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Editor's Preface

This book – the third volume in the series on the glacio-geomorphological investigation of "Tibet and High Asia", which was begun in 1988 – constitutes the continuation towards the north-west. Whilst southern Tibet was the subject of the first volume, and the second volume concerned central Tibet to the edge of this high plateau, the studies in the present volume present new research undertaken between the northern side of the Karakorum, the Aghil Mountains and the Kuenlun in the south, and the 1100 km-long, 400 km-wide, and 7439 m-high Tian Shan system in the north. This profile extends from 36°–43° N, ie from High Asia to the equally quasi-arid, areic Central Asian macro-region. The changes that have occurred here in the period since the last Ice Age have been so considerable that the investigations carried out by the authors have led to the emergence of an *entirely different* picture of this hitherto little researched area from the one older, deductive or preconceived view considered possible. Such a profile from Tibet to the north suggests a continuation from the Tian Shan to Siberia by way of the Altai and Sajan Mountains as prospective future research. The central problem of such a project would have to be the question of the Ice Age ELA gradient from the Tibetan to the North Siberian inland ice.

The research results published in this volume have been arrived at by applying glaciological, glacio-geological and glacio-geomorphological methods in the field as well as in the laboratory and the computing room since 1986 over the course of four research expeditions and surveying campaigns, which were financed by the Deutsche Forschungsgemeinschaft (DFG), the Max-Planck-Gesellschaft (MPG), the Academia Sinica, the Russian Academy of Sciences, as well as private donors. Carried out in the period August–November 1986, the most elaborate enterprise of all the expeditions led from the Tarim Basin to K2 on the north side of the main ridge of the Karakorum (Figure: group photograph of the participants). Due to the remoteness of the area, it was necessary to engage a camel caravan of 70 animals for the approach and return journeys. Representative measurements on the Tuyuksu Glacier, the Sary-Tor Glacier, and the Glacier No. 1 were carried out from 1984 to 1987, and in 1988 and in 1991 glacio-geomorphological expeditions were run from Kungey Alatau to the hitherto totally unexplored Dankov Massif in the Terskey Alatau.

The first paper (M. Kuhle) presents findings and evidence for a very extensive and thick glaciation of the mountainous edge around western Tibet during the last Ice Age, from whence an ice stream network, which infilled the relief, flowed down into the Tarim Basin to below 2000 m above sea-level. The second paper (M. G. Grosswald, M. Kuhle & J. L. Fastook), besides supplying the glacio-geomorphological evidence for a complete glacial cover of the Tian Shan Plateau during the Ice Age, together with some data provides model-based calculations of this glacial cover for various periods in the Late Glacial and Holocene. The authors M.B. Dyurgerov, V.N. Mikhalenko, M.G. Kunakhovitch, S.N. Ushnurtsev, Liu Chaohai and Xie Zichu explored the causes of glacier mass-balance variations in the Tian Shan Mountains. Undertaken on active glaciers, these investigations provide a platform for a future programme envisaging mass balance reconstructions of glacier areas extending over c. 3 million km^2 in High Asia.

The editor wishes to take this opportunity to express his sincere thanks to the editor-in-chief, Dr. Wolf Tietze, for another very generous and technically high quality publication of the new results, as well as for the expert advice and assistance generously given over the past decade by him and his wife in all manner of questions concerning the organisation of research. Last but not least thanks are due to all the authors for their contributions to this volume.

Göttingen, October 1993 *Matthias Kuhle*

Foreword Photograph: Members of the joint German-Chinese scientific expedition to the K2 and Karakorum N-slope, August to November 1986, photographed in the Yarkand valley on the return journey at the end of October 1986. From left to right and back-row to front-row: Chinese manager of the base camp; Rick J. Thwaites (Australian glaciologist); Prof. Huang Rongfu (Chinese botanist from Hsining); Chinese expedition doctor from Beijing; three Uigurian camel drivers and high altitude porters from Yeh Cheng; Jens-Peter Jacobsen (graduate geographer, Geographical Institute, Göttingen University and Institute for Scientific Film (IWF) Göttingen); Uigurian camel driver; Prof. Dr. Matthias Kuhle (project leader and leader of the German side of the expedition, Geographical Institute, Göttingen University); Dr. Ding Yongjian (Inst. of Glaciology and Cryopedology, Lanzhou); Uigurian interpreter; Dr. Qin Dahe (Inst. of Glaciology and Cryopedology, Lanzhou); Andreas Schulze and Holger Dietrich (graduate geographers from the Geographical Institute of Göttingen University); a camel driver; Rachman (caravan leader from Yeh Cheng); Chinese manager of the high-altitude camp; camp cook of the high-altitude camp; Prof. (Mrs) Feng Qinghua; Bernhard Dickoré (graduate botanist from the Botanical Inst. of Göttingen University); Prof. Xu Daoming (expedition leader on the Chinese side, Inst. of Glaciology and Cryopedology, Lanzhou); four Uigurian camel drivers and high-altitude porters from Yeh Cheng; Kuno Lechner (scientific cameraman, Inst. for Scientific Film (IWF) Göttingen). Photo: M. Kuhle, 23.10.1986

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Present and Pleistocene Glaciation on the North-Western Margin of Tibet between the Karakorum Main Ridge and the Tarim Basin, Supporting the Evidence of a Pleistocene Inland Glaciation in Tibet

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1. Description of the Project and Logistics of the 1986 Expedition

The author has been engaged in the reconstruction of the maximum extent of the Pleistocene glaciation in Asia since 1972, leading research expeditions to Tibet and to its surrounding mountain ranges since 1976 (Fig 1). Initially these were financially supported by the German Research Society (Deutsche Forschungsgemeinschaft = DFG), with additional support from the Max Planck Society since 1981. This, the sixth, expedition to High Asia took place between August and November 1986; the area under investigation was the extremely arid NW-edge of the Karakorum in the south and the Tarim Basin (Takla Makan desert) in the north (35°53'–39°N/76°–77°30'E) (Fig 1, No.5; Fig 138). Run as a joint Chinese-German enterprise of the Department of Geography of the Göttingen University and the Lanzhou Institute of Glaciology and Cryopedology of the Academia Sinica, the expedition was led by Professor Xü Daoming and the author, who had initiated the project. Eight Chinese and six German scientists took part in the expedition, together with a total of 30 technical support staff, camel drivers, high altitude porters and assistants, and last but not least a caravan of 70 camels. The area under investigation was the Karakorum N-side, where access is difficult; particularly the area of the K2 (Tschogori) and Skyang Kangri, including the Skamri, the Sarpo Laggo, the Skyang Kangri and the K2 glaciers, where little research had been undertaken. Though the glacio-geomorphological reconstruction of prehistoric glaciation is the subject of this paper, studies were carried out concerning current glaciology, botany and plant geography. In addition, four climatic stations were installed between 3980 m and 5330 m asl on surfaces of rock, debris and firn in the glacier catchment area, carrying out meteorological-climatological research into albedos and radiation balances, in order to calculate the *Ice Age energy losses* for the atmosphere on the basis of 80–90% of the reflection values of the expanses of ice, firn and snow which covered most of the rock and debris (cf. Kuhle 1988 c, 1988 d).

2. Characteristics of the Area under Investigation and their Methodological Significance

2.1 Topography and Climate

Extending over an area of 110 x 190 km, where key localities representative of prehistoric glaciation were investigated (Fig 138), the area covered by the expedition is divided into three diagonal mountain chains. They strike from NW to SE, with heights staggered from S to N, the

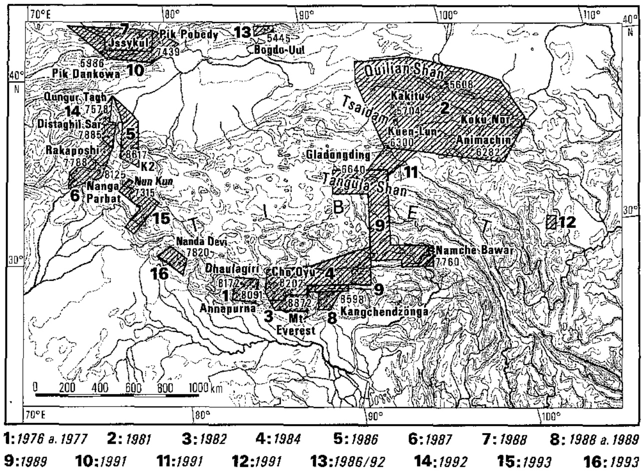


Fig 1 Research areas in High Asia visited by the author since 1976

highest being the main ridge of the Karakorum (8611 m or 8617 m) in the south; the next rising to 6858 m, but divided from the Karakorum by the Shaksgam valley, is the Aghil chain in the north. Still further north of this 20 to 30 km-wide mountain system, but divided by the Yarkand valley, is the here 60 to 70 km-wide mountain range of the Kuenlun, culminating at 6412 m in its western section. Longitudinal valleys and their transverse tributary valleys are embedded in these mountain chains. In the S/N cross-section of the area under investigation (Fig 138) their valley floor levels occur between 3400 and 4100 m asl (Fig 138). The mountain slopes, which rise up from there more or less abruptly, are thus 2000-4000 m high. The *concordant* elevation differences reach a maximum of about 5000 m. This is the *vertical distance* of the immediate catchment area, which made the local prehistoric glaciation drain off into the valleys. The two major *longitudinal valleys,* the Shaksgam and the Yarkand, drain the Tibetan plateau in the NW (Fig 1, No. 5). There the valley floors descend from altitudes which reach 5200 m or more on the Depsang plateau (an intact remnant of a western plateau) to give but one example, or even exceed 5570 m at the Karakorum pass. The convergence of the two mountain ranges, the Karakorum and the Kuenlun, is reflected in this W-Tibetan configuration of longitudinal valleys of the chosen study area. The highland of Tibet, which separates them further east, tapers towards the west, with the pointed ridge of the Aghil in the area under investigation forming its most westerly extension. The linking of a very high and - in the high mountains always immediately - *adjacent* local catchment area with *extensive, far distant* catchment areas by way of longitudinal valleys produces an *arrangement* which *favours* the reconstruction of Ice Age glaciation. On, that is in, the W-Tibetan Plateau areas proper, such a reconstruction would not be likely to succeed since they occur at too high an altitude, ie too close to the *present* ELA and thus too close to the present glaciation, to reveal either the *lowest former ice margins* by means of end moraines, or the *lowest glacial abrasions and* polishings (cf. Kuhle 1991 b, pp. 109-113; 1991 d, p. 134). Moreover, the glacigenic erosion forms which can only be brought about by guided, canalized run-off of the ice through the valleys, which cannot occur on plateaux but only in the adjacent valleys in the west, are most *clearly* recognized in the area selected for investigation. Trough cross-profiles are to be expected here which require a *shallow* profile of a *linear* valley, ie a *longitudinal* valley. The immediately adjacent, *steep cross-valleys* on the other hand, offer poor conditions for the preservation of U-shaped profiles as glaciated key forms (cf. Kuhle 1991 b pp. 1-5; 1983 pp. 154-155). Besides these two orographically *favourable factors* for the reconstruction of glaciers: (1) "the location on the edge of the high plateau, with immediate glacier run-off to far below the snow line" and (2) "canalization through valleys" this area as a *test-bed* for prehistoric glaciation is served by a *disadvantageous factor* related to this location: ie its *very marked aridity.* As far as it *can be assessed,* the aridity in the valley floors at about 4000 m nowadays amounts to 60-100 mm/year, and increases to 40 mm/year towards the Tarim basin below 2000 m asl. This fact indicates a *deterioration* of climatic conditions for *allochthonous* glaciation compared with those of the Tibetan highland, where precipitation at the 5000 m level is distinctly higher and would also have been so during the Ice Age. In western High Asia *precipitation* generally *increases* with *altitude.* In many places of the Karakorum, as well as the area under investigation in the stricter sense, it varies from 1200 to more than 1500 mm/year (max. 2000 mm/y) at altitudes between 5000 and 6000 m asl. Evidence of such quantities of precipitation was gathered by the 1986 expedition by snow accumulation measurements (snow profile pits) in the K 2 glacier catchment area and ablation measurements on the K 2 glacier, the Skamri glacier and the Sarpo Laggo glacier (cf. Ding Yongjian 1991, p. 244 and 1987, p. 25, Tab. 11; Xie Zichu et al. 1987 p. 7; Shen Yongping 1987 p. 1). These data indicate that in a cross-section stretching from W-Tibet to the adjacent mountain ranges in the west an ELA depression *did not assume the proportions* which the decreasing *mass elevation effect* would suggest. It was at least compensated in part by the *decrease in precipitation.* This line of thinking points to the fact that the local (autochthonous) glaciations of the Karakorum, the Aghil, and the Kuenlun mountains are *unlikely* to have been favoured climatically compared to the allochthonous glaciation experienced by the Tibetan plateau. *Primarily topographical reasons* also account for lesser productivity in glacier formation during the Ice Age among the mountain ranges on the western edge of Tibet. Though higher, their narrow crests and isolated peaks only offered *very small areas* for glacier development, whereas the Tibetan plateau, with its large undissected plateau, towered above the ELA depression. The resulting slow glacier drainage contributed in turn to a rapid *build-up* of the glacier surface, and thus to an *even higher* catchment area, leading to a further increase in the glacier feeding. These considerations suggest that

material	time	0°C-line (m asl) SEE	2SEE	temperature gradient (°C/100 m)	correlation	difference of altitude (m)	difference of temperature gradients (°C/100 m)
rock	0–24	6306	7781	−0.674	−0.567	---	---
rock	11–15	5701	7616	−0.827	−0.717	224	0.282
rock	15–11	5813	7392	−0.545	−0.536		
ice	0–24	4308	5018	−0.646	−0.837	---	---
ice	11–15	4566	5304	−0.671	−0.775	391	0.035
ice	15–11	4216	4913	−0.636	−0.852		

Tab 1 Statistically evaluated and summarized telemetric surface temperature measurements during radiation weather periods, taken with the aid of an infra-red long distance thermometer on mountain flanks with rock and ice surfaces on the Karakorum N-side (the K2 area in particular). The diameter of the measurement field is (Ø) 100 m for a measuring distance of 1500 m. Measurements: A. Schulze and M. Kuhle.

the traces of a *relief covering,* or a high-level relief-filling glaciation, could possibly be regarded as conclusive evidence of a total glaciation of western Tibet, in so far as its topography as well as its altitude ie lower temperatures and higher precipitation, draw Tibet closer to glaciation than the 1000 m-lower valley floors of the Shaksgam and Yarkand, which are already close to the permanent frost line.

The area chosen for investigation presents an overall *vertical distance* of 7200 m from K2 peak (8611 m or 8617 m) to the Tarim basin (1400 m asl). Composed of several mountain chains, the extremely high and wide Karakorum barrier *prevents* precipitation approaching from the Arabian Gulf and the Bay of Bengal in the south from reaching this study area. Strikingly heavy precipitation during the summer of the 1986 expedition provided evidence of some *monsoon influence* even here in western High Asia. A mean annual temperature of + 3.1°C was recorded at the "Tashikuergan" weather station (3090 m), as the one nearest to the area researched by the expedition.

2.2 Surface Forming Rocks

The bedrock in the area under investigation is important for the *glacio-geomorphological* glacier reconstruction, as well as for the registration of former catchment areas; the direction of run-off is indicated by erratica. In the region of its high peaks and crests, the Karakorum main ridge is composed of dark grey granites and gneisses with strikingly large feldspar crystals (sanidin) (Y 6). They present a very massive structure and show high resistance (Fig 3). Desio (1968) subdivides the K2 structure into Falchan gneiss (Fg), which forms the base up to 6700 m, and K2 gneiss (K2g), which takes over from these and extends to the peak. Further north are crystalline schists of varying degree of metamorphosis with interspersed banks of quartzite, which strike NE to SW and crop out vertically in the area of the K2 glacier and the Muztagh valley (Pt 1: Y δ 2/5, see Chinese Geological Map 1:1 500 000, sheet 1) (Fig 10, 11, 15). The analyses (A. Heydemann) of the samples which the author had collected here at altitudes between 4200 and 5200 m asl served to identify rocks like hydrothermally decomposed, marked granites and epidote schists. In the area where the Muztagh valley joins the Shaksgam valley series of limestone (calcites, Fig 117) (Fig 31), consisting of massive reef limestone outcrops, exhibit well preserved *glacial roundings, abrasions and polishings.* Besides granites severely marked technologically by diaphthorisis, the Sarpo Laggo and Skamri valleys (the catchment area of the Muztagh valley) yield siltite-biotite-schist, chlorite schist, biotite- and di-mica granite, mica schist, and silt- and sandstone, all of which are more or less tectonically or hydrothermally marked. North of the fold on which the Shaksgam valley is situated, the dolomites (P1; K2) of the Aghil range set in, building up the immediately adjacent, steep, orographic right-hand valley flank (cf. Fig 117; Fig 85). The peak region of the Aghil, too, is partly composed of granites (Y δ 2/5). Limestone as well as granite outcrop in the valley N of the Aghil pass (Fig 87; 90). The Aghil N-slope consists of granites, quartzites and reddish sandstone series (P1, K2), which slope down to the orographic left-hand tributary valleys of the Yarkand valley, and to the valleys and gorges of the Surukwat (Fig 138, 22). The Yarkand valley extends from the confluence with the main Surukwat valley up to the military station at Mazar through phyllites of differing degrees of metamorphosis (S2–3 and S1) (Fig 103; Fig 105). A granite zone projects into its orographic right-hand side from the NW (Y δ 1/5), forming the very *resistant core* of the Kuenlun main crest (Fig 60, Nos. 1, 2; Fig 61, 64), and enables narrow, steep-flanked gorges and V-shaped valleys to develop among the transverse valleys.

Within the area under investigation, not only the rock composition of the main crest of the Kuenlun range but also of those further down to the N alternate between granites and crystalline schists (metamorphites). 4950 m high, the "Mazar Pass" consists of metamorphosed sedimentary rock (phyllites) (S1) (Fig 54). According to the author's petrographic samples analysed by A. Heydemann, the metamorphic sedimentary rocks of the Yarkand valley E and W of Mazar contain quartziferous marble, tectonically marked greywacke and sandstone, chlorite schist, albite-amphibolite, chlorite-amphibolite, mylonite,

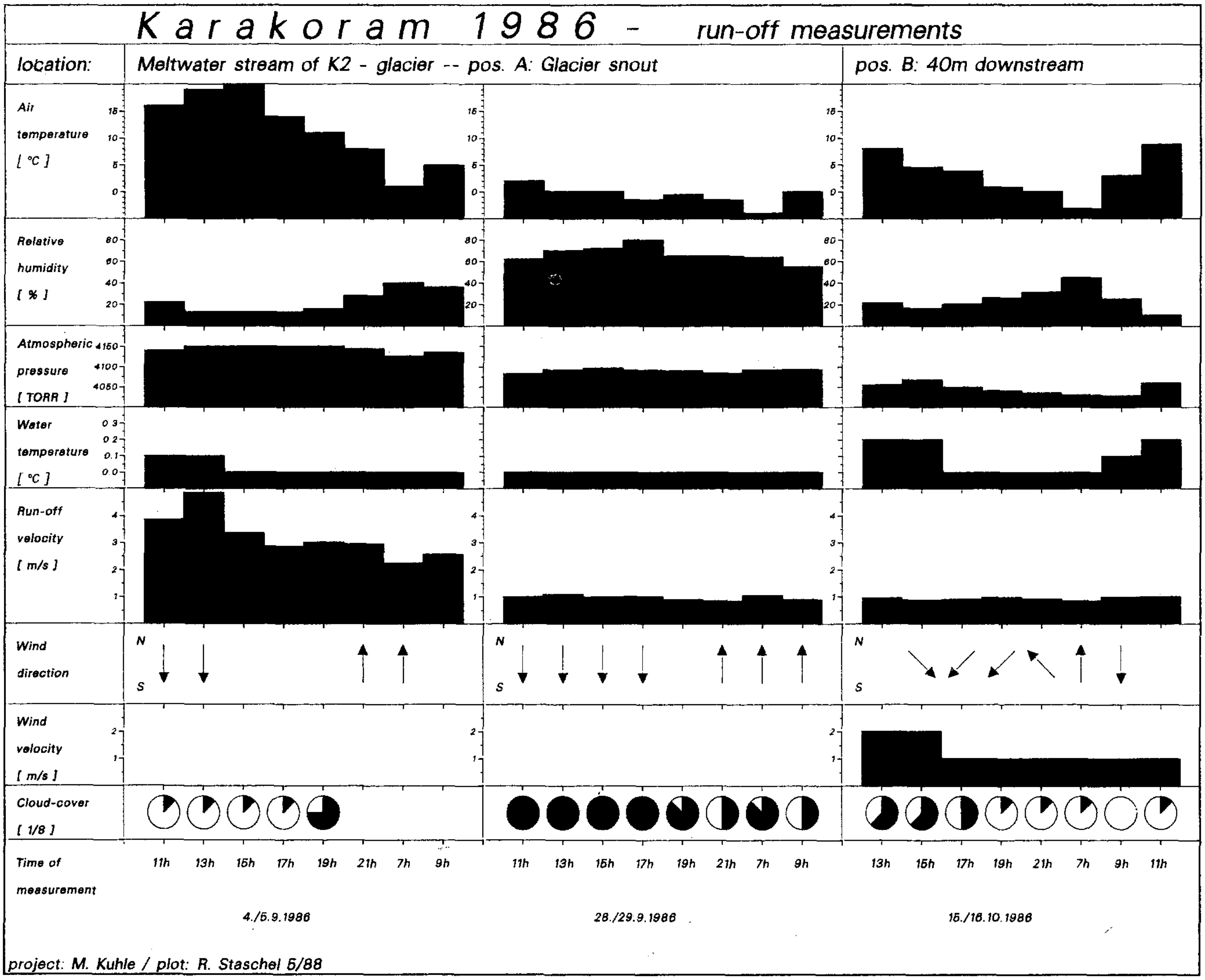


Fig 29 Climatic parameters and meltwater run-off on the K2 glacier (measurements by H. Dietrich). Representative interplay of the essential climatic parameters for the run-off situation in Karakorum glaciers of medium length (15–25 km). Increasing glacier length delays and mutes the reaction of the quantity of run-off on the weather.

di-mica schist, biotite schist, siltstone, muscovite-bearing quartzite, quartzitic sandstone and a quartz-pegmatite dyke. In accordance with resistance, the more than 6000 m-high main peaks of the Kuenlun are composed of the granites (Y 1/5; Pt 2/2) (Fig 21; 53; 58). In the N, the Kudi valley down-stream, crystalline schists in varying stages of development, metamorphic sandstones, marl, mud- and siltstone alternate with granites; together with limestones they present the surface-forming rocks on the northern edge of the mountains in places where the rocky ribs of the Kuenlun slope dip under the alluviones and foothill moraines towards the Tarim Basin (Pt 1/2; Pt 2/2; Pt 1/3:Z; Pt 3; P 2; C 3) (Fig 112). Information beyond the listed observations in the field and the laboratories and their petrographic classification may be found in the Chinese Geological Map 1:1 500 000 sheet 1 (see above).

3 Present Glaciers in the Karakorum, Aghil and Kuenlun and their Balance

3.1 The Karakorum North Slope

This is not the place to present a description of the present glaciers of the Karakorum, particularly since the exemplary research carried out by Visser (1934, 1935, 1938) during expeditions in the Shimshal area (NW Karakorum, leading to the S-slope) and to the Saltoro-Saser and Rimo-

Muztagh (E-Karakorum) has produced a survey of this glaciation. Here only *some* of the *main features* of the Karakorum glaciers in the area under investigation on the Central Karakorum N-slope will be addressed with regard to *qualities* of the local high mountain glaciers already present during the *Ice Ages.* In former times, in the face of an ELA depression of 1000 m (cf. below), these glaciers even flowed down from lower mountain groups now no longer covered by glaciers. It follows that these remarks contain the intention to present the situation by means of the *actuality principle.* In the case of the Karakorum glaciers especially - which are now forming the world's most extensive valley glaciation outside the arctic regions - *compound glacier feeding* (cf. Visser 1934, pp. 137-139; H. J. Schneider 1962, pp. 278-281) from primary precipitation in névé basins (Fig 1a, O and 2, O) or on firn stream sections prevails, together with the secondary feeding from avalanches (Fig 1a ↓ and 2↓, 3 ↓▽, 4 ▽, 12 ↓▽). Measurements carried out in the highest catchment areas above the snow line along the 23 km long K2 glacier between mid-September and mid-October showed c. 70 cm of freshly fallen snow (Fig 1a □ left). At the same time, medium-sized to *very large ice avalanches* (Fig. 1a ↓) were observed with considerable frequency (cf. Shen Yongping 1991, pp. 249-254). Discharges of this kind left from the séracs of the 3400 m-high NNE and NW walls of K2 (Fig 3 ▽ ↓; 4 ▽). In the course of their 1500 m drop the large *ice avalanches* disintegrate into powdery ice avalanches. They continue along the two glacier source branches for some kilometres, and even surge up against the opposite slopes. In some cases the particles of such giant avalanches even leapfrog across the next transverse ridge (ie Fig 1a between 3 and 6). The author observed table- to room-sized gneiss boulders which had been torn out of the wall by such avalanches (cf. Fig 5 O, 6 O). In all four 15 to 43 km-long glaciers, with a total area of 710 km^2 have been investigated; they range from the firn caldron to the firn stream type with compound feeding (cf. Kuhle 1988, pp. 561, 562, Figs, 4, 5).

The examples of the K2 and Skyang Kangri glacier, together with the *largest* glacier in the study area, the 43 km-long Skamri glacier (Fig 9, 47), present classic dendritic valley glacier systems, the tributary components of which are often linked by *ice falls* extending over several hundred metre-high *confluence steps* (Fig 1a, 6, 9, 10, 11, 12, 15, 44). The formations of ice pyramids, up to 25 m-high (Fig 10, 11, 12, 13, 15) are not only the result of the combination of overlying and subjacent glaciers as defined by Visser (1938, pp. 57-74), but also of the hanging and overlying *avalanche cones* (cf. Kuhle 1987a pp. 207-212) (Figs. 1a, 2, 5, 12, 19) which dip towards the glacier. *Ice avalanche transport* also explains the up to one metre-thick and approximately *horizontal inter- morainic strata* between overlying and subjacent glaciers (Fig 17 ■). The *mechanism of ice avalanches* is able to tear the till out of its embedment in the steep and up to several kilometre-high glacier walls (see above). This explains the accumulation of debris at the foot of the overlying glacier (■) *without* preceding melting

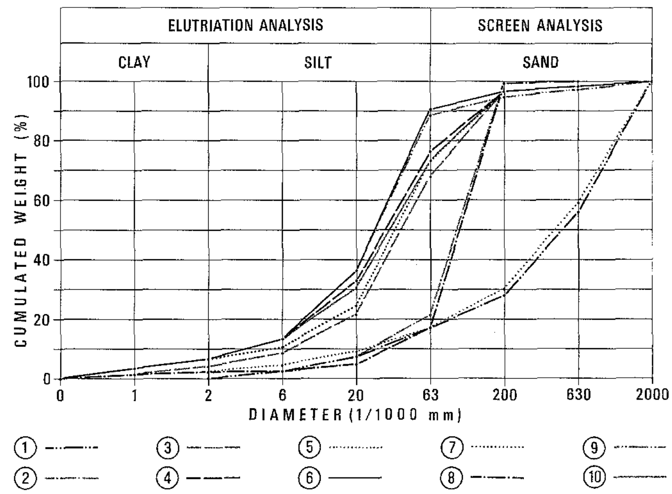


Fig 55 Cumulative Frequency Grain-Size curves from different clastic materials from the karakoram region. Quaternary sediments in the arid high-mountain environment between the Aghil main chain, over the western Kuenlun to the northern piedmont areas leading down into the Tarim Basin (37°46'N/36°24'N; altitud between 1480 and 3900 m asl), sub-2 mm fractions. Curves 2 and 8 show grain-size distribution from widespread, 1 to 2° inclined glaciofluvial fans on the piedmonts at 1480 m, more than 60 km away from the solid rock of the mountains. Curve 6 (till; 3740 m) shows the same characteristics as curve 9 (mudflow). In both cases the source rock is the Kuenlun granite, and the distance of transport amounts 15 km. Curves 1 and 5 are characteristic for alluvial terraces at the bottom of the Yarkand Valley (3800 m asl); the material contains granites from the Kuenlun slope and metamorphic rock (clay, silt and sandstones) from the Aghil slope.

processes of significance. It follows that it can also be brought about by the avalanche mechanism *only.* Having been torn out of the steep wall or wall gorge by the ice avalanches (cf. Fig 3 ↓) the boulder debris is taken up by the avalanche at the base, ie the bottom side. It will maintain this position among the downward-moving masses because its density is two and half times that of the falling ice debris, and is thus able to achieve a greater velocity of fall thanks to its *proportionately more limited* air resistance. The early *separation of the position* of the two materials as a result of their *specific gravity* during their fall from the wall is largely maintained right through to sedimentation in the ice avalanche cone at the foot of the wall, ie mixing does not take place.

3.1.1 The upper climatic glacier limit in the Karakorum

According to the detailed description of the Himalaya and the specific example of the Mt. Everest massif, an *upper climatic glacier* limit can be determined (Kuhle 1986, 1986 i). Wherever High Himalayan peaks rise above 7500–7700m, this limit can be seen distinctly not only between 7200–7400 m asl but can also be established at the same time on a numerical-statistical basis by regression analysis based on telemetric measurements of surface temperatures taken by infra-red thermometer. On Mt. Everest (27°59'16"N) a level is reached at 7200–7400 m where surface temperatures of the mountain flanks almost never reach temperatures near the melting point - it remains permanently frozen. On the contrary, average temperatures here remain at about -20 °C to -30°C, rarely rising above -10°C, so that a snow-to-ice metamorphosis as a result of sintering and ice bridge formation between the snow crystals *relatively warm* snow temperatures bring about, does not take place. At this altitude "temperature metamorphism" near the melting point is merely replaced by much slower bending processes of the *molecular diffusion* and by *pressure compaction,* so that the snow is blown off the steep mountain flanks before it is able to acquire sufficient local stability by adherence to the underlying rock (Fig 4 O). At this altitude *large* areas of rock are visible and the aforementioned *upper glacier limit* is reached. With ridges and summits rising to about 6500–7000 m asl, the peaks W of K2 have not penetrated that upper glacier limit (Fig 18) since even their steep parts are almost completely covered by *flank ice* (▽) and support permanent metre-to-decametre thick cornices of ice, firn-ice and firn (↑ ↓O). The same applies to the c. 6300–6500 m-high peaks of the orographic left-hand catchment area of the Skyang Kangri glacier (Fig 2). In spite of prevailing *high* wind velocities in the summit areas of the mountains, the ice and firn covers of the peaks are not diminished (Fig 12). Fig 4 shows the WNW wall of K2 (No. 1, 8611 or 8617 m), featuring a *vertical dissection.* The *lower* flank section is covered by steep flank ice which is pierced by similarly iced-up rock heads, ribs and precipices (Fig 4 ▽). On the *flattened* mountain shoulder, which borders on the lower steep section (Sla. Savoia, 6626 - c. 6700 m asl) decametre-thick superimposed layers of ice and firn can be identified at the break lines of the ice cascades (Fig 4 ▽ right). Higher up, at about 6850–6900 m, they are followed by almost ice-*free rock wall surfaces,* which form the final rise to the peak (Fig 4 O). *Small-scale* embedments of firn ice, firn and snow only appear in the ac- and ab- granite clefts and in its banking joints, on narrow rock ledges and cornices behind rock precipices, pillars and minarets, and in many scarfs left by smaller rock-falls (Fig 4 above - - - -). Up there the upper glacier limit has been reached and surpassed. The altitudinal value of the *upper glacier limit* at about 6900–7100m asl is, of course, only to be regarded as an approximation in this place (- - - -) since *steepness of walls* and *wind exposure* exert a modifying influence upon the *effective* upper limit between flank ice surface and exposed rock surface. *Corrasion* and *deflation* raise the snow particles from the underlying rock surface, while varying *inclinations of slopes and walls,* together with gravity, cause *downward displacement* of particles of snow and firn. The shifting of particles at the same altitudinal level which would take place on horizontally extended planes, is replaced by *downhill transport.* A glance into the N-wall of K2 (Fig 19) provides more detailed information on this correlation: in spite of even greater steepness (45- 70 ° overall) than that of the WNW-wall, 70% of its lower part is covered by thick flank ice which is forming a dump escarpment (Fig 19 ▽). But even the remaining 30% of the surface, where the rock is yet steeper and consequently unable to provide a hold for those flank ice ramps, is almost totally *encrusted* with ice and firn ice (Fig 19 below \/). Higher up, between the altitudes of 6600 m and 6800–6900 m, this glaciation of walls ceases though there is *no change* in the angle of the wall (Fig 19 \/). Lower temperatures on the N-wall are the reason for the upper glacier limit on the WNW-wall setting in at a higher altitude than on the N-wall. Even further above - and this provides a kind of counter-check - the wall recedes and becomes *flatter.* Yet once again the layers of ice and firn do not increase (Fig 19 over \/; Fig 3), but remain as they were before and the same applies to the WNW-wall above 6900–7100 m (Fig 19 above - - - - right). There is one striking exception: the hanging glacier (Fig 19 □) in the lancette-shaped and widened gorge in the wall which, starting immediately below the summit, descends about 2000 m, from 8600m to 6600 m asl (Fig 3, from O to ↓). The hanging glacier insert (□) appears in the upper 1000 metres into steep spalling at about 7650 m asl (Fig 19 ◆, 3 ▽). Below, confined on both sides by marginal furrows (Fig 3 ▽ to ↓), the wall gorge sets in, its here still relatively wide front eroded by spalling ice avalanches; this hanging glacier in a wall niche owes its existence to a *stable leeward slope position.* Sheltered from winds, this position prevents deflation, whilst causing the continued supply of *drifting snow* from SW, S and E mountain flanks through the formation of *lee vortexes.* An analagous situation occurs on the N-wall of Mt. Everest where, protected from winds by the Norton couloir, a comparable hanging glacier with balcony-like spalling was able to develop in an equally stable leeward position at 7450 m asl (Kuhle 1986, p. 152, photo 3). This evidence of a *climatic upper glacier limit* includes a *reduction* of glacier feeding areas beneath very high mountains. In this case *only some* of the snow supply from the NE, N and WNW-walls of K2 enters the secondary feeding of the K2 glacier. The other part of the snow masses in question is *blown out* of its catchment area, and drifts several kilometres further E on the prevailing W wind into the catchment area of the adjacent Skyang Kangri glacier. These glaciological-geomorphological observations were supplemented by telemetric measurements obtained with infra-red tele-thermometers during the warmest part of the year and day and statistically evaluated for the *thermal validation* of the climatic upper glacier limit (see Tab 1). The *climatic upper limit of glaciation* is becoming much more important for the main glacial one because it had been depressed by c. 1200 m, and thus almost lowered to the wall-foot area of the peaks towering above the present valley glaciers (cf. Kuhle 1986 i,

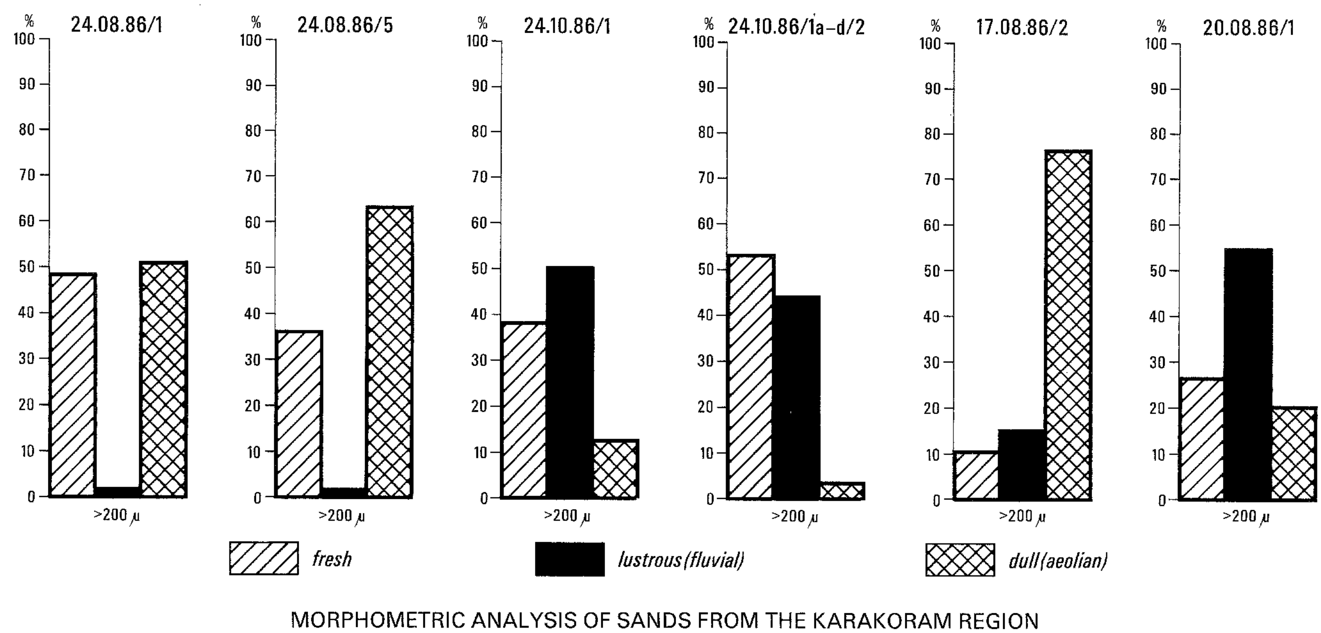


Fig 56 Surface texture of sediment grains greater than 200 μm. Samples 24. 8. 86/1 and /5 (cf. Fig 66, Nos. 1,7) show the characteristics of very briefly transported (less than 10 km) gravel and talus material from a canyon in the Kuenlun: fresh, angular grains (I) have the same portion as the aeolian grains (III), while the fluviatile-polished material (II) is almost completely lacking. Samples 24. 10. 86/1 and 24. 10. 86/1a–d/2 were all taken from outwash terraces in the semi-arid Yarkand valley; however, the distance to the source areas varies. With a greater transport distance, sample 24. 10. 86/1 shows a predominance of group II (fluvial grains), followed by the aeolian material (III) and has only a very small portion of group I. Due to the comparably long fluviatile transport of sediment 17.08.86/2(cf. Fig 55, No. 8) group I (angular, fresh) takes almost no part in it. In group III aeolian transport predominates which is typical for sediments of 200 μm in full-arid environments. Sample 20. 8. 86/1 (cf. Fig 55, No. 6) shows the characteristics of a till at a distance of up to 15 km from the source area. The fluviatile-polished grains (II) are predominant; it cannot be determined on each grain whether the group I grains must attributed to normal weathering or to glacial erosion and transport, but most of this substratum has been pounded and fractured by the glacier and shows cresent-shaped chattermarks on the grain surfaces. The aeolian material (III) still has a portion of 20% and seems to have been blown from a distance.

pp. 344/45). This resulted in a *widespread lack* of Ice Age valley feeding from the high peaks and steep flanks towering above the extensive glacier areas as "dry" rock structures, free of ice and firn. At the same time this was also validated by raised glacier levels. The relative height of the steep flanks was thus shortened from below. Hence *primary feeding* prevailed during the Ice Age, whereas avalanche feeding predominates now and is recognized as a characteristic feature of the Karakorum glaciers (cf. Ch. 7).

3.2 The Aghil Range and the Kuenlun Section of the Study Area

In contrast to the Karakorum, the present extent of the glaciation of these mountain groups is small. Higher peaks reach altitudes of max 6858 m, but there are no major expanses above 6000 m asl, so that significantly *smaller* catchment areas are available for glacier feeding when the climatic snow line runs at about 5300 m asl (Fig 138). In this respect even the *valley floor levels* are important; thanks to its central position in the mass uplift area of western High Asia, ie its greater distance from the relative erosion level of the piedmont areas, their mean elevation in the Karakorum is several hundred metres higher (at about 4100–5500 m on the Karakorum N-slope) than in the Aghil and Kuenlun. But even in the Karakorum *hanging glaciers* and *corrie (cirque) glaciers* (Fig 20 □) occur wherever catchment areas as lowered, upper denudation level on marginal narrow peaks (interferences of slopes and walls in the peak) rise somewhat above 6000 m. In contrast to the Karakorum, these small glaciers represent the *characteristic type* in the Aghil and Kuenlun (Fig 26 ○). The small, several hundred to a few kilometres-long hanging glaciers certainly *dominate numerically,* whereas in the Aghil only a single valley glacier extends to 8km; it is located on the N-exposition of the 6750m massif (36°20'N/ 76°20'E). The remaining 15 valley glaciers in the area under investigation in the Aghil are shorter, and have smaller surface areas (Fig 22 □, 138). In spite of the larger glacier area in the Kuenlun section of the study area, the 13 fully formed valley glaciers do not reach more significant length either (Fig 21 ○). The *small vertical distance* of the Aghil and Kuenlun glaciers results in glacier tongues dipping only a few hundred metres below the ELA, so that these small glaciers move within a relatively small climatic altitudinal belt. As these are *arid* mountain areas, the *mass turn-over* of these glaciers is *small* (Kuhle 1990 c).

These glaciers are chiefly maintained by *low* mean annual temperatures, and less so by the small quantities of precipitation. These conditions correspond to those of the extensive Karakorum glacier near the ELA where relevant measurements were carried out during the 1986 expedition. The ELA on the K2 glacier was determined by the Lichtenecker method (1938: uppermost middle moraine boulders emerging from glaciation plus 50m) as running at 5300 m asl. According to the gradient of 0.6–0.7 °C/100 m measured on the K2 glacier between 3960 and 5330m asl, the mean annual temperature at the ELA is −10.1° to −12.3 °C. This is the equivalent of the ice temperature +1 °C at a glacier depth of more than 10 m, and establishes the identity of *cold-continental* glaciers (Kuhle 1987 i, pp. 413–414). On 6. 10. 1986 Xie Zichu et al (1987, p. 10) measured −6 °C at an ice depth of 10 m at 5300 m asl.

In contrast to amounts of ablation measured in the Karakorum (cf. below) these values are transferable to Aghil and Kuenlun. These are evidence of a *cold type of glacier* found there, corresponding with the terminology of Lagally (1932) and Ahlmann (1935). Only glaciers with tongues that descend a little too deeply produce seasonal run-off, thereby changing from the typological categories of cold to *temperate glaciers*. This applies only to the *valley glaciers* of the Aghil and Kuenlun (Fig 21 ○, Fig 22 □). Unlike the Aghil and Kuenlun this *thermal* glacier type is *representative* of the Karakorum (Fig 9, 23 ◇, 45, 46 left quarter). This juxtaposition explains an essential dependence of the thermic glacier type on *vertical distance:* the further the glacier catchment areas rise above the ELA, the further the glacier tongue ends descend into warmer climatic belts, with the result, that their part in *melting ablation* is strikingly more important than that of the smaller glaciers of the Aghil and Kuenlun, which are almost exclusively concerned with *sublimation ablation* (evaporation ablation). Thanks to their small thickness, the very short glaciers and *firn shields scarcely flow at all,* and thus form soft, cushion-shaped, convex, marginally rounded, superimposed strata of ice lacking séracs (Fig 24, 87 upper margin, 1 a ◇ left 5 ◇); some of these contours are caused by *wind forms* such as wind flutes and snow ridges due to drifting (Fig 25 ○). The rather larger, more dynamic hanging glaciers, on the other hand, which already taper off in tongues ceased in downward direction with scarped margins, typical of cold glaciers (Fig 6 ◇◇, 20 □ left, 21 ○, 26 ○, 27 ○, 86 left side, above). Though typical of Karakorum glaciers there is a *lack* of ice pyramids in the Aghil and Kuenlun glaciers. This is not the result of climatic differences, as *global radiation* responsible for *ice pyramid formation* does *not* decrease significantly over the short distance of a few kilometres to the north. To begin with, this difference is due to the much smaller *length of the glaciers.* Ice pyramid formation requires a glacier tongue of more than c. 5 km *below* the snow line (ELA). A second factor is the necessity of a *steep* catchment area with avalanche feeding or over-thrust of a tributary glacier, so as to provide an overlying body of ice or a hanging glacier. Thirdly, glacier fractures (ice falls, séracs) preparing ice decomposition are favourable.

3.3 Further Features of the Mass Balance of Present Glaciers in the Area under Investigation

The *ablation* rate of the K2 glacier amounts to c. 1200–1300 mm/year, of the Skamri glacier c. 1300–1400 m/year and of the Sarpo Laggo glacier at least about 1500 m/year (Ding Yongjian 1987), whilst values of around 600–700 m/year are indicated by Ding (1987, p. 15, Tab 5) for the NE Karakorum glaciers. Evidence of 5 cm-thick *stalky water ice* among seasonal superimposed ice at 0.4–0.8 m below the firn surface ("infiltration congelation zone") at the level of the snow line (somewhat below a cold "recrystalline infiltration zone") also argues for the view that the classification of glaciers as presented above does not do justice to the *hygric change* in all its features when considering the altitude and the great *temperature amplitude* towards the snow line level. Compared with the glaciers of the Himalaya N-side and the Qilian Shan the more extreme division of the Karakorum glaciers into a *semi-arid, warm-in-summer zone* of glacier ablation next to a *cold-humid feeding area* is striking. These sub- *continental* characteristics apply to the glacier systems in question on the Karakorum N-side which terminate at about 4000 m asl. They are reflected in the measurements taken on the K2 glacier in the period 1. 9.–18. 10. 1986: 7 precipitation events at 4120m asl (glacier termination) produced a total of 1.0 mm; at 4660m asl in the period 11. 9.–8. 10. 1986: 15 events - a total of 33.5 mm; at 5150 m asl over the period 12. 9.–8. 10. 1986, 15 events - a total of c. 100 mm; snow profiles show *exponential* precipitation increases up to altitudes of 5500 m. In the almost 5 decades since 1937 (cf. Spender map 1:250 000) the K2 glacier (92 km^2) has receded by 1.8 km (36.7 m/y; Fig 23; 30a) (ice margin position: Fig 138 No. 1.; 36°03'N/76°27'E). During the same period the Skamri glacier (380 km^2) has receded by 0.2 km (Fig 9; 47; 49) and persists in this trend as does the Sarpo Laggo glacier (Fig 44, 46 left); whereas the K2 glacier has been advancing again since about 1983. The glacier continued its advance even in 1986, as shown by the *fully convex* front of the glacier tongue (Fig 23, 30a). The advancing tongues and submargins of the 50 *smaller* hanging glaciers, with surface areas of a few km^2 only, are also evidence of a continuously *positive mass balance* on the K2 north slope in 1986. In 1986 consequently the product of temperatures, precipitation and radiation favoured the glaciation (concerning the Alps cf. Kuhn 1983, p. 90). Below the altitude of 4600m asl, the *surface moraine* resulting from avalanche processes, influences the ablation process. At this altitude the K2 surface moraine exceeds 20–30cm (Fig 28, 10 x), and *reduces* the ablation rate to c. 40

mm/ summer month. These observations were carried out between September and October 10 th, 1986. With a superimposed moraine layer of only 3–8 cm (Fig 13 ■; 7 ○) the *most significant* ablation rate of 250–330 mm in the K2 glacier is achieved in this same period.

In the snow line profile short-term calculations of velocity showed K2 glacier movements (Fig 1a near □ on the right; 6 foreground) of at most 10 cm/day. Below the major confluence of tributaries on the orographic left-hand, at about 4750 m (Fig 10, tributary below No. 4) the daily yield in the late summer of 1986 rose to 20–35 cm/day. Fig 29 records the *meltwater run-off* of the K2 glacier from the beginning of September to the middle of October 1986, and establishes its relationship to the *atmospheric conditions* by measuring the climatic parameter, the interplay of which governs the run-off. The run-off of the melt-water stream analysed (Fig 23 ◇), which emerges from the glacier cave (○) not only contains water from the K2 glacier with its tributary ice streams, but also that from the 8.6 km-long Skyang Kangri glacier (Fig 2; 12) which fails to reach the main valley glacier (Fig 11; 30), so that it drains a total glacier area of c. 143 km^2. After flowing *sub-aerially* for more than 1 km (Fig 30 □-6) down-valley from its glacier outlet in the area of the tributary valley mouth, the meltwaters of the Skyan Kangri glacier enter a cave-like and evidently permanent glacier ice ponor and flow away *under* the main valley glacier (K2 glacier) (Fig 14 ∪). The meltwater discharge of the K2 glacier stream is *representative* of medium-length Karakorum glaciers on the N-slope as functions of air temperature, relative humidity, wind direction, wind velocity, cloud cover and air pressure. The parameters were recorded at the glacier outlet, at 4100 m asl, and again 40m lower down (4060 m asl) in a distance of 1.24 km from the glacier outlet. The late *summer run-off/graph* (measurements were taken – as far as the weather was concerned – on representative days, 4.–5.9.1986) represents *air temperatures* and the curve of *global radiation* with a *minimum of delay.* The more pronounced "run-off peak" between 12 am and 2 pm is explained by the fact that around midday the *insolation ablation* extends furthest up the glacier, whereas *air temperatures* reach their maximum one or two hours later in the afternoon. Though increasing with the flow velocity, the rate of discharge is unlikely to be greater at the height of summer (July and August) than during the prolonged period of fair weather recorded by the expedition in the early part of September, 1986. In October, the *high-continental winter* of the Karakorum N-side has already set in. The almost identical "run-offs" from 28./29.9 and 15./16.10.1986 are caused by *differing* weather conditions which *cancel out* the seasonal difference. For reasons of expedition logistics (the return journey over the pass had to be completed before the onset of wintry snowfalls at the end of October 1986) the discharge in winter could not be observed. It is likely to be almost nil in December and January (probably from November to February), considering the marked frosts of this season at altitudes above 4000 m. The exemplary K2 glacier can

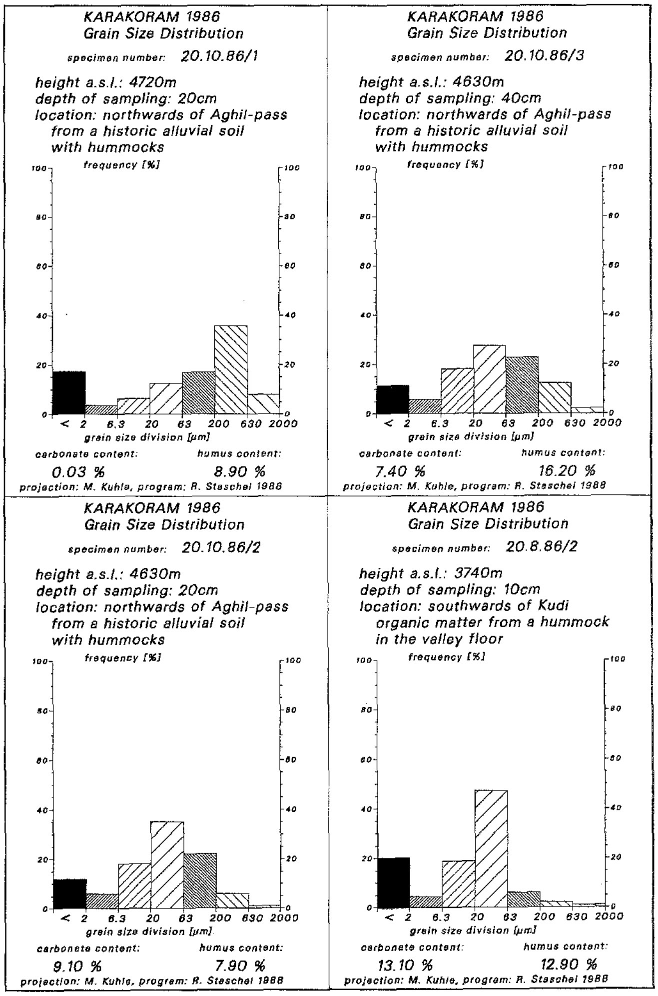


Fig 57 /a (20.10.86/1) Locality: 36°12'10"N/76°36'30"E; /b (20.10.86/2) and /c (20.10.86/3) Locality: 36°13'05"N/76°36'E (Fig 138, Nos. 25 and 26) /d (20.8.86/2) Locality: 36°40'25"N/77°04'05"E (Fig 138, No. 17).

therefore be classified as a *temperate* glacier with seasonal run-off.

4. Selected Observations Concerning the Sequence of Historical, Neo-Glacial and Late Glacial Glacier Positions

In the context of investigations concerning Main Ice Age cover, this section is mainly of methodological importance. The most profitable approach in matters of glacier reconstruction is the *reverse chronological order,* starting with the present state and tracing them back to their *maximum* prehistoric glaciation by way of their historical, Holocene and Late Glacial stages, in other words from fresh, well-preserved forms to the older more severely weathered ones. The intervening stages therefore serve as *bridges for the understanding* of the present mountains and valley forms, which have undergone partial transformation through the substantial effect of glaciers of

sample nr.	sample material	sampling	sampling depth	sample location	sample discovery circumstance	konv. ^{14}C-age (YB 1950)	^{14}C content (in % modern)
25.10.86/1	soil in alluvial terrace	exposure	0,30 m	36°27'20"N/76°57'45"E, Yarkand tributary valley, 4 km westward Mazar, Kuen Lun 3800 m asl	lower terrace below fescue of a spring niche root zone depth 0,6 m; granite gravel body, 1,3 m above the receiving stream	(40+/–80) indirect gravel field, glacier-stadium X	99,5+/–1,0
25.10.86/3	soil	exposure in terrace wall	0,30 m	36°27'20"N/76°57'45"E, Yarkand tributary valley, 4 km westward Mazar, Kuen Lun 3800 m asl	lower terrace below fescue of a spring niche s. 1/3 root zone depth 0,5 m; granite gravel body, above the receiving stream	155+/–65 indirect gravel field, glacier-stadium IX	98,1+/–0,8
20.8.86/2	organic material of earth hum-mocks	exposure	0,10 m	36°40'25"N/77°04'05"E, south-ward Kudi-valley up to Mazar-pass; Kuen Lun 3740 m asl	earth hummock in flood area of the mountain river in valley bottom; granitic sand; receiving stream about 2 m deeper; root zone depth 0,15 m	1610+/–90 indirect outwash, middle Dhaulagiri-stadium 'VII	81,1+/–0,9
24.10.86/4	peat horizon	exposure	0,62 m	36°24'N/76°52'E, Yarkand valley orogr. right	alluvial fan orogr. right of Yarkand river, high water bed with sand from metamorphic rock, root zone depth 0,3 m; reed and grass	110+/–60 indirect gravel field, glacier-stadium IX	98,6+/–0,7
20.10.86/1	organi-cally enri-ched soil horizon, peat from alpine grass	exposure at erosion edge, thermo-erosion in permafrost	0,20 m	36°12'10"N/76°36'30"E, Surukwat-valley, north-wards Aghil-pass, Aghil-range, 4720 m asl	erosion edge at spring grass in pre-recent alluvial fan material on morainic diamictites; granite-limestone-debris mixture; root zone depth 0,25 m	1655+/–180 indirect out-wash cone, middle-Dhaulagiri-stadium VI	81,8+/–0,8
20.10.86/2	organi-cally enri-ched soil horizon, peat from alpine grass	exposure at erosion edge, thermo-erosion in permafrost	0,20 m	36°13'05"N/76°36'E, Surukwat-valley, north-wards Aghil-pass, Aghil-range, 4630 m asl	erosion edge in pre-recent allu-vial fan material with earth hum-mocks on surface; granite-lime-stone-debris mixture; root zone depth 0,2 m	355+/–80 indirect out-wash cone, jounger Dhaulagiri-stadium VII	95,7+/–1,0
20.10.86/3	organi-cally enri-ched soil horizon, peat from alpine grass	exposure at erosion edge,	0,40 m	36°13'05"N/76°36'E, Surukwat-valley, north-wards Aghil-pass, Aghil-range, 4630 m asl	erosion edge in pre-recent allu-vial fan material with earth hum-mocks on surface; granite-lime-stone mixture; root zone depth 0,2 m	6205+/–145 indirect out-wash cone, Sirkung-stadium IV	46,2+/–0,8
15.10.86/1	lower moor peat	drilling core	0,20 m	36°03'N/76°25'20"E, Muztagh-valley, orogr. right, Karakoram 3950 m asl	spring fan, central position homo-geneous spring meadow peat; root zone depth max. 0,15 m; granite and quarzite debris, receiving stream about 12 m deeper, vegetation with earth hummock and hollow formation and alternative stagnating moisture	730+/–100	91,3+/–1,1
15.10.86/2	peat	exposure	0,80 m	36°03'N/76°25'20"E, Muztagh-valley, orogr. right, Karakoram 3990 m asl	riparian exposure below moor homo-geneous spring fen peat; root zone max 0,30 m deep; granite and quar-zite debris receiving stream 2 m deeper	2575+/–170	72,5+/–1,5
15.10.86/4	mud	exposure	1,00 m	36°03'N/76°25'20"E, Muztagh-valley, orogr. right, Karakoram 3990 m asl	moor, spring grass; pelites from granite and quarzite; receiving stream about 2 m deeper; root zone max. up to 30 cm deep; riparian exposure, 1 m below surface	12870+/–180 ground moraine late glacial period, Dhampu- or Sirkung-stadium IV	20,1+/–0,4
15.10.86/5	peat	exposure	1,20 m	36°03'N/76°25'20"E, Muztagh-valley, orogr. right, Karakoram 3964 m asl	riparian exposure lower terrace a more fluvial influenced area of sedimentation; gneis and phyllite gravel body with accumulated peat layer; root zone depth 0,2 m	3450+/–135 indirect out-wash, Nauri-stadium V	65,1+/–1,1

sample nr.	sample material	sampling	sampling depth	sample location	sample discovery circumstance	konv.^{14}C-age (YB 1950)	^{14}C content (in % modern)
24.10.86/1b	peat samples	exposure undercut by river	1,00 m	36°24'N/76°52'E, Yarkand-valley, orogr. right; Kuen Lun, 3760 m asl	alluvial flood soil with alpine meadow peat; erosion edge in the alluvial fan region; reed cover, sampling depth of the peat clay resp. the roots 100 cm receiving stream c. 60 cm lower; depth of root zone c. 0,3 m	4580+/−65	56,6+/−0,5
24.10.86/c	s. 1b					5935+/−85 indirect gravel field, Sirkung-stadium IV	47,7+/0,5
24.10.86/d	s. 1b/1c					1925+/−120	78,7+/−1,2

Tab 2 Samples for radiometric dating (C14) with their localities and thus detailed description in the area under investigation by the 1986 expedition. Laboratory analysis: M. A. Geyh in the Lower Saxony State Office for Soil Research, Hannover, Germany.

less than Main Ice Age size, although only exemplary localities will be mentioned.

The *reverse chronological* transition from the present glaciation of the transverse valleys (tributary valleys) in the Karakorum to the prehistoric (Late to Main Glacial) glacier infilling of the longitudinal valleys (main valleys) is best recorded within the *inter-connected* piedmont plains of the K2, Sarpo Laggo and the Skamri glaciers in the study area, as this is the place where the *former* confluence of these three medium to large-sized valley glaciers can be reconstructed in detail, such as to form a continuous dendritic glacier system. At the time of the confluence also the Shaksgam valley, as the N-Karakorum longitudinal valley, was linked to and partly in-filled by this glacier system.

4.1 The Moraines in the Forefield of the K2 Glacier

The historical glacier retreat from an ice margin position at a distance of 1.8 km from the present glacier end since the year 1937 has already been mentioned above (cf. Ch. 3.3). The ice margin is marked by converging *block ramparts* running conically to the talweg (Fig 138, No. 1).

An even more distinctive moraine locality occurs at 4060 m asl, the level which the K2 stream has cut down to into the very thick moraines. Here, at a distance of about 2.2 km from the present glacier tongue, the *lateral moraines* which have been preserved along both sides of the valley meet to form a *several* hundred-metre-high arc of end moraines (36°03'30"N/76°28'E; Fig 138 No. 2; 30 a ▸ top). The moraine outcrop walls uncovered by the stream cutting into them show the essential *characteristics* of glacier diamictites (Fig 31 ●). The outcrop front is divided into *diamictites* in the underlying bed (●) and *sorted* outwash sediments in the overlying bed (X). The diamictites show *glacio-tectonic* disturbances (minor faulting of materials of varying grain sizes and consistencies along upthrust slickensides and wedges thrown against one another ◂), flexures and readily distinguishable morainic strata with a typical salinebanking structure (◇). The overlying layers of gravel have been deposited in a small lateral valley, a lateral glacial trough between the valley slope and the right-hand glacier margin. The *compressions,* in particular, point to this moraine complex as *end moraine* rather than as lateral moraine - the latter being joined further up valley (on the right). The granites and metamorphites of the crystalline schists in the catchment area are completely preserved. Fig. 32 shows the lateral moraine terraces (V, VI) on the valley flanks, which provide the upstream link with the glacier surface of which this ice margin had been a part. These *impermeable* boulder clays disintegrate in *landslides* and *mudflows* during thaws and rainfalls and, following the retreat of the glacier and thus the loss of its resistance, have slipped down to the valley floor in recent times (decades) (Fig 23 ↙). There they are washed away by the valley stream (◇), so that none but the very coarse boulders remain on the valley floor (□). Fig 32 shows a *whole series* of lateral moraine edges (V) on the orographic right-hand, which are part of the *end moraine complex* of the outcrop described above (Fig 31). United in a *single* body of accumulations, these 5-7 generations of lateral moraine are younger than the 12 870±180 YBP (Tab 2 sample 15. 10. 86/4.) This C-14 date represents the *oldest* one in the sediments among those of the valley floor in the Muztagh valley (Fig 36 X). This is the locality of the confluence of the K2, Skamri and Sarpo Laggo glaciers which must have been ice-free even at that time.

This indication, this necessary classification in the *post-*Late Glacial period, suggests the neo-Glacial period between c. 4500 and 2000 YBP (cf. Kuhle 1986e, pp. 439-454; 1987c, p. 205, Tab 2; Shiraiwa and Watanabe 1991, p. 404) as the time of the formation of this moraine sequence.

No longer reached by ice, the valley floor level of the confluence area at 3920 m asl is evidence of the possibility that this *sequence of ice margin positions* reflects glacier positions in the ELA depressions of at most 100 m (present glacier termination: 4120 m; pre-historic glacier termination: no less than 3920 m asl; difference in elevation = 200 m : 2 = 100 m). It follows that these *ice margin positions* belong to a *neo-Glacial* to *historical (Little Ice Age?)* orographical snow line level at about 5200 m asl. In view of this ELA depression, the question arises of how old the last junction of the Skyang Kangri glacier with the K2 glacier is (Fig 11; 30; 138 No. 3; 30 a ▸ bottom). In *contrast* to the K2 glacier (main glacier) the Skyang Kangri glacier was *retreating* in 1986. With a debris cover measuring only a few decimetres in thickness, the tongue has *flat* longitudinal and cross-profiles at its end, thus indicating that back-melting is in progress. Squeezed against the southerly valley flank, its *asymmetric* position (Fig 11 ○; 30 xx on the right) corresponds to the shadow cast by the peaks to the S, which shield the ice from the markedly intensive midday radiation of this almost *sub-tropical* latitude. The remnants of *dead ice* on the orographic right-hand (Fig 11 ◇; 30 ◇) are indicators of the width of the tongue end *some years*, or a few decades, ago. At that time - and the *geometry of the ground-plan* of the former glacier tongue is evidence of this - the Skyang Kangri glacier reached the confluence with K2 main ice stream (Fig 30 a ▸ bottom). The ELA which was part of it ran just a few decametres below the present one. The *last confluence* of the two glaciers therefore belongs to the 1937 glacier edge of the K2 (Fig 138, No. 1) (see above), or to an even later advance, as suggested by the air photo (Fig 30 a). During the *preceding* historical to Holocene glacier positions (from the "Little Ice Age" back to the neo-Glacial period) there was a much heavier influx of ice from the Skyang Kangri glacier. It is most easily associated with the end moraines observed 2.2 km outside the present K2 glacier end (Fig 138, No. 2). Evidence of this *somewhat older* confluence is available in well preserved, orographic right-hand *flank abrasions and polishings* with a marked glacial polishing line (Fig 30 ▲, bottom right, 11 ◀ left). At that time the tongue of this tributary glacier *not only* reached the K2 glacier, but also turned entirely into the main valley, where it *settled down* on the right of the K2 glacier. The height of an end moraine gusset (Fig 30 X) provides evidence of the decametre-thick, historical to neo-Glacial glacier confluence. Thanks to *glacial flank abrasion and polishing* up to 200 m above the valley floor (Fig 30 ⋃, bottom right), largely without traces of weathering, and not even showing dark water lines (which only set in above), this confluence is attributed to the "Little Ice Age" (most recent Dhaulagiri Stage VIII or IX according to Kuhle 1982, p. 165 et seq.). In the central Himalaya the Dhaulagiri Stage VIII or IX is credited with an ELA depression of 40 m (Dhaulagiri-Annapurna-Himal; Kuhle 1982, p. 166) or 80–100 m (Khumbu-Himal; Kuhle 1986e, p. 454; 1987c, p. 205). The pertinent glacier level of the most recent Dhaulagiri Stage VIII or IX is preserved by a c. 150–200 m high lateral moraine ridge upstream on the orographic right-hand of the Skyang Kangri glacier (Fig 12 IX, right).

4.1.1. Some aspects of the preservation of prehistoric glacier levels: the case of historic glacier advances

Along the edges of the K2 glacier and its tributary ice streams only, and apparently *inevitably*, a very *sporadic* chain of indications of even the most recent historic glacier levels can be gained. The majority of the edge sections of the valley flanks have meanwhile been undercut by the lowered glacier edge, so that gullies, minor wall gorges and talus cones spilling out from them have become established on the new glacier edge within the short time of a few decades (Fig 34 X↘ ▲). Frost weathering as well as rock falls due to the steepness of slopes, and *denudation* through avalanche abrasion and polishing which also grinds out gullies and wall gorges, all combine to destroy the historical polishing and abrasion limits and the lateral moraine ledges, which are limited to the area below the snow line. The destruction of the latter is accelerated by *undercutting* (Fig 10, right hand quarter; 28 ▽▽; 47 ↑↑; 49; 23). These conditions *contrast* with the abrasion and polishing that is simultaneously taking place below; they apply particularly to the *more steeply-draining* tributary glaciers, the polished or abraded edges of which are topped by deeply gullied, strikingly precipitous rock walls; they have disintegrated as a result of frost weathering, but contain little loose material (Fig 15 ⇐ ⇒; 10 below - - - -; 7 below - - - -). Nonetheless remnants of lateral moraines have ben preserved in a *few* places (Fig 12 IX and X, centre), though at times not a genuine, morphologically-formed ledge, but merely a *ground moraine* or *lateral moraine material* that clings to the valley flank - like the only remains of the inner slope of a lateral moraine (Fig 9 ↓↓; 47 ↑↑; 23 ■). In the lower third of the valley glacier courses there are *ablation valleys* near the S- and W- facing valley sides, the increasing width of which require intermediate space between the glacier edge and valley slope (Fig 30 □-6 centre top), so that a direct glacial denudation process is no longer taking place, nor has it done so in historical times (Fig 14 ○○ -6; 16 ○ -6). In the area of these glacial bank valleys various manifestations of an increasingly *melting* ice margin occur in the form of stages of a valley glacier with a *negative* mass balance. At the K2 glacier (below the junction of the Skyang Kangri valley) *mudflow fans* (Fig 16 ◆; 14 X) from the orographic right-hand valley flank are being deposited in the embankment valley downstream below 4650 m. The floor of the embankment valley consists of glacio-fluvial gravels (-6), partly *covering* (○○) the ice of the K2 underlying glacier (□□). The K2 overlying glacier recedes for a few more decametres from the valley flank than the underlying glacier, thus forming a *hard shoulder* for some of the mudflow and gravel deposits (Fig 16 □). In the valley glacier sections where those para-glacial *kames* and *outwash-like* sediments accumulate now, present flank polishing and abrasion are *suspended*, and glacigenic undercutting has *lapsed.* This development of the K2 glacier set in after the Skyang Kangri glacier ceased joining it, and constitutes the *last stage* in the recent glacier development with *negative* mass balance. The Fig 1a (IX), 5 (IX) and 10 (X) show *another variant of preservation* of a glacier level which existed a few decades ago, but cannot develop unless it occurs *below* the ELA: the relative height of *the lateral moraine on the glacier.* Underneath this kind of *lateral moraine,* the several metres-high thickness of which has frequently been built up by the addition of local debris at the foot of the cliff of the valley flank, older, at times even *dead* (since no longer part of the glacier movements) glacier ice is preserved for decades, ie protected from ablation, thus indicating older, metre-to decametre-*higher, former* glacier surface positions.

In analyses of this kind, concerning indicators of prehistoric glacier levels, it is necessary to take into consideration the *basic fact* that the *level fluctuations* are greatest in the lower glacier sections, near the glacier tongue ends, whilst at the same time the surface height in the feeding area varies only *minimally,* even if the glacier tongue advances or retreats many kilometres. This *up-valley surface convergence* of prehistoric glacier levels *prevents* the initially seemingly obvious synchronocity of individual glacier levels on the basis of absolute differences in the altitudes of the upper edges of lateral moraines in the historic and Holocene glacier tongue basins. At best the respective *altitudinal relationship* of the lateral moraine ledges to one another - staggered upon one another on the valley slopes - would allow a stage to be recognized within the general order (cf. eg Fig 32 V). The *low,* prehistoric glacier levels just above the present glacier surface of the *feeding area* can consequently be synchronized with several tens to hundreds of metres-high *end moraines* at the *glacier end,* or several kilometres further down the valley, as in the case of the historical or even neo-Glacial moraines, in the forefield of the K2 glacier (cf. Fig 31 V). The stages under discussion here are the Sirkung Stage IV (late Late Glacial), the Nauri Stage V and the Dhaulagiri Stages VI to IX; according to the author's nomenclature (Kuhle 1982), they were first introduced for the Dhaulagiri and Annapurna parts of the Himalayas, and subsequently extended to the whole of High Asia (1986 e; 1987 c).

A detailed survey of the *sample valley flanks, selected here* from the K2 glacier valley in view of prehistoric *glacier level indicators* (Fig 1a; 5; 6), and the glacier sections between the upper framing walls of the surrounding ridges above the snow line, and further up to 4600 m asl, ie to c. 700 m below the snow line in particular, reveals that the preservation of prehistoric glacier horizons is *more than scarce.* In this place nothing but an *integral smoothing* of the valley flank on the orographic left (especially on the E-wall of the 7315 m high satellite peak on the E side) can be diagnosed (Fig 1a below No. 4 ▲; 6 below No. 1 ▲, left half of the photo), which is *less* pronounced higher up, where some of it disappears under flank ice and hanging glaciers with some ice balconies attached (◇◇).

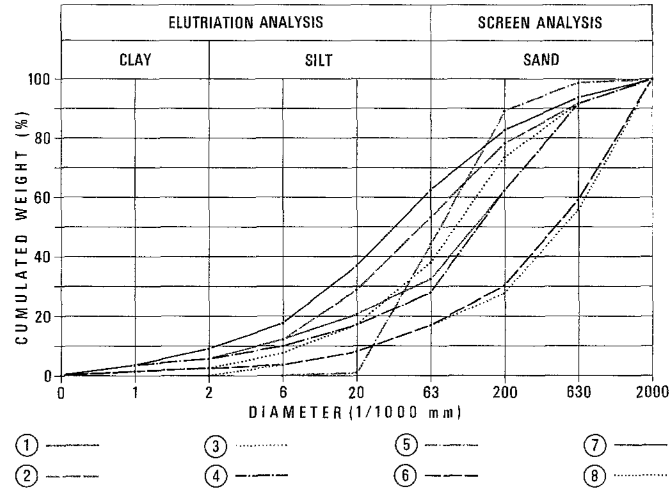


Fig 66 Cumulative frequency grain-size curves form different clastic materials from the Karakorum region. Grain-size curves of alluvial and glaciofluvial sediments. Samples 5 and 8 are from very large gently sloping (1°) glaciofluvial fans on the piedmont. The material is polymict. The distance transported amounts to more than 100 km (3850 m asl). Curves 1, 2, 4 and 7 show gravel deposits with differing admixtures of talus cone material on the bottom of a narrow Kuenlun valley. The granite and metamorphic detritus has been transported 10 km at the most. Curves 3 and 6 show also well-graded gravel deposits consisting of granitic and metamorphic detritus after transport of up to 100 km (middle Yarkand valley; 3800 m asl).

Glacigenic polishings or abrasions are also preserved further downstream. Thanks to the locally differing *interference* of valley wall and rock structure they vary considerably, though truly *significantly better* flank abrasions and polishings cannot be found anywhere (cf. Fig 7 ▲; 10 ▲; 11 ▲ below ----; 15 ◁; 28 ▲). Everywhere in this area, snow inserts and periodic meltwaters in *wall gorges and gullies,* and as a result, frost weathering, rock falls and avalanche processes, have *dissected* and *dissolved* the prehistorically intact rock slopes and areas, which could have shown glacier abrasions and polishings, down to the *present glacier surface.*

Only in places where the altitudes of the catchment area of the valley slopes are comparatively low and existing rock conditions favourable, can glacigenic flank polishings or abrasions be more *readily* reconstructed (Fig 32, ▲ right), albeit *without* actually preserved glacier striae anywhere above the present K2 glacier, Skyang Kangri glacier, Sarpo Laggo glacier and Skamri glacier (cf. Fig 44 and 9 below ----; 2 and 12 below ----). Though glacigenically rounded slope sections do occur in many places on the orographic right-hand valley flank of the K2 valley, large sections of them are cloaked by metre-thick debris covers (Fig 1a ◁; 5 ▽ left hand third of the photo; 6 ▷ right half of the photo;

11 ◁ left third; 10 △ true left; △ true right). Besides the detritus developed in situ, part of these *debris mantles* consists of in places very thick ground and lateral moraine (Fig 10 ⋃; 11 IV, V) deposited on glacigenic slip-off slope sections. Such preserved moraines from a time when the snow line occurred at least 100 m (at about 5200 m asl) below the present one (Fig 5 VI, V; 35 ▷) almost reach up to the snow line (up to 5100–5200 m asl). All these formations of debris and moraines on the orographic right-hand benefit from their W-exposition. Daily increase in temperature coinciding with afternoon radiation both favour *weathering* in the permafrost zone as well as *moraine production* by a process of forced melting.

These observations on glacigenic abrasion and polishing and their *preservation* on the valley flanks *above the present* glacier levels are essential for the assessment of glacial forms in the main valleys of the area under investigation which are a *long* way outwards and down-valley from the present glaciers. In many places of the main valleys glacial rock smoothing and glacier abrasions and polishings are much *better preserved* (cf. Fig 27 ▲; 37 and 38 ▲; 39 ▲; 40; 41; 42) *although* the hardness of rocks is the same. This *contradiction* is regarded as the main reason for the fact that some Chinese authors like Zheng Benxing & Li Jijun (1981), Shi Yafeng & Wang Jing-Tai (1979) and the "Quaternary Glacial Distribution Map of Qinghai-Xizang (Tibet) Plateau" (1991) by the same authors have argued for an only *marginally* more extensive valley glaciation in large parts of High Asia than there is today as the *maximal extent* of *glaciation* during the *Ice Age;* this view has been held for decades, and continues even to these days. The reasoning leading to this error runs as follows: 1. "since the valley flanks do not – or scarcely – show smoothing above the *present* glacier levels, glacier ice during the historic, neo-Glacial or Holocene times cannot have been significantly higher up on the valley flanks"; 2. "what ever was *sporadically* preserved in the way of minor flank abrasions and polishings, and always only a *few* decametres above the glacier levels (cf. above) has been so much transformed that it must be regarded as belonging to the Ice Age at the latest, otherwise the polishing forms ought to be much better preserved". On the other hand the author, who does *not* proceed from such a *linear extrapolation* of present conditions, but rather from *very well preserved,* old and almost completely reworked *young* forms of glacier erosion, takes the view of *qualitative leap-like discontinuities* in the prehistoric process; this view is founded upon very comprehensive findings in the field and the *comparative method.* The authors mentioned above completely reject the glacio-geomorphological key forms determining the entire main valley relief, simply ignoring glacier abrasions, polishings and striae, or vaguely enlisting periglacial "forms of convergence", which have *no empirical* basis on earth. According to the *alpine* example, which the author considers representative in this case, frost weathering has transformed glacigenic smoothings on the mountain flanks above recent glaciers and in their immediate forelands *past all recognition,* within a period of a few decades to centuries. In the Alps, the *purest* forms of U-shaped valleys set in tens of kilometres away from the present glaciers, and the *best* preserved ground abrasions and polishings and landscapes of glaciated knobs are related to transfluence passes which have been free from ice since Late Glacial times. Already beyond the actual glacier tongues in the confluence area of the Muztagh valley (Fig 138 No. 10) the glacial forms of the valley flanks are not only much better preserved in the same bedrock than those up-valley above the present Skamri glacier, which have been completely obliterated by the development of gullies (Fig 9 ▽), but possibly generally only there preserved as *unambiguous* forms (Fig 8 ▲; 36 ▲ ▲ ▲). Looking still further down-valley to the lowest main valley down into the longitudinal valley of the Shaksgam, the quality of preserved, glacigenic abrasion and polishing forms *continues to improve* even up to very high transfluence passes (Fig 138 No 12; 51 ▲; 37 ▲ right; 52 ▲▲; 38 ▲). At a *great* vertical and horizontal distance from present glaciation, glacier *striae* and *polished* surfaces of glaciated knobs in the Yarkand valley, the great valley system further north, are *perfectly preserved* as far down as 3700–3400 m asl (Fig 138 Nos. 46 & 33, 40, 41; Fig 93; 42; 128). This juxtaposition of a *bad* glaciogeomorphological state of preservation in areas of *ongoing* active formation through glacier ice run-off, and clearly well preserved forms from a greater spatial *distance* and therefore also *temporal* interval relative to present glaciation, leaves no alternative but to draw the conclusion that *glacial* transformation leads to a *faster* destruction of forms than fluvial and periglacial processes operating at lower altitudes. There is *no* way for the older and much more extensive forms to be in a fresher condition than the recent ones in the vicinity of ongoing glaciation. However, since in the course of deglaciation, those lower-lying areas must *also* have been *undercut* and *transformed* by smaller (narrower and less thick) valley glaciers and valley glacier tongues, especially on the valley flanks, the active time for those processes can only have been *very brief.* In any case, it must have been significantly shorter than the period of transformation activity meanwhile available to the *Holocene* and *inter-glacial* glaciers. It follows that the more than 1000 m thick Ice Age valley glacier – and ice stream network systems which filled the main valleys during the Ice Age (cf. Ch. 5; cf. for example Fig 37; 39 and 116 - - - -) must have thawed out completely within a short time, ie approximately down to the *present state* of the Karakorum glaciation. This fact is confirmed by the oldest C14 date of 12870 ± 180YBP of the valley floor of the Muztagh Valley (Tab 2 samples No. 15. 10.86/4) since Sarpo Laggo and Skamri glaciers *must* already have retreated from this valley cross-section at about 3900 m asl and terminated in the vicinity of the present ice margins *before* these sediments were deposited. It might even be the case that, in the course of post-Glacial climatic warming, the ELA rose *so rapidly* out of the completely glaciated valley relief, that glacial processes of under-cutting and transformation were not able to take place after a few isolated Late Glacial glacier advances. This is all the more likely, as deglaciation of *large volumes* of ice (which must be assumed for the Main Ice Age, according to the glacier reconstruction described in the following - see Ch. 5) must have taken place comparatively *slowly*, thereby allowing quite a long time for the raising of the snow line, without - considering the relatively minor periods of climatic cooling in between - the occurrence of significant glacier advances, causing glacial *undercutting*. This *fundamental* difference in the glacio-geomorphological course of the glaciated areas close to or at a greater distance from the present glacier has not only a *temporal*, but also an additional *thermal* component: the *altitude* of the glacier tongue ends depends on the snow line in its absolute, as well as its *relative*, altitude in respect of the mountain relief. When the snow line *descends* further into the relief, the glacier ends flow *progressively* - a factor of 1.5 to approximately 2 - further down into the valleys (Kuhle 1987, pp. 208–10; 1987 b, pp. 415–19; 1988, pp. 564–66). It follows that the glacier-free valley areas in the area surrounding the present terminations of relatively small glaciers near the ELA present a colder climate, with many more freeze-thaw cycles than those in the area surrounding Late Glacial ice-streams which have descended *far* into lower altitudes. Highly effective frost weathering in the vicinity of present glacier margins accordingly has a much more *transforming* effect than weathering in the area surrounding pre-Holocene glacier margins.

Summing up, one may generalize as follows: extreme increases in frost weathering near the snow line, which are even more severe in the vicinity of snow margins and on the *black-white* line of smaller glaciers, *transform* prehistoric glacial formation so *much faster* than in the lower regions of glacial advances that, in many places, *Main Ice Age* erosions are *better* preserved than the *Holocene to historic* forms of glacier abrasion and polishing. In this case the influence of the "weathering intensity" factor overtakes the "time" factor (cf. "Höhenstufe besserer glaziärer Formenerhaltung", Kuhle 1983, p. 161).

4.2 Moraines and Glacier Traces in the Muztagh Valley up to its Junction with the Shaksgam Valley (in the Confluence Area of the K2 Glacier, N-Skyang Lungpa Glacier, Sarpo Laggo Glacier and Skamri Glacier)

The confluence with the N-Skyang Lungpa glacier (Fig 138 No. 2; Fig 43 ▸▸) took place at a greater distance from the present K2 glacier tongue, down-valley from the striking end moraine, which can, at most, be classified as *neo-Glacial* (Nauri Stage V, c. 4000 YBP, or older Dhaulagiri Stage VI, c. 2000–2400 YBP) though its base may be of an even *more recent* date, belonging to the period of the middle to younger Dhaulagiri Stage 'VII-X (younger than 2000 to older than 30–80 YBP, cf. Kuhle 1982, pp. 162–66; 1986 e, p. 454; 1987 c, p. 205 Tab 2) (Fig 31, 32). A good 15 km long, this parallel glacier of the K2 ice stream now ends about 6 km from here. Fig 32 shows two *lateral moraine* levels on the orographic right, which developed at the time of the confluence of the two glaciers (IV, III; Fig 138 No. 3 a). At that time, their joint tongue terminated in the Muztagh valley, at about 3900 asl (Fig 43 X). It is probable that at the same time the joint area of confluence with the Muztagh valley was covered by the Skamri-Sarpo-Laggo glacier system, thus creating *another confluence* at this location. Though no direct end moraine has been preserved there, *lateral moraines* (Fig 30 a left hand top; 36) provide evidence of this fact. For chronological reasons the two lateral moraine terraces, which are located up valley at 400-500 m above the K2 valley floor (Fig 32, IV and III) are to be classified as Sirkung Stage IV and Dhampu Stage III (both older than 12 870 YBP), ie the *Late Glacial* period. Their extreme continuations outside the valley are the moraine ledges (Fig. 43, IV, III), a few decametres above the gravel floor of the K2 valley, which is set into the gravel floor of the Muztagh valley (Fig 36, -6). This chronological classification is *obligatory* in the sense that it follows from the fact that from 13 000 YBP at the latest (before 12870 YBP, Tab 2) the ice must have melted from the valley floor area of the Muztagh valley for the development of the *dated* peat clay in the bog with a spring meadow, which is 1 m below the present surface of the valley ground.

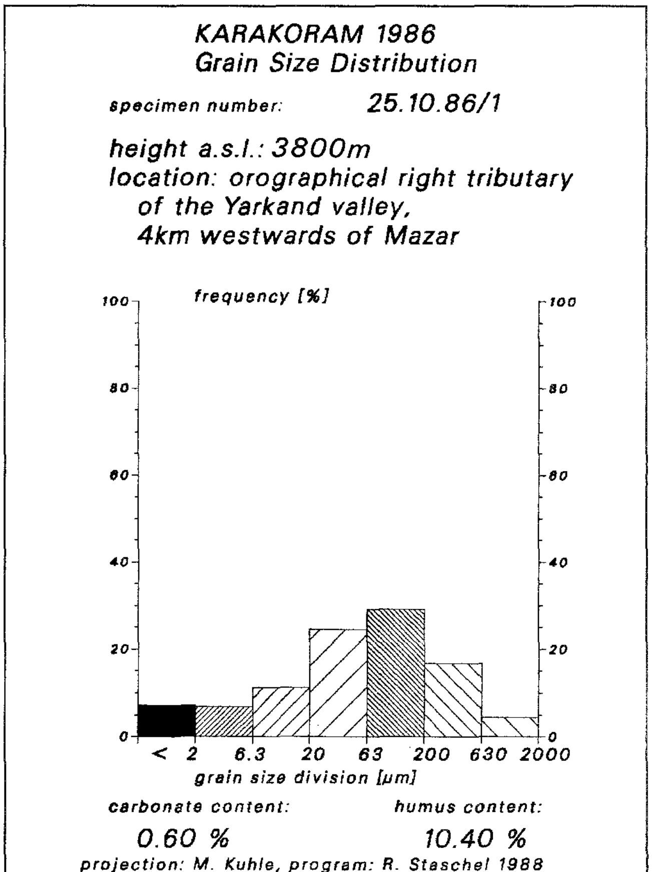
About 32 km long, in as far as its main branch is concerned, the Sarpo Laggo glacier has been in continuing, intensive retreat since 1986 (Fig 44; 46 left). Over a distance of c. 4 km its tongue is completely *covered* with superficial moraine, and split up into numerous large *dead ice complexes* below the glacier outlet that follows the live ice stream (Fig 44 ■). Still "live" in 1986, the glacier tongue end lies at 4200 m in the junction area of the last tributary valley on the orographic left-hand side of the Sarpo Laggo valley; a comparison with the Chinese map on the scale 1:50 000 (sheet 9-43 -ζ; air photograph base) shows that during the past 10–15 years it has *retreated* about 2 km, and 100 m upwards, ie over this distance the glacier tongue has *split up* into the blocks of dead ice mentioned above (Fig 138 No. 4). In its forefield the 1.5 km-wide glacier tongue left behind a complex meltdown landscape, or landscape of glacial accumulation forms, with meltwater capable of transporting sizeable loads meandering and *tunnelling* the dead ice complexes. This landscape of end moraine and thawing consists of three staggered very recent prehistoric fronts of ice margin positions (dumped terminal moraines) (Fig 44 X ■ ■), the oldest of which must been in contact with the glacier terminal as late as 1937 (cf. below 36°N/76°22'E). There is no trace at all of comparable outlines of fresh tongue basins down-valley, thus indicating substantial glacio-fluvial transformation of the valley floor. The dark metamorphite (phyllite) debris is mixed with massive crystalline components (granite, Falchan gneiss und K2 gneiss), which have come down from the Karakorum main ridge; removal and sorting of morainic material released from the ice takes place at many locations. In their area, and 3 km down valley of the disintegrated Sarpo Laggo glacier tongue, two tributary valleys filled with smaller glaciers join on the orographic right (Fig 45; 8). They are linked by confluence steps – one of which has meanwhile been dissected by a steep, V-shaped gorge (Fig 8 ▼) – and release substantial, though only a few decades old *mudflow fans* on to the main valley floor.

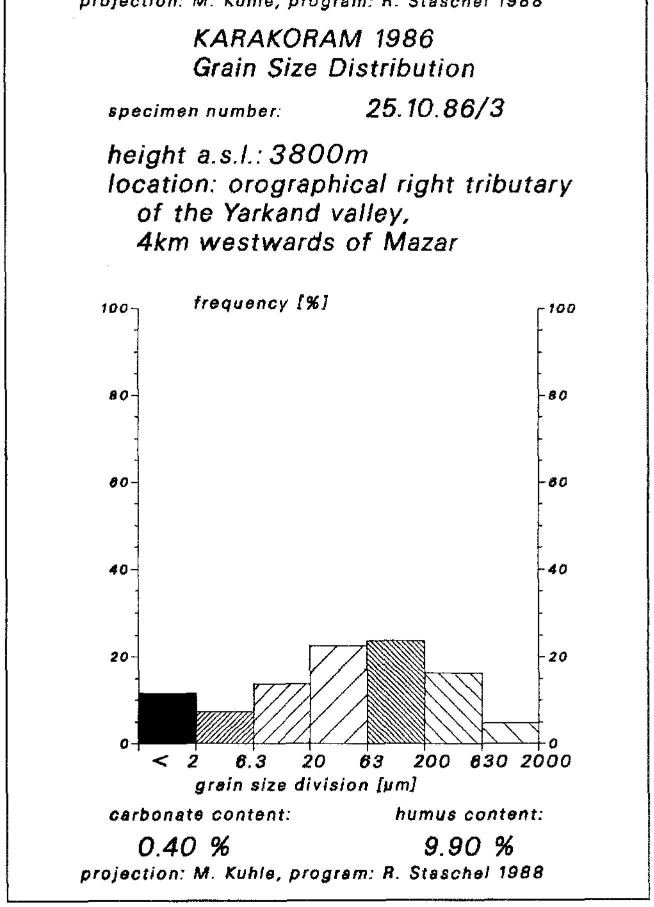
According to the Spender map of 1937 the Sarpo Laggo glacier was still close to reaching the lower of the two valleys (Fig 8), ie at that time its active glacier outlet was located 2.8 km further down-valley than at the time of collecting data for the topographical map (1:50 000), and in a 4.8 to 5 km lower position than the active glacier end in 1986 (Fig 44 & 46 X). It follows that the mud flow fan on the orographic right farthest up the valley can in fact not be more than at most 40-50 years old, ie it has been built up since 1937 (Fig. 45 ▽). Tributary alluvial fan deposits of this kind, as well as glaciofluvial transformation by glacier meltwaters in the forefield, prevented the preservation of an even older (ie than 1937) unequivocal end moraine (if there ever was one).

The previous (older than 1937) positions of the Sarpo Laggo glacier formed 100-300 m high *terraces of lateral moraines* on both sides (Fig 36 IV left; 46 IV). Their upper edges lie between 4100 m and 4400 m asl. On the intact inner slopes of the lateral moraines a total of at least 10–13 lateral moraine edges can be identified, thus giving an impression of the very *frequent* oscillations and variations of level which must have taken place at this glacier tongue. The moraine terraces are classified as belonging to the period of the "Little Ice Age", or younger Dhaulagiri Stages IX to VII, middle Dhaulagiri Stage 'VII, older Dhaulagiri Stage VI, back to the oldest neo-Glacial stage, the Nauri Stage V (Holocene, approximately 4000 YBP). Thus they belong entirely to the post-Glacial period. Up to the older Dhaulaigiri Stage VI, the Sarpo Laggo glacier always reached a confluence with the Skamri glacier, as demonstrated by the interlocking of lateral moraines in the triangular moraine inset of the confluence (Fig 36, IV, IV; Fig 138, No. 5). The only uncertainty of such a confluence concerns the final phase of the Sirkung Stage (IV) and the Nauri Stage (V). The Skamri glacier did indeed reach the Sarpo Laggo valley at that time and formed a *hammerhead* spit in its confluence, whilst the Sarpo Laggo glacier must still, or once again, have been in retreat in its valley embedment, as shown by the evidence of combinations of moraine configurations (cf. Fig 46 ▽ V; 36 ▽ V). Approximately nine lateral moraine dams of the Holocene Skamri glacier on the orographic right transform the left hand lateral moraines from the Holocene Sarpo Laggo glacier. A transformation of this kind has probably been caused by the *difference* in length, or size, of the two ice streams. Due to slower reaction to climatic cooling, the larger Skamri glacier can only have advanced with the end of its tongue when the Sarpo Laggo glacier tongue had already completed this advance, was in the process of retreat, and had thus vacated the mouth of the Sarpo Laggo valley. A corresponding difference in the glacier size-related behaviour of the tongues ends can currently be observed on the *advancing* K2 glacier in contrast to the *retreating* tongues of the Sarpo Laggo and Skamri glaciers (Fig 44 and 47).

Essential facts to be recorded are 1. the approximate *contemporaneity* of the two semi-level moraine terraces in the forefields of the two, formerly joined and still major valley glaciers, and 2. the *youthful* age of these lateral moraines, evidence of which exists in the good state of preservation of those small moraine ledges in the permafrost zone where solifluction is intensive. Only the uppermost areas of these late Late Glacial and Holocene lateral moraine terraces show clearly *periglacial* transformation in the *rounding* of ledges and the formation of *solifluction* tongues (Fig 36, IV; 46, IV). Hitherto only minor incisions of gullies into the soft moraine material, even *beneath* larger *source basins,* now filled with firn shields and hanging glaciers (Fig 36 \\) point in this direction. In the valley cross-profile of its present (1986) glacier tongue end these equivalent lateral moraine terraces on the Skamri glacier (sometimes referred to as Crevasse or Yinsungaiti glacier) reach up to 4450 m asl (Fig 36 IV), thus attaining altitudes of 420–430 m above the outwash plain at the glacier outlet at 4030 m asl. Below the level of the Sirkung Stage (IV) (Fig 36) another three, particularly striking edges of lateral moraine terraces can be discerned from top to bottom along these lateral moraine deposits on the orographic right: The Nauri Stage (V), the older Dhaulagiri Stage (VI) and the middle Dhaulagiri Stage ('VII) to the younger Dhaulagiri Stage (VII–IX). Fig 47 shows the panoramic view from the second highest level of the lateral moraine terrace (Nauri Stage V) at 4330 m asl across the 2.2–2.3 km wide tongue of the Skamri glacier. The Skamri glacier tongue is composed of tongues of at least three tributaries, each *over-riding* glacier streams (cf. Fig 49 □■●; this air photo was taken in about 1975). The three *overthrust* glacier tongues, the lowest of which forms as an underlying glacier tongue the actual and lowest end of the Skamri glacier, have remained *separate* ice bodies (Fig 9 X●). Each overlying glacier has pushed a small *frontal moraine* upon its supporting or respective underlying glacier (Fig 47 □□□), and moreover retrained its original tongue shape (cf. satellite picture NASA ERTSE-2653-04441-701; 0 69 / 0 70; 5 Nov. 1976). The glacigenic *truncated spurs* and *cuspate slopes* are evidence of prehistoric flank abrasions and polishings on the orographic left (Fig 47 ; 138 No. 6). Like the orographic right-hand lateral moraines, their formation dates back to the *later Late Glacial* period. The abrasions and smoothings are *unusually well* preserved in regard to outcropping metamorphites. Though on the same level the rock flank on the orographic right-hand is far more strongly weathered and *dissected* by gullies (Fig 9 & 48 ▽▽; 138 No. 8). Corresponding glacial flank smoothings and cuspate areas, offering evidence of in these places even higher, and thus older *levels of glaciers, filling the entire valley,* are preserved in the middle section of the Sarpo Laggo valley, which continues to be glaciated. There they reach up to the mountain spurs (Fig 44 ; 138 No. 7) and *facetted* the entire mountain shapes in the confluence area with

Fig 67a (25.10.86/1) and b (25.10.86/3); Location: 36°27'20"N/ 76°57'45"E; Fig 138, No. 38 (cf. Tab 2).





tributary glaciers to more or less complete *tetrahedrons* (Fig 46, bottom far left - - - -). On the orographic right of the Sarpo Laggo valley exist further truncated spurs with glacial cuspate areas and flank abrasions and polishings with rock smoothings are preserved (Fig 138 No. 8). The smoothings occur on vertical *outcrop bassets* of crystalline schist (Fig 8 ▲), and are interrupted by the acute incision of the *meltwater notch* (▼) of a hanging tributary valley glacier.

West of the Muztagh valley most of the peaks over 6000 m belong to the catchment area of the Skamri glacier. According to the Spender map (1937), their highest peak "The Crown" reaches 7295 m asl (Fig 46 and 47 No. 1). The present Skamri glacier terminates at 4000-4030 m asl (Fig 9) in ice covered by surface moraine. However, in common with most of the major tens of kilometres-long valley glaciers, the glacier *lacks* a frontal moraine, and the tip of the tongue, ie the body of ice, runs out on the rapidly growing *glacier outwash plain* (Fig 48 O). In 1986 the glacier tongue had been in *retreat*: evidence of this is available in the *flat*, longitudinal profile of the tongue end (cf. Fig 47 together with 49) and on the opening of the glacier outlet which set back some decametres into the tongue (Fig 9; 47 ⊍; 48 O left). In the interval between the collection of data for the Spender map (1937) and the Chinese maps from the seventies (Fig 49) the Skamri tongue has receded by approximately 0.9 km. By 1986 the distance from the 1937 position was c. 1 km. Even those somewhat *earlier* ice margin positions (cf. above) cannot be established by any remnants from frontal moraines. They are missing because of the *far superior* morphodynamics of the meltwater discharge, as against the comparatively minor and short-term effects of the advancing and scraping glacier tongue margin. The retreat of the ice is indicated by the position of the tongue end in relation to striking slope gullies or slope gorges which the earlier maps were consequently able to include (Fig 36 ⟍⟍). The next older stage of the Skamri glacier is marked by lateral moraines on the right which taper off at the gravel floor (Fig 36 'VII-IX). It was situated at a distance of about 1.5-1.7 km from the present glacier outlet position, and was reached 2000 YBP, or 440-80 YBP respectively (middle and younger Dhaulagiri Stage 'VII-IX). The next higher lateral moraine position on the orographic right is evidence of the previously maintained final confluence with the Sarpo Laggo glacier (Fig 36 IV; 46 IV). Although the tongue belonging to the Skamri glacier had presented an only slightly higher surface level, the volume of the glacier tongue had more than doubled, thanks to the *greater valley width* of 3 km (Fig 138, No. 9) in the area of the confluence with the Sarpo Laggo valley. Currently exposed ground moraine ramps higher up the valley, which stand out in Fig 9 as a result of a light cover of freshly fallen snow, show that, at the same time, the glacier level had been *raised* (Fig 9). The upper edge of the lateral

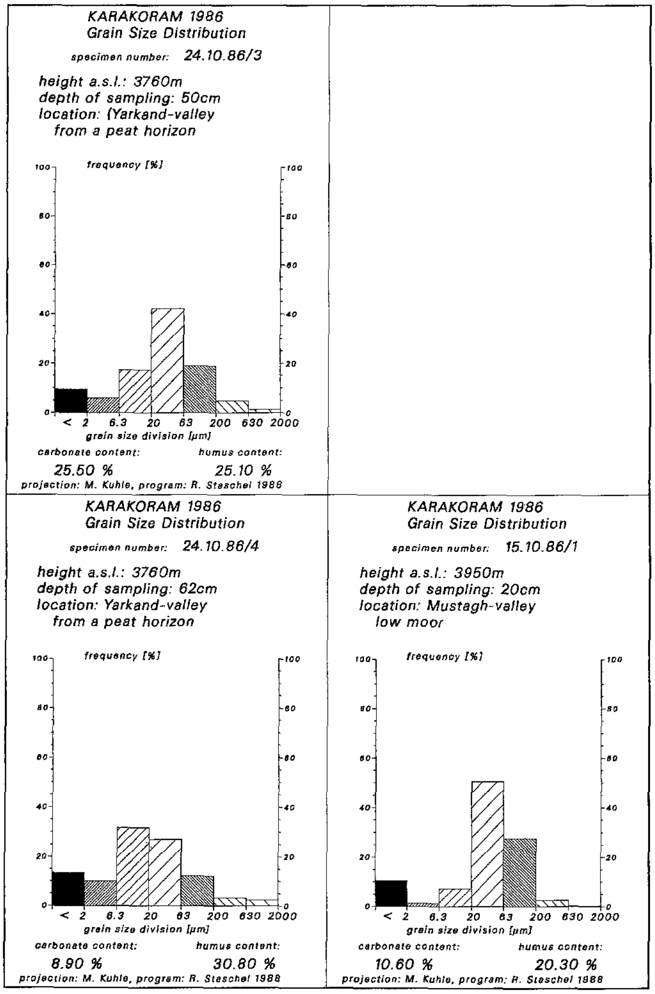


Fig 74 a (24.10.86/3) and b (24.10.86/4). Location: 36°24'N/ 76°52'E; Fig 138, No. 37; c (15.10.86/1). Location: 36°03'N/ 76°25'20"E. Fig 138, No. 10; Tab 2.

moraine of Nauri Stage V (Fig 36) lies at 4220 m asl, 190 m above the valley floor and can be classified as neoglacial (c. 4000 YBP). The level of the Sirkung Stage (IV) (Fig 36) will have been the *last* one to belong to the Muztagh glacier as a *joint* Skamri-Sarpo Laggo glacier, which *in turn* had joined the K2-N-Skyang Lungpa glacier system (cf. above). When the Muztagh valley was filled with ice, the locality of the samples (15.10.86 1-5; Tab 2; Fig 36 X; 138 No. 10¤) must have been under ice as well. This means that the peaty-clay on the ground moraine, which has been dated to 12870 YBP, is *younger* than this glacier stage, and that Stage IV consequently forms part of the Middle Late Glacial period. In the confluence area of the Muztagh and K2 glaciers large-scaled deposits of ground moraine from the Sirkung Stage IV are preserved on the right-hand slope of the Muztagh valley (Fig 43 IV). Its surface shows *exaration grooves* and *furrows,* which continue beyond the mouth of the K2 valley (Fig 43 ↓↓↓). Their interlocking with moraine material from the former K2 glacier is evidence of concurrent lateral deposition or thrust by this minor tributary glacier. The ground moraines (basal till) in the Muztagh valley lie upon a glacially *abraded* and *polished floor* on the bedrock (Fig 43 ▷), which has now been undercut by lateral erosion, following probably metre- to decametre-deep *subglacial* erosion accompanied by warming up and an arising snow line as far back as the Late Glacial period (Fig 43 ▷). Some of this erosion could have been caused by the subglacial stream of the K2 glacier at the time of the Sirkung Stage (IV). But even during the preceding Stage III (Dhampu Stage, according to the nomenclature employed by Kuhle, 1982) the surface of the Muztagh glacier in its area of confluence with the K2 glacier system must have been *below* the snow line (ELA). Evidence of this occurs in another and still higher remnant of lateral moraine with a preserved crest point (Fig 20 ■ III) and a slight trough-like deepening towards the valley slope (Fig. 50 III; Fig 138 No. 10).

As an upper *sequel* to the ground moraine covering of the lower valley floor slope at 4330 m asl, about 400 m above the valley floor, the over 100 m-long dam of lateral moraine (Fig 50 IV) has remained intact, thanks to the "denudation shadow" of a spur. At the time of the Sirkung Stage IV in the *Late Glacial* period the snow line was consequently *at most* 900 m lower than now (the present ELA being at 5300 m asl in this area). When *marking the stage* mentioned in the figures, the author has followed the concept outlined above, and decided on the *moraine classification.* Attention must, however, be drawn to some doubt concerning the classification of Stages III and IV. The *lower* lateral moraine terrace in Fig 50 may possibly have to be replaced by III in Fig 36. The question concerns relatively unimportant details. Both the moraine generations are undoubtedly older than 12870 YBP, and consequently belonging to the *middle Late Glacial* period. The c. 12 km-long chamber of the Muztagh valley from the tongues of the Skamri- and the Sarpo Laggo glacier to the junction with the Shaksgam valley is filled by *gravel* from the *glacier outlet,* which is up to 3.8 km wide and still building up (Fig 36, -6; 46, -6; 51, -6). Glaciated hanging valleys and major slope gullies continue to join from the orographic left-hand valley flank by way of *confluence steps,* with V-shaped meltwater dissections and large alluvial fan outpourings ("indirect outwash cones" after Kuhle 1983, p. 334 et seq.) on the valley floor (Fig 38 □; 51 □). In the flank on the left side of the Muztagh valley *glacigenic abrasions and polishings* are only preserved up to low levels of prehistoric glacier levels, and not very clearly either (Fig 46 ○○). The flank was last reached by glacier ice during Stage IV (cf. above). On the other hand, attention must be drawn to the almost total *absence* of glacially-smoothed flanks, or their *far-reaching disintegration* by frost weathering and gully formation (Fig 38 ▽), and that in extreme contrast to - at some points - very well preserved forms of glacier abrasion and polishing at significantly greater altitudes above the valley floor. Fig 52 shows a *glacial horn,* in the outcropping limestone, which is sharpened up to its peak at

4730 m. It is situated in the confluence area of the Muztagh and Shaksgam valleys (Fig 138 No. 13). Such forms of much more substantial glacial infilling are part of the early Late Glacial to Main Glacial plethora of forms, and will be considered in Chapter 5. During Stages IV or III the Muztagh glacier just reached the Shaksgam valley. Fed from the catchment areas of the Skamri-, Sarpo Laggo- and K2 glacier, the tongue of the Muztagh glacier flowed over and abraded and polished the limestone bar hill at the centre of the confluence into a *glaciated knob* (Fig 52 a; 138 No. 11; 36°08'N/76°22'E). Covered in parts by a thin sprinkling of moraine and gravel, the *polished* bar was partly sedimented over from its base by material from more recent gravel-fields. There are *no* remnants from end moraines or frontal moraines. This is a *typical* feature of ice margin positions of major valley glacier systems (cf. Kuhle 1991 b, pp. 75/76). The glaciated knob lies at 3800 m asl, a *snow line depression* of merely 200 m would enable the Muztagh glacier tongue to reach, and flow over it (as the present glacier ends of this catchment area lie between 4000 m and 4200 m asl). The question of a *Shaksgam glacier tongue* extending into the area of the Muztagh valley during Stage IV and/or III, or of the continuing existence of a Shaksgam-Muztagh glacier confluence, cannot be affirmed. A *snow line depression* of 400-600 m, or more, however, makes it likely, if not necessary, in the cases of Stages IV and III. The present glaciers in the catchment area of the Shaksgam valley, which reach down furthest, including the tongues of the S-Skyang Lungpa glacier, the N-Gasherbrum glacier, Urdok glacier, Staghar glacier, Singhi glacier and Kyagar glacier, which were sending a *joint Shaksgam glacier tongue* downwards into the Muztagh valley at that time, terminate between c. 4200 m and 4300 m asl, ie only 400–500 m above the said *confluence* with the Muztagh valley of the Late Glacial Muztagh glacier.

4.3 Exemplary Observations Concerning Historical, Holocene (Neo-Glacial) to Late Late Glacial Ice Margin Positions and Glacier Traces in the Catchment Area of the Kudi Valley on the Kuenlun North Side

Fig 21 shows the upper valley chamber of the "NW-valley of the 6328 m massif" (36°32'N/77°17'E) closest to the valley head, and its catchment area with the present 3.4 km-long glacier (O). This glacier possesses a *steep* tongue-front (O) and after years of retreat it has been *advancing* again since 1986. From the present and lowest glacier tongue end and there at 4800 m asl down to 4450 m asl (aneroid measurement) at least four more recent ice margin positions can be identified (Fig 21, X, IX, VI, V, IV; 138 No. 14–15). This sequence of end moraines and their *fresh* state of preservation permit conclusions as to a chronological sequence during the past c. 4200 YBP since the Nauri Stage (V), the second oldest one here. It is the *Holocene* moraine sequence since the *neo-Glacial* period. The highest mountain of the catchment area (Fig 21 No. 1) is probably (according to the Chinese map 1:50 000) the 6328 m peak. During historical times a tributary glacier coming from the orographic right still achieved a *confluence* with the main glacier. Today, it is still preserved as a typical transition from a "rock glacier tongue" to a *dumped terminal moraine* (Fig 21 X). Covered with a metre-thick layer of block debris, its tongue ends at 4700 asl (aneroid measurement), 2 km down valley from the uncovered glacier tongue of the main glacier in the valley axis. Inter-locking moraines of the two glaciers in the confluence area are evidence of such a confluence (Fig 21 below O). Small hanging valleys join on the orographic right; their valley heads contain remnants of dumped terminal moraines from historical to neo-Glacial corrie (cirque) glaciers, which are the result of a 100-200 m snow line depression. The present snow line in this valley runs at about 5200 m asl. All the rock glacier-like phenomena in the talweg area of the tributary valleys and corries may be traced back to *thawed-out white glaciers* and the surface moraines they left behind as dumped terminal moraines (cf. v. Klebelsberg 1948, pp. 157, 192/193). One cannot rule out the possibility that a *pseudomorphotic* change of some bodies of glacial block debris into a subsequently periglacially moving block mass with a permafrost core has taken place. In any case this valley area above 4400 m asl is situated above the *permafrost line,* evidence of which is available in the several metre-thick *flows of coarse block.* The next older pre-neo-Glacial end moraines occur between 4200 and 4400 m basal height (Fig 53; 21 ■, foreground IV) where their valley floors offer secure locations away from rockfalls and floods for pastures and housing of a major mountain pasture settlement (Fig 53 IV, centre). There are two, or even three, Late Glacial positions of ice margins, which are classified as Stages IV and III (Fig 138, No. 15). The *end moraine arcs* enclose two glacier tongue basins, which are lined up on the valley floor. They consist of diamictite with a lot of coarse block material. The distance between the lowest of these end moraine dams (III) and the present glacier margin is 10.5 km (Fig 53, III). The lateral moraine ledges on the flanks of tributary valleys (■■) shown in Fig 53 indicate that at the time of the Late Ice Age glaciation in question tributary glaciers from altogether 8 tributary valleys on the orographic right reached the main glacier; today a mere two tributary glaciers do so (Fig 21 above O between No. 1 and 2). Both are linked to the 6328 m peak. The catchment areas of the Late Ice Age tributary glaciers reached maximum altidudes of 5780–6095 m asl. Below the valley flank on the orographic left two generations of end moraine have been deposited (Fig 53 V) in the exit further to the east of the two N-facing tributary valleys; (at present, both of them contain 2.2 to 2.3 km-long valley glacier tongues). The end moraines lie at 4480 and 4300 m asl, are 2.3 to 2.6 km away from the tributary valley glacier tongues, and 570–750 m below them (below 5050 m asl). For the time being, these two tributary glacier end moraines are classified as neo-Glacial (Nauri Stage V). This chronology is based on the fact that these moraines in the exit of the tributary valleys are adjusted to the Late Glacial lateral moraines of the main valley. All the neo-Glacial to Late Glacial positions of ice margin Stages V to III described here concern *ELA depressions* of 300 to 400 m. Further descriptions of details of the Late Ice Age in the valley of the "6328 m massif" are not relevant to the Main Ice Age glacier-filling of the Kuenlun relief. The important fact is that only the *late Late Glacial* glacier ends terminated within this valley, whereas the early Late Glacial period – the Taglung (II) and the Ghasa (I) stage – are not represented in this valley (no more than in the Muztagh valley on the Karakorum N-slope, cf. Chapt. 4.2). The ice from the stages of the early Late Glacial period flowed as far as the main valley, the "Vale of Kudi", and continued for *tens of kilometres* further down the valley. It has hitherto not been possible to give precise details of either ice margin positions II or I. The *confluence* of the "NW-valley of the 6328 m massif" with the "Vale of Kudi" (named after the Kudi settlement at 3100 m asl) lies at 3900 m asl (Fig 53 below - - - - in the background; 25). Fig 53 is evidence of a large glacier valley form glacial abrasion and polishing transformed into a U-shaped valley, a glacigenic trough; a *glacier valley form,* in fact, that was abraded and polished almost up to the crestline, *far beyond* the late Late Glacial end moraines at the valley bottom. In some sections evidence of abrasions and polishings occurs in the form of almost perfectly preserved truncated spurs and *glacial cuspate* areas (Fig 53). On the orographic right-hand of the valley these glacial cuspate areas in coarse crystalline Kuenlun granites (Y1/5 Geological Map 1:500 000, sheet 1) and the adjacent abraded and polished rock areas above them extend 600–800 m above the valley floor, possibly without yet having reached the maximal Main Ice Age glacier level (Fig 53 - - - -). Fig 25 (Fig 138, No. 15) also shows glacial flank smoothings in the confluence area with the "Vale of Kudi" (). There, another tributary valley with its upper 5380 to 5526 m-high catchment area joins in even more directly, ie shortly. This is a typical, *Late Glacial* glacier valley, too. At the time *after the Main Ice Age* its confluence step was dissected by *sub-glacial meltwater* (Fig 25 →). This happened when the snow line in the SW-exposition had risen significantly above 4000 m asl. The rock floor remnants of the short Ice Age hanging trough valley are preserved as smooth glaciated *knob-like* denudation terraces (). The presence of firn shields at the crest of the catchment area (Fig 25 O) are evidence of an *orographic snow line* at about 5000–5200 m (5100 m) asl. An ELA depression of at least 600 m was required here for the Late Glacial tributary glacier to reach the valley exit. The 4950 m-high Mazar pass (Fig 138 No. 16; 36°35'N/77°00'E) ought to be mentioned in the topographical context; surrounded by large-scale high hollows, it is the pass route across the main ridge of the Kuenlun; together they form part of the uppermost catchment area of the "Vale of Kudi" (Fig 54). It continues to be close to the ELA, as is evident from the *perennial* snow areas and firn shields between 5000 and 5806 m (No. 1). As late as the Late Glacial period these *high hollows* (Fig 54) were completely filled with glacier ice, though the preserved forms manifest this only in parts. The reason for this are the metamorphic sedimentary rocks like the vertically outcropping phyllites Fig 54 shows (S1, T1+2, T3, P1; Geological Map 1:500 000, sheet 1). These conditions are not unlike those observed in similar rocks in the Arctic, for instance in Spitsbergen at 200–240 m asl, at 78°–80°N. In spite of the presence of a superimposed ice layer, which was 1500–2000 m-thick during the last Main Glacial period (Grosswald 1983, Fig 25 p. 108 and Fig 42, p. 159) extensive *frost debris slopes* (Fig 54 ∇) are now emerging there as here. They developed on frost (drifts) compensation slopes below frost cliffs, creating the shallow valley forms and hollow profiles, which could also be explained from the purely periglacial point of view. Data could be gathered from the N-slope of the 5806 m massif, including its up to 3 km-long hanging glaciers which is representative of the Late Glacial moraine deposits in the area of the "Vale of Kudi". Fig 55 No. 6 shows the *grain-size graph of an end moraine* which has been extracted from the "main N-valley of the 5086 m massif" and deposited in the confluence area in the "Vale of Kudi" at 3740 m asl (Fig 138 No. 17). Fig 56 (20.8.86/1) shows the *morphometry* of the sand fraction within the moraine sample. It contains a 55% proportion of fluvially-polished grains and 20% proportion of re-deposited aeolian material, the remaining 26% being a substratum pounded and fractured by the glacier, which shows crescent-shaped, shell-like chattermarks on the grain surfaces and must have been taken up close to the bedrock. Such an *assemblage of materials* is typical for the *Late Glacial* moraines from the catchment areas in the *Kuenlun granite* (Y1/5, Geological Map 1:500 000, sheet 1) which mean *transport distances* of 10–30 km. The maximum transport distance of sample 20.8.86/1 is 10–15 km. The predominance of 93% silt, 52% of which is solely coarse silt, is due to the coarse crystalline parent material, and is thus typical for end moraines in granite areas. The proportion of clay and sand in the matrix between the coarse boulders is well balanced. Such "tributary valley exit moraines" frequently contain *mudflow sediments,* which in turn bear almost all the *moraine features.* This is quite likely in so far as younger, valley-upward glacigenic diamictites (moraines) are regularly *taken up* by mudflows (for instance, when glacier lakes burst their banks), needing nothing but an *additional* mud-flow transport. Fig 55 No. 9 (20.8.86/1 a) shows such a mudflow deposit in the immediate environs of Fig 55 No. 6 (20.8.86/1). The cumulative graph of grain sizes within its 60% fine material proportion is *identical* with the one from the moraine material of the directly comparable catchment area. Its only *minor deviation* occurs within the 40% proportion of coarse silt and sand: the coarse grain in the *mudflow* sediment is only c. 2% greater than that of the *moraine,* a fact that is explained by the water in the mudflow, which is involved in the *erosion process.* For the sake of further comparisons the grain size composition of sample 20.8.86/2 (Fig 57/d) was also determined. It presents a purely fluvial gravel sediment from the floor of the main valley in the area of the episodic flood sedimentation located 2 m above the receiving stream (cf.

LEGENDE/ REFERENCE:

15 *Lokalität*/locality

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

Sedimentproben, C^{14} Analyse)/sampling, C^{14} 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 Unterschneichung/ 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)

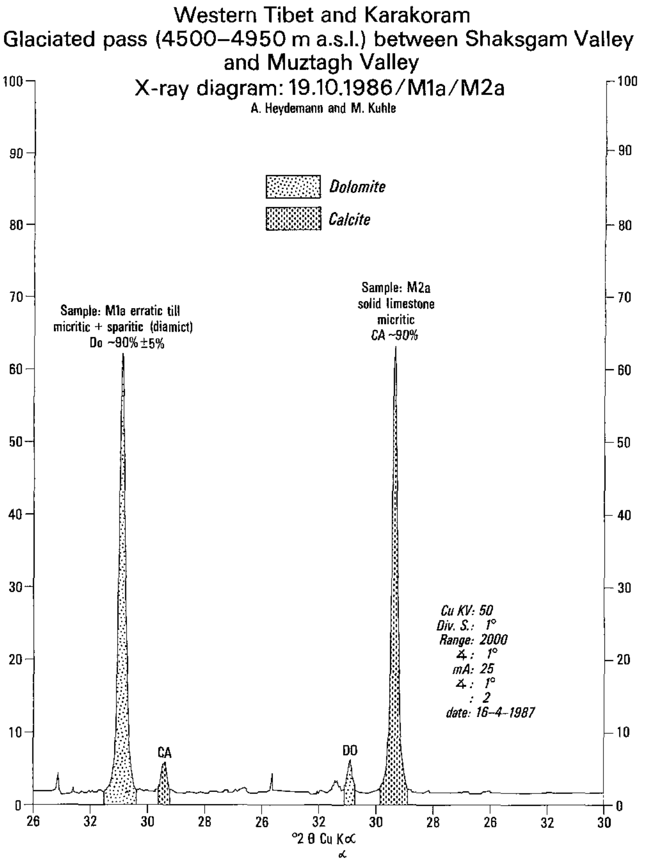


Fig 117 The erratic dolomite blocks and the dolomite scree (left peak) are mixed with up to 1.5 m-long mica-granite blocks (Fig 118) from the upper catchment area of the Shaksgam valley. The erratic dolomites lie on top of bedrock calcites (right-hand peak). The calcite rocks have glacigenically polished 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 glaciofluvial 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

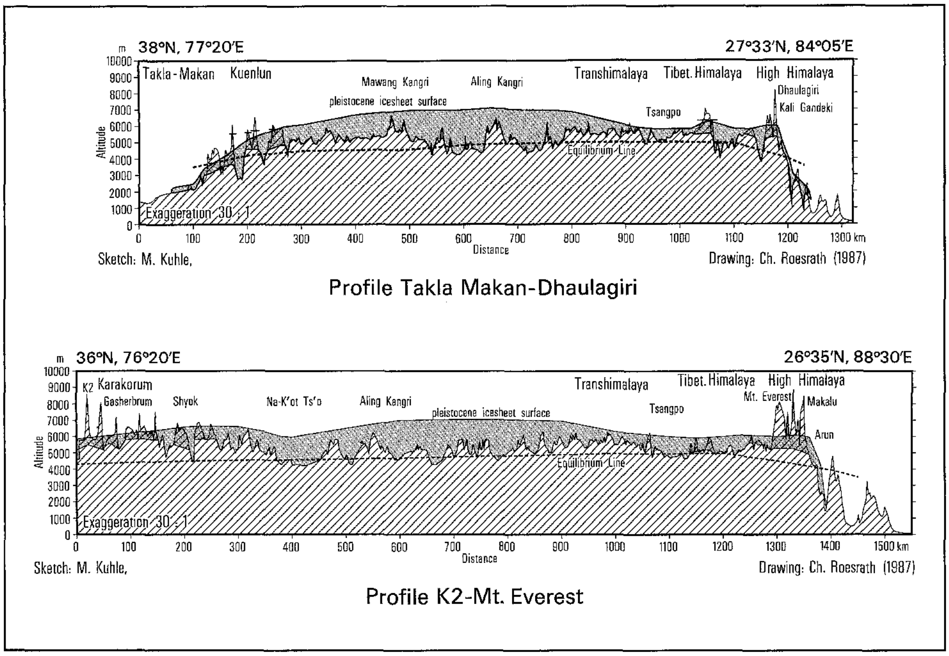


Fig 135 a/b
The two cross-sections give an idea of the Main Ice Age (Würm age) glacier cover of Tibet between a) the High Himalaya in the S and the Kuenlun in the N, down to the Takla Makan desert in the Tarim basin, and b) the High Himalaya and Karakorum, which is also in the north. This establishes 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

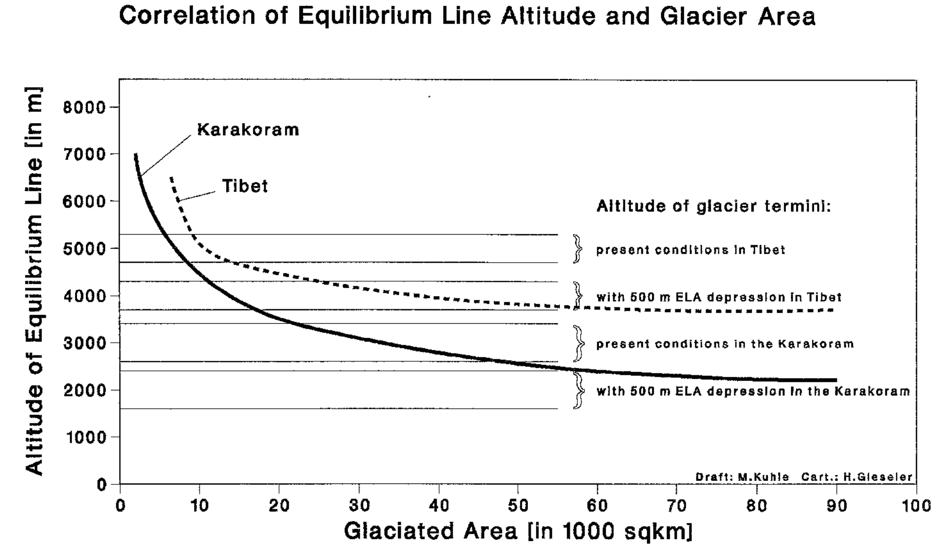


Fig 136
The diagram shows the increase in glacier areas in the particular area under investigation (in the Karakorum) and in the ice areas of the interior of High Tibet, which had been in contact with the prehistoric Karakorum ice stream network in the east. For this purpose a depression of the snow 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

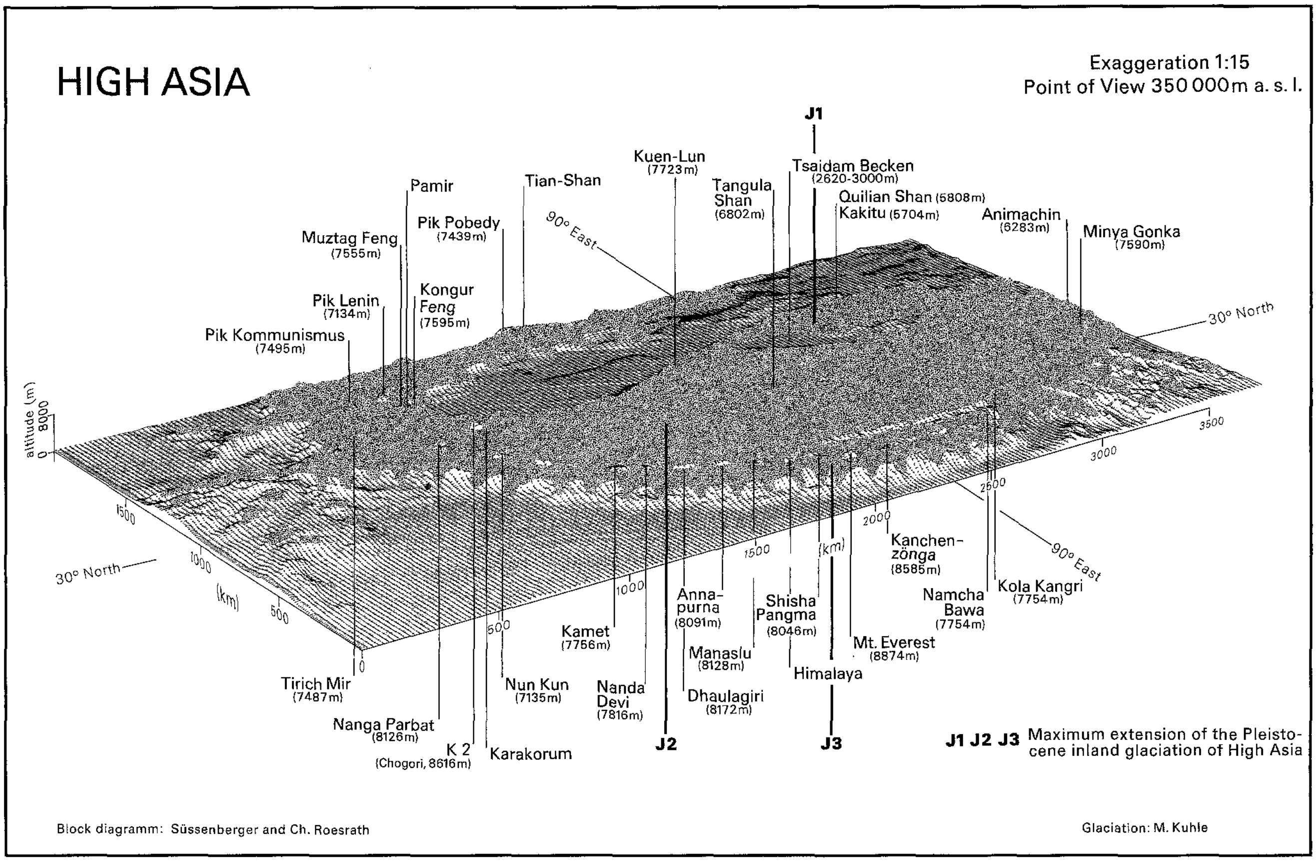


Fig 137 Towards the W the Würm period Tibetan ice merged with the ice stream network of the Karakorum (between Nun Kun and Tirich Mir), and formed a glacier system between Nanga Parbat and Kuenlun, which formed a bridge to the Pamir and even to the Tien Shan, in the NW.

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) =

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

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

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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Würm Glaciation of Lake Issyk-Kul Area, Tian Shan Mts.: A Case Study in Glacial History of Central Asia

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ABSTRACT: Recent field research and modeling experiments by the authors suggest that Würm glaciation of Tian Shan Mountains had much larger extent than it was previously believed. Our reconstruction is based upon the following evidence: 1. a till blanket with buried glacier ice occurring on mountain plateaus at altitudes of 3700 to 4000 m asl; 2. trough valleys with U-shaped profiles breaching the border ridges and thus attesting to former outlet glaciers spreading outwards from the plateaus; 3. morphologically young moraines and ice-marginal ramps which mark termini of the outlet glaciers at 1600–1700 m asl (near Lake Issyk-Kul shores) and farther down to 1200 m asl (in Chu River valley); 4. clear evidence of impounding the Chu River by former glaciers and turning Lake Issyk-Kul into an ice-dammed and iceberg-infested basin; 5. radiocarbon dates attesting to the Late Pleistocene age of the whole set of glacial phenomena observed in the area.

Our data on past glaciation provide a solution for the so called "paleogeographical puzzle of Lake Issyk-Kul", in particular they account for the lake-level oscillations (by ice dam formations and destructions), for the origin of Boam Canyon (by impact of lake outbursts), and the deflection of Chu River from Lake Issyk-Kul (by incision of the canyon and build-up of an ice-raft delta near the lake outflow).

The Würm depression of regional snowline was found to be in the range of 1150–1400 m. While today's snowline goes above the plateaus of Tian Shan touching only the higher ridges, the Würmian snowline dropped well below plateau surfaces making their glacierization inevitable. The same change in snowline/bedrock relationship was characteristic of the interglacial-to-glacial climate switches on the Tibetan Plateau resulting in similar changes of glaciation. The whole history of central Asian glaciations seems to be recorded in the Chinese loess sequences.

A finite-element model was used to test two climate scenarios – one with a gradual and another with an abrupt change in snow-line elevation. The model predicted that an equilibrium ice cover would form in 19,000 (first scenario) or 15,000 (second scenario) years of growth. It also yielded ice thicknesses and ice-marginal positions which agreed well with the data of field observations.

Introduction

Tian Shan – one of the largest mountain systems of Eurasia. Together with the Pamirs, Hindu Kush, Karakorum, Kunlun Shan, Himalaya and Tibet it enters the giant, orographically single, Central Asian mountain mass, or super-system. The most elevated core of the mass, confined within the 2000 m contour line, makes up a continuous high terrain having an area of 3.5 million km^2. Mountain ranges of the terrain form a huge asymmetric arch with its sharply convex side turned west and opening to the east to embrace the Tarim Basin with the desert of Takla Makan on its bottom.

Lake Issyk-Kul area belongs to Northern Tian Shan. The lake itself is the second largest mountain lake of the world, its areal extent is 6236 km^2, the catchment basin

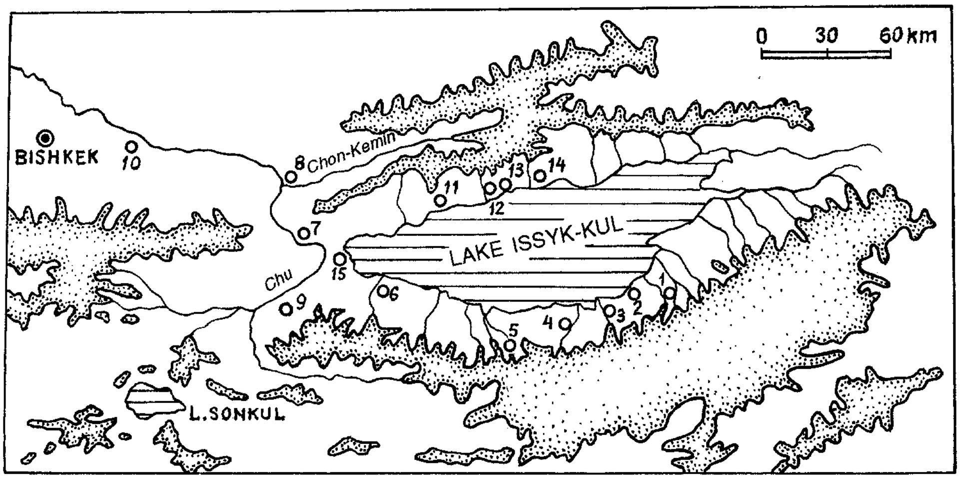


Fig 1
Würm glaciation of Lake Issyk-Kul area, Tian Shan Mts., after Bondarev and Pshenin (Atlas of Snow and Ice Resources [in press]), and the observation sites of the authors:
1 Dzhuku valley (val.)
2 Akterek settlement (stmt.)
3 Barskaun and Tamga vals.
4 Tosor val.
5 Kara-Ortok hill
6 Turasu val and Karashar stmt.
7 Kokpak-Kyrkoo val.
8 Chon-Kemin/Chu River junction
9 Kara-Kungey val.
10 Ivanovka quarry
11 Choktal val.
12 Cholpon-Ata stmt.
13 Sovkhoz "Progress"
14 Prishib hill
15 Bozbarmak hill

amounts to 22,080 km^2, water surface lies at 1607 m asl, the maximum depth reaches 668 m. Presently, the lake has no outflow; Chu River which used to flow into and out of the lake, today misses it by 11 km and enters the Boam Canyon, crossing the Kirgiz Range. Climate of the area is semiarid, the rate of precipitation increases from 115 mm/yr at the west end of the lake to 410 mm/yr at its east end, however on surrounding mountains of Terskey Alatau (up to 5216 m) (Fig 9, 16, 17, 22, 24, 26, 27, 29) and Kungey Alatau (up to 4770 m) (Fig 30, 33, 34, 38, 40) it reaches 800-900 mm/yr. An area of 650.4 km^2 is ice covered in the mountains (Fig 17), with 48 km^3 of water stored in the glaciers (Sevastianov 1991).

The mountains, being a part of the Caledonian orogenic belt of Northern Tian Shan, are built up largely of early Paleozoic rocks. Their structure and relief were rejuvenated during the Alpine orogeny. In the latter's course, a several kilometers of clastic orogenic sediments, the molasse, accumulated on the lake bottom. The beginning of the molasse accumulation, dated back to the Upper Neogene, marked the inception of the lake depression. A number of sedimentary units are recognized in the molasse section, of which the coarsest ones are traditionally attributed to the effects of tectonic activizations. Judging by studies of the southern coast sections, the Oligocene-Neogene beds of the molasses are 4000 m thick (Pomazkov 1972). Vigorous crustal faulting accompanied the tectonic activizations; this faulting is believed to be responsible for the Boam Canyon formation and deflection of Chu River from the lake.

Equilibrium line of present-day glaciers and ice caps of Northern Tian Shan lies within the altitude ranges of 3700–3800 m in the north and 3900–4200 m in the south (Krenke 1982). In Lake Issyk-Kul area the range is 3700 to 4100 m asl; the glacier tongues reach down to 3500–3900 m, while some of north-facing ones descend to 3000–3200 m (Sevastianov 1991). Widespread traces of former glaciations were also reported by many travellers, including such renouned naturalists as Severtsov, Mushketov, Davis, Berg, Prinz, Kalesnik and Gerasimov. Nevertheless, a number of paleoglaciological problems of Tian Shan remained unsolved. Practically unkown were the size and types of former glaciers, the range of snowline lowerings during the Würm and older coolings as well as glacial/interglacial climate changes. Possible role of glaciations in reorganizations of hydrographical systems, landform reshaping and Cenozoic lithogenesis was never considered and discussed. However strange it may seem, and despite the scarcity of factual data, a belief in small-scale glaciation became deeprooted, having been based on mere speculations about the probable consequences of climate aridity in Central Asia. This especially applied to the last glaciation of Tian Shan, which was believed to have been the smallest. Some relatively new data on altitudinal position of end moraines, which appeared to have been formed during the last glacial maximum (LGM) (Alyoshinskaya et al. 1983), along with "old" radiocarbon dates obtained from shores of such high-plateau lakes as Sonkul and Chatyrkul (Sevastianov 1977, 1991), seemed to corroborate this concept. As far as pre-Würm glaciations are concerned, they were envisioned as having larger extent, in fact, the larger the older their age was (Fedorovich 1960; Kachaganov 1979; Prinz 1929; Selivanov 1990).

Würmian depression of snow line was estimated as moderate to small, and geographically differentiated. In particular, it was pointed out by Kalesnik and Epstein (1936) and Kalesnik (1936) that on Ak-Shiyrak Range and in upper reaches of Naryn River the depression amounted to 500 m or, according to Prinz (1927) - to 450–500 m. It was further assumed that the range of snowline lowering increased in westerly direction to 600 m, reaching on Terskey Alatau Range and Mount Khan Tengri 700 to 800 m (Prinz 1927). These estimates seemed to have been

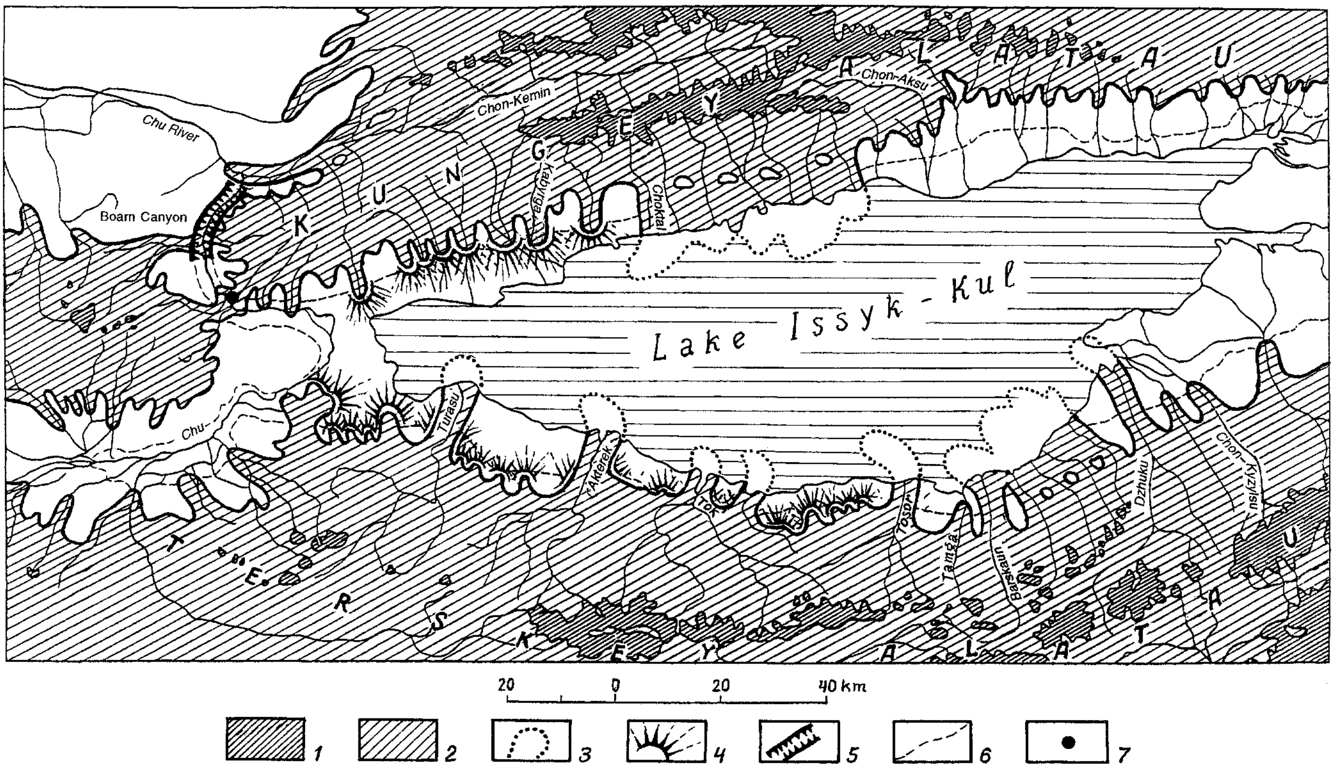


Fig 2 Würm glaciation of Lake Issyk-Kul area. Based on the authors' observations. Also published in: Grosswald et al. (1992). 1 present-day glaciers; 2 Würm ice covers; 3 floating ice lobes (tentative reconstruction); 4 ice marginal ramps; 5 erosional breach - Boam Canyon; 6 tentative limit of Issyk-Kul transgressions; 7 site of sampling for radiocarbon dating - Kokpak-Kyrkoo mouth

confirmed by recent research: the University of Moscow team also inferred that the Würmian snowline depression in Lake Issyk-Kul basin had not exceeded 600 m (Markov 1971). The smallest range of that depression, 200 to 350 m, was inferred from the altitudes of empty cirques on Ak-Shiyrak Range (Bondarev 1965, 1982), and the largest - 1100 to 1200 m - in Kirgiz Alatau Range (Maksimov 1980).

The opposite views had their advocates, also. Over a century ago Severtsov (1877), who spotted "bona fidi" moraines in Chu-River basin at altitudes of about 1500 m asl, inferred that the former glaciations of Tian Shan were quite extensive. This view was shared by Kassin (1915) who, after inspection of boulder till and outwash masses overlying the Issyk-Kul terraces, speculated on "... the former Malaspina-style glaciers that completely inundated all the foothills of the Terskey Alatau and Kungey Alatau Ranges".

Some geologists believe that all, or nearly all, glacial landforms, which are in evidence on Tian Shan, represent only Würm glaciation. According to that concept, all the glaciers formed during different cold epochs of the Pleistocene, were of about the same size, so that the latest of them were to destroy or mask evidence of the earlier glaciations. While the systems of heterochronous moraines known from quite a number of mountain valleys were all considered to have been stadial formations marking ice-terminal oscillations dated from the last deglaciation, not the traces of several independent glaciations. This view was shared by Maksimov (1983, 1985), Serebryannyi and Orlov (1988). It also tallies with the results of our studies.

Lake Issyk-Kul and its "Paleogeographical Puzzle"

Issyk-Kul area of Tian Shan comprises the basin of Lake Issyk-Kul, adjacent Terskey Alatau and Kungey Alatau Ranges and mountain plateaus in upper reaches of Chu and Naryn Rivers, as well as several internal ranges, including Ak-Shiyrak, towering above the plateaus. This area is of the key relevance for solving paleogeographical problems of Tian Shan as a whole. In addition, there are unsolved problems of that particular area itself.

Among the problems are the above mentioned lack of outflow from the lake, clear evidence for its recent connection with Chu River, and the traces of considerable oscillations in lake level, in particular its rises to the altitudes which strongly surpassed the levels of rocky sills

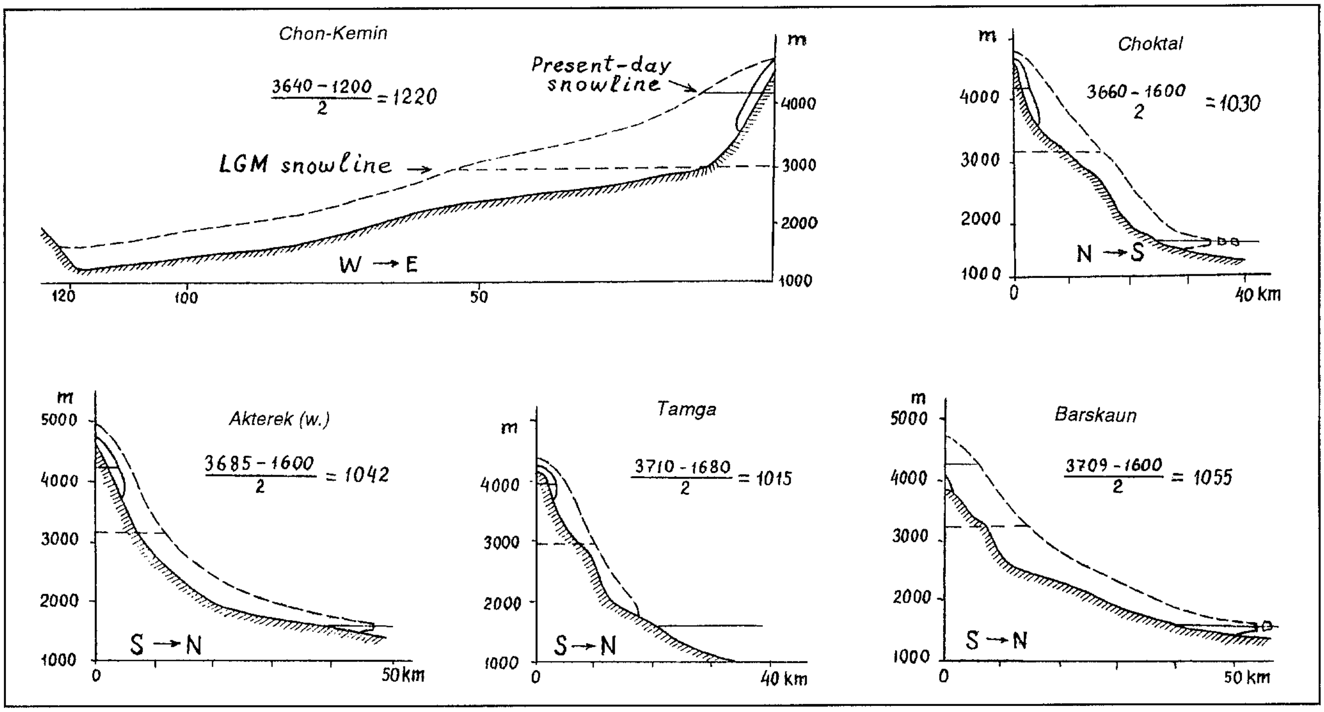


Fig 3 Lowering of snowline during the LGM in the several representative valleys of Lake Issyk-Kul area: Chon-Kemin, Choktal, Akterek (west), Tamga and Barskaun. Determined by the method of Höfer (1879)

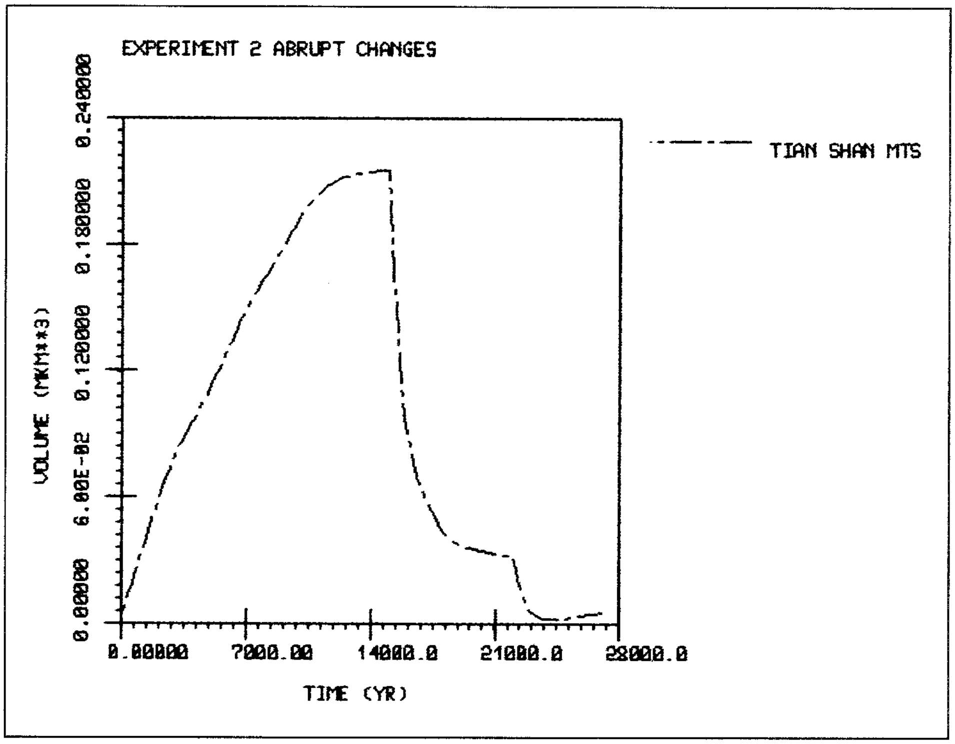
within Chu River valley. A giant Boam Canyon formed where the river breaks through the Kirgiz Range, is an example of another problem, as its origin is under debates for a century. It was this specific bundle of unsolved issues, interconnected with each other into a single complex problem, was ment by Semionov Tian-Shansky and Berg, and later by Gerasimov, Bondarev and Maskimov, when they discussed the "paleogeographical puzzle of Lake Issyk-Kul" (Berg 1904; Bondarev 1958; Gerasimov 1953, Maksimov 1985).

Hence, the unsolved puzzle of Lake Issyk-Kul comprised the following pieces. What caused the past lake-level oscillations resulting in formation of high terraces or, in other words, what kind of natural agency could repeatedly block the lake outflow? Why present-day Chu River does not flow into Lake Issyk-Kul, what forced it to turn away into mountains? And what sort of forces could create the Boam Canyon, what was the specific mechanism of the canyon's formation? Until recently, there was no theory which could account for the whole bundle; at best, only some *ad hoc* explanations were advanced for some of the problems in question.

One of the approaches stemming from the views of Semionov Tian-Shansky and Berg, tended to treat all the above phenomena in terms of gradual geomorphological evolution of the Neogene intermontane basin which predated the present-day Issyk-Kul and its environ. The ancient lake basin was believed to have been much larger than the present-day Lake Issyk-Kul, and the ancient lake level - much higher than the presently existing level. Erosional incision by an overflow stream that, from the outset, discharged water across the lowest saddle in the Kirgiz Range was taken for the chief mechanism both for incision of Boam Canyon and the lake level changes. Consequently, Boam Canyon was considered to be a conventional trough valley produced by erosion of "normal", or relatively uniform, equable stream flow, while all the changes experienced by the lake level were reduced to one-way lowering. Another version of this view, shared by Kvasov and Seliverstov (1960), suggested that Boam Canyon is not a Neogene-age, but a younger, late Quaternary, feature.

Another view originated from Mushketov and Fedorovich. As we already pointed out, it suggested that the leading role in the canyon formation had been played by tectonic, in particular seismotectonic, faulting. Gerasimov, Shnitnikov and Maksimov were among the partisans of this concept. Namely, they all believed that the northern deflection of Chu River and its divorce from Lake Issyk-Kul had resulted from a tectonic rift which subsequently developed into the Boam Canyon. They further maintained that, in addition to tectonism, a natural dam created by Upper Chu's deltaic accumulation in western part of Lake Issyk-Kul also contributed to the

Fig 4
Volumetric growth and shrinkage of the Northern Tian Shan ice cover with time (modeling experiment 2: abrupt lowering and rise of snowline by 1200 m)



deflection (Gerasimov 1953; Maksimov 1985; Sevastianov Shnitnikov 1980; Sevastianov 1991).

Mushketov-Fedorovich concept is still quite popular and has many supporters (Sevastianov 1991). The same can be said about the older concept of Semionov Tian-Shansky and Berg. Whatever their differences, they both have one major feature in common: neither gives credit for reorganizations of the "Lake Issyk-Kul-Chu River" system to former glaciations. By contrast, our concept suggests that those reorganizations were virtually caused by glaciers. It has been already several years ago that we, basing on geomorphological evidence from Lake Issyk-Kul area, came to realize that "late-Pleistocene snow-line depression in northern Tian Shan Mountains had amounted to 1100–1200 m which forced the glaciers of Terskey Alatau and Kungey Alatau to advance down to Lake Issyk-Kul and to fill and block Boam Canyon thus turning the lake into an ice-dammed basin" (Grosswald 1989, p. 42). A little later, the same idea was put forth by Selivanov (1990) who got involved in discussion of the past Lake Issyk-Kul level oscillations. Selivanov concluded that the oscillations had resulted from past streamflow dammings caused by repeated fillings of Boam Canyon with ice and till. Unfortunately, for some unclear reasons he related those damming events not to the Late, but to Early Quaternary, having for this neither stratigraphic nor geomorphological grounds. It is noteworthy (but not necessary correct) that, according to or Kvasov and Seliverstov (1960) and to Maksimov (1980), no canyon existed in the Kirgiz Range during the Early Quaternary. As well as, possibly, had not existed Lake Issyk-Kul itself (Gerasimov 1953).

How come that the causal links between reorganizations within the Lake Issyk-Kul-Chu River system, on the one hand, and former glaciers, on the other, which seem so obvious, were overlooked for so long? An answer has been provided by Fig 1. It shows a reconstruction of Würm glaciation in Lake Issyk-Kul area, recently produced by Bondarev and Pshenin (not published) for the World Atlas of Snow and Ice Resources. In comparison with previous works, the reconstruction looks maximalistic, and still it reads that not a single Würmian glacier ever approached the lake's shore line nor reached the Boam Canyon. Thus the glaciation was thought unable to directly interfere with evolution of the system. No wonder, the virtual role of glaciation in reorganizations of the system was not assessed, not even acknowledged.

The case is, as now became obvious, that our predecessors, in their majority, simply failed to identify the clear evidence of former ice marginal positions at low altitudes, such as end moraines on Lake Issyk-Kul shores or within Boam Canyon. Hence, they just had to resort to other explanations, such as tectonic or seismotectonic hypotheses.

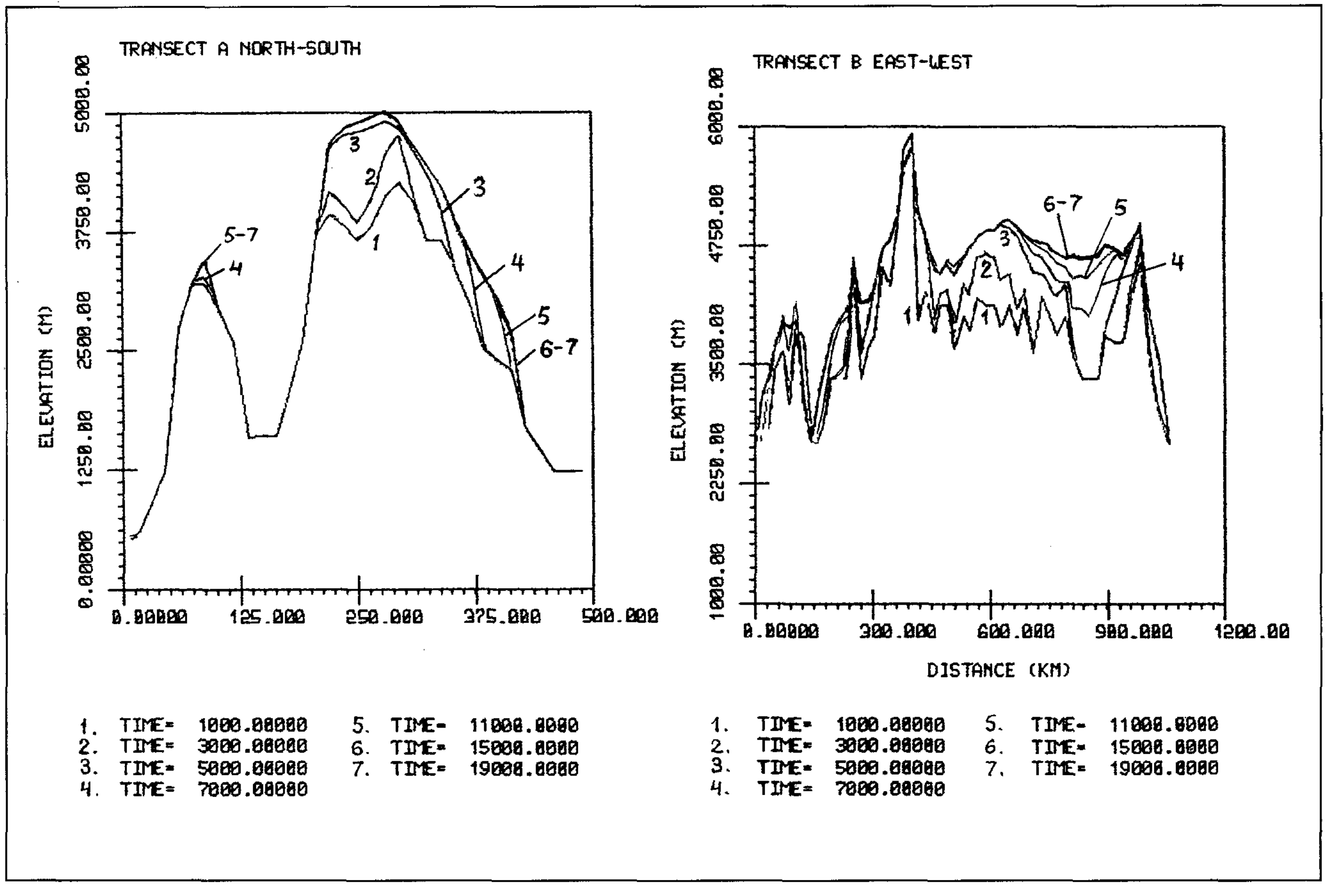


Fig 5 Ice thickness changes on the plateaus of northern Tian Shan with time (modeling experiment 1: gradual lowering of snowline – by 1200 m in 6 ka)

Evidence for Würm Glaciation

Our observations conducted in Lake Issyk-Kul area during 1986–1991 field seasons resulted in establishing the broad occurrence of morphologically fresh glacial landforms – end moraines, glaci-fluvial fans and terraces as well as a variety of ice scour features, lying at surprisingly low altitudes – near the lake's shore line at about 1600 m asl and within Chu-River valley down to 1200 m asl. It was also found that a specific kind of end moraines, namely ice marginal ramps (IMRs), or *Bortensander*, (Fig 9, 24, 27, 29, 33, 34, 36) which are peculiar to the glaciated semi-arid mountains (Kuhle 1989, 1990) broadly occur throughout foothills of Terskey Alatau and Kungey Alatau Ranges facing Lake Issyk-Kul.

Fig 1 shows the most important sites of our field observations. All the sites lie beyond the area which was formerly shown as glaciated. Nevertheless, fresh looking glacial landforms, in most instances end moraines, were identified at each one of them (Fig 9, 10, 18–41). The largest field of till deposits and glacial landforms occurs south of the lake, on high plateaus (Fig 17), or so called "syrty", and mountain ranges (Fig 13–16). Both the plateaus and adjacent mountain slopes are mantled by lodgment till containing giant erratic blocks (Fig 11) and rafts. Tabular bodies of ground ice were found within the till in several excavations. Plateau geomorphology is dominated by a number of recessional moraines and dead-ice ("kettle-hole") landform complexes.

The plateau surface has the altitudes of 3700 to 4000 m asl (Fig 17), the mountain ranges rise to 5000 m and more (Fig 17, 25, 27, 29), while mean present-day snow line in the area is at about 4100 m asl. This implies that extensive Pleistocene glaciation was there inevitable: even in case that the snow line lowering was as small as 500–600 m (which was assumed by Prinz and Kalesnik), the whole area of plateaus and ranges south of Lake Issyk-Kul had to turn into a continuous field of net snow accumulation. Which, in turn, led to formation of a big ice cover (Fig 17) with its margins descending to the mountain foothills (Fig 9, 10, 22, 24, 25–27, 29) where its net ablation was possible. As suggested by specific topography of the area, the former ice

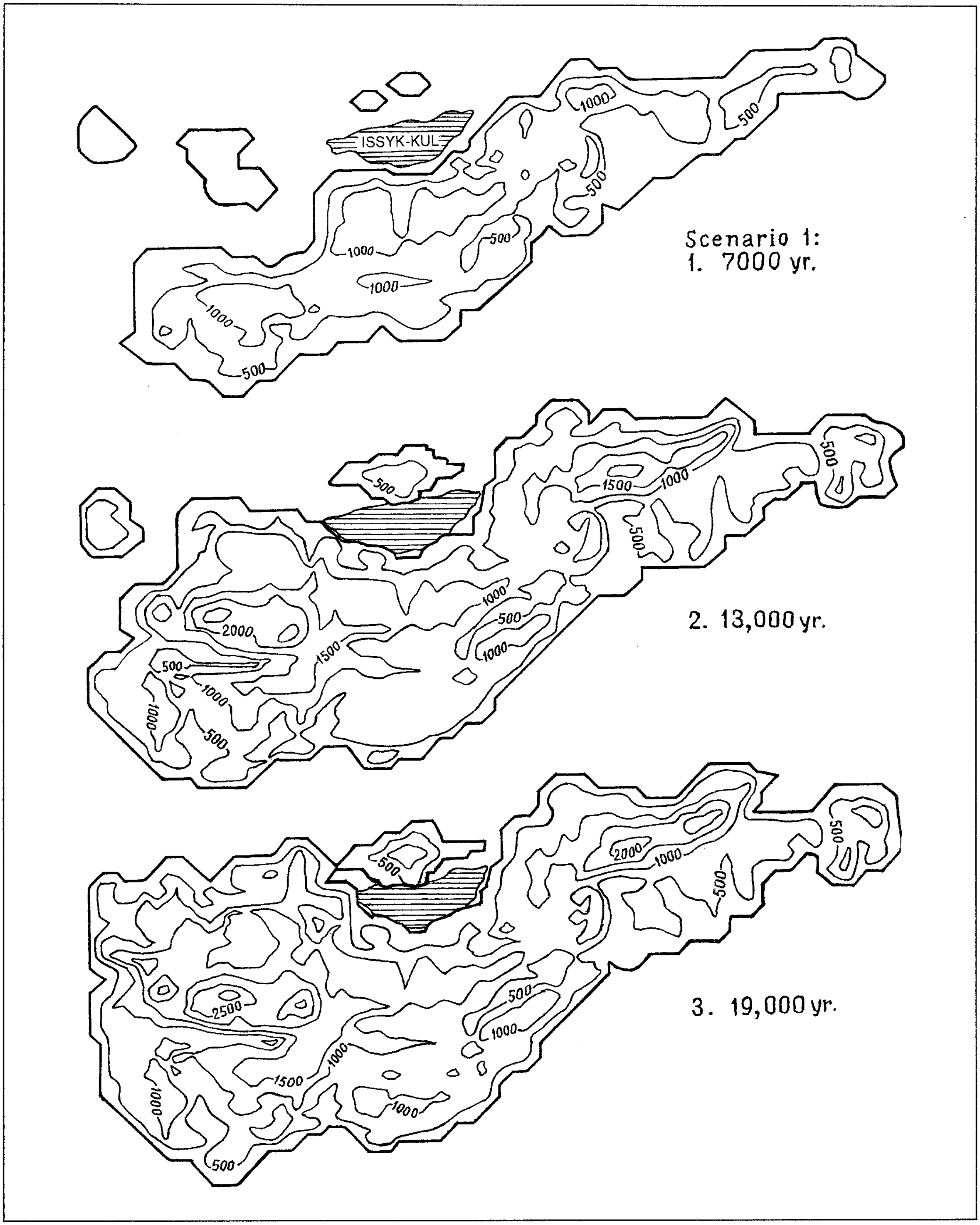


Fig 6 Changes in ground-plan configuration and thickness of the Northern Tian Shan ice cover with time (modeling experiment 1: gradual lowering of snowline – 1200 m in 6 ka). Note the rapid ice-cover formation at 7 ka (only 1 ka after full lowering of the snowline) and relatively small difference between ice configurations at 13 and 19 ka. Issyk-Kul is not filled in with ice due to the assignment of high (–5 m/yr) ablation rate within the lake's boundary.

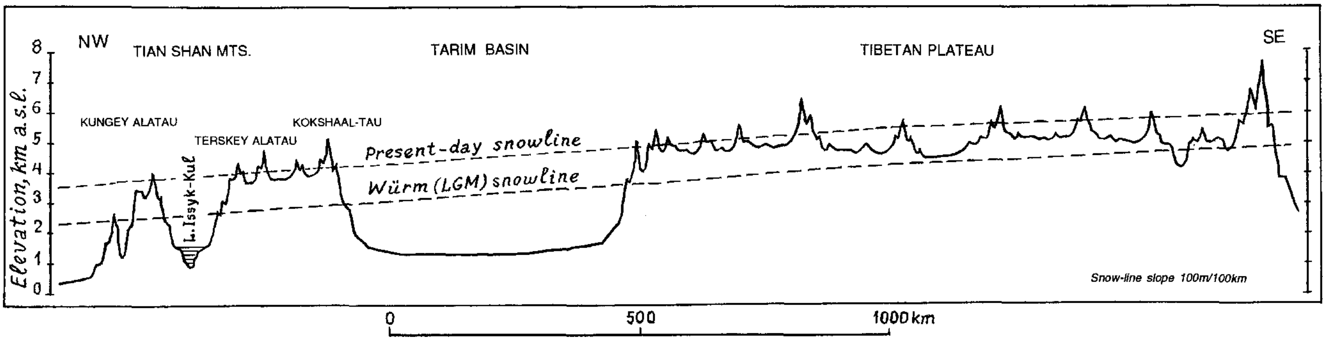


Fig 7 Generalized NW-SE profile across northern Tian Shan, Tarim Basin, Tibetan Plateau and Himalayas showing relationships between present-day and Würmian snowline, on the one hand, and the earth-surface topography, on the other.

flow directed there mainly due southwest and west into broad valleys of the Naryn River system, while minor outlets of the ice cover moved in eastern and northern directions penetrating mountain passes (Fig 17) and trough valleys (Fig 13–16).

Specifically, several north-flowing outlets, which discharged the plateau ice into Lake Issyk-Kul, breached the Terskey Alatau Range forming the trough valleys of Dzhuku and Barskaun (Fig 22). The upper and middle reaches of the valleys have a morphology of typical glacial troughs. Their long profiles look like successions of basins and reagels and the cross sections are U-shaped (Fig 13–16, 22). The surface of the plateau grades into the valley floors through broad ice-scour funnels (Fig 17), the valley mouths open up to the lake (Fig 9, 22, 24, 25); there are no terminal moraines there, only lateral moraines occur on both sides, each represented by a set of a few short, echelon-spaced ridges with the relief of about 200 m. Judging by observations in several exposures, the ridges are made up of tightly packed boulders of various granites and sandstones cemented by buff-coloured silty sands, many boulders are really big, measuring tens cubic meters; they are often faceted and grooved (Fig 26, 30, 31, 33, 34, 36, 38,41). At the trough mouths the moraines diverge and flatten out while thick accumulations of late-glacial and post-glacial gravels bury the moraine bases (Fig 9, 22, 24, 25). This geomorphology suggests that during the last glacial maximum the ice streams of the Dzhuku and Barskaun valleys advanced well into Lake Issyk-Kul, getting there afloat and producing icebergs.

Pronounced glacial geomorphology is characteristic for the rest of the Terskey Alatau valleys as well. In particular, this is true for troughs of the Tamga, Tosor, Kara-Ortok and Turasu rivers which we studied (Fig 24–27, 29). These troughs also open up to Lake Issyk-Kul and, judging by morphology of their moraines, conveyed glacier ice into the basin. In fact, only a few glaciers did not reach the lake, among them – the Tamga and Chon-Kyzylsu glaciers (Fig 9). Not only lateral (like in all the valleys) but also end moraines occur in the said troughs. In Tamga valley the moraines are in evidence down to 1900 m asl; they look like steep-slope ridges made up of big blocks and boulders cemented by sandy matrix, with a 5 to 10 m-thick sheet of milk-coloured loess-like sand overlying their crests. In ground plan, the moraines form a system of a few subparallel arches of which the proximal one is only a few miles from Lake Issyk-Kul. Longitudinal gradients of Tamga lateral moraines are much gentler than those of the valley floor, so that in some 3 km the moraines rise to the crests of inter-valley divides and merge with moraines of neighbouring valleys. The same style of moraine/bedrock relationship holds for all other valleys of the area: everywhere the lateral moraines quickly ascend to watersheds. Thus they strongly suggest that the northern slope of Terskey Alatau Range was glaciated by a continuous ice cover, not by a number of separate alpine glaciers. Only the very lake-side margin of the ice cover was divided into lobes, and higher up, in the zone of the plateau break, a chain of nunataks probably protruded through the ice cover (Fig 11, 13, 17).

Some evidence for glaciers/lake basin interaction was uncovered on the coast between settlements of Akterek and Zhenish. There, in an a partly submerged zone, extending for 15 km along the present shore line and having apparent width of 150 to 200 m, a multitude of loose giant boulders of granite and sandstone rocks occur. The boulders are glacially shaped-faceted, polished and scarred by crescentic gouges. Basing upon geomorphological setting and petrographic composition of the boulder field, we concluded that the boulder field is a residual of a large end moraine, resulting from a long lasting wash-out process. The moraine appears to have been deposited by a piedmont glacier which moved far into the basin and, at least partly, went afloat.

We counted another five valleys west of Barskaun, that used to contain Terskey's outlet glaciers that were calving into the lake. Among the valleys – the troughs of Tosor (Fig 24), Tok (Fig 25) and Turasu. No end moraines could be

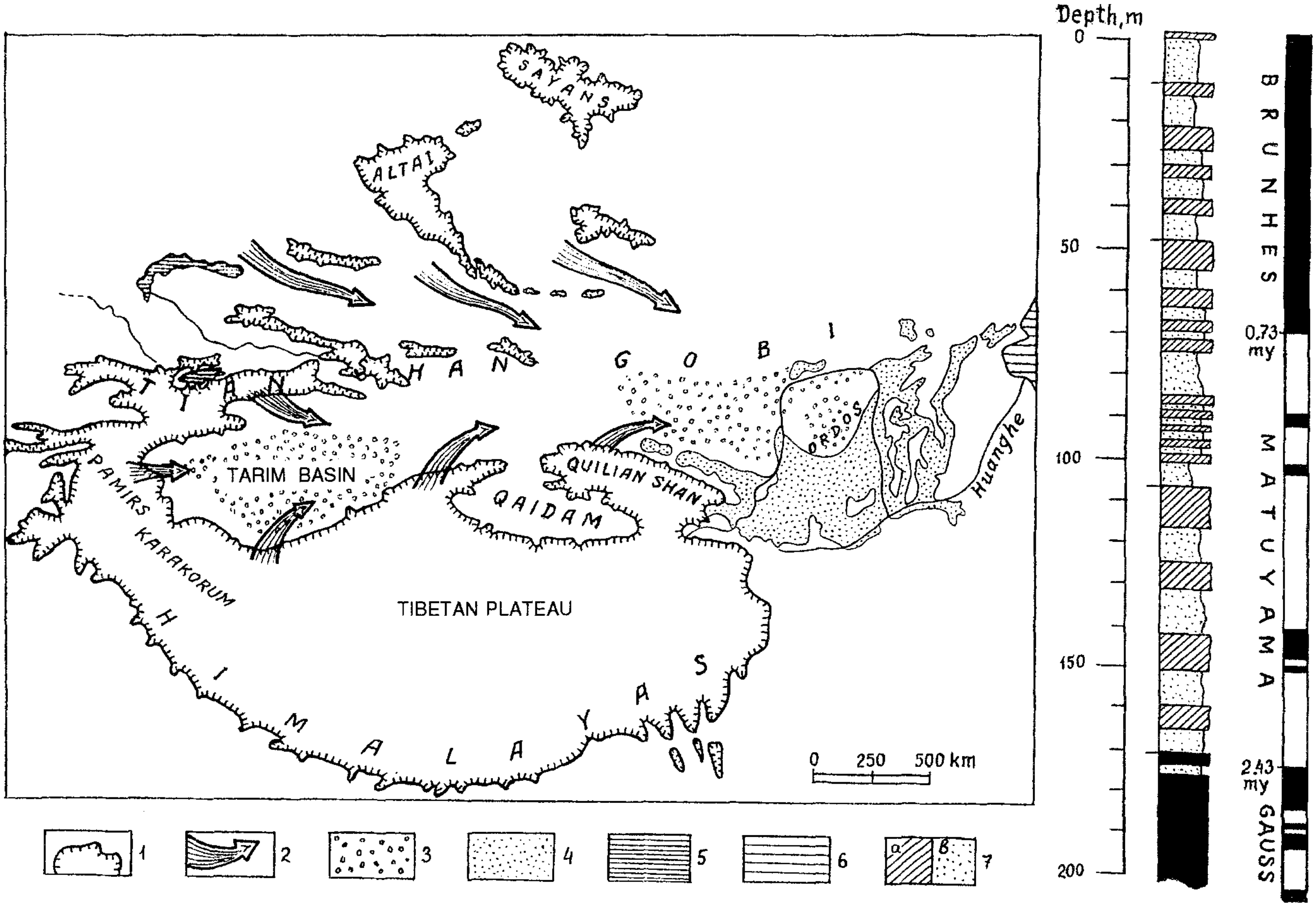


Fig 8 Ice covers, deserts and loess accumulation in Central and Eastern Asia during the last and older glaciations: the map shows directions of winds and their relationship with loess sources and major loess accumulation grounds. On the right: the Chinese loess stratigraphy after Kukla (1988).
1 ice covers; 2 directions of katabatic and general circulation winds; 3 largest deserts; 4 main field of the Chinese loesses; 5 lakes; 6 the sea; 7 major elements of the loess stratigraphy: a - paleosol beds, b - loess beds, as compared to the paleomagnetic epochs and events (extreme right)

spotted in the valleys (except Turasu), while the lateral moraines are virtually truncated on the boundary with the former - high-level - Lake Issyk-Kul. Only in the lower Turasu valley, some 100 m south of the settlement of Karashar and the circum-Issyk-Kul highway, a chain of asymmetric hills identified as remnants of ice marginal ramps were found at 1650–1680 m asl. This "Karashar Moraine" is erosionally dissected into a few asymmetric hills, aligned along an arch turned by its convex side to the north; the tops of the hills reach the altitudes of 1720-1740 m, which is about 130 m above the present-day level of Lake Issyk-Kul. The hills are made up of glacially shaped boulders and pebbles of granites, quartzites and other rocks, cemented by buff-coloured clayey sands, their southern (stoss) slopes are steep, around 30°, the northern (lee) slopes - much gentler, only 10° and less. Huge accumulations of grooved and faceted boulders mantle the western part of Terskey Alatau down to Chu River valley. The boulder masses, normally concealed by surface sediments, were open to inspection in Kara-Kungey valley due to fresh incision produced by a recent mudflow.

Wealth of evidence attests to past ice incursions into the lake from the opposite side of its basin, too. Morphologically fresh end moraines and ice marginal ramps are widespread in the southern foothills of the Kungey Alatau Range (Fig 30–38). In particular, such a moraine with apparent till thickness of 80 m occurs in lower reaches of the Choktal valley (Fig 30, 31), at the head of a piedmont outwash fan. Much larger moraines were found on eastern outskirts of the settlement of Cholpon Ata, at Sovkhoz "Progress" and Mount Prishib. Ample accumulations of large glacially faceted and striated boulders and blocks form there an impressive ridge at 1650 m asl (Fig 33–36), at the very shore line of the lake (see Fig 32 in: Kuhle 1990).

Of utmost importance for explaining the environment reorganizations in question are moraines of Chu River valley. One of a key sites was found to be a few kilometers

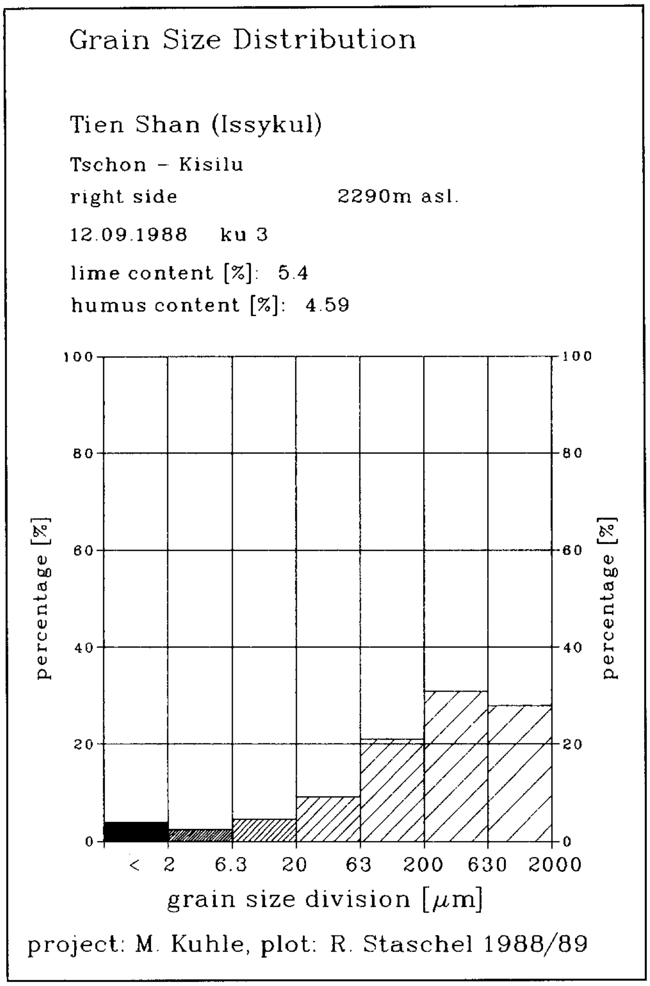


Fig 12 Analysis of the finely grained moraine substratum between the erratic granite blocks (Fig 11) on the orographic right-hand flank of the Chon Kyzylsu (Tschon Kisilsu). It occurs on bedrock phyllites as a fine matrix between these large moraine blocks. Taken from a depth of 0.20 m. The composition of grain sizes shows the characteristic proportion of grain sizes, and can be identified as moraine material on the basis of the two peaks, ie of a coarse-grained and a fine-grained peak. Bi-modal grain-size graphs of this kind are typical for ground moraines. Location of the sample taken: 42°17'N/ 78°08'28"E; Fig 1 right-hand of No. 2

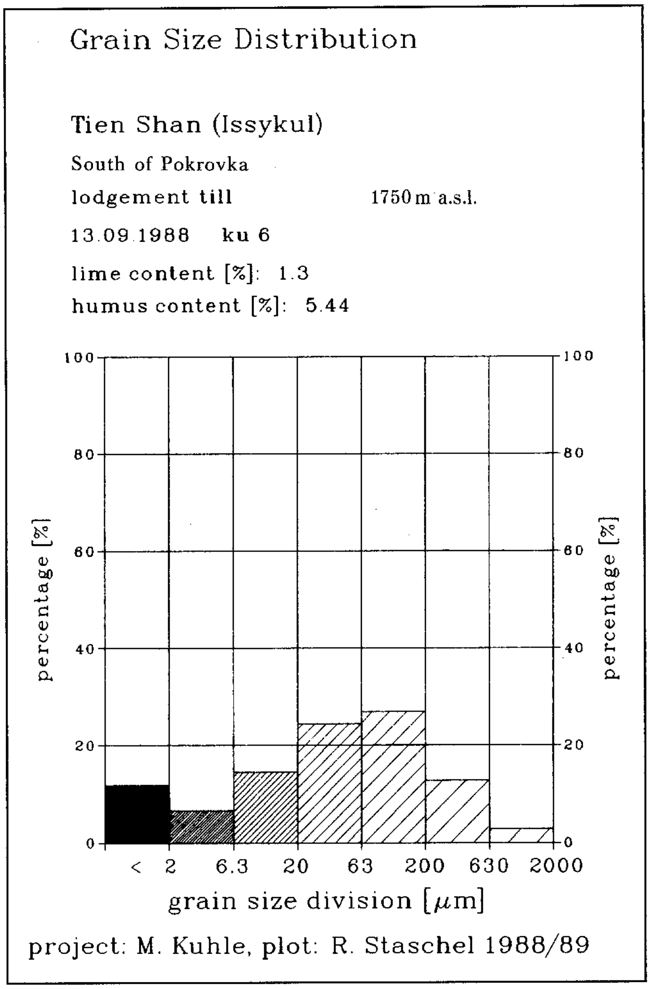


Fig 19 Locality of sample, cf. Fig 18 X. The bi-modal course of the grain-size graph, ie renewed rise to a peak of fine grain sizes, shows that this is typical ground moraine material (basal till).

upstream of Boam Canyon, at the junction of Chu River with its right-hand tributary Kokpak-Kyrkoo. The Kokpak-Kyrkoo valley was glaciated by an outlet glacier of the last ice cover of the Kungey Alatau Range, the outlet's signature being two tiers of lateral moraines on both sides of the valley. The moraines grade into lacustrine terraces, resting on the bottom of Chu River valley, the lower (60 m), and the upper (120 m) ones. In fact, the masses of till are replaced there, in a way of interfingering, by lacustrine sediments - loess-like yellow-white silts containing frequent dropstones; this all was clearly seen in bluff sections of the Chu terraces.

Another morainic lobe enters the valley on its opposite side, from the Bailamtal valley, the eastern Kirgiz Range. That lobe is dissected by a number of fanning ravines and turned into a system of short radial ridges, which also grade into the lacustrine terraces of Chu River. The moraines of both sides of the river merge into each other on the valley floor to form a single continuous field of glacial deposits extending along that floor for about 10 km - just till the upper entrance into the Boam Canyon. Obviously, the masses of ice flowing from the Kungey Alatau and Kirgiz Ranges and joining on the valley bottom had to form a dam which was high enough to cause the Lake Issyk-Kul level rise by hundreds of meters (Fig 39). And thus to account, with big margin, for all the lake-level oscillations recorded by Alyoshinskaya and Bondarev (1970), Gerasimov (1953) and others.

Several additional ice dams were formed within the Boam Canyon itself. This is suggested by river-cut masses of till entering the canyon from its hanging tributaries. Of particular importance is the moraine which was detected on the left-hand slope of the Chu valley at its junction with Chon-Kemin river, some 200–220 m above the valley bottom. The moraine was deposited by a former glacier of the Chon-Kemin valley (Fig 40, 41); this fact suggests that the glacier was by 100 km longer than today, and its snout lowered to the altitude of 1200 m asl (Grosswald et al. 1992). All the evidence for ice marginal positions during the LGM (Last Glacial Maximum) are presented in Fig 2.

To determine the age of the discussed ice-damming event, two samples of organic detritus were collected from the base of 60 m glaciolacustrine terrace in Chu River valley at the mouth of Kokpak-Kyrkoo (Site 7 in Fig 1). Their radiocarbon dating yielded the following results:

26,100± 600 yr BP (IGAN-616) and
32,390±1780 yr BP (IGAN-971),

which warrant the conclusion – basing on these dates along with unambiguous position of the sampling site relative to the till and lacustrine beds – that the ice-damming event started *later* than 26,000 yr BP, and probably climaxed during the LGM.

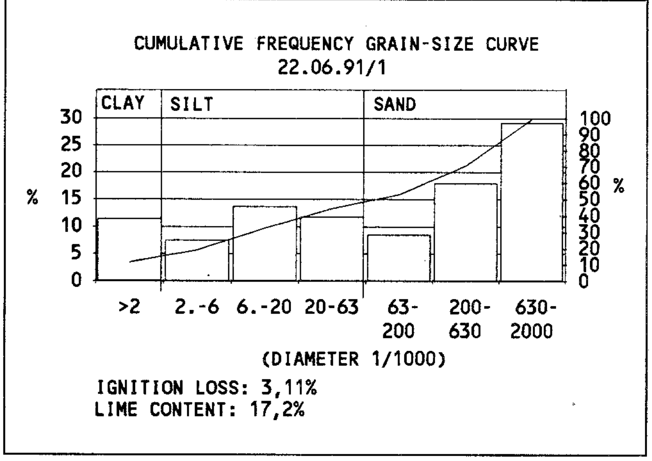
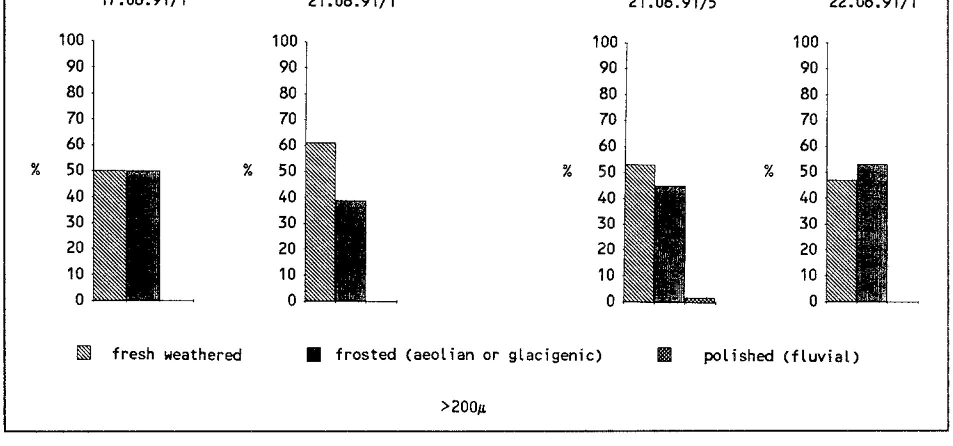


Fig 20 1650 m asl, grain size distribution in the boulder clay of a ground moraine, which reaches the south bank of Lake Issyk Kul between the ice-marginal ramps (IMRs) before disappearing below the lake level. This is matrix material from between large erratic blocks of granite which originated in the Terskey Alatau chain (Fig 1, 2, 3). As shown in Fig 2, the northern outlet glaciers of the Tian Shan inland ice cap between Dzhuku in the E and Barskaun in the W reached Lake Issyk Kul here, and calved into the lake (cf. Fig 21). Localitiy of samples taken: 42°12'N/77°43'E.

Discussion

Snowline depression and the extent of glaciation. Our map (Fig 2) provides information on the extent of last glaciation in Lake Issyk-Kul area (Grosswald et al. 1992). It implies that the glaciation of the area was twice as extensive as was previously believed (compare with Fig 1). The glaciers extended not only over the high-mountain zone, but on the mid-altitude zone, also. Obviously, this gave the glaciation a new dimension since ice discended to such low levels that it just couldn't help but invaded the Lake Issyk-Kul and Chu River valley. Which, in turn, had to cause the formation of ice-dammed lake and choking it with icebergs. This implies that the interaction between glaciation and the said elements of paleoenvironment turned out much more immediate and vigorous, than it was previously deemed possible. This further implies that not only melt water, but also glacier ice, till and outwash materials were brought into direct contact with the lake and river, producing floating ice tongues and icebergs, dumping the diamictons, impeding the surface runoff.

Fig 21
Morphometry of some fine moraine grains in the northern foreland of the Terskey Alatau chain between Bishkek in the W and Chon Dzhuku in the E; among them also the sample 22. 6. 91/1 shown in Fig 20. It is a characteristic feature of the ground moraine that glacigenically weathered material predominates over recently weathered material. The reverse obtains in end moraines (21. 6. 91/1/5). If the sample contains polished material, as in sample 21. 6. 91/5, it is a case of ice-marginal ramps (IMRs) and implies meltwater involvement in their construction as a matter of course. Sample locality 22. 6. 91/1, Fig 20.



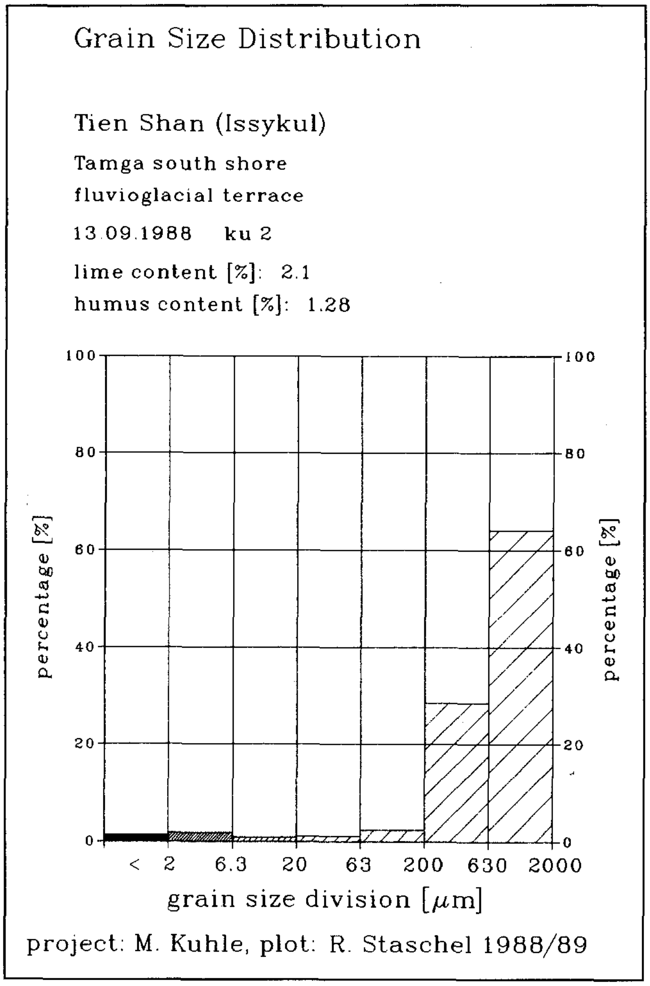


Fig 23 The grain size analysis shows the typical picture of Ice Age mountain foreland outwash plains (glacifluvial sediments) with their characteristic lack of fine material and the rapid increase in the proportion of sand to coarser material (630–2000 μm). Location of extraction: area of the mouth of the Tamga river (-valley) in the mountain foreland near the line of the south bank of Lake Issyk Kul; 1660 m asl (Fig 1, No. 3; see Fig 2); 42°09'20"N/77°33'E.

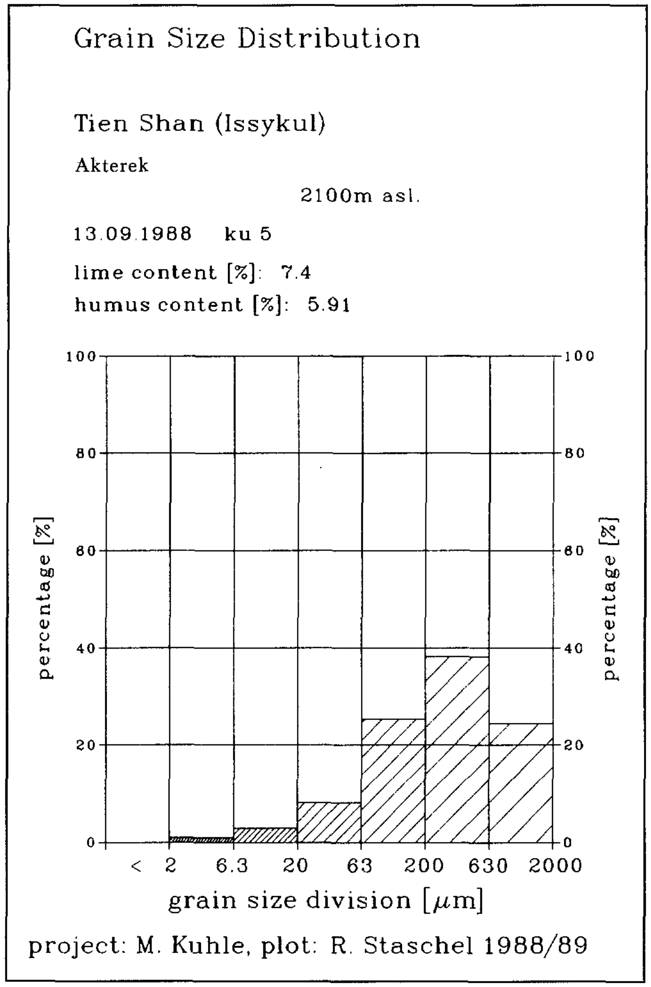


Fig 28 The composition of grain sizes is typical of the kind of outwash moraines occurring on the surface of ice marginal ramps behind the end moraine: the fine grain peak of the moraine is missing – having obviously been the first to be washed out fast. For locality of samples taken on the ice marginal ramp see locality of the photograph in Fig 27 (42°07'20"N/76°45'E) and Fig 29 with the IMR outer slope concerned.

Our data on the altitudes of past glacier snouts made it possible to determine the range of Würmian snowline depression by means of Höfer's techniques (Höfer 1879). To this end, several large glaciated valleys were selected, having "live" glaciers in their headwaters and *prima facie* end moraines of LGM in low reaches. Then, we had to determine the snout altitudes for the present-day and former glaciers of the valleys; to calculate the altitude differences; and to divide the differences by two. Specifically, this was done for the valleys of Chon-Kemin, Choktal, Akterek (west), Tamga, and Barskaun rivers. The results are presented in Fig 3, which implies that at Chon-Kemin the snowline lowering amounted to 1220 m, whereas at the rest of the valleys it was some smaller, though exceeded 1000 m, anyway. Moreover, three (of those four) glacier tongues were cut short by calving, which definitely prevented them from reaching even lower altitudes. Considerable depression of past snowline characterized the westernmost Kungey Alatau Range, too. This comes from the fact that ice-free cirques lie in the range at 2700–2800 m asl, while the equilibrium line of existing glaciers – at 3800–3900 m asl, suggests that the past, probably Würm, snowline depression measured there 1000 to 1200 m.

On the average, the Würmian lowering of snowline in the Lake Issyk-Kul area of Tian Shan was about 1150 m (Grosswald et al. 1992). It turned out to have been at least twice as large as it was accepted by our predecessors. On the other hand, our result is remarkedly close to the same parameter determined for the Tibetan plateau - 1180 m (Kuhle 1988). At it is becoming increasingly clear from mounting evidence, during the LGM snowlines were about one kilometer lower than they are today everywhere on the Earth. Their lowering shows no change between hemispheres and remarkably little change with latitude, on the wet side of the mountains and on their dry side, on the margins and in the interiors of continents. This conclusion, ranking among the major global generalizations, has been based on a wealth of research data from Europe, Americas, Africa, and the entire Pacific basin (Broecker and Denton 1989). Hence, reported here data on Central Asian LGM-snowline depression seem to be in line with corresponding glacial change in the rest of the world.

The LGM snowline lowering values, obtained in Lake Issyk-Kul area, were deducted from the present-day equilibrium line altitudes within the whole Northern Tian Shan, and the result compared with land topography, and a tentative, working map of LGM ice covers was compiled for further comparison with modeling experiment results. In the process, all available means of checking up the ice-marginal positions, such as airphotographs, information from field descriptions by Prinz, Fedorovich, Kalesnik and others, as well as by our own, were employed to verify and refine the map. After modeling, the map gave a state-of-the-art portrait of the glaciation and its growth history. In particular, of the size and thickness of specific ice covers, and of their true relationships with intermontane basins, rivers, Lake Issyk-Kul and smaller high-plateau lakes.

The southern margin of the ice cover extended to the lowest foothills of Kokshaal Tau Range, to the elevation of 1500 m asl, close to the bottom of the Tarim Basin; that elevation was assigned by Fedorovich to the Lower Quaternary ice margin. The northern margin reached the levels of 1100–1200 m in its eastern part (Kungey Alatau and Zailiysky Alatau Ranges), and 900–1000 m in its western part (Kirgiz Range). Over the plateaus of "syrty" the ice covers were continuous; the protrusions of nunataks were possible only where the alpine peaks rise, while all the plateau depressions, including the ones now occupied by the Sonkul and Chatyrkul lakes, were buried under a kilometer-thick ice.

Ice covers and Lake Issyk-Kul. The essence of that relationship, as depicted in Fig 2, can be reduced to the following facts: the mountain ice covers surrounded the basin and the outlet glaciers converged on its bottom. At least 25 outlet glaciers focused their flow into Lake Issyk-Kuk and discharged icebergs. We can speculate on the further fate of the icebergs: provided, the mean air temperature in the basin was below 0 °C (since today it is about 6 °C, and the above snowline depression translates into 7 °C to 9 °C of cooling) while the amount of heat stored in water was quickly spent on melting. Hence, there could be no ice melting in the lake, and the mass of icebergs was building up until, time permitting, a floating ice shelf formed.

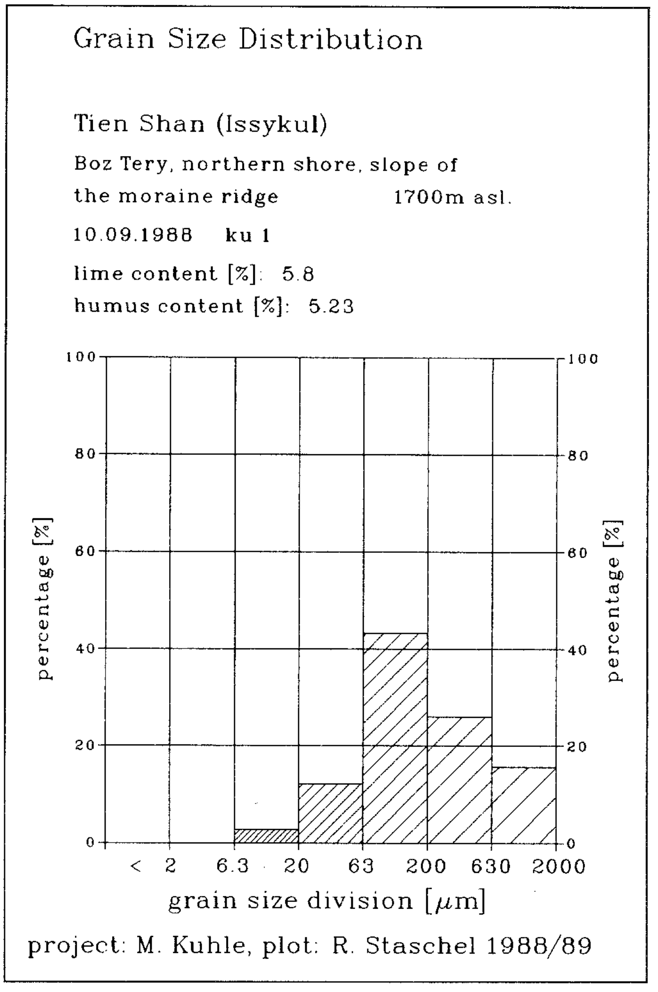


Fig 32 Taken from the surface of the end moraine ridge c. 100 m above Lake Issyk Kul, this moraine material (Fig 1, No. 12) is relatively coarse thanks to its high proportion of granitic sand. On the surface the fine material has moreover been washed out by excessive meltwater and rain-wash (however, cf. Fig 35). Locality of the sample origin: 42°40'N/77°12'E; (Fig 33).

The latter, being squizzed and dragged largely westward, in the direction of intermittent water flow, carried glacial debris to the outflow, built there an ice-raft delta pushing it into a dam. This mechanism of building the ice-raft deltas has been described from present-day glacial lakes (Gilbert and Desloges 1987). A group of hills built of deformed layers of ice-rafted sands and till lenses, called Bozbarmak, seems to be a remnant of the dam. Possibly, the deltaic sediments of Upper Chu River contributed to the dam also. In our view, the formation of that dam was one of the factors explaining the deflection of Chu River from Lake Issyk-Kul. The role of another factor was played by formation and deepening of Boam Canyon

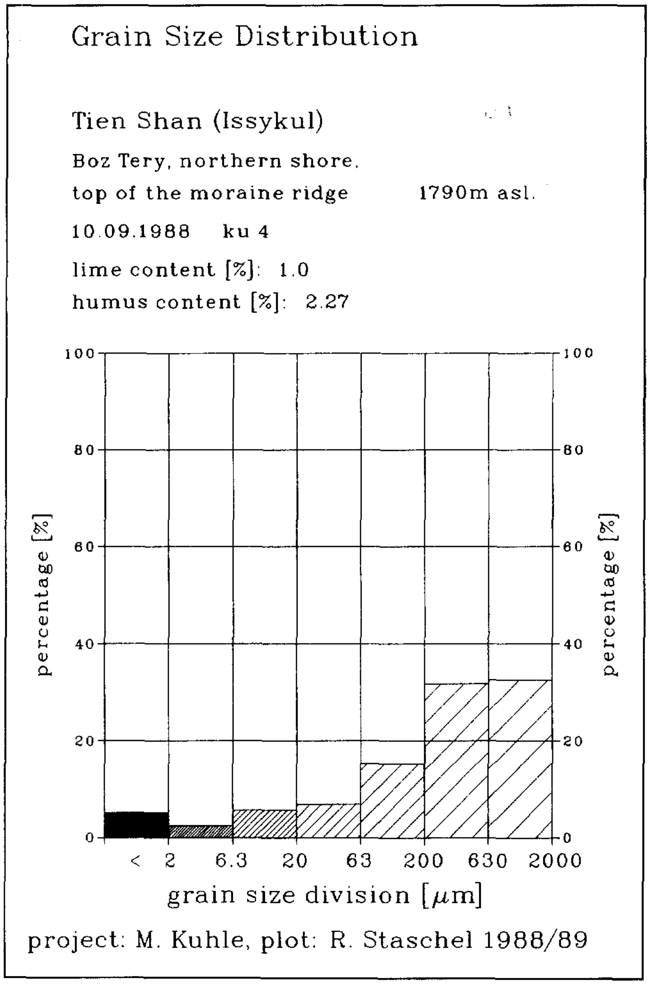


Fig 35 Main Ice Age moraine material from a depth of 20 cm (locality: Fig 1 No. 12; Figs 34 and 36), showing the typical "fine grain size peak" in the bimodal course of the graph of the columnar diagram. Locality where the samples were taken: 42°40'N/77°12'E.

which provided a channel for easy outflow, urging the river to take a shortcut.

Another alternative suggested by modeling experiments (see below) is the complete filling in of the basin by glacier ice. Indeed, it was in intermontane basins where ice thicknesses attained their maximum values which were well in excess of the lake's depth. If this was the case, some of the above evidence should be related to the sequence of deglaciation events, not to LGM.

Origin of Boam Canyon. The establishing of LGM ice-damming of Lake Issyk-Kul provides a natural explanation both for lake-level oscillations and mechanism of the Boam Canyon formation. The lake water balance during glaciation turned positive, which could have been a mere result of diminished (by ca. 50%) evaporation. Hence, the water discharge from the lake was inevitable. On the other hand, judging by the LGM situation in the basin as reflected in our map, the only way of this discharge was by breaking through the ice dams. This sort of lake outbursts are typically caused by rises in water level, while the specific rate of the rises depends on the ice dam height (Nye 1976). As the dam on the Kokpak-Kyrkoo profile had an estimated relief of about 350 m, the expected lake-level rises could reach 300 m, well in excess of the heights recorded in lacustrine terraces. The established fact of the past Lake Issyk-Kul ingressions in Kochkor Basin which now hosts Upper Chu River supports this estimate. Minimum elevations within the basin - 1900 m asl; nevertheless, the glaciers, which invaded the basin from the east, went afloat and produced icebergs.

Relatively high levels of "glacial" (or, rather, late-glacial) Lake Issyk-Kul have been suggested by the structure of above-mentioned ice marginal ramps, also. Judging by the sections studied between lower reaches of Turasu and Akterek rivers, in particular in the Kara-Ortok hill (Fig 1 and 2), the ramps are made up of gradedly bedded silts and sands, similar to the parallel-bedded deep-sea turbidites. It is probable, that those silt and sand masses were accumulated in underwater environment, which could only be possible if the lake level stood much higher than today.

The difference between the highest level of "glacial" or late-glacial Lake Issyk-Kul and the altitude of a rocky threshold in its outflow suggests that the volumes of water, which were repeatedly discharged during lake-outbursts, were equivalent to a 200–250 m water layer, ie amounted to about 1300 km^3. This implies that during the outbursts occurring at intervals of 100–150 years (another of our estimates), giant water discharges (on the order of a million m^3/s), and extreme flow velocities (equal to or exceeding 20 m/s), typical of such catastrophic events, did take place. This, in turn, implies very high erosional potential of the outbursts. Hence, it were the outbursts which can account for the canyon formation, making tradtional resorts to arguments of neotectonics superfluous. As for the debris, produced by hollowing the canyon, it went on building of an enourmous outwash fan which blankets the bottom of Lower Chu Basin. A section of the fan is exposed in a quarry near the Settlement of Ivanovka (Fig 1), 85 km west-northwest of the Boam Canyon mouth, where a sequence of cross-bedded gravel, loess and coarse-grained sands outcrop; their total thickness appears to be in excess of 80 m.

The above information gives a chance to see the geological "potency" of Tian Shan glaciation in proper perspective. Specifically, it makes us re-assess the part played by glacial and glaciofluvial deposits in sedimentary sequences of the region. For instance, we expect that upcoming detailed studies will result in re-interpretation of the sediment sequences from Lake Issyk-Kul bottom and coast, in acknowledging the important role played in the sequences by glacial debris.

Ice thickness and surface elevation. To determine thickness and surface elevation profiles of the ice cover, which formed on plateaus and ranges south of Lake Issyk-Kul, a finite-element program solving the continuity

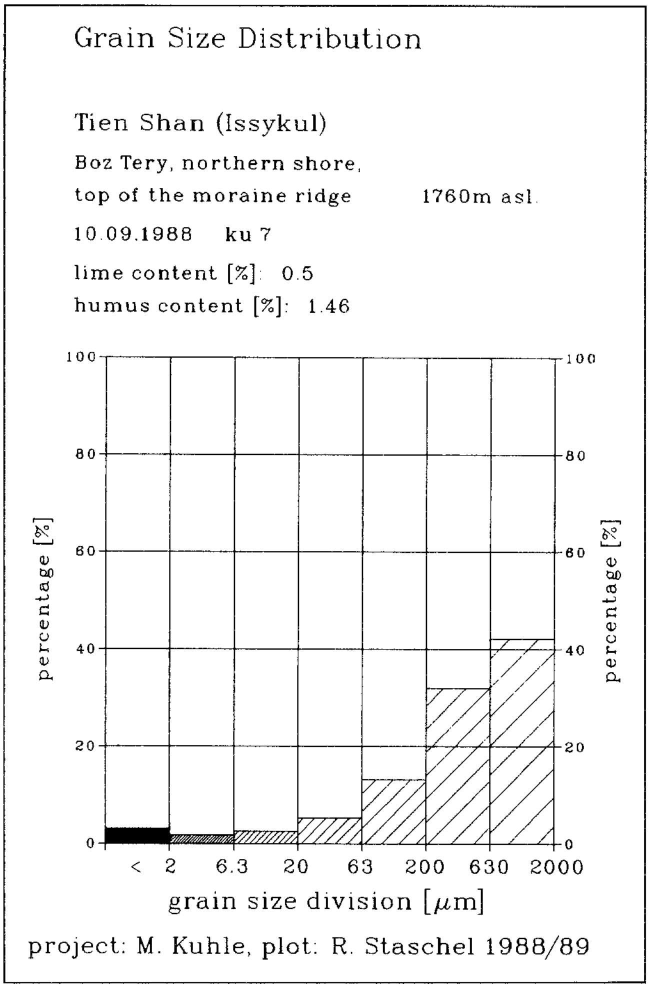


Fig 37 Somewhat richer in sand, but less well endowed with pelite (cf. Fig 35), this end moraine sediment belongs to the same Main Ice Age moraine complex as shown in Fig 36 (Fig 1, No. 12), with the bimodal course of the column levels, including a distinct "fine grain size peak" and a "coarse grain size peak" characteristic of moraine substrata. Locality where the sample was taken: 42°39'59"N/ 77°12'01"E.

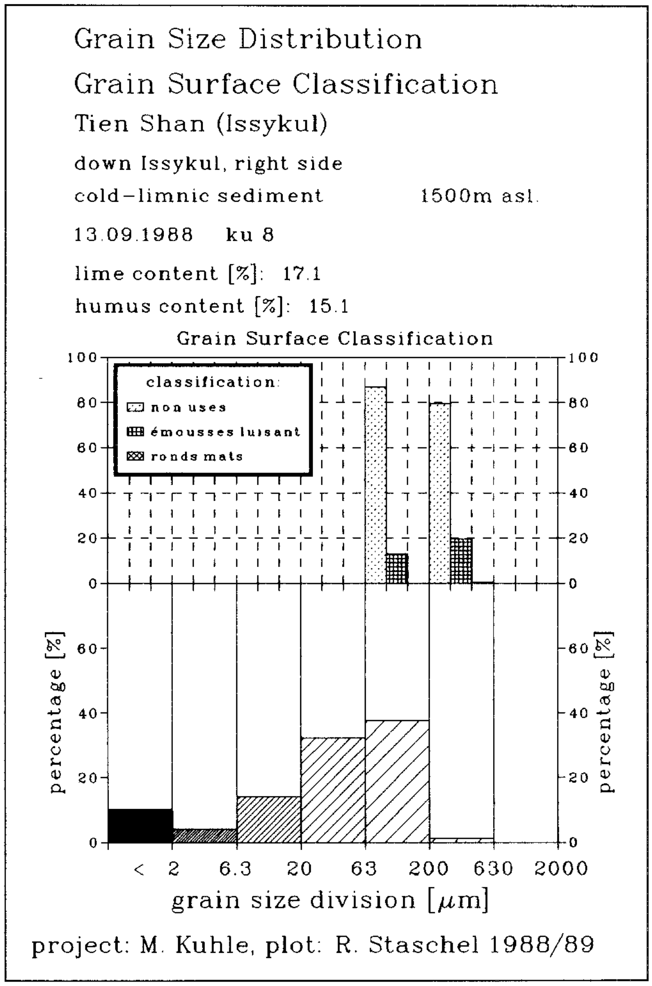


Fig 39 Held back by the glacier in the Chu valley above the Boam canyon, the lake sediments are tens of metres above the present talweg (Fig 1 on the left above No. 15; Fig 2). Their grain size distribution has the characteristics of till. A "fine grain size peak", or bimodal course, is shown in the columnar diagram. Under the key word "none uses" (ie not affected by fluvial or aeolian influence) the grain surface classification shows grains broken by glacier force as predominating in at least 80% of the sand fraction analysed here. 13–20% of the grains reveal fluvial influences, which can be explained by the involvement of glacio-fluvial dynamics. Locality where the sample was taken: 42°30'20"N/75°52'30"E.

equation for ice flow was employed. Input to the model consisted of the spatial coordinates of the nodal points and the values of each of the following material properties at the nodal points: bedrock elevation, flow law constant, sliding law constant, present ice surface elevation, accumulation rate and percent of the flow due to sliding. Output consisted of time-dependent ice elevations and isostatically adjusted bedrock elevations at each nodal point, as well as column-averaged ice velocities at the centro id of each element defined by a group of nodal points (Fastook and Chapman 1989).

The finite-element grid consisted of 4-node quadrilaterals with material properties defined at each of the nodal points. In this modeling experiment, the grid had 1735 nodes and 1639 quadilateral elements, with a grid spacing of approximately 16 km. The mechanism controlling the velocity was assumed to have been ice flow (without sliding), and the flow law constant was taken as 3.0 Bar m^{f}.

All nodes except those within the boundaries of Lake Issyk-Kul were assigned mass balance relationship typical of polar continental regions. They have an unadjusted equilibrium line at 312 m, a maximum ablation rate of −1 m/yr ice equivalent, and a peak accumulation rate of 0.26 m/yr at 650 m. Beyond this, the accumulation rate declines to less than 0.10 m/yr at elevations above 4000 m. This mass balance curve could be slid up or down to reflect local snowline elevations. Since snowline elevation is largely a function of latitude, we needed only define snowline adjustment at the pole and its slope (100 m per 100 km) to calculate the snowline elevation at any gridpoint. Knowing that, we could use the mass balance curve to obtain a net accumulation or ablation rate for the node.

The nodes within the boundary of Lake Issyk-Kul were assigned ablation rates of –5 m/yr, which reflected the calving mechanism of ice wastage. If the lake were to freeze solid or to be chocked with icebergs then this calving would cease and the lake nodes be assigned a mass balance similar to the one described above. In the light of Lake Issyk-Kul paleoclimate and geomorphology, the latter assumption looks realistic. Hence the modeling results, where they describe the lake basin, should be regarded as minimal.

Modeling experiments were conducted to test two climate scenarios - one with a gradual, and another with an abrupt change in the regional snowline elevation. In first experiment, the snowline was lowered in 200 m intervals every 1000 years, beginning with the present elevation of 4100 m, until the LGM value of 2900 m was reached at 6000 years. This value was maintained until the volume of the ice cover equilibrated at 19,000 years. Then, to simulate the deglaciation, the snowline was raised 200 m every 2000 years until its present value was reattained at 29,000 years. Again it was held here until it equilibrated at 33,000 years. Surprisingly, the ice cover at this point turned out more extensive than at present. The snowline was further raised to 4900 m, at which value the ice cover disappeared at 41,000 years. After that, the snowline was returned to its present level once more, allowing the ice cover (slightly larger than its minimum size) to equilibrate at 45,000 years.

In second experiment, the snowline was abruptly lowered to its LGM value and held at this level until attainment of equilibrium at 15,000 years. The snowline was then returned to its present position, and again held until equilibrium was attained at 22,000 years. Since this did not result in the ice cover shrinkage to its present configuration, the snowline had to be further raised to 4700 m, and held until the ice disappeared by 25,000 years (Fig 4).

The abrupt lowering of the snowline from 4100 m to 2900 m resulted in the widespread formation of a thin ice cover that then thickened and extended its margins gradually with time. At its equilibrium configuration, the ice cover was found to have increased in size to such an extent that it reached the limits determined by our field research. In particular, the experiments suggested, that it took the ice cover only 10,000 to 13,000 years of growth, depending on type of assumed scenario, to reach and intrude the water body of Lake Issyk-Kul. The maximum ice thicknesses localized over intermontane depressions where they reached 2500–3000 m, while over high plateaus ice grew to maximum thicknesses of 1200–1250 m (Fig 5 and 6).

In both the experiments, the time of ice-growth to *nearly* maximum volumes turned out about two times shorter than the time required for ice-cover equilibration. This is clear from comparison of ice-cover configurations 2 (7 ka after completion of snowline lowering) and 3 (a full equilibration which took another 6 ka). But in general, whichever scenario we may find more realistic, our modeling experiments warrant the conclusion that the time required for building the Tian Shan ice cover was sufficiently short to fit into the brackets set by the last ice-age chronology. The same can be said about the time of full ice-cover equilibration which took only 15 to 19 ka.

Tian Shan ice cover vs. Tibetan Ice Sheet. It may be argued that the ice cover of Tian Shan was a perfect paleoglaciological analogue of the great Tibetan Ice Sheet. There is a number of arguments favouring this contention. Indeed, both Tian Shan and Tibet have similar - dry and cold - climate; they have the same style of geomorphology with vast high-level plateaus lying at the elevations of 4000 and 5000 m, respectively; present-day snowline in both the regions goes some 300 m to 700 m above the plateaus getting in touch only with the higher ridges towering over the plateaus; the same rate of snowline depression during LGM, amounting to the mean value of 1200 m, characterized the regions. Hence, during the ice age, the snowline over both Tian Shan and Tibet lowered well below plateau surfaces making their continuous glacierization inevitable (Fig 7). Our modeling experiments aimed at simulating the Tian Shan glaciation seem equally applicable to Tibetan Ice Sheet, although the ice-sheet build-up and equilibration probably took some more time on Tibet than on Tian Shan.

In our view, the LGM Central Asian glaciation was represented by a continuous bow-shaped chaine of mountain ice covers and sheets, which semi-confined Tarim Basin with Takla Makan Desert on its bottom. Paleoclimatically, this setting strongly suggests that powerful winds born by the westerly jet-streams coupled with katabatic air-flows focused on Takla Makan and the neighbouring deserts (Fig 8). Which, in turn, implies that the deserts, first of all Takla Makan, were the source areas of the magnetic dust masses that were transported and deposited by wind to form the Chinese loesses. It seems improbable that Gobi Desert alone was that source area as believed by Kukla (1988); it is only natural to assume that, to account for as major a loess field as the Chinese, commensurately large ice sheets and corresponding wind systems were needed.

With all said in mind, we may further speculate that the history of the Central Asian glaciations, including ice-sheet inception and changes on Tian Shan and Tibet, had to be recorded in the Chinese loesses – just like the history of the Greenland and Antarctic Ice Sheets were recorded in bottom sediments of the surrounding oceans. If this inference is upheld by further studies, then the Chinese sequences which are containing the record of about 20 loess/paleosol alternations and are underlain by the Gauss/Matuyama boundary (ca. 2.43 my) should be considered a *curriculum vitae* of the Central Asian glaciations. The Tibetan Ice Sheet and its satellites appear to have been incepted simultaneously with the ice sheets surrounding the North Atlantic.

Conclusions

Judging by the authors' studies in Lake Issyk-Kul area, the horizontal and vertical extent of Würm glaciation in Tian Shan Mountains was an order of magnitude larger than shown by all previous reconstructions. At the LGM, continuous ice covers buried the systems of mountain ranges and plateaus, their outlet glaciers breached border ridges and descended to foothills, till elevations of 1500 m asl in the south and 900–1000 m asl in the north. The glaciers originated from high plateaus flowed into intermontane basins; as a result, in some basins ice thickness exceeded 2500 m; a number of glaciers from Terskey Alatau and Kungey Alatau Ranges focused their flow on Lake Issyk-Kul and invaded its basin with ice which led either to ice shelf formation or, possibly, to a complete ice filling of the basin.

The environmental impact of the glaciation turned out much stronger than it was formerly admitted, also. The past glaciers reshaped the mountains by scouring trough-valleys and glacial cirques and by breaching the ridges bordering high plateaus, while the glacially-derived debris were transported downslope to form moraines and ice marginal ramps, to fill in Lake Issyk-Kul hollow and the rest of intermontane basins. The glaciers interfered with hydrographic systems, too. In particular, they impounded Chu River and chocked Issyk-Kul with icebergs, which caused big-range lake-level oscillations, the formation of Boam Canyon (by the impact of lake outbursts), and the deflection of Chu River from Lake Issyk-Kul (by incision of the canyon and build-up of an ice-raft delta at the lake outflow).

The ice sheet/water basin interactions, as studied in Lake Issyk-Kul area, may help in understanding the ice-age behaviour of the variety of different basins surrounded by calving glaciers. In particular, they provide a working model for explaining the glacial evolution of Lake Baikal (and its unique faunas) or even the Arctic Ocean which got confined by grounded marine ice sheets.

The average LGM lowering of regional snowline in Tian Shan was found to equal 1200 m. Since the present snowline in Lake Issyk-Kul area has the altitudes of 4100–4200 m, passing only a few hundred meters above the high plateau surfaces, the Würmian snowline dropped well below the surfaces making their glacierization inevitable. An analogous snowline/earth surface relationships were characteristic of present-day and ice-age Tibet, which had to result in similar changes of glaciation. The whole history of the Central Asian glaciations seems to be recorded in the Chinese loess sequences. The beginning of the loess accumulation dating back to the Gauss/Matuyama boundary (2.43 my) may be considered as a milestone marking the onset of the Central Asian glaciations.

Judging by our finite-element model, ice covers would form and equilibrate on Tian Shan in 19,000 years (first climate scenario with a gradual change in snowline elevation) or 15,000 years of growth (second scenario with an abrupt lowering of the snowline). The modeling experiments yielded ice thicknesses (up to 2500–3000 m in intermontane basins and 1250 m on plateaus) and ice-marginal positions; the latters conformed to the results of the authors' field observations.

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On the Cause of Glacier Mass Balance Variations in the Tian Shan Mountains

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ABSTRACT: Over the period of more than 30 years the mass balance observations have been carried on the several representative glaciers of the Tien Shan. But these data are not sufficient to explain the asynchronous degradation of the glaciation in different parts of the mountains. Special field observations were undertaken to study asynchronous changes in mass balance. The programme used an imitation model of the mass balance monitoring (Dyurgerov 1988). According to the model the annual values of glacier mass balance b_n and equilibrium line altitude ELA_n were replaced by the current values of b_t and ELA_t measured during the ablation season. The imitation model was evaluated on Tuyuksu Glacier in 1987–1989. The results showed that the function $b_n(ELA_n)$ can be replaced by $b_t(ELA_t)$ which is nonlinear and can be approximated by the hypsographic curve of the glacier. After these tests the similar measurements were accomplished in summer of 1989 on the Tuyuksu Glacier, Sary-Tor and Glacier No. 1. It was established that the asynchronous changes in mass balance were primarily due to the increase of summer snow accumulation eastward the Tien Shan. The second reason was the difference in the altitudinal position of the glaciers.

Introduction

In mass balance studies are usually operated with annual values (Kasser 1967; 1973; Muller 1977; Haeberli 1985; Glacier mass balance bulletin, 1991). But annual data do not permit to trace back mass balance formation over a year. We can only assume that the process of mass balance changing varies according to climatic conditions, aspect, size, shape and altitudinal range of a glacier. Thus, every glacier responds individually to climatic changes.

To study variations in glaciers behaviour we used the imitation model of M. Dyurgerov (Dyurgerov 1988). In the model the annual values of mass balance b_n and corresponding ELA_n are replaced by the current values (b_t, ELA_t) measured during one ablation season. Assume

$$b_n(ELA_n) \approx b_t(ELA_t) \quad (1)$$

This relation allows to obtain the additional information about mass balance formation for the whole altitudinal range of the glaciation.

We chose Tuyuksu Glacier, Sary-Tor, and Glacier No. 1 as they represent three different areas of the Tien Shan (Fig 1). Tuyuksu Glacier is located on the northern windward slope of the Zailiysky Alatau Range, which is the most humid part of the north-western Tien Shan. Sary-Tor is one of the major glaciers of the Akshiyrak compact glaciation on the Interior high plateaus of the Tien Shan. Glacier No. 1 located on the Tianger Range represents a dispersive glaciation of the most arid Eastern Tien Shan. Tab 1 shows the main characteristics of the glaciers.

Glacier	Cordinates	Area, km^2	Heigt, km upper	lower	Mean mass balance gr/cm^2
Tuyuksu	42°N; 77°E	3.1	4.23	3.45	−36,4
Sary-Tor	41°49'45"N; 78°10'30"E	3.6	4.75	3.85	− 9,0
Glacier No. 1	43°05'35"N; 86°49'48"E	1.8	4.87	3.74	−11,7

Tab 1
Characteristics of the glaciers

Verification of the Mass Balance Monitoring Model

To test the assumption (1) one needs long-term records of ELA_n and b_n. Such records are available for Tuyuksu Glacier (Fig 1) where observations have been carried on since 1957. The records are presented in (Muller 1977; Haeberli 1985; Makarevich et al. 1987; Haeberli, Muller 1988; Glacier mass balance bulletin, 1991). The data were obtained from 160 snow stakes and several pits. Mass balance of the whole glacier for the i-th instant (b_i) was calculated as the weighted sum of mass balance for each 50-meter altitudinal interval in the ablation zone and 100-meter interval in the accumulation zone:

$$b_i = 1/S \Sigma b_{ij} s_j \quad (2)$$

where bij, sj - mass balance and area of j-th altitudinal interval for the i-th monent; S - total glacier area; n - number of altitudinal intervals.

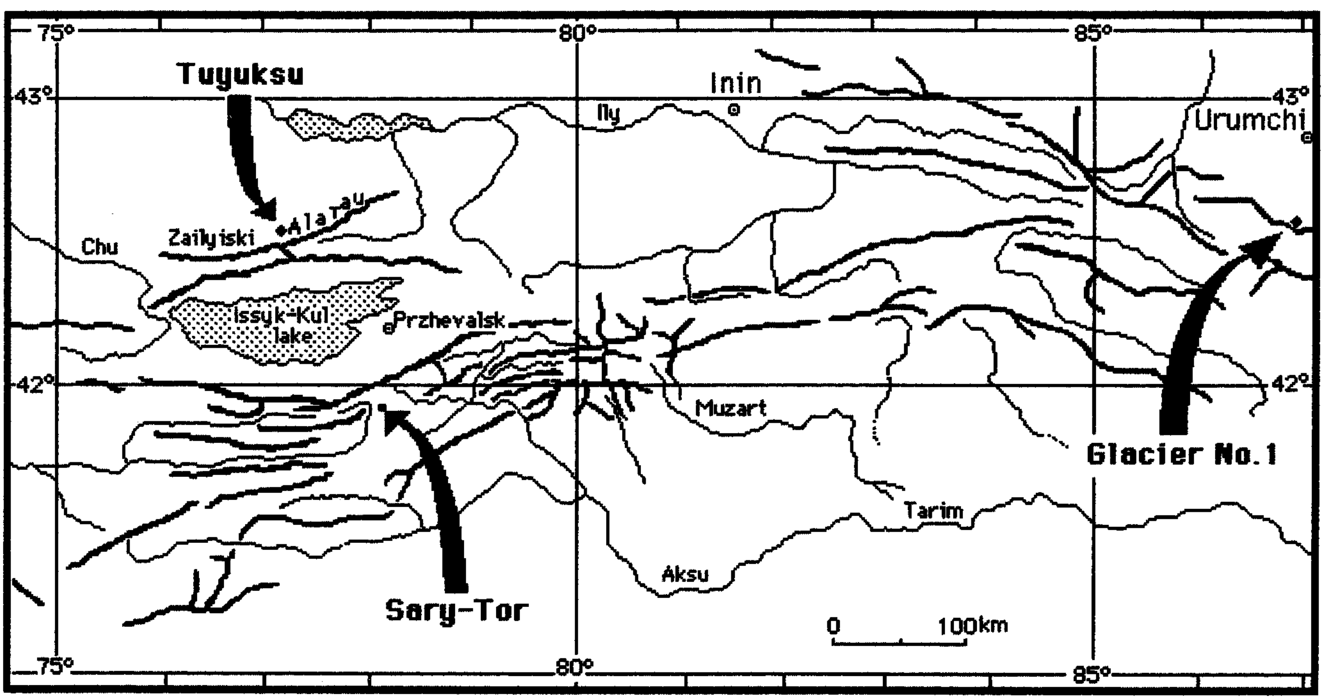
The equilibrium-line altitude was determined, on the one hand, while taking readings from the snow stakes and, on the other hand, by fixing the elevation of the zero balance point on the balance curves bt(z), where z stands for the altitude a.s.l.

As a result, we obtained 40 values of b_i and bt(z) which enabled us to establish the association between the current mass balance and ELA_t (Fig 2). The function $b_{ti}(ELA_{ti})$ characterizes the glacier area from the terminus at the altitude of 3400 m up to the altitude of 3835 m where the mass balance in 1988 made up 61 gr/sm^2.

So, the answers to two main questions were obrtained:

1. Generally we can replace $b_t(ELA_t)$ by $b_n(ELA_n)$. But in cold snowy winters when b_n and b_t are positive and extremely high considerable discrepancies are possible.
2. The associations between mass balance b_n or b_t and glacier equilibrium line altitude ELA_n or ELA_t can be approximated by the glacier hypsographic curve. The correlation coefficient between bn and ELA_n is 0.90. However, when the equilibrium line occupies its highest position, the error of the linear approximation can be up to 70 gr/sm^2. Such conditions took place in 1978 when ELA_n was 4210 m and the annual mass balance was equal to −148

Fig 1 Location sketch map of Tuyuksu, Sary-Tor glaciers and Glacier No. 1



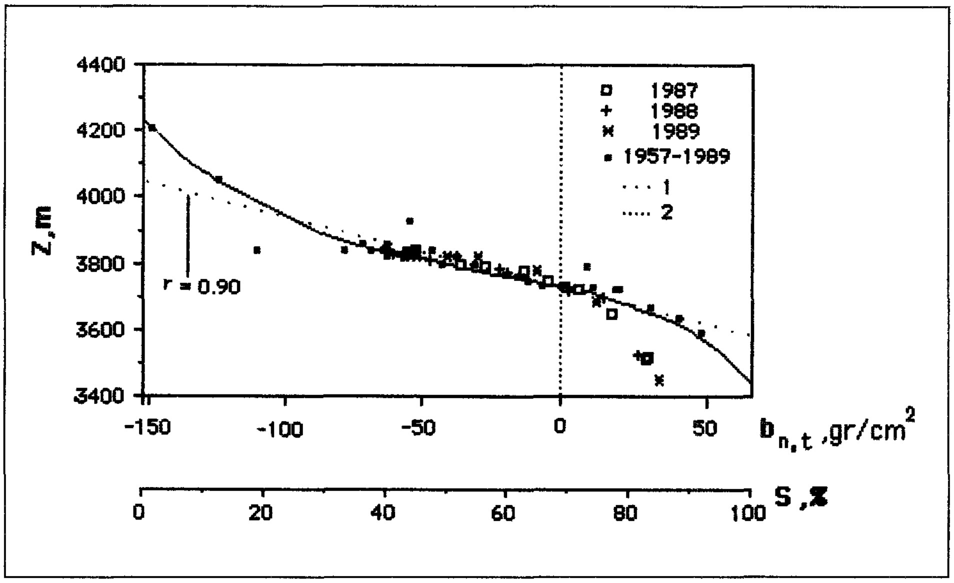


Fig 2
Glacier equilibrium-line altitude (z) versus current (1987-1989) and annual (1957- 1989) values of the mass balance (b_t, b_n correspondingly)
1. linear approximation
2. zero mass balance

gr/sm². Using the approximation by the hypsographic curve we assume that the linear association occurs between accumulation area ratio (AAR) and b_n which holds for the majority of glaciers (Mikhalenko 1990).

Simultaneous Monitoring in 1989

The results of the mass balance and ELA measurements in summer of 1989 are summarized in Tab 2-4. The mass balance was changing simultaneously on theglaciers before July, 14-16 that was shown in Fig 3. By that time the equilibrium line had already appeared on the surface of Sary-Tor and Glacier No. 1 (Fig 4). Tuyuksu had been still covered by winter snow. In the period between July, 16-Aug., 11 the mass balances of Tuyuksu and Sary-Tor were decreasing synchronously reaching zero value by the beginning of August. Mass balance of Glacier No. 1 was positive at that period and practically did not change until the end of August due to heavy snowfalls in the accumulation area (Fig 4). Since early August the mass balance began to decrease rapidly on Tuyuksu and slower on Sary-Tor where snowfalls also occurred like on Glacier No. 1.

Thus, the simultaneity of mass balance changes was broken in the middle of the ablation period. The most probable reason for that is the increase of the contribution of summer precipitation to the annual sum eastward from Tuyuksu to Sary-Tor and further to Glacier No. 1. Differences in the air temperature are essential either (Tab 5).

As the upper boundary of runoff (R_{up}) was above Tuyuksu glacier, the liquid runoff came down from the whole glacier area along with the infiltration inside. On the

Tab 2
Current mass balance of Tuyuksu glacier in 1989 (g/cm²)

Z, m	Date / S, km²	6.03	6.17	6.29	7.14	7.20	7.30	8.10	8.21	8.30	8.09	9.20
3400-3450	0.046	23	31	21	10	-5	-40	-89	-122	-154	-164	-201
3450-3500	0.09	14	27	17	10	0	-35	-82	-113	-134	-141	-163
3500-3550	0.137	19	26	19	12	4	-29	-69	-105	-120	-127	-147
3550-3600	0.181	20	30	24	17	10	-24	-61	- 92	-102	-106	-134
3600-3650	0.116	21	35	28	22	14	-15	-48	- 74	- 87	- 88	-107
3650-3700	0.178	26	44	37	31	23	- 6	-36	- 59	- 66	- 62	- 91
3700-3750	0.422	35	50	44	39	33	10	-13	- 41	- 53	- 49	- 70
3750-3800	0.449	35	54	50	48	44	22	0	- 20	- 29	- 25	- 44
3800-3900	0.299	38	58	55	58	59	44	27	14	7	11	1
3900-4000	0.303	34	51	48	52	53	40	25	15	9	13	5
4000-4100	0.3	24	42	40	45	46	36	23	15	10	15	7
4100-4219	0.142	13	33	31	36	38	31	20	13	9	14	8
total	2.663	32	45.2	40.4	38.9	25.7	14.4	- 8.6	- 29	- 39	- 37	- 53
ELA, m		-	-	-	-	3450	3685	3780	3820	3820	3820	3825

* Glacier is covered by snow

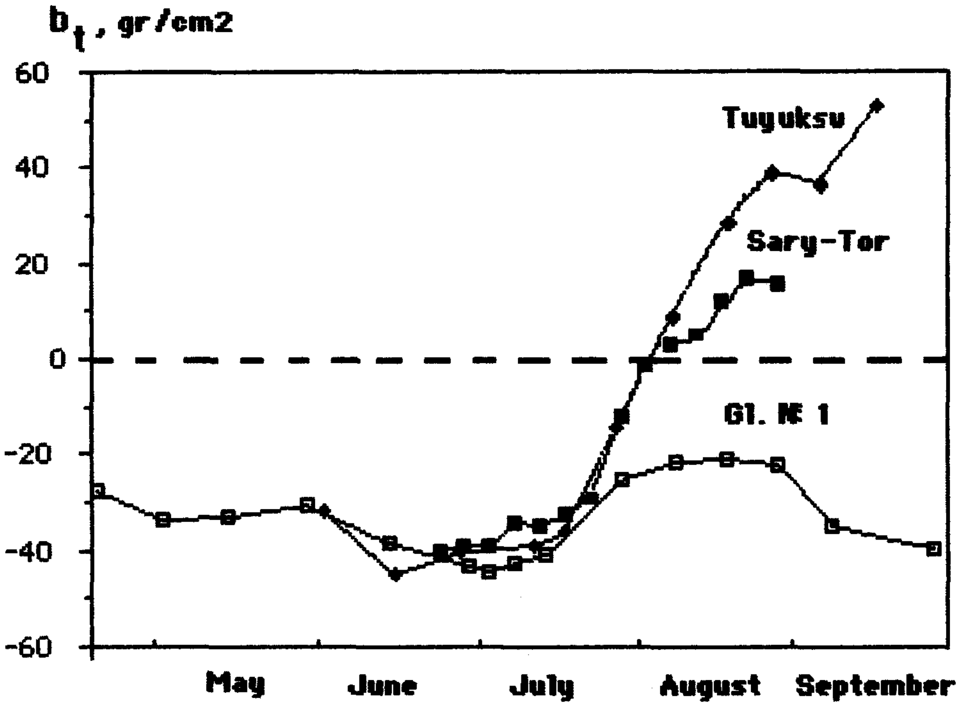


Fig 3
Current mass balance during 1989

contrary, the runoff boundary of Sary-Tor and Glacier No. 1 was located rather low, and the mass balance above it was continuously growing in the cold firn zone (Shumskii 1964) due to summer snowfalls. At the same time, the mass balances of these glaciers were decreasing beneath R_{up}.

Altitudinal Structure of Mass Balance

To study the behaviour of mass balance from year to year or during the ablation period, we consider the term "altitudinal structure of mass balance" (AS) which describes the changes in current or annual balance curves (Dyurgerov 1991; Kunakhovitsh 1991).

The ratio of balance curves for two different years can be written as

$$b_2(z) = ab_1(z) + c \tag{3}$$

where b_2, b_1 are the consequent mass balances (balance curves), a and c are linear coefficients. There are two extreme types of mass balance altitudinal structure.

In maritime conditions with dominating winter snow accumulation and high air temperature coefficient a can be

Tab 3 Current mass balance of Sary-Tor glacier in 1989 (g/cm²)

Z, m	Date / S, km²	6.25	6.30	7.05	7.10	7.15	7.20	7.25	7.31	8.05	8.10	8.15	8.20	8.25	9.01
3.86–3.90	0.1	7.3	–5	–15	–37	–46	–59	–81	–109	–127	–140	–152	–164	–177	–175
3.90–3.95	0.173	18.2	12.2	6.6	– 3.4	–11	–19	–33	– 66	– 83	– 94	–105	–113	–124	–124
3.95–4.00	0.18	25	20.4	14.4	3.7	– 5.6	– 6.4	–21	– 53	– 70	– 78	– 92	– 99	–108	–107
4.00–4.05	0.211	32.2	27.5	23.4	13.6	8.7	6	– 3.8	– 36	– 52	– 58	– 70	– 78	– 86	– 89
4.05–4.10	0.266	33.4	34	30.2	18.7	14.5	12.2	6.2	– 23	– 40	– 47	– 58	– 67	– 77	– 80
4.10–4.15	0.273	39.5	36.2	39.1	26.8	15.7	14.8	9.3	– 20	– 39	– 43	– 56	– 65	– 74	– 76
4.15–4.20	0.336	34.4	31.2	28.8	20	22.9	17.5	8.5	– 15	– 28	– 35	– 44	– 53	– 62	– 62
4.20–4.25	0.273	38.9	39.3	41.8	43	47.4	42	34.9	17.2	10	6.8	1.6	– 0.4	– 4.6	– 2.1
4.25–4.30	0.289	41.1	41.8	45	43.6	50.8	50.7	51.4	30.6	14.5	12.2	7.6	4.9	2.8	6.1
4.30–4.35	0.422	37.1	39.3	37.4	35.8	40.3	35.4	36.2	30.4	19.6	16.3	12.6	9.7	8.4	11.2
4.35–4.40	0.375	42	40.2	38.9	37	42.2	37.4	37.3	31.4	20.8	18.6	21.4	26.3	28.4	31.5
4.40–4.45	0.180	64.6	39.3	74.8	78.6	85.1	91.5	96.1	95	95.8	96	96.4	96.8	97	99
4.45–4.50	0.156	74.8	78.1	85.9	93.4	102	103	108	103	104	106	107	108	109	113
4.50–4.55	0.102	60.3	62.8	65.3	66.1	70	72.1	74.5	68	68	68	68.4	68.7	69.3	71.4
4.55–4.60	0.070	48.1	47.3	46.4	47.6	49.7	51.4	52	55.2	53.4	59.2	60.2	62.4	62.8	64.3
4.60–4.80	0.208	35.7	36.1	36.4	37.2	37.3	38.9	45.4	46.3	43.8	40	40.8	41.4	42.6	44.9
total	3.614	40	39.1	38.8	34.3	35	32.7	28.8	11.9	0.9	– 2.8	– 4.2	– 12	– 17	– 16
ELA, m		*	3894	3912	3950	3995	4003	4045	4198	4214	4219	4225	4233	4258	*

* The galcier is covered by snow

Z, m	Date S, km²	4.18	5.02	5.15	5.31	6.16	7.01	7.05	7.10	7.16	7.31	8.11	8.21	8.31	9.11	9.30
3.81–3.85	0.018	5.6	14.6	5.9	−14	−12	−19	−18	−22	− 37	− 88	−107	−126	−143	−137	−139
3.85–3.90	0.024	13.5	22.3	17.7	0.2	3.8	− 4.1	− 2.2	− 7.7	− 17	− 63	− 84	− 98	−115	−108	−109
3.90–3.95	0.047	15.1	24.3	17.3	7.7	12.5	5.5	7.8	2.6	− 5.4	− 42	− 57	− 69	− 83	− 73	− 73
3.95–4.00	0.073	18.3	24.1	21	19.6	28.4	26.7	29.4	25	207	− 4.1	− 10	− 14	− 20	− 10	− 4.1
4.00–4.05	0.107	16.5	24.2	23.1	18.9	27.9	28.4	31.4	27.7	26.6	6.6	1.8	1.6	1	15	21
4.05–4.10	0.116	22.3	32	32.6	32.4	38.6	38.3	41.5	38.4	39.4	21	20.2	21.7	22.5	39.5	45
4.10–4.15	0.055	32.9	35.8	36.5	36.2	42.1	54.1	55.8	56.1	54.3	50.6	49.8	51.8	54	66.8	73
4.15–4.20	0.044	35.4	38.5	39.4	39.2	46.6	62.2	57.8	55.7	56.8	53.6	52.7	55.4	59.7	75.3	83
4.20–4.25	0.037	40.2	44	46.1	46.7	54.8	65.7	71.8	71.4	70.7	67.8	67.1	70.6	77.6	91.7	95.5
4.25–4.30	0.037	43	47.1	48	49.3	58.2	70.6	71.4	72.4	70.6	69.8	69	72.8	81.3	94	98.6
4.30–4.35	0.041	38.3	42	43.1	44.2	53.8	64.9	61.2	62.3	61.3	60.1	60.9	64.8	77.3	89.4	93.3
4.35–4.40	0.040	50.8	55	56.3	58.1	67.6	85.5	86.2	92	90	89.4	90.8	95.8	108	116	121
4.40–4.45	0.027	56.1	60.3	61.6	64.5	73.2	88.6	92.4	93.5	98	97.4	99.2	104	115	127	132
4.45–4.49	0.011	30.3	35.1	36.3	39.6	46.1	52.2	54	55.2	54	57.9	54	62	72.6	82.3	87.3
total	0.677	27.6	33.9	33.1	30.8	38.3	42.9	44.7	42.8	40.9	25.5	21.7	21.2	22	34.6	39.4
ELA, M		*	*	*	3833	3888	*	*	*	3927	4003	4014	4035	4036	3997	3987

* The glacier is covered by snow

Tab 4 Current mass balance of Glacier No. 1 in 1989 (g/cm²)

equal to 1. Then $b_2(z) = b_1(z)+c$. Such situation was described by M. Meier for South Cascade glacier (Meier 1962). This is the additive type. Perfect balance curves would be parallel in this case.

In the continental arid conditions coefficient c is equal to zero. Then $b_2(z) = a\ b_1(Z)$. This is the multiplicative type.

In 1989 three Tien Shan glaciers had different mass balance (Tab 5) with quite different altitudinal structures. On Tuyuksu it was closer to the additive type. The extreme multiplicative type was observed on Glacier No. 1 (Fig 4). The position of the upper boundary of liquid runoff depends on the type of AS. The closer is AS to the multiplicative type, the lower is the position of R_{up}, and vice versa.

According to the imitation model we consider each transient balance curve $b_t(z)$ as representing the annual curve $b_n(z)$. So we can estimate the long-term altitudinal structure of mass balance and annual curve $b_n(ELA_n)$ over one ablation period.

As the altitudinal structure of mass balance in continental conditions is much closer to the multiplicative type, the curve $b_n(ELA_n)$ has the smaller vertical gradient than any curve in maritime climate (Glacier mass balance bulletin, 1991). We must note that the approximation of $b_t(ELA_t)$ by the hypsographic curve may have considerable discrepancy when the equilibrium line takes its lowest position. The major discrepancies were observed on the continental glaciers. As the mass balance on Glacier No. 1 was still increasing in early summer of 1989 despite the rise of equilibrium line, three lowest points on $b_t(ELA_t)$ curve characterized the beginning of melting period but not the ablation (Fig 5). The situation considered here is quite typical for the conditions of low winter accumulation and high contribution of the summer accumulation to the annual mass balance. In this case, the form of the curve in its lower area differs from the hypsographic approximation. But there are no annual values of such extremely high mass balance among the long-period records on these glaciers.

Discussion

The regularities of the mass balance formation found in summer of 1989 for the whole observation period on Tuyuksu and Glacier No. 1 can be extrapolated. As well as in 1989, the average mass balance for the thirty-years period was negative and lower on Tuyuksu than on Glacier No. 1 (Tab 5). The annual precipitation decrease along with the increasing contribution of the summer snowfalls in the annual sum results in the increasing continentality of the

Tab 5 Meteorological data from the nearby stations of Tuyuksu, Sary-Tor glaciers and Glacier No. 1 (respectively) for 1989:
P_a – annual precipitation
P_s – summer precipitation
b_n – net mass balance

Meteo station	H, m	Pa, mm	Ps, mm	Ps, %	ts, °C	bn, g/cm²
Tuyuksau	3450	1071	485	45	3.7	-46.0
Tian Shan	3614	160	80	50	3.0	-16.7
Dashiku	3539	487	314	65	3.1	11.0

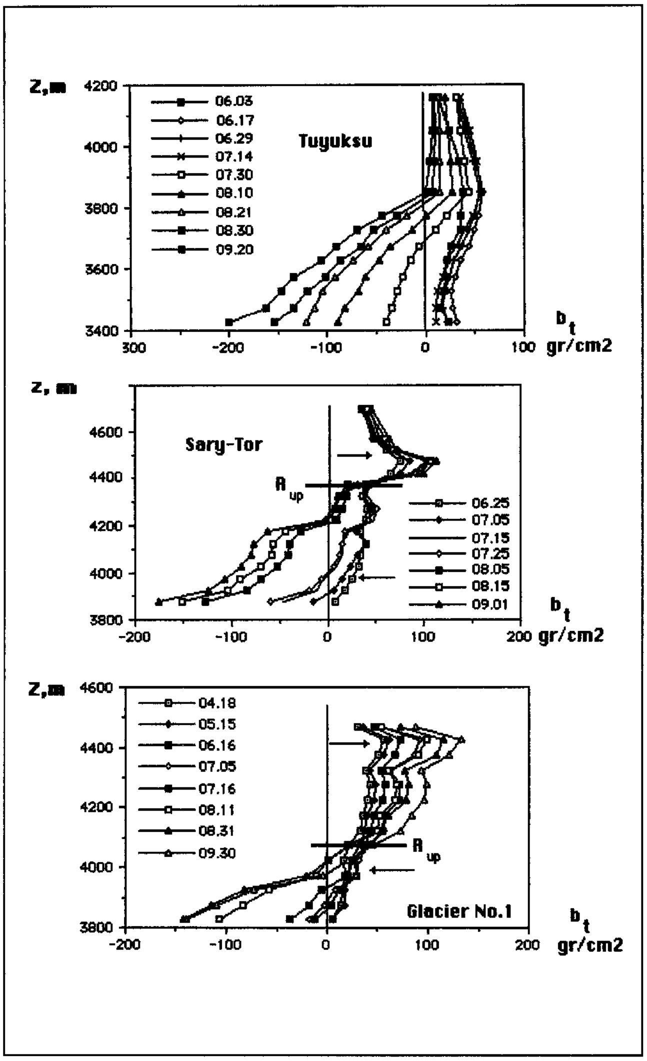


Fig 4 Mass balance curves for 1989

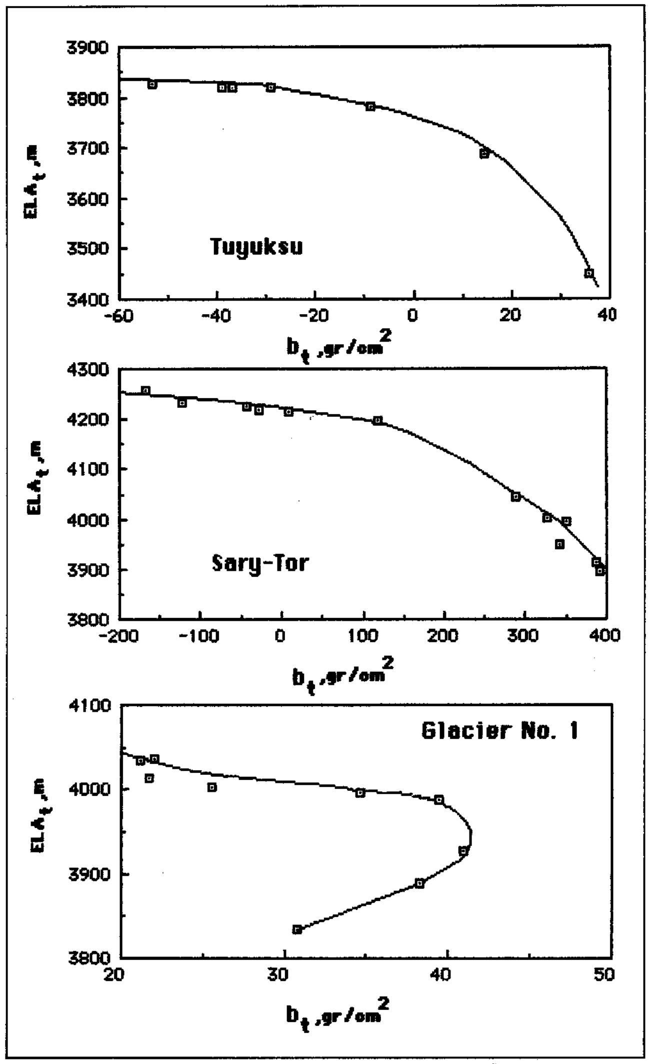


Fig 5 Glacier equilibrium-line altitude versus mass balance for 1989

climate eastward the Tien Shan mountains (Tab 5). The peak of the accumulation on the majority of the continental glaciers is on summer when the ablation is maximal (Ageta, Higuchi 1984). The altitudinal structure of the mass balance reflects all the peculiarities of the climatic conditions. In the continental conditions AS is closer to the multiplicative type and the curve $b_t(ELA_t)$ differs from the hypsographic curve, especially for the lower part of a glacier.

We can account the differences between transient and annual values of a mass balance for the differences in the altitudinal structure during an ablation period. The rate of the change of the vertical mass-balance gradient during the ablation period and the position of the upper boundary of the liquid runoff are the main elements of AS. The lower is the level of the liquid runoff, the closer to the multiplicative type is the altitudinal structure of mass balance and the slower is its response to the climatic changes. AS of the glacier system of the Tien Shan depends on two factors:

1) the distance from the windward periphery of the mountain massif which determines the continentality of the climatic conditions
2) the altitude asl of glaciers or altitudinal distribution of glacier area. This corresponds to M. Kuhn's assertion that the differences in the mass balance and the rate of reaction of the nearby glaciers depend on the differences in the glacier area distribution by altitude (Kuhn et al. 1985).

So the imitation model can be used for the mass balance monitoring of the continental glaciers. The substitution of $b_t(ELA_t)$ function for $b_n(ELA_n)$ is the most effective in the western part of the Tien Shan where the amount of the annual precipitation is larger. When the climate is more continental and the ablation goes on simultaneously with the accumulation, this substitution cannot be applied when the equilibrium line takes its extremely low position. This is the principal limitation of the model (Kunakhovitch 1991). In the Tien Shan the imitation model of the mass balance monitoring is correct for the glaciers with mass balance exceeding 15g/cm^2.

Conclusions

1. The multiplicative change of the vertical mass balance gradient db_n/dz is typical for the glaciation of the Tien Shan which is accounted for the presence of the upper boundary of the liquid runoff on a glacier. Thus, even a little change in accumulation alters db_n/dz that never occurs on alpine glaciers (Kuhn et al. 1985). The higher the location of a glacier and the larger its accumulation area above R_{up}, the greater the influence of the changes in summer accumulation on db_n/dz. As the summer accumulation prevails in the Tien Shan and its contribution to the mass balance grows eastward, the fluctuations of the summer accumulation are of the prime importance for the changes of annual and summer mass-balance gradients.

2. Monitoring based on the imitation model allows to obtain the whole range of the mass balance changes related to ELA during one ablation period. The function $b_n(ELA_n)$ can be replaced by $b_t(ELA_t)$ in the continental areas with the summer maximum of precipitation when the mass balance of a glacier is more than 159 g/sm^2. Above this level of the corresponding equilibrium line $b_t(ELA_t)$ can be extrapolated to the upper glacier boundary by the hypsographic curve. The linear approximation popular with many investigators (Palgov 1969; Braithwaite 1984; Glacier mass balance bulletin, 1991), brings to considerable mistakes.

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