 ~ 18 Ka B.P.) only in some areas - so on the Deosai Plateau (Fig.1, 2), whilst for the most part it applies to that during the late glacial period (Stages I ~ IV ibid.).

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 past glaciation of the Shigar valley as far as down to the Skardu Basin (Fig. 3 and 4), are to be acknowledged. Both authors have also introduced personal field observations on the Ice Age glaciation of the Braldu- and Basna valley, merging into the Shigar valley. Oestreich (ibid.) was the first researcher who recognized the complete glaciation of the Deosai plateau.

5 Conclusion: the Postglacial Rock Avalanches as a Confirmation of the Glaciation of the Last Glacial Period

The valuable reconstruction of rock avalanches in the Karakoram (Hewitt 1999), is an objection to the glaciation concept of Dainelli (1922) and the later literature (see above), which misinterpreted postglacial rock avalanches as end moraines of the last glacial- and late glacial period. Due to their geomorphologically fresh status of preservation, Hewitt (ibid.) classified the rock avalanches (Fig.1) as being Holocene. This corresponds chronologically with the crumblings of glacigenically abraded trough flanks (see e.g. Fig.1 Nos. 36, 45, on the right above 60, between 63 and 77, on the left above 77, below 51, above 91, left of 85 and 86) which can unambiguously be diagnosed. Hewitt introduced three examples of misinterpretation in our investigation area. The author completely agrees as to these isolated cases. At the same time, however, he did not recognize a contradiction to the glaciation history of the Karakoram, dicussed here (see 2. and 3.), but, in contrast, a confirmation.

According to the author's opinion, which contradicts to Dainelli, especially in the Central Karakoram, no end moraines of the last glacial period (3rd espansione) are verifiable, because the glacier network had only one joint ice margin position in the lower Indus valley. In addition, the Late Glacial end moraines were not been preserved, but were completely removed by the abundant and very erosion-effective meltwater in the forefield of the melting-back glacier network in the narrow valley receptacles. For this the forefield of the current and young-historic Biafo glacier (Fig. 1 between No.36 and 35) provides a representative example. Instead of an end moraine, a gravel floor has only been preserved there. Owing to this, the author considers the concept of the search for end moraines in the Karakoram - pursued by many researchers - as methodologically wrong. Thus he has concentrated his work on the sedimentological and geomorphological analysis of the valley flanks (Fig. 3 Profile 45, 46) and the search for ground moraines and lateral moraines (Fig. 1), as well as on the observations of glacigenic flank abrasions and polishings (Fig.1). It is mainly the reconstruction of the trimlines with the help of highest erratics, remnants of ground moraines (lodgement till), glacier striae and -polishings, abrasion roundings with upper lines of the polishing (Fig. 1), but also ends of wall gorges which enables a past glaciation in the high mountains to be estimated. The analysis of end moraines, however, is due mainly to the mountain forelands and lowlands. The glaciogeo-morphological analysis of valley cross-profiles (Fig.3) introduced here, which has led to the reconstruction of the Indus-Karakoram glacier network (Fig. 2), confirms the destabilisation of the valley flanks by the realization of glacigenically over-steepened trough valley cross-profiles (Fig. 3). Accordingly, rock avalanches as have been described by Hewitt (ibid.) confirm the 1000 to 2900 m thick glacier network as the condition for those prolific postglacial crumblings (Fig. 1 e.g. on the right of No. 51, left above 53, left below 73, left above 77). As for the NW-Karakoram Kuhle, Meiners & Iturrizaga (1998) came to a corresponding conclusion.

Andersen (2000) introduced several examples of Late- to postglacial rock avalanches from Norway, as e.g. the Gloppedalsura rock avalanche in the glacigenic Gjesdal. These can also be explained by reshaping of the Ice Age glacial landscape.

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