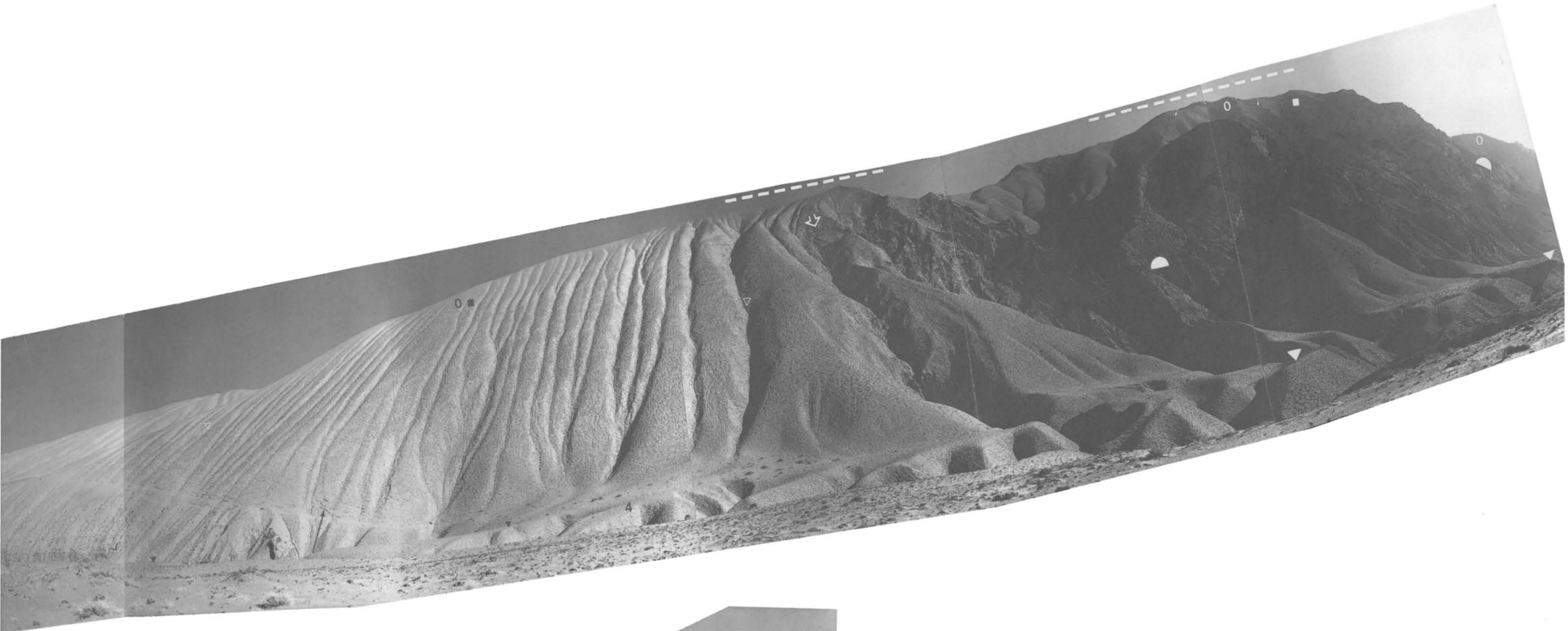


◄ Fig 116
Panoramic view across the middle Shaksgam valley, taken at 4880 m asl a little above the Aghil pass (4863 m asl; Fig 138 right-hand below No. 25) and ranging from SE (left edge of the photo) to SSW (right-hand edge); No. 6 = Apsarasas, 7245 m; Nos 5 and 8 = Teram Kangri group, 7462 m; No. 3 = Sia Kangri, 7422 m; No. 4 = Urdok c. 7300 m; No. 1 = Gasherbrum I (Hidden Peak), 8068 m; No. 2 = Gasherbrum II, 8035 m or Broad Peak group, 8047 m; No. 7 = 6210 m peak, N-satellite of the Skyang-Kangri group. □ marks the Shaksgam valley floor at 4100 m asl. The glacial flank abrasions have polished the valley shoulders and glaciated knob-like rock heads in the bedrock limestone up to at least 5500 m asl (- - - - left and right). - - - - (in the centre) marks the early Late Glacial glacier level at 5000–5200 m asl. The Aghil pass (where the photo was taken) was consequently overflowed by a 600 m-thick Main Ice Age glacier arm which moved NW towards the Yarkand ice stream. The high peaks of the Karakorum towered 2000–2500 m above the Shaksgam ice stream network which had the Tibetan inland ice flow-off towards the NW. Viewpoint: 36°11'N/76°37'E. Photo: M. Kuhle 20.10.86.

Fig 115 ▲
View from 1480 m asl towards W across the varied talus sanders (drift floors) of the last (Würm; 5) and the last but one (Riß; 6) Ice Age (○) (Fig 138 above No. 45). Laid down by an anastomosing system of channels in the form of a fan extending of many tens of kilometres (1–2° inclination) during the cold periods (cf. Fig 55 Nos 2 and 8; Fig 56 [17.8.86/2]; Fig 134), these drifts are now dissected by "microfluviatile channels" (▽) in the wake of heavy rains. This almost barren pebble desert (Serir) in the Tarim basin is part of the Takla Makan desert, and shows the formation of deflation pavement and ventifacts on the larger stones. In the background offset, neogenic sediments from the area of the Tarim basin boundary fault are visible. Burdened by the loose sediments of the surrounding high mountains, isostatic down-warping of the basin begins here. Viewpoint: 37°46'N/77°26'E. Photo: M. Kuhle 17.8.86.

◀ Fig 118
Erratic block of glimmerite granite or gneiss originating from the transfluence pass of the Shaksgam-Muztagh valleys at 4450–4500 m asl, c. 550–600 m above the present valley floor (Fig 138 No. 12) (cf. Fig 37 ■ on the right). Facetted by glacier transport, the smaller block, which was found in otherwise dolomitic moraine material together with other, up to 1.5 m-long blocks of this kind, shows a gneiss-like banding of the light feldspar crystals on the opened-up side, as one finds in the bedrock of the main Karakorum ridge. These are blocks which have been subject to long-distance transport by the Late Glacial Shaksgam glacier. The immediate slope catchment area of the site where this block was found consists solely of calcite rock. Sample locality: 36°05'N/76°28'E.

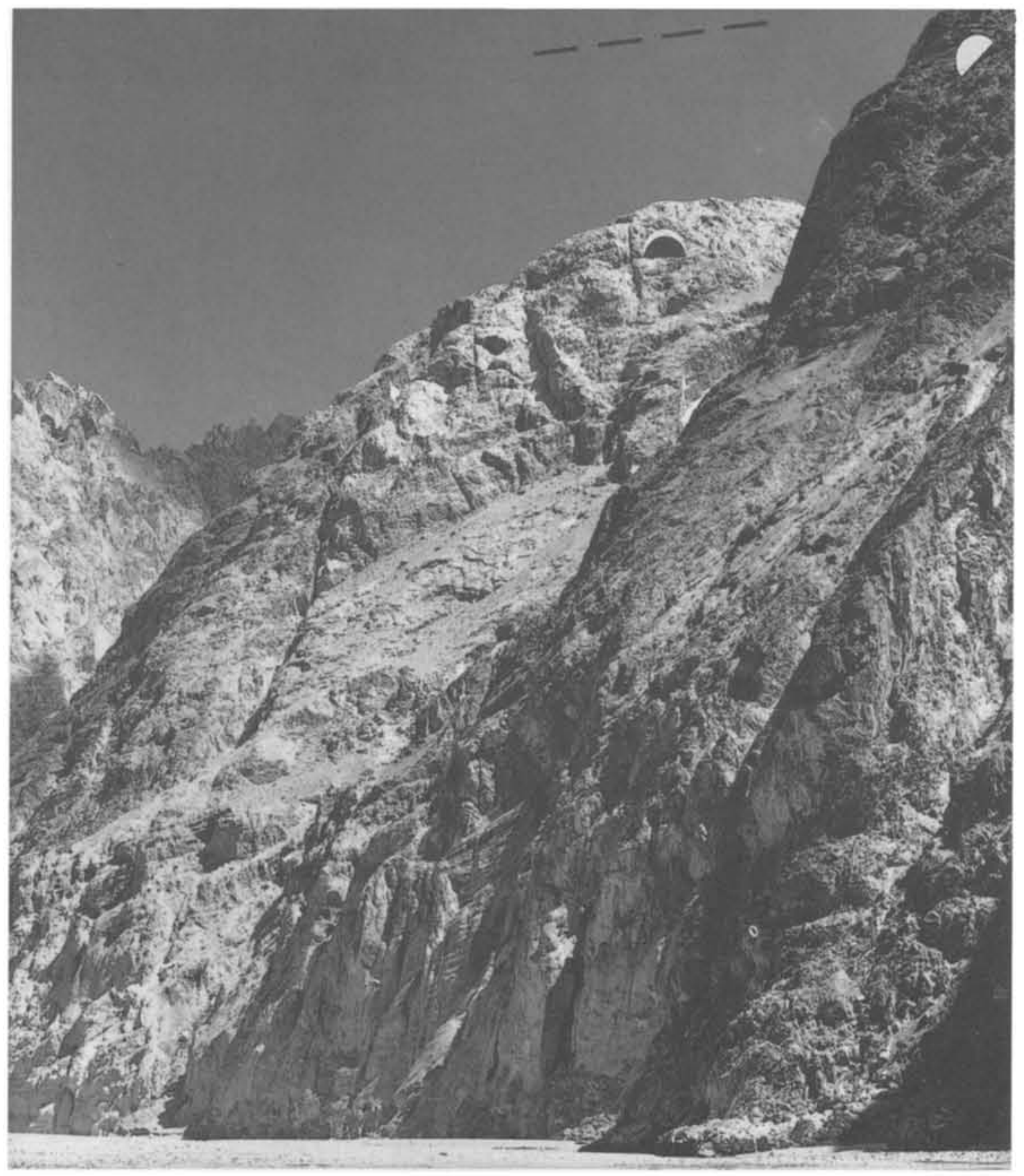


Fig 119 ▲
View from the drift-floor of the Shaksgam valley from c. 3950 m asl towards the SSE into the orographic left-hand limestone flank and up-valley (Fig 138, to the right of No. 12). This steep valley wall shows the features of a trough flank, as the glacigenically abraded and polished (◗) wall first becomes steeper from top to bottom, and subsequently changes from a convex polished profile above the drift floor to a concave one below the drift floor. The abraded and polished areas (◗) show little splintering or crumbling. The drift floor of the Shaksgam valley in its entire width is activated by seasonal floods, evidence of this being shown in its uniformly light colouring extending to the very edge of the rock flank. Viewpoint: 36°06'20"N/76°28'E. Photo: M. Kuhle 1.9.86.

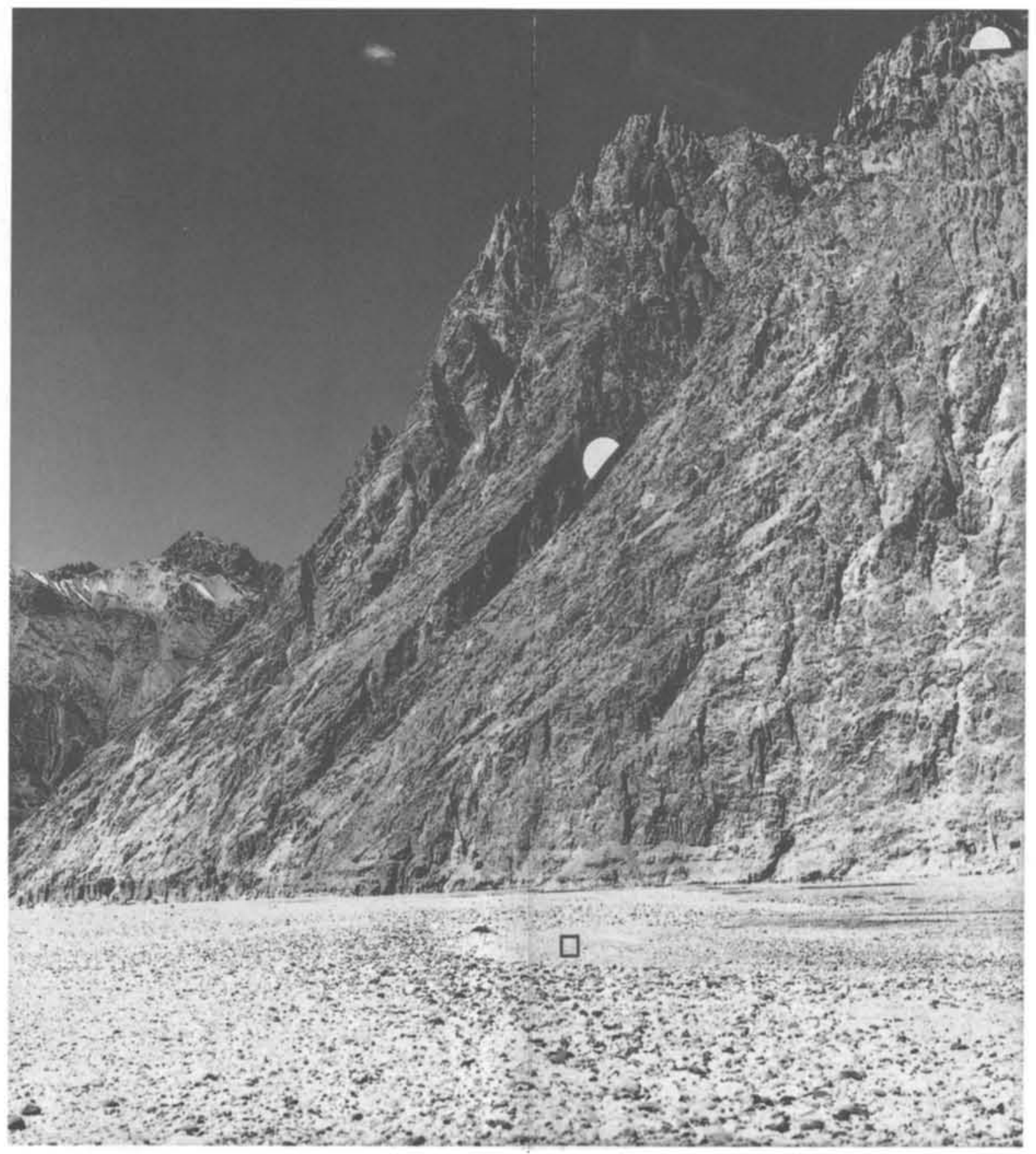


Fig 120 ▲
View from the drift floor (□) of the Shaksgam valley at 3990 m asl into the orographic right-hand trough flank down-valley towards the W (Fig 138, left below No. 21). This photo shows a Main- to Late Glacial glacier scarp, which follows a marked bend of the valley to the left. Hitting this bend at an obtuse angle, the glacier ice, thus diverted, has scoured out a groove-like concave bow into the rock which reaches high up the wall (◗). Viewpoint: 36°07'N/76°32'E. Photo: M. Kuhle 1.9.86.

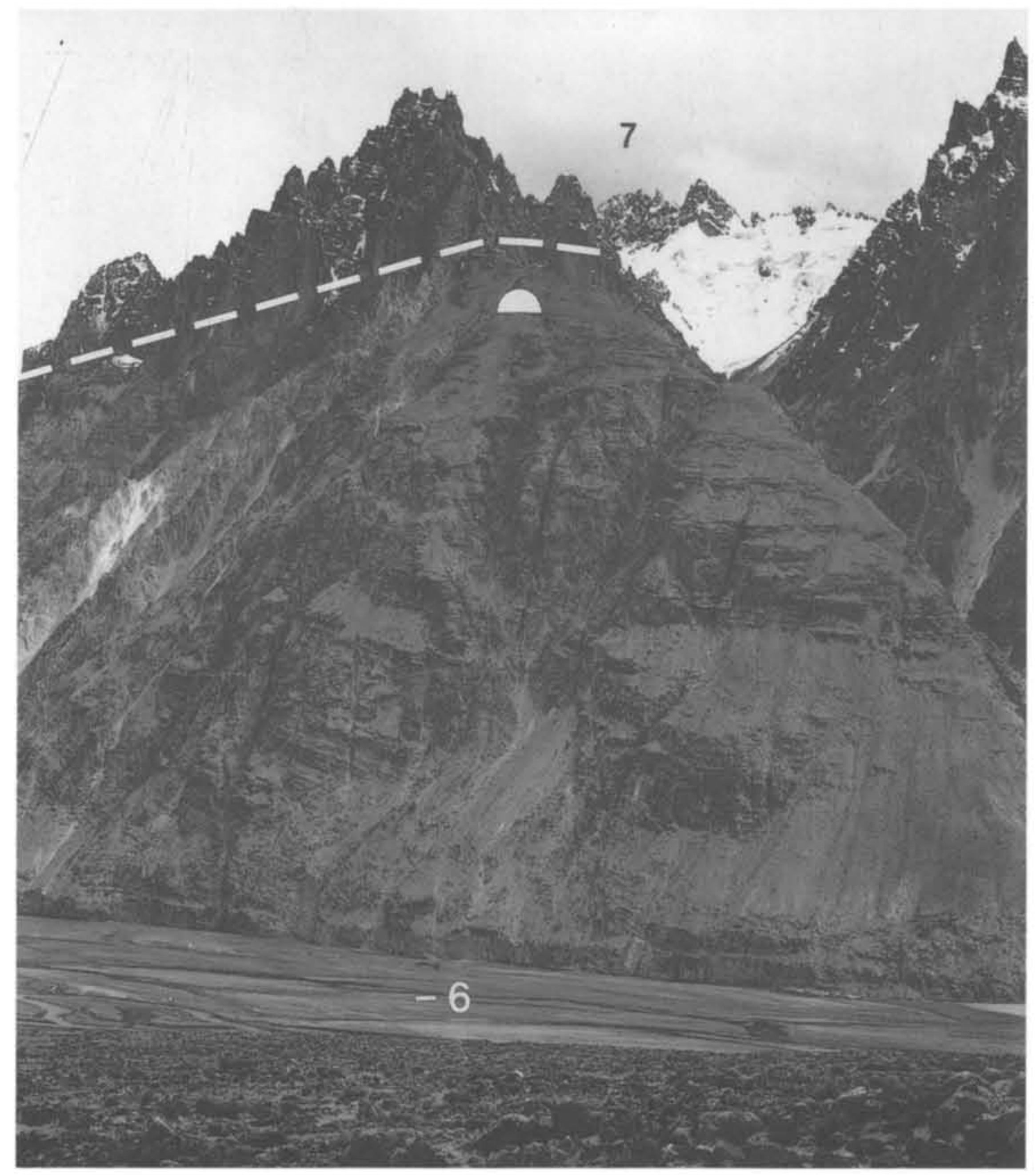


Fig 121 ▲
View from a mudflow cone (with granite blocks in the foreground) at 4180 m asl into the orographic left-hand flank of the Shaksgam valley towards the SSW (Fig 138, right of No. 23). No. 7 = summit superstructure on the 6210 m massif; -6 present drift floor of the valley. From a genetic point of view it must be regarded as a current ice cave drift floor (ie. a sander which the valley canalizes). It was deposited by glaciers (Staghar-, Urdok-, N-Gasherbrum- and S-Skyang-Lungpa glaciers) which reached the floor of the Shaksgam valley. Late Glacial flank abrasion and polishing (◓) on these outcropping edges of limestone extends to a polish line (- - - -) at an altitude of 5000–5100 m, ie. to 1000 m above the drift floor. Viewpoint: 36°08'N/76°37'45"E. Photo: M. Kuhle 29.8.86.

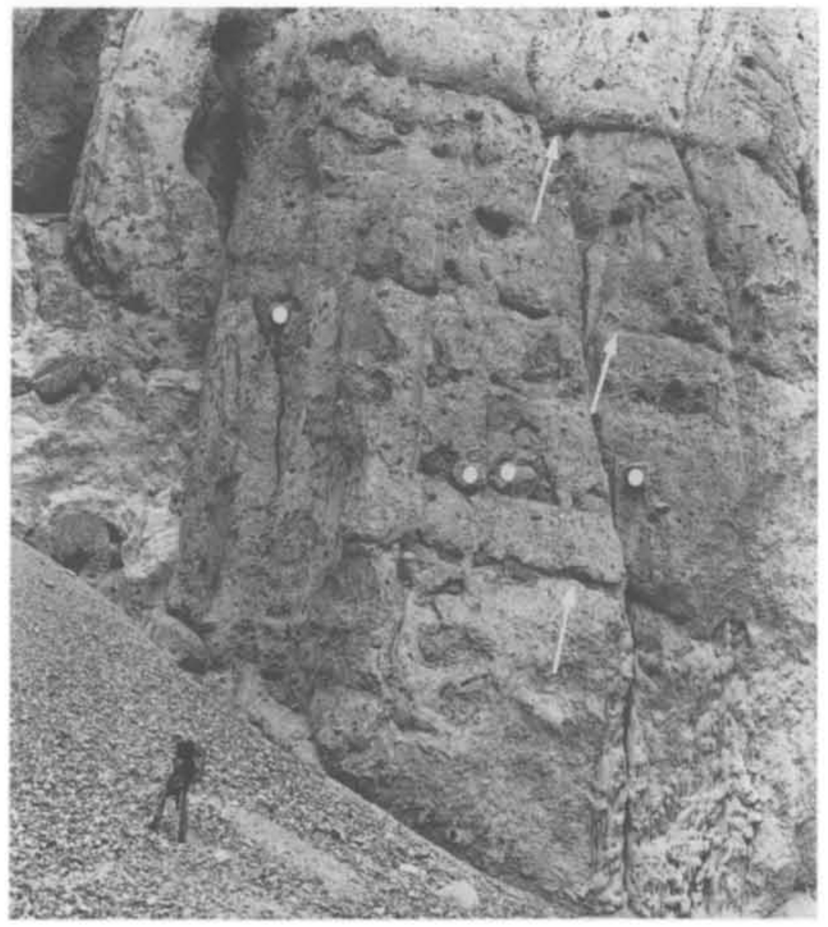


Fig 122 ▲
View from 4270 m asl into the orographic right-hand flank of the Shaksgam valley in the exit of the "southern Aghil pass-valley" on to Late Glacial ground moraine material (●). In this place it has been preserved over several hundred metres up the valley slope (see Fig 71 ◄; cf. Fig 83 ■; Fig 138 on the right above No. 23). (●●) mark large polymict blocks. They are rough-edged, rounded and facetted; isolated from one another, they "swim" in a pelite-rich matrix. Their wealth of fine material is an indication of the intense grinding by the heavy pressure exerted by a very thick and fast-flowing ice stream burden. Banding of crumbled holes and grooves in the exposed wall (↗) is indicative of banking of the ground moraine, pointing to typical strata sedimentation. For a comparison of size see the 1.80 m-tall person. Viewpoint: 36°08'20"N/76°37'43"E. Photo: M. Kuhle 20.10.86.

Fig 123 ▲
View from 4100 m asl down the middle section of the western Surukwat valley ("lower northern Aghil pass-valley") towards the N (Fig 138 left, above No. 27). No. 1 = 6094 m peak in the northern Aghil mountains; a hanging glacier tongue which, from a thermal viewpoint, belongs to the "cold" type, emanates from this glaciated mountain and flows down south (right of - - - -, background). It ends in a rough, steep drop which is characteristic for this type. (- - - -) marks the Ice Age glacier surface in the area of the orographic left-hand flank of the valley. The orographic right-hand flank abrasions (◗) have left a concavely scoured, glacigenic trough valley profile in the bedrock granite. (▽) mark present talus cones, (□) a talus and mudflow fan, and (■) a rough block moraine of granite scree rock, which has been transported by an orographic right-hand side valley glacier. Viewpoint: 36°16'20"N/76°34'45"E. Photo: M. Kuhle 21.10.86.

Fig 125 ▲
View from the orographic left-hand side valley at 3550 m asl towards the SSE up the lower Surukwat valley (Fig 138, right of No. 33). In the foreground is the mudflow-fan surface (X), which is shown in the right-hand third of Fig 126. It is covered by blocks which have been transported over short distances from the immediately adjacent valley flank (●). Their forms are accordingly rough. Below the crest ridges (No. 6) of the Aghil mountains which rise to c. 5600 m here, truncated spurs and glacial cuspate slopes (◓) set in. The polishing of their inclines and the rounded ridges above mark an Ice Age minimum glacier level (- - - -) about 1000 m above the valley floor in the confluence area of the eastern (left) and western (right) branches of the Surukwat valley. The tributary valleys (○▽) which, now free from glaciation, divide the main valley flank into these cuspate slopes, are the source of alluvial fans in the Late Glacial drift floor terrace (sander terrace ▼, 3, Taglung Stage II). Viewpoint: 36°22'N/76°40'50"E. Photo: M. Kuhle 28.8.86.

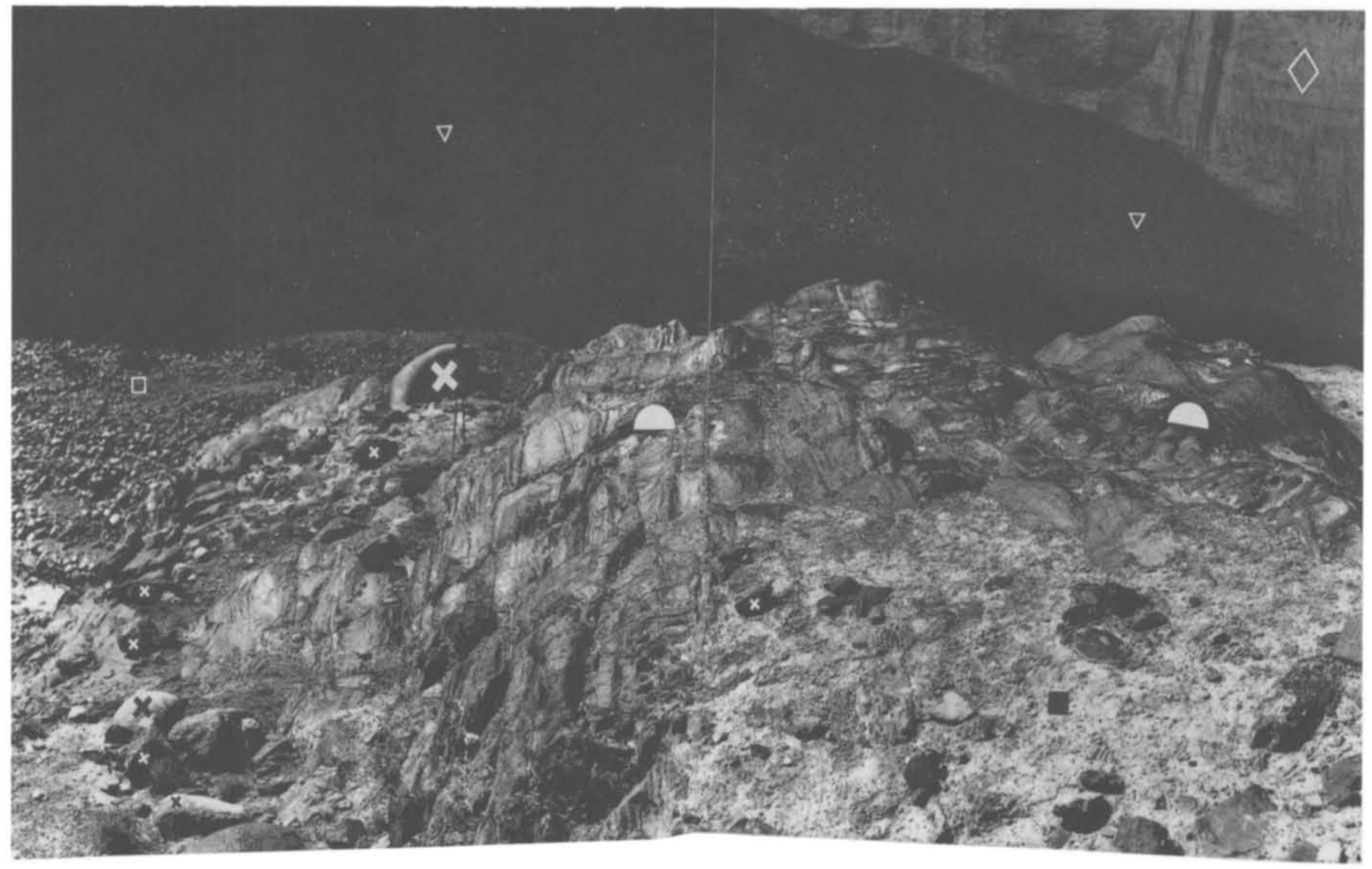


Fig 124 ▲
Glacially abraded, polished and rounded bedrock quarzite rocks (◓) in the orographic right-hand flank of the western Surukwat valley at 3700 m asl (Fig 138, No. 46). Affected by polishing and also by sub-glacial meltwater, the rock surfaces bear a ferro-manganese crust. In parts they continue to be covered by boulder clay (■) and more or less rounded polymict blocks (X). (□) marks the high-water bed of the present Surukwat river; (▽) shows fresh talus slopes, which have accumulated from material crumbling down the walls of glacifluvial drift terraces (♦). Viewpoint: 36°18'N/76°35'15"E. Photo: M. Kuhle 29.8.86.

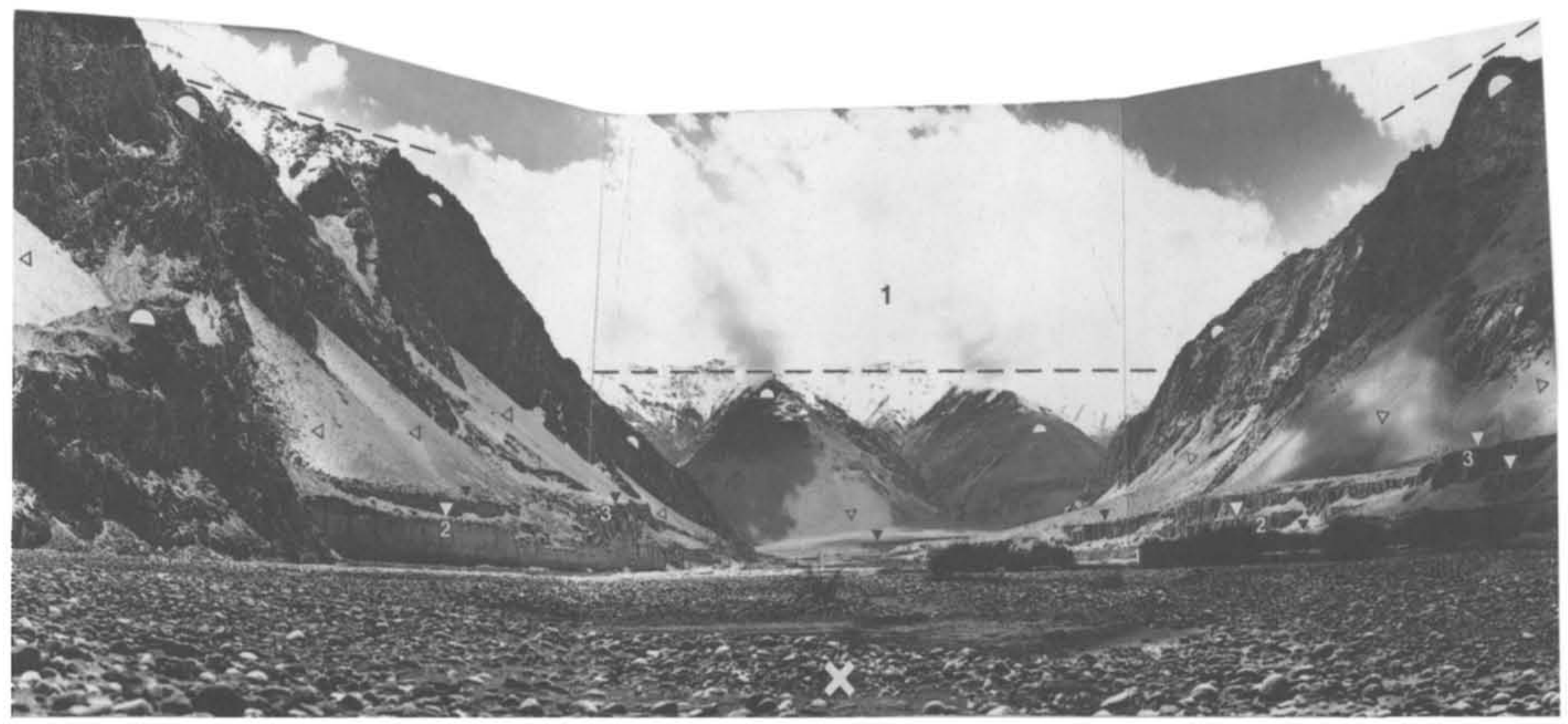


Fig 126 ▲
View from the present drift floor of the lower Surukwat valley (X) at 3450 m asl (Fig 138, between Nos. 33 and 49), facing S towards the Aghil mountains (near No. 1 = 5250 m-high crest ridge). - - - - mark the surface of the prehistoric ice stream network. Glacially abraded up to more than 1000 m above the valley floor, the valley flanks (◖◗) of bedrock metamorphic rocks (phyllites) have been roughened by crumbling since deglaciation, and frost-induced talus slopes (▷▷) as well as detrital embankment, followed by terrace formation (▼ 2,3) have flattened and stepped its cross-profile. Viewpoint: 36°23'10"N/76°41'E. Photo: M. Kuhle 28.8.86.

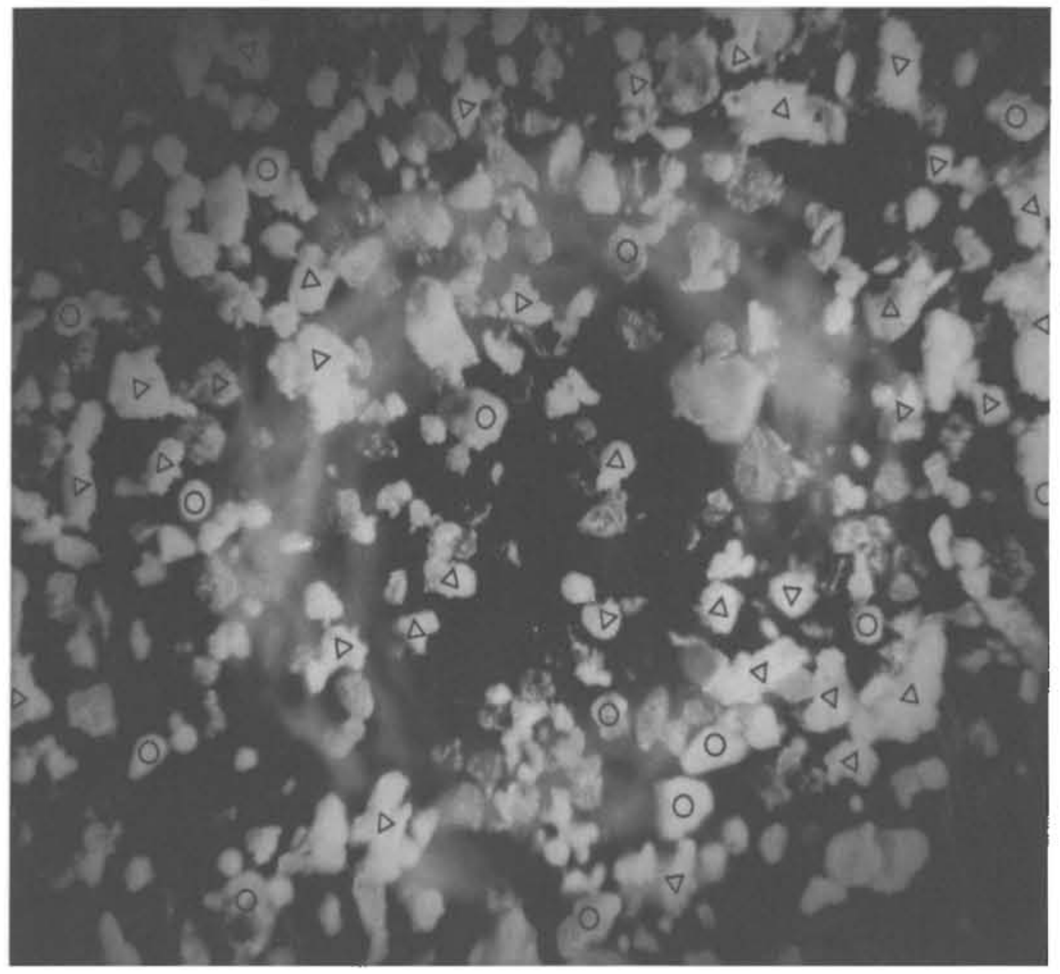
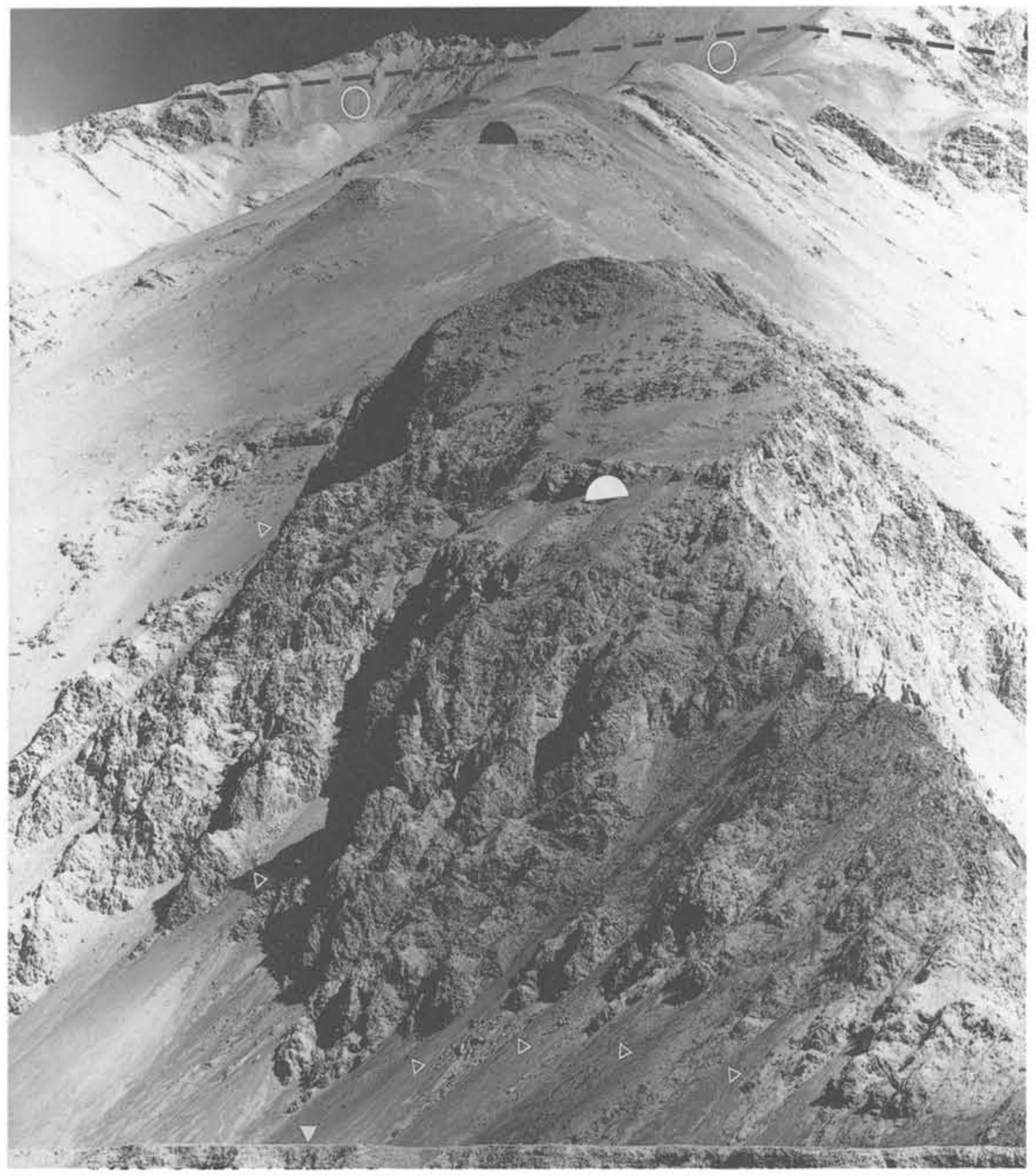


Fig 127 ▲
Microscope picture of the grains >63 μm from sample 17. 8. 86/1. The sample was taken from the major cone sander in the northern foreland of the Kuenlun mountains at 1480 m asl (Fig 138 ¤ right-hand above No. 45). Cf. further analysis information in Fig 55 (curves 2 and 8), Fig 66 (curves 5 and 8), Fig 56 (17. 8. 86/2) and Fig 115 (○, 5 and 6). The morphoscopic analysis (Fig 56) (surface texture) of the quartz grain surfaces shows 75% dull aeolian grains (▽). The fact that, in spite of a glacifluvial distance of many tens of kilometres, the grains had to be transported, only about 15% of them appear to be fluvially lustrous (○) is evidence of the substantial aeolian, ie. here also cold-arid, syngenetic transformation of the SiO_2 grains.

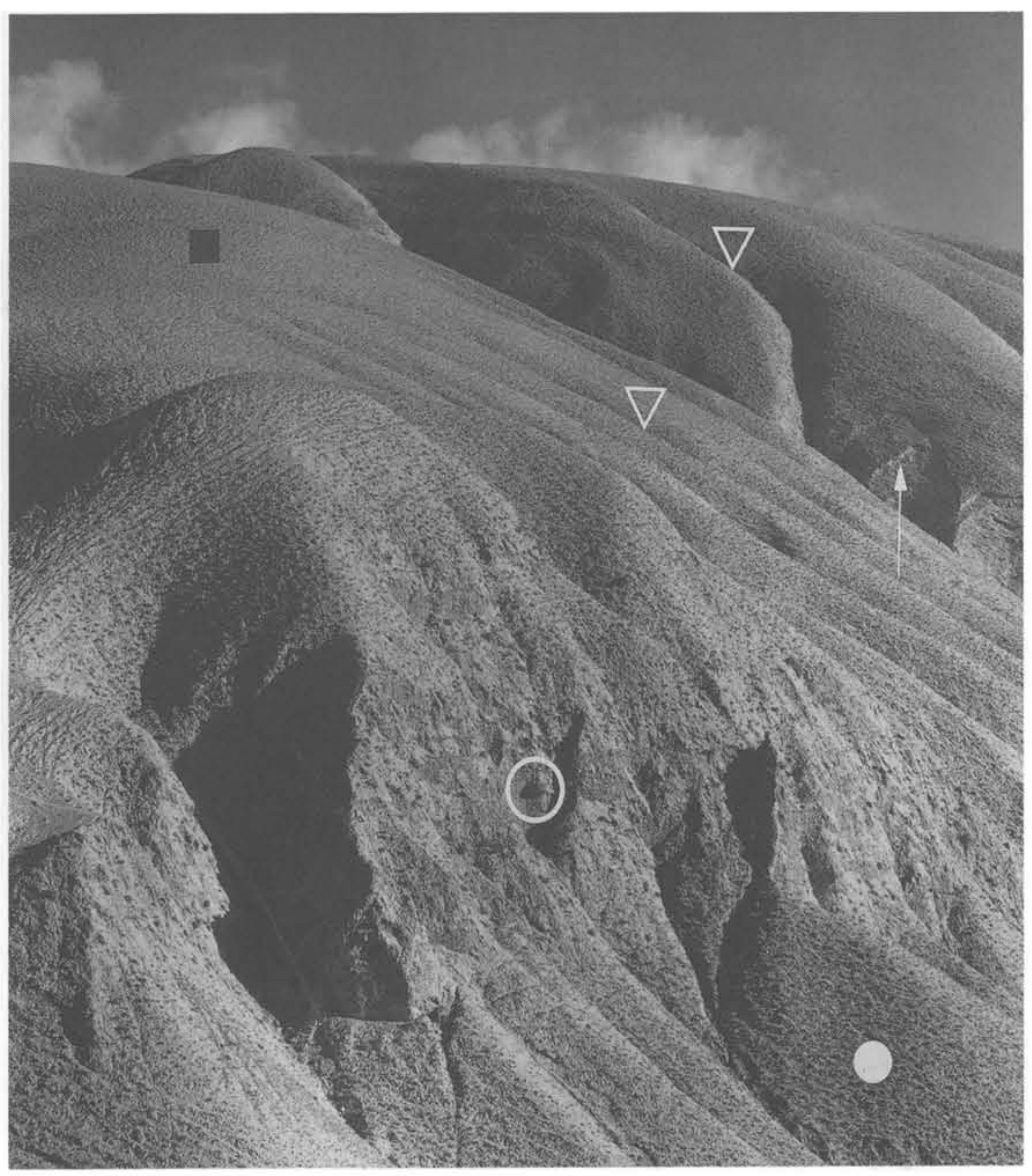
Fig 128 ▲
Main to Late Ice Age glaciated knobs (◓) with preserved polishing (↓) in vertical pelitic metamorphites (phyllites). These glaciated knobs are in the confluence area of the Surukwat valley and the Yarkand valley (Fig 138, No. 33). Their surface is about 3580 m asl, approximately 100 m above the present gravel floor of the two valley bottoms (cf. Figs 99 and 42). Though shallow, the fluvially-formed plunge pools constitute evidence of the contribution to the rock surface formation made by sub-glacial meltwater (↓). The surface of the glaciated knobs continues to be partially covered by ground moraine (∪). It has conserved the slightly frost-weathered rock surface since deglaciation, and prevented, or at least reduced, its frost-splintering. Late Glacial glacier outlet gravel-fields set these glaciated knobs into sediments from the bottom up (□). Accordingly, younger, ie. post-Glacial or Holocene debris and mudflow cones (▽), are in turn set into the present surface of these gravel-fields. Viewpoint: 36°23'N/76°41'E, facing E. Photo: M. Kuhle 27. 8. 86.

Fig 130 ▲
View from a rock wall fan (X) above the valley floor at 3800 m asl, looking SW into the orographic left-hand flank of the Yarkand valley (Fig 138, No. 37). Late Glacial moraines (■) lie on the exposed rock base of more or less horizontally-bedded sedimentary rocks up to 700 m above the valley bottom. They are topped by rock heads (◥) with abraded flanks, which have been formed above the snow line during the Main Ice Age. At that time the level of the ice stream network (- - - -) was at 5000 m asl, ie. 1200 m above the valley floor. Compare the levels of moraine and polish lines on the opposite (right) valley flanks shown in Figs 60 and 106. A snow-cut gorge (⬠) dissects the confluence step, which forms the junction with an orographic left-hand side valley (hanging valley) from the 5880 m massif (Aghil). ▼▼ mark Late Glacial glacifluvial drift ledges. Viewpoint. 36°24'N/76°20'35"E. Photo: M. Kuhle 26. 8. 86.

◀ Fig 129
View from the area of the mouth of this orographic right-hand side valley of the Yarkand (Fig 138 left-hand, below No. 50) facing NNW up this side valley. No. 1 = 6136 m peak in the valley catchment area (Kuenlun). The valley floor is situated at 3750 m asl. Above it on the right there are two terrace levels (▼▼ centre of the photo), which are the work of two generations (X) of mur fans. Their substratum contains large, rough-edged, smooth-edged, and also rounded blocks (●●), which point to a formation from dislocated local moraine from the immediately adjacent small, second-order side valley (its junction may be found on the right-hand edge of the photo). (□) marks a rock base topped by glacifluvial drift material (▼) and kame-like moraine material (■). Above the talus slopes (▷▷), which mask the foot of the slopes, and are part of the present development of periglacial slopes of frost compensation, remnants of Ice Age flank abrasions and polishings (◥◤) are preserved in the crystalline bedrock slates (phyllites). (◓) on the left is a transfluence pass, leading over from the main (Yarkand) valley. (- - - -) marks an early Late Glacial polish line at 4750 m asl. Viewpoint: 36°24'25"N/76°48'E near Bazar Dara. Photo: M. Kuhle 27. 8. 86.



◄ Fig 131
View from the Mazar military station at 3800 m asl into the orographic left-hand flank of the Yarkand valley, towards the SE (Fig 138, below No. 39). At the bottom the terrace edge (▼) of a Late Glacial, glacifluvial drift floor (of a valley sander) can be seen. Formation and preservation of Ice Age flank abrasions and polishings (◓) varies according to the rock fabric: softer strata higher up experienced much more rigorous back-abrasion and back-polishing (below the upper ◓) than the layer edges of relatively resistent metamorphites at the base (◓ ■ below). Only at the harder strata associations do the slopes become steeper again (◓ above). ○○ mark Late Glacial corrie forms. Viewpoint: 36°26'10"N/77°01'E. Photo: M. Kuhle 21.8.86.



◄ Fig 133
For locality see Fig 132. View of the orographic right-hand end moraine of the "Pusha moraine valley" in the northern Kuenlun foreland, facing S up-valley (Fig 138, centre, between Nos. 43 and 44). The moraine ridges (■) reach relative heights of 400–700 m. The time of their formation must be described as polyglacial, since outlet glacier tongues, which repeatedly reached the Tarim basin in the course of several Pleistocene ice ages, contributed to it. The moraines consist of polymict, partly rounded and facetted blocks (of limestone, phyllite, crystalline slates), in part of significant dimensions (○). Isolated from one another, these blocks "swim" in a fine matrix. The moraine ridges carry a primary layer of loess, which is dissected by gully washings (▽). On particularly steep gully slopes the loess slips off in the form of more than one metre-thick "loess boards" (↑). At the gully exits at the foot of the slope, the down-washed secondary loess from higher up is sedimented in the form of shallow cones (●). Viewpoint: 37°18'N/77°07'E. Photo: M. Kuhle 30.10.86.

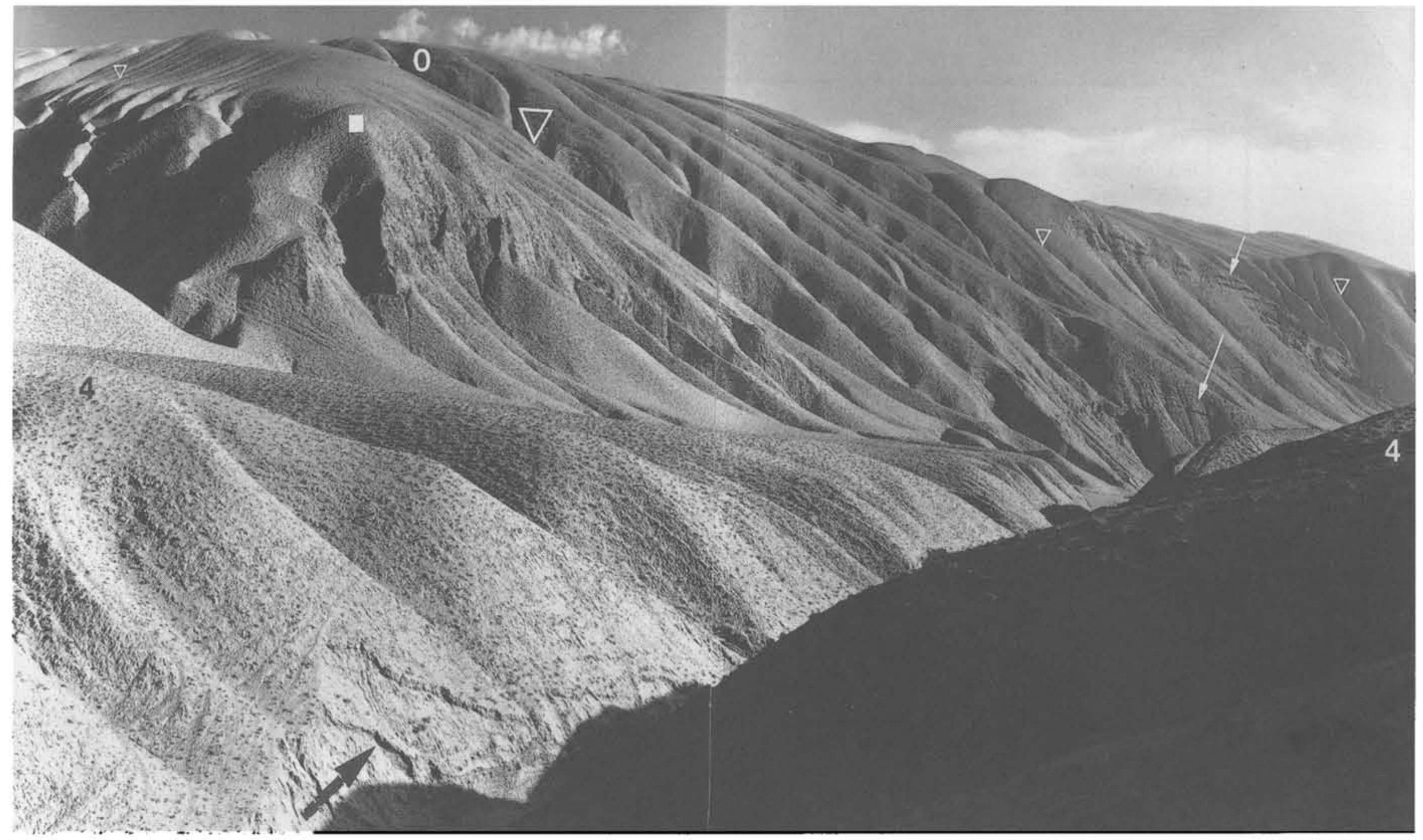


Fig 132 ▲
View from the floor of the "Pusha moraine valley" (cf. Figs 112 and 113) at 2000 m asl, facing S up-valley to the orographic right-hand lateral moraine (or median moraine, which is also the end moraine here) (Fig 138, centre, between Nos. 43 and 44). The point where the photograph was taken is on the glacifluvial ice cave drift floor terrace No. 4 of the Late Ice Age (Ghasa Stage I) (terrace of a sander which became canalized in the moraine valley). (■) marks the moraine of the Late Ice Age (Main Ice Age) (No. 0), which has been cut into by arid gully washing (▽▽). (For details see Fig 133). The sedimentation of these end moarines from the far-reaching W-Tibetan outlet glaciers, and those from the Kuenlun which advanced as far as the Tarim basin, took place with significant involvement of glacier meltwater, so that in many places kames from glacifluvial drift are incorporated into the moraine (↙) (see Fig 134). Viewpoint: 37°17'N/77°08'10"E. Photo: M. Kuhle 30.10.86.

Fig 134 ►
View from c. 2000 m asl into the orographic right-hand flank of the "Pusha moraine valley" in the northern foreland of the Kuenlun, facing SSW (Fig 138, between Nos. 43 and 44). Late to Post-glacial backward erosion during and since deglaciation has cut small and very steeply flanked side valleys (small gorges) into this moraine valley flank (■) which had been formed by moraine material. They allow the diamictite glacial material to be analysed (macroscopically) in the exposure detail. Several hundred metres high, the moraines are built up of polymict blocks, which had been deposited in a matrix of fine materials (○) (cf. Fig 133 ○ and ongoing text). In some places conditions for deposits result in a banking. Even interspersed glacifluvial (kame-like) material of drift and sand, which is sorted and stratified (↓), has been exposed. (↑) marks a typical break in the loess cover on top of the moraine, which allows a morphological insight into the thickness of particular primary loess strata. Viewpoint: 37°17'N/77°08'10"E. Photo: M. Kuhle 30.10.86.

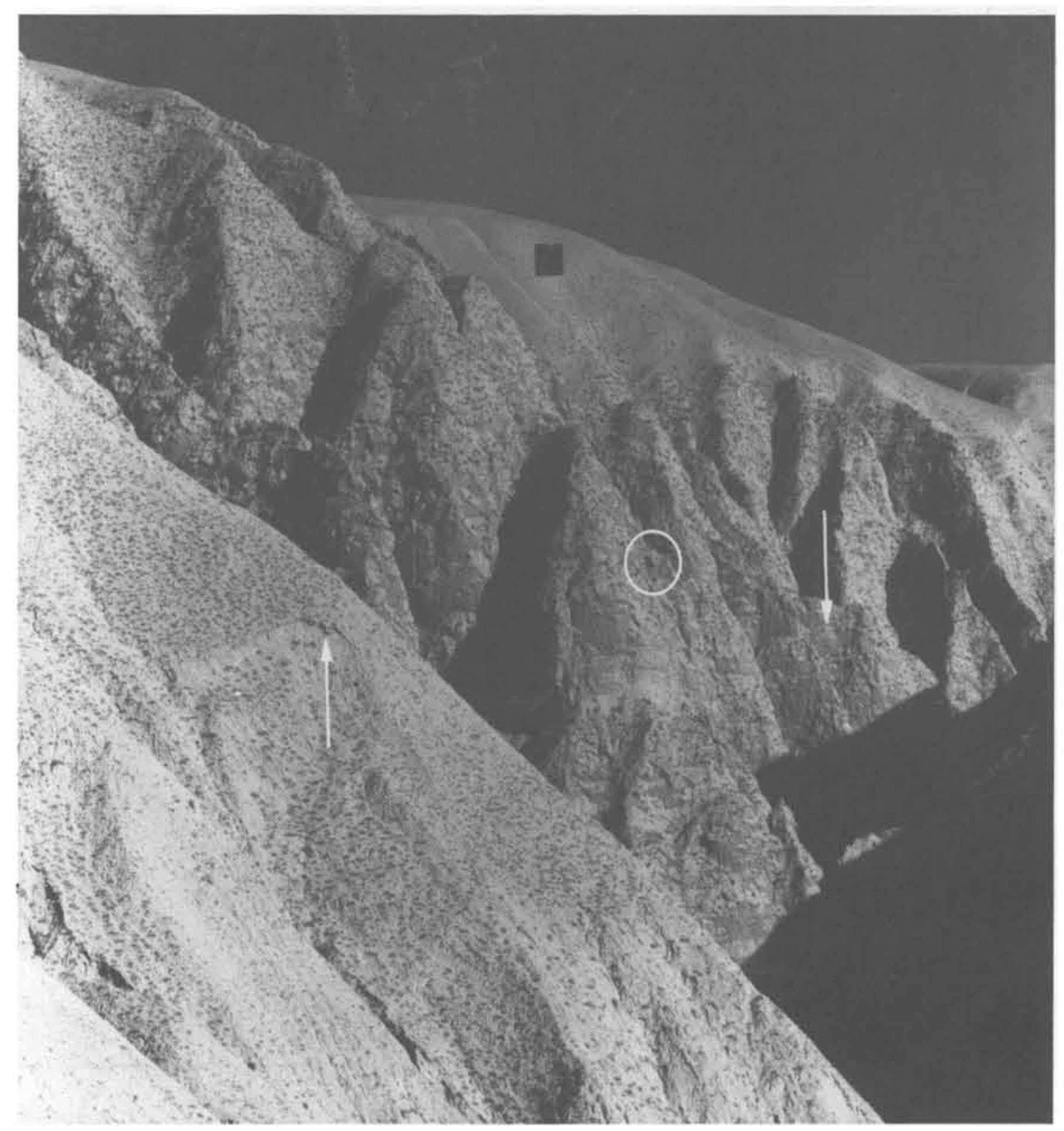


Fig 138 Glacio-geomorphological indicators for the reconstruction of the maximum prehistoric glacier cover in the area under investigation by the 1986 expedition between the main Karakorum ridge and the Tarim basin. Design: M. Kuhle.

LEGENDE/ **REFERENCE:**

15 *Lokalität*/locality

Rundhöcker und ähnliche glaziäre Schliffformen/ roches montonnées and related features of glacial polishing

Sedimentproben, C14 Analyse)/sampling, C14 analyses

Grundmoräne mit erratischen Blöcken/ basal till with erratics

Gletschertorschotterflur-Terrassen/ alluvial terraces in contact with moraines

no.-2 *Gletschertor-Schotterflur-Stadium*/ stages 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

mittlere klimatische Schneegrenzhöhe im Hoch-Würm/ mean elevation of the ELA during the last glacial maximum (m a. s. l.)

Sattel, Pass/saddle, pass

Transfluenzpaß/transfluence pass

glaziärer Flankenschliff/glacial flank polishing

glaziäre Dreieckshänge/glacial valley-slopes (triangular shape)

Kar/cirque

Endmoränen von Talgletschern im Gebirge/ 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)/ big blocks (erratic or not erratic)

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

Gletscherschrammen/glacier striae

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)

Gipfelpunkt/peak

Fluß/river

Gletscher/glacier

Firn und Eis/firn and ice

Entwurf/Draft: *M. Kuhle* *Kartographie*/Cartography: *A. Flemnitz*

0 5 10 15 20 25 30 km

Tab 2, 20.8.86/2), also at 3740 m asl. The "Vale of Kudi", though a petrographically comparable granite catchment area, also shows a *predominance of coarse silt* of about 46%, which is similar to that in other deposits (cf. above), and due to the coarse-grained granite parent material. The essential difference consists, however, in the proportion of 20% of clay, indicating *still water sedimentation.* This sediment was deposited more than 1610 ±90 YBP ago (Tab 2, 20.8.86/2) - this being the age of the fluvial valley floor. Thanks to the presence of glaciers in the former and the present catchment areas, it is possible to interpret these gravel deposits as prehistoric "direct" and contemporary "indirect" *outwash gravel fields* (Kuhle 1983, pp. 336–339) (cf. Fig 58 □; 59 □).

4.4 Observations on the "6532 m-Massif" with Regard to Categorical Late Glacial Glacier Traces and Moraine Deposits in the Kuenlun S-Slope (Fig 138 No. 18; Main Peak: 36°27'N/76°52'E).

Fig 60 shows the 5610 m mountain spur (No. 2) south of the 6532 m main peak. The main peak is approached by the icy rock ridge further north (No. 1). Behind the ridge the most significant glacier flows down S from the glaciated mountain group, which currently covers a glaciated area of c. 8 x 8 km^2. Nourished by two small source branches, this 6 km-long glacier (measured along the main branch) now ends at about 5000 m asl. Three of the six more than 6000 m-high peaks of this massif are among its on average 5900-high catchment area. These figures allow an orographic snow line at 5450 m asl to be calculated for the Kuenlun S-slope. About 2 km down-valley from the 1986 glacier end a more than 150 m-deep, about 1 km-long, *glaciofluvial gorge* in monolithic granite sets in (Fig 61). Its talweg contour terminates down-valley at 4100 m asl. Above the glaciofluvial gorge there is a hanging valley bottom, which has been widened by glacigenic scouring; on its rocky floor coarse boulders are deposited almost to the gorge edge. The valley flanks rising from this floor beyond the glaciofluvial gorge incision have been smoothed and shaped in the way of glacigenically polished and abraded surfaces up to the crest at 4700 m. The valley bottom, into which the gorge has been cut by sub-glacial meltwaters, terminates at c. 4250 m asl, thus forming a 150 to 170 m-high *confluence step* to the main valley, namely the Yarkand. Adjacent to this steep step there are two *Late Glacial lateral moraine generations* (Fig 62 ■ IV, III; 138, No. 18) both 3.1 km away from the recent glacier tongue. The highest lateral moraine deposits occurring on both valley flanks reach up to 4340 m (Sirkung Stage IV). The older lateral moraine (Dhampu Stage III) sets in at 500 m down valley. Its glacier attained the *confluence* with the *main Yarkand valley glacier,* which continued to exist at that time as documented by terraces of lateral moraine on the orographic right at the corresponding level of the Yarkand valley (Fig 62 III, left). The moraines of tributary valley exits, moreover, show that the *late Late Glacial* glacier of the Sirkung Stage (IV)

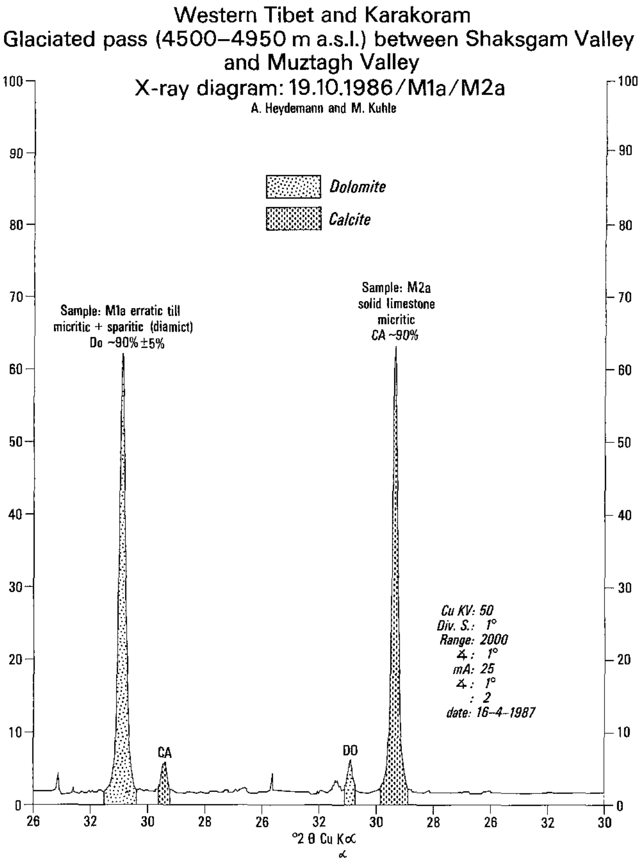


Fig 117 The erratic dolomite blocks and the dolomite scree (left peak) are mixed with up to 1.5 m-long mica-granite blocks (Fig 118) from the upper catchment area of the Shaksgam valley. The erratic dolomites lie on top of bedrock calcites (right-hand peak). The calcite rocks have glacigenically polished surfaces and the form of glaciated knobs (Fig 38 ▲). They are on the 4500 m-high Shaksgam-Muztagh valley transfluence pass (Fig 37 ▲ right, and Fig 51 ▲ right). Locality of samples: 36°05'N/ 76°28'E; Fig 138, No. 12).

extended down to 3750 m in the main valley floor (Fig 62 IV, left; 63 IV). The position of the ice margin, which had met with an already *glacier-free* main valley, is evidence of a *snow line depression* of c. 625 m (present glacier tongue end at 5000 m asl, the prehistoric one at 3750 m asl; difference 1250 m. 1250 : 2 = 625 m). This is an orographic ELA depression to c. 4800 m asl, and almost *half* the amount that can be proved for the *Main Ice Age* snow line depression (cf. Chapt. 6). The almost 200 m-high outcrop shown in Fig 62 (IV, on the right) presents the characteristic *banking structure* (■) of the *lateral moraine,* with a coarse layering in the upper parts, indicating the increasing *influence of meltwater* (▼). There is reason to believe that a *sub-glacial* incision of the gorge (Fig 61) coincided with the building of these Late Glacial moraines, since both gorge and moraines were situated more than 400 m *below* the snow line of the time. The lowest moraine section of the ice margin of Stage IV (Fig 63) shows the compact coarse boulder packing *without* layering or even banking; this is typical of *end moraine sedimentation,* ie of a deposit *without* any remaining major *glacier movement impulse.* Fig 63 (●) depicts large moraine boulders lying close to, or just a little above, the stream bed. The granite blocks are slightly *weathered* and are coated with *iron-manganese crusts,* with decimetre-deep finely weathered, yellowish-loam detritus in between. The incrustation with iron-manganese takes several hundred to several thousand years to develop. This is an indication that during the Holocene, re-depositing of material near the stream bed, or alternatively, by mud flows, must have been of *minimal importance.*

Another valley to be presented as belonging to the "6532 m massif", and draining S into the Yarkand valley, is the parallel valley to the E (Fig 138 No. 19). It has the form of a "gorge-shaped trough" (cf. Kuhle 1983, p. 155; 1991 b, pp. 1–8). Fig 64, taken at a distance of 3–4 km from the present glacier end, depicts the upper section of this increasingly - upward - narrowing gorge through which the Late Ice Age glacier had flowed. In 1986 the tongue was at 4850 m asl. Remnants of *neo-Glacial to late Late Glacial* moraines are found down to that level where the gorge passage begins at 4200 m asl. In the gorge passage itself *remnants of ground moraine* are preserved in situ in some cornices in the walls and in wall gorges (Fig 64 ■ and ▽ (right)). They consist largely of local granitic, coarse angular boulder debris. Glacigenic and slightly concave scouring of the gorge walls appears up to many hundreds of metres above the valley floor; this is the main feature of the *"gorge-shaped trough form".* Fig 65 shows the lower part of the gorge reaching down to c. 3800 m asl. Here, too, the valley is too *narrow* to be able to preserve end moraines indicating Late Glacial ice margin positions. There are none down to the trough bottom of the Yarkand valley (Fig 65, background), although in the W-parallel valley such moraines are preserved almost intact (cf. above). The ground floor within the gorge passage is some metres to decametres wide (Fig 64 □; 65 ●). It shows a representative mixture of present and Holocene *gravel floor (outwash) deposits.* During their passage through the narrow gorge they become channelized and receive fresh detritus from the autochthonous talus slopes (Fig 64 ◁ left; 65 ◁). Fig 66 Nos. 2 and 7, both show the cumulative grain size graph of the glacio-fluvial sediments with large *quantities of moraine.* The considerable proportions of fine sand and silt are attributable to the properties of the granite parent material. These samples, together with those from Fig 67/a and /b were taken from *glacio-fluvial* terrace formations 1.3 to 3 m above the present stream (Fig 138 No. 38), which are already showing more or less developed reddish-brown traces of weathering. The clay peak of sample 25.10.86/3 (Fig 67/b) is evidence of a proportion of a *fine moraine material matrix* purely fluvial material does *not* have. The datings listed in Tab 2 (25.10.86/1 and /3) are evidence of the recent age of this body of glacier mouth outwash plain of the "indirect outwash" (sander) type. The characteristic morphometry of fine grains can be gathered from Fig 56

(24.8.86/5) and 68: the considerable proportion (37%) of fresh and glacigenic transformed material with fractures and chattermarks on the grain surfaces indicates *glacier transport* and the supply of *fresh debris* in the gorge, whereas the dominant influence (63%) of *aeolian* processes confirms the aridity and absence of a meadow vegetation cover: the lower proportion of merely 2% of fluvial rounded and polished fine grains demonstrates the short distance they are transported (less than 10 km in this instance), which is typical of glacier mouth outwash plains (for sample localities see Fig 65 ↓ and 105 ↓).

4.5 The Post-Main Ice Age (Late Glacial and Holocene) Glaciofluvial Gravels and their Terrace Forms in the Skaksgam Valley, in the Yarkand Valley and in the Vale of Kudi (Fig 138, Nos 20–24, 30–32, 36, 40, 41 and 43).

The presentation of the following findings and data resulting from observations have *not* been drawn up as an end in themselves, but for the purpose of directing the attention to the *post-Main Ice Age* formation of outwash gravel fields. Following the presentation of the historic to Late Glacial inventory of glacial forms, this sub-chapter in a final section sets out to prepare the concluding introduction of the *maximal glacier infilling* of the mountain relief up to the northern forelands of the Kuenlun.

4.5.1 Loose rocks in the Shaksgam valley

The gravel floor of the Muztagh valley has already been described (Chapter 4.2). Four sediment samples from the level of historic to Late Glacial outwash gravel floor (Fig 36 X; 138 No. 10) on the orographic right between the junction of the Sarpo Laggo valley and the K2 valley are radiometrically dated (Tab 2, 15.10.86/1, 2, 4, 5). The samples were taken from altitudes between 3900 and 3950 m asl, and are evidence that no glacier can have been in this valley cross-profile between 730 ±100 and 12870 ±180 YBP any more. Any intermediate or subsequent over-run is *excluded,* since all the datable materials (peat and peaty-clays; Fig 74/c) are *soft loose rocks* which have been taken from *near the surface,* from a depth of only 0.2 to 1.2 m. In view of the *intensive exaration processes* (cf. Fig 43 and 50 ↓↓) set in train by the ground polishing of an ice stream several hundred metres-thick the loose rock could *not* have been *preserved.* The important conclusion to be drawn from this is that the Late Ice Age (and in any case the Main Ice Age) Muztagh glacier must already have *melted down* by 12870 YBP (cf. Chapt. 4.2). As has been mentioned above, the Muztagh glacier of Stage IV or III must have been the last one to pass this locality. This data basis *also shows* that, lying at roughly the same altitude above sea-level and drawing on glacier catchment areas at comparable altitudes, or even feeding on the prehistoric Muztagh glacier, the Shaksgam valley floor (Fig 138, No. 11) was *freed from* ice at approximately the *same time.* This is conclusive evidence that all the undisturbed loose rock material in this steep-flanked trough of the Shaksgam valley, in so far as they are not morainic but of glaciofluvial origin, must have been deposited after about 12870 YBP, and are thus *very young* (cf. Fig 27 □; 37 □ -6, 38 □; 39 ■ -6; 51 □ -6; 52 □; 52 a □; 69 ◆; 70 -6; 71 □ X ▽; 72 -2-6; 73 □; 75 □; 76 □; 77; 78 X; 79 ■; 80 ■; 81 X; 82 X; 83 ◆ -6; 84 □ ◆; 85 □; 116 □; 120 □; 121 -6). The present valley gravel floor has been laid down by the Shaksgam river. Every summer the floodwater arms resulting from the enormous *discharge of glacier meltwater* sweep across the present valley gravel floor over its entire width of c. 1 km (Fig 37 -6). Evidence of this exists in the form of an orographic left-hand *undercutting caveto* in the bedrock, about 2–4 m above the fresh gravel bed (Fig 75 ◀; 138 No. 20). This is a matter of lateral erosion on the *outer bank,* which is caused by the rise in the water level of c. 2m. Nearby Fig 76 shows the formation of a *basal rock socle* (■) below the cavetto (⤵), thus indicating a development in three phases.

Phase one: the *gravel level* occurred c. 2 m higher than now, overlay the rock base (■), and the rock cavetto (⤵) was subject to regular undercutting; Phase two: increased lateral erosion on the escarpment slope (undercut slope) led to denudation of the gravel floor, huts lowering the gravel level by c. 1 m and exposing the bare rock floor (■); Phase three: further lowering of the gravel level by another metre and undercutting of the rock floor, thus creating a small *rock terrace* or rock socle (Fig 76 ■). From now on, the upper caveto is only reached by *exceptionally* high floods; however, regarding the fresh smoothing of the rock, such floods are not *rare* (Fig 76 ◀) – these *seasonal* peaks of meltwater run-off from snow and ice occur between May and the beginning of August. During our 1986 expedition the hydrological data of the Shaksgam and the Yarkand valley were collated and examined by Professor Feng Qinghua (Feng Qinghua 1991, pp. 255–263). The lateral erosion diagnosed above in the bedrock of the Shaksgam valley had at the same time also been undercutting all the *talus fans* and *mudflow cones.* This led to steep terrace steps and rock slide faces, both tens to hundreds of metres high (Fig 27 □ above; 37 □; 39 ■; 52 □; 69 ◆; 71 X; 72 □; 73 □; 77 ⬋; 78 X; 79 ■; 80 ■; 81 X; 82 X; 83 ■; 85 ■). *Besides* the meltwater discharge within the normal seasonal cycle, there are occasions when substantial *glacier lake outbursts* lead to movements on the gravel floor of the Shaksgam valley over its entire breadth, and to such *undercutting* of adjacent loose rock material.

4.5.1.1 An excursus concerning the hydrology: the Shaksgam floods induced by meltwaters and glacier lake-outbursts

The cause of these outbreaks are the 21 km- and 28 km-long glacier tongues of the Kyagar- and Teram Kangri glacier, which advance as far as the Shaksgam valley. Both act as *glacier lake dams* (Fig 138 right below No. 47; Fig 70 below Nos. 6, 5, 8, 3). The Kyagar glacier lake sends regular discharges through a *sub-glacial* meltwater tunnel, the Teram Kangri glacier lake chiefly discharges on its margins but also has a subsidiary, sub-glacial escape route in an ice tunnel (cf. the studies concerning hydrology carried out by Feng Qinghua 1991, pp. 258–263 in the course of the 1986 expedition: "These glacier outburst floods in the Shaksgam valley [Yarkand river system] are characterized by high peak discharge, big rising rate, relatively small total volume and short duration" [as above p. 263]). The nearest hydrographic measuring station is situated at Kagun, far below the junction of the Shaksgam with the Yarkand (1420 m asl; 37°59'N/76°54'E in the northern Kuenlun foreland, leading down to the Tarim basin. The *great distance* (520 km) from the positions of these episodically-discharging glacier lakes is the reason for some of the actual run-off being lost for measurement. Sixteen floods due to *glacier lake outbursts* were observed over the period 1954–1984. They regularly took place from summer to early autumn (June to October). The floods tended to reach run-off peaks of 2000–3000 m^3/sec. On three occasions, lastly in 1984, it rose to 4500–4700 m^3/sec. The highest run-off peak, recorded in 1961, was 6270 m^3/sec. Normal summer *meltwater peaks* registered during the same period varied between 1000 and a maximum of 2220 m^3/sec, with the year of the 1986 expedition seeing the second highest discharge ever measured during the 32-year period of observations standing at 2000 m^3/sec. The *glacier lake outbreaks* produce *flood volumes* of 0.19×10^8 m^3 to 1.5×10^8 m^3, which is about 1/10 of the comparatively continuous annual run-off. In order to give an idea of the *morphodynamic potential* for the Shaksgam valley floor, or the undercutting of its edges, the mean "rate of flood travel" of 11.1 to 16.6 km/h is also significant, although the highest flood discharge peak did *not always* coincide with the greatest velocity. Nonetheless, it is possible to note a general *increase in speed, together with water quantity.* The mean water volume of the Shaksgam river throughout the year is calculated to be around 130–150 m^3/sec. The erosion module for the entire area eroded by the Yarkand system is stated by Feng Quinghua (1991, p. 262 Tab 8) to be 1260 t/km^2/year. As far as *glacier lake outbursts* are concerned, one must imagine the discharge of two lakes with a maximum of 3.23×10^8 m^3 (Kyagar dammed glacier lake on the Shaksgam valley floor at 4760 m asl) and 1.92×10^8 m^3 (Teram Kangri dammed glacier lake on the Shaksgam valley floor at 4520 m asl). The lakes were dammed back by respectively 60- and 90 m-high and 0.3- and 1.5 km-long ice dams, which the glacier tongues had formed. These lakes drain within a few hours, so that the actual passage of the flood wave through the valley takes about 18 to 22.5 hours.

4.5.1.2 Large young mud fans, debris slopes, alluvial fans and terrace remnants in the Shaksgam valley and their mutilation: an example of diametric, syngenetic morphodynamics in the Karakorum

It has not been possible to provide unequivocal evidence of the maximum level of run-off reached during a dammed glacier lake outburst, but 4 metres above the normal flow level per 1 km of valley width are *probable; considerably more* is to be expected in areas of narrow, *bottleneck-like,* parts of valleys. This is, however, the place to draw attention to the Myricariae bushes in the foreground of Fig39. Only 1-2 m above the receiving stream (though in the recess of an inner bank), they were nevertheless not swept away entirely with roots and all, but were able to recover after the flood on August 30th, 1984. The problem remains unsolved. The question of the *water-level* does not really affect the effectiveness of the process of *undercutting* these large cones and fans (Fig 27 □; 37 □; 39 ■; 52 □; 69 ◆; 71 X; 77 ➘; 79 ■; 82 X; 85 ■). In every case it amounts to a retreat, which, starting at the *distal base,* continues right to the top in the form of these crumblings on the edges of cones and fans (Fig 77 ➘; 78 X). The importance of undercutting can be *directly* assessed from these crumbling masses: metre-high and very fresh (one year old) *debris piles* of material from the steep cliffs above have accumulated at the foot of gravel- or debris-cliffs. They are removed by summer floods or glacier lake outbursts (Fig 27 ▽; 52 ▽; 78 ▽; 79 ▷; 82 ▽). At times the steep gullies and "organs" in these escarpments produce small special debris cones or fans, which are only a few months (Fig 81, below X) or a few years old (Fig 80 below ■; 27 between □▽□). Corresponding special cones also emerge from gravel and debris caves (Fig 82 ↓) in karstic dolomite debris. In any case *all* these *tributary* debris infillings, from steep tributary valley gorges or wall gullies, are characterized by substantial *distal* transformation (Fig 37 □; 73 □; 78 X; 80 ▽). The *height* of the debris- and gravel cliffs present a measure (Fig 69 ◆; 77) of their considerable intensity, as this is the *equivalent* of the degree of *mutiliation* of these accumulations in their *ground plan,* and at the same time for the considerable removal of *masses or cubatures.* In the face of such *denudation dimensions,* the observation becomes more important that these cones and fans are nonetheless actively *engaged in formation.* In early summer the snow melt causes mudflows and sediments from mountain torrents to be deposited on the floor of the Shaksgam valley, from which they are *syngenetically* removed. These accumulations consequently present *stable* forms at the *peak of their development.* Supposing that such ruins of cones and fans are stable, the processes of *synchronous* aggradation and degradation become the direct expression of *extreme morphodynamics* in this longitudinal valley of the Karakorum (Fig 84, right-hand side of the photo). In extreme cases the fresh escarpments of cones and fans reach heights of 100 to 120 m (Fig 27; 39; 69; 77; 82; localities: Fig 138 Nos. 20–23), though having been deposited against the Shaksgam glacier (cf. below) many began their development as *kames,* and therefore possessed escarpments from the very start. There is no doubt that the mudflows and alluvial fans mentioned above are the most striking forms, since they constitute the most significant post-Late Glacial (younger than 12870 ±180 YBP, Holocene) debris deposits laid down in the Shaksgam valley. They tend to be *autochthonous* or rather deposited directly in the mouth of short tributary talwegs. But there are also gravel terraces which have built up along the valley and are preserved at three different levels at least, namely at 20, 60–40 and 120 m above the present gravel floor of the Shaksgam river (Fig 72 -2; 73 -1 bottom right; 37 -0; 138 No. 20). The lowest and most recent level is represented on a larger scale (Fig 51 -2; 52 -2; 138 No. 11) in the junction of the Muztagh valley, as well as SE of the mouth of the "southern Aghil pass-valley" (Fig 70 -2; 138 No. 22), whilst minor remnants of the 40–60 m terrace (still reaching 60 m here) have been preserved in this valley chamber of the Shaksgam valley on the orographic left (Fig 70 -1; 138 No. 24). Another minor example of this 40–60 m terrace has been preserved a little further down valley, again on the orographic left in the area of the junction of the two still glaciated Karakorum gorges (Fig 83 on both sides below No. 7), which run down to the Shaksgam valley from the S (Fig 83 -1; 80 -1; 79 -1; 138 No. 23). Deposited from the glacier outlet positions of three Holocene stages of *Shaksgam glaciers,* these three *gravel floor terraces* are classified as Nauri Stage (V), older Dhaulagiri Stage (VI) and middle Dhaulagiri Stage ('VII) and are numbered -0, -1, and -2. In accordance with the nomenclature for High Asia, which is applied here (Kuhle 1982, p. 118), the assignment of moraines (numbering on the left) to glaciofluvial gravel floors (numbering on the right) is as follows:

Main Ice Age	0 - No. 5	Older Dhaulagiri Stage	VI - No. -1
Ghasa Stage	I - No. 4	Middle Dhaulagiri Stage	'VII - No. -2
Taglung Stage	II - No. 3	Younger Dhaulagiri Stage	VII - No. -3
Dhampu Stage	III - No. 2	Stage	VIII - No. -4
Sirkung Stage	IV - No. 1	Stage	IX - No. -5
Nauri Stage	V - No. -0	present glaciation	- No. -6

The fact that terraces set in in the Shaksgam valley is evidence of the maximum Holocene extent of the Shaksgam glacier. The glacier stage concerning the gravel floor terrace -1 (Fig 70 -1) is to be classified as the Older Dhaulagiri Stage (VI), and terminates at a maximum distance of 15 km from the lowest glacier tongue, that continues to reach the Shaksgam valley (Gasherbrum glacier tongue, Fig 83, visible in the background below No. 1 on the left). At that same time the three large Shaksgam glaciers further up-valley - the Kyagar-, Teram Kangri- and Urdok glaciers - were still flowing into the Gasherbrum glacier, being the one with the lowest tongue end position. The same applies to the outward-lying Skyang glacier. Together the glaciers formed a *dendritic valley glacier system, verging* on a minor ice stream net. It follows that, belonging to the Middle Dhaulagiri Stage 'VII, terrace -2 extended some kilometres *further up valley* than -1, where it reached the *next younger* glacier outlet of this glacier system. The lowest remnant of terrace -2 reached by the author is shown in Fig 83 (-2) (cf. Fig 70 -2). Fig 37 shows the *most up-valley* terrace remnant of a glacier mouth outwash plain -0 in its topographical context. The distance to the *present* Gasherbrum glacier end is 35 km (Fig 138 No. 20). The *next older,* and therefore *higher,* terrace of a glacier mouth outwash plain must be assumed to be *down-valley* from the junction of the Muztagh valley. In accordance with the nomenclature which has been introduced here, it must be numbered 1, as it belongs to the Sirkung Stage IV, the last *Late Glacial* glacier position. Older than 12870 ±180 YBP, this stage must have had a much more depressed ELA than all the other Holocene stages, the most powerful advance of which had been the neo-Glacial Nauri Stage V (c. 4000–4500 YBP). The terrace -0 mentioned above is part of it. Regarding the development of both the Muztagh glacier and the Shaksgam glacier during the Sirkung Stage IV, it follows that a confluence of the two ice streams to a joint superior *Shaksgam glacier system* is also likely in view of the absence of terraces No. 1 in the confluence area (cf. Chapt. 4.2). Up to the terrace remnant No. -0 furthest up valley (gravel floor of the Nauri Stage V) in the Shaksgam valley at 3900 m asl there is an altitudinal difference of 400 m, and thus an *ELA difference* of 200 m (ie 400:2) from the present Gasherbrum (Shaksgam) glacier end at 4300 m asl. Fluctuating between 240 and 560 m, it thus fits into the pattern of neo-Glacial ELA depressions which can be observed in many places in High Asia (Kuhle 1987 c, p. 205 and 1986 e, pp. 441–452; Shiraiwa & Watanabe 1991, pp. 404–416). According to this, the *depression value* of c. 200 m for the Nauri Stage V is on the *low side;* it is explained by simultaneous *important glacier elongation* of almost 35 km (cf. above), resulting in *enlargement of the surface,* which may have replaced *part* of the vertical descent of the glacier tongue by the increase in its ablation.

The presence of gravel floor terrace remnants must not lead to the conclusion that the up-valley Shaksgam valley glacier occupied the valley over its entire breadth, nor that all the alluvial debris fans and mudflow cones are younger than the particular glacier stage in which the valley cross-profile in question was still reached by the glacier. This applies particularly to the progessively lower and relatively young outwash gravel floor terraces (-1 and -2) further up valley. On the contrary, by being laid down against the body of the valley glacier as a base, these fans and cones were *syngenetically* deposited as *kame formations.* This approach allows the entire period to be available for the formation of all these *very substantial* and *thick* formations of fan and cone forms in the Shaksgam valley (Fig 27 □ above; 37 □; 39 ■; 69 ◆; 71 X; 73 □; 77; 82 X; 84, right half ▼ inter alia) during which the surface of the valley glacier in particular valley cross-profiles clearly remained *below* the snow line. This was just the case during the youngest Late Glacial period, or, in the case of the Shaksgam valley section in question, the floor of which now stands between 4000 and 3850 m asl (Fig 138 between Nos. 22 and 20) approximately during the Sirkung Stage IV. The idea of such a prehistoric process of sedimentation using the valley glacier ice as a base is supported by the above-mentioned observations from the bank basins (margin valleys) of the K2- and Skamri glacier (Chapt. 4.1.1.). Where ever the glacier tongues leave out a relatively wide *bank basin* or an *ablation gorge* - as a rule on *one* valley side only; on *both* sides only in the vicinity of the tongue end - the space between valley flank and glacier margin is infilled by debris cones and talus slopes from the immediately adjacent valley slopes, and by mudflows and alluvial fans from the tributary talwegs (Fig 10 ▷ far right; 12 ▽; 14 X; 16 ◆; 23 a ■; 28 ▽; 30 ▽; 34 x; 35 ▷; 47 ▽). On its ENE bank the Shaksgam glacier developed a *wider* area of bank basins in which those fans and cones were *syngenetically* filled. It is the WSW-facing side of the valley which, thanks to its *exposure to radiation,* has thawed out more (Fig 39 ■; 69 ◆; 71 x; 82 x). In the area of the tributary gorge exits on the orographic left-hand side of the valley, *kame-like* alluvial debris fans (-1) could not be accommodated before the older Dhaulagiri Stage (VI) (Fig 79, 80 and 83 -1). This is the reason for the exposure of *material from lateral* and *ground moraine* over long distances at this flank. The form of the moraine has *not* been preserved because it is covered by small debris cones and debris talus which accumulated from above (Fig 83 ▽, centre third of the photo 84 ▼ right half). On the other, orographic right-hand, valley side moraine material *from the same period was deposited against* the intermediary cones and fans; this, however, took place in a position (on the outer bank) which was exposed to a later Shaksgam river, so that moraine material of such recent origin has *not* been preserved. Markedly older - ie in this case *Late Glacial* - morainic material (Sirkung Stage IV and earlier) - belongs to a considerably *wider* Shaksgam glacier: that fact explains why, in some places, corresponding *ground moraine material* (basal till) can be found at the *base* and in the *core* of fans and cones (Fig 81 x; 82 x; 85 ■). Cones and fans consequently began to develop on the orographic *right* (Fig 138, from No. 22 to No. 21) - though further down the valley on the *left* side as well, where 5 to 10 km *down the glacier* a bank basin developed at the same time (Fig 73 □; 138 from No. 23 to 12). As *deglaciation progressed* and the ice receded, they spread out over the ground moraine. It follows that their formation *began* when the ELA was raised above the level of the glacier surface of particular valley cross-profiles. When the snow line was noticeably lower than now, the construction of these *kame-like lateral glacial margin accumulations* was initially most actively intensified by the intervening layers of glacier, firn and eventually snow in the catchment areas of the tributary valleys, gorges, and wall gullies (Fig 27 O; 37 O; 69; 80; 84 below No. 2). Activities on the orographic right-hand flanks of the Shaksgam valley have meanwhile largely been reduced to the seasonal melting of snow (Fig 37 O left), whereas on the orographic left the N-facing Karakorum hanging glaciers in the heads of tributary valleys continue to be effective (Fig 80 and 83 on both sides of No. 7). Besides the large fans and cones which were first started in *late-Late Glacial* times as kame-like bank formations and continue to develop through current glacifluvial undercutting, much *more recent* debris fans and debris cones can be identified as well. Fig 72 provides an example for this: small, recent as well as larger debris cones (▽) have been set into the outwash gravel floor terrace -2, and must therefore be younger than c. 2000 YBP. The older gravel floor terrace -1, together with material from older ground moraines, has been removed from this outer bank *before* the sedimentation of -2 began. Further down valley, the adjacent alluvial debris fan (□) shown in Fig 72 must be *even younger* than terrace -2, since it has been deposited in the denudation space previously occupied by terrace -2, which had been placed in front of the junction of a tributary valley. Set into its surface, the larger debris cone contributes, albeit on the margin, to the covering of terrace surface -2 (▽ above □). This cone is therefore bound to be still younger than the alluvial debris fan (□). Its comparatively large size can be explained by the size of the wall gorge with its high-rise catchment area (↓).

The question of the *overall thickness* of debris fillings and gravel floor, which is showing the most recent glacio-fluvial traces on its surface, ie the thickness of loose material in the present valley floor (Fig 84 ◇ □; 85 □ ■), can only be answered *vaguely*. Judged by the *steeply flanked* trough form (Fig 116 □), or by the *obtuse angle* formed where the *box-shaped* insert of the gravel floor meets the trough flanks (Fig 120 □), the valley floor of the Shaksgam should lie *several hundred* metres below the corresponding thickness of loose material (Fig 37 □ -6; 39 -6; 71 □). It is doubtful whether the loose material consists solely of weathered detritus and gravel. Rather more likely are insertions of ground moraine layers from the Early Glacial, Late Glacial and Main Ice Ages (cf. below). Regular post-Main Ice Age *gravel infilling* during deglaciation, which can be observed in these longitudinal valleys of High Asia, has only been partially reversed by Late Glacial to Holocene incision (cf. above), and argues for the fact that both the *moraines of several* Pleistocene *ice ages* and gravel from intermediate *inter-glacial periods* have filled the Shaksgam trough.

4.5.2 Loose rock in the Yarkand valley and its tributaries

"The N-Aghil pass-valley" or upper Surukwat valley N of the Aghil pass (36°10'-17'N/76°32'-40'E; Fig 138 No. 25) visited by the 1986 expedition belongs to the upper catchment area of the Yarkand valley. The 4863 m-high Aghil pass forms a flat valley-head (Fig 86 ↷; 87 ↷). This valley-head area has not been reached by glacier ice since at least 1655 ± 180 YBP (Tab 2, sample No. 20.10.86/1), though the sample locality, which suggests the *absence* of glacier ice, is situated 140 m N *below* the pass culmination (Fig 87 O). The age of the sample affects the position *near the talweg*. This had been previously reached by the *tongue ends of hanging glaciers,* which flowed down from the good 6000 m-high Aghil main ridge to the W and E. Minor hanging glaciers and firn shields *continue* to be present on the E and W exposition of the two valley flanks (Fig 86 below ----; 87 near the bottom left-hand margin and in the centre; 88). The sample was taken from an erosion edge of the present meltwater stream (talweg of the upper Surukwat valley, N of the Aghil pass; Fig 138 No. 25). It had been uncovered by thawing permafrost. The *permafrost table,* too, is evidence of the *climatic proximity* to glaciation in this location. In the light of what has been said above, it is possible, though not certain, that glacier ice was still reaching this valley cross-profile during the older Dhaulagiri Stage VI (c. 2400-2000 YBP). It is, however, certain that the *outwash cone* (gravel floor) of the glacier of that stage was deposited here. This implies that the valley cross profile was probably last reached by the glacier end during the *neo-Glacial* Nauri Stage (V). There is another sample locality somewhat further down valley (Fig 138 No. 26), close to the talweg at 4630 m asl (Tab 2, sample No. 20.10.86/3), which must have been free from ice since at least 6205 ± 145 YBP (Fig 87 ↓↓↓). In accordance with the age of the peat on the *outwash cone* (glacio-fluvial gravel floor), the latter is to be classified as belonging to the early Holocene or late Late Glacial period. In the same locality, and again above the permafrost level, more recent and cover-forming outwash cone material in an overlying peat-horizon can be *dated as at least* 355 ± 80 YBP. It is likely to be from the middle or younger Dhaulagiri Stage ('VII or VII), ie to be less than c. 2000 or 440 YBP (Kuhle 1987c, p. 205 Tab 2). The analyses (Fig 57a/b/c) show that the three samples (20.10.86/1/2/3) are directly *glacially-induced* sediments consisting of a mixture of granite and limestone debris with another fine grain-size peak with the clay fraction, which is typical of moraines (Vagners and Dreimanis 1971). The area down-valley from the upper Surukwat valley (Fig 88; 89; 90) has consequently been free from ice for more than 6200 years. Since deglaciation during the late Late Glacial period (Sirkung Stage IV) its valley floor has been filled by *gravel-fields* (Fig 88 □; 89 □◇) from the adjacent small valleys as well as from younger *mudflow cones* (Fig 88 ○; 89 ○; 90 ○). The *highest*, ie *late Late Glacial*, lateral moraines are preserved at 4300 m asl on the orographic right (Fig 88 ■; 138 No. 27). They are classified as Sirkung Stage IV, and belong to a *snow line depression* of c. 500 m (at present, the lowest ice margins occur at 5200 m asl, IV ice margins at 4200 m asl ⇨ ELA depression of 500 m). Here there are both *lateral moraines* with crests (◆) and *kames* which, being glacigenic lateral formations, left behind *kame terraces* (■). Fig 91 provides an insight into the next lower valley chamber where the above-mentioned upper Surukwat valley, leading down from the Aghil pass, joins a western source branch of this valley system. The valley floor consists of interlocking *gravel-fields* or *outwash cones* (−2 to −6) of *historic* to *present* glacier ends of this catchment area (Fig 138 No. 28).

Developing approximately since the middle Dhaulagiri Stage 'VII, ie for c. 2000 years, these gravel-fields (−2 to −6) possess terrace steps up to 12 m high (■). In places where the valley floor drops below the 4000 m-line, very thick *Late Glacial* gravel-fields, which by dissection are transformed in terraces, take over *abruptly* (Fig 92 Nos. 3, 2, 1; 138 No. 29). They are superimposed by more recent *local moraines* from small, steep hanging glaciers of directly contiguous catchment areas (Fig 92 ■). Further up valley, at 4000 m asl, these gravel-fields dovetail with *Late Glacial main valley terminal moraines*, which must be classified as Dhampu Stage III (Fig 138 No. 29). Three distinctly separate terrace levels can be discerned (Fig 92); No. 1 belongs to the Sirkung Stage IV, representing the youngest level and lying 50 m above the present talweg. No. 2 (Dhampu Stage III) reaches c. 130–140m, and No. 3 (Taglung Stage II) 200–260m. The two older, ie higher, terraces in particular consist of glacio-fluvial gravel deposits and re-deposited gravel, with even *older moraine material* (from the Taglung Stage, and before) worked-in or washed-out by the activities of *outwash cones* near the lateral ice margin (Fig 92 ↓ inter alia). The picture shown in Fig 92 was taken from a ridge of outcropping slates, which juts out abruptly towards the middle of the valley, followed by a glacigenic *ravine-like*, narrow and steep passage (Fig 138 No. 46) of the Surukwat bottom contour line.

Interspersed with several metre-high cataract steps it shows *potholes* in the rock-floor of the talweg and on its flanks, suggesting a *sub-glacial* formation caused by hydrostatic meltwater pressure. Since their formation requires a rise of the ELA *beyond* the 4000 m line, the *Late Glacial Period* of our chronology is the *only* one possible. Down valley the ravine-like, steep and narrow passage – which could be interpreted as a Late Glacial *ice margin position* – the now even richer *glacio-fluvial gravel deposits* continue in the form of terraces up to 300 m-high (Fig 138, Nos. 30, 31). The highest terrace begins with a steep surface incline (Fig 26 ▽ No. 3), which subsequently flattens out (Fig 22, No. 3). The enormous glacio-fluvial infilling of the W-Surukwat valley begins here, at 3750 m basic height asl. The greater part of its gravel bodies belong to the Late *Glacial period*. All in all eight generations of terraces can be discerned (Fig 22 ▼▲). The lower terraces, only a few metres to decametres-high, are younger. They belong to the post-Late Glacial period, and were deposited in a *deep* erosion incision (Fig 22 Nos. −5 to −0) in the form of thin gravel *bands* or gravel *field segments*. The incision had been created in the Late Glacial gravel deposits by the meltwater discharge of the melting remnants of the Late Glacial ice stream net (Fig 22, Nos. −5 to −0). A *typical* feature of glacier outlet gravel floors (valley outwash) canalized by a valley is the rapid *loss* in sediment thickness and the vertical displacement these terraces experiences on their way down the valley. Over a distance of 10 km, their heights, whilst retaining their *proportionality* to one another, thus decrease to *less than half* (Fig 22, cf. No. 3 on the far right with No. 3 on the far left, cf. 26 No. 3 with 94 and 95 No. 3). The *same* glacio-fluvial terrace sequence occurs (Fig 22 ▼▼, far left; 95 Nos. 1–3 left) in the opposite, eastern, source branch of the Surukwat valley, which is five times longer than its western counterpart (Fig 138 No. 32). The gravel body likely to be the oldest one (Fig 22, far right, at the base of terrace section No. 3), has been TL-dated at its base as approximately 12 Ka (dating by E. Drosdowski, Torun Academy, and S. Fedorowicz, TL-Laboratory, Gdansk University). In this valley very well preserved *glacier striae and polishings* (Fig 40, 41, 93, 124) have been found *below* the level of the gravel terrace surfaces, at locally exposed outcropping valley flank surfaces on the orographic right

(Fig 138 No. 46). These polishings must consequently be classified as older, ie as belonging to the *older Late Glacial* to the last *Main Glacial period* (cf. Chapt. 5.3). It is highly likely that a large part of such striae is *buried* under these gravel deposits. Over the 5 km stretch from the confluence of the two Surukwat source branches (Fig 95; 126 background) to its junction with the Yarkand valley, the Surukwat valley shows continuations of the *glacio-fluvial* gravel terraces (cf. Fig 96 ▼ No. 3; 126 ▼ Nos. 1-3). These forms appear as steps on three levels, and dove-tailed with *mud-flows,* which are in consequence also part of the *middle - to late* - Glacial period (cf. above) (Fig 97 ▼ Nos. 1-3). Here, on the lower Surukwat valley and in its continuation down the Yarkand valley, the maximum terrace height (No. 3) decreases to c. 100-80 m (Fig 98 ▼ No. 3). In this area there are outcrops of thin strata schists (mica schists). The glaciated knobs of the Yarkand valley, which are formed in these rocks of the Main to early Late Glacial period (Ghasa Stage I) are partly buried by sediments (at times up to half their height), ie covered by gravel (Fig 138 No. 33; 96 ▲; 98 ▲; 99 ▲; 100 ▲; 101 ▲ left). In the ensuing down-valley section of the Yarkand valley, the oldest terrace areas take up a strikingly large amount of room (Fig 98 □ right, No. 3) and have been preserved over almost the entire width of the valley. This is explained by the fact that the river was confined by a gorge. From this point a *glacigenic ravine* has been cut into the rock threshold with the aforementioned glaciated knobs (Fig 102, 96 ↑). This ravine lies at 3400 m asl (Fig 138 between Nos. 33 and 34), and thus *so far below* the Late Glacial snow line that *sub-glacial meltwater* was able to form and carve out this cut in this area, though the valley glacier surface was much higher at the time. The glacier retreat from the valley cross-profile was *initially* followed by glacio-fluvial infilling, and subsequently, when the Surukwat river cut into the valley floor, by its removal. From now on, the river followed the ravine; it was therefore confined to this narrow valley floor section, with the result that these *further* terrace areas remained intact. In this confluence area the Yarkand river, too, has been similarly confined by a ravine (Fig 138, No. 34). Here it passes through an extremely narrow trough profile (Fig 99, middle section of the photo).

From here, and over a distance of 40 km up the Yarkand valley (10 km beyond the Mazar military station, till No. 36 in Fig 138), the valley floor infilling with gravel terraces remains conspicuously small by comparison with the Surukwat valleys (cf. Fig 105 ▼, 110 ▼). They are only partly preserved in this valley, and their thicknesses do not exceed a few decametres (Fig 60, 103 and 104 ▼). The late *Late Glacial* gravel floor deposits dovetail with *alluvial fans* and *mudflow cones* from adjacent steep tributary valleys and from valley flank gullies, or have in part even been transformed or replaced by them (Fig 138 Nos. 35 and 36; 105, O). The mudflow cones and fans tend to be noticeably younger rather than of comparable age (Fig 109, O). Predominantly from the *Holocene,* they *continue to develop* (Fig 106 x, 107 x, 129 x). The two largest preserved terraces attain heights of several metres above the present river level (Fig 104 and 105 ▼).

Organogenic material (Fig 108 □, Tab 2, samples 24.10.86 1b/1c/1d) indicates that the age of these gravel deposits is more than 1925 to 5935 C14 years, thus making them the *oldest* dated gravel sediments in the Yarkand valley. Moreover, dating of lower surfaces of gravel terraces (Fig 60 ▼ in low elevation above the receiving stream; 110 ▼ far left) and surfaces of alluvial fans in the Yarkand system of the section mentioned before established them as being approximately 110 (even 40) - 155 YBP (localities: Fig 65 ↓; 108 x □), as shown by the historic to current glacio-fluvial activities between the Aghil mountains and Kuenlun (Tab 2, sample 24.10.86/4; 25.10.86/1/3). Fig 55 (Nos. 1, 3, 4, 5, 7, 10), 66 (Nos. 1, 2, 3, 4, 6, 7); 67 (a/b); 74 (a/b); 56 (24.8 and 24.10.86) and 68 describe the *sedimentological* features of these *Holocene* gravel floor terraces and even younger deposits of the Yarkand river (localities: Fig 138 Nos. 37, 38, 39).

Attention is to be drawn to the significant difference between the graph of cumulative grain sizes taken from *glacio-fluvial* terrace sediments (Late Glacial to Holocene gravel fields, Fig 55 Nos. 3, 4, 7 and 10) to those sediments of predominantly fluvial genesis in the recent talweg area of the Yarkand valley (Fig 55 Nos. 1 and 5). The proportion of pelites, typical of *glacier milk,* is significantly reduced in the *river sediments,* so that 82% of its fraction remain the *sand spectrum range.* The difference between *gravel-field features* and the *examples from moraines and mudflows* (the graphs of which are almost the same; Fig 55 Nos. 6 and 9) is far smaller than between *outwash gravel fields* and purely *river sediments.* Most of the gravel-field samples of the Yarkand valley and its tributaries continue to show the fine grain peak typical of moraines (Dreimanis & Vagners 1971) (Fig 67b; 74 a/b). In this case it is evidence of the sedimentological *proximity* of the moraine, and thus of the glacier, as an *essential* feature of glacier outlet outwash plains or *sanders.* The first four columnar diagrams of Fig 56 (24.8.86/1/5; 24.10.86/1/1a-d/2) illustrate how much the *morphometric* features of quartzite grains in the terraces of the Yarkand system vary in relation to the *transport distances.* 24.08.86/1/5 (cf. also Fig 68) are taken from *short* tributary valleys. 24.08.86/1/1a-d/2 are from the *long* Yarkand main valley. The high proportion of "fresh" material is explained by the omnipresent supply of solifluidal and denudation *slope debris* but also by glacigenic fractures caused by small glaciers in the tributary valleys. The significant proportion of "dull" material indicate aeolian processes typical of arid, or *semi-arid* climatic regions.

However, this requires qualification in so far as the characteristic of "dullness" can *also* be observed in glacigenic processes, like the grinding of lower and ground moraine, which produce grain surfaces with polished edges and a dull grain surface not unlike that of SiO_2-grains transformed by aeolian processes only (Fig 68◁◁). In this respect the 50-65% of "dull" components of the tributary valley samples 24.08.86/1/5 are more informative about the *short* distance of only a few kilometres to the present glacier ends of the Kuenlun than about aridity. As stressed before, Fig 105 shows very clearly how relatively little debris material and terrace material fills the floor of the Yarkand valley section higher up. In this context attention must be drawn to the considerable present *denudation and solifluction* processes which shift the detritus on the slopes by c. 2–6 cm/year downhill (Fig 105 ▷). It would have filled the valley floor disproportionately more, had there not been *glacial* removal during the ice age – with *several* repeats during the Pleistocene (cf. Kuhle 1991d, pp. 139, 141, 170). Even these tributary valleys without a *short* connection to higher mountain massifs of the Kuenlun, and consequently with no glacigenic supply of material from ice margin positions on the edge of larger valley glaciers during the Late Ice Age and Holocene, are noticeably *poor in debris and gravel* as compared with present material weathering (Fig 109, 110 left half of the photo). It is essential for this approach to bear in mind the spatial differentiation of valley sections with *directly* adjacent high mountains, and thus Late Glacial *supplies of glacigenic material* as it occurs, for instance, in the area of the W Surukwat valley, where glacio-fluvial accumulations of vast dimensions are a typical feature (Fig 138 No. 31; 22 ▼ Nos. 1–3). Here it belongs to the *post-Main Ice Age* (cf. above). It is necessary to bring to mind the details of this strange, but at the same time *characteristic,* distribution of the quantitites of debris within these valley courses in their dependence on the distance from Late Glacial and Holocene glaciers in order to draw the right conclusions in the summary of this paper (cf. summary chapter 4.5.3).

4.5.3 Loose rocks in the "Vale of Kudi" and the gravel-fields up to the N-foreland of the Kuenlun (Tarim basin)

This sub-chapter has to fill an additional systematic function. In doing so it closes the spatial gap between the Yarkand valley area and the Kuenlun S-slope together with the Kuenlun N-escarpment, down to the mountain foreland in the Tarim basin. The Fig 54, 21, 53, 25, 59, 58 and 111 represent the S to N down-valley succession of cross-profiles of main valleys and tributary valleys with their *infilling of loose rock* between 5000 m, or – as the case might be – 6000 m (Fig 21) and 3000 m asl. Fig 21 and 53 show valley cross-profiles adjacent to areas of the Kuenlun main ridge, which are still glaciated, including the Holocene and Late Glacial moraine discussed above (cf. Chapt. 4.3, Fig 138 No. 15). The gravel-fields here possess *narrow* linear outlines along a talweg (↓) confined by high moraines. Fig 54 depicts an equally high valley floor (as Fig 21 and 53) though it is almost entirely *lacking* in moraines and loose gravel rocks (Fig 138 No. 16). The talwegs of this most southerly catchment area of the "Vale of Kudi" extend up to high hollows or troughs, the glacier of which is melted away quite *suddenly* after filling them completely throughout the Ice Age. They have *not* experienced *any* Late Glacial to Holocene glacier activity since. This is due to the fact that these high hollows or troughs are not flanked by *any* peaks which rise substantially above their level, ie by potential late-Late Glacial to Holocene glacier catchment areas. That is the reason why at present there are no glaciers, but *only* snow patches on these mountain ridges (Fig 54). The sample locality No. 17 (in Fig 138) has already been mentioned with regard to the Late Glacial moraine sequences in the "Vale of Kudi" and its tributary valleys (Chapt. 4.3). The C14 dating of sample No. 20.08.86/2 shows that the lowest, only 2 m-high terrace is more than 1610 ± 90 years old (Tab 2). Apart from alluvial fans of tributary valleys there are no higher terraces in this location at 3740 m asl. The same terrace representing the *characteristic level* not only emerges upstream at 4000 m asl (in the area of Fig 25◀) but also down-valley in the main valley below near the Kudi settlement at about 3000 to 3200 m (Fig 58 ▼) and repeatedly (Fig 138 No. 40) in the tributary valleys (Fig 59 ▼). Fig 57/d provides information about the fine grain composition of this gravel terrace at 3740 m. At this height- interval, the loose rock valley floor in the main valley, including the present river bed (□), is only a few hundred metres wide (Fig 58). Judged by valley walls, which drop steeply below the valley floor, the *gravel thickness* might be estimated at one hundred metres or more (Fig 58 and 59). But a deposit of unclassified rock, like diamictites, is not to be excluded either. The undercutting of the outer banks in the bedrock (Fig 58 ↗) is an indication of a *different* prehistoric genesis of the valley profiles from that of the present fluvial processes – a *glacigenic* one, in fact (cf. Chapt. 5.4). The autumn waterflow (late October 1986) of this mountain river or stream in the "Vale of Kudi" near Kudi is about 0.7–1.2 m^3/sec. In the lower course of this valley further north, up to about 20 km down stream from the Kudi settlement remnants of several metre- to decametre-high glacio-fluvial gravel-field terraces with sporadic *accumulation of rough boulders* are preserved above the present valley floor in narrow ledges along the rock-walls. This is quite evidently Late Glacial moraine material (outwash) which has been washed out *near ice margins* – a fact that can be deduced from the size and packing density of the boulders (Fig 138 No. 41).

For reasons of expedition logistics, the northern continuation of this Kuenlun valley system up to the mountain foreland, ie into the Tarim basin, was subsequently studied further east, in the "Vale of Pusha" beyond the Akaz pass (3270 m asl; Fig 138 No. 42). The Fig 112 and 113 show complimentary perspectives of the lower section of the "Vale of Pusha" (Pusha settlement: 37°20'N/77°08'E) and of its *glacio-fluvial gravel field,* which forms the valley floor (Fig 138 No. 43). These are *Late Glacial* gravel- fields of the Ghasa Stage I (Kuhle 1982, pp. 154–55), which cover the floor of the glacier tongue basin between the Main Ice Age moraines in a cord-like gravel field (Fig 112 No. 4; 113 No. 4). It is possible to discern at least two terrace levels at 10–6 m (Fig 112 ▼ No. 4) and 25–20 m above the present talweg (Fig 113 ▼ No. 4), which are evidence of two *early Late Glacial* accumulation phases of the Ghasa Stage I. The terrace heights decrease towards the valley exit as such glacio-fluvial accumulations tend to do. Hundreds of small *"special alluvial fans"*, which have been pushed out of the gullies of the inner moraine slopes (Fig 112 ▽), are set into their surfaces (Fig 113 ●). *Late Glacial gravel-field segments* of this kind run through the tongue basins of the piedmont glaciers of the main ice ages in repeated patterns of *parallel stripes*, extending from W to E over a distance of more than 100 km S of the Yehcheng settlement (Fig 113 No. 0; 114 No. 0). These gravel-field cords pour into the mountain foreland proper of the Tarim basin (Fig 114 No. 4) from the exit of these *Main Ice Age end moraine valleys* or *tongue basins* from an altitude of approximately 1900- 2100 m asl. Having been channelled by the parallel striped end moraines over a distance of 15-30 km, the gravel-fields were now able to *fan out* widely (Fig 138, Nos. 44, 45) from here, ie from the position of the Main Ice Age glacier margins and glacier outlets (Fig 114) and to settle down as *very extensive gravel fields* with scarcely any relief, which extend as far down as c. 1500-1400 m asl (Fig 115 Nos. 5 and 6). They reach the present *circle of irrigated oases* in the interior of the Tarim basin at the point of their transition to *limnic* sediments in *terminal basins*, which used to be lakes even in Late Glacial times. This is the case in the S-N profile near the Yehcheng settlement. In the more constricted N-Kuenlun mountain foreland beyond the *very extensive* ranges of Main Ice Age end moraines (see below, Chapt. 5.4), some of the younger Late Glacial sections of gravel-fields spread over the much more extensive Main Ice Age gravel-fields (Fig 138 No. 44 above No. 45), which in turn must have covered those of older Pleistocene ice ages (Fig 115 Nos. 5 and 6). The graphs 2 and 8 in Fig 55 (Fig 138 ✧ N of No. 45) show the *grain size composition* in these gravel field fans in the foreland with surface inclinations of 1-2° over a distance of 60 km from the bedrock at the foot of the mountains, as well as the *qualitative* difference between them and the *channelled* gravel-fields with shorter transport distances in valleys (graphs 3, 4, 7 and 10). On the other hand, the contrast with graphs of *purely fluvially-transported* deposits (graphs 1 and 5) is just as evident. Fig 66 demonstrates the *essential contrast* of the grain size composition of two more samples (17.10.86/1/1A) from the same locality of *gravel-fields* in the *foreland* (Fig 138 N of No. 45) (graphs 5 and 8) to other channelled gravel-fields from tributary valleys of the Yarkand (graphs 1, 2, 4 and 7). Showing a proportion of more than 75% of *dulled* quartzite grains, Fig 56 (17.08.86/2) stresses the high proportion of *aeolian* processes involved in shaping the grain surfaces of the pelitic components of this glacio-fluvial sediment (Fig 138 N of No. 45). The reason for this is to be seen both in the *aridity* of the climate in the Tarim basin, which continues to expose the bare gravel-fields to deflation and corrasion and the time available since at least 18 Ka BP (Fig 115 ○, No. 5 and 6). Fig 127 gives a directly visible rendering of the high proportion of dull grains in the range of >63 µm.

It must be remembered that, within this study, which focuses on the *Main Ice Age glaciation*, the detailed descriptions and analyses of *outwash gravel-fields* from the Main Ice Age to the post-Glacial period via the Late Glacial to historic times (Chapt. 4.5.1.-4.5.3) have the function of providing evidence to demonstrate how *fresh* and *youthful* these glacio-fluvial infillings of valley floors are, and how entirely *different* these post-Main Ice Age processes must have been from the preceding ones which gave the valleys their forms. It has been shown that the *present formative processes* in the main valleys *cannot explain* their shaping (cf. Chapt. 5.4).

5. Maximum Prehistoric Glacier Infilling of the High Mountain Relief between the Karakorum Main Ridge and the North Kuenlun Foreland (Fig 138)

In the course of the description of geomorphological and sedimentological traces of maximum prehistoric glacier infilling, the general topographical sequence from the highest, *present,* glacier areas of the Karakorum to the *lowest,* former ice margins in the mountain foreland of the Kuenlun is here being *repeated* for the fourth time. Related to the description of present glaciation (Chapt. 3), the Holocene to Late Glacial glacier positions (Chapt. 4.1 to 4) and the outwash gravel-fields (Chapt. 4.5.), this S-N sequence has hitherto dealt with the mountain thresholds of the Aghil and the Kuenlun, which have been inserted, running from W to E. This has been repeated three times in this S-N profile, beginning each time *at the top* again, at the present glaciers, and subsequently following three downward profiles every time. This would even have to apply to the *maximum* ice infilling of the relief, unless its level, the *glacier surface,* had moved across these ridges at some *transfluence passes* and maintained a S-N descent. Disregarding the possibility of such glacier transfluences, even glaciation - which is, after all, never superior to the relief but within its surface - would have had no alternative other than following the direction of the present valley inclination with its *ice flow.* It follows that, in this respect, the twin thresholds of the Aghil and the Kuenlun ridge also continue to hold their own for the maximum glaciation. The description has to start *twice* again at the top, descending from heights of more than 5000 m asl first to 4000 m (Shaksgam valley floor), then to 3300 m (Yarkand valley floor), and finally to less than 2000 m asl (Kuenlun foreland). It should be mentioned in advance there is no lowest glacial margin position between the Karakorum N-slope and the Kuenlun, since the altitude of the entire area is too high. For this reason, reconstructions of the maximum height of *abrasion* and *polishing lines* and *ice levels* were carried out in the valley systems of the Shaksgam and the Yarkand. At that time the *entire* valley relief had been *infilled with ice.* The *lowest* glacier end positions were only formed beyond (north of) the Kuenlun mountains where the foot of the mountain ends in the Tarim basin.

The investigation consequently begins in the Karakorum N-slope as the highest, more than 8000 m-high feeding area, by examining the *highest preserved* of the abrasions and polishing lines, more or less unambiguous evidence of which is found in smaller and larger remnants. Though the local Ice Age glacier levels are the highest, and far above the present glaciers, they also show the *smallest* height difference to the valley floor levels there. Only the main valley floors and longitudinal valley floors further down, which are now *free of glaciation,* manifested the greatest ice thickness during the Main Ice Age, whereas up the main valley, and especially up the tributary valleys the gap between the Main Glacial ice level and the present one has been *closed save* for a vertical distance of a few hundred metres. This observation gives some idea of a Main Ice Age valley glacier net with very well balanced, and by contrast with the present glacier surface inclinations, markedly *flatter,* surface inclination curve. The overall picture of the *Ice Age glacier surface geometry,* which *flattens* the mountain relief in this way, shows that it was underpinned by the *greatest* thicknesses of ice where the valleys are incised most deeply. In other words, that the ice was thinner where the valley floors extend far beyond the *Ice Age snow line,* and far into the present glacier region. Glaciation of peaks and ridges *above* the valley ice level was, if anything, *less* than now, as the *upper climatic glacier line* had been lowered equi-directionally with the climatic snow line (cf. Chapt. 3.1.1; cf. Kuhle 1986i: pp. 344/45). However difficult it may be in a particular case to arrive at an exact classification of remnants of flank abrasions and polishings or even abrasion and polishing limits of the valley flank sections, the *concept of an overall level* of Ice Age glacial relief-infilling succeeds quite well, providing two-, or even three-dimensional *interpolations* of height- levels for intervening areas from both sides, even from the opposite valley flank for more or less great horizontal distances.

5.1 The Maximum Prehistoric Glacier Levels which can be Shown to Have Existed in the Karakorum North of K2 (Catchment Area of the Muztagh Valley)

Beginning with the K2 valley, where the 1986 expedition was able to carry out investigations into the highest valley heads in the Karakorum main ridge, the highest preserved abrasion and polishing limits (----) above the recent glacier feeding areas were found to be running 400–600 m above the present glacier surface (Fig 1a) This is a value which polishing limit remnants in the feeding area of the Aletsch glacier in the Bernese Alps also indicated for the reconstruction of the Main Ice Age raising of the Jungfrau firn, thus confirming the comparatively *minor heightening* of the catchment area levels in question. Higher values (around 600 m) can be recognized at the N-wall of K2 itself, thanks to increased steepness at the base following glacial undercutting of the banded Falchan gneisses. In Fig 5 (---- centre) the *undercutting* is particularly clear at the shadow line of the K2 N-spur. On the W-source branch of the K2 glacier, which leads up to the eastern "Sarpo Laggo pass" (Fig 1a No. 6), the ice thickness during the Main Ice Age had been reduced to 400 m and less as a result of the over-running of the aforementioned c. 5800 m-high pass (Fig 1a ---- between Nos. 2 and 6). In the entire catchment area of the K2 glacier the glacier surface was thus located between 5800 and c. 6000 m asl.

Fig 5 (---- left) shows the reconstruction of the level on the orographic right-hand valley flank. Its continuation down-valley is shown in Fig 6 (---- right). In this exposure the Ice Age increase in the glacier thickness to the S, up to the middle coupe of the present K2 glacier, is very noticeable (---- background). Convex glacigenic abrasion and polishing forms () are characteristic here. At the same time the intervening destruction of outcropping metamorphic sedimentary rock (quartzites and Baltoro black slate, cf. Geological Map of K2, 1:25 000, Desio 1968) by frost weathering, is striking. After the deglaciation of the rock flanks it led both to the *formation of debris covers* with solifluction dynamics (▷) and of young *talus cones* (Fig 35 ∇; 10 △). Weathering is aided by the positioning of the *outcrop layers* up to the slope surface. Fig 1a, 6 and 10 show the maximum height of the prehistoric K2 glacier surface as far as it is preserved at the valley flank on the left side (Fig 6 ---- left side; 10 ---- centre). Towering up to 2000 m above the present glacier surface (Fig 1a No. 4, 6 Nos. 1, 2; 10 Nos 2, 3) this steep, shaded ENE-exposed valley flank continues to be *transformed* by ice and snow-induced rock falls and is forming a dense *network* of *wall gorges* and *gullies.*

Small hanging glaciers and firn shields, some of which developed only in post-Glacial times, contribute to the process of breaking-up the valley wall (Fig 1a, 5, 6 ◇). Fig 7, 10, 11, 28, 30, 32, 35 (----) assist in the reconstruction of the maximum prehistoric ice level at the great *ice barrier* to the N, the Karakorum main ridge formed between the S-slope and N-slope. Eight kilometres away from the K2 peak the ice thickness (----) increased to about 700–800 m. This is the confluence area of the third orographic left-hand tributary glacier (Fig 10 and 11 below No. 4). During the Ice Age and even now the tributary glacier system on the orographic left-hand has in turn been divided into three source branches, and connected with the main glacier (K2 glacier), by a *confluence step.* Relatively steep even during maximum glaciation this *steep step* in the rock floor explains the *ascent of the glacier surface* into this tributary glacier system (Fig 10 and 11 ---- in the background, below Nos. 3, 4, 5, or 4, 5). There is a 7315 (7330) m-high mountain at its source (Fig 1 No. 4); *during the Ice Age* the 1000 m-high scarp of this structure descended steeply to what was then the glacier surface. Two and a half kilometres further down valley another tributary glacier joins on the orographic left; it used to flow into the Ice Age K2 glacier with a 500–600 m greater thickness than now (Fig 10 ---- right, below No. 7; 11 ---- below No. 6), where *exaraction grooves* and abrasion and polishing lines caused by detersion and detraction are recognisable (----) above some glacial *abrasion* and *rock polishings* (). The catchment area of this tributary stream is certain to be higher than 6000 m (No. 7 = 6040 m); indeed, it is likely that even peaks of more than 6500 m (Fig 11 and 12 No. 6; 6540 m according to T. Myamori's map (1978) Baltoro Muztagh in Mountaineering. Maps of the World p. 128) follow on here. Up to now, this area has remained completely unexplored. In spite of this, the Italian 1929 Geographical Expedition described a 6540 m mountain here as Monte Chongtar. Between the two left-hand tributary glaciers, *two levels* of abrasion or polishing lines can be reconstructed (Fig 10 below No. 6), the lower one of which corresponds to the oldest Late Glacial stage, the Ghasa Stage I (nomenclature Kuhle 1982, p. 154). Further down the valley, where the most northerly left-hand tributary valley joins, another *two levels* of abrasion and polishing lines can be made out; this is particularly clear when viewed from different angles (cf. Fig 11 ---- (below No. 7 and up to the right-hand edge of the picture) with Fig 28 ----). Further perspectives of this valley flank with its considerable *roughness* (cf. Chapt. 4. 1. 1.) are shown in Fig 7 and 14 (----). On the orographic right-hand side, directly opposite that tributary valley on the left side, the 8 km-long Skyang Kangri valley joins the K2 valley directly from the E without any steep steps (Fig 11, left third; 30 right). A 1 km-wide tributary ice stream from this valley drained into the main ice stream of the next higher order (Fig 138 No. 3). In the confluence area the abrasion and polishing lines are evidence of an ice thickness of c. 800 m (Fig 2 right, 11 left, 12 right and 30 ----). A diversion into the Skyang Kangri valley up to its upper end, with the 7544 m-high Skyang Kangri, is to facilitate the reconstruction of the glacier level of the Main Glacial period (Fig 2; 12). Directly N of the Skyang Kangri twin- peak (Fig 2 and 12 No. 1; left 7544 m, and right c. 7500 m asl) the Ice Age glacier level was *higher* than the intermediate valley ridge between the Skyang Kangri and the western branch of the N-Skyang Lungpa glacier (Fig 2 and 12 ---- far left). Since the ridge separating the two glacier systems now was *not* rounded, it can be concluded that a glacier transfluence was largely *missing* at that time. The significant *supply of ice avalanches* (↓↓) from scarp walls, which had, and still has, an effect upon not only the valley head but also over many kilometres from the up to 6640 m-high peaks (Fig 2 No. 3; 12 Nos. 2 and 3) down to the orographic left-hand valley flank, has *obscured* the prehistoric abrasion and polishing lines. The preserved remnants of the truncated spurs and polished *triangular slope facets* (Fig 2---- left half) indicate a prehistoric ice level between 6000 and 5600 m asl, corresponding to a prehistoric glacier surface 600 to 400 m above the present one. In the directions of the confluence area of the Skyang Kangri valley with the K2 valley, an *oldest* – ie *highest* – abrasion and polishing surface bordered by an *ice scouring groove* has been preserved up to 600 m above the present glacier surface level (Fig 2---- right half, 12 ---- right half; 30 a ---- bottom right). Now and in the most recent historical period of glaciation (during stages IX and X), the right-hand side of the Skyang Kangri glacier, together with its margin in the form of an *ablation gorge* (Fig 2 □□ ; 12 foreground and IX, right), which are exposed to the SW and thus to increased radiation, have kept their distance from the valley flank, whereas the Ice Age glacier, as the largest tributary branch of the K2 glacier, on the orographic right, attached itself to the walls of this valley, reaching up to a height of at least 600–800 m (Fig 12 ---- on the right-hand edge; 30a ---- bottom right). The K2 main valley glacier continued down valley at a correspondingly high level below the Skyang Kangri valley junction (Fig 10 right-hand edge of the picture; 30 left half; 30a). Fig 10 and 30 both show the two glacial flank smoothings of the main valley (Fig 30) lying *directly opposite* each other, together with the abrasion and polishing lines which run up to 1000 m above the valley floor (----). Due to parallax shift (----), the perspectives of these pictures, which look down on the K2 valley to its junction with the Muztagh valley, render the Ice Age *increase of glacier thickness*, which occurred in this direction, less distinct. Evidence of this was supplied by the author by means of the *loss in height of the valley talweg,* which is greater than *that of the ice scour limits.* Fig 32 and 43 select the area of the junction of the K2 valley with the Muztagh valley in respect of their ice scour limits (----) from opposing observation perspectives. The ice scour limit on the orographic right runs here at about 5200 m asl, somewhat more than 1000 m above the valley floor (□ ◇) in the forefield of the present K2 glacier (Fig 23; 23a). The reason for the absence of moraine deposits from the Late Glacial stage I (= Ghasa Stage, after Kuhle 1982, pp. 154/55) below this ice scour limit, is the persistent lack of height of the snow line (ELA) during the relevant period of the early Late Glacial time. In the same way as the *Main Ice Age* glacier surface (Fig 32 ----) is to be reconstructed *solely* on the basis of the smoothly abraded and polished valley flank rocks (Fig 28 on the left-hand edge; 32) and the *suspension* of these smoothings towards the top by means of the *ice scour limit*, since the relevant glacier surface – far above the snow line – must have been part of the *denudation area* of the glacier, and consequently *no* moraine deposition was possible in these valley cross- profiles of the glacier feeding areas, they are still missing here during the *early Late Glacial* period (stage I). The oldest moraines belong to the *middle Late Glacial*: the Taglung Stage II and the Dhampu Stage III (nomenclature for High Asia after Kuhle 1982, pp. 155–57) (Fig 32 II, III). The moraine deposits following on down valley from stage III become more and more *voluminous*, jut out into the valley space like terraces, thereby *diminishing* its volume (Sirkung Stage IV, Nauri Stage V, older Dhaulagiri Stage VI in Fig 32 and 43). The vertical continuation of the moraine deposits of the valley down to its floor can be gathered from Fig 23 and 23a, showing the moraine stage IX and X (■■).

A little above the junction of the Ice Age K2 tributary stream with the Muztagh valley, the formerly 21 km-long N-Skyang Lungpa glacier joined the K2 valley as the then most *important* tributary glacier (Fig 138 Nos. 2, 3a). Its source was located in two branches running parallel to the Skyang Kangri glacier (cf. above) on the N-flank of the 7544 m-high Skyang Kangri (Fig 138). Fig 32 (---- left third

II, III, IV) shows the highest provable *ice level* including the lateral moraine of the Late Glacial period on the orographic right in this junction area. The confluence point of the talwegs of both valleys is marked in Fig 30a (➤, top), together with *ice scour* limits in the confluences of N-Skyang Lungpa valley, K2 valley and Muztagh valley (----), which are close to one another. During the *Neo- Glacial* glacier advance (c. 4000–4500 YBP, by analogy with datings by Kuhle 1986e, 1986c), the *mountain spur* of the two converging flanks of the N-Skyang Lungpa valley and the K2 valley was *extended* by 1 km (Fig 32 V, V right and V left) through a middle moraine, which was formed by the two lateral moraines of two united valley glaciers. Fig 43 (----) depicts the *highest* provable prehistoric glacier level in the K2 valley – Muztagh valley junction with its gravel floor, which appears as the ice scour limit (---- on the upper edge of the air-photo) in Fig 30a. In the course of the 1986 expedition Xü Daoming found "hard till" at 5000 m asl, and 1100 m above the valley floor in this region (Fig 138 No. 12). It was subsequently dated in the TL laboratory of Gdansk University as $56 \cdot 10^3 \pm 8.4 \cdot 10^3$ YBP.

In order to present the *ice scour limit*, it is now proposed to follow the orographic right-hand flank of the *Muztagh valley* (Fig 43 to the right), from the mouth of the K2 valley up to the Sarpo Laggo glacier (Fig 30a ---- near the left-hand edge; 50 ----). Passing above some *Late Glacial* lateral moraine ledges (III, IV ↓↓), it delimits a heavily scoured, preserved abraded and polished edge ( ) towards the top. The lateral moraines are preserved in places where it was easy to press them into the shadow of the small tributary valley below the 6040 m peak (Fig 50 No. 2). The source branch of the Muztagh valley on the orographic right-hand side, the Sarpo Laggo valley, continues to be glaciated like the K2 valley. It takes the form of a *trough* valley with a *broad* floor and *steeply* ascending abraded and polished flanks (Fig 36, left third; 44 left quarter; 46). Besides sharing the classic *over-steepened* and slightly *concave* abraded lower flanks (Fig 36, far left below), it even has the strikingly *straight* valley flanks, which are typical for many V-shaped valleys, and *smoothings*, which are *direct indicators* of glacial development (Fig 138, No.7; 44). Between the confluent tributary valleys of the Sarpo Laggo system perfectly *triangular* glacigenic *slopes* – in the true geometrical sense – have developed here. Such polished rock surfaces have been formed in many places *irrespective* of the nature of the rock or their bedding conditions. Fig 8 even shows crystalline schists with quartzite layers vertically outcropping on the right flank of the Muztagh valley ( ), which have been *whittled down* to form a *smooth surface*. This is a rock sequence exhibiting a *wide variety* of abrasion and polishing resistances. The main valley axis of the Sarpo Laggo valley can be well surveyed over 30–32 km, up to the 6544 m-high Kruksum peak at its head (Fig 36 and 44 No. 1), where a mountain ridge which was *abraded* and *polished round* during the Main Ice Age, the still glaciated, 5931 m- high Karphogang W of the E-Muztagh pass marks an Ice Age minimum glacier level of about 5900–6000 m asl (Fig 44 ----). This was the location of a central, slightly *dome-shaped* heightening of the ice stream network, which was in contact with the Karakorum S-side through several hundred metres-high transfluence thicknesses including three confluence passes (E-Muztagh pass, 5422 m; W- Muztagh pass 5370 m, and the Sarpo Laggo pass 5685 or 5645 m, directly at the valley head). There have been several of such *transsection glacier domes* along the Karakorum; the author suggests reference to this one as the "Sarpo Laggo dome". In the S it was connected with the Baltoro glacier system, and in the W with the Panmah glacier system. Both of the southern ice stream networks belong to the Main Ice Age catchment area of the Indus glacier (cf. Kuhle 1987h, pp. 606/607; 1991, pp. 297–299). It was lying opposite the Shaksgam-Yarkand glacier system, which includes the ice stream networks in question. Fig 44 shows the direction of origin of the two main components (→←, background), which lead down from the Karakorum passes mentioned above. The arrow (→ left) marks the inflow of the ice stream which the 1929 Italian Geographical Expedition had named the Meridional Chongtar glacier. Being the largest tributary stream of the Sarpo Laggo glacier, it was linked with the K2 glacier during the Ice Age via a continuous surface (ie without any breaks in slope) that extended over the c. 5800 m-high "E-Sarpo Laggo pass" W of the K2 (cf. Chapt. 5.1; Fig 1a No. 6; 138). In this way the massif of the 7315 (7330) m-high peak (Fig 11 No. 4) which includes the 6540 m-high Monte Chongtar (or Chongtar peak Fig 11 and 12 No. 6; 50 No. 1) was *enclosed* by the *more* than 1000 m-thick ice streams of K2, the Muztagh and the Sarpo Laggo glacier. Further higher peaks of this Sarpo Laggo catchment area are the mountains of the Lobsang group, including the 6745 m-high Thyor and the 7275 m-high Muztagh tower, which were connected via the Moni glacier. The *maximum thickness* of the upper Sarpo Laggo glacier which it was possible to reconstruct geomorphologically on the basis of mountain forms and abrasion and polishing grooves (Fig 44 ----; 46 ---- left side) reached 700–1000 m *higher* in the present snow line limit (at 5000–5200 m asl; Fig 138, No.7) than the present ice stream. On the way to the confluence with the Skamri glacier in the Muztagh valley (Fig 36 ----; 46 ---- right side; 50 ----) the glacier thickness increased to well above 1000 m. Here the glacier level was around a *minimun* altitude of 5000–5200 m, whilst the level of the present gravel-floor in the valley bottom (No. –6) lies at 3900–4000 m asl. Twice as broad as the Sarpo Laggo valley, the Skamri valley with the presently about 40 km-long Skamri glacier, (also known as Crevasses or, at times, as the Yinsugaidi glacier) flowing off the Panmah Muztagh, reaches the Muztagh valley from the W (Fig 36 and 46, right-hand side; 9; 47; 48; 49; 138 No. 9). Linked with the Sarpo Laggo glacier in the SE, the 7045 and 7090 m peaks of the Chiring group belonged to the S-catchment area of the Ice Age Skamri glacier. Their marginal satellite peaks are depicted in Fig 36, 46 and 47 (Nos. 2, 3, 4). In the west, as an entirely glaciated area of *great mean altitude*, the Drenmang group, which culminates at 6736 m, *continues* to

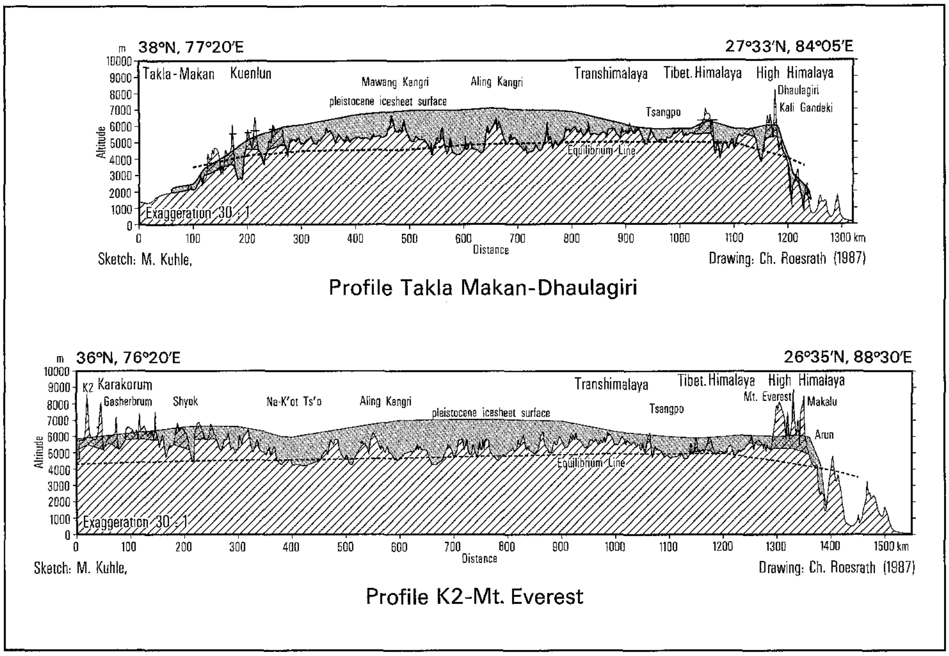


Fig 135 a/b
The two cross-sections give an idea of the Main Ice Age (Würm age) glacier cover of Tibet between a) the High Himalaya in the S and the Kuenlun in the N, down to the Takla Makan desert in the Tarim basin, and b) the High Himalaya and Karakorum, which is also in the north. This establishes the connection between the ice of the Tibetan interior and the particular area under investigation in this paper, which includes the Karakorum, Kuenlun, and the Tarim basin with their contemporaneous ice stream network. Profile /a shows the outlet glacier tongues in the northern Kuenlun foreland, which flow down to below 2000 m asl (cf. Fig 112, 113 and others; also Fig 138, Nos. 43, 44, 55 and 137).

be part of it (Fig 9 and 47 background). The catchment area to the N has its highest point on the 7265 (7295) m- high "Crown" as the highest peak of the Skamri glacier system (Fig 46 and 47 No. 1). By contrast with the K2 glacier system and in view of the *much more extended present* ice surfaces of both the Sarpo Laggo and the Skamri glacier system at *considerably lower* altitudes of the peak catchment areas, it is evident that from a certain minimum height of the relief it is the *valley bottom level* rather than the height of the peaks - but in any case above all the *mean* altitude of the relief - which determines the construction of the present and the large prehistoric glacier areas. The difference from the present glacier system is a *significantly higher* glacier level during the Ice Age. It lay 21 km into the Skamri valley from its junction with the Muztagh valley at 5300-5400 m asl, ie c. 850-900 m *above* the present glacier surface (Fig 9 - - - -). During the Ice Age, the Skamri glacier had a two-pass link - the one further S being 5475 m high - with the Nobande glacier on the Karakorum S-side in the Indus catchment area. The level of the upper ice stream was thus kept *below* 6000 m. Fig 47, 49, 46 and 36 (in that sequence) show the vertical distance of the *Ice Age abraded* and *polished ledges* ( ) up to the *polishing grooves* and *ice scour lines* from altitudes of 5400 to 5200 m asl downwards (- - - -). They are indicators of a *minimum ice thickness* of about 1000 m. It is likely that the ice thickness was *much greater* still, and that for two reasons: 1: Due to its *short-term* effectiveness, which must be assumed here, the *maximum height* of prehistoric glacier abrasion and polishing was hardly sufficiently marked in geomorphological terms to be preserved; 2. The valley floor level on which the assessment of the thickness is based, lies *too high*, as it is *still upgrading* now (Fig 9, 36, 46, 47, 48, 49 No. –6 □). The *box shape* of the Muztagh valley, which is a result of *infilling of the valley floors* with loose rock (moraine and glacio-fluvial gravels), suggests a far more than 100 m-lower altitudinal position of the *rock bottom* when it was abraded and polished by the Main Ice Age glacier. This approach is justified in view of the abraded or polished valley flank slopes striking the valley floor gravel at an *acute* angle, or *their continuation* below the loose rock infilling (Fig 9, 36, 46, 47, 49 each of them near the right-hand edge). Fig 47 shows with classic clarity (in the centre, below) the *interference* of steeply joining, hanging V-shaped valleys and the polishing of main valley flanks into *glacial cuspate surfaces* (cf. also Fig 49). In summing up this section, it must be said that a *1000-1200 m-thick ice stream* has left the Muztagh valley, ie reached the junction with the Shaksgam valley, the main valley of the next higher order.

5.2 The Reconstruction of the Maximum Ice Thickness between the Karakorum and the Aghil Mountains (the Ice Age Glaciation of the Shaksgam Valley)

The Shaksgam valley is the major northerly longitudinal valley of the Karakorum, and thus one of the original branches of the Yarkand valley. It does not only *drain* the central and western Karakorum, but also drains part of western Tibet. The Shaksgam valley directed the *Ice Age* glaciers and their melt-water run-off north to the Tarim basin, one of the interior basins of Central Asia in an *areic* and arid environment, whilst the Shyok and Nubra valleys, the two transverse valleys of the eastern Karakorum, took theirs into the southern slope and down to the Indus valley. The subject of this chapter is the central section of the Shaksgam valley (Fig 138, No. 24–11); extending beyond the junction with the Muztagh valley (Fig 37, 51, third on the far left), it is 125 km long. The side valleys and present glaciers of the upper Shaksgam catchment area have already been mentioned above concerning the glacier lake outbreaks (Chapter 4.5.1.1) (Fig 138, No. 47). The uppermost catchment area of the Shaksgam valley includes the northern and eastern flanks of two more than 8000 m-high and six more than 7000 m-high mountains. In descending order, or from east to west they are Apsarasas (7245 m), Teram Kangri (7462 m), Sia Kangri (7422 m), Urdok (c. 7300 m) (Fig 70 and 116 Nos. 6, 5, 8, 3, 4; Fig 83 and 84 No. 1), Gasherbrum I (8068 m; Fig 116 No. 1), Broad Peak (8048 m or Gasherbrum II 8035 m; Fig 116 No. 2) and Skyang Kangri (7544 m, NW aspect, Fig 2 No. 1). Much *more effective* for the Ice Age feeding of glaciers than this uppermost catchment area were the *extensive* plateau spurs between the upper Shaksgam and the upper Yarkand valley in Western Tibet. Evidence of this is present in *high valley floors* between 4000–5000 m asl and superimposed summits like those of the Aghil Mountains, ranging from 6000 to almost 7000 m without too *great* relief energies and vertical distances from the highest points down to the talwegs (Fig 138, lower section corner on the right-hand side). The areas continue to be glaciated – though only in the form of *intermittent* small-scale mountain glaciation. At the time of the Ice Age the *snow line* (ELA) (cf. Chapter 6) *fell below* the *median* relief level with the result that there was a rapid *qualitative change* in the growth of the *feeding areas* thanks to a relief-specific self-reinforcing effect in the form of self-increasing *ice covers,* ie *plateau-ice* formations like those the author had also been able to detect in Central Tibet (Kuhle 1988, p. 566, Fig 9 II; 1991d, p. 144 ff, 211). Returning to Chapter 5.1, the author draws attention to the glacial *forms of identification* in the area shown in Fig 30a to the N (upwards) and those of Fig 32 and 43 (to the left) towards the exit of the Muztagh valley. The area of its confluence with the Shaksgam valley is marked by a calcite crossbar mountain. It is rounded, and has all the features of a *glaciated knob* (Fig 52a , 138 No. 11). The glaciated knob is about 150 m high, covered with *scatterings of ground moraine* and by moraine material deposited along the edges (■). (Left of the right-hand ▷) the Shaksgam river undercuts the glaciated knob at a rock cliff which has become an escarpment. The base of the glaciated knob set into sediments from the glacio-fluvial pebbles of the high-water bed (□). In the place where the left Shaksgam valley flank joins the right-hand one of the Muztagh valley, a *glacial horn* has been formed, ie *a total* wrapping of ice has created a pointed open peak (Fig 52; 138 No. 13). *Composed of calcite* (Fig 117), the horn is 4730 m high, and has preserved its glacially polished form including the *finely smoothed* polishing ( ). With a valley bottom at an altitude of 3800 m asl, the horn is evidence of a *minimum ice thickness* of 1000 m. Being broader up- valley the mountain ridge between the two valleys provides a good overview across the *macro-forms* of the Muztagh – Shaksgam confluence (Fig 51). The glaciated knob described above (Fig 52a) marks the point of the confluence (Fig 51 in the third on the left between ▼ and ■). Smoothed by flank polishings ( ) the main Ice Age trough with its almost completely retraceable ice scour limit (- - - -) extends down valley. Fig 51 (in the foreground and middle-ground) shows the polished calcite rocks of that glacial horn (Fig 52 No. 4) up to its peak at 4730 m asl (Fig 51 No. 4). Besides *classic convex forms of flank polishing*, which cut here steeply outcropping layers of limestone (foreground, left-hand side), striae polishings on outcrops (on the right, below No. 4) and glaciated knob-like polished ridges have been preserved at altitudes above 4600 m which provide evidence of the ice confluence over to the Shaksgam valley (Fig 51 far right; cf. Fig 38). Late Glacial lateral moraines are represented by at least two ledges on the orographic right-hand valley slope (Fig 51 ■■). At least 1000 m thick, the *ice covered* and *reshaped* the entire spur of this intermediate valley divide (Fig 51 - - - - above No. 4).

What remains is the question of the *direction of flow* of this glacier during the Main Ice Age, ie whether ice emanated only from the Muztagh valley, or did the Shaksgam valley hold a corresponding glacier at the time. To start with the *similarly significant* altitude of the catchment areas of the two adjoining feeding areas points to the latter. Fig 38 shows *classic* forms of glaciated knobs in calcite, which occur on the *transfluence pass* between the Muztagh and the Shaksgam valley at 4500 m asl ( ). Since their steep *lee-side* points to the Muztagh valley on the left, a dominant ice transfer from the Shaksgam to the Muztagh valley must be assumed, at least during the final phase of their reworking. This implies a *larger* ice filling, and consequently greater flow pressure in the Shaksgam valley. Evidence of the still *very considerable ice burden* at the level of glaciated knobs 700 m above the valley floors of the two adjacent valleys (Fig 38 □, on the left) which has created these forms of polishing is found in their *finely chiselled* polishing. It can only occur in polishing dynamics where there is a *film of water* underneath the glacier ie in the case of a *temperate* glacier bottom. This can only be achieved by so substantial an ice burden in conjunction with intense flow dynamics that melting point pressure was reached. In order to understand the soft, *finely chiselled* form of these glaciated knobs at very high altitudes, *enormous glaciation* together with the *highest flow velocities* must be assumed. Another variation to be considered in the explanation of these forms would be that of an only *moderately thick* ice over-flow during the Late Ice Age by a merely *temperate* glacier tongue – it leads however to the following dichotomy: there has either been an overflow, with *cold* ice at this altitude, since there is no other way for a sufficiently large ice filling the major valleys on both sides of the glaciated knobs, or the ice was warm (temperate). This, however, would require an already *substantial vertical distance* to the altitude of the *snow line* at the time (during the 1986 expedition a temperature of −6 °C was obtained at the level of the snow-line in the K2 glacier of an ice depth of 10 m, while the annual mean temperature was found to be −10.1 to −12.3 °C (cf. Chapter 3.2) – both evidence of a cold glacier) which argues *against* a sufficiently high glacier filling of the valleys. If, however, a sufficient glacial filling of the two major valleys is assumed for the time of an ice overflow during no more than *minor Late* Glacial snow line depression, it would speak for a much more substantial glacial filling and overflow thickness at the time of the *Main Ice Age,* so that *both* ways of explanation – no matter which side one comes to – speak for the previously stated *enormous glaciation* of the entire extreme high mountain relief with consequently *extreme flow* velocities. A *direct* indication of the considerable ice thickness immediately above this transfluence pass (Fig 138 No 12) is given by the calcite rocks (Fig 37 ▲ to the right of No. 2) which are rounded by glacigenic flank polishing at altitudes of up to 5200 m asl (Fig 37 ----). The level of the ice was probably even higher in the Main Ice Age (cf. Chapter 5.1: observations on the prehistoric Skamri glacier). So far, however, it has not been possible to obtain direct evidence of this. In correspondence with the two Late Glacial moraine walls on the orographic right-hand Muztagh valley flanks (Fig 51 ■■) there are also several *Late Glacial moraine walls and ledges* deposited at approximately the same level this side (north-east) of the transfluence pass on the orographic left-hand of the Shaksgam valley (Fig 37 ■■). These glacigenic deposits contain *polymict* substrate consisting of blocks of calcite, dolomite and gneiss. Even higher above, in the culmination area of the transfluence pass, the author himself has found up to 1.5 m long *gneiss blocks* (Fig 118) on polished limestone rocks of those glaciated knobs (Fig 38a ▲) (sample No. 19.10.86/1). These *erratic* blocks have been transported there, over kilometres, even tens of kilometres, from the gneiss areas of the Karakorum main ridge by the valley glaciers of the prehistoric ice stream network. *Dolomite rocks,* too, are found as widely scattered *erratic* blocks on the transfluence pass between 4450 and 4950 m. They lie on top of the calcite bedrock (Fig 117). There is solid dolomite at a short horizontal distance, as on the right, northerly, side of the Shaksgam valley, where the superstructure of the Aghil mountains (Fig 37 left-side) consists of dolomite. The exact origin of these erratic blocks of gneiss and dolomite eludes *precise* definition. Nonetheless, finding them amounts to confirmation of the ice transfluence across this pass between the valleys of Muztagh and Shaksgam, and thus at the same time, of an *ice thickness* of at least 1150 m (3800 m to 4950 m asl). The *polishing lines* at 5200 m asl (Fig 37 ----) even prove ice levels of 1400 m above the floor of the Shaksgam valley. Up to c. 5200 m asl even the dolomite outcrops of the orographic right-hand Shaksgam flank in the cross-section of the transfluence pass show glacial polishing in *decreasingly preserved quality* (Fig 37, left side 72, 85, 120 ▲). The quality of preservation of flank polishing decreases *rapidly* towards the highest polishing line (----), so that an *older* (higher) and a *more recent* polishing line can be identified as belonging to the main, or as the case may be, Late Ice Age (Fig 85 ---- above and ---- below). Fig 85 shows the glaciated knob in the exit area of the Muztagh valley (▲ far left, cf. Fig 52a). It is in a comparatively better state of polish preservation than the Shaksgam flank of the same valley cross-section profile (Fig 85 and 52a). The reason for this difference is more *intense* post-glacial *reworking* due to the far *higher catchment areas* of the valley flank slopes. Fig 73 shows the 4730 m high glacial horn from a more up-valley perspective (No. 4, cf. Fig 51 and 52 No. 4) and on the left the transfluence pass (Fig 38, 37 and 51 right-hand) described above in topographical context. Fig 76 and 119, adopting the point of view of Fig 73 (Fig 138, No. 20), follow the orographic left-hand flank of the Shaksgam valley with its glacial polishing (▲) up the valley, and show the Ice Age glacier surface (----); on the orographic right-hand the flank polishing (▲) and polishing levels (----) join up with Fig 72. Another few kilometres up the Shaksgam valley the *convex* flank polishings which predominate on the orographic right- hand (Fig 37 in the two-thirds of the exposure on the left and Fig 72, 73 ▲) will be making way for *concave* flank polishing (Fig 120). This is a scarp section in the abrupt left-hand bend of the valley (Fig 138 No. 21; 83 right-hand third ▲ in the background). Opposite this glacigenic scarp a currently still glaciated orographic left-hand side valley from the 6210 m-high summit of the Karakorum stack joins the Shaksgam valley (Fig 27), showing the *mixture* of very abrupt forms with rounded, glacial flank polishings typical of steep side valleys ( ). The adjacent up-valley chamber of the main valley is shown in half section of Fig 83 (right- hand half of Fig 138 No. 23) and presented in semi-profile in such a way as to give a clear image of the *highest* preserved polish line (----) c. 1400 m above the pebble floor (4100 m asl, Fig 84 □, right side). Fig 39, 69, 71 and 82 show flank polishing (▲) on dolomite outcrops ("glacial band polishing" = "Schichtkopfstreifenschliffe" after Klebelsberg 1948, pp. 338–340) from above and gullies, side valley cuts and post-glacial fan and cone accumulations (∇● × ■◆) stand out. In spite of gullying and disintegration of polished areas the glacigenic valley flank has remained almost *intact.* Apart from pure polishing this flank is an example of the feature known as *glacigenic capping* (Fig 82a ▲). Here strata and outcrops have been hit *diagonally* by the glacier so that the resulting abrasions created polishing areas *discordant* with strata and clefts. Fig 121 shows a section of the orographic left-hand valley flank (Fig 83, centre) in shaded neutral light, so that the *uniformity* of flank polishing brought about (▲) stands out clearly, whilst Fig 83 (at wall pillar left of ▲) highlights the *roughness* by casting a shadow. Gorge-like and sharply incised, the side valley on the Karakorum side (Fig 80), where the vertical distance from the summit (No. 7) to the floor of the Shaksgam valley is 2000 m, shows in an effect *extremely great vertical distance* can exert against the preservation of glacier polishings, though the horizontal distance may be negligible. Leaving behind steep gorge walls, the resulting *linear erosion* includes extreme dissolution of polished flanks, the relicts of which remain visible in nothing but valley-shoulder degradations (▲). The glacigenic flank forms which continue south-eastwards, up to the Shaksgam valley (Fig 79; 83 left side and a few kilometres further up-valley 70; 84) remain separated *like boxes* by an approximately 1000 m-wide pebble floor (□). Their thickness, which has continued to increase thanks to the most recent *historic glacier pebble floor deposits* (–6), must be added in to the reconstruction of the thickness of the Ice Age valley glacier. The *loose rock thickness* of the valley floor can only be approximately estimated as being 200–500 m, thought its width, including the mur fans (♦) on the fringes even exceeds 1.5 km (Fig 83, 84). The *flank polishings* (▲) reach about 5500 m asl, ie up to the marked polish lines (----), which occur at 1400 m above the valley pebble floor. It follows that in this section of the middle to upper Shaksgam trough a prehistoric ice thickness of 1400 m plus 200–500 m is likely. In the circum-Tibetan mountains of Ice Age High Asia, where – as in this case of western Tibet – *outlet glaciers* of the Tibetan inland ice are concerned, there is evidence of valley glacier thicknesses of this kind in several places, such as the Dhaulagiri Himalaya (Fig 1, No. 1) on the southern edge of Tibet (Kuhle 1982, Vol. I, p. 57, Vol. II Fig 124a) and in the Namche Bawar Massif (Fig 1, No. 1) on its SE edge near the meridional stream furrows on the Tsangpo- Brahmaputra break through (Kuhle 1991 d, p. 189 Fig 64). In the border area of prehistoric valley glacier surfaces arose the 5466 m-high summit of the glacial horn the *Late Glacial* valley glacier had sharpened (Fig 83 and 84, No. 3 "Shaksgam Horn") in the centre of the valley chamber under discussion (Fig 138 No. 22). During the main Ice Age it had been totally covered by glacier ice (---- above No. 3), resulting in its rounded shape in the top section of the summit. Flank polishings are *strikingly well* preserved, wherever – as in the immediate environs of the "Shaksgam Horn" (No. 3) – mountain spurs jut out towards the Shaksgam valley, *narrowing* its silhouette through rock barriers (Fig. 84 ▲▲ between Nos 3 and 1). *Truncated spurs* and *polished barriers* of this kind can also be observed further up-valley (Fig 70 ▲; 138 No. 24) and serve to form valley *chambers* up the Shaksgam valley to the section in which the present larger side valley glaciers coming down from the Karakorum main ridge reach the floor of the main valley (Kyagar-, Teram Kangri-, Urdok- and Gasherbrum glacier; cf. Fig 138 No. 47). In the overall context of extensive glacigenic working and far-reaching polishing of the orographic right-hand valley flank (Fig 71 ▲) there are considerable striking *roughnesses* in the short, relatively steep, V-shaped side valleys (Fig 71 below Nos 1 and 2; 79 background). In particular it is clear how, parallel to slope gradients, post-glacial destruction of the glacial flank polishings visibly increases here (Fig 24) together with *increasing altitude of the catchment area*, as under the c. 6500 m-high summit No. 2. Below steep ice on flanks and temporary snow patches more or less linear – effective mur activities occur frequently so that side valley heads are *carved up* into further small cuts. They are separated from each other by mountain spurs with stretched rock slopes. During the Main Ice Age ice masses of *great* thicknesses (cf. above) from the main valley glacier swept *across* these side valleys and their subsidiaries rounding the ridges of these mountain spurs (Fig 24 ▲▲, reduced to half size). This is the way in which the present day *inter-glacial* morphodynamics, working *along lines* and at *right angles* to the *extensive* nature of the Main Ice Age, conflict with those of the Ice Age. In side valley topographies of this kind glacial and inter-glacial morphodynamics *cancel each other out* (cf. Fig 84, on the left, below No. 2).

In this valley chamber of the Shaksgam trough south of, and below the Aghil pass (Fig 138 Nos 22–23) on the polished flanks *covers, veneers and remnants of ground moraine* are preserved everywhere (Fig 71 ◄; 83 ■, far right and left; 84 ■). In a manner typical for *ground moraines* the outstanding feature of the material composition is the strikingly high proportion of *pelitic* parts of the matrix (Fig 122). This is a consequence of the enormous *friction* caused by the Shaksgam ice stream which, while bearing down with, a thickness of at least 1300 m, must be assumed to have had a relatively *high* run- off speed, thanks to the western Tibetan ice then pushing down from above. The ground moraine detail from Fig 122 is below ◄ in Fig 71 and close to the location (■ far left) in Fig 83. In addition to the ground moraine material, which has been transformed into *earth pyramids* to a greater or lesser extent (Fig 83 ■, right and left; 71 ◄) remnants of the *sharp upper edges* of the *lateral moraine*, where the moraine came to a halt on the slope, have been preserved (Fig 84 ▼▼, left half of the photo). Such *ledges* occur on at least three different levels on lateral moraine slopes: at about 250 m (▼) above the valley pebble floor (□), and at c. 400 m and 650 m relative height (▼▼). These are bound to be *Late Ice Age* levels of valley glaciers, since the flank polishings (▲) extend more than twice as far, up to the polishing line (----). Only those of Late Glacial glacier surface levels are preserved, though, as they remained *below* the accompanying snow-line (ELA) in this valley cross-section. The formation of a *lateral moraine crest ledge* can *only* be explained by way of this conclusion. On the basis of the lateral moraine ledge even an – in part – more recent than Main Glacial period age of the adjacent lower ground moraine slope coverings must be assumed (Late Glacial). At the time of that *Late Glacial* glacier surface at 4750 m asl (see above) the ELA must have been at about 4800–4900 m asl, ie only 300–400 m lower than now. Desio (1936) thought this level of the Shaksgam glacier to have been the highest during the Ice Age, insofar as he regarded an ice thickness of 500–600 m as a distinct possibility in the area of the Muztagh–Shaksgam confluence. According to the findings of the author, however, the level belongs to the late (older) Late Glacial period.

Mason (1930 p. 263) assumes a *transfluence* of the Shaksgam glacier through the Aghil pass, which leads Mason to the conclusion of a confluence of an upper Shaksgam glacier with the Urdok glacier; the author *fully agrees* with this view, asserting its validity even for periods of much smaller thickness of the Shaksgam glacier, considering that the present Urdok glacier continues to reach the Shaksgam valley floor.

In the following the *transfluence* of ice into the adjacent northerly Yarkand valley system will be discussed in view of its significance for the thickness of the Shaksgam glacier (Fig 138, No. 25). The rise on the orographic right-hand flank of the Shaksgam valley up to c. 4900 m asl alters the *perspective* to the upper polishing lines (----), as they are shown in Fig 83 and 84 from the valley floor, and shows clearly how much higher, ie up to c. 5500 m asl, the level of the Shaksgam glacier must have been (Fig 116 ----). From here – that is, from a short distance above the Aghil pass (4863 m asl) – the Karakorum main ridge can be surveyed from Gasherbrum I (No. 1, Hidden Peak 8068 m) to the Terram Kangri group (Nos 5 and 6 at 7400 m) and *below* the Shaksgam trough with its perfectly preserved glacial flank polishings ( ). *Below* the pass depression on the Aghil pass proper (below 4863 m asl), the flank polishings of the right-hand Shaksgam trough flanks turn off to the west, ie to the left, and follow the main valley. *Above* they continue the upper main valley axis towards the NNW, across the Aghil pass and into the "northern Aghil pass valley" (Fig 86 up to ----). This *transfluence pass* has a wide *trough-shaped* cross-section. On the left it is cut out of massive limestone (Fig 86 , centre and left side) and on the orographic right-hand out of granite (far right). The highest point of the pass is *covered with ground moraine* containing nests of granite blocks (||).

5.3 Glaciation and its Maximal Level between Aghil Mountains and Kuenlun (the Glaciation of the Yarkand Valley System during the Ice Age)

Beyond (ie seen from the Shaksgam valley north of) of Aghil pass (Fig 138 No. 26) there is a highland valley, the "northern Aghil pass valley" running to the Surukwat valley, a large side-valley of the Yarkand valley. A few metres below the top of the Aghil pass (Fig 86) there are two small *pass lakes* occupying two glacigenic *over- deepenings.* Still further beyond the Aghil Pass (Fig 87) there are *glaciated knobs* on the orographic left-hand and *classic* extensive flank polishings (in the two middle quarters of the photo) forming the *highest* polishing line (----) in *abrupt* contrast to roughnesses and rubbings higher above. Unless visible to the far right and far left of Fig 87, the corresponding right hand valley flank is presented from another perspective, showing its flank polishing in a bird's eye view (Fig 90). They are formed in granite, with a surface roughened by weathering with coarse blocks. The glacier polishings have been applied to characteristic *triangular slopes* or *glacial slope facettes* across the curb exits with truncated spurs of the side valleys. Flank polishing (----) reached the level of the spur summit here at 5500 m asl, and 800 m above the valley floor. Down valley (Figs 89 and 88 ----) corresponding with the valley gradient, the Ice Age glacier surface must of course be assumed to dip towards the NNW or N, though with a shallower gradient than the valley one, since *ice-stream net surfaces* also always grow towards the cuppola form, which is not in entire agreement with the basal run-off gradient of the underground. The cuppola form was always subject to *trough formation* in the direction of the great ice run-off, the Shaksgam and Yarkand valleys, but to heightening in the areas of the main ridges and the high level valley floors of the mountains. One of these *central cuppola areas* was the Karakorum main ridge area (cf. Chapter 5.1), a second the Aghil mountains north of the Shaksgam valley with the "northern Aghil pass high-level valley" under discussion (Fig 138, Nos. 25–27). Ice scour ledges and glacial flank polishings (Fig 88; 89 and 90) are evidence of the *valley being almost totally filled with ice* (----) up to the cross- section, with a valley floor descending from 4200 m asl. The valley floor is composed of coarse block ground moraine (Fig 89 ◇), modifying super-imposed mur fans in many places (Fig 88, 89 and 90 ○○). More Recent to Present glacio-fluvial pebble floors (□) are inserted into these accumulations. On the orographic right-hand, ranging from 200 m to a mere 20 m, Late Glacial *lateral moraine terraces* are preserved (Fig 88 ■ background and foreground; 138 No. 27). At the lower end the confluence area joins an orographic left-hand side valley (Fig 138 No. 28; 91) which runs east from the still considerably glaciated, 6750 m-high Aghil massif (Fig 22 No. 5) down to the Surukwat valley. The valley floor is the area of confluence at about 4100 m–4200 m asl; its *glacio-fluvial* sediments have been previously presented (see Chapter 4.5.2). The main Ice Age erosion forms are *classically* preserved here (Fig 91). Two *glaciated knobs* of quartzite rock were preserved (on the left), including their *polished* surfaces. A glacial horn rises in the confluence area like the back fin of a perch (centre; Fig 138, No. 28). On the orographic left-hand glacial triangular slopes ie truncated spurs (right) are preserved. There is an Ice Age moraine remnant below (◆). A reconstruction on the basis of these indicators, the prehistoric *minimum ice level* (----) is up to 1000 m above the pebble floor (---- in the background) at 5100–5500 m asl. Four kilometres down-valley a glacial U-profile with concave rising polished flanks has been formed in the bedrock granites (Fig 123). Other briefly connected side valley bays supplied Late Ice Age packs of rough blocks in the form of *pedestal moraines* (Fig 123 ■), which were deposited on the talweg of the Surukwat valley (Fig 138, No. 29). Down-valley and below the *Late Ice Age* ice-marginal location of the main valley (Fig 138, No. 29) which has been established on the evidence of *end moraines*, the Ice Age trough has been filled with moraines, and above all by glacio-fluvial drift floor *pebble terraces* (Fig 92, Nos. 1–3). On the orographic right-hand flank of the Surukwat valley flank polishing ( ), though *substantially affected* by falling stones and avalanches, show up to levels of at least 5000 m asl, where a supreme polish line (----) is clearly visible. A side valley on the right (below No. 1) serves as a short connecting link between this valley chamber (Fig 92) and

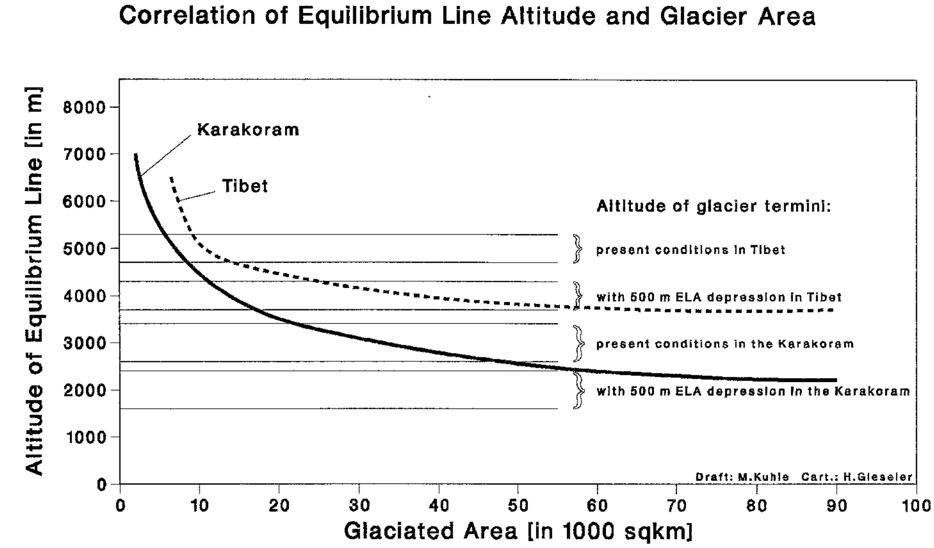


Fig 136
The diagram shows the increase in glacier areas in the particular area under investigation (in the Karakorum) and in the ice areas of the interior of High Tibet, which had been in contact with the prehistoric Karakorum ice stream network in the east. For this purpose a depression of the snow line (ELA depression) of 500 m was selected as an example - an ELA depression which had indeed existed in Early Glacial periods of an initial ice age, and again in the Late Glacial period of the last ice age (Würm). The graphs show that even during such a comparatively small depression of the ELA (the snow line depression of the Main Ice Age, however, amounted to c. 1300 m - see ongoing text) an ice stream network of c. 100,000 km^2 formed in the Karakorum, or an ice cover maintained in the interior (Tibet). Although the two ice formations were connected (with the valleys of the Shaksgam and Yarkand draining the W-Tibetan outlet glaciers) the different conditions of their reliefs became clear: the difference between the altitudes of the Karakorum valley floors and the Tibetan plateau was and is being compensated by the then and now lower ELA. In other words: though the altitudinal difference between the two graphs is c. 1000 m, they are approximately parallel, and striving for the same increase in glacier area.

the in this place c. 6300 m (Fig 22, No. 3) high Aghil main ridge with its slopes and minor valley glaciers (Fig 138, No. 29). The *Late Glacial* confluence of main and side glaciers is established through the evidence of the highest lateral moraine terraces (■). The older, pre-Late Glacial, ie *Main Ice Age* confluence occupied the entire silhouette of the valley. It left no lateral moraines, because the ice surface was *higher* and the ELA *lower* than in the Late Glacial age. When the ELA runs *below* the surface of the ice stream network, *neither* formation of lateral moraines *nor* moraine restriction of the valley cross-section worth mentioning can take place. The valley is closed off by a Late Glacial age *gorge* which is cut into the ground of the trough (cf. Chapter 4.5.2). Down valley *ice cave drift floors* set in; they are several hundred metres thick (Fig 138, Nos. 30/31; 26 ▽). Two kilometres outside it is joined by an orographic left-hand side valley, which supplied the Surukwat valley with another ice stream from the 6750 m-massif, more exactly from its 6125 m-high northern tributary peak. Its left side *ice scour limit* (----) is shown in Fig 26. The summit above (Fig 26, No. 1, 6094 m) is another, still glaciated feedingpoint of the catchment area of this Ice Age tributary stream. Fig 22 (▲ far right) shows metamorphites which this tributary stream has polished from the *confluence level* down into the main Surukwat valley. Moving up the main valley (location: Fig 26, dark corner bottom right) the author has found perfectly preserved *glacier striae* some decametres above the present valley drift floor (Fig 40, 41, 93). These glacial *polishings* (Fig 41 ▲) with extensively preserved sheafs of smaller (↓) and larger (Fig 40 and 41 ↑) *glacier striae* converging and diverging at acute angles occur in solid quartzite rock between 3700 and 3790 m asl (Fig 138, No. 46). The rock surfaces show crests of iron-maganese. Splintering is rare in this very hard rock, where breakage occurs through the grains, whereas larger, more often *concave* than convex *mussle-breaks* (Fig 93 ←), the fault scarps of which point downhill towards the ENE, dissect and terrace the polished areas. In addition, there are lunate fractures, crescentic gouges and crescentic fractures (Fig 93 ↙, above) as well as more or less symmetrical round to oval "chattermarks" (↙↙ below) with crumbling edges (cf. Flint 1971 pp 95/96). The reason for the *excellent* state of preservation of polishing and striae is, apart from the very resistant rock, to be seen in the *extensive covering* with moraine material up to the Holocene. Some of it is still in place (Fig 41 ■). On the same valley flanks, polished areas have formed *discordantly* to the steep to inclined rock strata, and been given a soft-wavey resurfacing by *sub-glacial*, and therefore especially *fast-draining* melt waters (Fig 124 ▲). This discharge far *above* the valley floor (□) with the talweg is evidence of a prehistoric aquifer through a *sub-glacial meltwater tunnel*. Its effectiveness in the kind of *lateral erosion* for glacial formation of *trough valleys* was recognized and described by Tietze (1958 and 1961) on the basis of studies of sub-glacial aquatic erosion in Scandinavia more than 30 years ago. The same may be observed in numerous places in the Alps (like the south slope of the Bernina; Valais, Matter valley head, approach to the Boden- and lower Gorner glacier). In parts these glacial work surfaces inside the rock also show deposits of *boulder clay* (■) and of uncovered, at times rounded blocks (X). To facilitate a better insight into the overall topographical context of the Surukwat valley the location of the *glacier striae* in question is shown in Fig 22 (▼ far right). Turning further ENE, down the Surukwat valley (Fig 22; 138 No. 48), the marked Ice Age level of the glacier (----) is almost completely preserved in the form of orographic right-hand *flank polishings* (◓) and rounded intermediate valley divides between the side valleys. Here, at the junction of V-shaped side valleys, which interrupt the continuous line of valley flanks on the right, glacial cuspate areas or, better, *truncated spurs* have been prepared for their typical shaping by flank polishing. The heads of the short side valleys, which descend from the c. 6300–6500 m high Aghil ridge continue to be glaciated up to valley glacier dimensions (Fig 22, □ small). (Only the summits Fig 22 Nos. 3 and 5 have hitherto been mapped with approximate altitudinal data of 6300 and 6750 m asl shown in Fig 138; even among the few semi-nomadic herders in this area all these peaks have remained *unnamed*). Late Glacial lateral moraine terraces (■■ below No. 3) have only been preserved over a distance of 1.5 km on the orographic right-hand c. 700 m above the valley floor at about 4200 m asl. They belong to the Ghasa Stage (I), ie to the early Late Ice Age. The moraines have undergone *substantial solifluidal* re-shaping by powerful block-flow tongues (Fig 22 ↓), a fact that amounts to an indication of *permafrost*. The present *lower permafrost line* ought therefore to be below 4200 m asl. During the main Ice Age the ELA evidently fell far *below* the glacier surface, as there are no lateral moraines. In the main Ice Age the mapped area of the ice stream network (----) was at 4500–4800 m asl, ie a good 1000 m above the present valley floor filling. This glacier surface is confirmed by the orographic left-hand glaciated knobs and flank polishings (Fig 22 ◓ fore- to middle ground). Fig 94 represents a detail of the right- hand flank of the Surukwat valley (cf. Fig 22 above No. 3, far left) with a *polished* and *rounded* triangular slope in outcrops of red sandstone (◓).

Set down against these flank polishings on solid rock like textbook examples, *ground moraine covers* (■) can easily be discerned in this locality. Constructed from larger components in finer intermediate masses, these glacial *diamictites* form lighter protruding edges on slopes, the modelling of which is to be attributed to *exaration* by the glacier bottom when pushing across the plastic material. The origins of the gullies, which have since formed along the slope fall-lines go back to post-Late Glacial deglaciation. In Fig 22 (near the left-hand edge of the photo) the orographic right-hand (eastern) original branch of the bilateral Surukwat valley within its glacially *abraded* flanks can be seen up-valley (Fig 138, No. 32). Above the drift floor (X) and Late Glacial drift floor terraces (▼) described above (see Chapter 4.5.2) is an area of confluence with the previously discussed original western branch (Fig 95) showing these cuspate sandstone areas (◓ right side of the photo) and orographic right-hand flank polishings of the eastern Surukwat branch (◓ left third). The *flank rounding* of the mountain's spur in the intermediate orograhic right-hand area of the confluence (Fig 95 ◓◓ far left) is particularly rich in indicators of glacier activity. Relevant indicators are provided by *live edges*, which "nibble" at the diagonally laid out *glacially abraded roundings* through under-cutting and resulting crumbling (below ◓◓, on the far left). In a different light this *glacial undercutting of the escarpment* (Fig 125 below ◓◓, on the left third of the photo) becomes even more evident, as it appears to be slightly concave, thanks to the way the shadows fall: a relatively acute live edge is swinging up to smooth glacier polish above and comes to a halt (Fig 95 and 125 ◓◓, far left). From a perspective 4 km further down the valley (Fig 126) the entire valley cross-section immediately below the two original branches (the eastern and the western ones) of the Surukwat valley comes clearly into view as that of a classic trough, and the corresponding orographic left-hand flank of the valley with its rock polishings and glacier abrasion (◓◓ right-hand side) in crystalline schists (phyllites) stands out well.

Reconstruction of the *maximum* glacier levels for this valley section (Fig 95, 125, 126 ----) reach up to 4000 m asl, ie c. 900–1000 m beyond the present - 1 km-wide, including mur fans and drift terraces (▼▼) on its edge - drift floor (X □), so that an ice-stream with a thickness in excess of at least 1000 m must be assumed. Flank polishings which improve *rapidly* towards the lower levels are a clear indication of more recent, ie *Late Glacial* glacier levels (Ghasa Stage I?) (Fig 97 ---- far right). The Yarkand and Surukwat valleys merge at the place known as "Ilik" (Fig 96; 138 No. 33). The drift floor (□) is at about 3450 m asl there. In the area immediately adjacent to the confluence of the two valleys there are some more or less well preserved *glaciated knobs*, which have been shaped out of easily weathered, upright phyllites (Fig 96 ◓ centre, below). Some of these Ice Age polishing forms (◓) were dressed in Late Glacial, glacio-fluvial drift (▼), or at least *surrounded by sediment* at the base (Fig 99 and 100 ◓▼ below, 101 ◓▼, left). From the Holocene to the present day meltwater activities of the Surukwat river, which dissected the Late Glacial drift fillings in a *terraced landscape*, have at the same time peripherally undercut, and partially *destroyed* some of these glaciated knobs. This process is exemplified by the steep wall which meets the glacially rounded dome of the glaciated knob with a fresh active edge (Fig 126 ◓, left-hand bottom; 101 right-hand of ◓ left side near edge). Further down-valley in the same area of the confluence glaciated knobs, which have been undercut in the same fashion as their western side may be observed (Fig 98 ◓◓ right of □). Even the *polish* of the peaks of some of these glaciated knobs has been partially preserved (Fig 42, 127 and 128 ◓↓). This is all the more remarkable as they are easily *splintering* phyllites, with *little* resistance to frost weathering; standing *upright*, they are, moreover, optimally placed for allowing water to infiltrate. Though rather *difficult* to smooth, this glacial polish has evidently even polished these glaciated knobs. However, the majority of the polished areas of the glaciated knobs has been roughened-up and largely destroyed by weathering since deglaciation (Fig 42 and 128 ◓ in the background). Partially still not washed off the polished rocks, a veneer of loam is evidence that the formative basement polishing was accompanied by *ground moraine sedimentation*, or at the final stage was *taken over* for a short while (Fig 128 ⇓). Two aspects point to a relatively short temporal interval to deglaciation: 1. the preservation of polishing despite that extreme susceptibility to weathering, and 2. the *small-scale* and *faint* features etched into the relief of the polished rock surfaces (Fig 42 ↓). Here especially the concave and merely decimetres-wide *ice-scoured* basins (cf. Fig 127 and 128 ↓⬇) are evidence of an extremely *plastic*, and with that the warm underside of the glacier as well as very *careful* scraping during the polishing process, of the sort typically occurring only in the *presence* of *"lubrication"* in the form of an intermediate film of water (cf. above Chapter 5.2). It must consequently be assumed that basement polishing took place in the middle to lower *ablation area* of Late Glacial valley glaciers, which continue to flow together. In order to reach the confluence locality known as "Ilik" (Fig 138 No. 33) from the nearest, and still glaciated region, ie the Aghil ridge in the catchment area of the Surukwat valley with the Surukwat glacier, it is necessary for the ELA depression to drop to 600–700 m (max. 800 m) (presently lowest glacier ends: 4600–4800 m asl; "Ilik" is situated at about 3400 m asl; difference in altitude: 1200–1400 m : 2 = 600–700 m). During the Main Ice Age, however the ELA was more than 1000 m lower (c. 1300 m; cf. below, Chapter 6), so that the thickness of the ice here in the area of the confluence was 1000 m or more (cf. above) and the glacier surface at the valley cross-section of "Ilik" remained *above* the ELA. Evidence for this can be seen in the polished and abraded slopes (Fig 98) and their upper edges, the ice scour limits (– – – –) at an altitude of about 4400 m (cf. Fig 42, 96, 99, 100). Not only the Surukwat valley described above as *trough valley* (Fig 101 inter alia), but also the section of the Yarkand valley above the area of the confluence has a glacial *U-shaped cross-section* (Fig 96 below No. 1; 99 middle ground; 138 No. 34). Below the confluence the Yarkand valley being the main valley, and of greater significance, also shows the *classic* form of a trough (Fig 96, left side; 98). To be precise, it should be classified as a *box-shaped* trough, since the abraded and polished rock floor has been built up and almost levelled by the Late Ice age *glacio-fluvial* drift floor (□). The concave transitional slopes which form the link to the valley flanks higher up, consist of nothing but *secondary* accumulative forms like mur cones (Fig 98 X) and talus slopes (∇). The true, original *primary* trough flanks in the bedrock are preserved as *polished slopes* or *cuspate slopes* and truncated spurs in the rocks of the upper parts of the slopes (). Here the profiles of the slope gradient appear as *flat* to – still higher up – *convex* lines (Fig 98). The latter applies predominantly to mountain spurs in areas of valley confluences (Fig 96, right-hand side of the photo, top). Apart from these *secondary* accumulation there are also well-preserved stretches with previously developed *kame-like* bank formations with scree fans and *moraine core*, which owe their existence to the *Late Glacial* glacier filling of the valley (see above); their opening-up in fresh rips took place after deglaciation (Fig 98 ■). The mountain stream of the Surukwat valley (Fig 101) had to erode itself through the rockbar mountains, which were polished to glaciated knobs in the confluence area described above, and down to the lower level of the Yarkand river (Fig 102 O). Gorge-like and *narrow*, and thus shown to be of *very recent* origin, this cutting in the bedrock suggests itself as Late Pleistocene, and therefore probably as an indicator for *glacial over-deepening* of the lower Surukwat valley chamber – which is also confirmed by the bulging, broad embayed outline of the Recent drift floor (Fig 96 □; 101 X). This linking of *genesis and timing* of Ice Age formation is substantiated by a geomorphological *discordance* of glacier polishing forms and their fluvial destruction: Fig 102 shows convex-concave *glacier polishings* () near the upper edge of the gorge; on the left side (▲) of the gorge walls they have been *incised* vertically. Thus evidently older than the gorge, the polishings are from the Late Ice Age (cf. above), so that the gorge must be of largely post-glacial origin. However, a first gorge of just Recent, Late-Glacial *subglacial* origin suggests itself, since melt-water accumulating under hydrostatic pressure beneath the ice at the end of the tens of kilometres-long valley glacier must have been particularly beneficial to the gorge formation. An immediate indication of *cavitation corrosion* in the course of the formation of the gorge has been found in the remnants of *whirlpools* and related indicators of fast-flowing water (Fig 101 ⇩, foreground).

From the confluence with the Surukwat valley (from "Ilik") through the narrow *trough profile* (Fig 99) of the Yarkand valley and upwards, unambiguously characteristic forms of Ice Age glacier infilling of the relief occur (Fig 138, No. 34). On the orographic right-hand a *polished slope* has been preserved up to much more than 1000 m higher than the drift floor of the valley bottom, ie to c. 4600 m asl or even further up (Fig 99 - - - - background). The *blurred lines* of this ice scour limit can be explained by the ground polish of prehistoric flank glaciation *parallel to the slope gradient*, which must have existed above 4400–4600 m asl. Being at least 400–600 m above the Ice Age ELA there is the risk that the "bergschrund line" (it forms in places where the frozen-up flank ice on the bergschrund turns into glacier ice that flows down the slope according to Kuhle 1982, Vol. II, Fig 113; 1983, pp. 134 et seq) is *mistaken* for the valley glacier level proper. The "bergschrund line" merely marks a flattening of the ice surface on the edge of the valley glacier. Lying lower, the valley glacier level proper, however, is also *above* the ELA, its nearest documentation being the valley glacier polish line, which runs at right angles to the slope gradient (ie almost horizontally). The Late Glacial glacier surface (or one of the Late Glacial glacier surfaces), on the other hand, was *below* the ELA, and evidence of it is found in the glacigenic *accumulations* of typical bank formations (Fig 99 ■). Fig 103 shows the perspective into an orographic right-hand side valley mouth as seen from the same Yarkand valley chamber (Fig 138, No. 34). The maximum altitude of the catchment area of this side valley is just under 6000 m (5994 m, Fig 96 No. 1). It formed a major Main Ice Age Yarkand *glacier embayment* with an ice surface level at c. 4600 m asl (Fig 96 ---- below No. 1; 103 ----). On the left side valley flank, as well as on the slate bar mountain in the centre of the side valley exit there are not only *glacier polishings* (▲ at the centre of the middle ground), but also Late Glacial *moraine deposits* sedimentation (■■) of decametre thickness (cf. Fig. 99 ■). In this mouth of a side valley, this *vertical* change from top to bottom, ie from an *ice scour limit* to *moraine material* deposits in the area of a glacier bank is tangible evidence of the Late Ice Age snow line's rise from the mountain relief in *relation* to the particular glacier surfaces of the time. On the orographic left-hand (Fig 107 ----; 138 No. 49) the glacier polish line, in correspondence with the opposite flank, runs about 1000 m above the several hundred metre-wide drift floor (Fig 103 ▼), so that this, too, makes it likely that Ice Age glacier thickness *exceeding* 1000 m (probably with an additional drift thickness of approximately 1200–1300 m). Here, too, in accordance with the underlying climatic dependence of prehistoric glaciation, a polished slope *without accumulations* (Fig 107 ▲) appears on the upper slope, whereas *ground moraine covers* with polish lineaments and scouring grooves (↓↓) as well as *lateral moraine-like* formations (■■) are deposited on the rocks further down. Descending from a catchment area (No. 1) well above 5500 m, a steep cut (∇ large) joins and breaks through the Yarkand valley flank, its heavily roughened rock slopes in sharp contrast to the latter's glacigenic *polish*. At the same time a *Late Glacial lateral moraine* from the side glacier that flowed from this valley was piled up against the main glacier (Yarkand glacier) (■■, on the right). Already much roughened-up, the flank polishings in the side valley (Fig 107 ▲▲ below No. 1) belong to the same phase of glaciation. *Moraine deposits*, as well *as steep linear erosion* which shaped the talweg-near part of the side valley (∇ large), are both part of a snow line which rose in Late Ice Age times, producing *subglacial meltwaters* and causing that deep erosion. The shape of the debris deposits at the mouth of the river follows the orographic left-hand of the tributary glacier that turns off into the Yarkand valley, leaning up against the main glacier (Fig 107 ∇∇ ,far right).

The confluence area of Bazar Daran (Fig 138 No.50) lies about 9 km up the Yarkand valley. The orographic right-hand side valley is several tens of kilometres long (Fig 129); it is the product of three branches, the catchment area of which is 6136 (No. 1) – 6340 m high, and part of the Kuenlun. Here, too, the Bazar Daran valley chamber (cf. Fig 107 X inter alia) is characterized by at least two generations of mur cones, which fill the valley floor and raise it in places (Fig 129 X). The varied, and in part rounded and *facetted* block material (●) is *dislocated* moraine material from higher sections of the slope, and from the valley heads of smaller side valleys. Following the same principle, the author observed in the side valley down-valley (cf. Fig 103) there is a glacigenic *polished floor* and *trough bottom* in the bedrock below the polish line (Fig 129 ----), the polished slope (▲) and underneath the Late Glacial moraine material (■) which has been fluvially dissected to a depth of 20–50 m (□). There (Fig 103 □) as here (Fig 129 □) *subglacial* meltwater erosion took place in the valley floor at an altitude of about 3750 m asl, and dissected the trough floor. The present drift floor deposits (below ∇) point to prehistoric erosion and a regime *different* from the present, ie a *sub-glacial* one. *Prehistoric* erosion is also indicated by the formation of talus cones and talus slopes (∇) at the foot of the rock base, which is not only evidence of *limited* lateral erosion but also of the *absence* of vertical erosion. *Pebble deposits*, which have been preserved on that rock base, are further evidence of Late Glacial *meltwater activities* (Fig 129 ▼ left side of the photo, on both sides of ■). As far as details are concerned, it must remain open as to which phase of the *melt-down process* their accumulation or *leaching process* belongs, which distilled them from the previously sedimented moraine material and endowed them with their glacio-fluvial character. The general sequence, however, was as follows: 1. Late Glacial rise of the ELA above the glacier surface, together with amounts of meltwater in conjunction with linear erosion *underneath* the ice; 2. disintegration of the body of the valley glacier, accompanied by collapse of the sub-glacial melting vault and, as a result, *filling* of the valley talweg *with ice*; at the same time, moraine deposits are washed out along the edges of the valley glacier and modified to sanders on the bank. At the same time there was also the usual dovetailing with hanging solifluction covers and sediments (Chapter 4.1.1). This second phase of the disintegration of the ice (2.) also includes the island- or *kame-like* build-up of moraines and pebbles in the valley centre, a position that is *isolated* from both the valley flanks (Fig 129 ■). Preserved without any morphological *support* from the valley flanks, such a deposit requires a *"cake mould-like"* enclosure of disintegrated glacier ice for its genesis. Theoretically the only alternative to be taken into consideration is *erosive* stripping in the valley centre by formation of talwegs on *both sides*, but this must be discounted because of the *absence* of a second talweg (behind ■■). Instead there is a flat saddle of loose rock. Moreover, a several decametre difference in height from the main valley talweg below the rock base is too great by far for the formation of a second talweg (Fig 129). A comparable glacio-geomorphological situation with several metres thick glacio-fluvial pebble strata on a rock base, which has been formed by subglacial dissection, was observed in the course of the 1989 expedition to Central Tibet in the large valley SE of the Nyainquentanglha (Kuhle 1991d. p. 164). However, in that case topographical reason would allow a *purely* subglacial formation variant, which has been excluded here. Even in the "Bazar Daran" area of confluence, where Ice Age *glacier thicknesses* exceeded 1000 m (Fig 129 ----), the ice surface (at around 4750 m and more asl) was far *above* the snow line of the time. This is the reason for finding here only blurred polishing lines, ie *syngenetic transformations* brought about by *local hanging* flank glaciation. The true maximum level of the ice stream net surface should accordingly have been noticeably above the one entered here (----). Thanks to the considerably *smaller* gradient of the flanks compared with the Karakorum N-slope described above, the right reconstruction of the *highest ice-scouring* lines in this area between Aghil and Kuenlun is much more difficult, for *groove formation* on valley glacier surfaces along steep flanks is strengthened by *enforced* crumbling and remains more visible after deglaciation as debris covering is absent. Two to three km further up the Yarkand valley (Fig 138 No. 51) there is another valley chamber (Fig 60), the floor of which has two levels (▼ at the very top, and at the very bottom), which in turn permits another reconstruction of Late Ice Age *sub-glacial* meltwater erosion. There is a *polished* or *abraded floor* (▲ ▲, centre) with at least 200 m-deep fluvial cuts and subsequent build-up of pebble floor (▼▼). Beyond this the usual abraded slopes (▲ ▲) of metamorphite (phyllite) rock extend to 4500 m asl (---- below = Late Ice Age) and 5000 m asl (---- above = Main Ice Age) (cf. below). Here the Yarkand glacier had received an influx of ice from two tributary valleys from the 6532 m massif on the orographic right-hand (Fig 60 Nos 1 and 2; 138 Nos 18 and 19) (cf. below). The sub-glacial cut, which transferred that wide rock floor *typical of troughs* (▲ ▲ centre and ▼, very top) into a rock base or denudation terrace is best interpreted as a *polyglacial* formation. During the *Late Glacial* rise of the ELA towards the end of every Pleistocene ice age, the rock base, already rounded by Main Ice Age ground polishing (▲▼, centre top), continued to be residually and steadily more carved out by ever deeper-reaching subglacial meltwater erosion. During interglacial periods – as at present – the rock base makes its appearance time and time again, surrounded by deposited relicts of gravel terraces (▼▼ below and half-right) and *moraines* (▼▼ centre, top). Still significant in Late Glacial times, the thickness of the Yarkand glacier is well documented by orographical left-hand *lateral moraine sediments* for another 2.5 km up-valley (Fig 130 ■) (Fig 138 No. 37). Somewhat flattened here towards the top, the middle slopes are covered by glacial diamictites up to an altitude of about 700 m above the pebble and mur sediment-filled valley floor. The formation of *lateral moraine* ledges with well-preserved ridge lines or crests (↓↓) is clearly visible. *Lateral sander-like*, graded pebble bands are preserved a good 100 m above the valley talweg (▼). The Late Glacial glacier surface at 4500 m asl evidence of which is found in the lateral moraines, was here – as can again be concluded from the moraines – already *below* the simultaneous climatic ELA. Approximately established in this way as running at 4600 m, the ELA was 600 to at most 700 m *below* the present ELA (5200 asl), having already *risen* 600–700 m compared with the Main Ice Age snow-line. The *Main Ice Age* flank polishings (▲) stretch at least 500 m further up than the moraines (Fig 130 ----), thus resulting in a Yarkand glacier thickness of about 1200 m or more. Coming from the south and via a very high and steep confluence step, a *hanging valley* supplied the main valley with a tributary glacier, which left behind more *Late Glacial lateral moraines* (Fig 130 ■■ far left). The opposite, orographic right-hand main valley flank is somewhat dissected by the mouths of side valleys from the 6532 m-high massif of the Kuenlun range (Fig 106 Nos. 1 and 2). When seen from here, as well as observed from a viewpoint further down valley (Fig 60 peak No. 2 equals No. 1 in Fig 106), *two* main valley glacier levels (---- below and above) can be discerned. The *lower* level corresponds to the highest *Late Glacial* level of the immediately opposite valley flank (Fig 130 ■↓), which has been established on the basis of moraines; the higher level (Fig 60 and 106 ----) recurs on both the valley flanks c. 1200 m above the talweg (cf. above). Below the orographic right-hand (Fig 106 ▲) *Late Glacial* glacier work – which is further characterised by largely *accumulation-free* abraded flanks – *lateral moraine terraces* with very fresh, ie well preserved sharp edges (Fig 106 ■), begin. At least two generations, or levels, are to be discerned here, with the now c. 200 m-high *moraine* or *glacier bank formation* (small ■■ ones on the left) appearing from the side valley, whilst the glacial bank formation, which follows the main valley (■■ right-hand) is about twice as high. The following interpretation is obligatory: the recent Late Glacial Yarkand main valley glacier had already melted away from the section of the valley, when the high catchment area of the 6532 m massif continued to let its side glacier flow down to the main valley. Its tongue spread on the main valley floor in a hammer-headed form (Fig 106). After deglaciation several generations of *mur fans* were thus deposited on the main valley floor in this area (X). The *wealth of granite* blocks (○) from the mudflow-like sediments transported out of the same side valley on the right points to dislocated *moraine* material. Dissected *cone sanders* or *ice cave drift floors* that have been transported from this valley and deposited as glacio-fluvial terraces which are now several decametres high, serve as documention of post-Late Glacial and Neo-Glacial to Historic retreat of the side glacier by virtue of their various levels (▼▼ centre, background) down to the drift filling of the present side valley floor. The terraces of *graded* pebbles must not be mistaken for those from *dissected mur fans* on the main valley floor (▼ left, just above X). The latter have *uneven* surfaces. Another 5 km up the Yarkand valley there is another junction on the orographic right-hand side with the tributary valley from the 6532 m and 6008 m-high massif of the Kuenlun (Fig 138 No. 18), the main valley confluence of which has been investigated in greater detail. It is a *hanging valley*, the 150–170 m-high *glacial* granite *confluence step* has been cut into a much more than 100 m- deep *glacigenic gorge* (Fig 61). Started in Late Glacial times, it is now in the process of *subaerial* development (cf. Chap. 4.4). In prolongation of the level of the *trough threshold* or confluence steps, where the incision of the glacigenic gorge set in, the *Late Glacial moraines* from Stages III and IV, which were described above, have been deposited by the glacier of this hanging valley in a kind of *pedestal moraine* (Fig 62 ■, below) while undergoing *syngenetic* dissection by meltwaters (Chap. 4.4).

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 (= 1 950)	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 1 950)	c. 30– 40
XI	= stadium XI		30– 0 (= 1 950)	c. 20
XII	= stadium XII = recent resp. present glacier stages		+0– +30 (1 950–1 980)	c. 10– 20
				M. Kuhle 1993

Tab 3 Glacier stages between Karakoram and Tarim basin from the last High Glacial (Würm) to the present ice margin positions with gravel fields (sander, resp. sander-terraces) belonging to it and their approximated age (s. numbering and figures in the photos concerning moraines and debris fields) (eg III or 2; s. also Tab 2).

As usual, this took place in *two phases* which accompanied the thawing down of the main valley glacier: 1. the ice stream net surface initially was at about 5000–5100 m asl (cf. above), far above the ELA and, as part of the system, a side glacier with the same level was laid down *beside* the main glacier; 2. during the Late Ice Age the ELA rises above the sinking ice stream net surface, and the main valley glacier melts down faster than the side glacier, which has moved *across* the main glacier - partly because of its steeper gradient and consequently greater flow velocity - thus becoming a *hanging glacier.* This second phase saw the *moraine base* being formed (Fig 62 ■, bottom) through the *overthrust* mechanisms of a *ground moraine ramp* (cf. Kuhle 1983, p. 76f; 238f; 1991b p. 89, 104f). 3. As the ELA *continues to rise*, the *subglacial* meltwater erosion, which is cutting into the gorge is becoming so intensive that it *also washes out* the moraine base and contributes to its lowering and dissection (Fig 62 ■ bottom). Dissection *followed* the thawing of the main valley glacier as it presented the relative basis of erosion. 4. From this phase onward the side valley glacier tongue increasingly took over the *eroded debris* of the moraine base, until the side valley glacier tongue *finally* adjusted itself directly to the floor level of the main valley. This was the case ever since the main glacier had *melted away* from the confluence area. During this phase of the *recent Late Ice Age* (Sirkung Stage IV, see Chap. 4.4) the *recumbent* pedestal moraines became *enclosing* lateral moraines (Fig 62 ■■ bottom, left side, two thirds of the photo). As indicators of this development the end moraines of this final phase of the side glacier in the area of the confluence (Fig 62 ■ IV) are *adjusted* to the main valley floor (Fig 62 IV ■ far left). The *minimum* period this section of the main valley was *free from ice* is established by radiometric dating of an organogenic horizon in the valley floor as 4580+/–65 years BP (Tab 2 sample No. 24.10.86/1b; Fig. 108 □). On the orographic left-hand side, somewhat further up the Yarkand, a steep and just under 10 km-long *V-shaped valley* from the Aghil mountain range comes down from the 5880 m-high massif. Set into relatively soft metamorphites with *little* resistance to weathering, the valley had a *sharply* marked talweg. Only some *vaguely recognizable* remnants of polished slopes indicate glacigenic formation (Fig 138, No. 52). Some 1.5 km upstream from this side valley exit, on the orographic left-hand of the Yarkand river, a rock base with moraine blockwork deposits has been preserved which is comparable to the one described above in respect of its height and outline (Fig 60 ▲▲, centre). Like the one 10 km outside the valley, which has been interpreted above, it is regarded as a subglacially *eroded remnant of a trough floor* with a prolonged history of glaciation and deglaciation.

The valley floor of the highest section of the Yarkand valley to be investigated in detail lies between c. 3780 and 3850 m asl (Fig 105; 138 Nos. 38, 39, 36). It is a *classic trough valley*; in the Holocene it was transformed by talus slopes and sloping sheets of debris (▽), on its flanks and by alluvial fans (○), accumulation terraces (▼) and an up to 1 km-wide valley floor (□), which is covered by glacio-fluvial braided rivers. On the orographic right-hand on a *transfluence pass* leading to a small adjacent valley in the north, the *glacier polishings* can be most unambiguously established in the oblique exposures of the *glaciated metamorphite outcrops* (Fig 105 ▲ bottom right). Similar polishing of glaciated knobs are found on the rounded mountain ridges (Fig 105, all the remaining), though they have been formed by far more modest ice masses, merely decametre-thick. The valley flank relief of the Yarkand trough between Aghil and Kuenlun is much flatter than the containing walls of the Shaksgam trough between Karakorum and Aghil (see Chapters 4.5.1.2), so that the *valley glacier ice scour limits* here have been transformed by the slope polish of more *recent* flank glaciations with *greater* ice thicknesses.

If the Yarkand valley (Fig 110; 138 No. 53) is followed up another 10 km, the *maximum* altitude of the abrasion line is *not* only shown *from below* by the rounding of still polished mountain ridges (as in Fig 105), but also *from above down* through the *suspension* of pointed peaks and acute ridges (Fig 110 ---- above). *Abraded* mountain ridges are of course also visible here from below, ---- (Fig 110). In this uppermost area of the middle Yarkand valley (Fig 138 No. 53) the ice thicknesses exceeded 600 m even on the abraded transfluence pass (its maximum altitude being 4420 m asl: Fig 105 bottom right, 110 big), so that the maximum *ice thickness* would have been at least 1100–1200 m, assuming an Ice Age glacier level around 5000–5100 m asl (----) and a valley floor (□) at about 3920 m. Fig 104 provides the detailed analysis of a down valley orographic left-hand section of a flank (Fig 138, No. 36), which has been *abraded* and *rounded* by that more than 1 km-thick Yarkand glacier (). Fig 131 shows a section of the parts of slopes, frost-weathered metamorphosed sedimentary rocks have roughened up; they rise far above 1000 m and beyond the present *lower permafrost line* that runs at about 3800 m asl (evidence through occurrence of block glaciers, for instance in Fig 110 ■). The remaining and probably particularly resistant *polish facette* () is striking. Its glacially abraded surface dissects the strata deposits at an *acute-discordant*, ie diagonal angle. Details of this down valley prolongation of this left-hand valley flank (Fig 138, No. 38), including the maximum extent of the glacier surface are shown in Fig 65 (----).

During the Ice Age the V-shaped glacigenic *side valley* in the foreground (Fig 65 ; 138 No. 19, cf. Chap. 4.4) conducted a similarly substantial tributary stream towards the Yarkand glacier. Further upstream this side valley (Fig 64) from the 6008 m-high massif assumes more and more the form of a *"gorge-shaped trough"*; it has been cut out from the red Kuenlun granite, which preserves the *large abraded forms* very well (), but precipitated much Holocene *crumbling* (Fig 64) among the finer structures of the rock surface. Pictured on the right in Fig 104, the orographic left-hand side valley (Fig 138, between Nos 38 and 36), is a 16-km long glacigenic V-shaped valley that descends from the south side of the 5880 m-high massif, the summit region of which continues to be glaciated (○). Its flank abrasions () suggest a prehistoric glacier level which confirms to some extent that of the main valley glacier (----). Joining the main valley on the opposite side, the orographic right-hand "Mazar side valley" (Fig 138 No. 39), which descends in a southerly direction from the 4950 m-high "Mazar pass" (which crosses the Kuenlun main ridge, cf. Fig 54) as a *trough valley* with a wider drift floor (Fig 109 □), has also been abraded, smoothed and rounded far up the flanks in the metamorphite or phyllite substrate, at least in comparison with its *specific weathering* roughening-up (). Here, too, the Yarkand valley glacier level recurs (----). The *corresponding* maximum main glacier level in the confluence area of the "Mazar valley" has been entered in Fig 105 (---- in the middle).

5.4 Some Observations on the Maximum Glacier Cover of the Kuenlun North Incline Down to the Lowest Prehistoric Sites of the Overall Study Area in the Tarim Basin

Following the "Kudi valley" and its side valleys down towards the north, with reference to their *highest* prehistoric ice scour levels amounts to a repetition of the route covered in pursuit of the reconstruction of the Late Ice Age and even more recent glacier covers (cf. Chapters 4.3 and 4.5.3). The "Mazar pass" (4950 m asl; Fig 138 No. 16) is laid out in upright deposits of metamorphites, which are very prolific producers of frost-weathered scree, and consequently presents (Fig 54 ▷◁) a relief with moving periglacial moraines and strip form material sorting, which is largely understandable in purely solifluidal terms (Kuhle 1985, p. 189, Fig 2). The forms of the mountains and rocks on which the *present* style of formation is based are *glacigenically* rounded () and covered in part by *moraine sheets* (■). This mountain landscape is very *typical* for the prehistoric inland ice areas of western Spitzbergen (Kuhle 1983a, p. 48 Fig 6; p. 53 Fig 9 background) and the Ice Age glacier areas of Alaska in the area of *sedimentary rocks*. Here, at 5000 m asl, *near* the altitude of the present snow line (5200 m asl), where *perennial* patches of snow persist to this day, the *glacier thickness* (Fig 54 ----) depended solely on the steeply linked, deeply sunk valley relief, especially the "Kudi valley", which transported the ice directly to the north. As the *representative* orographic right-hand side valley of the "Kudi valley", the 6328 m massif north west valley (cf. Chap. 4.3) acts as the glacio-geomorphological link with the high summit of the Kuenlun (Fig 21). On the orographic right-hand *glacial cuspate areas* ie truncated spurs (◁◁ left side of the photo) with a *thin* deposit of frost talus are preserved as indicators of flank polishing. In this, the *highest*, section of the valley (Fig 138 Nos. 14; 21), the more recent, ie *post-Late Glacial dissolution* of the glacio-geomorphologically formed relief of the Ice Age is generally evident in contrast to the well preserved form of the trough in the *middle* and *lower* section of this valley (Fig 138 No. 15; 53). It is a case of *destruction of form* by the small Neo-Glacial to Historic glaciers (Fig 21 ○) and their oscillating ice margins, which undercut the valley flanks by *small-scale lateral erosion*. Besides *cuspate areas* () and *truncated mountain spurs* between the side valley mouths () Fig 53 shows the best developed *ice-scour limit* (---- on both edges) 850–950 m above the valley floor at about 5100–5300 m asl. The reconstructed glacier level in the background is evidence of the confluence of this Ice Age tributary glacier and the main "Kudi valley" glacier (Fig 53 - - - - centre). Here the thickness of the ice has *increased* to more than 1000 m, with a glacier surface of at least 4900 m asl. Following this glance into the left flank of the main valley, Fig 25 shows *abrasion forms* (▲) in the right main valley flank of this confluence area, which opens into another small *steep* side valley. This valley is more of a *cut*, or *V-shaped* valley, whilst the adjacent 6328 m massif's north west valley is a *trough valley* that has been widened by glacigenic concave flank abrasion. The former is partly the result of increased *traction power* in the ice due to the steepness of the valley (cf. Visser 1938, pp. 138/9) and partly of initially *subglacial* and then continuing as *subaerial* fluvial linear erosion (Fig 25 ⍀) of glacial meltwaters (O = present glaciers) with steep curve of the gradient of the hanging valley. This points to a formation in *two phases*: 1. when the ELA was *below* the level of the ice, *extensive* glacial ground abrasion created a rather *wide* valley floor in a hanging valley; 2. when the ice melted down further, ie the ELA rose *above* the glacier level, ground abrasion *receded* by comparison with the onset of the subglacial meltwaters' effectiveness, ie the abraded floor was dissected along *linear* lines. The geomorphology of these confluence areas moreover presents an opportunity to draw attention to a *glacial-genetic sequence* of forms which one regularly finds in places like, for instance, the northern Limestone Alps and the Dolomites, where towers of *sheer, polished* rock and related forms (X) rise *abruptly from rounded rocky ridges* with a thin scattering of scree (Fig 25 ▲ left). "Abruptly" means that lines of steep rock faces contrast sharply with the gradients below, with an angular *bend without* a concave transitional arch. The most likely explanations for this are *soft* rocks at the base, and *resistant* hanging ones above, ie a *petrographic* division, the geomorphological effectiveness of which remains entirely *independent* of the continuous processes of flank polishing. From the confluence area in question down the "Kudi valley" maintains the character of a deep-cut, narrow *trough valley* to *"gorge-like trough"* over a distance of more than 20 km, in the course of which another orographic left-hand side valley joins (cf. Chap. 4.3; Fig 138 No 17). In the area of the next confluence (Fig 138 No. 40) with a 27 km-long orographic left-hand side valley the *glacigenic* main valley character was adopted, which is shown in Fig 58. In this cross-section the valley floor has been set down into the bed rock granite to about 3000 m asl. The trough flanks have been *polished* to a height of many hundreds of metres (◥ ▲ ◤). Thanks to *ferro-manganese crusts* of varying intensity the *roughening-up of polishings* by crumbling, which are typical for massive crystalline rock, can be *relatively* dated as belonging to either the Recent or the Earlier Holocene. In all the irregularities that are *associated with rocks* – such erosion processes along areas of instability which occurred syngenetically with and due to the structure of the glacier filling, besides minor crumbling and more substantial rockfalls – there is evidence of the *typical widening* ice *scouring* near the valley bottom everywhere in the "Kudi valley". This led to the proportionally wide valley floor area which merges in a *concave* line with the wall-like scouring, *extended* middle sections of the trough flanks. The latter are below the *convex* arch (Fig 58 ◥ ◤), which has become the regular "back rest" for the highest slope part of the trough (cf. Klebelsberg 1948, pp. 353/54). Here, too, the main Ice Age thickness of the glacier (- - - -) exceeded – it would appear by far – 1000 m. The numerous short and steep side valleys (for instance, Fig. 111) dissect the polished flanks into in part *extremely pointed* cuspate *glacial wall facettes* (▲). The 27 km-long left hand side valley mentioned above has the same glacial-geomorphological features as the main valley (Fig 59 ◥ ◤) and shows a comparable thickness of ice filling (- - - -). The orographic right-hand side valley which, descending from the 5486 m massif (Kuenlun), joins right next to the Kudi settlement (Fig 138 No. 54), presents the repeatedly found *vertical partition* of the cross-section: flank abrasions above (Fig 111 ◥ ◤) have transformed the fluvially V-shaped valley into a *glacigenic V-shaped valley*; below the talweg cut has *consequently* been set into the upper valley cross- section, as shown by the bends in the gradient, which are the *edges of undercuttings* (⍀) of the talweg (∪). Since the subglacial linear erosion, which led to the deepening of this talweg, can only have been effective as from the late Late Ice Age – where the ELA ran far *above* the concerning glacier surface – without being subsequently destroyed by ground abrasion, it presents the geomorphological *change of regime* proper from the glacial valley that was eroded by abrasion and polishing to the steeper V-shaped formation of the fluvial valley.

In general there is, moreover, *indirect evidence* of the glacial genesis of the upper section of the valley cross-section in the present, fundamentally *different* morphodynamics, which focuses on the talweg, and could not have led to the valley form as a whole. The loess cover in the same area of the "Kudi valley" was found to be only 6–17 cm thick in the cross-section at 3000 m asl, thus pointing to only recently completed deglaciation (recent Late Ice Age). 25 km down the main valley ("Kudi valley") the definite *trough character* persists, but for logistical reasons on the expedition it proved impossible to follow the remaining 32 km into the foothills. However, in order to be able to complete the glacial-geomorphological profile to the north as far as the Tarim basin (Takla Makan desert), the line of investigation will be transferred a little to the east across the 3270 m-high Akaz pass (Fig 138 No. 42), which has immediate access to the northern Kuenlun foreland.

The "valley of Pusha" in particular was the object of two field trips devoted to investigating it from the exit to the foothills of the mountains (cf. Chap. 4.5.3). Though the valley continues to extend over crystalline slate (phyllite series) higher up, it is set in limestone outcrops near the exit. Here, at an altitude of the valley floor of about 2550 m asl, the latter sometimes show *glacial abrasion and polishing and forms of ice scouring* on the orographic left-hand. Corresponding *glacigenic rock abrasions and* *polishings* on the orographic right-hand side of the valley are shown in Fig 112 (▲). One and a half kilometres down-valley they merge with equally polished stretches of rock, which are now totally covered by *ground moraine*, and higher up increasing cloaked in *lateral moraines* (▲↓). It is in this locality (Fig 138 No. 43) that *lateral, or medial moraines* (Fig 112, 0■, far right and far left) set in, as the ones furthest up the valley to be dated as belonging to the *Last Ice Age* (cf. below). These *moraine roots* or proximate moraine beginnings are up to 800–900 m above the valley floor, ie they occur up to approximately 3400 m asl. In places accumulations of rough blocks are exposed (↓). The typical character of moraine accumulations can even be diagnosed from a distance on the basis of delicate wall forms (Fig 112 ↓ between 0 and ■) and slope accumulations running along the valley, as well as irregular undulations of dimensions larger than solifluction is capable of producing (↓ left of 0). It is, moreover, *characteristic* for the deposits of moraines that they are equally laid down on rock platforms at varying heights of the rock wall (under the second 0 on the right and a little lower under ■) without forming a continuous accumulation level, as is the case in river terraces. Down valley, the moraine material cloaks the bedrock valley flank completely (from ↓ to left). The left valley flank, too, is covered by *glacial diamictites* in the same way (0 and ■ far left). Following the valley down from here to the NNE there are *no further* outcrops of bedrock, nor is any reached by the metre or even decametre-deep gullies on the slope (▽▽). This leads to the conclusion that, beneath the moraine cover, the bedrock limestone of the northern Kuenlun mountain edge drops under the valley floor (▼▼), and that, starting from here, moraine walls *without* a rock core have built a 12 km prolongation of the valley into the foothills of the mountains, ie have created an exclusively *moraine valley*. It was consequently prolonged into the foothills by an *outlet or piedmont glacier* (as far as ■0 in the centre background). Down valley the height of the moraines decreases. If their levels in the area or in the vicinity of moraine contact with bedrock can be used to reconstruct an ice thickness (Fig 112 - - - -) of 800–900 m above the valley floor, it decreases to 700–600 m over a down-valley distance of c. 5 km, and to 400–300 m over 10–11 km down valley. This *decreasing thickness*, or altitudinal reduction of the moraine landscape of the foothills in the "Pusha valley", is represented in a larger survey in Fig 113. It shows how *large the scale of this moraine landscape* is. An entire series of parallel valleys in the foothills thus forms an *end moraine landscape of parallel strips* (cf. Hövermann & Kuhle 1985, pp 30–31; Kuhle 1991b, p. 74, Photo 43, illustrating a Late Glacial example from NE Tibet). Fig 113 shows five parallel moraine valleys at distances of tens of kilometres (Fig 138 No 43), and their *roots* on the mountainside are in the places where the moraine contours in the west are just visible in the *loess-bearing* air (0 left side of the photo). Fig 114 shows a moraine valley in the foothills 16 km east of the "Pusha valley" (between 0 in the foreground and 0 in the background), together with a kilometre-wide terminal basin (□). It is one of four moraine valleys up to the Yawashih or Tess settlement (37°25'N/77°26'E). This valley runs west of and parallel to the "moraine valley of Tess" (Tess is an irrigation oasis like Pusha; Fig. 138 No. 55). All these moraine ridges have decimetre-thick *loessial covers*; sheet-wash and gullies (▽) transform the primary loess of the upper slopes into alluvial loess at the base (Fig 113 ●●) so that they even assume a thickness of some metres towards the foot of the slopes. With aeolian loessial deposits amounting to *several millimetres a year*, the thicknesses observed here could easily be realized with the 20,000 years available. According to detailed observations during the 1986 expedition (when A.Schulze made the measurements) present *loess sedimentation* is carried out by NNE winds blowing at a rate of 5–7 m/sec towards the mountains. Thanks to the dust in the air they reduce vision to 40–10 km (cf. Fig 113 and 114). It is, moreover, not necessary to assume a loess sedimentation that set in after the Main Ice Age, since even during their formation by the pushing outlet glaciers the highest end moraine ridge areas (0) had remained *partially free* from ice, and thus open to *loess* precipitation. Aeolian loess covers cause wind slab-like slippages and step slippages, ie *loess slabs* on the steep slopes (Fig 114 ↖ ; 132 ↗ ; 133 ↑; 134 ↑). They permit a *direct* insight into the strata construction of the aeolian sediment which is spread *concordantly* across the moraine relief. The loess cover *induced* the ubiquitous *gulley cuttings* (▽▽) in the form of *converging* runways and – starting from the lower slope – via *regressive erosion* on its talwegs, moves upwards to uncover the *deposited* moraine material (Fig 132, 133, 134). These gullies therefore produce mixtures of alluvial loess and moraine material (Fig 113 ●; 133 ●). In places even glacio-fluvial drift terraces (Fig 112 ▼) call to mind the formation of *microfluviatile rills* in arid environments (Meckelein 1959) and are, of course, autochthonous forms on these slopes, due to rarely occurring heavy rains. Fig 113 (foreground, left), showing areas where the roots, or at least some of the growth of dwarf bushes, have been *denuded*, is evidence of these erosion processes, which are concentrated in gullies below, setting in on the moraine ridges with *sheet flood-like* departure. In parts this aerial erosion gave rise to "humpback"-like, though residual, pedestals, which are now colonized by cushions of dwarf shrub (Fig 113 foreground, right, beneath No. 4). In places where the gullies of the slopes dissect moraine terraces which are tied to the *melt-down process* of the ice (Fig 113 X, 132 and 134 ■), *moraine material* has been uncovered in large decametre-high exposures (Fig 132 below ■; 133 O; 134 O; 10). The exposures are on the orographic right-hand flank of the "Pusha moraine valley" described above (Fig 112, centre, right-hand from 0; 113 beneath No. 4, right) at 2000 m asl. The moraine ridge rises here to 400–700 m above the valley floor (Fig 113, O left; 132 0). The exposure in Fig 133 shows typical glacial diamictites with *large* to *very large* (O) *polymict* blocks (limestone, granite, crystalline slate) "swimming" in a fine ground mass (matrix), thus *isolated* from one another. In some places 35% of the fistsize components are *aligned* to the NNE (group I = 22°), and more than 50% of them are standing upright (group IV). The other exposure (Fig 134) with large to very large *polymict* blocks (O) shows corresponding conditions. The intermediate mass is extremely *condensed* here, and is interrupted by, or interspersed with, somewhat *sorted* glacio-fluvial bands of eroded moraine material (↓). The decametre-high exposure near the valley exit (Fig 10) is representative for the orographic left-hand moraine slopes of the "valley of Pusha". The material has *almost all* the features of the sedimentological structures of moraine (cf. Woldstedt 1961, pp 27–31; v. Klebelsberg 1948, Vol. I, pp. 252–292; Schwarzbach 1974, pp. 30/31 Tab 6). Besides the only occasionally chaotic structure - moraine material is *slightly* stratified here, or at least *laid down in banks* - the *polymict* load of blocks of varying cubatures (■) occurs in the *immediate* environs, though *isolated* from each other by *loamy* to *fine sandy* ground mass. Apart from rather rough-edged, though *occasionally* rounded blocks, the *mixed* appearance of which is typically glacial, relatively many blocks are *facetted*, ie somewhat polished down on one or several sides. An essential feature, however, are the strata of *compressed* sand (X below) down to the typical *flexures* and *sand-silt nests* (slices) (X above), which have been squeezed out of their original strata formation. Once again, the matrix is *tightly packed* (2 g/cm^3) and has a markedly *higher* ramm resistance than unbaked mudflow material. This can be attributed to *compacting* by upthrust glacier ice. This *reduction* of pore volumes is an *internal* compacting which *reduces* the dimensions of glacial- tectonic flexures - they are joined by *shearing planes*, which run more or less *diagonally* to the direction of impact, as shown in the corresponding crumbling lines (↘) in Fig 10. They are *not only* diagonal, as in the case of grain-sized, homogenous folded clay stones, but the *moraine-specific*, internal structural irregularities have led to *shell-like, bent*, sickle-shaped shearing planes and corresponding crumbling breaking out from the exposure wall (↙). Concluding this analysis, attention should be drawn to the evident similarity with the moraine material in the zone in front of the present K2 glacier (cf. Fig 31). Or, from the other point of view: the *significant* difference as compared to the equally diamictic mur cone and mudflow material from the Shaksgam valley (Fig 77; 78 X) in comparison with the local material here it amounts to a *further* confirmation of these very extensive and deep *end moraine* findings in the northern Kuenlun foreland. The most significant differences concern the *lower density* of the mur material (1.8–1.9 g/cm^3) and the regular, *lenticular-shaped* cross-sections (X) of the particular consecutive sedimentation events. The *histograms* showing *grain sizes* on the other hand, resemble those of end moraines down to the last detail (cf. for instance Fig 55 No. 6 with 9), and the *morphometry* of the fine grains of murs is largely *dependent* upon the original rock (cf. for instance Fig 56, 20.8.86/1); this applies far less to local moraines, and not at all to the distant ones in question. In any case a genesis of murs or mud-flows, however, is out of the question thanks to the *overall* geomorphology of the foothills of the mountains and the up to more than 700 m-high extensions of the mountain valleys, both of which contain diamictites, which are the only ones to bear *glacio-geomorphological* characteristics. The settlements like Pusha and Tess, to name just two, are *irrigation oases* which are tied to *spring horizons* in the mountain forelands at the very place where the moraine valleys end and the deposited ground moraines, with their water-impounding boulder clay, come to the surface. Another particularly instructive index, since *unambiguously* glacio-geomorphological circumstantial evidence, are the *pressure and abrasion or polishing forms* in the loose material of the walls in the parallel strips of the foothill moraines. They are particular *exaration rills* which the subglacial moraine impregnating the glacier bottom has *ploughed* into the less inclined ground moraine (Fig 114 ↓↓). The *more detailed* analysis of forms in the moraines of the "valley of Pusha" and the medial moraine walls to the west of it is struck by *pressure grooves* (Fig 113 ⇩⇩) which post-glacial gullies have not, or scarcely been able to dissect. This must be attributed to greater *firmness of the moraines* near the surface thanks to *increased linear compaction* through polishing or abrasion pressure. But even in places where there was back-cutting erosion by gullies, *pressure or polish grooves* came out through the lines of the gullies, thus making them into wavy lines (↓↓) retracing the differences conditioned by *compaction in resistance*. Evidence of particularly intensive *moraine material induration* of this kind can be seen in the occasional *total suspension* of larger gullies towards their *lower end* (Fig 113 ↑↑), whilst most of them merely undergo a restriction, ie a *narrowing* of the gully in the area of that groove induration. *Horizontal striping* (at right angles to the fall-line) of the moraine slopes and the continuation of these *abrasion or polish and pressure lineaments* on the other side of large gullies generally also point to a temporal *sequence* of the two forms, ie the dissection of these approximately horizontal forms *after* deglaciation.

Regarding the dating of foothill moraines: during the 1986 expedition samples for TL dating were taken (by Xü Daoming) from the exposure face (Fig 10). Their examination by the TL Laboratory at Gdansk University (supervisor: S. Fedorowicz) indicated ages of 32.9+/–4.9 Ka BP and 22.0+/–3.3 Ka BP. In spite of the circumspection indicated towards the possibilities of dating diamictites with this method, it points to the *Würm ice age* as the formative period of these moraines by the last piedmont glaciation of the foothills.

Though these end moraines of the *Last Ice Age* extend as far down as 1900–2000 m asl, *older*, more severely transformed end moraines can be mapped still further out into the foothills and some decametres deeper down (1750–1850 m asl; 37° 26'–32'N/77° 10'–45'E). A down- tilting to the north of their rough primary stratification or, better, banking up, has been induced *post-genetically* by *tectonic* movements (24–30/10). This old moraine material has been seized by the *subsidence area* of the Tarim basin (cf. Norin 1932; Machatschek 1954 p. 266; v. Wissmann 1959, p.

1335). Chen (1988, p. 30 Fig 3) mentions a depression of 2–3 mm/y for this area. The *coincidence* of the foothill at large with the edge of the depression which must have started the tilting of what at the outset were normally flat lying moraines to the north is explained by the process of depression, which the *load* of rock waste sediments from the mountains had *induced*. The changes in the position of the moraines are not only shown up by the inclination of the banking structure, but also by the tectonic fractures on some moraine surfaces. In the extended foothills of the exit of the "valley of Pusha", for instance (Fig 138, near No. 44) a moraine ridge can be observed that has been pierced by *up-tilted outcrops*. It must consequently be assumed that the *fault* in the northern *edge of the Kuenlun*, and the fault in the southern Tarim basin, which runs in a WNW/ESE direction, are precisely in this area. These *old moraines* – the substantial shifts of which required a rather extensive period of time – are classified as belonging to an older ice age, and thus to the *Riß period*. Situated about 100–200 m lower than the Würm moraines, the Riß moraines correspond with the worldwide slightly *lower* snowline of the Riß ice age, as compared with the Würm period. The fact that old moraines have been preserved *at all*, contradicts a uniform uplift of Tibet and its northerly mountain fringe in the course of the Pleistocene. If the Tibetan plateau and the Kuenlun had risen further since that older, probably Riß period glaciation, even very small rates of lifting of only 3 mm/year (cf. the higher rates of uplift according to Chen 1988, p. 30 Fig 3) during the 120,000 years of the Riß-Würm interglacial period would have resulted in a 360 m higher position. Having been higher during the Würm ice age, these *glacier feeding areas* would have led to an equally depressed lowest marginal location of the foothill ice, with the result that the only 100–200 m lower *Riß moraines* would have been *over-run* and *destroyed* (Kuhle 1989c, p. 283). The author therefore considers the *preserved* old moraines to be significant indicators of his Tibet-uplift- model (Kuhle 1993a), which *contradicts* the old ideas. It is based on a first uplift of Tibet *above* the *snow line* during the early Pleistocene, and a repeat *inland glaciation* during the Pleistocene (Fig 135 and 137) which led to *glacio-isostatic depression* during glacial periods and to *glacio-isostatic uplift* during the deglaciation of the inter-glacial periods. The old moraines contrast with the *parallel strip* end moraines (cf. above) of the last Ice Age through their *wide-ranging, lobe-shaped* course. The older ice rims surrounded a much larger and, *on the margins, largely continuous* piedmont glaciation. The *extremely* voluminous glacier deposits, including more than 700 m high moraine ridges (Fig 113, 114) as well as that *change* in the outline of the form of terminal basins point to *poly-glacial* formation of this wealth of forms in the foothills (cf. the example of the Alpine foothill glaciations according to Schaefer 1981). According to this model *every new ice age* occurring during the Pleistocene brought fresh supplies of drift material from the terminal or outlet glaciers of NW Tibet and its surrounding mountain ranges – the Kuenlun in this case – to the moraine landscape. Every new advance modified its growing buttresses, which in a *feedback* loop, in turn modified the run-off from the foothill glacier itself. The *development* traced here is one from initially almost *buttress-free, broadly lobate* terminal basins to really- reduced tongue basins of the *Würm period*, parallel median strip moraines have canalized and squeezed in (Fig 113). The resulting *substantial* ice thicknesses of far more than 700 m (cf. above) at the expense of larger expanses, though lesser thicknesses of ice confirm the *increasing* obstruction as a result of the *canalization* of the glacier run-off by the pile-up of detritus from *very large* catchment areas. Outside the Würm age end moraines (as for instance, in Fig 114) the Main Ice Age *ice cave drift floors* (No. 5) spread outside like a fan in the way described above (Chap. 4.5.3), though slightly canalized at first, thanks to the barriers set up by the old moraines (Fig 138 No. 45). Once outside the old moraines, the Würm drift entered the area of the Riß drift (No. 6). Here, however, the relief (ie gradient), is *so slight* that the two drift floors *mingled* (Fig 115, Nos 5, 6). Thanks to the slight gradient the more recent meltwater flows only made *shallow* cuts in the older body of drift material (Fig 115 ∇), so that they could not develop into permanent small valleys. On the contrary, under the circumstances of such an *unstable* structure of burden and energy from the Würm age meltwater run-off neither this area of older drift floor preserved a large-scale *anastomosing* channels, in which the – in principle – overlying, more recent drift materials (cf. Troll 1926, sketch of the Munich "inclined plain", together with the map and cross-section on the formation of talus fans, quoted after Woldstedt 1961, pp. 142–145, Fig 70, 71) were *mixed* with the Riß drift in those Riß-age drift floor sections. Probably even greater was the mingling of different *glacilimnic* deposits near *Yehcheng* settlement, on the WSW centre rim of the Tarim basin, an area *without run-off*, though with large-scale main Ice Age *terminal lakes landscape*, where the meltwaters collected. The dry deltas and recent lacustrine sediments in the famous example of Lop Nor 1100 km further east, in the northern foothills of the heavily glaciated Kuenlun (cf. Norin 1932, Fig 6, "Tschunak Stage") give an idea of conditions of sedimentation which *alternated* until the Late to Post-Glacial periods. The Riß-Würm age drift floor fans stretch from the lowest end moraines (see above) to the Yehcheng settlement c. 50 km further north, where they merge with these *glaci-limnic* sediments at 1470 m asl. Their *water-impounding* qualities – in contrast to the drift material – lead to *spring outflows* and favour the *sinking of wells*. They constitute the ecological basis for the founding and development of this town. This was helped by the fact that these fertile pelites are *easily worked*. Though, in principle, important, because of certain details, observations concerning the wealth of forms occurring among the in parts far more than 100 m-deep Quaternary cuts into at times tectonically displaced, much more substantial and precisely sorted drift sediments at about 1650 to 1750 m asl will not be gone into here, although there are similarities with the outlines of prehistoric glacier tongue basins. In the view of the author

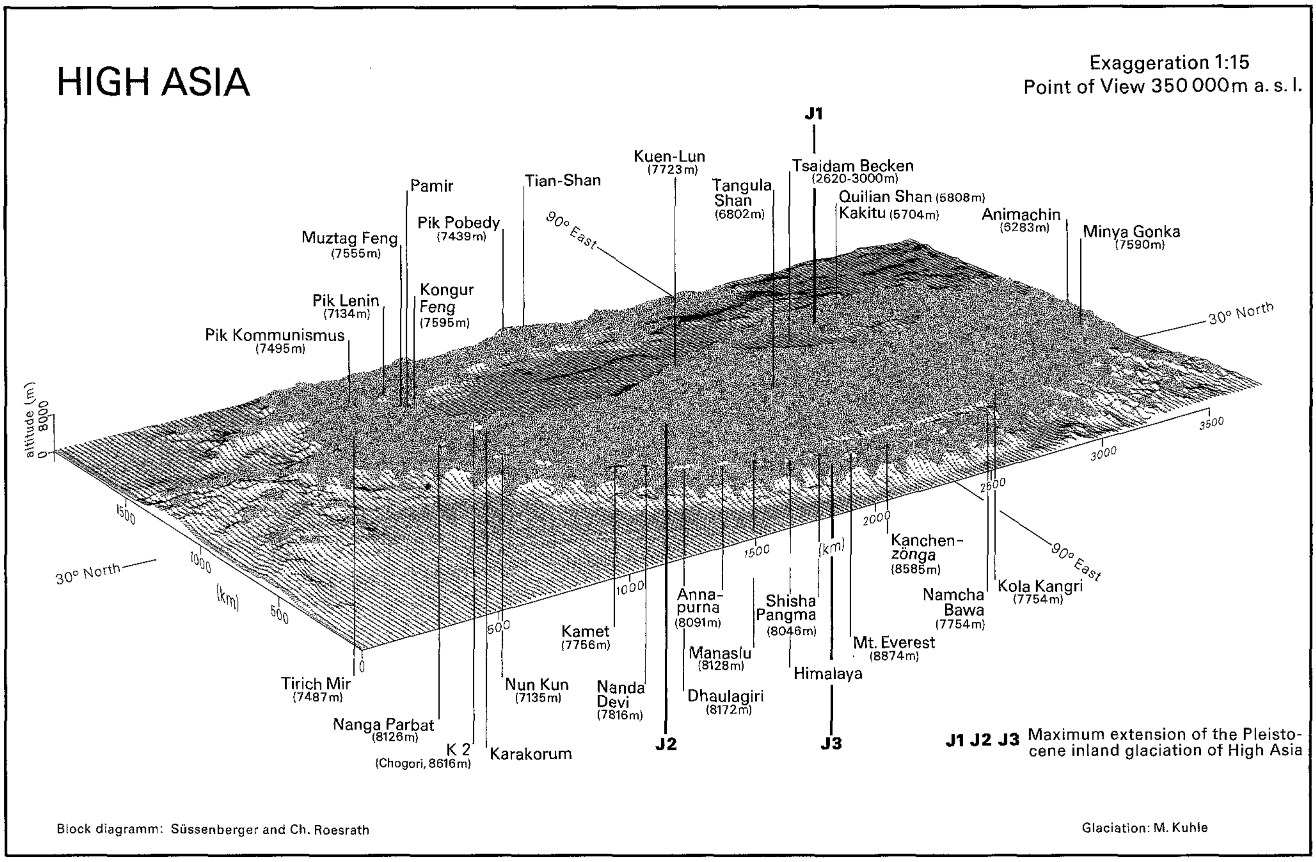


Fig 137 Towards the W the Würm period Tibetan ice merged with the ice stream network of the Karakorum (between Nun Kun and Tirich Mir), and formed a glacier system between Nanga Parbat and Kuenlun, which formed a bridge to the Pamir and even to the Tien Shan, in the NW.

they do *not* touch upon the topic under discussion, which concerns the evidence of the *lowest established* prehistoric marginal sites of ice. Inexplicable, however, and thus an open question in need of an explanation is the genesis of the numerous *cubic metre-sized*, partly *granite blocks* (37° 45'-48'N/77° 27'E) from the Kuenlun north slope which are scattered along the piste south of Yehcheng at about 1500 m asl. These blocks are *erratica*, and far too big for purely fluvial transportation. Even the rather narrow diamictite bands they belong to are most likely to suggest transport agents like "mudflows" or similar *self-propelled wet masses*, which were initiated by Ice Age *glacier courses* or outflows from impounded *moraine lakes*. A wealth of glacial forms, however, is entirely absent.

6. The Altitude of the Snow Line in the Last Main Ice Age and the Amounts of their Depression in the Area of Investigation

In the part of the Karakorum that falls within the area under investigation (Fig 1 No. 5, on the southern edge) the *climatic snow line* (ELA) on the present glaciers described above (Chap. 3) was defined on the basis of the *snowlessness lines*, the setting-in of *surface moraines* and *ice pyramids* as being at 5200-5300 m asl. In the Aghil mountains and on the Kuenlun ridge (Fig 1 No. 5, centre to northern edge) it runs at. c. 5200 asl. Moraine findings from the *Würm Ice Age* (among others Fig 112, 113 and 114 No. 0) at about 2000 m are evidence of the drop of the *lowest* ice-margins by at least 2600 m, since the biggest contemporary Kuenlun glaciers (Fig 21 O) flow down to an altitude of about 4600 m asl. By far the most of the present medium and smaller glaciers end a few hundred metres further up. This altitudinal difference permits a *snow line depression* of approximately 1300 m to be established (ELA depression (m) =

$$\frac{\text{present altitude of tongue end (m asl) - prehistoric altitude of tongue end (m asl)}}{2} =$$

$$\frac{4600\text{-}2000}{2} = 1300 \text{ m}$$

According to this calculation the ELA of the *Last Ice Age* ran at about 3900 m asl (5200 − 1300 = 3900). The same value is arrived at when v. Höfer's method (1879) is employed. If the most closely linked, ie the nearest, highest catchment areas of the Kuenlun main ridge with a mean (glacier catchment area) altitude of 5800 m are based on this calculation, the arithmetic mean to the lowest prehistoric ice margin locations at 2000 m asl also proves to be 3900 m asl (ELA [Würm] m asl) =

$$\frac{\text{mean altitude of frame-ridge (m asl) - altitude of tongue end (m asl)}}{2} + \text{altitude of tongue end (m asl)}$$

$$= \frac{5800-2000}{2} + 2000 = 3900 \text{ (m asl)}$$

and the ELA depression again 1300 m. Based on the local topography of the Kuenlun in the area under investigation, but subject to extreme simplification, these reconstructions of the ELA must be regarded as *approximate values*. Based on the author's (Kuhle 1988, pp. 546–563) method of ELA correction, which takes the topography into account, the assumed value of the depression is, if anything, *too small*, since 1. the mean altitude of the *catchment area decreases* together with the downward move of the snow-line, and 2. the share of *ablation area* per altitudinal interval increases as glaciation moves down into the foreland. As the author (Kuhle 1987d, pp 409–415; 1988b pp. 590/591; 1989c, pp. 275/276; 1991d, p. 211) – at times in co-operation with others, who carried out glacio-climatological 3-D model calculations for the simulation of the Ice Age Tibet ice, its mass balances (Kuhle, Herterich, Calov 1989, pp. 203–205 Fig 4–7) and the resulting flow dynamics, in accordance with the field data they had collected – attempted to show, the filling of the relief with glacier ice that took place was so considerable that *not only* the autochthonous glaciation of the Kuenlun N-slope, but also the ice of the Tibetan inland ice, which had flowed in from further afield (Fig 135a), had flowed down to the Tarim basin (Takla-Makan) at 2000 m asl in the form of *outlet glaciers*. If this is so, the build-up of ice would have led to an enlargement of the *secondary* mean altitude of the catchment area of the ice dome from the interior and the margins of the ice stream net, and thus to a change in the mass balances through a *positive feed-back*. With an Ice Age climatic snowline at about 3900 m asl and a mean altitude of the valley floor at 3800–4200 m asl in the area under investigation in the valleys of the Shaksgam and Yarkand, a far-reaching *glacier filling* of the entire relief between Karakorum and Kuenlun, which are linked with these main valleys, must be assumed in view of the very *slightly inclined* and *winding courses* of the West Tibetan outlet valleys (Fig 135 a/b). But even the *ice transfluences* into the *Karakorum S-slope* (Chap. 5.1) were a consequence of the *impounding* of the ice stream net (Fig 135b) which accumulated backwards in the still glaciated side valleys from these main valleys. The degree to which an ELA depression of more than 1000 m must have caused the *filling* of the relief *with ice* on reaching the presently ice-free main valley floors is shown by the fact that with a mere 1000 m higher ELA contemporary glaciers like K2, Sarpo Laggo and the Skamri glacier already stretch down to about 4100 m asl. To aid comparative orientation, the exponential *increase in the glacier area* during an ELA depression of only 500 m – as it occurred during the later Ice Age (Stage IV or III) – is given for the Karakorum in Fig 136.

7. Type and Order of Main Ice Age Glaciation from Karakorum to Kuenlun, their Relation to the Tibetan Inland Ice and their Indicator Value for High Asia during the Ice Age

The type of glaciation was that of ice stream net (Fig 135a/b), the Karakorum S-slope to Kuenlun N-slope surface section of which was in *continuous* contact with gradients of, in parts, the same dimension, thanks to transfluence passes (for instance, Muztagh-Shaksgam valley transfluence and Shaksgam-Yarkand valley transfluence by way of the Aghil pass; cf Fig 138 Nos. 12 and 25), and with the counter-gradients by way of glaciated saddles (for example, Mazar pass; see Fig 138 No. 16). In the Karakorum area alone this ice stream net extended over an *area* far in excess of 100,000 km^2 (cf. the Late Glacial situation Fig 136). Glacial forms of erosion, abrasion and polish lines are merely evidence of the *minimum ice thicknesses*, the upper limits of which in turn prove the occurrence of ice thicknesses of more than 1000–1400 m in the main valleys. As the chief drains the valleys of the Shaksgam and the Yarkand, the main valleys of the first order, brought about a large-scale *inclination of the glacier surface* towards the north west. Insofar as they are high enough, the intervening summits, crests and higher mountain ridges formed nunatak-like breaks in the surface of the ice stream net ie they interrupted an approximately even level of a valley glacier. They, in turn, were covered with *shallow* glaciation on their flanks, and even steeply draining hanging glaciers, the run-offs of which were adjusted to the level of the ice stream net, and contributed to the supplies for the valley glaciers by the *avalanche feeding* that is typical for their steep relief. In the area under investigation the *surface level* of this dendritic valley glacier system, which joins to form a large network, lies between approximately 6000 and 4200 m asl. Only on the Kuenlun N-slope of the area under investigation did the surfaces of the *outlet glacier* tongues, which appear here as ice streams, dip down *below* the ELA, and having extended over tens of kilometres, still reach the deepest marginal ice locations at about 2000 m asl. As can be seen in Tab 1 and has been confirmed by field observations of the construction of the K2 summit, thus setting an example for the Karakorum (see Chap 3.1.1), the present *upper limit of the glacier* is at about 6900–7100 m asl. In accordance with the telemetric measurements of surface temperatures during the *warmest* time of the year and day (11 to 3 o'clock) even in situations of radiation weather and on prevailing rock (ie dark) surfaces, temperatures of less than 0 °C occur from upwards of 7616 m asl (Tab 1, line 2, with double SEE and correlation coefficient of −0.827). Starting from the highest value for the *upper glacier line*, which has been arrived at and thus substantiated by measurements, and consequently following a main Ice Age ELA depression of 1300 m for the upper glacier line as well, the latter must have been at a maximum of 6500 m asl (7616 – 1300 = 6316) during the Ice Age. The above-mentioned *avalanche feeding* from the peaks that rise steeply above the ice stream net was *restricted* to the still glaciated area below 6500 m asl down to the level of the ice stream. In contrast to the present time, the *activities of ice avalanches* was *greatly* reduced. During the Ice Age *primary* feeding by snow precipitation dominated. The *change* in the glacier feeding situation from the present *Inter-Glacial* to the *Main Glacial*, ie from the preferential *avalanche feeding* to *primary feeding* on the raised surface of the ice stream net leads to the conclusion that in prehistoric times the great mountain heights of the Karakorum, Aghil and Kuenlun had become *unimportant* for the feeding of glaciers, and that *area extension* at a lower altitude had taken their place. This perspective opens the way to a better understanding of the *direct eastern link* of the Tibetan glacier cover, where extensive prehistoric glaciations have hitherto been reconstructed by Li Chichun and Cheng Penhsing (1980), Norin (1982), and especially by Trinkler (1932), although *no-one* but the author has so far published anything concerning the *inland ice*. It has been accepted that 1. the mean altitude of valley floors and – areas in western Tibet including the Depsang Plains and the high peneplains extending towards the east, the lowest, though still 4900–5100 m-high depression of which are filled by the lakes of the Aksai Chin and Sarigh Jilganang Köl, are situated c. 1000 m *higher* than the main valley floors of Shaksgam and Yarkand investigated by this study, which carried a minimum load of a 1400 m-thick glacier during the *Ice Age*. But even these main valley floors were 2. *above* the Ice Age snow-line (ELA). It follows that here, in the valleys of the Shaksgam and Yarkand, the climate must have been much more *humid* than in Western Tibet, so that there, 1000 m further up in areas much more favourable to glacier feeding, no ice masses could have been built up at that time, too. Though Western Tibet is relatively dry, it still receives *more precipitation* than the valleys of the Shaksgam and Yarkand, where at 4000 m asl scarcely 60–100 mm/year are recorded in the *rain shadow* of the Karakorum main ridge, nor for the Yarkand valley or even the Aghil mountains. Measurements of firn accumulations carried out on the K2 glacier in the summer of 1986 have established that at altitudes between 5000 m and 6000 m precipitation is higher to the 10th power and rises to c. 1500–2000 mm/year. Though precipitation does not increase to the same degree towards W-Tibet, it nonetheless amounts to c. 100–200% more than in the middle sections of the Shaksgam and the Yarkand valleys, and increases to 439 mm/year near the Lhasa station (3760 m asl) in Central Tibet further east. The snow profiles on the Geladaindong-E glacier in the western Tanggula Shan even yielded about 700 mm annual precipitation at 5800 m, c. 200 m above the ELA (Kuhle 1991d, p. 137). In providing glacio-geomorphological evidence of the very extensive *glaciation of the ice stream net* in the semi-arid rain shadow of the Karakorum this analysis of *interferences* through topography, altitude and precipitation provides the access to the probability of a *glaciation of the interior of western Tibet*, together with a link with this ice stream net, the main arms of which drained the edge of the inland ice to the NW in the form of more than 1400 m-thick *outlet glaciers* (cf. Fig 137). During the Ice Age there was no basis for the controversial argument of the *dome-like uplift of the snow line* above Central and Western Tibet, as has correctly been demonstrated by v. Wissmann (1959 Fig 14 and map 1:5 000 000) and in the same way by Shi Yafeng et al "Map of (....) the snowline elevation in China" for temporary courses of snowlines (cf. Chap. 2). As has been shown in the case of Central Tibet (Kuhle 1991d, p. 139), the *mass uplift effect* responsible for the *present upvaulting* did not exist during the Ice Age because the *energy transformation of the incoming* sunshine from *short-wave to long-wave radiation* failed to materialize, thanks initially to snow patches that remained throughout the season, followed by floors of firn and ice in the broad bottoms of high valleys (cf. present forms in Visser 1938, Fig 89–91, "flood ice"), and finally by shields of firn and larger ice caps which built up as cooling down proceeded, and increasingly coalesced. *Even now* the *inter-glacial* frost drift areas of Tibet with their albedo values of about 14–20% (Kuhle 1987d, p. 409 Fig 23, 416; Kuhle & Jacobsen 1988, pp. 597–599) lead to the formation of the most significant macro-climatic *heating-up area* (Flohn 1959, p. 323). This *climatic "heat dome"* collapsed into itself above the firn areas of the Tibetan highlands, due to the 90% *reflection of the solar radiation* into the stratosphere, and changed into a thermically inverse *"bowl of coldness"*, so that the isolines of the climatic snowline *dropped down* from the periphery to the centre of the plateau. Amongst other things, this resulted in the complete build-up of the ice of the Tibetan interior by a *positive feed-back*: it extends over an area of 2.4 x 10^6 km^2, and is more than 1000 m thick (Fig 137). *Inverse to the present snow line* vault of Tibet, this *inverse snow line bowl* was in theory thermally raised a little further by a reduced *föhn* effect caused by in the Ice Age reduced luff-like precipitation from the higher edges – in fact, however, *over-compensated* by the formerly reduced humidity which reached the centre of the plateau, and made to flatten out. At present the ELA difference between the Shaksgam and the Yarkand valleys, the area under investigation, and the western plateau of Tibet (Depsang Plateau and Aksai Chin) is 900 m (5250 to 6100 m asl, to be precise). The altitudinal difference is 100 m less, since the western Tibetan Plateau is higher than the area between the Karakorum and the Kuenlun, ie the study area. According to these altitudinal relationships, *even* a continuing existence of that heat dome above the highland could have allowed the surface of the plateau to tower *above* the snow line and *even the Tibetan inland ice* could have built up on its own accord. Fig 136 places the Tibet *graph of the increase in glacier areas* beside the one for the Karakorum, with depressions of the snowline amounting to only 500 m. During the Main Ice Age this depression amounted to even more than 1000 m

(1100–1300 m; cf. Chap. 6 and Kuhle 1988b, pp. 588–590). Even during an *Early Ice Age* ELA depression of only 500 m (which must subsequently have existed in a post-Main Ice Age as well, ie have returned during the Late Ice Age) a plateau ice built up in the Tibetan interior, as shown in that horizontal course of the graph. In this context attention must be drawn to the fact that, *besides* the *in-situ* build-up of firn and ice, which has been described above as a consequence of a snowline depression with an effect upon the plateau surface, there has at the same time been a previously effective *coalescence* of glacier ice from the mountain ranges of western Tibet and the more than 5500 m-high mountain ridges down in the high valley floors and basin-shaped depressions. Assuming a snow line depression of 500 m, a plateau ice *had* to build up gradually in this way in the early Ice Age, since the lowest edges of the *initial valley and hanging glaciers* concerned flowed down approximately the *same* altitudinal distance *below* the ELA, the highest reaches of the feeding area towered *above* the ELA. At first confined to the level of the high plateau below the ELA, this glacier outflow proceeded to superimpose the continuing snowline depression and supported the development of plateau ice by building up a *secondary region of feeding areas*. This might have taken the form of an ice cover which spread, like a hammer-head, on the plateau at the foot of the mountains and, fed by numerous hanging glaciers and small valley glaciers, not only reached the lowest points of the plateau at an early stage, but also *built up* the area of the high plateau, thus *enlarging* the glacier feeding area. The insight into this further factor of *successive* gradual development and self-augmentation completes the entire *causal linkages* of glaciation in the Tibetan interior: 1. ELA depression; 2. Albedo reinforcement by initial areas of snow, firn and ice: 3. Built-up of a higher and thus colder glacier surface, favouring further glacier feeding.

One must accordingly envisage a Main Ice Age ice covering the Tibetan interior (Fig 137), the substantial NW-*outlet glaciers* of which flowed down the Shaksgam and Yarkand valleys. At the same time there was an ice stream net in the area under investigation, which was pierced by high mountains similarly to all high mountain ranges with ice thicknesses in places of more than 1000 m, which surround the high plateau of Tibet (Kuhle 1988b, pp. 590–591). Thanks to its *extremely low precipitation* in the N-shadow of three more than 6500–8600 m-high mountain systems the area investigated in 1986 occupies a *key function* for the Central Asian arid region, which required the present detailed study. The fact that it was possible to present evidence of ice stream net *glaciation of extreme dimensions* allows a better reconstruction of the *Tibetan* Glacial inland ice (Fig 137).

Summary

Between August and November 1986 an expedition attempted the glacio-geomorphological and glacio-geological reconstruction of the *maximum Ice Age glacier cover* between the Karakorum in the south and the Tarim basin in the north. For methodological reasons the field analysis was carried out in a *reverse-chronological order, from above to below*, ie starting with the present glaciers of the high regions, down-valley via the reconstruction of historic, Neo-Glacial (Holocene) and Late Glacial ice margin locations to the lowest *Main Ice Age end moraines*. Geomorphologically classified as belonging to the last Ice Age (Würm), though their absolute date is the same, the lowest moraines extend from the Kuenlun N-slope to the mountain foothills and into the Tarim basin to at least 2000 m asl. These glacigenic diamictites have *enormous* thicknesses. Some of the moraine walls rise to a relative height of 700 m. Classified as belonging to the *Riß age*, even older foreland moraines reach heights of c. 1800 m asl. They evidence that during the Riß-Würm interglacial period the large glacier feeding areas of NW Tibet and of the Kuenlun were not raised above their Riß period level, or they would have been *over-run* by the glaciers of the last Ice Age (Würm). The data established by these observations contradict the traditional approach of an approximately continuous (same direction) uplift of Tibet and mountain ranges surrounding it in the north since the early Pleistocene. The data thus obtained point rather to *glacial-isostatic rise and fall* in the course of the Quaternary ice ages. The *present snow line* runs at 5300–5200 m asl, and experienced a 1300 m *depression* to about 3900 m above sea-level during the *last Ice Age*. Assuming an intermediate altitude of the main valley floors between 3900 and 4200 m asl of the Shaksgam and Yarkand valleys, an *ice stream net-like glacier filling* of the entire relief between Karakorum and Kuenlun was inevitable. *Empirical* and detailed evidence of the glacier filling was provided by trough valley profiles, polished and abraded slopes with polish lines, transfluence passes with glaciated knobs and erratica, and glacier striae. The *main valley glacier thicknesses* exceeded 1400 m. The *level* of this ice stream net, which had a dome-like vault immediately above the mountain ranges in the vicinity of the ice sheds, was situated in the feeding area between c. 6000 and 4000 m asl; within the study area it was only the Kuenlun N-slope that it fell below the ELA and down to the lowest ice margins. The Main Ice Age ice level of the area under investigation was broken by mountain ranges and individual peaks, which towered at most 2500 m (K2 summit) above the ice stream net. As the *climatic upper glacier line* had been depressed to at least 6500 m asl, and the ELA by 1300 m, hanging glaciers developed in this merely 500 m-wide altitudinal belt in the direction of the surface of the ice stream net, which joined the main glacier arms and despatched ice avalanches. Slopes and wall faces towering beyond 6500 m asl were *permanently frozen*, and contributed little to the feeding of the glacier. While *secondary* glacier feeding through ice avalanches predominates now, the Ice Age experienced chiefly *primary* feeding through snow falling upon the extensive surfaces of the ice stream net. The *study of the Ice Age* in High Asia regards this extremely leeward, most *arid*

edge of Tibet with a precipitation which amounts to only 25% of the Central Tibetan humidity at 4000 m asl, as having the character of a *paradigm*. At the same time its extreme Ice Age glaciation makes an approximately total glaciation of the Tibetan *inland ice* likely (Fig 137), the more so as its valley floors and areas are 1000 m higher. During the Main Ice Age period this inland ice – the reconstruction of which has concerned the author since 1976 (Fig 1) – had an extension of about 2.4×10^6 km^2 (without the Tian Shan range). The main branches of this enormous ice stream network, the Shaksgam and the Yarkand glaciers, acted as northwesterly *outlet glaciers*. The potentially *Ice Age triggering* effect of this *subtropical* – and therefore *extremely energy-effective* – inland ice has been the subject of the author's detailed radiation- and relief-specific hypothesis of the Ice Age since 1982. In contrast to all the researchers publishing work on loess, who assign the deserts of the Asian interior as its place of origin, or call upon the services of periglacial genesis, the author attributes these *very considerable loess deposits* – like those in Europe – to *glacial genesis*, and in this case to the Tibetan inland ice. Loess was and is being blown from the glacio-fluvial and glacilimnic sediments of the foothills of the North Tibetan mountains.

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