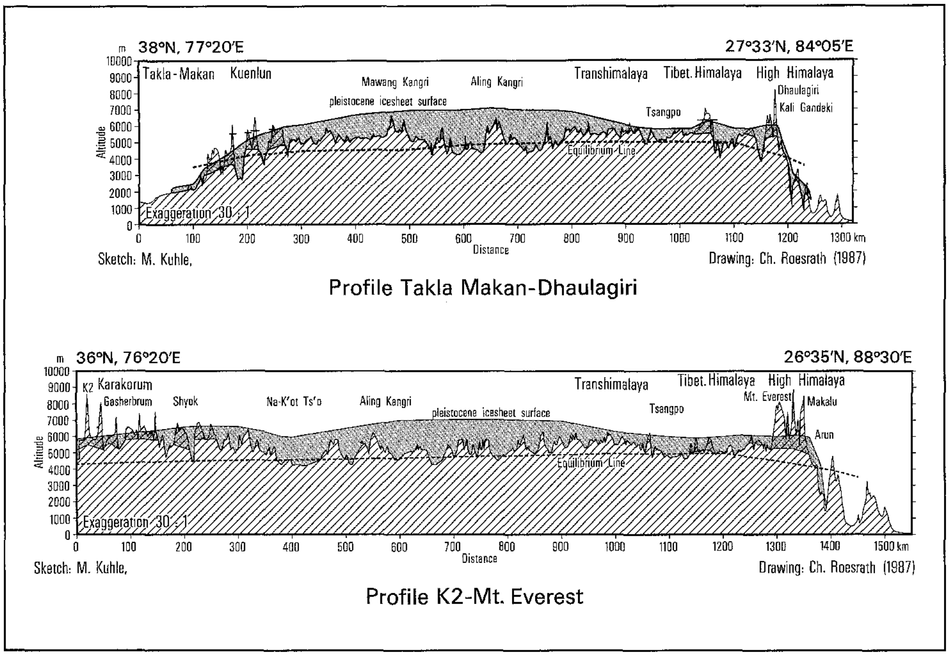

Fig 135 a/b
The two cross-sections give an idea of the Main Ice Age (Würm age) glacier cover of Tibet between a) the High Himalaya in the S and the Kuenlun in the N, down to the Takla Makan desert in the Tarim basin, and b) the High Himalaya and Karakorum, which is also in the north. This establishes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

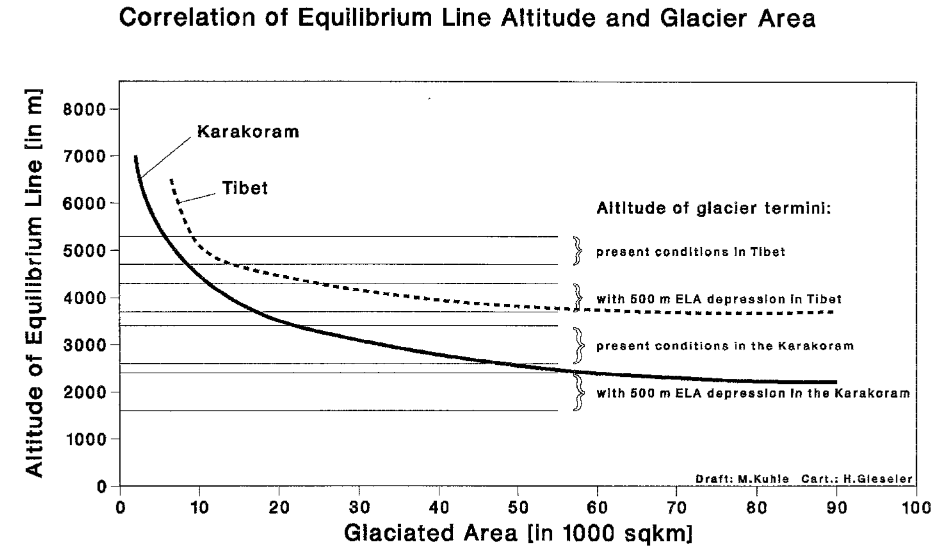

Fig 136
The diagram shows the increase in glacier areas in the particular area under investigation (in the Karakorum) and in the ice areas of the interior of High Tibet, which had been in contact with the prehistoric Karakorum ice stream network in the east. For this purpose a depression of the snow 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 1950)	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 (O). 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 O) 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 ○; 134 ○; 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, ○ left; 132 0). The exposure in Fig 133 shows typical glacial diamictites with *large* to *very large* (○) *polymict* blocks (limestone, granite, crystalline slate) "swimming" in a fine ground mass (matrix), thus *isolated* from one another. In some places 35% of the fistsize components are *aligned* to the NNE (group I = 22°), and more than 50% of them are standing upright (group IV). The other exposure (Fig 134) with large to very large *polymict* blocks (O) shows corresponding conditions. The intermediate mass is extremely *condensed* here, and is interrupted by, or interspersed with, somewhat *sorted* glacio-fluvial bands of eroded moraine material (↓). The decametre-high exposure near the valley exit (Fig 10) is representative for the orographic left-hand moraine slopes of the "valley of Pusha". The material has *almost all* the features of the sedimentological structures of moraine (cf. Woldstedt 1961, pp 27–31; v. Klebelsberg 1948, Vol. I, pp. 252–292; Schwarzbach 1974, pp. 30/31 Tab 6). Besides the only occasionally chaotic structure - moraine material is *slightly* stratified here, or at least *laid down in banks* - the *polymict* load of blocks of varying cubatures (■) occurs in the *immediate* environs, though *isolated* from each other by *loamy* to *fine sandy* ground mass. Apart from rather rough-edged, though *occasionally* rounded blocks, the *mixed* appearance of which is typically glacial, relatively many blocks are *facetted*, ie somewhat polished down on one or several sides. An essential feature, however, are the strata of *compressed* sand (X below) down to the typical *flexures* and *sand-silt nests* (slices) (X above), which have been squeezed out of their original strata formation. Once again, the matrix is *tightly packed* (2 g/cm^3) and has a markedly *higher* ramm resistance than unbaked mudflow material. This can be attributed to *compacting* by upthrust glacier ice. This *reduction* of pore volumes is an *internal* compacting which *reduces* the dimensions of glacial- tectonic flexures - they are joined by *shearing planes*, which run more or less *diagonally* to the direction of impact, as shown in the corresponding crumbling lines (↖) in Fig 10. They are *not only* diagonal, as in the case of grain-sized, homogenous folded clay stones, but the *moraine-specific*, internal structural irregularities have led to *shell-like, bent*, sickle-shaped shearing planes and corresponding crumbling breaking out from the exposure wall (↗). Concluding this analysis, attention should be drawn to the evident similarity with the moraine material in the zone in front of the present K2 glacier (cf. Fig 31). Or, from the other point of view: the *significant* difference as compared to the equally diamictic mur cone and mudflow material from the Shaksgam valley (Fig 77; 78 X) in comparison with the local material here it amounts to a *further* confirmation of these very extensive and deep *end moraine* findings in the northern Kuenlun foreland. The most significant differences concern the *lower density* of the mur material (1.8–1.9 g/cm^3) and the regular, *lenticular-shaped* cross-sections (X) of the particular consecutive sedimentation events. The *histograms* showing *grain sizes* on the other hand, resemble those of end moraines down to the last detail (cf. for instance Fig 55 No. 6 with 9), and the *morphometry* of the fine grains of murs is largely *dependent* upon the original rock (cf. for instance Fig 56, 20.8.86/1); this applies far less to local moraines, and not at all to the distant ones in question. In any case a genesis of murs or mud-flows, however, is out of the question thanks to the *overall* geomorphology of the foothills of the mountains and the up to more than 700 m-high extensions of the mountain valleys, both of which contain diamictites, which are the only ones to bear *glacio-geomorphological* characteristics. The settlements like Pusha and Tess, to name just two, are *irrigation oases* which are tied to *spring horizons* in the mountain forelands at the very place where the moraine valleys end and the deposited ground moraines, with their water-impounding boulder clay, come to the surface. Another particularly instructive index, since *unambiguously* glacio-geomorphological circumstantial evidence, are the *pressure and abrasion or polishing forms* in the loose material of the walls in the parallel strips of the foothill moraines. They are particular *exaration rills* which the subglacial moraine impregnating the glacier bottom has *ploughed* into the less inclined ground moraine (Fig 114 ↓↓). The *more detailed* analysis of forms in the moraines of the "valley of Pusha" and the medial moraine walls to the west of it is struck by *pressure grooves* (Fig 113 ⇩⇩) which post-glacial gullies have not, or scarcely been able to dissect. This must be attributed to greater *firmness of the moraines* near the surface thanks to *increased linear compaction* through polishing or abrasion pressure. But even in places where there was back-cutting erosion by gullies, *pressure or polish grooves* came out through the lines of the gullies, thus making them into wavy lines (↓↓) retracing the differences conditioned by *compaction in resistance*. Evidence of particularly intensive *moraine material induration* of this kind can be seen in the occasional *total suspension* of larger gullies towards their *lower end* (Fig 113 ↑↑), whilst most of them merely undergo a restriction, ie a *narrowing* of the gully in the area of that groove induration. *Horizontal striping* (at right angles to the fall-line) of the moraine slopes and the continuation of these *abrasion or polish and pressure lineaments* on the other side of large gullies generally also point to a temporal *sequence* of the two forms, ie the dissection of these approximately horizontal forms *after* deglaciation.

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

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

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

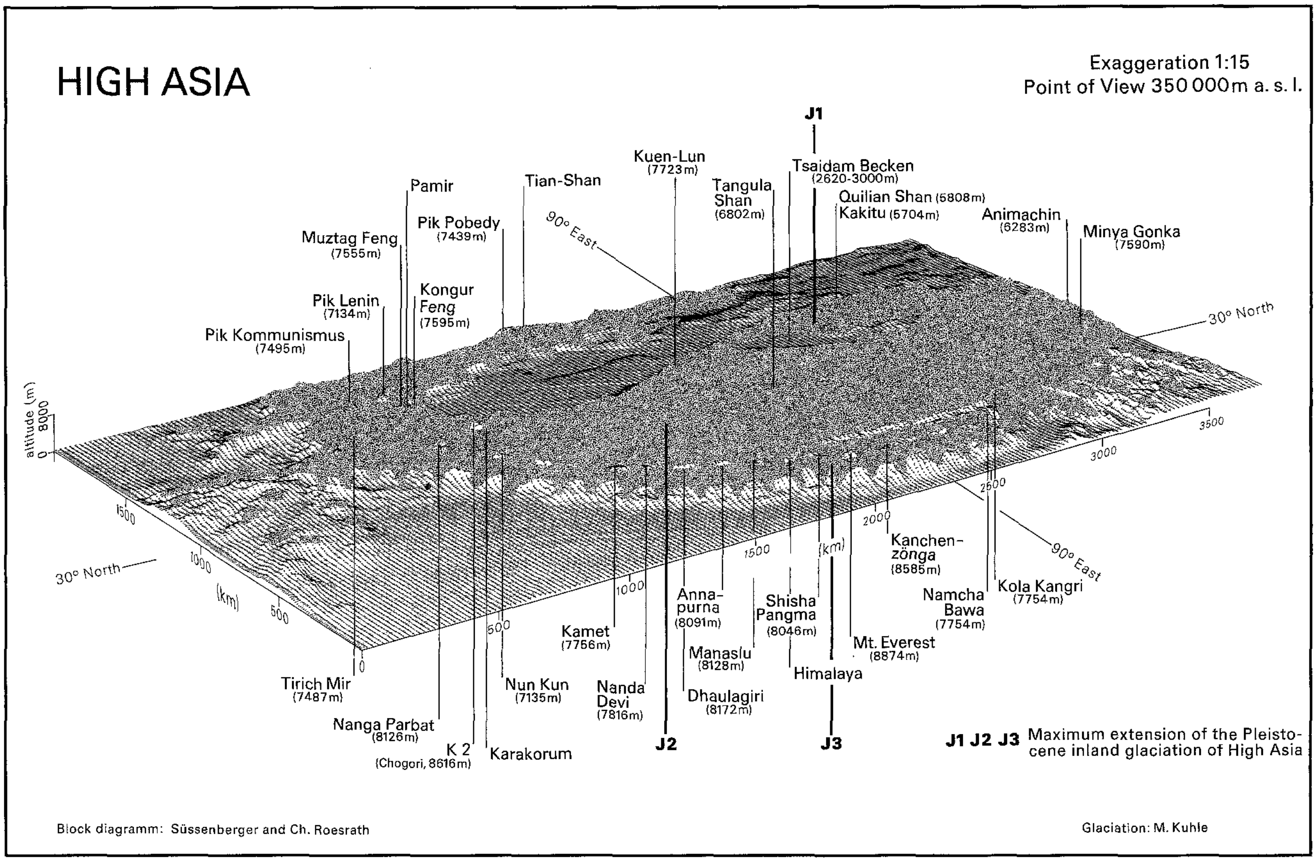

Fig 137 Towards the W the Würm period Tibetan ice merged with the ice stream network of the Karakorum (between Nun Kun and Tirich Mir), and formed a glacier system between Nanga Parbat and Kuenlun, which formed a bridge to the Pamir and even to the Tien Shan, in the NW.

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

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

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

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

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

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

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

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

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 2.4×10^6 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

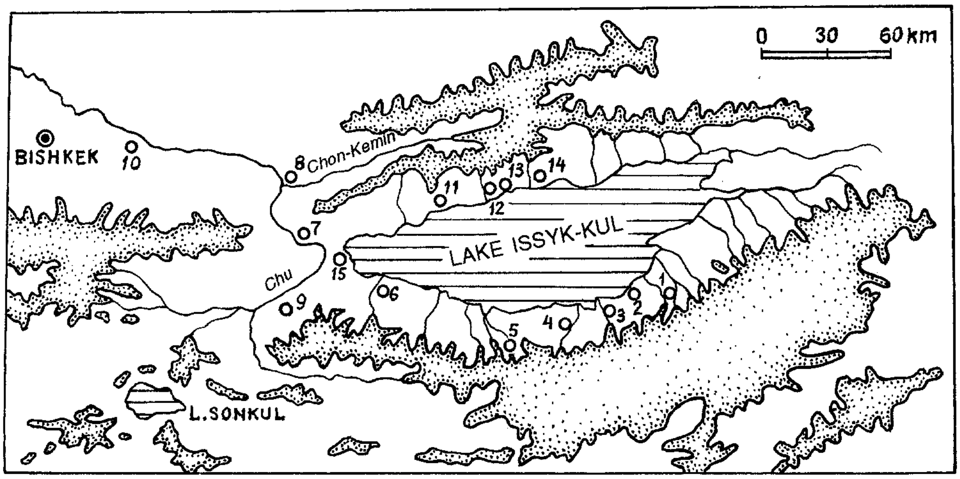


Fig 1
Würm glaciation of Lake Issyk-Kul area, Tian Shan Mts., after Bondarev and Pshenin (Atlas of Snow and Ice Resources [in press]), and the observation sites of the authors:
1 Dzhuku valley (val.)
2 Akterek settlement (stmt.)
3 Barskaun and Tamga vals.
4 Tosor val.
5 Kara-Ortok hill
6 Turasu val and 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

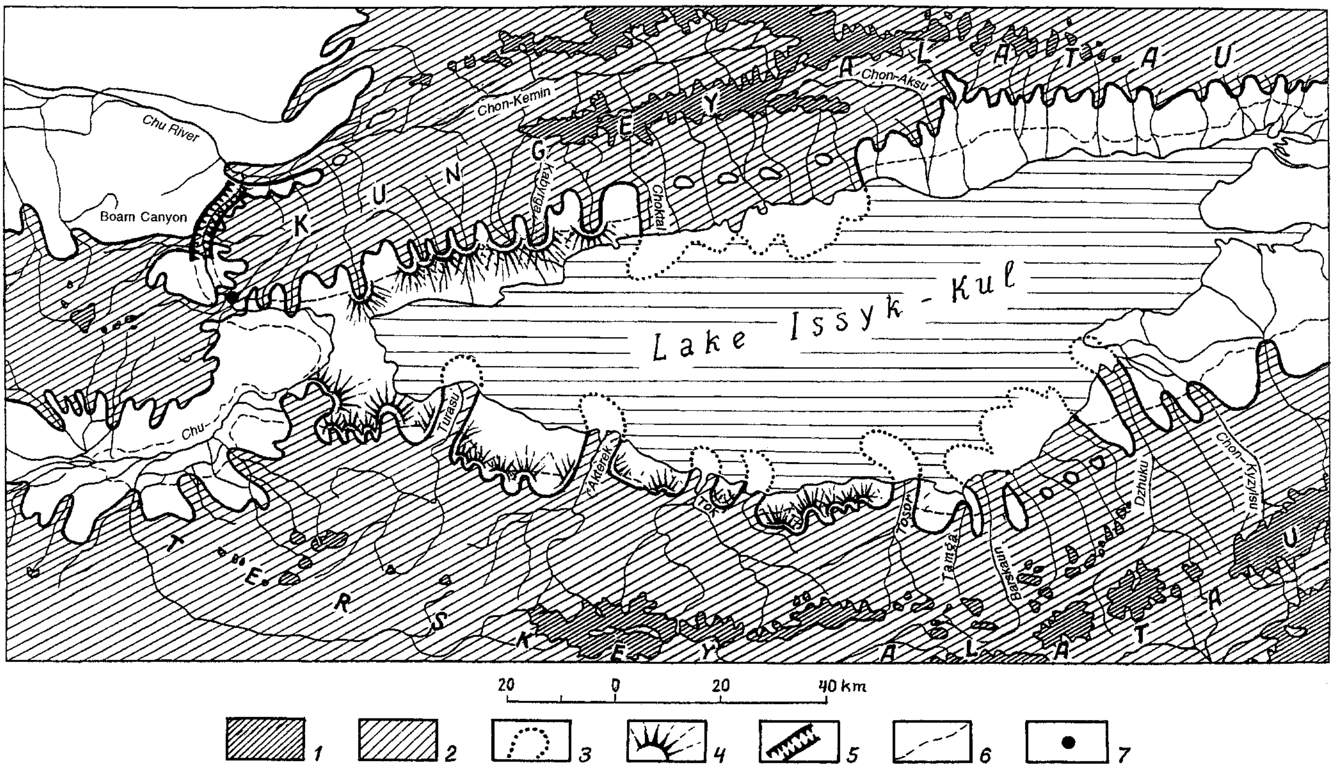


Fig 2 Würm glaciation of Lake Issyk-Kul area. Based on the authors' observations. Also published in: Grosswald et al. (1992). 1 present-day glaciers; 2 Würm ice covers; 3 floating ice lobes (tentative reconstruction); 4 ice marginal ramps; 5 erosional breach - Boam Canyon; 6 tentative limit of Issyk-Kul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

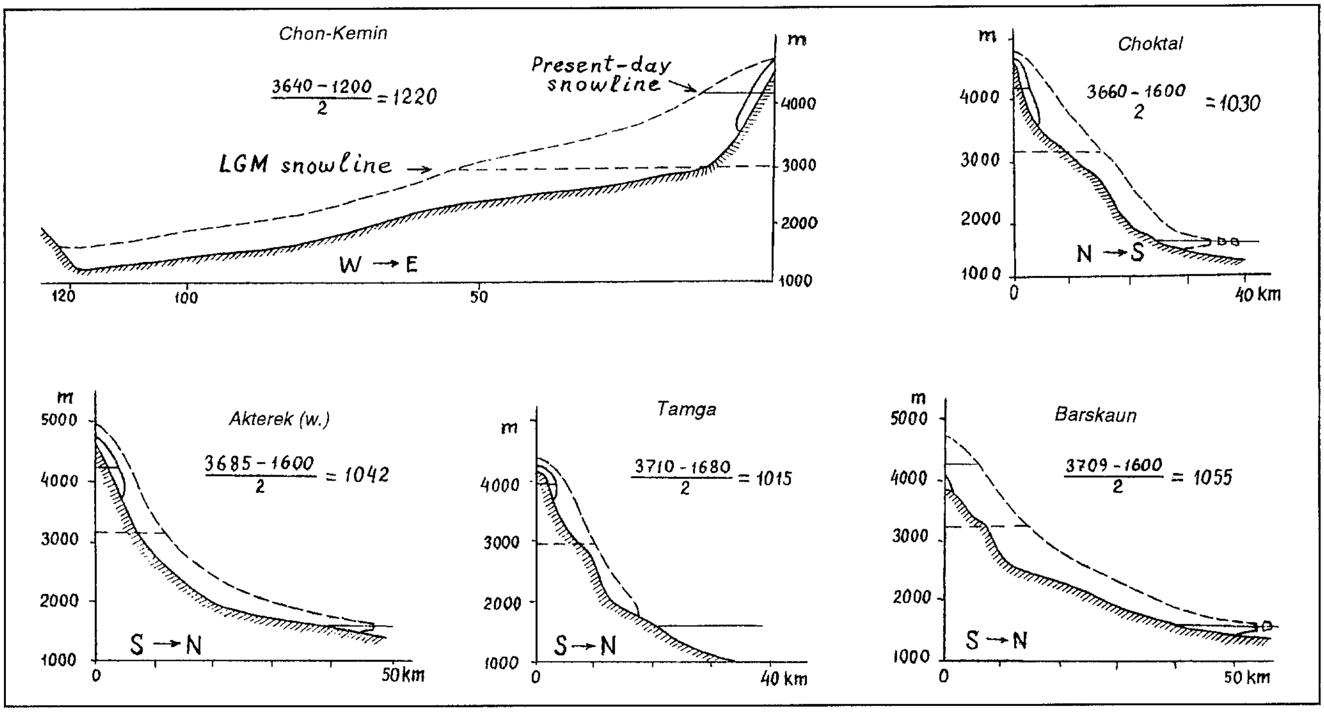


Fig 3 Lowering of snowline during the LGM in the several representative valleys of Lake Issyk-Kul area: Chon-Kemin, Choktal, Akterek (west), Tamga and Barskaun. Determined by the method of Höfer (1879)

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

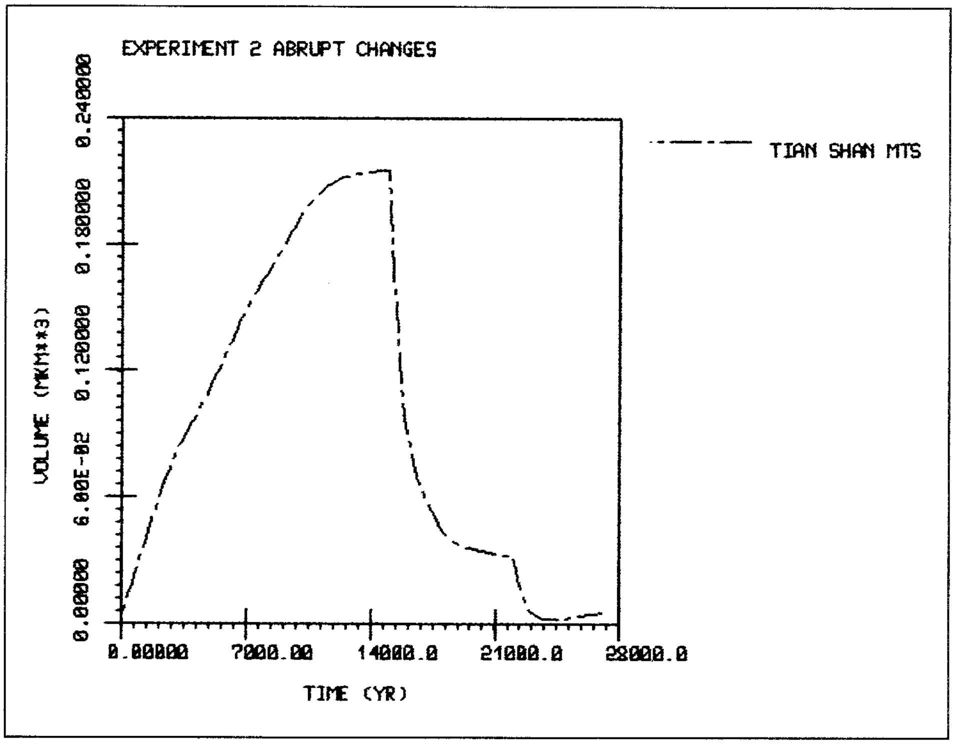


Fig 4
Volumetric growth and shrinkage of the Northern Tian Shan ice cover with time (modeling experiment 2: abrupt lowering and rise of snowline by 1200 m)

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.

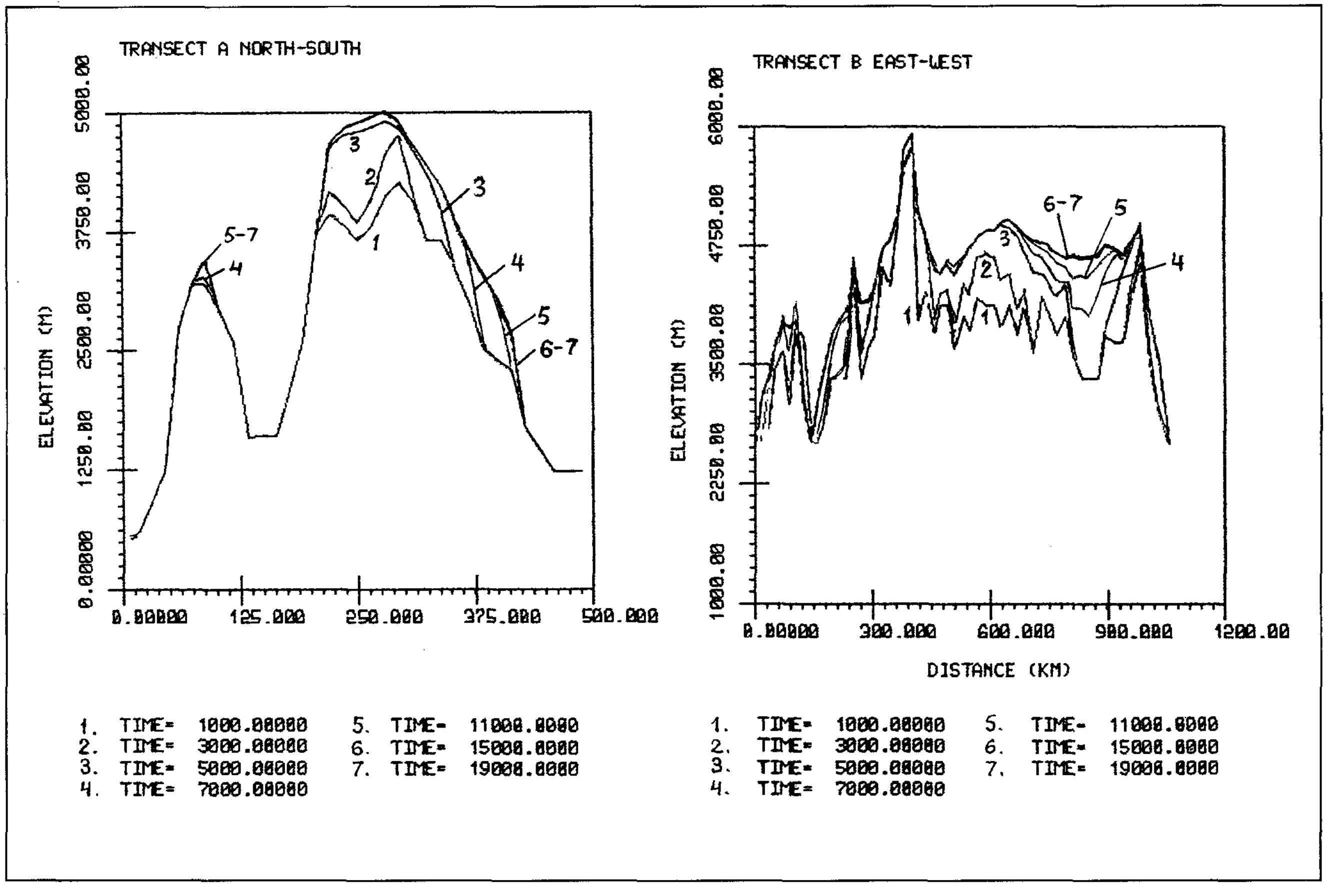


Fig 5 Ice thickness changes on the plateaus of northern Tian Shan with time (modeling experiment 1: gradual lowering of snowline – by 1200 m in 6 ka)

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

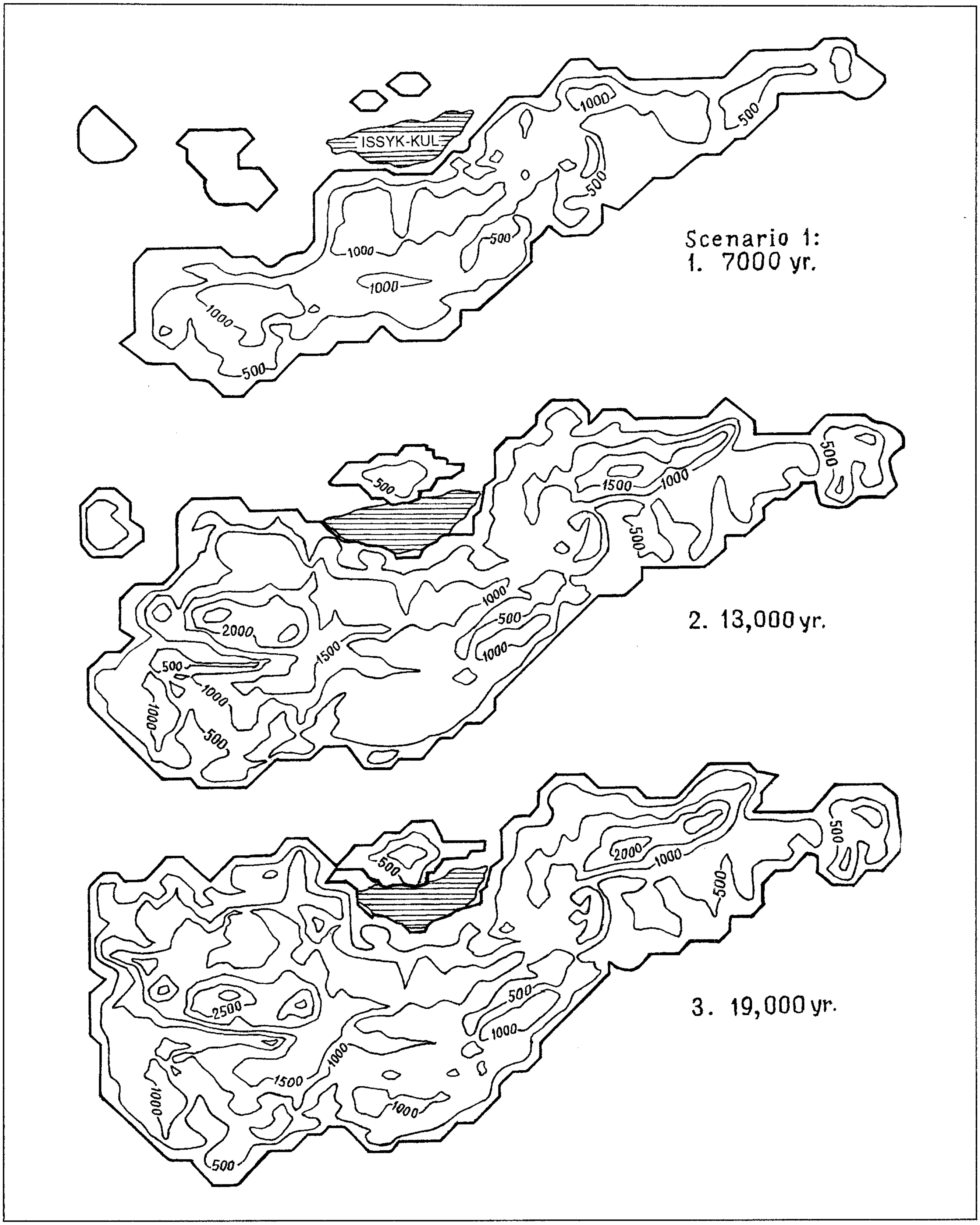


Fig 6 Changes in ground-plan configuration and thickness of the Northern Tian Shan ice cover with time (modeling experiment 1: gradual lowering of snowline - 1200 m in 6 ka). Note the rapid ice-cover formation at 7 ka (only 1 ka after full lowering of the snowline) and relatively small difference between 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.

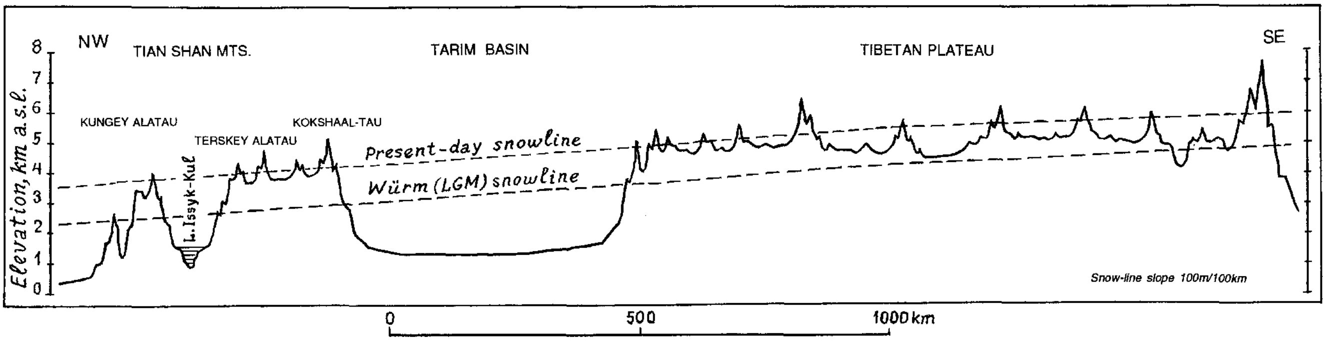


Fig 7 Generalized NW-SE profile across northern Tian Shan, Tarim Basin, Tibetan Plateau and Himalayas showing relationships between present-day and Würmian snowline, on the one hand, and the earth-surface topography, on the other.

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

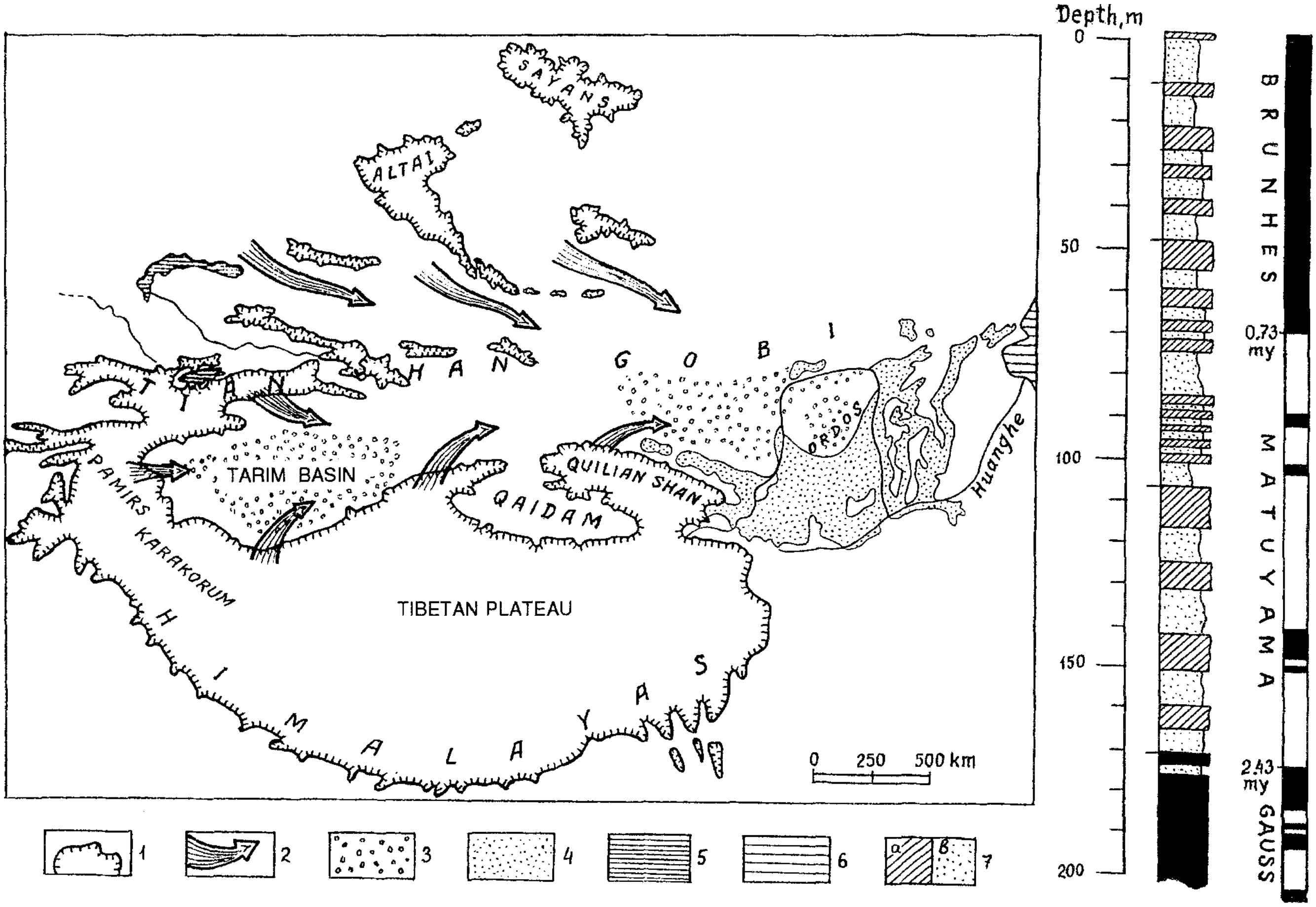


Fig 8 Ice covers, deserts and loess accumulation in Central and Eastern Asia during the last and older glaciations: the map shows directions of winds and their relationship with loess sources and major loess accumulation grounds. On the right: the Chinese loess stratigraphy after Kukla (1988).
1 ice 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

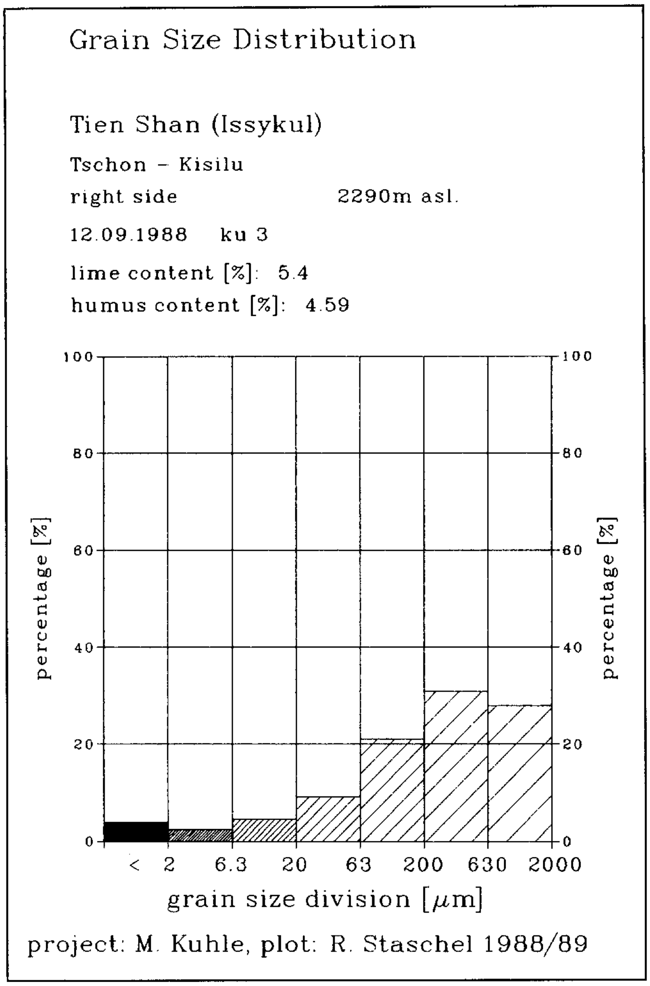


Fig 12 Analysis of the finely grained moraine substratum between the erratic granite blocks (Fig 11) on the orographic right-hand flank of the Chon Kyzylsu (Tschon Kisilsu). It occurs on bedrock phyllites as a fine matrix between these large moraine blocks. Taken from a depth of 0.20 m. The composition 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

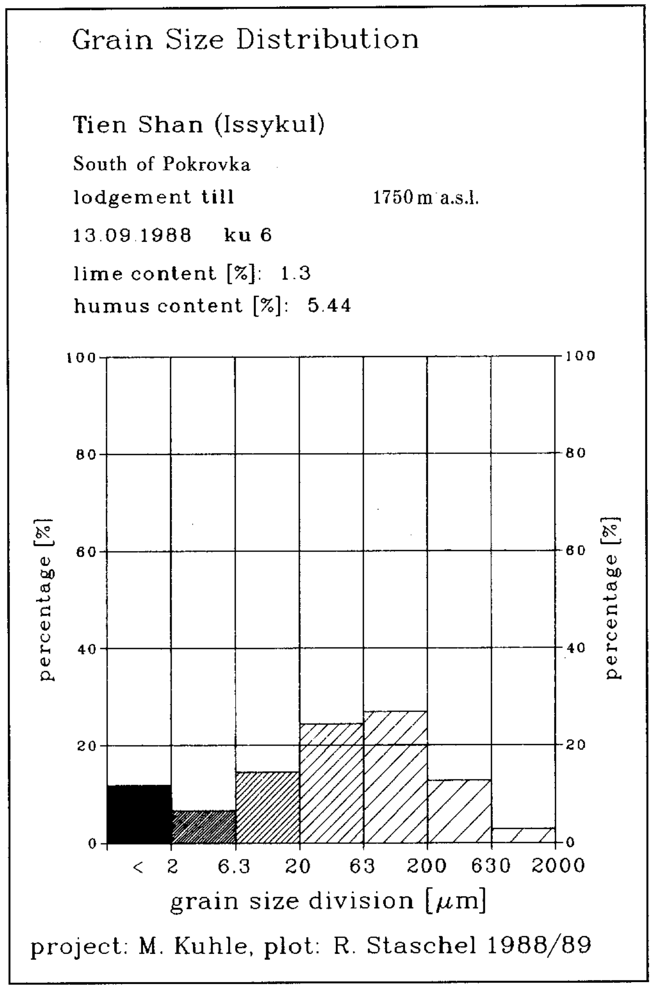


Fig 19 Locality of sample, cf. Fig 18 X. The bi-modal course of the grain-size graph, ie renewed rise to a peak of fine grain sizes, shows that this is typical ground moraine material (basal till).

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.

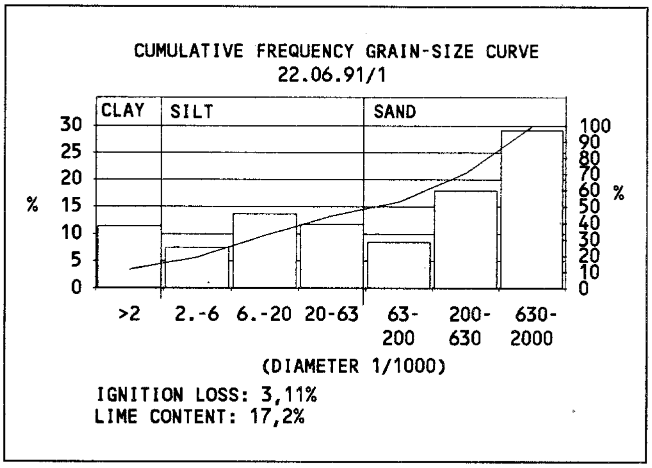
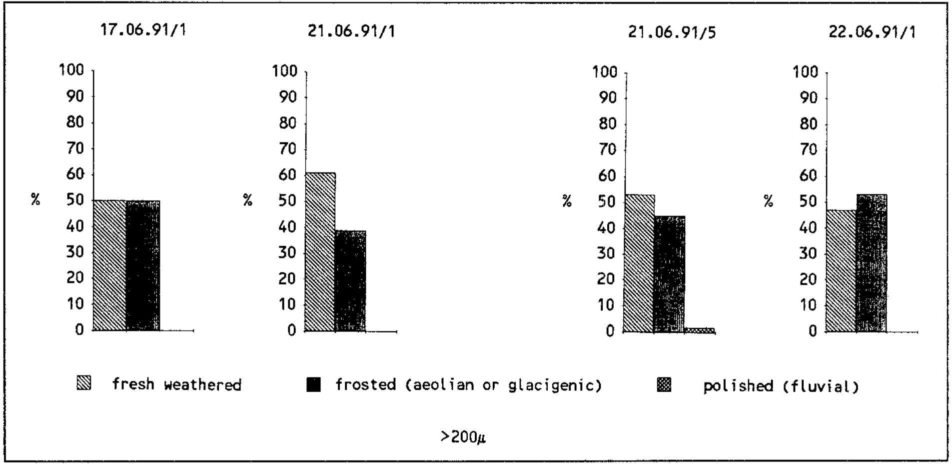


Fig 20 1650 m asl, grain size distribution in the boulder clay of a ground moraine, which reaches the south bank of Lake Issyk Kul between the ice-marginal ramps (IMRs) before disappearing below the lake level. This is matrix material from between large erratic blocks of granite which originated in the 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.



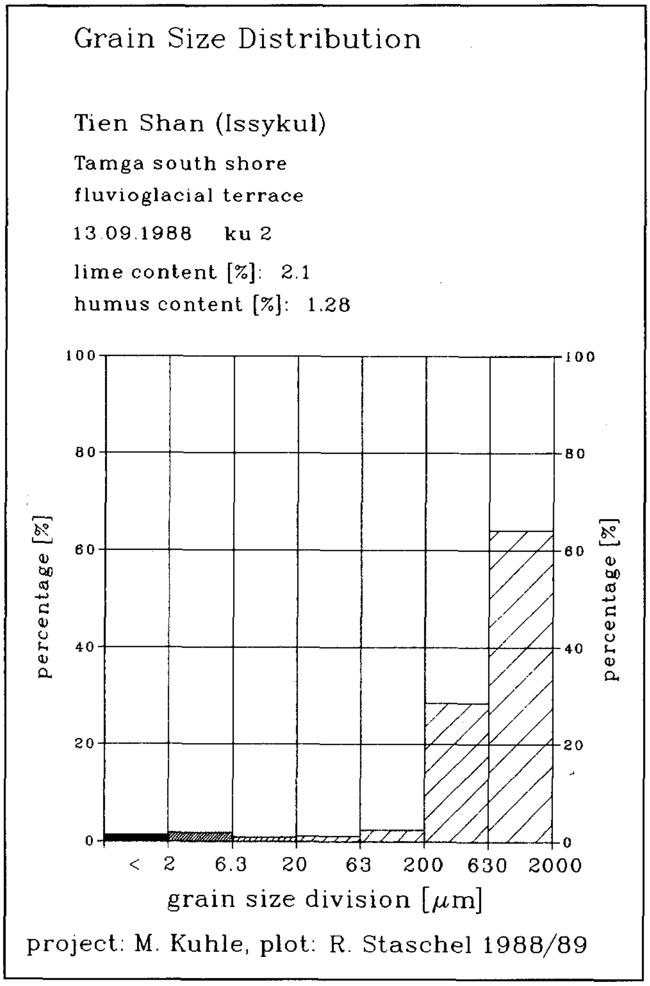


Fig 23 The grain size analysis shows the typical picture of Ice Age mountain foreland outwash plains (glacifluvial sediments) with their characteristic lack of fine material and the rapid increase in the proportion of sand to coarser material (630–2000 μm). Location of extraction: area of the mouth of 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.

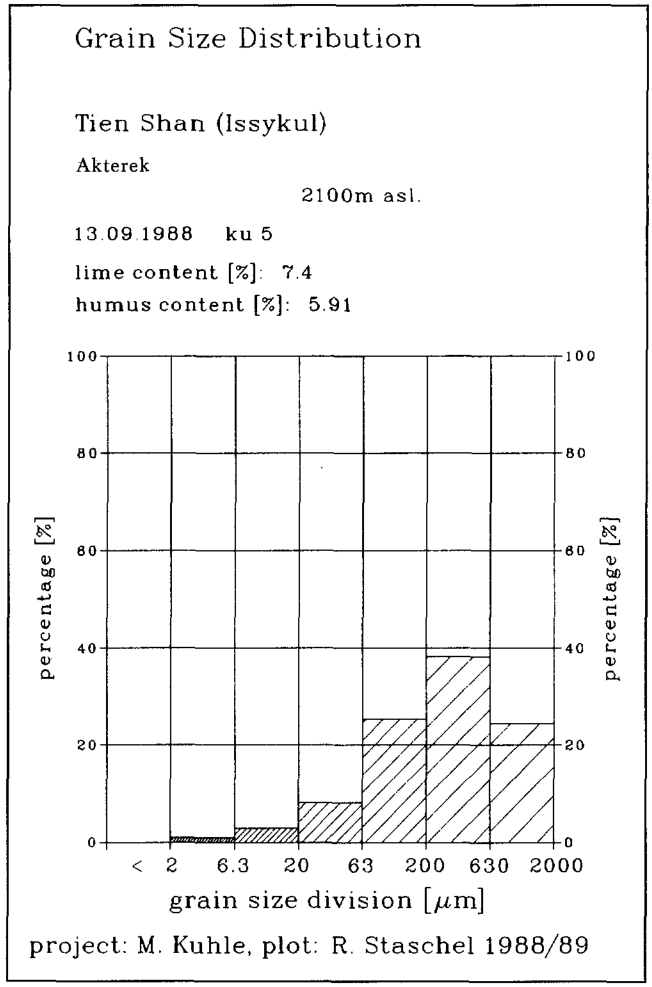


Fig 28 The composition of grain sizes is typical of the kind of outwash moraines occurring on the surface of ice marginal ramps behind the end moraine: the fine grain peak of the moraine is missing – having obviously been the first to be washed out fast. For locality of samples taken on the ice marginal 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

Fig 9 ▲
View from the south bank of Lake Issyk Kul at 1650 m asl facing SE towards the eastern Terskey-Alatau chain and up into the Chon-Kyzylsu valley (cf. Fig 1 on the right-hand side of No. 1). Numbers 1, 2 and 3 mark 5216 m-, 4845 m- and 4590 m-high massifs, which continue to be under glaciation. ∇ indicates persisting hanging glacier tongues. Several 100 m- high ice-marginal ramps (IMR) are evidence (■) of the main Last Ice Age piedmont glacier lobe which flowed out of this valley towards the NW and to within at least 13 km of the present edge of Lake Issyk-Kul (cf. Fig 2). Initially, at the height of the Würm Ice Age (0), the Chon-Kyzylsu outlet glacier bordered on (↙) the highest end moraine, or IMR level (■0), and found its surface undergoing a step-by-step melt-down during the Late Ice Age. This allows another four lower IMR levels (■■) to be identified, which descend in steps. The more recent, Late Glacial glacier positions within the Main Glacial location of the ice margin (■0) can be seen from the mountains down in the tongue basin (I, II, III) as shown in Fig 10 (■). —— shows the glacifluvially formed and relatively steeply inclined (7–9°) IMR surface, which falls away from the glacier and further into the mountain foreland. ● indicates the sweep of the oldest tongue basin, which is covered with more recent glacifluvial drift. ▼▼ mark post -Ice Age glacier rill washings and formation of small valleys. ---- indicate the Ice Age glacier level reconstructed on the basis of flank polishings, and thus the considerable glacier thickness. ↙ marks the position from which Fig 10 was taken, and indicates the locality of the erratica findings of Fig 11. Locality: 42°24'N/78°04'E; Photo M. Kuhle 10. 9. 88.

Fig 10 ▶
View from the orographic right-hand rock spur (consisting of phyllites) at the exit of the Chon-Kyzylsu valley at 2400 m asl, facing NW across the Late to Main Glacial piedmont moraines (■) towards Lake Issyk-Kul (O). (Fig 1 right-hand of No. 1). The orographic right-hand valley flank, which is in view here, was polished by the Chon-Kysylsu outlet glacier, during the last Main Ice Age (Würm or Waldai) when the terminal moraines marked ■ 0 were laid down by the piedmont glacier (cf. Fig 9). This glacial flank polishing is 350–400 m above the valley talweg (above the river on the right edge of the section) at 2160 m asl. The Main Ice Age piedmont glaciation of this valley spread like a hammerhead spit, so that it formed a foreland ice together with similarly extended foreland ice masses of the adjacent parallel valleys (cf. Fig 2). However, this applied only to the Main Ice Age. During the Late Glacial stages I (Ghasa stage), II (Taglung stage) and III (Dhampu stage) the tongues of the outlet glaciers not only became shorter (ie failed to extend so far out into the foreland), but also steadily narrowed (see sequence ■I, ■II, III). Lateral moraines contained them on their sides (■). Glacifluvial kames (X) have been piled up against the lateral moraines of Stage I. ↓ mark Holocene runnel washings in the moraine material; ▼ shows the position of erratic moraine material (cf. Fig 12, sample 12. 9. 88/3 and Fig 11). Evidence of large erratic blocks of granite occurs in the moraines ■I up to levels of 2300–2400 m asl, and indeed up to the moraine ridge. These moraines are deposited upon crystalline slates. The Main Ice Age moraines (■ 0) rest on more or less disordered neogenic and older sedimentary rock. Locality: 42°17'N/78°08'30"E; Photo: M. Kuhle 12. 9. 88.

Fig 11 ▼
View from the orographic right-hand valley flank of the Chon Kyzylsu (or Tschon Kisilu) at 2290 m asl, facing S up into the valley (Fig 1 right-hand of No. 1). In the foreground medium-sized to very large light-coloured erratic blocks of granite (●) have been deposited by this Late Glacial outlet glacier of the Tian Shan ice-cap on the glacigenically polished rock flank, which consists of greenish, metamorphic sedimentary rock (phyllites) (cf. Fig 12). The largest of these blocks is 5 m long, 4 m wide and 3 m deep (for comparison of size see the 1.75 m tall person). Corresponding to these erratica deposits, the Late Glacial lateral moraine terrace (□ III) is visible as a wooded, flattened area on the opposite side of the valley on the orographic left-hand side (cf. Fig II, III). The erratica lie 130–150 m above the present drift floor of the valley. The Main Ice Age level of the ice stream network (- - -), which can be reconstructed on the basis of the increasing absence of glacigenic flank polishings, runs above the timber-line (2850–3050 m asl), and thus c. 1000 m above the valley floor. Far down (□IV), another and more recent Late Glacial ledge of lateral moraines is visible (Sirkung stage). No. 2 = 4000 m-high peak with satellite rocks; No. 1 = 3985 m - high peak; both these mountains are ice-covered. Locality: 42°17'N/78°08'28"E; Photo: M. Kuhle 12. 9. 88.

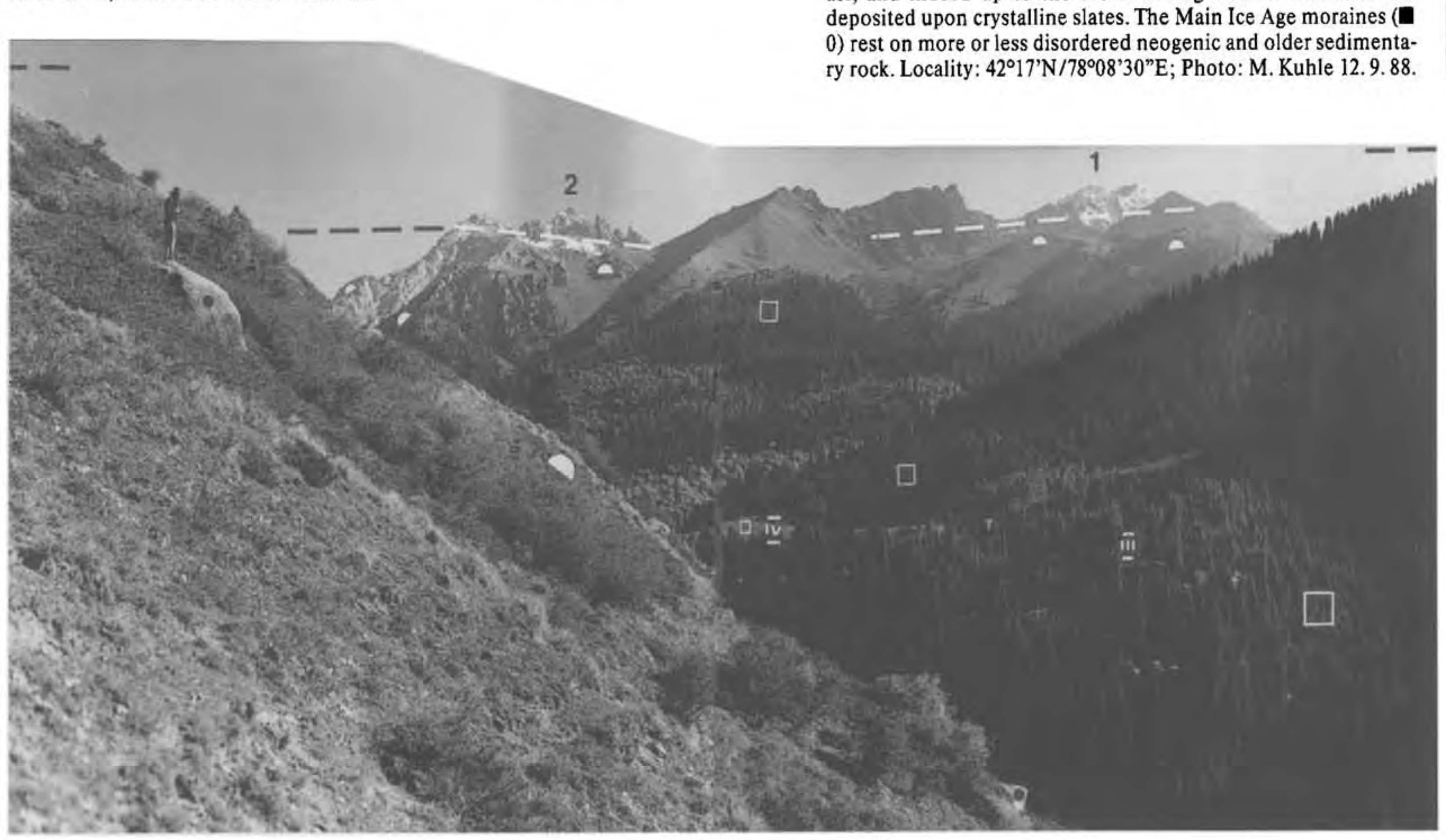


Fig 13 ▶
View from the valley bottom of the upper Chon Kyzylsu at 2700 m asl towards the S, looking into its valley-head with the confluence area of three headwaters on the NNE slope of the Terskey Alatau (Fig 1 on the right, below No. 1) (cf. Fig 14–17). As the evidence of flank polishings on the bedrock gneisses and metamorphites show, the main valley and its tributaries had been filled with glacier ice up to c. 4000 m (----). At least two different ice-scour limits, ie prehistoric glacier levels, can be discerned: the lower one (---- below) is definitely from the Late Glacial period, but even the upper ice-scour limits marked here (---- above) are likely to belong to the early Late Glacial time. The highest, ie Main Glacial, ice stream level cannot be reconstructed here. □ mark talus cones from post-Late Glacial rock falls from the glacially polished valley flanks. ∇ indicates a mudflow cone on the valley floor. No. 1 = 4590 m peak, still with substantial ice superstructure; No. 2 = its c. 4300 m-high satellite N-peak. Locality: 42°08'N/78°12'30"E; Photo: M. Kuhle 11. 9. 88.

Fig 14 ▲
View from 3150 m to the E into the orographic right-hand flank of the Chon Kyzylsu main valley in the confluence area of the tributaries near its valley head (Fig. 1 right-hand side, below No. 1), where bands of ice-scouring occur up to the very ridges (☗). Glacially polished to great altitudes, the cuspate-shaped mountain flanks have undergone periglacially concordant transformation. ---- marks the minimum height of the Ice Age glacier level at 3700 m asl. At the bottom of the trough-shaped tributary valley, Late to Neo-Glacial dumped end moraines are found (□). The flattened parts of slopes in the main valley must be regarded as Late-Glacial lateral moraine ledges (■). Locality: 42°07'30"N/78°12'E (cf. Fig. 15). Photo: M. Kuhle 11. 9. 88.

Fig 15 ▲
View from 3200 m asl towards SW, facing up into the orographic left-hand headwaters (Chon Aschuto) of the Chon Kyzylsu (Fig 1 on the right below No. 1). This is a hanging valley, the bottom of which is joined to the main valley by a c. 400 m-high confluence step. Flowing down from the Tian Shan plateau, one of the numerous outlet glaciers of the last Ice Age inland ice caps transformed the valley into a classic trough valley. After the late Glacial deglaciation only smallish talus cones (∇) were formed below the glacial flank polishings (◥☗). ---- marks a probably Late Glacial glacier level at 3700–3600 m asl. Due to the absence of unambiguously preserved ice-scour limits, there is no evidence of higher ice levels. No. 1 = c. 4000–4200 m - high glaciated secondary peaks of a northern Terskey Alatau secondary ridge. Locality: 42°07'29"N/87°11'59"E (cf. Fig 16); Photo: M. Kuhle 11. 9. 88.

Fig 16 ▲
View from 3650 m asl, from the tongue of the Aschuto glacier (X) which is still under snow, towards the NNE down the Chon Aschuto valley (Fig 1 on the right, below No. 1). No. 1 = 4000–4200 m-high, only just glaciated northern peaks of the eastern Terskey Alatau chain. This trough valley has been scraped out (cf. Fig 15) by a northern outlet glacier of the Tian Shan ice cap (Fig 2) which flowed across the Aschuto transfluence pass (cf. Fig 17). The well preserved Late Glacial flank polishings (◥◤), with an ice scour limit at 4000–3750 m asl in the bedrock metamorphic rocks are evidence of a prehistoric glacier level (----). The main Ice Age glacier level was noticeably higher. ∇ marks post-Glacial talus cones. □ (small sign further back) points to an historic ice margin location (Little Ice Age?) at 3300 m asl. The next younger ice margin location with a dumped end moraine, which is still likely to be holding some dead ice (□ larger sign more to the front) lies at 3400 m asl. The present glacier tongue ends at 3480 m asl, and is covered by debris (not visible, as it is in the blind spot of the exposure). Locality: 42°05'N/78°12'20"E; Photo: M. Kuhle 11. 9. 88.

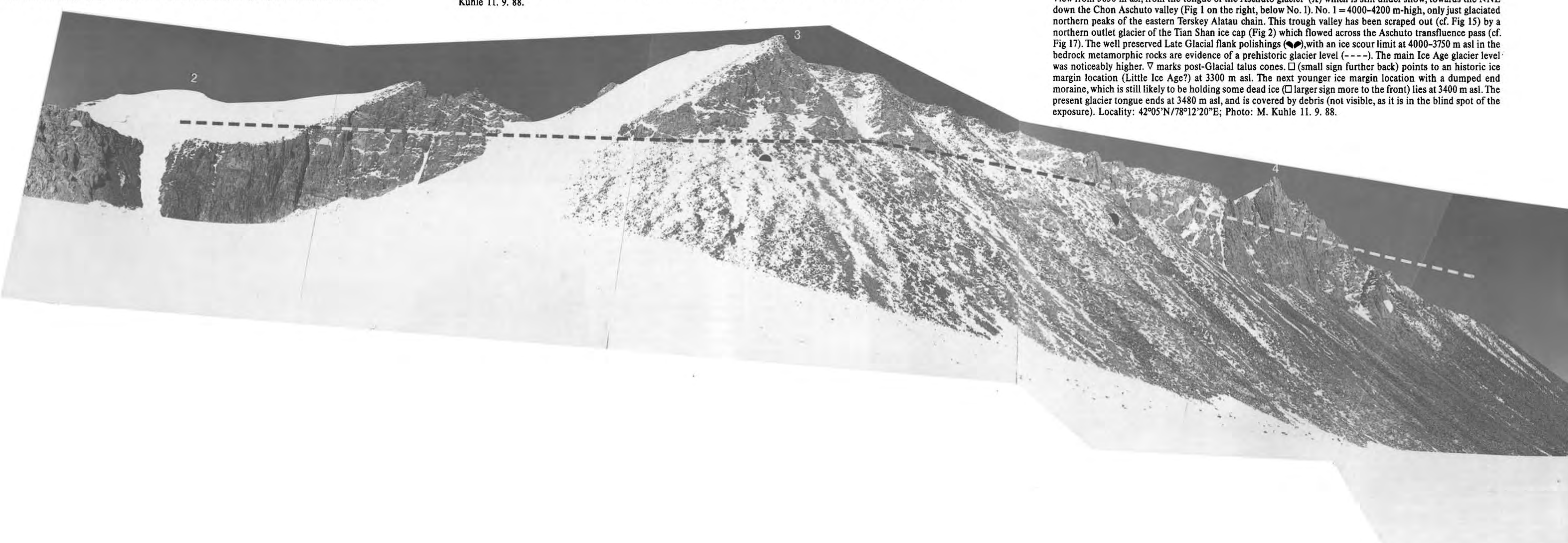


Fig 18 ▲
View from the southern edge of the town of Pakrovka near the south bank of Lake Issyk Kul in the area of the valley exit of Chon Kyzylsu (♦) in 1700 m asl towards the SE to three Main (?) to Late Glacial ice margin indicators (■) (Fig 1 right-hand of No. 1). The tongue of the Chon Kyzylsu outlet glacier emerged from the Terskey Alatau mountains and entered the mountain foreland (♦). Here, outside the hitherto lowest end moraines (cf. Fig 10) a renewed advance led to hammerhead-shaped spread. The three kame terrace-like moraine levels (■ or IMR levels) were built up during this process. In the three levels, which are laid down here, their edges bordered on the same number of glacier surface levels. The extreme steepness of the edge gradients (towards the right) is evidence that these were deposited along the edge of a glacier or against it (remnants of alluvial fans would have a much gentler slope). The glacier tongue may have reached Lake Issyk Kul (though persisting uncertainty on this point argued against an entry down to Lake Issyk Kul in Fig 2). X marks the locality of the sample described in Fig 19. Locality: 42h19'N/78°00'E; Photo: M. Kuhle 10. 9. 88.

Fig 17 ▼
View from the Ice Age transfluence Pass (Aschuto Pass, No. 11) from 4100 m asl across the Terskey Alatau main ridge (Nos. 1-4), which formed the NNW boundary to the ice cap cover of the Tian Shan plateau (Fig 1 on the right, below No. 1; cf. Fig 2 in the right-hand corner). The outlet glacier of the Chon Kyzylsu flowed over this pass. Some of the ice persists to this day (cf. Fig 16X), though only locally, over some kilometres to the north and south. This panoramic view has been taken from NE (Nos 2 and 3) to SE to SW (No. 1), showing the central Tian Shan plateau and above it, culminating at 5125 m asl, the Ak Shiyrak mountains (Nos. 5-10). The massif (Nos 2-4) is 4590 m-high (the main peak is out of view). The visible peaks at the Ak Shiyrak massif (Nos 5-10), 20 km away, are 4500-5100 m high. This massif is the presently most heavily glaciated area of the central Tian Shan plateau (glacier area c. 400 km^2), and has five up to 15 km - long valley glaciers (Petrov glacier, c. 60 km^2) (▼). Peak No. 1 is c. 4800 m high. The Tian Shan plateau (X) was covered by a c. 1000 m - thick ice cap. This minimum thickness has been reconstructed on the basis of glacier polishings of mountain spurs (⪆). During the main Ice Age the ice level there (- - - -) was at about 4500-4700 m, asl. The Tian Shan plateau is covered by thick ground moraine (X). The transfluence thickness of the Chon Kyzylsu outlet glacier above this pass was at least 300-400 m, so that its surface level reached 4400-4500 m asl (- - - -). Some glacial flank polishings and smoothings have been preserved in the coarse crystalline rocks of some small areas to this day (◥▲). Locality: 42°03'20"N/78°09'E; Photo: M. Kuhle 11. 9. 88.

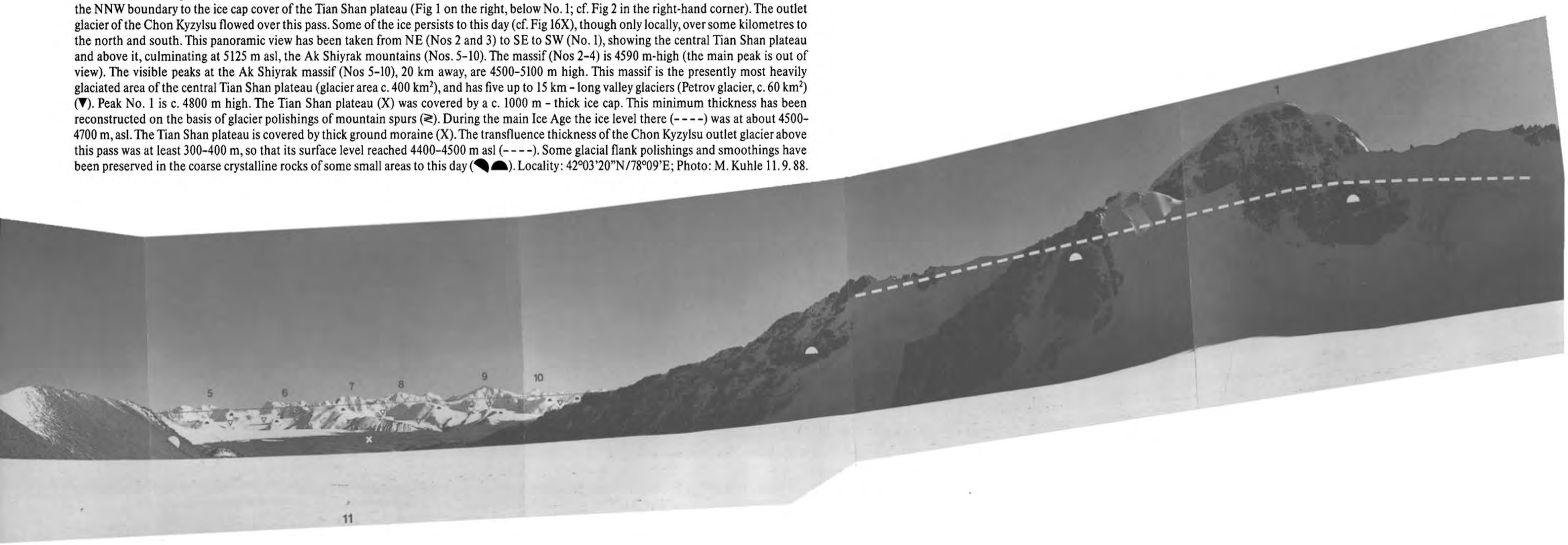


Fig 22 ▲
View from the northern foreland of the Terskey Alatau near the S shore of Lake Issyk Kul at 1630 m asl towards the S to the exit of the Barskaun valley (Fig 1, No. 3). Nos 1–3 = 4200–4550 m-high peaks of the northern Terksey Alatau chain. - - - marks the surface of the Ice Age Barskaun outlet glacier of the Tian Shan ice cap (Fig 2), which can be reconstructed on the basis of glacier polishings ( ). Lateral and end moraines in the highest locations (■O) belong to the Main Ice Age piedmont glacier, which calved into Lake Issyk Kul. ■I and II belong to the Late Ice Age (Ghasa and Taglung stages). Even at that time the glacier tongue spread like a hammerhead into the foreland to c. 1650 m asl. The Main Ice Age moraines (■0) are of the ice marginal ramp (IMR) type, while the Late Glacial moraines (■I and II) leant against the inner slopes of the ice marginal ramps in the form of narrow lateral moriane ledges. ● indicates a more recent (than I and II) Late Glacial drift floor, which had been laid down by a glacier cave that already receded far back into the valley (cf. Fig 23). O marks corries and hanging valley heads. Locality: 42°09'50"N/77°39'E; Photo: M. Kuhle 13. 9. 88.

Fig 25 ▲
View from 10 km S of the shoreline of Lake Issyk Kul at 1800 m asl towards the SE into the Tok valley and over to the Terskey Alatau chain (Fig 1 No. 5). No. 1 = main or western subsidiary peak of the 4509 m massif; No. 2 = 4023 m peak on the E side in the 4478 m-high massif. Above the late Late Glacial drift floor (O = sander), the Late Ice Age lateral moraines (■) of the Taglung stage (II) and the Ghasa stage (I) are to be differentiated. mark the Main Glacial flank polishings. The corresponding piedmont glacier calved into Lake Issyk Kul (cf. Fig 2). III indicates the location of the ice margin of the Late Glacial Dhampu stage (III), which remains within the valley. Locality: 42°07'15"N/76°56'E; Photo: M. Kuhle 13. 9. 88.

Fig 24 ▲
View from a Late Glacial drift floor (outwash plain) surface (●) near the south bank of Lake Issyk Kul at 1640 m asl towards the SSW (No. 3) up the Tosor valley (Fig 2; Fig 1 No. 4) No. 1 = 4672 m-high peak; No. 2 = c. 4400 m-high western satellite peak; No. 3 = eastern satellite peak of the 4509 m-high massif, all of which belong to the Terskey Alatau chain and continue to be ice-covered. mark the Ice Age polishings on rock ridges and flanks up to the ice scour limits (- - -). ■0 refer to the main Ice Age piedmont moraines and IMRs of the same period. ■I are the corresponding Early to Late Glacial forms (Ghasa stage). The former extend as far as into Lake Issyk Kul. ▼ indicate talus cone-like, Late Glacial glaciofluvial drift floors. Locality: 42°10'N/77°25'E. Photo: M. Kuhle 13. 9. 88.

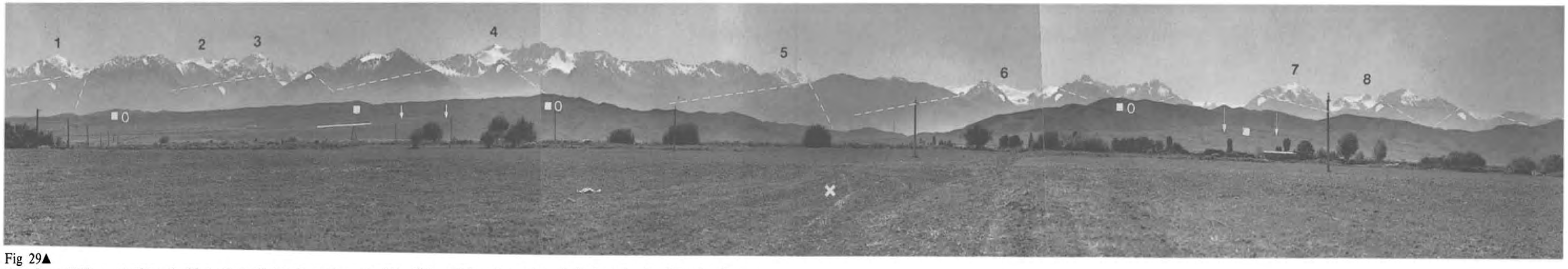


Fig 29▲
View from 1800 m asl, 10 km S of Lake Issyk Kul to the outer wall of the 300 m-high end moraines. Laid down by the Akterek valley piedmont glacier, they are of the ice marginal ramp (IMR⟋■0) type (Fig 1 between Nos 6 and 5; Fig 2). Panorama taken form SE (No. 1 = 4478 m-high massif) via SSE (Nos 2 and 3 = 4763 m-high massif, the highest feeder area of the immediate Akterek outlet glacier), S (No. 4 = W satellite of the 4763 m-high massif), SSW and SW (Nos 5 and 6 = 4159 to 4200 m-high peaks) to approximately WSW (Nos 7 and 8 = c. 4250–4300 m-high peak). - - - - mark the levels of Ice Age valley glaciers of the Terskey Alatau chain, which entered the foreland. The preserved glacial flank polishings (◣◢) extend to these altitudes. The central outlet glacier tongue found its way through the gap between the two main glacial ice marginal ramps (between ■ 0 on the left and ■ 0 on the right) and down to Lake Issyk Kul, into which it calved (cf. Fig 2). ↓↓ point to runnels in the IMR outer slopes. They are the result of meltwater which had reached over the upper edge of the moraine. The outer slopes of the IMR (——) have a 7°–15° incline. X marks drift-covered ground moraine. Locality: 42°11'50"N/76°41'10"E; Photo: M. Kuhle 13. 9. 88.

Fig 30 ▲
View from the N bank of Lake Issyk Kul at 1615 m asl towards the NE across a mountain foreland on to the Kungey Alatau mountains (Fig 1 No. 11). This is the area of the Chok valley mouth, where the main glacial ground moraine (X 0), with its large granite blocks which are "swimming" (●) in a fine matrix, reaches the lake with a 42 km-wide front (cf. Fig 31). This was the W edge of the widest glacier calving front into Lake Issyk Kul (Fig 2). The oldest Late Glacial moraines (■ I = Ghasa stage) are to be found in the immediate valley exit. All the mountain ridges which are visible from here were polished by the glacier as it entered the foreland (◣◢). Locality: 42°36'50"N/76°46'30"E; Photo: M. Kuhle 10. 9. 88.

Fig 34 ▶
View from an end moraine ridge (■ 0) in the area of the same ice margin location as Fig 33 (Fig 1 No. 12) at 1770 m asl, facing north as far as the S-foot of the mountain range, when the glacially polished, bedrock rock is visible (◢). Further down a several hundred metre-thick mass of moraine material (■ I = Ghasa stage) has been pushed up against it. The end moraines of the Main Ice Age (■ O) are separated from them by a several km-wide tongue basin (◇). Here they form an island of median moraines (Fig 2, on the right-hand side of Choktal) between the two glacier tongue ends which calved into Lake Issyk Kul. The largest blocks (● granite) have been left in the highest positions on the moraine ridge – a typical feature in the construction of end moraine walls. The surface of some of these blocks shows signs of weathering and exfoliation (desquamation) (● right-hand side). Locality: 42°39'59"N/77°11'59"E; Photo: M. Kuhle 10. 9. 88.

Fig 31 ▲
View from a place adjacent to the one used for taking the photograph for Fig 30 at 1615 m asl, facing SE towards Lake Issyk Kul (O) (Fig 1, No. 11). The Main Ice Age ground moraine (basal till) (X 0) with its large granite blocks (●) is submerged beneath the surface of the lake waters (0). Locality: 42°36'48"N/76°46'31"E; Photo: M. Kuhle 10. 9. 88.

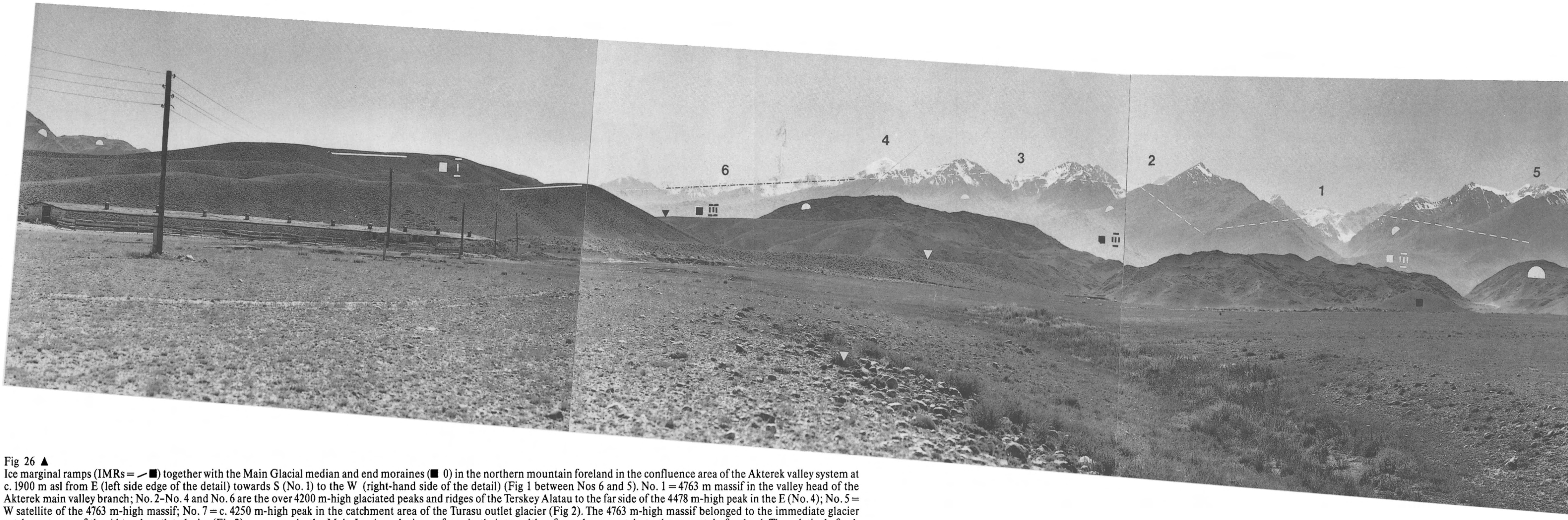
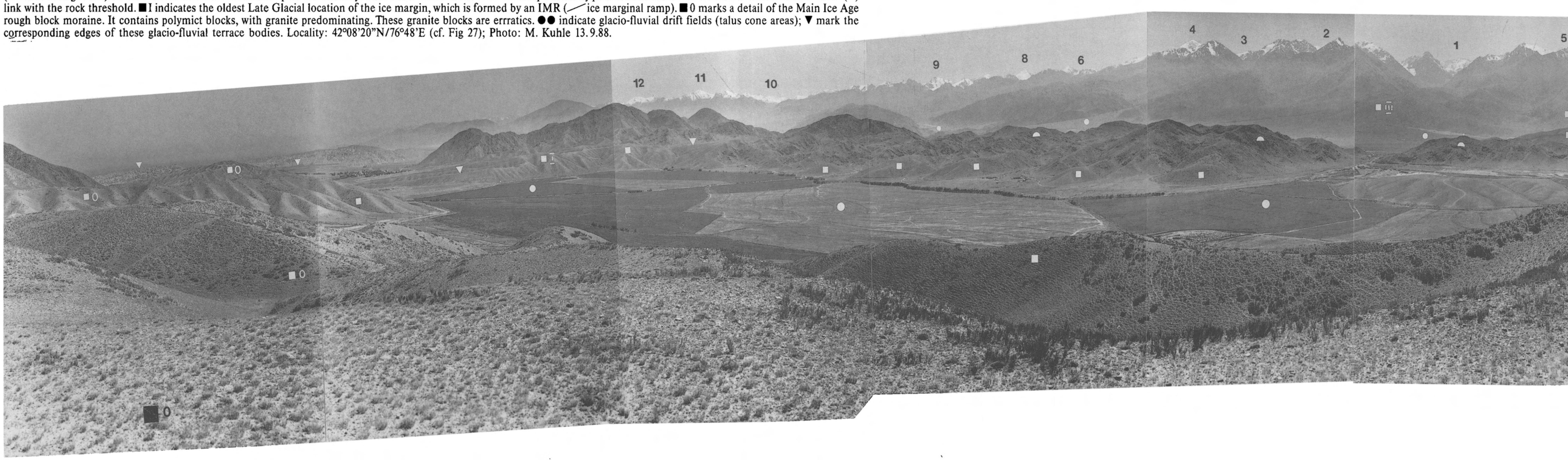
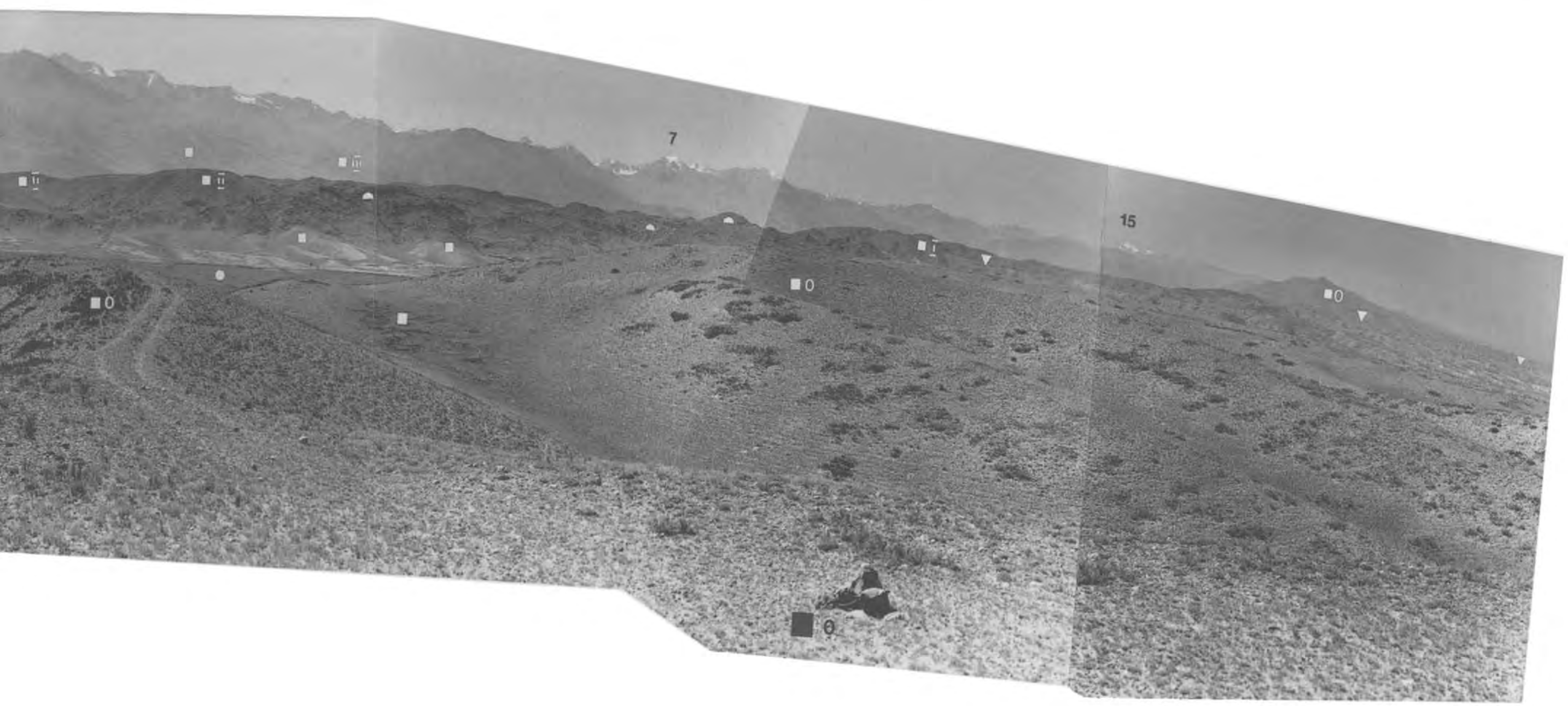


Fig 26 ▲
Ice marginal ramps (IMRs = ⟋■) together with the Main Glacial median and end moraines (■ 0) in the northern mountain foreland in the confluence area of the Akterek valley system at c. 1900 m asl from E (left side edge of the detail) towards S (No. 1) to the W (right-hand side of the detail) (Fig 1 between Nos 6 and 5). No. 1 = 4763 m massif in the valley head of the Akterek main valley branch; No. 2–No. 4 and No. 6 are the over 4200 m-high glaciated peaks and ridges of the Terskey Alatau to the far side of the 4478 m-high peak in the E (No. 4); No. 5 = W satellite of the 4763 m-high massif; No. 7 = c. 4250 m-high peak in the catchment area of the Turasu outlet glacier (Fig 2). The 4763 m-high massif belonged to the immediate glacier catchment area of the Akterek outlet glacier (Fig 2). - - - - marks the Main Ice Age glacier surfaces in their transition from the mountain to the mountain foreland. The relatively fresh glacial flank polishings (in the background) stop at this altitude. The Late Glacial end moraines (■ III = Dhampu stage) in the valley exits below are of much more recent origin. (in the middle ground) indicate a rock threshold of metamorphic sedimentary rocks which has been flowed over and polished by piedmont ice. In the flow shadow small moraine "tails" (■) link with the rock threshold. ■ I indicates the oldest Late Glacial location of the ice margin, which is formed by an IMR (⟋ ice marginal ramp). ■ 0 marks a detail of the Main Ice Age rough block moraine. It contains polymict blocks, with granite predominating. These granite blocks are errratics. ●● indicate glacio-fluvial drift fields (talus cone areas); ▼ mark the corresponding edges of these glacio-fluvial terrace bodies. Locality: 42°08'20"N/76°48'E (cf. Fig 27); Photo: M. Kuhle 13.9.88.





◄ Fig 27
View from the 200 and more metre - high Main Ice Age end moraines of the IMR type (■ 0) at 2100 m asl across the very extensive tongue basin of the Akterek outlet glacier. It is in the northern Terskey plateau foreland (Fig 1 between Nos 6 and 5; see Fig 2) (cf. Fig 26). This is a panorama which extends over 100 km from E (on the left side edge of the detail) via ESE with the 4672-high massif in the upper Barskaun valley (No. 12) and some peaks (Nos 11–9) to the 4509 m-high massif (No. 8) SE, showing the peaks of the 4478 m-high massif (Nos. 6, 4, 3), via S with the mountains of the 4763 m massif (No. 2, 1, 5), via SW with the c. 4250 m-high massif No. 7 in the catchment area of the Turasu outlet glacier, towards W with the 4502 m-high peak (No. 15), the most westerly glaciers of which flowed from the Terskey Alatau down to the Chu valley. The central Akterek glacier emerged from the valley under No. 1, and joined parallel valley glacier branches in the foreland to form one large piedmont ice. There are late Late Glacial end moraines (■ III = Dhampu stage) in the valley exits with corresponding talus cones of the same period just below (● background). ▲▲ mark a bedrock metamorphic rock threshold which the foreland ice has polished and rounded. Superimposed upon it are earlier Late Glacial moraines (■ II), which are classified as belonging to the Taglung stage. At the foot of the mountains (■ in the background) as well as at the rock threshold (■ middle-ground) Main to Late Glacial moraine "tails" are deposited in the flow shadow of the glacier. On the outside they are followed by the main glacial tongue basin with the glacio-fluvial drift fields (● talus cones, outwash fans) filled in by the Late Ice Age. ■ I marks old Late Glacial (Ghasa stage) ice marginal ramps (IMRs). Their lee-side ramps consist of outwash slopes with a 7–15° incline (▼). In the foreground the Main Ice Age end moraine or IMR edge can be seen (■ 0) running along the northern edge of the tongue basin. Once close to it, the edge of the glacier has given way to small meltwater valleys, which start here (■ in the middle foreground) and lead N down to Lake Issyk Kul (cf. Fig 28 and 29). The piedmont ice surface of the Akterek outlet glacier is entered in Fig 26 and 29 (- - - -). Locality: 42°07'20"N/76°45'E; Photo: M. Kuhle 13. 9. 88.

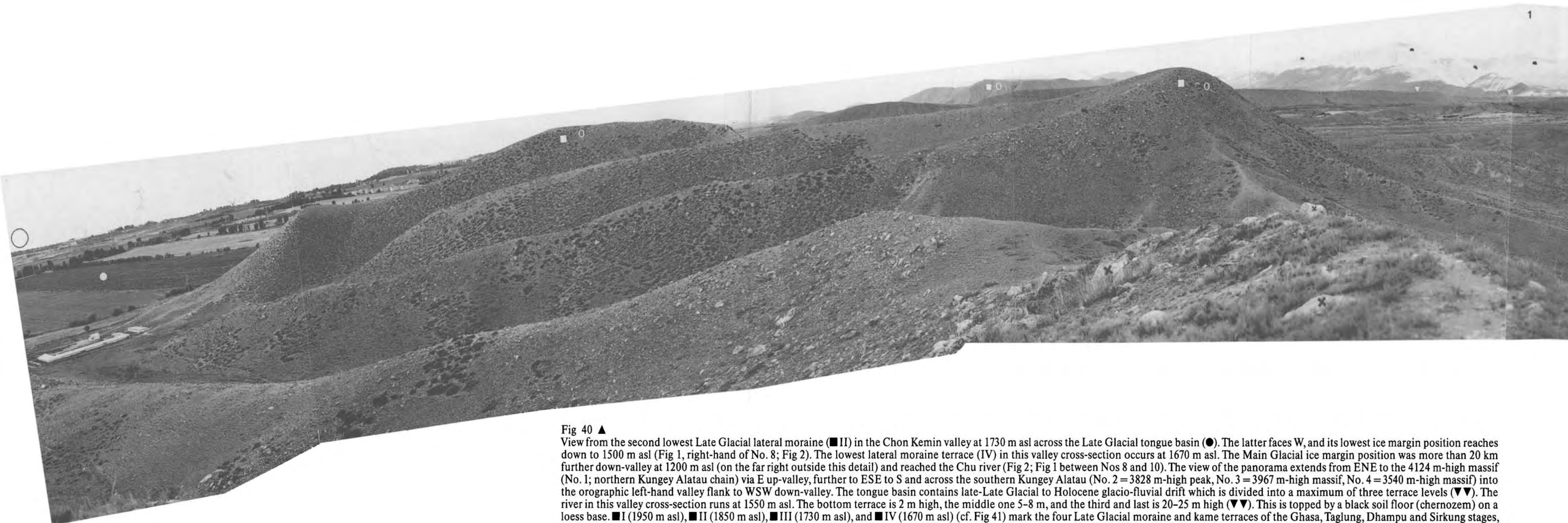


Fig 40 ▲
View from the second lowest Late Glacial lateral moraine (■II) in the Chon Kemin valley at 1730 m asl across the Late Glacial tongue basin (●). The latter faces W, and its lowest ice margin position reaches down to 1500 m asl (Fig 1, right-hand of No. 8; Fig 2). The lowest lateral moraine terrace (IV) in this valley cross-section occurs at 1670 m asl. The Main Glacial ice margin position was more than 20 km further down-valley at 1200 m asl (on the far right outside this detail) and reached the Chu river (Fig 2; Fig 1 between Nos 8 and 10). The view of the panorama extends from ENE to the 4124 m-high massif (No. 1; northern Kungey Alatau chain) via E up-valley, further to ESE to S and across the southern Kungey Alatau (No. 2 = 3828 m-high peak, No. 3 = 3967 m-high massif, No. 4 = 3540 m-high massif) into the orographic left-hand valley flank to WSW down-valley. The tongue basin contains late-Late Glacial to Holocene glacio-fluvial drift which is divided into a maximum of three terrace levels (▼▼). The river in this valley cross-section runs at 1550 m asl. The bottom terrace is 2 m high, the middle one 5–8 m, and the third and last is 20–25 m high (▼▼). This is topped by a black soil floor (chernozem) on a loess base. ■I (1950 m asl), ■II (1850 m asl), ■III (1730 m asl), and ■IV (1670 m asl) (cf. Fig 41) mark the four Late Glacial moraine and kame terraces of the Ghasa, Taglung, Dhampu and Sirkung stages, while ---- show the corresponding glacier levels. Here, in a southern exposition, the ■III moraine terrace (far left and far right) is only 1.5 km wide. ◣◓ indicate the valley shoulders, rock terraces, mountain spurs and glacigenic cuspate slopes (◣ below No. 2) the Last Ice Age glacier cover polished, truncated and rounded. Now under periglacial climatic conditions, the corries (○) present post-Glacial solifluidal moving drift covers with a large proportion of blocks. Locality: 42°47'N/76°07'15"E; Photo: M. Kuhle 13. 9. 88.

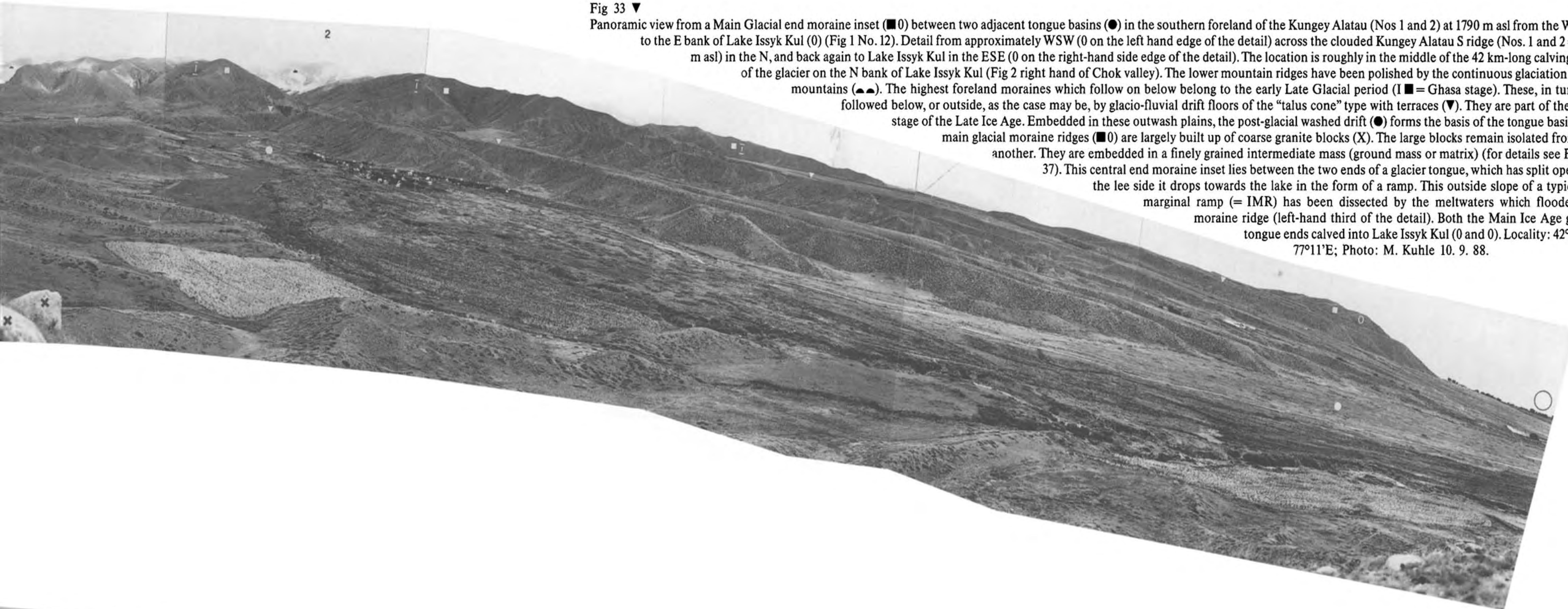


Fig 33 ▼
Panoramic view from a Main Glacial end moraine inset (■0) between two adjacent tongue basins (●) in the southern foreland of the Kungey Alatau (Nos 1 and 2) at 1790 m asl from the W bank to the E bank of Lake Issyk Kul (0) (Fig 1 No. 12). Detail from approximately WSW (0 on the left hand edge of the detail) across the clouded Kungey Alatau S ridge (Nos. 1 and 2 = 4771 m asl) in the N, and back again to Lake Issyk Kul in the ESE (0 on the right-hand side edge of the detail). The location is roughly in the middle of the 42 km-long calving front of the glacier on the N bank of Lake Issyk Kul (Fig 2 right hand of Chok valley). The lower mountain ridges have been polished by the continuous glaciation of the mountains (▲▲). The highest foreland moraines which follow on below belong to the early Late Glacial period (I ■ = Ghasa stage). These, in turn, are followed below, or outside, as the case may be, by glacio-fluvial drift floors of the "talus cone" type with terraces (▼). They are part of the same stage of the Late Ice Age. Embedded in these outwash plains, the post-glacial washed drift (●) forms the basis of the tongue basin. The main glacial moraine ridges (■0) are largely built up of coarse granite blocks (X). The large blocks remain isolated from one another. They are embedded in a finely grained intermediate mass (ground mass or matrix) (for details see Fig 34–37). This central end moraine inset lies between the two ends of a glacier tongue, which has split open. On the lee side it drops towards the lake in the form of a ramp. This outside slope of a typical ice marginal ramp (= IMR) has been dissected by the meltwaters which flooded the moraine ridge (left-hand third of the detail). Both the Main Ice Age glacier tongue ends calved into Lake Issyk Kul (0 and 0). Locality: 42°41'N/77°11'E; Photo: M. Kuhle 10. 9. 88.

Fig 36 ▲
Photograph taken within the same Ice Age end moraine complex (■ 0) like Fig 34 (Fig 1, No. 12). View from 1700 m asl across the Main Glacial tongue basin with the Late Glacial drift floor terraces (outwash plain terraces) (▼) on to the early Late Glacial end moraines (■ I = Ghasa stage) and the lower, glacially polished mountain ridges to the south. ↑ marks the glacio-fluvial dissection of the Late Glacial moraines (■ I) in the valley exit. ● are blocks of granite (cf. Figs 32 and 35), which have been transported over long distances. X indicates a more recent Main Glacial ice margin, evidence of which is shown in a moraine ledge deposit. A glacier tongue passed a somewhat older moraine complex here before descending to Lake Issyk Kul (Fig 2 right of Choktal). Locality: 42°39'59"N/77°12'01"E; Photo: M. Kuhle 10. 9. 88.

Fig 38 ▼
View from the N bank of Lake Issyk Kul in the S mountain foreland of the Kungey Alatau at 1630 m asl towards the N across the extensive ground moraine areas (X 0) (Fig 1 No. 13). The ground moraine contains large blocks of granite (●). It extends into Lake Issyk Kul, ie further down than 1600 m asl (Fig 2 right-hand of Choktal). The foot of the mountain has been glacially polished (▲). ▼mark late glacial drift floor terraces (talus cone terraces) which run down from the ice margin locations at the immediate mountainfoot. Locality: 42°39'20"N/77°13'E; Photo: M. Kuhle 10. 9. 88.

Fig 41 ▲
View from 1720 m asl on to the inner slope of the orographic right-hand lateral moraine terrace (■ III; Fig 40) with large polymict, rounded to facetted, in parts even striated blocks over unstratified crystalline blocks (X = granite and syenite) (Fig 1 on the right of No. 8). No. 1 marks the catchment area of the valley and at the same time the position of the valley head in the central Kungey Alatau in the E. ▲◢ mark orographic left-hand Late Glacial flank polishings. Locality: 42°47'N/76°07'15"E; Photo: M. Kuhle 13. 9. 88.

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.

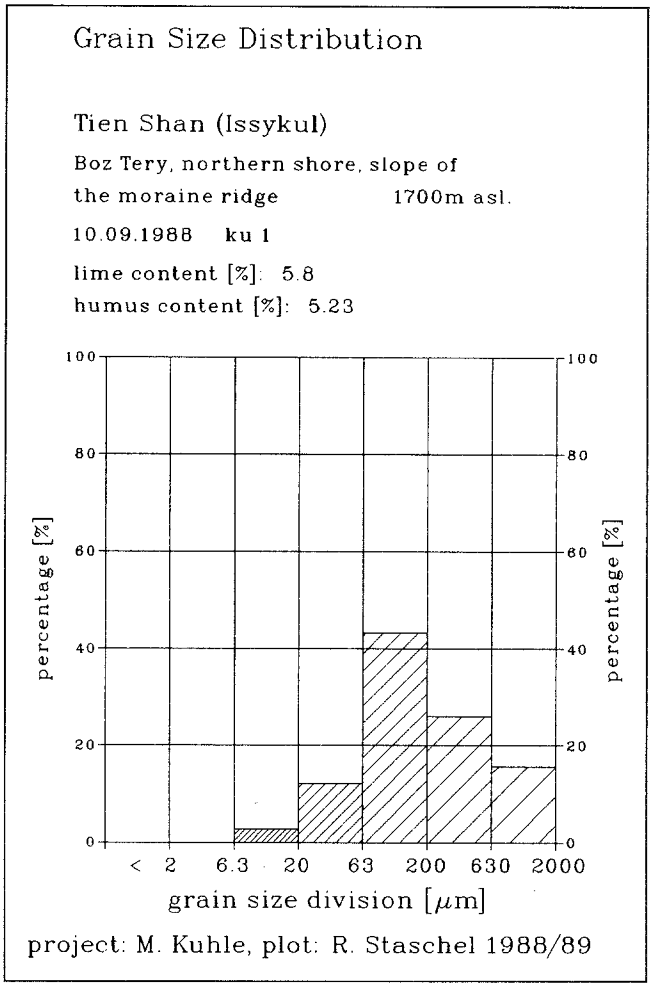


Fig 32 Taken from the surface of the end moraine ridge c. 100 m above Lake Issyk Kul, this moraine material (Fig 1, No. 12) is relatively coarse thanks to its high proportion of granitic sand. On the surface the fine material has moreover been washed out by excessive meltwater and rain-wash (however, 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

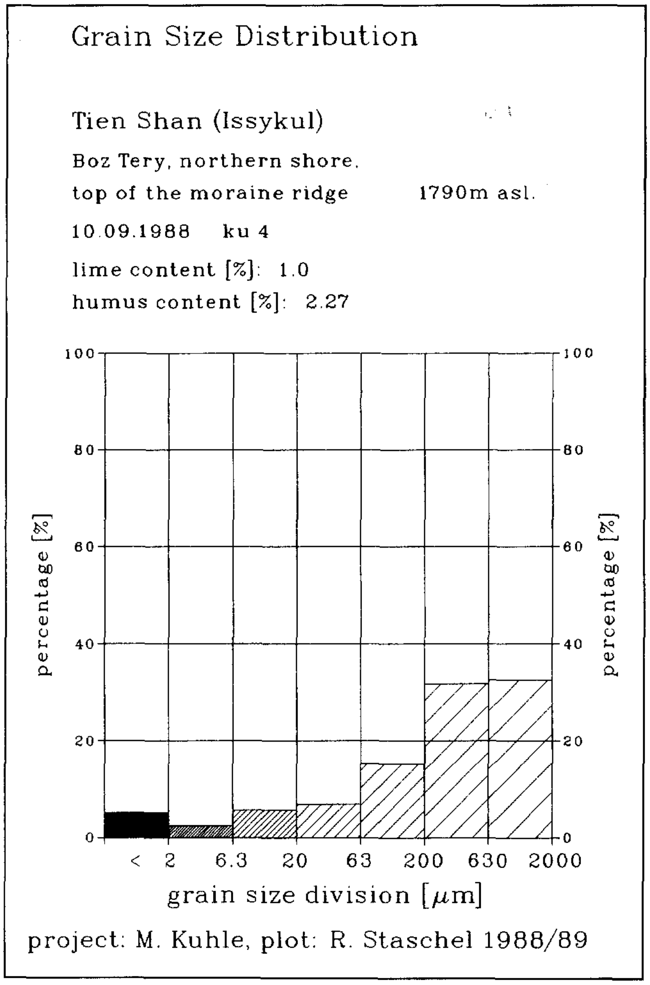


Fig 35 Main Ice Age moraine material from a depth of 20 cm (locality: Fig 1 No. 12; Figs 34 and 36), showing the typical "fine grain size peak" in the bimodal course of the graph of the columnar diagram. Locality where the samples were taken: 42°40'N/77°12'E.

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

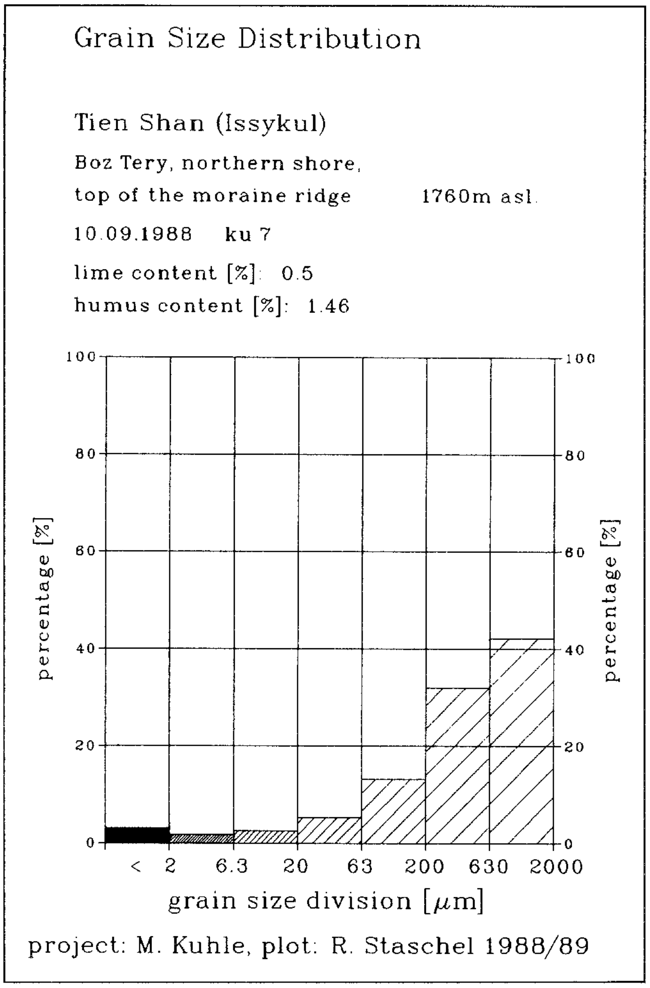


Fig 37 Somewhat richer in sand, but less well endowed with pelite (cf. Fig 35), this end moraine sediment belongs to the same Main Ice Age moraine complex as shown in Fig 36 (Fig 1, No. 12), with the bimodal course of the column levels, including a distinct "fine grain size peak" and a "coarse grain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

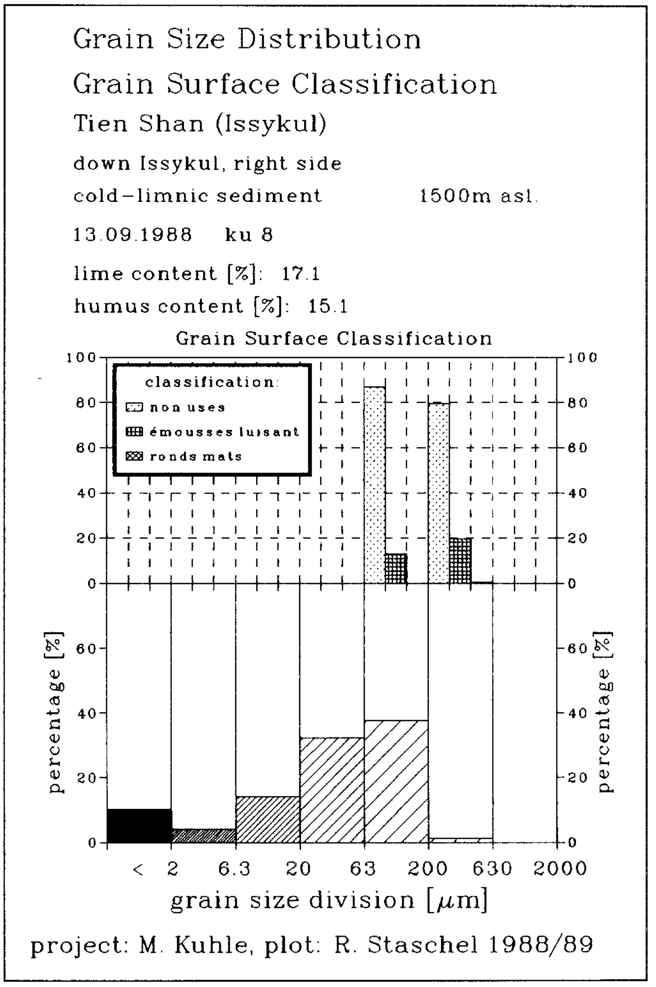


Fig 39 Held back by the glacier in the Chu valley above the Boam canyon, the lake sediments are tens of metres above the present talweg (Fig 1 on the left above No. 15; Fig 2). Their grain size distribution has the characteristics of till. A "fine grain size peak", or bimodal course, is shown in the 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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