

Geomorphological Findings on the Build-up of Pleistocene Glaciation in Southern Tibet and on the Problem of Inland Ice –

Results of the Shisha Pangma and Mt. Everest Expedition 1984

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The Area under Investigation and Presentation of the Problem

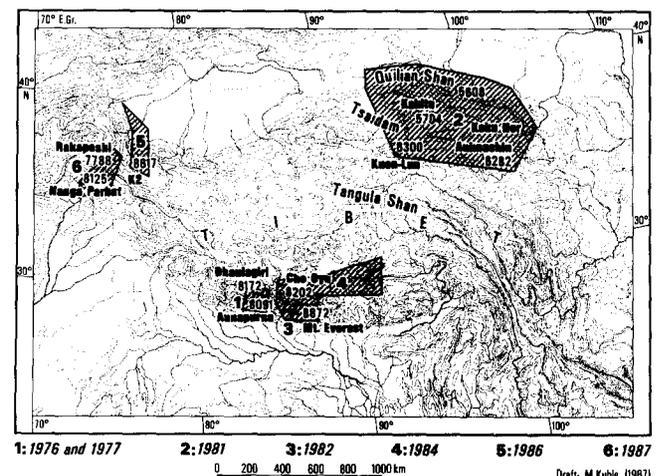
The evaluation of the findings made in 1984 is part of a complex of problems which the author had also investigated in the course of six other research expeditions to Tibet and the High Himalayas (1976, 1977, 1981, 1982, 1986, 1987). Extending from August till November 1984¹⁾, the expedition permitted previously unexplored parts of the Tibetan Himalaya, of the Tsangpo Depression and the Transhimalaya rising to the N of it to be included in the investigations of another section of the N slope of the High Himalayas, i.e. the N slopes of the Shisha Pangma and Mt. Everest (Fig 1).

The aim of the expedition was to attempt a reconstruction of the lowest pre-historic glacier margins in the N slopes of Mt. Everest and Shisha Pangma, as well as in seven mountain groups of the Transhimalaya and the Tibetan Himalaya N and S of the Tsangpo between 28° and 29° 50' N, as well as between 85° 24' and 91° 13' E. In the area under investigation, the High Himalaya rises to 8874 m, with the massifs of the Transhimalaya and the Tibetan Himalaya explored here reaching 6000 to 7200 m. In the Tsangpo valley the lowest areas descend to 3700–3800 m²⁾.

At this still great altitude of the lowest surfaces of S Tibet, which constitute a topographical limit for the climatic possibility of even lower locations for ice margins, the problem of recording those which are actually the lowest, i.e. glacier locations of the High Glacial, arises. This applies especially to very great

altitudes of glacier-catchment areas. However, it must be taken into account, that the upper limit of glacier-catchment areas underwent a climatic lowering during the Ice Age which ran parallel to, and at a constant distance from, the equilibrium line (Kuhle 1985, 1986a, b). A topographical limit of the lowest glacier margins is only missing in the S slope of the Himalaya where the Bo Chu (Sun Kosi) descends to 1000 m asl. This problem did not occur in the remaining hitherto researched areas under investigation, such as NE Tibet (Fig 1, no. 2) where the Tsaidam depression descends to about 2900 m, and the Gobi desert N of the Quilian Shan (Richthofen Mountains) extends below 1400 m. In the areas covered by the expeditions of 1976–1987 (Fig 1, nos. 1, 3 & 6), and in the NW (Fig 1, no. 5), the Ice Age glaciers were also

Fig 1 Location of the area of the 1984 Expedition (No. 4) in relation to those of the author between 1976 and 1987 (Nos. 1, 2, 3, 5, 6).



1) The expedition was financed by the Deutsche Forschungsgemeinschaft, the Max Planck-Gesellschaft and the Academia Sinica.
2) Altitudinal data in accordance with ONC H-9, 10 1:1,000,000 and aneroid measurements.

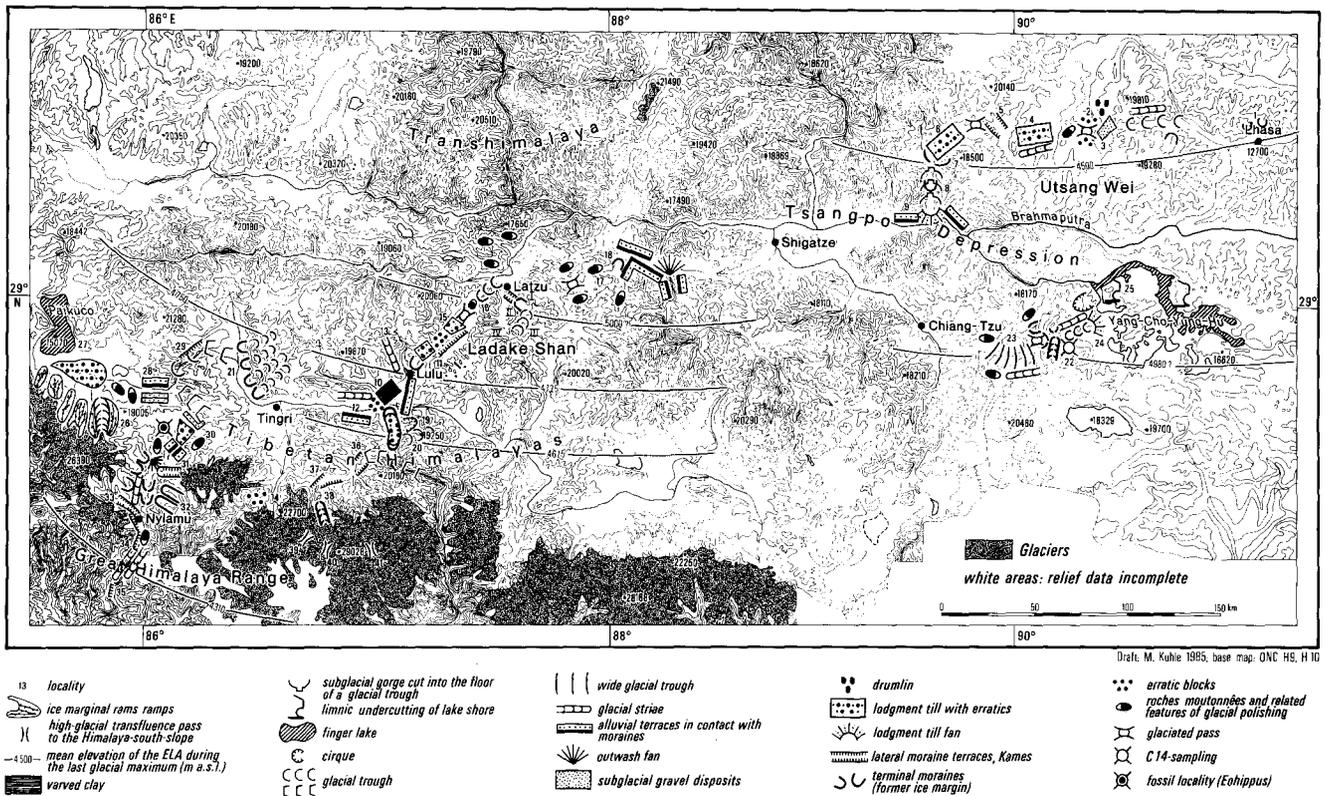


Fig 2 Glacio-geomorphological map of the 1984 investigation area in S Tibet; (Explanation in text).

able to deposit their climatically lowest moraines along the entire S slope of the Himalayas, which in some places drops to 600 m. In this way the findings of the fieldwork carried out there complement the results presented here. On the other hand, the area explored in 1984 plays a significant role as a gap in Central Tibet, which touches upon the Transhimalaya and follows the Tsangpo depression.

The Method of Calculating the Line of Equilibrium as a Means of Reconstructing Pre-Historic ELAs

The calculation of the levels of equilibrium lines can be carried out with the methods used by v. Höfer (1879), Brückner (1886), Richter (1888) and Kurowski (1891), and their variations developed by Louis (1955), Kuhle (1980, 1982, 1986b), that is by Gross et al. (1977), since only they are based upon factors which are more or less obtainable in the absence of glaciers. These factors are the mean altitudes of the crest frames (v. Höfer) or the highest summit altitudes (Louis; Kuhle) and the lowest location of ice margins (terminal moraines), as well as the superficial extent of the accumulation and ablation areas (Kurowski; Gross et al.), but also the difference of the angle between the average slope of the area of accumulation and ablation. (Kuhle 1988c).

Due to insufficiently exact large-scale topographic maps, methods based on the division of areas are not suitable for the area under investigation. Moreover, the possibility of variations are too great and range from $AAR = 0.5$ (1:1) to 0.85 (1:5.66) from accumulation area to ablation area, as was demonstrated by Meier (in Müller 1980, p. 93) in the case of North American glaciers with an even balance. The accuracy of the v. Höfer method can therefore not be improved upon in the case of the area under investigation. Like its derivative methods of calculation (cf. Kuhle 1986b, p. 41), its application provides sufficient precision in conjunction with the topographical maps available for S Tibet. The author therefore intends to confine himself to the v. Höfer method and, in cases of pre-eminent avalanche feeding, to its variation according to the "maximal method" (Kuhle 1982). The maximal method substitutes the mean altitude of the crest frame with the highest point of the catchment area, and sets it in arithmetical relation to the lowest location of an ice margin. The chosen approach recommends itself by its lucidity, and is intended to avoid a mixing of methods, being based upon the principle of supra-regional comparability: not only all the equilibrium line reconstructions (ELA, GWL) carried out by the author in High Asia, but also the majority of calculations of other authors, or their compilations in v. Wissmann (1959), were conducted in accordance with v. Höfer (1879).

Lowest Ice Margin Locations and Equilibrium Lines in the Transhimalaya (North Utsang Wei, to the North of the Tsangpo) (Fig 2 and Tab 1)

Joining the Lhasa valley immediately N of the town, a northerly tributary valley contains moraine complexes which run from two third-order valley ends in a southerly and subsequently, from the larger basin, in a SW direction down to 4250 m and 3950 m asl ($29^{\circ} 43' N / 91^{\circ} 04' E$; Fig 3), even reaching a valley floor of a higher order in the last case. With a mean altitude of c. 5200 m for the catchment area, the *orographic equilibrium line* for the SE aspect is calculated as occurring at 4725 m, and for the SW exposition at 4575 m asl. The *climatic equilibrium line* near Lhasa at the time of the *last High Glacial period* accordingly ran at about 4500 m (Fig 2, no. 1).

It was noted in the Tibetan Himalaya N (i.e. to the leeward) of the Dhaulagiri-Himal that an expositional lowering of the equilibrium line by 300 m from the S to the N aspect had occurred (Kuhle 1982, p. 170, 1983). At a corresponding latitude, under comparable conditions of incoming radiation and equal annual precipitation (Jomosom 270 mm/y, Gyantse 271 mm/y and Lhasa 437 mm/y) the climatic equilibrium line would here, too, be found to run 150 m below the orographic equilibrium line on a S exposition. In the case in hand the climatic equilibrium line is thus to be assumed as being 75 m lower than the orographic equilibrium line on a SW exposition.

Extending over tens of kilometres, the E and W slope of the Shüke oder Chalamba La (pass to the W of Lhasa, $29^{\circ} 41' N / 90^{\circ} 15' E$) furnish firm indicators of the existence of pre-historic ice flow networks with *glacier thicknesses* of c. 1200 m in valley-branch flows in the form of erratic findings (Fig 2, no. 2). Biotite-granite blocks with K-feldspar (albeit without chlorite) over a dark Mesozoic rhyolite bedrock (with chlorite) have been found up to 100–150 m above the height of the pass (5300 m) (Fig 4 & 5). Analogous parent rock of probably pre-Miocene age does not occur in the area. According to the geological map presented by Gansser (1964, Plate 1) they only appear at the surface in the Lhasa area, 80–100 km further to the E. This implies transport from the E over tens of kilometres. The topographic arrangement of the erratics also indicates an ice-sheet which swept across the watershed of the pass without a break in the characteristic manner of an ice-stream-network. Rounded and polished mountain ridges rising to 5600 m and steep rock walls towering above them are confirmation of the reconstructed ice level.

Up to 50 m high remnants of ice marginal ramps, esker-dams or kame remnants found to the E of the actual culmination of the Chalamba La at an altitude of 4300 m are already regarded as phenomena of the sub-glacial disintegration of the High Glacial ice cover, just as transverse and lateral moraines between glaciers breaking up into several lobes are considered as belong-



Fig 3 Moraine complex some decametres thick in the form of a terminal moraine between 3950 m in foreground and 4250 m in the middleground of the photograph (X) in a tributary valley near Lhasa (S exposure $29^{\circ}43' N 91^{\circ}04' E$; cf. Fig 2, No. 1). The incisions in the moraine are cutting up to 50 m deep through the diamicton. It lies on granite. The glacier tongue flowed from a corrie-like valley-end at 5200 m from the upper right and pushed the terminal moraine like a pedestal. Photo: M. Kuhle.

ing to the Late Glacial age ($29^{\circ} 41' N / 90^{\circ} 15' E$) (Fig 2, no. 3) On the W fringe of this wide, intra-montane basin there are drumlin-like accumulations.

All of these accumulations are in the centre of the basin. On the edges where tributary valleys present themselves at the 4500 m-high basin floor, ice marginal ramps (*bortensander*) (Kuhle 1984, 1989) and intersecting *outwash cones* are evidence of later, Late Glacial glacier stages. The mean level of the catchment area equals 5400–5600 m here. Higher up in these valleys there are still more recent *dumped end moraines*, which resulted from the block glacier phases as the last stage before the retreat of the ice (v. Klebelsberg 1948; Höllermann 1964; Kuhle 1982). This does not refer to periglacial “block glaciers”.

W of the Shüke or Chalamba La, and down to at least 4430 m asl, the floor of the descending valley and the adjoining N/S basin is covered with ground moraines containing erratics and a lot of interstitial material (Fig 2, no. 4) ($29^{\circ} 35' N / 90^{\circ} 00' E$). In some places the boulder clays are tens of metres thick. Though more than a kilometre wide, the entire floor is covered by the ground moraine deposits. The small present-day stream has not yet succeeded in a fluvial transformation of the ground moraine above 4580 m. Only downstream has

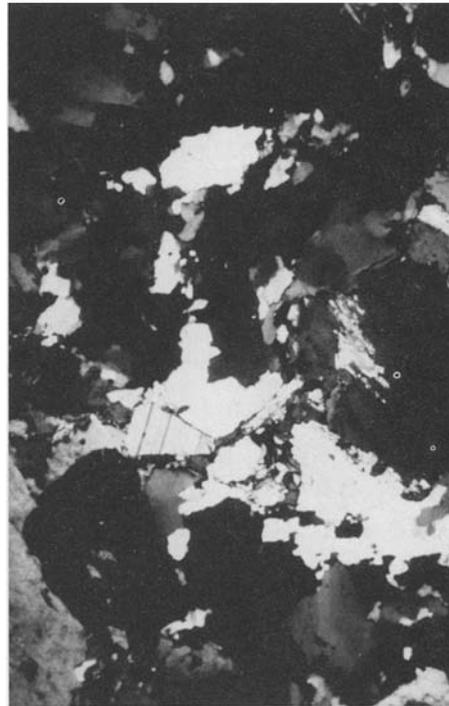


Fig 4
Thin section of the solid rock of the Chalamba La (Col 5300 m asl 29°41'N 90°15'E; cf. Fig 2, No. 2) upon which lie the erratics of Fig 5. This is a rhyolitic rock with crystals of quartz, albite-oligoclase, mica and chlorite and is hydrothermally altered. Its eruptive character is shown by its macro-crystalline features (xenocrysts of twinned feldspar) embedded in an amorphous matrix. Laboratory photograph: A. Heydemann and M. Kuhle.

Fig 5 Thin section of an erratic block from the Chalamba La (see Fig 4 and Fig 2, No. 2). This is a biotite-granite i.e. an acid plutonic rock (light block) of quartz (dark, wavy extinction), biotite, albite-oligoclase and potash feldspar crystals with a macro-crystalline structure. This structure and its different mineral composition (the granite with potash feldspar, the solid rock rhyolite with chlorite) show that these blocks are glacially transported erratics. Laboratory photograph: A. Heydemann and M. Kuhle.

the increasing discharge led to the flushing out of some of the fine material between the boulders, though even here they cannot be distinguished from an ablation moraine with absolute certainty.

As far as the glacier retreat stages of the Late Ice Age are concerned, two tongue basins of 4800 m and 4650 m are closed off by frontal moraines, which are particularly well preserved (29° 38' N/90° 10' E) (Fig 2, no. 4). The present-day glacier, which had caused the formation of the two tongue basins in the past, is situated at the foot of the dominating and well over 6000 m high peak. A tributary valley on the right hand had provided a link with the valley which descended in a westerly direction from the Shūke La.

Fifteen km S of the 6138 m (20, 140 ft) high summit there is a valley leading W and up to the 4900 m high Zu Ka La, where kame terraces or paraglacial embankments have been preserved at altitudes around 4550 m (29° 44' N/89° 54'E). Besides boulder clay, they tend to be sandy accumulations, indicating a prolonged, Late Glacial ice deposit, followed by the formation of dead ice (Fig 2, no. 5).

In the valley floor areas on both sides of the Zu Ka La this is followed by covers of ground moraine extending over a distance of 10 km S of the culmination (to approximately 29° 43' N/89° 45' E; Fig 2, no. 6).

Away from this valley, which is called Orio Matschu, towards the S and in the direction of the Tsangpo valley and further down, a significant bipartite valley cross-section starts at about 4380 m; its lower part is box-like, while the upper part presents a trough-like extension;

such bipartite profiles are to be attributed to glacial formation in the upper part and to simultaneous sub-glacial meltwater erosion, together with lateral undercutting, in the lower part. It is a formation which developed only below the level of the High Glacial equilibrium line (Kuhle 1982 p. 50, Fig. 97, 1983a, p. 117, 227). Its extent is explained by a major discharge of the Ice Age network of ice streams, i.e. of the Tibetan inland ice, for the box shape appears to be set 50–100 m deep in the trough floor. Valley shoulders (rock terraces) separate it from the trough flanks, which are at times only slightly concave or straight.

Moraine accumulations some 4340 to 4300 m high occur 202 km away from Lhasa (by road) (29° 37' N/89° 39' E, Fig 2 no. 7). Below them slightly terraced, stratified glacio-fluvial drift, with many rough blocks, take over. They are evidently outwashed moraine deposits. Outwash moraine material may be observed in the openings of all the side valleys up to 4250 m.

About 20 km down valley at the basal level of about 4070 m, there is a moraine accumulation on the true left-hand bank, which forms a polymict terrace, with rounded as well as faceted blocks embedded in an ample matrix of fine material (not outwashed) that runs down-valley over several kilometres (29°28'N/89° 37'E). On the valley side of these moraines, large mudflow- and alluvial fans from side valleys have been piled-up in a kame-like fashion on the probably Late Glacial ice fringe of the valley glacier – itself probably already reduced. Reaching heights of 100–120 m, these conical shapes break off more or less immediately, though sometimes in

Fig 6
Late-glacial end-moraine (××) in the Orio Matschu valley with its base at 4020 m (29°27'N 89°38'E; Fig 2, No. 8). The walls of the moraine are formed of large blocks in a matrix of fine loamy material. The true left wall (× right) carries a wash of loess. Below this terminal position lies an outwash fan forming a low terrace (◆). Photo taken into the valley by M. Kuhle.



Fig 7
High or Late-glacial outwash fan (No. 5) forming 35 m to 60 m high terraces at the outlet of the Orio Matschu valley (29°21'N 89°36'E, 3820 m; Fig 2, No. 9) in the Tsangpo valley; the Tsangpo on the right. If in fact the outwash fan No. 5 is of maximum glaciation age this shows that then the Tsangpo valley was free of ice (cf. Fig 8). Photo: M. Kuhle.



three or four steps, towards the bottom line of the main valley.

Four kilometres down-valley from that moraine there are wall moraines with basal altitudes of about 4020 m (29° 27' N/89° 38' E, Fig 2, no. 8; Fig 6). Two lower gravel terraces of more recent origin penetrate the walls, which continue to block half the valley.

Further on down-valley (29° 25' N/89° 36' E) and interlocking, and at times modified by mudflow substratum, or by single mudflow events from a side valley on the left, there are accumulations which attain thicknesses in excess of 200 m. They consist of polymict blocks of mixed degrees of roundness without suggesting any sorting. In up-valley direction, as well as down-valley, they change into graded valley fills with four terrace levels. In this profile the valley floor of the Orio Matschu shows considerable deposits. The lowest, and therefore belonging to the High Glacial stage, of these ice margins of the glacier systems concerned have been found at about 3900 m asl, on the S slope of what older maps present as the Nien-tschen-tang-la mountain group, 6–10 km from the Tsangpo, and with peaks rising to 6138 m (Fig 2, no. 9). With a mean altitude of 5400 m for the catchment area, the *orographic line of equilibrium* becomes calculable at 4650 m asl. The precise and lowest ice margin has been established by terminal moraine walls, followed by terraces of outwash plains or ice contact stratified drift, respectively.

The terraces of outwash plains of this true left-hand tributary valley of the Tsangpo continue in the 35–60 m high terrace system of the main Tsangpo valley (Fig 7).

Together with traces of glaciers in the Tibetan Himalaya S of the Tsangpo a picture evolves of two ice-stream systems of the High Glacial stage (Fig 74, 12 and 13), both of which just failed to reach the Tsangpo with the lowest edges of their outlet glaciers. Between the two glacier areas, though itself free from ice, there was thus a glacio-fluvially marked valley trough, with floors of stratified drift at altitudes of about 3800–3900 m running parallel to the latitude. Two terraces can be clearly distinguished (Fig 8). The upper one (60 m,



Fig 8
Outwash terraces No. 5, 60 m (probably maximum glacial) and No. 4, 20 m (late-glacial) in the central Tsangpo valley (29°21'N 89°34'E; cf. Fig 2, No. 9) at 3800–3900 m asl (cf. Fig 7). These contain well rounded components in part well-sorted gravels in the outwash (sandr). Tsangpo in the foreground. Photo: M. Kuhle.



Fig 9
The Tsangpo trough; alluvium with flood plain and river (29°20'N 89°33'E; 3830 m). Where the flood plain ends a terrace a few metres high has developed (▼) and above, running out on it, is an alluvial fan (■). The patches of aeolian sand on the slopes (●) show the semi-aridity of this environment (with an annual precipitation of 200 mm) at an elevation a little below the potential tree-line (4300 m) and the permafrost-line (4700 m). The pleniglacial glacier flowed from the ice-smoothed rock domes and ridges (◆) down into the Tsangpo valley.
Photo: M. Kuhle.

no. 5) has hitherto been attributed to the last High Glacial, whereas terrace number 4 (20 m) is considered to be the result of the late glacial Ghasa Stadium (I) (Kuhle 1982, p. 118). According to computer modelling of the ice dynamics by Herterich and Kalov (1988) based in turn on the author's glacio-geological data (Kuhle 1986c, 1987a), this E section of the Tsangpo valley could well have been filled with glacier ice during the last High

Ice Age. Geomorphological evidence for such filling is not, however, available so far (Fig 9). Since the Tsangpo valley bottom line exceeds 4600 m, an ice bridge between the Transhimalaya ice (I 2) and the most southerly complex of the Tibetan ice of the Tibetan Himalaya (I 3) must be assumed as being W of 86° E.

Lowest Positions of Ice Margins and Equilibrium Lines in the Tibetan Himalaya (South of the Tsangpo) (Tab 1)

Just as alluvial fillings of the Tsangpo valley point to lack of glacier cover during the last Ice Age as long as they remain undated, so do varves in the Lulu basin indicate a complex in the Tibetan Himalaya that remained free of ice during the late Ice Age in any case (Fig 2, no. 10). The floor of this basin is situated at 4300 m, and the glaciolimnic terraces are preserved time and time again over tens of kilometres. They end in a front with fan-like deltaic deposits towards the basin (Fig 10 ↓).

The Glaciolimnic Sediments in the Lulu Basin as a Representative Sample

The up to 60 m thick varves were subjected to detailed investigation. Fig 11 shows the annual stratifi-

Fig 10 Glacial-lake deposits of a late-glacial ice-dammed lake (▼) on the N slopes of the Lulu basin. The lake terrace is 60 m high and lies with its edge in a terrace flight of deltaic deposits from tributary valleys (↓). The lake deposits are easily eroded and the deeply eroded channels contrast with the very small gullies above its floor (28°38'N 87°04'E; Fig 2, No. 10; cf. Fig 11–16). Photo at 4300 m. M. Kuhle.



cation of 0.2–0.5 cm thick layers. Medium silt (MS) and clay prevalence (C) are characteristic indicators of strata laid down in summer and winter (Fig 12, above right and below left). Frenzel (in a private communication dated 13. 6. 88) had subjected the samples to electron-microscopic screening and, by comparing them with fermented loam material from N Scandinavia, interpreted and then confirmed them as being glaciolimnic. He recognized characteristic “chattermarks” (Fig 13) and curved breaks with sharp, glacial fracture edges (Fig 14 and 15) in the angular quartz grains of the strata put down in summer and winter. These sharp edges also indicate the absence of aeolian polishing, the results of which may range from tarnishing to rounding-off down to the grain size of the summer strata. The author regards the wealth of small clay mineral plates in the summer strata as the typical “mineral grain peak” (fine grain peak) of the bimodal grain size distribution of glacial material (Dreimanis & Vagners 1971). This peak is even more evident in the fine spectrum of this secondarily (i.e. limnically) enriched material than in the primary boulder clay of the actual moraines (Fig 16). In general the wealth of clay minerals is initially dependent on the existing bedrock in the denudation area and its pre-glacial periglacial conditioning process. The comparison with the samples Frenzel (cf. above) had taken in N Scandinavia are evidence of this; in spite of pedogenesis (they come from a depth of only 4–7.5 cm) they present a smaller proportion of clay mineral. It must, however, also be noted in principle that these combinations of sediments, which are proportionally rich in fine grains, can only be the result of the grinding glacier action. In the Tibetan varves under consideration, the new formations of clay mineral through pedogenesis can be excluded, as the material has been taken from a depth of 45 m. There is no indication of soil formation (Fig 10 & 11).

These paraglacial (glacio-fluvial and glaciolimnic) and solifluctual sediments on the metamorphic bedrock of the slopes became Holocene and are still being blown out. Thus blown sand floors with ripple markings and hummocks came to cover the pebble terraces with thicknesses ranging from a few decimetres to several metres (Fig 12, top left). The aeolian character is revealed through the comparison with purely glacio-fluvial sediments (Fig 17).

Some Observations on the Glaciation of the Lulu Basin and the Southern Slope of the Lazu Massif (Ladake Shan)

A striking terrace of outwash material (15–20 m high) has been deposited in the Lulu basin from the Lulu valley (Fig 2, no. 10). It dovetails with the glaciolimnic sediments but, being about 40 m lower, it does not attain their height of deposits. Composed of plutonic and metamorphic gravel, some the size of a head (60%

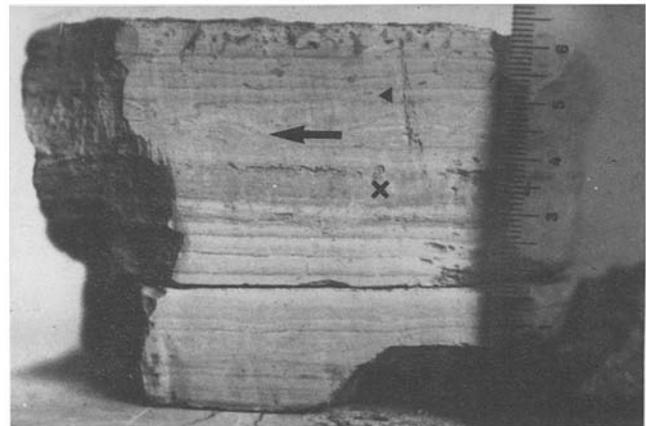
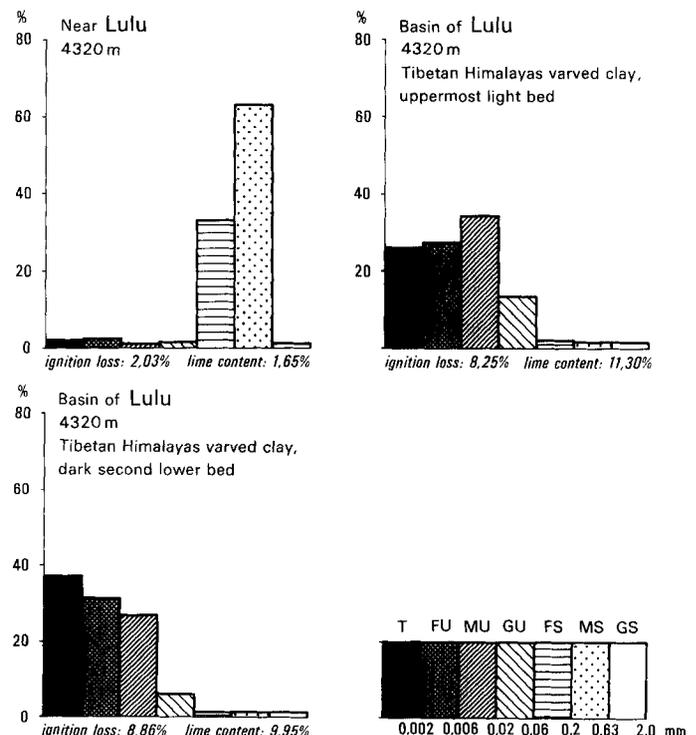


Fig 11 The lake deposits in the Lulu basin are up to 60 m thick (cf. Fig 10). They have all the characteristics of rhythmites of glacial origin and are obviously varve deposits. The yellow intercalations (10YR 7/6) are 1–2 mm thick (▲) and alternate with light grey bands (5Y 6/1) 1–5 mm thick (×). They contain weak colour transitions to light grey (cf. Fig 12 grain size analysis of winter and summer layers). Some of the bands show push features (←). Laboratory photo: M. Kuhle and J. Peters.

rounded, 38% with smooth edges: Reichelt Method, 1961), the terrace becomes a diamictitic boulder clay cover towards the valley. This change in material is evidence of the position of a valley glacier ice margin at an altitude of 4350 m (S slope of the Lazu massif, Fig 2,

Fig 12 Particle size analysis from the Lulu basin (28°38'N 87°04'E) of the aeolian sediments in mounds on the outwash terraces (upper left) and late-glacial varve sediments (cf. Fig 11).



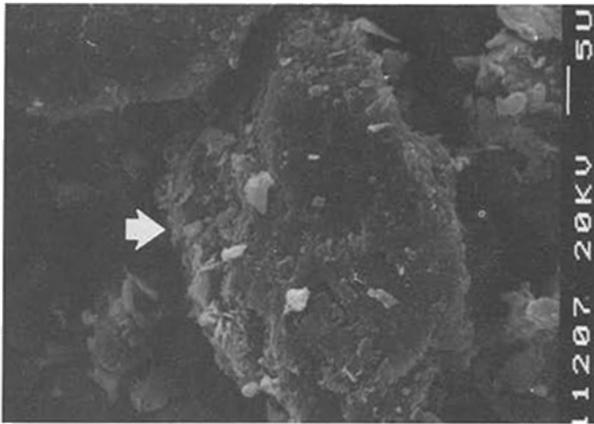


Fig 16

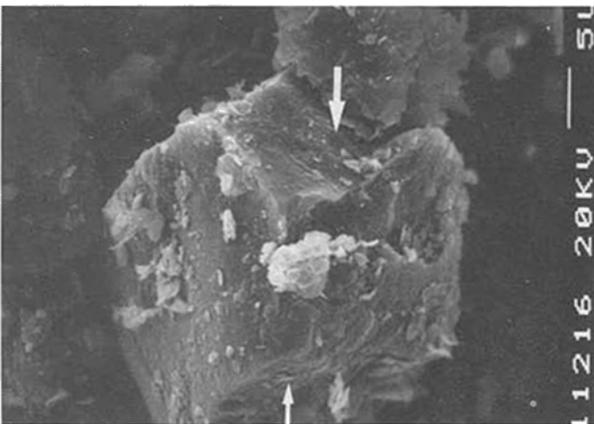


Fig 15

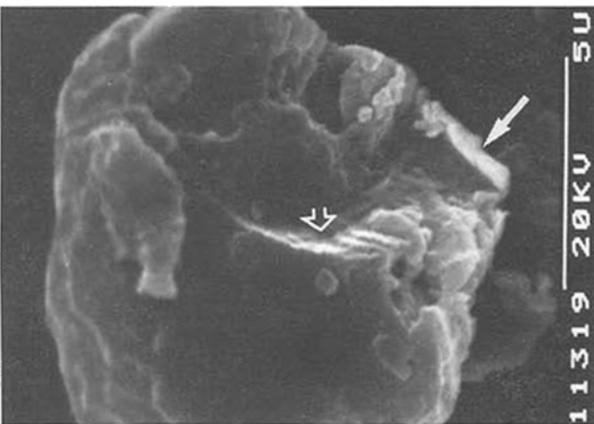


Fig 14

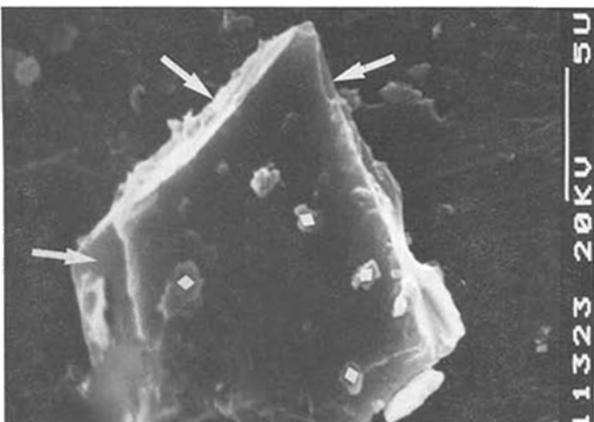


Fig 13

Fig 13 Glacially modified quartz particles with some little-modified clay mineral particles (◆◆) from the winter layer T (clay) of the varve sediments of Lulu (4320 m) (cf. Fig 11 and 12). There is no sign of aeolian or periglacial working. Fracture surfaces of this kind (←) on quartz particles are typical of the 'Gärlehme' of Scandinavian ground moraine. Scanning-electromicrograph and analysis: Laboratory B. Frenzel.

Fig 14 Quartz particle, also from the winter varve layer of Lulu (4320 m) dominated by clay. The grain shows typical crescentic cracks (↵) and impact surfaces (←; chattermarks). Here, too, there are no rounded edges which would denote aeolian or periglacial modification (cf. Fig 11 and 12). Scanning-electromicrograph and analysis: Laboratory B. Frenzel.

Fig 15 and 16 Samples from the Lulu basin summer varve deposit (dominated by medium silt fraction; Fig 11 and 12) at 4320 m. The large quartz grains show smooth impact surfaces (Fig 15 ←) and are like wedge-shaped chattermarks typical of glacial transport. The grains are plastered with clay minerals derived from the glacier milk suspension. These also form even an overall-layer on those of the summer deposits (Fig 16 ↗). Newly formed clay minerals do not occur since these samples have never been affected by pedogenesis (weathering) due to the great depth – 10–45 m below the surface – from which they were taken. Scanning-electromicrograph: Laboratory B. Frenzel.

no. 11, 28° 45' N/87° 14' E). It shows an orographic ELA-lowering up to a maximum of 1275 m to an altitude of 4725 m. However, for the time being and as a matter of care the position of the ice margin is still to be classified as of High Ice Age, although it is likely that the entire Lulu basin was filled with a glacier. This is indicated by very large granite blocks (longitudinal axes of up to 3 m) in a matrix of fine material at the present level of the river (4300 m) (Fig 2, no. 12). They present the picture of an outwashed moraine, and are W of and outside the axis of the mouth of the Lulu valley. It may in fact be the moraine material transported by mud-flows, since it cannot be entirely ruled out that moraine lakes in the catchment area burst their banks. Under conditions of an ice-filled Lulu basin however, the striking rock polishing on the upper slopes, even on outcropping metamorphic slopes in the Shegar Tsong basin (4400 m) adjoining in the NW, could be explained by the polishing action of glacier masses which were several hundred metres thick (Fig 2, no. 13). The under slopes are now being dissected by regressive linear erosion. The notion of the Lulu valley being completely filled with ice is supported by Odell's (1925, p. 331) reconstruction of the complete glaciation of the Tingri basin (28° 35' N/86° 35' E) to the W. On the Phusi La (pass at 5411 m) (Fig 2, no. 14), S of the Tingri basin Odell found some boulders with ammonites 300 m above the Kyetrak glacier, which is now taking a N course from the Cho Oyu massif. The fossil permits the rock to be unequivocally identified as being Jurassic. It was an erratic, as pre-Jurassic metamorphic-crystalline bedrock series occur at that particular place. The boulders therefore originated in the Jurassic area 30 km to the N. Odell concluded from the erratic and the geological situation that there must have been ice-flow powerful enough to over-run watersheds and to reverse the direction of the Kyetrak glacier towards the S; coming from

the High Himalaya, it now flows N. He takes this as an indication of the ice-cover of S Tibet during the High Glacial stage. The author's own findings appear to point in the same direction. In the local context it implies that the component which ran off to the S and reached the Phusi La must have come from the Lulu basin. A landscape of glaciated knobs immediately on the edge of the Lulu or Tingri basin – like the one near Shegar Tsong – is equally evidence of considerable glacier cover and filling. Attention must also be given to the fact that the recent tongue of the Kyetrak glacier terminated at 4980 m. This is only approximately 700 m higher than the floor of the Tingri basin (4300 m). Thus an ELA depression of merely 350–400 m already made it accessible for the Kyetrak glacier. This implies that a depression of a maximum of 700 m is required for the filling of the basin with glacier ice (see below). The reorientation to the S would have taken place at this point, together with some of the run-off via the Phusi La, not only the Nangpa La (5700 m) into the Rongshar valley.

Just as the erratic on the Phusi La provides evidence, so can other findings of erratica in the Lulu valley not be explained in any other way than by glaciation that filled in the relief (Fig 2, no. 15). Here, above 4350 m (cf. above) there are boulder clays in glacial bank formations (wall forms, Fig 18), as well as valley floor fillings with very large biotite and two-mica granite blocks (longitudinal axes 1.5–4 m). These diamictites are on top of bedrock basalts into which the valley has been cut, and were deposited at least 170 m above the valley floor on the flanks of a trough and on rock ledges (Fig 19–23). More and more widely spread out, and almost reaching the 5200 m high Latzu pass (28° 54' N/87° 27' E), these ground moraines are unambiguous evidence thanks to sedimentological criteria: the rounded and light-coloured blocks of acid-plutonic rock are embedded separately in a fine, sandy to loamy-clayey matrix. This moraine cover is at least 10 m thick and superimposed upon those hydrothermally decomposed basic bedrock volcanics.

Although of comparable significance, the connection (the geomorphological sequence) with the corresponding wealth of denudation forms has generally been neglected, due to the difficulty of visualizing (and describing) topographic-geomorphic relationships: the Latzu pass borders on a classic landscape of polished basins and swells with wide-ranging trough profiles (Fig 24). It presents all the elements of Scandinavian glacial denudation landscapes (such as the Kebnekaise or Sarek massifs at 67° N, albeit 4000 m lower). The glacially scoured rocks are dressed with a shallow (i.e. decimetres thick) layer of conglifraacts. Steeper slopes experienced a more forceful periglacial transformation to frost-smoothed slopes (conglifluction slopes). Glaciated ridges have been slightly pointed by frost cliffs where the slopes of partition valley interfluvies intersect. As in Scandinavia, the periglacial morphodynamics have tended to leave a concordant and thus slight mark on the glacial relief (Fig 24).

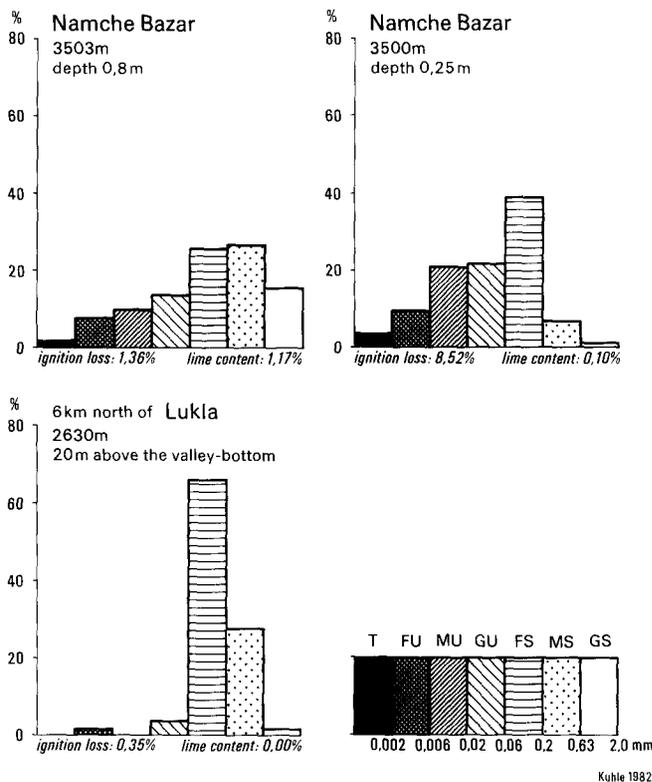


Fig 17 Representative example of glacio-fluvial sediment from the paraglacial sandr on the S edge of Tibet (Khumbu Himal; 27°42'–51'N 86°43'E). These are late-glacial paraglacial sands (Ghasa I stadium or Pre-Ghasa stagnation ● 1/2) deposited in the marginal depression alongside of the some ten-kilometre outlet glaciers originating from the S Tibetan ice stream system (Fig 75 and 76, 13). Characteristically fine sand is prominent, while in case of aeolian sands medium sand is dominant (cf. Fig 12). The microscopic analysis showed that most grains in these sandrs are unreworke and they have travelled only a short distance.

Fig 18 Ground moraine ridges in Lulu valley (Latzu massif S slope 28°48'N 87°14'E 4950 m; Fig 2, No. 15, cf. Fig 19–23). The large, light coloured morainic blocks are of granite with biotite and two-micas. The country rock shows them to be far-travelled erratics. They lie on basic volcanic rocks and 'swim' in a fine loam-rich ground mass. This is typical of ground moraine (lodgement till). Photo to N: M. Kuhle.



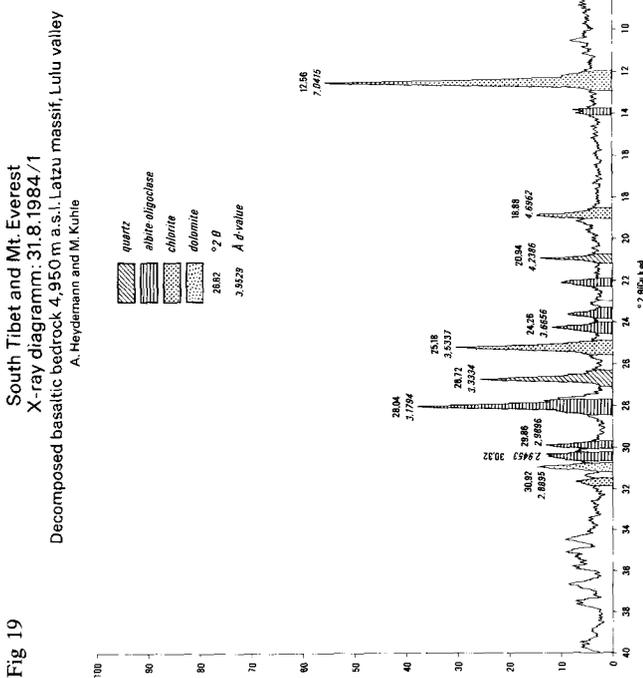
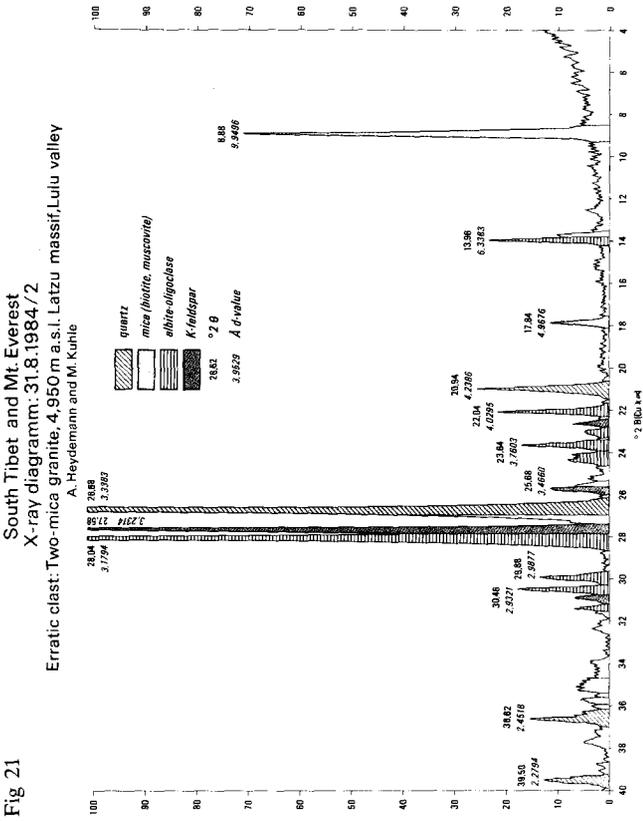


Fig 19 Country rock into which the Lulu valley is cut (28°48'N 87°14'E cf. Fig 22).

Fig 20 and 21 Petrography of the erratic granite blocks which lie upon the in situ basalts (Fig 19 and 22) of the Lulu valley (28°48'N 87°14'E cf. Fig 23).

The following considerations in relation to the geomorphological state of preservation are to be taken into account: annual mean temperatures of -9° to -12.8° C recorded at the present ELA level (Ding Yongjian 1987, p. 10, Table; Kuhle 1988a; Sun Zuoze 1987, p. 59) in Tibet are also likely to have caused very cold ice here, at least 500 m above the ELA, during the Ice Age. Cold and highly viscous ice, stretches of which are frozen to the rock surface, leaves less distinct traces of erosion.

Glacier Traces on the North Side on the Lazu Massif

On the N slope of the Lazu massif a box valley with remnants of troughs in crystalline slates leads down to the Tsangpo (29° N/ 87° 30' E). Though the forms of erosion (Fig 25) are evidence of great ice thicknesses, there are no end moraines. The Mangar (or Phu) valley, the next valley to the E, presents few definite forms of

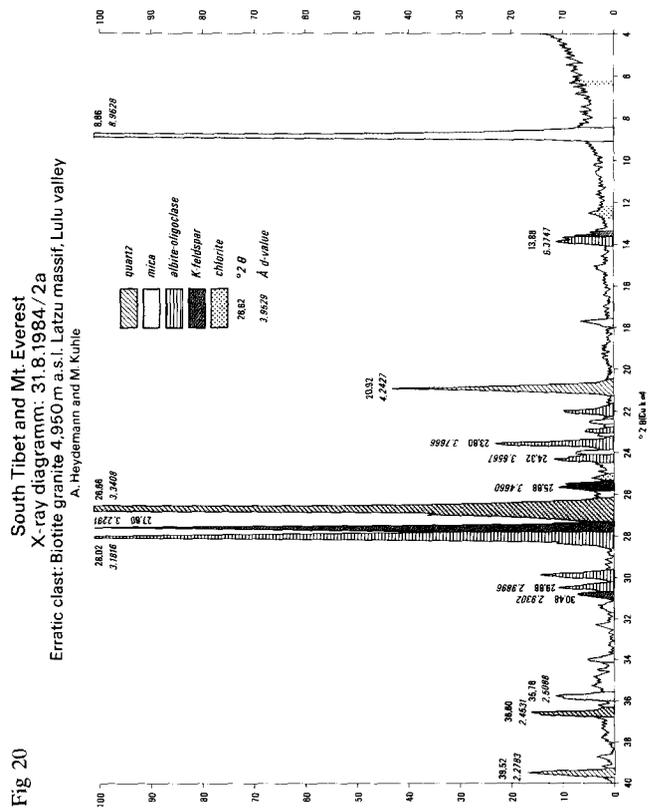
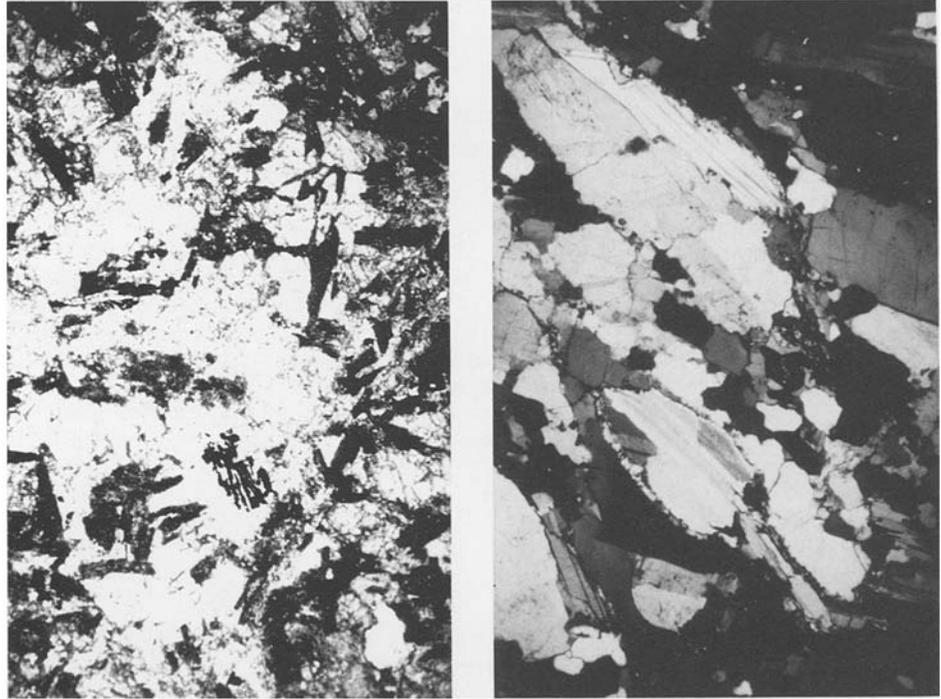


Fig 22

Thin section of the basalt country rock of the Lulu valley (Fig 19). It is a much decomposed basaltic rock in which the pyroxenic components are hydrothermally chloritised and carbonatised. Laboratory photograph: A. Heydemann and M. Kuhle.

Fig 23

This thin section of the erratic two-mica granite (Fig 21) shows the diagnostic macro-crystalline structure of plutonic rocks (Fig 22 and 23 have the same scale). Laboratory photograph: A. Heydemann and M. Kuhle.



accumulation below 4200 m, which may be of glacial origin, but evidence for this is lacking. The lowest undoubtedly glacial end moraines are to be found below the Phu settlement at an altitude of 4200 m (Fig 2, no. 16); (28° 58' N/87° 38' E). The mean altitude of the catchment area of the adjacent summit region being at 5500 m, the ELA depression would be 900 m. The orographic ELA ran at 4850 m asl (Tab 1). The ice margin is classified as belonging to the Late Glacial Ghasa Stage (I) (after Kuhle 1980, 1982). The entire landscape here has gentle features, and is provided with glacially scoured upper slopes and granite block coverings so that the relief appears to have once been completely dominated by ice (Fig 25). There is no evidence, however, of an even lower ice margin position. Higher

up the Phu valley, at 4500 m and 4700 m, and again at 5000 m asl, more Late Glacial tongue basins join in; they are surrounded by striking granite block walls (Fig 25 and 26). They are classified as the Taglung Stadium (II), the Dhampu Stadium (III) and the Sirkung Stadium (IV) (according to Kuhle 1982) with ELA depressions of 750 m, 650 m and 500 m respectively.

Strikingly rounded mountain tops and ridges were observed in metamorphics in the areas to the S of the military station at Latsu (29° N/87° 45' E) and on the 4500 m high Suo La (Solotse La) pass from the Latsu valley to Shigatse up to altitudes of 4700 to 5500 m asl (Fig 27, Fig 2 no. 17). Reference to such forms in soft sedimentary rock introduces contradictory arguments: 1. Soft sedimentary rock can also easily be rounded

Fig 24

Glacial landscape of eroded hollows and ridges in the pass over the Latsu massif (Ladake Shan) at 5200 m (cf. Fig 2, No. 16 28°57'N 87°27'E). The Latsu pass is set in a broad trough-like profile that reaches up to 6000 m. The area was completely ice covered. Photo to NW: M. Kuhle.



Tab 1 Table of Snowlines (Equilibrium Line Altitudes = ELA) in S Tibet and on the N Slopes of the High Himalayas

- (1) mountain chain and its highest summit altitude (m)
- (2) locality of the former ice margin
- (3) exposition
- (4) average altitude of the mountain crests (catchment area) (m)
- (5) recent arithmetical snowline ($S_{(m)}$; method: v. Höfer 1879) (m)
- (6) recent snowline corrected according to a thermodynamical upper limit of glaciation (m)
- (7) deepest former (high glacial, Würm?) ice margin (m)
- (8) late glacial ice margin (m)
- (9) holocene (neoglacial and historical) ice margin (m)
- (10) deepest former (high glacial, Würm?) arithmetical snowline ($S_{(m)}$; method: v. Höfer 1879) (m)
- (11) late glacial arithmetical snowline ($S_{(m)}$) (m)
- (12) holocene (neoglacial and historical) arithmetical snowline ($S_{(m)}$) (m)
- (13) high glacial (Würm?) snowline depression (m)
- (14) high glacial (Würm?) snowline corrected according to a thermodynamical upper limit of glaciation (m)
- (15) corrected high glacial (Würm?) snowline depression according to a thermodynamical upper limit of glaciation (m)
- (16) climatic snowline (m)
- (17) Notes

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)
Transhi- malayas near Lhasa 5684	29°43'N 91°04'E	SE SW	5200 5200	5700 ¹ 5700 ¹		4250 3950			4725 4575			1200 1200			4500 4500	recent $S_{(m)}$ not reached
Nien- tschen- tang-La Transh. 6138	29°25'N 89°36'E	S	5400	5750? 5800 ¹		3900			4650			1100?			4500	type of glaciation: ice netw.; transfluence pass with erratics: Schyke or Chalamba La 29°41'N 90°15'E
betw. Latzu & Lulu Tibetan Himal. 6391	28°58'N 87°38'E 28°45'N 87°14'E	N S	5500 5400	5750? 6200 ¹ 6150? 6200 ¹		4200 4350			4850 4875			900? 1275?			5000 4725	Local. near Phu $S_{(m)}$ -depr. 1075? in rel. to climat. $S_{(m)}$ accord. to v. Wissmann - 1475 m
Pang La Massiv Tibe- tan Himal. 5867	28°33'N 87°05'E 28°26'N 87°06'E 28°27'N 87°07'E	N S S	5400 5400 5400				4500 4250 4500		4950 4825 4950						5100 4675 4800	local. near Pang La recent $S_{(m)}$ not reached; local. near Tushidzom local. Nyomdo
	28°27'N 86°08'E	N	7100	6325				5530			6310				6250	$S_{(m)}$ -depress. 15 m 20th century?
	28°26'N 86°08'E	N	7100	6325				5560			6315				6265	$S_{(m)}$ -depress. 10 m 20th century?
Shisha Pangma- N slope plateau; glacier on ice- marginal ramps in foreland	28°26'N 85°45'E 28°26'N 85°48'E 28°27'N 85°48'E	NE NW NNE	6050 6120 6080	5915 5990 5940												present ice margin down to 5780 m present ice margin down to 5860 m present ice margin down to 5800 m

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)	
Lan-Kazi-Massiv Tibetan Himal. 7193 (7223)	28°52'N 90°08'E	N	6430	5705											5855	present ice margin at 4980 m; name of gl.	
	28°53'N 90°08'E	N	6430	5705			VII	4790			5610				5760	"; Kang-Chüng gl.	
	28°53'N 90°08'E	N	6430	5705				4800			5615				5765	S _(m) -depression 95 m "; S _(m) -depression 90 m	
	28°53'N 90°08'E	N	6430	5705			VII	4810			5620				5770	"; S _(m) -depression 85 m	
	28°53'N 90°08'E	N	6430	5705				4830			5630				5780	"; S _(m) -depression 75 m	
	28°53'N 90°08'E	N	6430	5705			VIII	4860			5645				5795	"; S _(m) -depression 60 m	
	28°53'N 90°08'E	N	6430	5705				4870			5650				5800	"; 17th cent.?	
	28°53'N 90°08'E	N	6430	5705			IX	4880			5655				5805	"; 19th cent.? S _(m) -depression 50 m	
	28°52'N 90°08'E	N	6430	5705				4900			5665				5815	"; 19th cent.? S _(m) -depression 40 m	
	28°52'N 90°08'E	N	6430	5705			X	4920			5675				5825	"; 20th cent.? S _(m) -depression 30 m	
	28°53'N 90°11'E	N	5900	5705?			V	4730			5315				5465	recent S _(m) taken from glacier in the west parallel valley (Kang-Chüng glacier)	
	28°53'N 90°11'E	N	5900	5705?			VI	4820			5360				5510	S _(m) depression 337 m	
	28°52'N 90°11'E	N	5900	5705?			VI	4836			5368				5518	S _(m) depression 337 m	
	28°52'N 90°11'E	N	5900	5705?			VI	4870			5385				5535	S _(m) -depression 320 m	
	Everest-N slope	28°25'N 87°00'E	N	7300	6235	5910		4380			5840					5990	local. lower Zambu S _(m) -depression 245 m near Zambu
		28°17'N 86°58'E		7300	6235	5910		4550			5925					6075	
28°15'N 86°48'E		N	7300	6235	5910		4780			6040					6190	S _(m) -depression 200 m opposite to a recent glaciertongue in 5180 = rec. climat. S _(m) at Everest main ridge	
Dzakar Chu, upper Arun valley; 8874	28°15'N 86°48'E	N	7300	6235	5910		4780			6040					6190	S _(m) -depression 200 m opposite to a recent glaciertongue in 5180 = rec. climat. S _(m) at Everest main ridge	
	28°15'N 86°48'E	N	7300	6235	5910		4780			6040					6333		
E of the basin of Man-ko-pa Tib. Him. c. 5800	28°44'N 86°27'E	W	5300	5900		4400			4850			1050			4750	glaciation type: ice stream network; deeper ice margin, covered by sediments possible	
Shisha Pangma-E-SE slopes 8046; middle Bo-Chu (transverse valley)	28°17'N 86°01'E	E	5600				4100			4850					4900	steep hanging	
	28°09'N 85°59'E	W	5800				4100			4950					4900	short valley glacier	
	28°06'N 86°00'E	SE	7100				3670			5385					5450	after Kuhle (1982) a recent valley gl., 35 km long Nylam Phu glacier; near settlement Choksum	
	28°06'N 86°00'E	SE	7100				3380			5240					5305	after Kuhle (1982) a recent snowline difference of 130 (128) m from N to S exists at the main ridge of the Himalayas (Dhaulagiri-Annapurna)	
lower Bo-Chu (Sun Kosi Khola) 8046	27°56'N 85°56'E	S	7100	6250		1600			4450			1900		4310	gorge-glacier up to Friendship Bridge-local. = recent climat. S _(m) at Shisha Pangma main ridge		
															6313		
															-6260		

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)		
Shisha Pangma N slope 8046	28°46'N 86°09'E 28°35'N 85°45'E 28°32'N 85°07'E 28°30'N 85°07'E 28°28'N 85°08'E 28°28'N 85°08'E 28°27'N 85°08'E	NE N N N N N N	6500 7100 7100 7100 7100 7100 7100 7100 7100 7100 7100 7100 7100	5705 6325 6325 6325 6325 6325 6325 6325 6325 6325 6325 6325 6325	•1/2-I I-III II-IV II-V V-VI VI-VII VII-VIII	4500 5015 5210 5300 5415 5455 5495			5500			850? 5375	950 5415 6260 -6313 5993 6090 6135 6193 6213 6233			(4) is uncertain, bec. high gl. ice marg. is 60–70 km away from the highest catchment area S _(m) -depression 170 m S _(m) -depression 125 m S _(m) -depression 67 m boundary of hist. stage S _(m) -depression 47 m 17th cent.? S _(m) -depression 27 m 19th cent.?		
N slope of the whole Lankazi massif 6679	28°51'N 90°11'E 28°52'N 90°06'E	NNE N	5900 6000 6000 6000	5705 5595 5595 5595													5855 5745 5585 5605 5640	present ice marg. at 5510 m; local. 6084 m peak NNE-face; Kang-Chüing-N-gl.; present ice marg. at 5190 m; S _(m) -depression from 160, 140, 105 m;
E slope of the whole Lankazi massif 7193	28°44'N -29°10'N 90°21'E -91°04'E	E	5500 -5855	5745 -5855		4460			4980			765 -875					4980	20 m is the assumed depth of the Yang- Cho-Yung Hu, a glacial piedmont lake but presumably it is much deeper. It is not clear whether it is the deepest part of a high glacial ice margin or the deepest part of a later glacial stage

1) after v. Wissmann (1959): climatic snowline

The present day overall climatical snow line (here used equivalent to equilibrium line altitude = ELA) in S Tibet and the N slope of the Himalayas between 28° to 29°50'N and 85°24' to 91°13'E runs at approx. 5900 m asl based on fifteen values. Assuming that the deepest ice margin positions found in this area belong to the last High Glacial (Würm) the snow line during the Ice Age ran at about 4820 m based on eight values. When, however, an Ice Age *climatic* snow line is extrapolated from these eight values we come down to 4720 m. Thus the High Glacial snow line (ELA) depression was as much as 1080 m or 1180 m respectively.

Fourteen late and post glacial stages were identified although the features were not everywhere found in completeness. The boundary between late and neo-glacial stages is most likely situated at a snow line depression of 67 m to 95 m. The onset of glacial stages in historical times is assumed to coincide with a snow line depression of some 55 m.

Presupposing stable humid conditions a snow line depression of 1180 m would mean a Pleistocene cooling of the warmest month by 7.1° C in S Tibet. However, according to own measurements in this area a gradient of 0.8° C/100 m could be correct, and thus a cooling of about 9.4° C would not be unlikely.

periglacially; 2. their very softness should have encouraged and left behind different forms, such as acute dissection by linear erosion and the outwashing of ravines which, in spite of periglacial influences, had a dominant effect in the more recent past, and continues to do so – if the previous glacial formation had not come down to us (Fig 27).

Sedimentary rock and glacier polish: a comparison with the Arctic

Here, as in numerous other places of the S Tibet ice cover, sedimentary rock indicates a geomorphological landscape character which presents an equivalent to the sedimentary rock of W Spitzbergen (i.e. Dicksonland on Svalbard, cf. Kuhle 1983b) including its post-glacial, periglacial-fluvial reshaping with funnel erosion and box valleys with braided river gravel floors. The only difference is that, in oceanic Spitzbergen, at 79° N, perennial patches of snow occur far below the present glacier tongues, despite a comparable annual precipitation of 250 mm. In subtropical S Tibet they rarely occur a little below or even at the equilibrium line level.

The rough texture of the metamorphic sedimentary rocks in place can be the result of 1) a post-glacial frost-weathering which always has a powerful effect on them in the continental highland. Granites in glaciated knobs on the Kakitu massif in N Tibet, though still beneath ice during the Late Glacial stage, were shown to be affected by considerable post-glacial rock disintegration (Kuhle 1987a, p. 205, Fig 17; p. 231, Fig 32). Sugden and John (1976, p. 196, Fig 10.4) even instance it for less weathering-intensive areas like Scotland, where glaciated knobs have by now become nothing short of anti-flow dynamic castellated rocks. Although not during an initial periglacial fragmentation but at a later point, in the course of a disintegration into rough blocks, the gross voluminous structure of the granite even proves to be of advantage, whereas finely weathered metamorphics retain the large structures more effectively (Fig 29, 41). 2) In the case of the Tibetan glaciation areas there is no need to assume that initially the surfaces of the rock ridges were substantially polished by glacier action; on the contrary, over vast stretches, the ground-ice of the semi-arid and therefore very cold Tibetan ice (cf. above) was firmly frozen to its rock base, so that the traces of abrasion must already be *primarily* rated as small within the specific context of the Ice Age climate.

Observations on Prehistoric Ice Covers between Shigatse (29° 17' N/88° 54' E) and Suo La (29° 09' N/88° 02' E)

Situated at the same level as the Tsangpo valley, the S parallel valley (between the Changma and Silung settlements; see ONC 1:1,000,000, H-9, and Fig 2, no. 18) is filled with large alluvial fans side-valleys and a

broad alluvial floor, now used for agricultural purposes. 70–90 km (by road) W of Shigatse discordantly and shallowly deposited reddish-brown weathered glacio-fluvial – and possibly even glacial – sediments occur on diagenetically consolidated, disturbed and probably Late Tertiary (or even Neogene?) sedimentary rocks. They definitely belong to the Pleistocene and form ridges tens of metres high. At km 93 (29°14'N/88°14'E) and at the altitude of 4160 m the track reaches the level of a terrain of conspicuously softly rounded forms, executed in more or less distinct banks of limestone and to be interpreted as a classical landscape of glaciated knobs. It is essential to note that the landscape has undergone recent dissection into gullies and gorges, which confirms the prehistoric formation of the glaciated knobs.

Here at 88° E the N outlet glaciers of the Tibetan Himalaya stretched down to about 4100 m (Fig 2, no. 18), failing to reach the Tsangpo level by a small amount. This confirms (see above) that, with rising valley levels of the Tsangpo and its immediate parallel valleys further W between 87° and 86° and the existing, correspondingly great altitudes of the catchment areas, the Tsangpo furrow had been reached by the outlet glacier tongues of the ice-stream network and the inland ice, and W of 86° E had also been infilled. This is the area from whence I3 was connected with I2 (Fig 75).

Glacier Traces in the Pang La Massif, South of the Lulu Basin (28° 30' N/87° 07' E)

On the N side of the massif, in the valley descending to the N of the Pang La, covers of rough quartzite blocks reach down to at least 4500 m, and form dumped end moraines (28° 33' N/87° 05' E) (Fig 2, no. 19). The entire broad valley floor up to the pass at 5200 m consists of moraine fields, the substrata of which experienced solifludial shifts of a few tens of metres during the Holocene. With a mean altitude of the catchment area of 5400 m, calculations show an orographic lowering of the equilibrium line to 4950 m, and corresponding ELA depression of at least 800 m. On the S slope moraine deposits in the shape of walls of quartzite blocks tens of metres thick (Fig 28; 28°26'N/87°06'E) and slope dressings with glacial diamictites were found as far down as 4250 m asl (Fig 2, no. 20). This moraine substratum already flanks a somewhat larger side-valley (Fig 29) which the Pang La valley joins, and which in turn reaches the Dzakar Chu at the settlement of Tushidzom. Valleyward findings of erratics 270 m above the valley-bottom line near the Nyomdo settlement confirm the position of the ice margin there. This results in a glacier surface slope which does not allow the glacier to terminate above 4250 m. From the confluence of the tributary valleys at the Nyomdo settlement a mature trough-profile manifests itself (Fig 29). The trough valley leads down from the highest peak (5867 m)



Fig 25



Fig 26

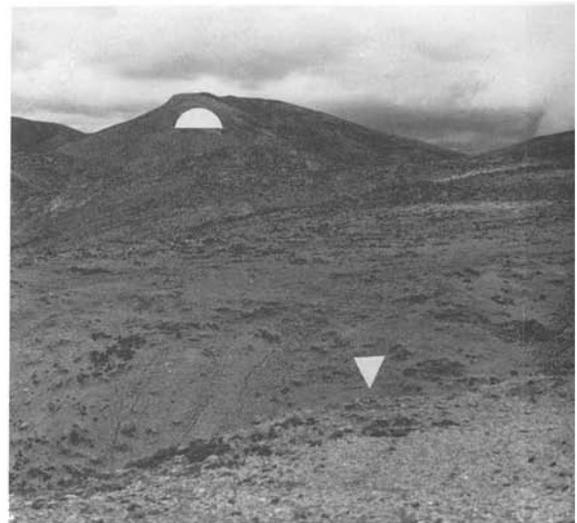


Fig 27



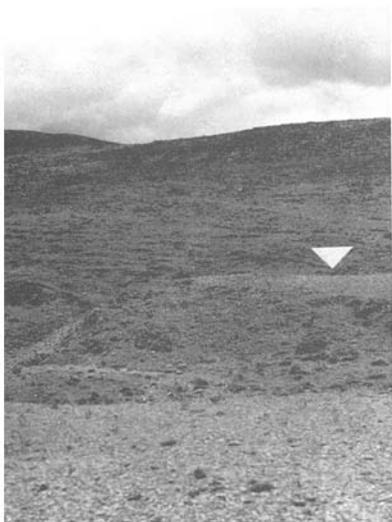
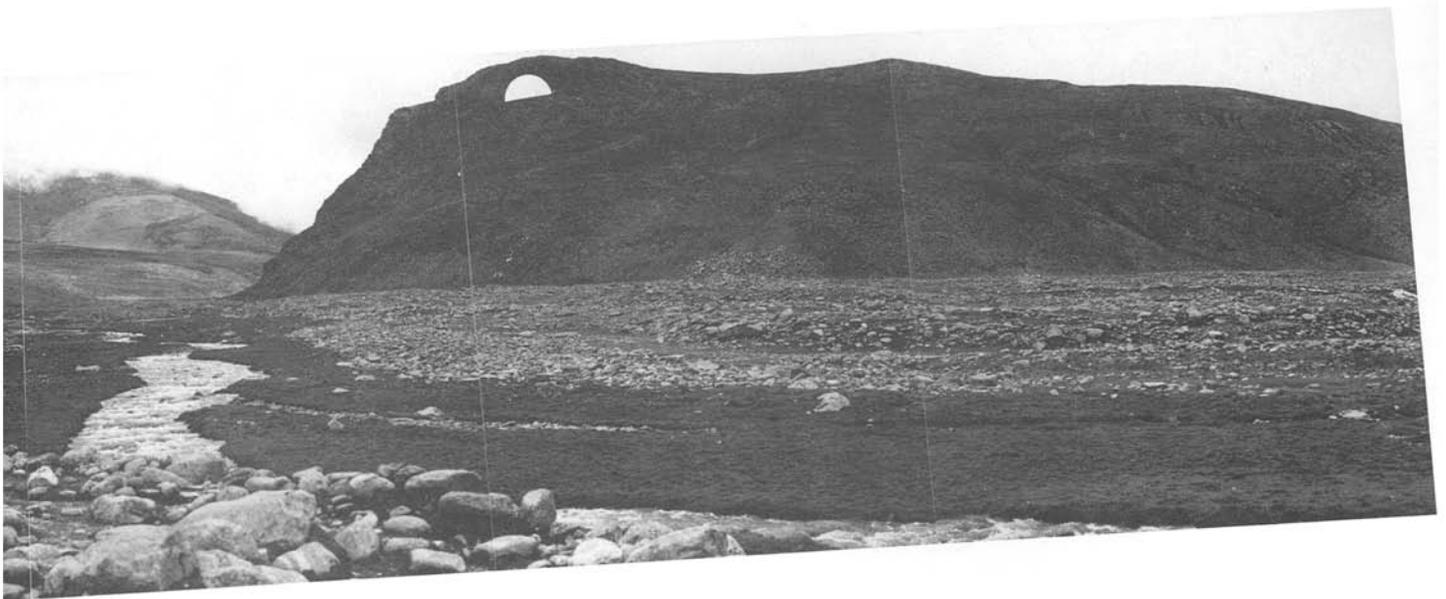


Fig 28

Fig 25

The glacial landscape of the central Mangar valley (Phu valley) in the north slope of the Latzu massif (Ladake Shan 28°54'N 87°39'E; viewpoint 4420 m, centre of panorama up-valley to WSW, cf. Fig 26). The valley floor (fore- and middle-ground) is covered with a granite block loam (ground moraine deposit) of the late-glacial stages I (Ghasa stadium) and II (Taglung stadium). The metamorphic rock ridges here (●●) have the characteristic shape of roches moutonnées and were completely overridden by the ice (Dhampu and Sirkung stadium). They project through ground moraine remnants; (cf. Fig 2 left of No. 16 III, IV). Photo: M. Kuhle.

Fig 26

Looking down the middle valley of the Mangar (see Fig 25) towards the NE from 4700 m. In the middle-ground are late-glacial morainic deposits modified glaciofluvially (▼). III shows the end moraine of the Dhampu stadium (late-glacial) which is also shown in down-valley direction in Fig 25. II marks the remains of a lateral moraine of the Taglung stadium (late-glacial). During the maximum glaciation the area was completely ice-covered which produced the whalebacks (●). Photo: M. Kuhle.

Fig 27

Roche moutonnée landscape (●) at 4500–4700 m on the Solotse La (Suo La) S of the Tsangpo valley (Fig 2, No. 17; 29°04'N 87°59'E). Periglacial solifluction sheets lie on the glacially scoured weak metamorphosed sedimentary series. Today they are dissected by deeply incised gullies (▼▼). Photo: M. Kuhle.

Fig 28

View from 4400 m to the NNE (centre of panorama) upwards to the area of the confluence of the two component streams of the Ice Age Dzakar Chu glacier (Fig 2, No. 20; cf. Fig 29); right is the settlement Nyomdo, left are lateral moraines with large blocks of quartzite (●). In the background similar erratic blocks overlie the in situ flysch series (↓). The true left valley side is glacially scoured, the etchings being smeared with ground moraine (▼). Glacial striations occur generally, as well as on the true left side (●). Photo: M. Kuhle.



Fig 29

Fig 30



Fig 31





Fig 29
Panorama (centre to the E) from 4400 m (Fig 2, No. 20; 28°27'N 87°09'E; cf. Fig 28). The valley of Nyomdo is a glacially formed trough valley. The fields of the settlement are located on impermeable ground moraine (◆◆). The valley is picked out by well-preserved striations along sedimentary layers (→). The smoothing of the surface (▲) up to and over the surrounding heights indicates a complete covering of glaciers of the 'Panga La massif'. Photo: M. Kuhle.



Fig 30
From Panga La (28°30'N 87°06'E) at 5200 m in the S Tibetan mountains to the S to the Himalayas and Mt. Everest. In the middle a large open corrie (●) is cut into metamorphic rocks N of Dzakar Chu (Fig 2, No. 19). The recent glacier tongues (●●●) lie only 1000 m above the valley floor of the longitudinal Dzakar Chu where it is most deeply incised (⚡). As the results of the (reconstructed) depression of the equilibrium line altitude by about 1200 m at the glacial maximum the relief was then completely full and overflowing with an ice stream network flowing from the Himalaya (Fig 75 and 76, I3). Photo: M. Kuhle.

Fig 31
View from 4430 m towards the ENE (centre of panorama) to the W slope of the massif in the E of Man-ko-pan (28°44'N 86°27'E; Fig 2, No. 21). Both the trough valley (above ●) and the rounding of the summits show that the relief has been completely covered by a network of ice streams or an ice sheet (Fig 75 and 76, I3). Small ice streams flowed from the hanging trough valleys (above ●) to combine in the basin of Man-ko-pan (foreground). Photo: M. Kuhle.



Fig 32 From Tschü Sü La (4880 m) above the Yang-Cho-Yung-Hu (Yamdrok lake) towards the WSW to the 7193 m (7223 m) Lankazi massif (28°56'N 90°07'E; Fig 2, Nos. 24, 25); (---) shows the recent orographic snow line (equilibrium line altitude); (●●) mark the recent glacier tongues (in the ablation zone) facing E. The lake (at 4480 m) is a postglacial filling of a glaciated surface (shown by rounded hills ●, and corries ×). This is shown by the steep shores freshly cut into the slopes (↘). Photo: M. Kuhle.

of the Pang La massif. Its profile polishes metamorphic stratified rock. In the valley running S from the Pang La, as well as in its tributary valleys, broad, flat, corrie-like glacier beds have been formed (Fig 30, foreground). Flysches with marls and argillaceous schists occur with vertical stratification. They have been smoothed by basic glacial polishing across the outcrop curvatures and are superimposed upon by ground moraine tens of metres thick (Fig 28, background). These are drifts of large, rounded quartzite blocks with a crust of ferro-manganese (up to $2 \times 3 \times 3$ m), which are separated by a fine, sandy, at times also clayey matrix, lilac blue in colour. Well-rounded blocks occur beside others with rounded edges. A solifluidal layer of moving drift has formed within the uppermost metre of the moraine. On the approximately 5500 m high arêtes and peaks, the affected light-coloured quartzites occur. Having filtered through from higher altitudes, or been broken down into block covers, they occur in sections, superimposed upon partially eroded soft flysches as relatively weathering-resistant residues. The moraine blocks were picked up by the glacier at this point, and transported well over 16 km away. Besides the lowest, distinct ice margin position at 4250 m (see above), another one had been formed at 4500 m, so that equilibrium lines at 4825 m and 4950 m result. The higher one is classified as belonging to the Late Glacial stage (Ghasa Stadium I) (see Tab 1). Since they occur on southerly aspects, the appropriate climatic ELAs are to

be stated as occurring 150 m lower at 4675 and 4800 m asl.

Due to the absence of present glaciation in the Pang La massif it is difficult to estimate the recent climatic ELA. However, it does run S of the Dzakar Chu at about 6000 m (Fig 30, background), so that 5900 m constitutes a likely figure, which points to an Ice Age ELA depression of 1225 m (5900–4675 m) and a Late Glacial one (Stage I) of 1100 m (5900–4800 m).

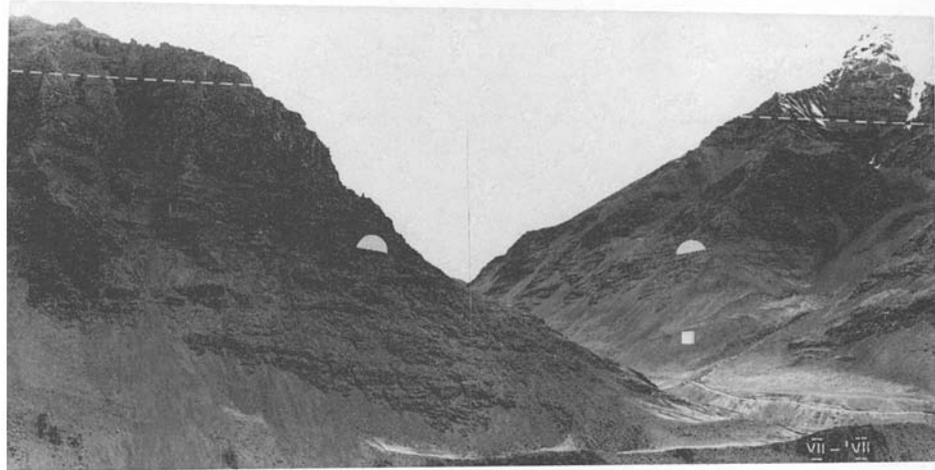
Greater equilibrium line depressions, i.e. lower ice margin positions, cannot be dealt with within the representative section of the Pang La massif under investigation here, since at the altitude of 4250 m the Dzakar Chu, a W branch source of the Arun valley (main valley running in a W–E direction) has been reached. This valley has been glaciated from the Himalayan N slope (Mt. Everest group; see below; Fig 30, background) during the Late and High Glacial Period. This implies that there was an ice-stream network without further ice margin locations N of the Himalayan ridge.

Traces of Glaciation in the Massif East of the Man-ko-pan Basin (28° 44' N/86° 27' E)

This c. 5800 m-high massif N of the Tingri basin is now likely to be entirely free of glacier (Fig 31), or at most occupied by very small corrie glaciers and perennial snow patches. The recent climatic equilibrium line thus runs at 5900 m (no higher, however). During the Ice Age there was glaciation of the type of an ice-stream network which filled the entire relief; all valley troughs are evidence of this (Fig 31). At the valley exits to the S and W ground- and end moraine ramps take over (Fig 31 ●●). There is a similarity between these and remnants of ice marginal ramps (cf. Fig 41) and pedestal moraines (dam moraines), on which the glacier builds up its own drift foundation by transporting and depositing lateral moraines (Kuhle 1982, vol. 1, p. 84, vol. 2, Fig 82; 1983a, p. 238). The trough-profiles extend down to 4600 m asl, and at 4400 m the moraine ramps become submerged below the more recent (recent Late Glacial?) glacio-fluvial drift sediments of the Man-ko-pa basin or the Tingri basin (Fig 31, middle ground; Fig 2, no. 21). With the mean altitude of the catchment area at 5300 m, the orographic equilibrium line on the W side is calculated as being at 4850 m, in accordance with a climatic ELA around 4750 (Tab 1). It has hitherto not been possible to prove in situ that this is a Late Glacial lowering of the equilibrium line, though the author records it as likely as the basins of Tingri and Lulu which join on in the E must have been filled with ice during the Ice Age (cf. above), and this catchment area is situated here at a similar altitude. Moreover, the ELA depression of 1050 m during the maximum Ice Age would be disproportionately low in comparison with lowerings in the vicinity.

Fig 33

View from Kang Chüng valley at 5200 m of the main valley leading down eastward (right) from the Lang La (Lankazi massif 28°57'N 90°11'E). At the glacial maximum this accidented area was filled up to the summits with a network of ice streams at least 1000 m thick (Fig 75 and 76 I3). This is shown by the lateral striations on the laminations of the metamorphic rocks (●). (---) indicates the minimum thickness of the ice. VII-VII marks the moraines of the Kang Chüng glacier known historically; these are younger than 790 ± 155 and older than about 320 years. (■) shows other moraines known historically and up to neoglacial age (younger than 4160 and older than 320 years). Photo: M. Kuhle.



The Glaciation of the Lankazi Massif (28° 56' N/90° 07' E; Tab 1)

On the N-facing Kang Chüng glacier in the Lankazi massif (Fig 32) which rises to 7193 (7223) m in the 5150 m high area of the Lang La (28° 57' N/90° 11' E) in the Tibetan, or Inner Himalayas, it was possible to distinguish at least nine historical glacier locations, all of which were more recent than 790 ± 155 BP (Fig 33). All these stages occurred within a maximum equilibrium line depression of less than 100 m as compared to the glacier tongue of the time, which does not descend further than to 160 m above the lowest of these moraine formations (Tab 1). In accordance with the Kuhle nomenclature (1982, p. 118) they are classified between Stadium X to

the more recent Dhaulagiri Stadium (VII) and down to the Middle Dhaulagiri Stadium (VII).

Neo-Glacial moraine chains were observed and mapped as extending into the mouth of the main valley of the parallel E valley (feeding into the Lang La valley, which descends in an easterly direction) (Fig 34, 35). They occur as far down as 4730 m (Fig 2, no. 22). The unnamed high peaks of the immediate catchment area rise to 6600 m, and the mean altitude of the ridge surrounds will be about 5900 m. The equilibrium line of the N face accordingly ran at about 5315 m, and at the time of that ice margin position thus 390 m lower than the recent equilibrium line (which is at about 5705 m). With reference to this equilibrium line depression a classification as Neo-Glacial or late Late Glacial is

Fig 34 A valley E of and parallel to the Kang Chüng valley seen up-valley towards the S; at its end is a completely glaciated peak at some 6600 m (left background: Fig 2, No. 22). Viewpoint at 4960 m. The sides of the valley here, as in all valleys parallel, are glacially scoured to considerable heights (●, cf. Fig 33, 35, 36, 38, 39). The well-developed lateral moraine terraces with sporadic alpine vegetation on each side of the centre line of the valley and the end moraines (foreground) are from a neo-glacial stadial position (Nauri stadium V or Older Dhaulagiri stadium VI ?). Photo: M. Kuhle

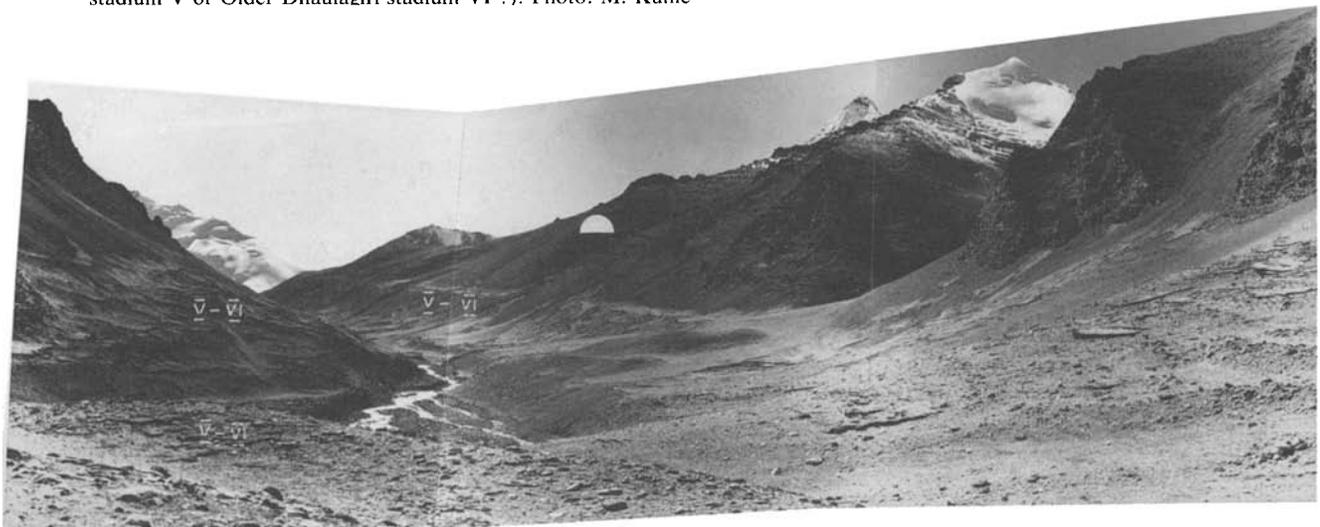




Fig 35 The valley parallel to the Kang Chüng valley to the E seen from 5300 m (see Fig 34; 28°54'N 90°10'E) looking down-valley to the N. The high peak (◆) reaches 6500 m. (▲) shows the remains of the lateral striations on the laminations of the easily weathered metamorphic rocks. The bluffs and arêtes (▼▼) are the remains of the earlier slopes of this trough valley landscape eroded by ice in maximum glaciation times. The moraines in the valley floor are of neoglacial age (V–VI). (↓) shows the location of C14 dating (790±155 BP; Fig 2, No. 22 ○) in the main Lang La valley. The neoglacial end moraines (×) are at 4730 m. Photo: M. Kuhle.

appropriate. It has not been possible to establish the minimum age of the moraines with the aid of C-14 analyses. Under the cold-arid conditions, there is a lack of sufficient organogenic substratum. At the exit of the E parallel valley of the Kang-Chüng North Glacier to the Lang La valley (28° 53' N/90° 11' E) it was possible to date one of the samples, taken at a depth of 0.6 m from a few metres-high outwash plain terrace, at 790±155 BP (14C content 90.6±1.7% modern; analysis by M. A. Geyh, 30. 5. 85) (Fig 2, no. 22; Fig 35 ◀). Both the position of the ice margin down valley at 4730 m and the ELA depression of 390 m are thus shown to be older, and therefore classified as Nauri Stadium V (Kuhle 1982, p. 159, 160). This also provides a clue for the fact that the ELA depressions of 390 m and more are of Neoglacial age at least (for Himalaya cf. Kuhle 1986d, p. 438–454; 1987b, p. 200–205). The formation of humus from alpine turf which has been dated here may admittedly be significantly more recent than the ice margin position in question, for the amount of shifting in the area affected by the confluence of the alluvial plains of two valley glaciers is considerable (Fig 36). It is true that this is the body of a terrace the top of which must have been dispensed within the course of the dissection that goes with the retreat of a glacier, and thus from further interference and the opportunity of providing the most favourable locality for taking samples. The ice margin positions of 4820 m asl, with ELA depressions of 345 to 320 m, are situated further up-valley from the sample locality, i.e. higher, so that they must be classified as probably more recent and thus historical. In principle, however, it cannot be ruled out that they are much older than the formation of soil and humus in question, since soil destruction may not only be due to a crossing glacier, but also to meltwater erosion. For this

reason, though also because of equilibrium line depressions around 300–350 m, they are being classified as three stages of the older Dhaulagiri Stadium (VI) (Kuhle 1987b, p. 205).

W of the Lankazi massif and the Lang La there is a very large and more than 10 km-wide high valley with a valley floor between 4400 and 4800 m (Fig 37). It runs in the direction of the Chiang-Tzu settlement and is enclosed by more than 6000 m high mountain massifs which are currently glaciated. Included amongst these are the main peaks of the Lankazi massif in the region of the Lang La (Fig 37). All the main valleys of the Lankazi massif (where the Lalung and Sutu settlements are located), which run westwards towards the large valley, bear characteristics of relief-filling glaciation. These include glacial scouring on outcropping layer edges in easily splintered metamorphics, as well as glacial grooves several hundred metres above the 4800–5100 m high valley floors (Fig 33–36, 38, 39). Although there has been no certain evidence that also the large longitudinal valley had been completely filled by glaciation, it is highly likely considering its altitude. From both the flanks, large, flat alluvial fans descend (Fig 38), like those observed in glaciated longitudinal valleys of Würm age in Alaska (i.e. the McKinley river N of the Alaska Range, 63°30'N/151°W), in W Greenland (Sarqaqdaalen, Nugsuaq, 70°03'N/52°W; Kuhle 1983c) and in Dicksonland (Western Svalbard, e.g. Idodalen, 78°03'N/52°W; Kuhle 1983b). During the Ice Ages, ice in valleys of this kind would build up to thicknesses of several hundred metres without leaving behind striking glacial traces in the metamorphosed sedimentary rocks, apart from polished triangular areas between the exists of side-valleys and softly-rounded forms (Fig 37 ●●; Fig 2, no. 23).

Fig 36

Looking down the Lang La valley from 4800 m (Lankazi E slope, cf. Fig 33, 35 (in distance), 39). The cross profile of the trough valley is box-like. Above the ground moraine there lie the up to 790 ± 155 year old outwash deposits of the historic glacier (foreground); the river has cut down 1–2 m into the gravels (\downarrow). The valley sides are glacially scoured (\blacktriangle). Photo: M. Kuhle.



The E slope of the Lankazi massif is characterized by prehistorically totally glaciated, glacialic box valleys with distinctly polished flanks (Fig 33, 35, 36, 39). As in the case of the Alps, Late Glacial and High Glacial glacier-levels cannot be differentiated with precision, but a surface of an ice-stream network at altitudes of around at least 5500 to 5750 m has been established by ridges that have been polished into roundness and intermediate valley divides with upper ice-scour limits (Fig 33, 34, 39 -----). This implies ice thicknesses of about 1000 m. Higher peaks like the Kang Chüng, the Lankazi peak and other high mountains (Fig 33, 35, 37) reach altitudes of 6500 to 7193 (7223) m, and rose another 1000 m above the surface of the ice-stream network. In the opening of the large valley which descends in an E direction and is being accommodated by the track from Lang La, there is a large moraine fan of ground moraine at 4530 m ($28^{\circ}55'N/90^{\circ}24'E$; Fig 2, no. 24). It is tentatively being attributed to the more recent Late Ice Age (Sirkung (IV), Dhampu (III), or Taglung Stadium (II)). As a valley lake, the wide and branching Lake Yamdrok (Yang-Cho-Yung Hu in Chinese; Fig 32, Tab 1) to the E of the Lankazi massif has a certain similarity to the Alpine Vierwaldstätter Lake. It is interpreted as a lake in a terminal basin of the High to Late Glacial genesis (Fig 2, no. 25). Its level is at an altitude of 4480 m, and assuming a minimal lake depth of 20 m, it would have required an ELA depression of 765–875 m in order to have a glacier reach the lake floor. This ELA depression rather suggests a Late Glacial (Ghasa Stadium (I)) rather than a High Glacial classification (Tab 1). The post-glacial filling up of a lake in a former terminal basin is confirmed by the limnic undercutting of the shore with the subsequent formation of striking steep banks (Fig 32 $\blacktriangledown\blacktriangledown$) as a new, previously not yet established formation element. Gentle glacialic slopes have been worked upon concordantly by interglacial and post-glacial periods of solifluction (Fig 32 $\bullet\bullet$), and have now been given a new formation of a different kind by the undercutting of the lake shore.

The Lankazi E face with an equilibrium line at around 5800 m (Fig 32 ----) and an upper glaciation line at the level at which a shallow relief sets in (Fig 32 $\bullet\bullet\bullet$; cf. also Fig 30 $\bullet\bullet\bullet$); make the exponential glacier increase clear. Even an ELA depression of only 500 m brought the equilibrium line to 5300 m asl, the glaciers then terminated only 300 m above the level of the lake and the gain in ice area was considerable. It is the altitudinal interval within which large, bowl-shaped corrie basins were polished out (Fig 32 \times).

Lowest Ice Margin Locations and Equilibrium Lines in the High Himalaya

Prehistoric glaciation of the High Himalaya cannot be separated from that in S Tibet and in the Tibetan Himalaya N and E of the Shisha Pangma massif. The two will therefore be treated together.

The Glaciation of the Shisha Pangma Massif with Bo Chu (Sun Kosi) and its Catchment Area (Tab 1)

In the course of our joint 1984 expedition Zheng Benxing included the Shisha Pangma N slope in its 'Geomorphological Map of the Mount Xixabangma Region' (1988), and the author explored it as well (Fig 40, 41). The states of glaciers recorded by the two fieldworkers proved to be identical. Apart from the nomenclature, differences occurred in the estimates of the age of glacialic diamictites.

Some notes on the nomenclature and age classification of the stadia

Zheng Benxing designed a local nomenclature of the state of glaciers, with the massifs of Shisha Pangma and Mt. Everest in mind. Here the author attempts to extend his exemplary Dhaulagiri-Annapurna-Himal nomenclature so as to include Tibet and its adjacent mountain



Fig 37

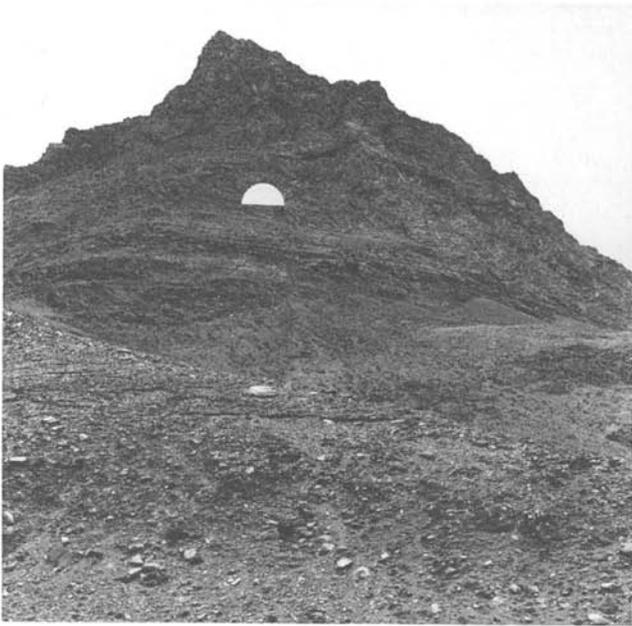


Fig 38



Fig 39



Fig 40

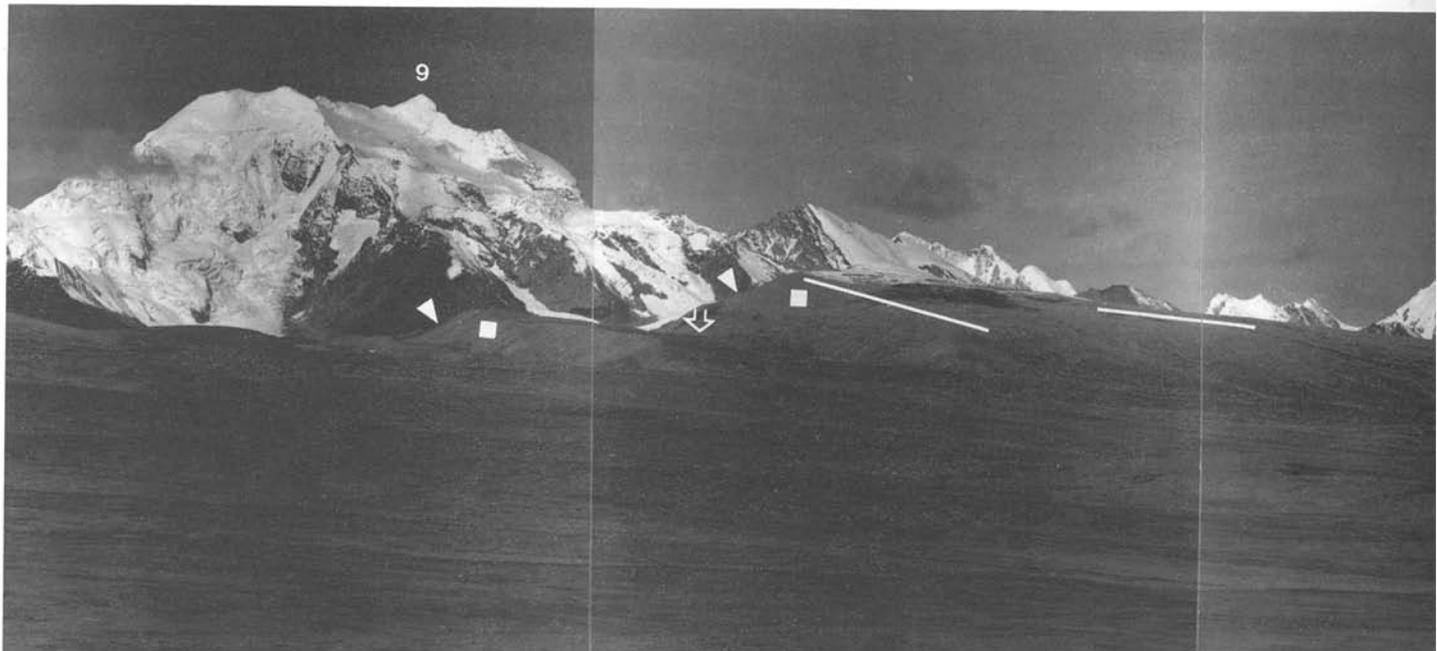




Fig 41

Fig 37
View looking S (centre of panorama) from 4630 m in the large E–W running longitudinal valley that leads from the Lankazi massif (28°51'N 89°54'E; Fig 2, No. 23). The glaciated peak at c. 6600 m to the left lies E of Lang La (c. 5150 m). The view right shows the broad trough-like valley contours towards the W. The valley floor is filled with late-glacial and postglacial alluvium (with portions of outwash ◆ and historical alluvial fans (▼)). The alluvium lies above lodgement till. The slopes show glacial striations (●●) on the weakly resistant metamorphic rocks. These are modified but not removed by solifluction sheets and gully wash (↓). The aspect of the relief recalls that of the ice sheet eroded region of West Spitsbergen or e.g. that of the Kluane mountains in the West Yukon (Canada).
Photo: M. Kuhle.

Fig 38
Detail of the state of preservation of features produced during the maximum and late-glacial. Sculpturing of strata edges (●) on finely stratified and thus easily weathered crystalline schists (phyllites). True left tributary valley of the Lang La valley (Lankazi massif; Fig 2, between No. 23 and 25; 5100 m). Photo: M. Kuhle.

Fig 39
Glacial shaping of the upper Lang La valley viewed toward the W. (Lankazi massif; viewpoint 5050 m; Fig 2 E of No. 23). The lineations of glacial scouring are well preserved in spite of the slight resistance of the rocks. This shows them to be young and thus probably late-glacial in origin. The level of the ice during the glacial maximum lay above the rounded rock bosses (---). The lateral kamiform terrace (◆) must therefore have been formed along the young, late-glacial valley glacier.
Photo: M. Kuhle.

Fig 40
N slope of Shisha Pangma (No. 8) seen from 5250 m (Fig 2, No. 26 28°23'N 85°47'E) on the S Tibetan plateau. The high plateau in the foreground is in direct contact with the High Himalayas without interposition of the Tibetan Himalayan mountains. I–VI shows late-glacial to neoglacial stadal positions of the Yepokangara glacier which flowed from Shisha Pangma. Its tongue ends today at the foot of the 8046 m mountain (↓; cf. Fig 42 ×). The moraines I–VI deposited between the lateglacial piedmont glacier tongues are marginal sandr (ice marginal ramps = IMR, × and I). The IMR were raised above the equilibrium line in the post-glacial so that today they have individual ice patches (× and I). This new glaciation shows that Tibet has been uplifted more rapidly than the Himalayas because the recent tongues of the Yepokangara glacier (↓) no longer reach these lowest end moraines.
Photo: M. Kuhle.

Fig 41
The N slope of the 7299 m Kang Benchen (No. 9) photographed towards the W (centre of panorama) from 5250 m. Its glaciated slopes lead down over 1700 m vertical distance to the High plateau of S Tibet (Fig 2, N of No. 27; 28°40'N 85°34'E). The recent (↓) glacial tongues reach exactly to the base of the marginal sandr (ice marginal ramps or IMR ■■■). These were deposited in the late-glacial. Their ramp-like contours (▼) with a steep front slope facing the mountains (—) are characteristic of such glacial and glaci-fluvial outwash features of semi-arid piedmont glaciers. Photo: M. Kuhle.

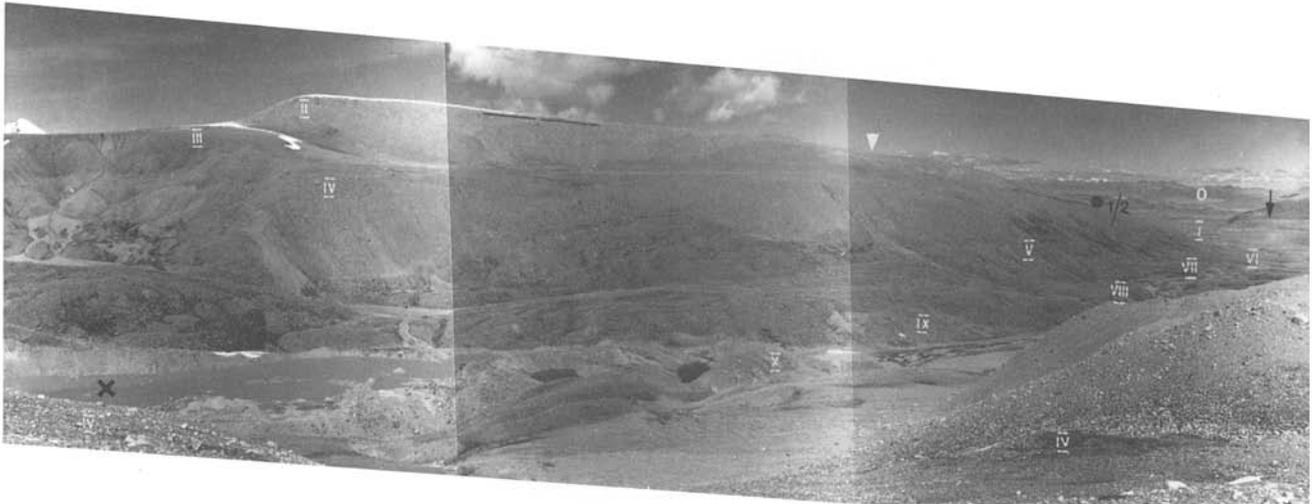


Fig 42 Marginal sandr (ice marginal ramps = IMR, —) seen in cross section from 5820 m towards the W (centre of panorama). These late-glacial (I–IV) and as recent as historical morainic ramps (V–IX) dominate the valley floor (×) by more than 500 m. × marks a glacial lake in dead ice on the tongue of the Yepokangara glacier of Shisha Pangma N slope (Fig 2, No. 26; cf. Fig 40 ◀). The checker-board intermixing of kamiform features and lateral moraines of progressively younger ages is characteristic. This results from the progressive downward contraction and narrowing of the glacier tongues between the marginal sandr deposits (see the succession from II, ● ^{1/2}, III, IV, V, IX, X). (○, extreme right) marks the area of the ice sheet engulfing the relief (↓), an area of classical roches moutonnées. Photo: M. Kuhle.

systems (Kuhle 1980, 1982, 1983a, 1986d, 1987b). In what follows a comparison of the two systems will be presented.

Zheng Benxing

Highest moraine platform of Xixabangma glaciation (middle Pleistocene?)	_____
Higher moraine platform of Nyanyaxungla glaciation (middle Pleistocene?)	_____
Oomolangma glaciation I (Fuqu glaciation I)	_____
Oomolangma glaciation I (Fuqu glaciation II)	_____
Oomolangma glaciation II (Pulao glaciation)	_____
Neoglaciation (all stadia up to the sub-recent terminal moraines are included)	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____
“ ”	_____

M. Kuhle

Ghasa Stadium (I) (Late Glacial)	_____
Taglung Stadium (II) (Late Glacial)	_____
Dhampu Stadium (III) (Late Glacial)	_____
Sirkung Stadium (IV) (Late Glacial)	_____
Nauri Stadium (V) (Neo-Glacial, c. 4000–5000 BP)	_____
Early Dhaulagiri Stadium (VI) (Neo-Glacial, c. 2000–2500 BP)	_____
Middle Dhaulagiri Stadium (VII) (Neo-Glacial, more recent than 2000 BP)	_____
Late Dhaulagiri Stadium (VII) (c. 440 BP)	_____
Stage VIII (c. 320 BP)	_____
Stage IX (more recent than 320 BP)	_____
Stadium X (c. 80–30? BP)	_____
Sub-Recent – Recent Stadium (30–0 BP)	_____

In spite of a subdivision into 13 stages, the author's scale still proved to be too small in the forefield of the recent glaciers of the Shisha Pangma N slope (Fig 2, no. 26). It is even possible to establish altogether 16 to 18 steps and stages (Fig 40, I–VI). This applies to the forefield of the Yepokangara Glacier (Fig 42, O–X), the immediate catchment area of which includes the Shisha Pangma main peak.

The author differs from Zheng Benxing essentially in that his assessment of the time for the stage process in the Shisha Pangma foreland is different. There are no absolute data for the area concerned. It is too cold and arid to produce sufficient quantities of organogenic substratum for radio-carbon analysis. Whilst the Chinese age-dating attributes the oldest moraines to the Middle or even the earlier Pleistocene, the author – assessing the freshness of forms – assumes an at most Late-Glacial age, i.e. c. 15,000 BP (Ghasa Stadium I and younger). All the more recent oscillations in the course of the glacier retreat have, accordingly, taken place in post-Glacial times. This agrees with the glacier retreat of the past 13,000 years given for other areas in High Asia (Karakoram, Quilian Shan, Kun Lun, Himalaya) Kuhle 1986, 1987 a, b).

Outside the late-glacial glacier margins mentioned above, which descend at most down to only 5015 m asl (Ghasa Stage I, cf. Tab 1; 28°35'N/85°45'E), marked and well-preserved forms of glaciated knobs occur widely (Fig 42 ▼; Fig 2, no. 26). They rise to fully 5250 m asl, and are evidence of a complete inland glaciation, which subdued the S Tibetan relief. It must be attributed to the last Glacial Maximum of the Ice Age, 18,000 to 30,000 years ago (Fig 42 ○). In accordance with the great altitude of the plateau of S Tibet, which does not fall

below 4500 m anywhere here, a maximum ELA-depression of 850 m to 5415 m can be ascertained (Tab 1; 28°46'N/86°09'E). During the pre-Ghasa stagnations (● 1/2), two glacier locations, which are to be found entered between the High and Late Glacial (Kuhle 1982, p. 153), the glacier tongues which had moved furthest to the N from the High Himalayas down to the plateau, terminated in the tongue basin of the Paiküco (Peiku Tso) lake at about 4500 m asl (Fig 42 ● 1/2; Fig 2, no. 27). They present stages of the retreat from the situation of the all-covering inland ice to a Himalaya N slope-piedmont glaciation that remained into Late Glacial times.

The complex of problems concerning the ELA-depression and glacial isostasy on the Shisha Pangma

Compared with equilibrium line depressions of the Late Glacial stages of the Himalayan N face, which attain 700–1250 m (Kuhle 1982 p. 154 et seq.), those of the Shisha Pangma N side are very small. The Ghasa Stadium (I) thus came up with an ELA-depression of from 6325 m to 6060 (6058) m, i.e. of only 265 m (Fig 42, I).

To begin with, this results in objections to the age of the moraines, establishing them as being still more recent than Late Glacial and possibly even as Neo-Glacial (cf. Kuhle 1986d, 1987b). In any case their age can by no means be interpreted as High Glacial or even as Middle Pleistocene, as Zheng Benxing tries to do (see above). Thin section analyses of moraine blocks are evidence even of what are, microscopically, clearly visible iron hydroxide formations which follow hairline cracks as lines. Blocks of coarse-grained Shisha Pangma eye-gneiss were selected (Fig 43). Under the cold and arid conditions the formation of weathering lines of this kind requires more than the 2–4 Ka of the neo-Glacial

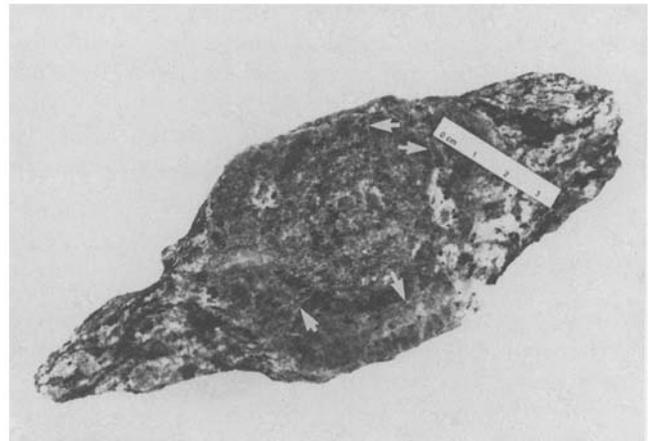
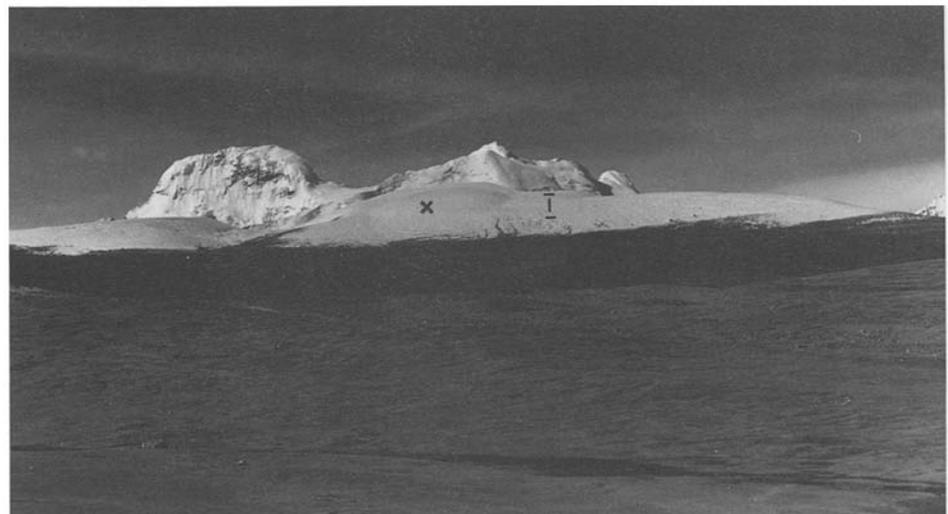


Fig 43 Polished sectioned hand specimen of Shisha Pangma augengneiss. The rock portion is a morainic component of the late-glacial ice-marginal ramps in the Shisha Pangma N slope (locality ● 1/2, Fig 42). Weathering discloses streaks of iron-hydroxides (↓) produced over 9000 to up to 15000 years. They may be followed between the crystals of albite oligoclase and potash-feldspar. The light weathering streaks (↓) are positioned on the cross-wires. Laboratory photograph: M. Kuhle and J. Peters.

fluctuations of the glaciers. An age of 9–15 Ka, however, seems to be right (see below). This raises the question of how it is here that in the immediate vicinity of ELA-depressions of up to 1250 m (on the S side of the Himalaya) of the same age, the Late Glacial (the Ghasa Stadium I, for instance) reached no more than 200–300 m. The answer to this may be helped by the observation that some of the terminal moraines from the Late Glacial are now covered by a glaciation of their own, i.e. are now within or even above the present level of the equilibrium line (Fig 40 × and 44 ×). Evidently the post-Late Glacial uplift has subsequently pushed up the terminal moraines of the rising equilibrium line so

Fig 44 N slope of the western Shisha Pangma mountains (the nameless 6500 m peaks between Shisha Pangma and Risum to the W: 28°38'N 85°39'E, Fig 2, N of No. 27). End moraine of the marginal sandr (IMR) type of the late-glacial (Ghasa stadium I). The surface of the moraine has been lifted above the equilibrium line in the post-late-glacial period so that an hour-glass-like plateau glacier has been produced (×). This uplift is the result of a very intense glacio-isostatic unloading as the Tibetan ice sheet melted. Photo: M. Kuhle.



that the recent glacier tongues terminate fully a mere 500 m above the lowest end moraines of the Ghasa Stage (I) (Fig 42, distance of recent dead-ice lake \times to I). This leads to the postulate of an uplift since the Late Glacial period, which raised the terminal moraines by several hundred metres, i.e. by perhaps 600 m within 15 Ka, for it is accepted that

$$ELA_{\text{Depr.}} = \frac{t_p - t_i}{2} \text{ (m asl),}$$

with t_p representing the recent termination of the glacier tongue and t_i for that of the prehistoric one. This is equal to an ELA-depression of 500–600 m at that time ($600:2 = 300 + 265 = 565$ m). This is much too low for the Ghasa Stadium (I), which tends to show depressions clearly above 1000 m (Kuhle 1982). This amounts to an indication that either the assumed uplift is still being set too low, or that the lowest immediate moraines of the piedmont glacier on Shisha Pangma belong to distinctly younger Late-Glacial stages than had hitherto been assumed (as shown in Fig 40, 42 and 44). A clarification of this problem calls for the investigation of the cause of this evidently very considerable average uplift of some 40 mm/year or more, and one that is three to eight times greater than Gansser (1983, p. 19, and an oral communication in January, 1983) states for the High Himalaya. Evidence for the much more rapid uplift of the Tibetan plateau, as compared with the 8046 m high massif during the Holocene, can be found here on the Shisha Pangma. Although the large terminal moraine wedges or broad moraine ridges (Fig 40, I and 44, I) of the ice marginal ramps type (ice marginal ramps = IMR; see below) have been raised above the equilibrium line and are superimposed upon by a glaciation of their own (Fig 40 \times and 44 \times), the mountain glacier tongues (Fig 40, \downarrow), the predecessors of which deposited the terminal moraine ramps in prehistoric times, no longer reach the lowest ice margin locations (Fig 42, I). It follows that the Shisha Pangma massif, the catchment area of the mountain glaciers, was not uplifted in the same way as the terminal moraines. The tongue of the Yepokangara glacier (Fig 40 \downarrow) has now retracted 15 km from these terminal moraines.

The finding of the Eohippus, a denizen of the lowlands, at an altitude of 4900 m ($28^{\circ}33'N/86^{\circ}10'E$), E of the Shisha Pangma, provides general evidence for an uplift of S Tibet of about 5000 m during the past 40 million years. However, at the same time the High Himalaya rose faster than the Tibetan plateau over a longer period of the Late Tertiary and Pleistocene. The antecedents of the Himalayan transverse valleys, which originated in Tibet and – synergetically with the uplift of the Himalaya – cut through the main range, are evidence of this (Hagen 1968; Kuhle 1982, p. 18, 20). How, then, can the suddenly accelerated *post-glacial* uplift be explained in an otherwise constant geological-tectonic situation? Had it not be confined to the post-glacial period only, but persisted over prolonged periods

of the Quaternary, the altitude of the Himalayan peaks would already have been surpassed. The hypothesis of glacial-isostatic uplift offers an explanation – precisely also because of the extremely high rates of uplift (see above), which are otherwise unknown in primary vertical tectonics. Creating an increasing impact, the inland ice on the Tibetan plateau achieved a much more remarkable thickness than the ice in the mountains on its fringe, such as the Himalaya, where the comparatively narrow valley glaciers flowed steeply down to below 2000 or 1500 m asl, losing much of their thickness. This explains the much more significant glacio-isostatic deformation, i.e. the greater uplift as compared with the Himalayas which goes together with de-glaciation. The once faster tectonic uplift of the Himalaya, like the compensatory glacio-isostatic uplift of the S Tibetan plateau edge will have taken place at that time along the same N Himalayan marginal fault as a slip plane.

By contrast with the Himalaya, morphological features of such forced plateau uplift are stepped accumulations of ice-marginal ramps, which slope towards the mountains (Fig 41 \blacktriangledown). The amounts of uplift previously deduced are limited: partly by the extent of the ice burden in the Main Ice Age through the thickness of the ice. In turn, this is topographically limited by the ice overspill into the south slope of the Himalayas, i.e. by the lowest passes plus several hundred metres. But the large Himalayan gaps like the E Bo Chu (Sun Kosi or Po Ho), or the adjacent Chilung Chu (-Ho) in the W have also by way of outlet glaciers contributed to the lowering of the inland ice level.

A glacio-isostatic depression of 600 m implies a thickness of about 1800–2000 m ($0.9:3-3.3$ [g/cm^3]) of ice. That postulated 600 m of uplift may admittedly also contain 100 m of normal tectonics, so that the necessary thickness of ice could be reduced to 1500 m. On the other hand, it must not be assumed that the glacio-isostatic compensatory movement is already completed, so that the end result will be an uplift of more than 600 m. According to Chen (1988, p. 30, Fig 3) an uplift of 7–10 mm/year was measured in central Tibet. Values such as these are cited by Mörner (1980, Fig 6 and 20) for current vestiges of uplift in the central area of the N European inland ice on the Gulf of Bothnia during the Vistula (Weichsel) period, and by Andrews (1970, Fig 2 – I; Tab. V-1) in the equally central areas of the Laurentian ice in North America. Mörner (1978, Fig 1) assumes uplifts of 500 mm/year for the time soon after the deglaciation 10,500 to 7000 years ago. The analogy for Tibet shows that comparable final uplifts continue to occur, and also that by far the largest part of the isostatic uplift had occurred thousands of years ago and shortly after the melting of the ice.

Thanks to the very thick continental crust in Tibet, which is more likely to have reacted to glacio-isostatic depression en bloc than in a spatially much differentiated way, the 1500–1800 m thick ice here on the very edge of the plateau is not necessarily the pre-condition for the

recorded amount of uplift. Ice-scour limits and moraine deposits with erratics are evidence of minimum ice thicknesses in Central Tibet of 700 m (NE Tibet; Kuhle 1987a) to 1200 m (Transhimalaya, see above) and of 1600–2000 m in the contiguous mountains (Karakoram and Himalaya; Kuhle 1982, 1983a, 1988b, p. 146). Simple surface interpolations according to the pattern of a low dome shape (like those hitherto always hazarded in the case of the N inland ice, as there are no inspired clues there either) give substance to central Tibetan ice thicknesses of up to 2700 m (Kuhle 1988b, p. 143, Fig 2).

Such thicknesses are confirmed by the inland ice models for Ice Age Tibet, established thermally and in terms of flow-dynamics, which Herterich and Kalov (1988; Mainz Lecture, 1987) had, with the aid of a computer, calculated based on data supplied by the author. They show that, after a 10 Ka phase of ice development, with a mean ELA level of 4250 m and an annual precipitation of 120 mm already produce ice thicknesses in excess of 2000 m, though the morphological final equilibrium of the ice dome surface is not yet reached at that stage. If the same conditions persist for a longer time, the ice thickness will increase even further (Herterich, Kalov and Kuhle 1988; in preparation).

The assumption is therefore that there is a more or less rigid, integral glacio-isostatic overall depression and uplift which had manifested itself on the Shisha Pangma N slope in considerable short-term altitudinal differences of the terminal moraine ramps (Fig 44).

Some observations on terminal moraines and ice marginal ramps on the Himalayan north slope

On the N slope of the Langtang-Himal between Bo Chu (Sun Kosi) in the E and Chilung Chu (-Ho, Gyirong Zangbo or Upper Trisuli valley) the Tibetan High Plateau borders immediately on to the High Himalaya for 60 km (Fig 2, no. 27), which rises 1700 m (Fig 41: Kangpenqin or Kang Benchen 7299 or 7211 m) to 2500 m (Fig 40) above the plateau. This is a rare topographical feature. The norm, as on the Dhaulagiri, on Annapurna and even on Mt. Everest, is that the formation of the S Tibetan Highland proceeds from the Tibetan or Inner Himalaya without intact plateau areas. At the Shisha Pangma the high plateau forms an area at the foot of the mountain on to which 12 recent glacier tongues flow out (Fig 40 ↓ ; Fig 41 ↓↓). In prehistoric, i.e. Late Glacial to historic times a more or less distally continuous foreland glaciation had been formed (cf. Tab 1, final glacier levels in the Yepokangara glacier forefield between 5015 and 5495 m asl; Fig 42). At the time the Late Glacial foreland glaciers advanced 10–15 km into the pediplain they formed ice marginal ramps over 600 m-high (Fig 41 ■; Fig 40 × and I; Fig 42 II). They have been deposited as large middle moraine-tricornes. Ice marginal ramps as characteristics of semi-

arid foreland glaciation have also been described in detail from other areas of Tibet (Quilian Shan and Kun Lun) as well as for the Zagros Mountains (Iran), from Alaska, and from the subtropical Andes (Aconcagua Group) (Kuhle 1984, 1987a, 1989). Here, too, they form a combination of moraine diamictites and stratified outwash deposits. Ice marginal ramps tend to have surface inclines of 7–15° (Fig 41 —; Fig 42 —) their fringes consist of several terraces of moraines and paraglacial formations which were once bordered by progressively melting and retreating generations of glacier tongue surfaces (Fig 42 III, IV etc.). This regressive development can be traced back down to the recent glacier tongues at the foot of the mountains (Fig 42 ×). Towards the mountains the ice marginal ramps accompanying the basins of the glacier tongues outcrop into the air (Fig 41 ◀◀; Fig 42 left of II; Fig 40 behind I). The tongue basins of adjacent ice flows coalesce on the edges (Fig 2, no. 26). The sudden decrease of the pushing force of the ice where the glaciers issue from the steep mountain slopes upon an open foreland plain is essential to the formation of IMRs. The glaciers spread over a large area here. Considerably reduced in consequence, the oscillations of the glacier fringes led to relatively stable locations of ice margins with polygenetic sedimentation of heaped diamictites and outwashed moraines from the meltwaters. More or less well sorted, the re-deposited psamites of the latter are laid down on the outer slopes of the terminal moraines. They were transported down the ice marginal ramps into the foreland (Fig 41 —). The ice marginal ramps, which contain Shisha Pangma eye-gneiss blocks up to the size of a room, and some of which are well-rounded components, have dark phyllites at their core. It was surrounded by glacier ice and buried in a bed of moraine and outwash material (Fig 40 ▼ shows the locality of an outcrop).

The surface areas of the ice marginal ramps are now the scene of macroforms of solifluction. Apart from large block flow tongues on the steeper slopes, there are macro-structural frost patterns with diameters of 1–2.5 m (Fig 42 ▼ shows one of the localities). This takes place more than 1000 m above the permafrost line, which is established at 4600 m through the presence of pingos near the Paiküco (Peiku Tso) (Fig 2, no. 27).

Further indications of prehistoric glacier covers in the northern and eastern Shisha Pangma foreland

Glacier termini (the altitudes of the final position of their tongues) and their ELA-depressions from the pre-Ghassa stagnations (● 1/2) of the early part of the Late Glacial to the Neo-Glacial (V, VI) and historic (VIII–X) glacier locations (the latter are concentrated in the example of the prehistoric Yepokangara glacier) are given in the table (cf. Fig 40; Fig 42; Fig 2, no. 26). The glaciated massive-crystalline knobs which rise



Fig 45 The S Tibetan high plateau N of Nylamu (Fig 2, No. 31; $28^{\circ}19'N$ $86^{\circ}03'E$) seen towards the N from 4800 m; in the left middle ground the irrigated cultivation of the Bo Chu cut down to 4300 m; (●) the roche moutonnée country of the 5200 m Sho La plateau (Sho La pass 5060 m Fig 2, No. 30). This completely glaciated (at the glacial maximum) area is characterised by the post-glacially formed frost debris slopes (periglacially frost smoothed slopes = Glatthänge ↓ and foreground). These forms are a function of the easily weathered metamorphosed sedimentary rocks. Photo: M. Kuhle.

between 5000 and 5250 m or 5300 m ($28^{\circ}40'N/85^{\circ}47'E$) have already been mentioned above (Fig 2, N of no. 26). They present high continental frost-weathering and exfoliation along relaxation joints. Although these destroyed glacial striae and polishings, they did not affect the classically-fashioned form of the glacial knob (Fig 42 ↓).

Further E ($28^{\circ}40'N/85^{\circ}58'E$) there are kame-like deposits running W to E in the area of the Kolmangcheng Hu valley (ONC 1:1,000,000 H-9), which ends in the Mankopa valley in the E (Fig 2, no. 28). Only a few kilometres long, these small-scale gravel ridges consist of very well washed gravels (hard, non-stratified crystalline rocks) and can be regarded as sub-glacial esker-like deposits. The five steps of the gravel ridge terrace, ranging in height from 2–3 m (1) to 5.6–7 m (2), 12 m (3), 35 m (4) and 100–120 m (5) present a secondary undercutting of what was originally a homogeneous sub- or paraglacial gravel body. Corresponding gravel deposits in the form of terraces flanking a valley are to be found to the true left of the 4550 m high gravel floor of the valley. Kame terraces have also been preserved at the openings of small hanging valleys on the true left about 27 km outside the valley ($28^{\circ}48'N/86^{\circ}09'E$) (Fig 2, no. 29). They are situated 150–250 m above the Mankopa river. Deposited inside the banks of the main Mankopa valley glacier (minor flow), these deposits are *so recent* that they have not yet been cleared out by the

streams of tributary valleys. They have not even been cut down to base-level. These are post-High Glacial or Late Glacial accumulations, and evidence of an already considerably lowered ice-level. The glacial knob-like rounding at all levels of the valley flanks is evidence of glaciation which covered the entire relief during the High Glacial period. In this case it is the outcrop curvatures of sedimentary rock which have been preserved in their glacier-polished state. These indications make it likely that the Mingchü Ho (valley N of the Mankoyang massif; ONC 1:1,000,000 H-9), as well as the Tingri basin, which lies 100–150 m deeper down, were filled with ice (see above).

The upper catchment area of the Mankopa valley is a 4900–5300 m high plateau, which is dissected by several shallow valleys, leading to the Jalung Sho La (pass at 5060 m) (Fig 45, background). The plateau (centre: $28^{\circ}35'N/86^{\circ}11'E$) forms the watershed for the Bo Chu (the upper Sun Kosi), the Himalayan transverse valley draining southward. Everywhere on the plateau slopes between 4500 and 4800 m there are exposed remnants of gravel bodies and gravels layers with thicknesses varying from 5–10 m. In the bowl-shaped valley sources on the S slope, for instance, and down to the Bo Chu they are dove-tailed with rough block moraine covers (Fig 2, no. 30) ($28^{\circ}30'N/86^{\circ}10'E$, 4800 m asl). They have the character of ice contact stratified drift of the Late Glacial period. At the altitude of 4900 m well-rounded erratic blocks of up to 1.2 m in length may be found. A local ice-cap, or plateau glacier, had evidently remained on the plateau. Triangular in outline, its glacier tongues overlapped into the peripheral valleys from all sides. Lack of thrust from the ice led to deposited end moraines or outwashed glacial sediments in the form of alluvial cones. Whilst on the plateau chalks, chalk-marls and lightly metamorphosed rocks came to the surface, these ramps consist of well-rounded gravel with quartzite, granite and gneiss components. They are the substratum of repeatedly shifted allocthonous moraine material (Fig 2, no. 30). Other than glaciogenic interpretations of the gravel would require the assumption of a complete infilling of the valleys on which they occur to a thickness of several hundred metres. However, this would contradict the mass-balance between the relatively small catchment area available for the processing of scree and the considerable quantity required for the filling-in of the relief. Sedimentary rock at about 4900–5100 m on the plateau is superimposed upon by ground moraine covers of unstratified, crystalline erratics, several metres in length, that have been transported here over considerable distances (Fig 2, no. 30). They can be observed from the track to Nylamu.

Cut down a few hundred metres into the plateau, the curving box valleys or troughs with wide valley floors are typically glacial. They may be compared with valleys in the sedimentary rocks of Spitzbergen (Kuhle 1983b).



Fig 46 Southern Tibet with Tibetan Himalaya N of Nylamu (Fig 2, No. 31; 28°17'N 86°02'E) looking up from 4850 m towards the E. The valley was completely glaciated (----- glacier level). This is shown by the glacial scouring and roches moutonnées (●). The trough-like box valleys formed then from the easily eroded metamorphosed sediments were strongly modified by deposits from gullies and stone chutes (↓). Photo: M. Kuhle.

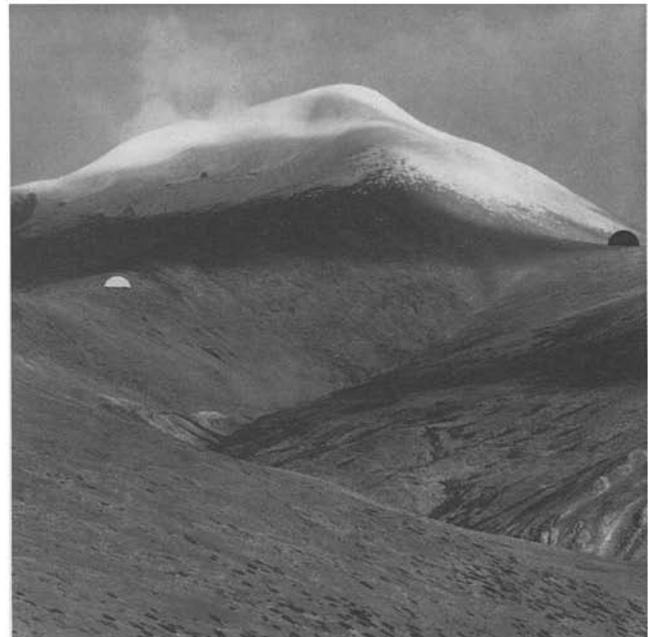


Fig 47 S Tibet, Tibetan Himalayas N of Nylamu with a 6652 m massif (28°18'N 86°20'E; Fig 2, No. 31) seen from c. 5050 m towards the E. The peak carries recent glaciers related to an orographic snow line (equilibrium line altitude) of more than 6000 m. A depression of this line by only 400 m would easily cover the whole area with glacier ice. The scoured transfluence passes (●) indicate a maximum glaciation ice thickness of more than 1000 m related to a depression of the equilibrium line of about 1200 m. Photo: M. Kuhle.

Sub-glacial melt-water erosion is likely to have played a part in their formation, too (Fig 2, no. 30).

The character of the landscape in the area of this plateau and its S slope (Fig 45, Fig 46) is altogether similar to that found in the central areas of Spitzbergen (cf. above), as for example in the sedimentary rock plateaux of Dicksonland (79° N). They, too, have been superimposed upon and shaped by a more than 1000 m-thick inland ice cover during the High Glacial stage (Kuhle 1983b; Troitzky et al. 1979; Boulton 1979). Built up by more or less concordantly bedded sedimentary rocks (sandstones, quartzites), the plateau area is topped by 100–200 m-high, almost scree-less glacial knobs (Fig 45 ●). In the depressions the bedrock is superimposed on by ground moraine boulder clays and in the valleys periglacially smoothed slopes (conglifluction slopes) have developed. They consist of amorphous conglifluction scree covers (Fig 45, foreground), which are interrupted by more resistant outcrop steps (Fig 45 ↓). It is worth noting in this comparison that, after great similarities in their glaciation history, their conformity is primarily based on petrographic foundations. Apart from that an oceanic-arctic climate of freeze-thaw cycles contrasts with a high-continental subtropical one. Not the geomorphology, but the denser plant cover (tundra, dwarf bush heath) as well as the more frequent patches

of perennial snow of the corresponding altitudinal level in the oceanic Arctic, mark the difference whilst precipitation totals (200–300 mm/year) are the same (Fig 45, 46).

A glaciation of this plateau between the Longlonqu river in the N and the Keya river (near the Yali settlement in the upper Bo Chu) in the S requires a lowering of the equilibrium line by at least 800 m (from 5900 to 5100 m). However one must take into account the fact that in pre-glacial times the plateau was situated at a higher level than now (cf. above). The greater altitude of the surface of the ice domes makes it likely that *deglaciation* took place only when the line of equilibrium rose to a level which differs by a mere 600 m from its present location. This points to a deglaciation of this plateau area at about 15–20 Ka BP.

The prehistoric glaciation of the Bo Chu (Bote Chu, Sun Kosi) and its tributary valleys (Sisha Pangma group's E and SE slope; W flanks of the Rolwalling-Himal and the 6652 m high massif further N in the Tibetan Himalaya: 27° 52' N–28° 30' N/85° 50'–86° 20' E)

From the 6652 m high massif (28° 18' N/86° 20' E; Fig 47) to the vicinity of the main valley floor (Bo Chu)



Fig 48

Looking down an E tributary valley of the Bo Chu towards the W from 5350 m (Bote Chu, S Tibet $28^{\circ}20'N$ $86^{\circ}07'E$ Fig 2, No. 31). The glacier level (---) shows that at the glacial maximum the whole area was filled with a network of ice streams. Transfluence passes (▲) in many places were formed by the thick ice masses. Lateral (▼) and end moraines and ground moraine (x) deposits remain from the glacial stadia of late-glacial, neoglacial and historical times. Photo: M. Kuhle.

valley glaciers, probably early Late Glacial, and at most 9 km long, descended in a W direction to about 4100 m asl ($28^{\circ}17'N/86^{\circ}01'E$); (Fig 2, no. 31). The sole evidence of this ice stream are the at most 120 m high terraces of lateral moraines (Fig 48 ▼ ▼). With a mean altitude of 5800 m for the catchment area, the orographic equilibrium line is calculated as lying at 4950 m (Tab 1). In this instance the lower of the two ice margin locations was computed. A higher one happened to lie 1 km further up-valley at an altitude of 4250 m, and would have corresponded to an equilibrium line c. 75 m above the first one. Corresponding deposits of ice margins also occur in the S parallel valleys, as well as to the W of the Bo Chu on the E slope of the Shisha Pangma massif.

Now separated from the three recent, up to 3.7 km long, glaciers in the NNW-facing source basins only by neo-glacial or historic dumped end moraines, these glacier locations were classified as belonging to the High Glacial by Zheng Benxing (1988). This is like describing the openly wooded terraces of the lateral moraines of the middle Arolla valley and Zinal valley, or the glacial deposits near Täsch and Saas Grund in the Valais (Alps), as belonging to the High Glacial period because for tens of kilometres there are no marked moraines – or none at all to be found in the large main valley, the Rhone.

Similarly to the alpine region mentioned above, the entire relief had been filled with glaciers to at least 1000 m above the bottom of the main valley. Evidence of this is found in, for instance, the rounded 5030 m-high transfluence pass ($28^{\circ}20'N/86^{\circ}07'E$) of the aforementioned valley to the adjacent one further north (Fig 48 ▲). There are distinctly polished flanks and rounded valley shoulders in many places (Fig 46, 47, 48 ▲, ---). Due to the splintering frequently encountered with metamorphics, and to the frost weathering, well preserved polishings on the flanks are rare. Nonetheless, they

have been preserved in the main valley (Bote Chu) at 4300 m asl (Fig 2, no. 31). Fundamentally speaking, this section of the Tibetan Himalaya presents a large-scale landscape of glaciated knobs, of which morphologically speaking the most foreign element is linear and fluvial lateral erosion (Fig 46, 47). If it had been effective throughout the Quaternary in its present intensity, there would now be a steeply-shaped landscape of pointed ridges, steep rock flanks and narrow V-shaped valleys with long stretches of gorges and walls much cut up by the action of water. 10,000 years after deglaciation these form elements already contrast with the periglacial macro-forms conforming to glacial geomorphology. Even small, periodic funnels without catchment areas worth mentioning in precipitation areas of only 150–300 mm/year have since cut deeply into the bedrock once the mantle rock of the practically vegetation-free slopes had been removed (Fig 46, ↓ ↓).

Following up the history of the Late Glacial glaciation further down the valley reveals the tongue basin of the Tsangdong (Fig 2, no. 32). The Tsangdong settlement ($28^{\circ}15'N/86^{\circ}01'E$) is located on the valley floor at about 3750 m in the Bo Chu. At Late Glacial times the main valley glacier received its chief supply of ice by way of the Keya valley (confluence of the Bo Chu ($28^{\circ}22'N/86^{\circ}04'E$)). It descended from the highest catchment areas of the Shisha Pangma E flank, which must be put at 7100 m asl. The Tsangdong tongue basin is probably as old as the lowest moraines of the W slope of the 6652 m high massif (Fig 48 ▼ ▼); the comparable geomorphological state of preservation also seems to indicate this. A somewhat more recent and less representative glacier tongue basin is situated in a valley chamber about 7 km upstream from Tsangdong at 4050 m (Fig 2, no. 32). It must be regarded as the more recent one of the two Late-Glacial stages occurring on the 6652 m massif (Fig 48 × ×).

Both of the tongue basins are characterized by their more recent stratified drift (outwash plain) fillings. They function as sediment traps and retain the glacio-fluvial washed drift of the immediately preceding stage. During further glacier retreats they were cut up into terraces.

Large terminal moraines were pushed from the mouths of two tributary valleys on the true left into the tongue basin of the Tsangdong (Fig 2, no. 32; 28° 13' N/86° 01' E and 28° 17' N/86° 04' E).

These indicators of ice margins also belong to the Late Glacial system (more recent than the Ghassa Stadium (I)). With the help of the indicator of the relative dating of red-weathering at comparable altitudes above sea-level, the echelons of lateral and end moraines at Nylamu (Nyalam, Fig 2, no. 33), which occur as far down as 3670 m (Fig 49) can be regarded as parallel in time to those of Tsangdong. Pushed out from the Fuqu valley as far as into the main valley (Bo Chu) (28° 06' –09' N/85° 58' –86° 00' E) these diamictites extend up to 3900 m and contain blocks up to the size of houses (Fig 49 ● ●). There is no marked humus stratum, so that the moraine surfaces must have weathered under cold-semi-arid climatic conditions during the Holocene. The abrupt climatic division towards the monsoon-humid Himalaya S slope occurs 5 km up-valley from Nylamu. The tongues of the recent and some 14 km-long Nylamu Phu glacier (Fuqu or Shisha Pangma SSE glacier) and the 7 km-long glacier at the foot of the wall, which flows down an E parallel valley from the Shisha Pangma, are 21 and 23 km away from the Nylamu stadium and more than 1000 m higher. The equilibrium line at about 5400 m (Tab 1, 5385 m) for the Nylamu stadium is evidence of an ELA-depression of 500 m. This corresponds, to a rather small equilibrium line depression on the Himalaya S side for the Late Glacial (cf. Kuhle 1982, 1987a). Even the moraines of the Nylamu stadium the author would possibly attribute to the Sirkung or Dhampu Stadium (III of IV; Kuhle 1982) (Late Glacial) Zheng Benxing (1988) places in the middle Pleistocene. Absolute datings have not been obtained up to now.

In the valley chamber of Nylamu still more, and substantially higher, walls of lateral moraines begin. They are particularly well preserved on the true right-hand side (Fig 49 ■ ■; Fig 2, no. 33). Like the moraine material deposited further up-valley, they are characterized by angular, round-edged, or rounded block components. The moraines are completely covered with green meadow vegetation; their petrography is akin to that of the moraine from the Nylamu Stadium; the polymict composition of blocks include varieties of gneiss including eye gneiss and varieties of granite. There is thus evidence of the Shisha Pangma catchment area belonging to this older glacier stage. These lateral moraines correspond to the end moraine which closes the valley chamber of the Choksum (Shü Shang) settlement at 3380 m asl 5 to 7 km down-valley (28° 06' N/86° 00' E, see Tab 1). The end moraine juts out from the true right-hand valley flank (from the W) towards the deepest

point of the valley (Fig 2, no. 34). The stage is to be regarded as the Taglung or Dhampu Stadium (II or III), and thus as being from the Late Glacial period. Compared to its present position the equilibrium line was depressed by about 650 m. The extreme steepening of the Bo Chu valley profile (Bote Chu or Sun Kosi) sets in between Choksum and Nylamu where the Fuqu or Nylamu Phu valley ends. Above this steepening, e.g. in the tongue like basin of Tsangdong, mud-flows, along with alluvialfans, are reshaping the moraine landscape, whilst below the glacial forms are modified by substantial linear incision (Fig 50). Barrier mountains below Nylamu mark the old level of the valley floor. Glacial polishing had rounded their tops. They received their finish as late as the Late Glacial and are accordingly well preserved, glaciated knobs (Fig 49 ▲). All the post-Glacial slides and mud-flow activities are swallowed up by the very deeply incised valley long profile which continued its formation during the sub-Glacial period, and subsequently rapidly removed by the ample waters of the Sun Kosi river and distributed tens of kilometres further down the valley. The mudflow event on August 28th, 1983 is an example of such a discharge. Caused by a moraine lake in a tributary valley on the true left of the Bo Chu at an altitude of more than 4800 m asl, it killed more than 100 people. It reached the main valley at about 3000 m and proceeded another 2000 m further down. Covering a vertical distance of about 4000 m, cut through the track of the mountain road in two places and destroyed the large suspension bridge (the Friendship Bridge) 20 km down-valley.

The main valley gradient (Bo Chu) over 30 km equals a 2000 m height loss; this explains the enormous tractive powers in the prehistoric valley glacier. They had the effect of polishing the floor much more intensively than the polishing on the flanks which led to the formation of a trough-shaped gorge, or – depending on the intensity – a gorge-shaped trough (Visser 1938, Vol. II p. 139; Kuhle 1983a, p. 155 et seq.). The glacial formation of this valley section is established through the frequently concave profile of the flanks in the gorge, which have been smoothed by glacial polishing (Fig 51 →). The crumbling away of the rock (Fig 50 ◆) and even larger rockfalls have occurred frequently. They are presented here as outcrops of the Khumbu and Kathmandu covers (KU 1–3, KN 2, 3) which consist of gneiss metamorphics (Hagen 1969, p. 129) and slope northward at 20–35° (20–35/10). Obsequent or, respectively insequent in its course, the valley tends towards a V-shape profile, thanks to the structure of its rocks and the collapse of undercuttings induced by crevices. Having reached down to at least 1600 m, that is to the present Kodari settlement (27° 56' N/85° 56' E) (Tab 1) this High Glacial valley glacier remained with its tongue below the equilibrium line (4450 m) for 22 km. The end of its tongue was about 2800 m below the ELA. It follows that the meltwater was able to cause sub-glacial erosion over the same distance. With a surface

Fig 49

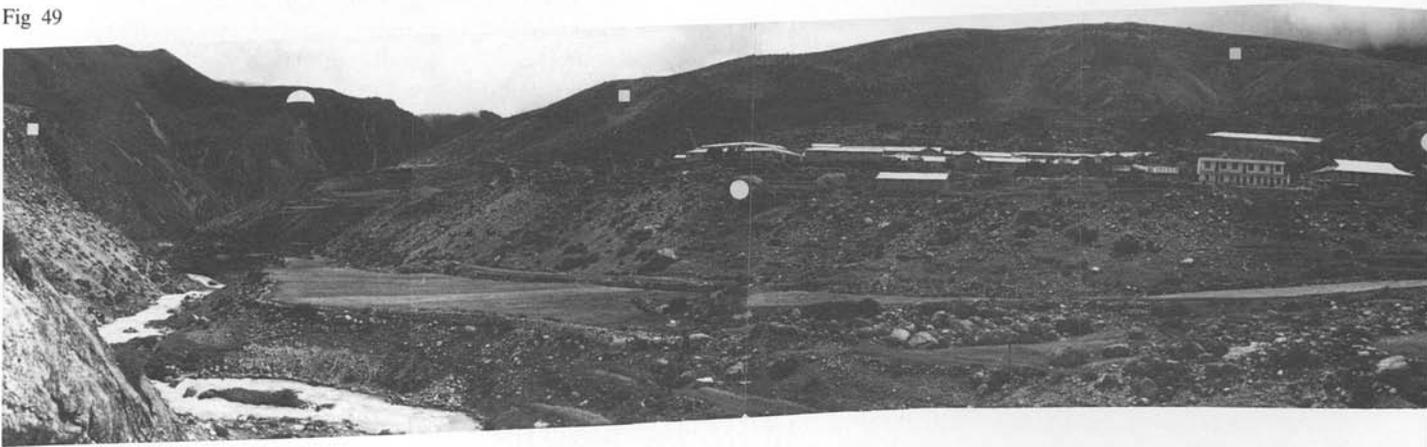


Fig 51

Fig 53





Fig 52



Fig 49

An area of lateral moraines (■●) by the settlement of Nylamu in Fuqu Chu (Fig 2, Nos. 33, 34). View from 3690 m towards the SW (centre of panorama) in the area of the Bo Chu-(Sun Kosi-)transverse through the Himalaya main range. The glaciers from the SE Shisha Pangma (Fuqu glaciers) formerly flowed together in the Fuqu valley (Nylamu valley) and even in the late-glacial reached (■) the Bo Chu main valley (■ well to the left and ▲) and thus the previous Bo Chu glacier. Fig 2 shows (Nos. 33 and 34) the course of the lateral moraines in the valley system. A few kilometres below the confluence of the Fuqu and Bo Chu valleys the glacial accumulations set out due to the steepened relief (well left in the background). Thence far into the S side of the Himalaya lateral scouring and roches moutonnées (▲) indicate the extent of the glaciation at the maximum (Fig 50, 51 and 52; Fig 2, No. 35). Because of the content of the augengneiss and granite these montane moraines are poor in fine material and contain coarser components (● well to the right, outcrop). Photo: M. Kuhle.

Fig 50

The glacially produced gorge of the middle Bo Chu where it breaks through the main Himalaya seen looking southward down the valley from 2700 m (Fig 2, between No. 34 and No. 35). The erosive force of the steeply downhill-flowing maximum glaciation glacier resulted in bottom erosion exceeding lateral scouring. This resulted in a slightly trough-shape concavely scoured V-shaped valley. At the same time at 1500 m below the equilibrium line altitude (snow line) the subglacial melt-water had enhanced erosive force which exaggerated the V-shape. (♣) marks a typical rock fall of post-glacial origin. Photo: M. Kuhle.

Fig 51 Looking up the Bo Chu from the same viewpoint as Fig 50 (28°02'N 85°59'E; 2700 m). (↓) shows the concave scoured slopes of this glacial gorge-like trough. Both contemporaneous and later rock falls account for the roughness of the rock walls (▶▶). Photo: M. Kuhle.

Fig 52 Ravine in the valley floor of the gorge-like trough of the lower Bo Chu or Bote Chu (Fig 2 No. 35) at 1900 m. This ravine with almost straight right-angled walls with later rock falls (↑↑) is cut into the glacially scoured rock floor of the valley (▲). It follows a fault shatter zone. The incision resulted from erosion by subglacial meltwaters under hydrostatic pressure and continues today subaerially. Photo: M. Kuhle.

Fig 53 The Dzakar Chu seen downwards towards the N from 4450 m (Fig 2, No. 36; (28°23'N 86°57'E)). This box-like valley is a trough with a wide outwash-gravel-filled bottom. The valley slopes of metamorphosed sediments are completely rounded up to the inter-valley watershed by glacial scouring (▲▲) (cf. Fig 54). (↓) the settlement Zambu; behind is a similar glaciated relief. Photo: M. Kuhle.

split up by crevasses and ice-falls, the steep glacier tongue had the effect of a very immediate discharge of meltwater from the ice surface to the rocky floor of the valley. As a result of the hydrostatic pressure of the intraglacial water column there was a very intensive erosion and deepening of the valley floor (Tietze 1961). This was further intensified by cavitation corrosion. Easily visible remnants of slickensides show that the Sun Kosi follows a fault which facilitates linear erosion. This permits an explanation of the bipartite cross-section of the Sun Kosi: beneath a trough-shaped gorge profile (gorge-shaped trough) there is an acute and c. 40–100 m deep gash-like cut with almost parallel steep flanks (Fig 52). They form suggestive working edges towards the hanging glacier bed (Fig 52 ▲ ▲). Glacigenic polishings on the upper metres of the gorge and further down into it are evidence of the simultaneity of glacial erosion in the upper section of the profile (Fig 52 above ▲ ▲) and meltwater erosion in the lower one (Fig 52 below ▲ ▲). As a matter of fact the two associated processes are linked in the way of a geomorphological sequence. The sub-glacial erosion through meltwater precedes the glacial polishing. It prepares the detraction and detersion carried out by the glacier by way of renewed resistance to attack, thus making glacigenic deepening come to the fore instead of flank polishing. The gorge-shaped trough profile is therefore not only to be attributed to the considerable gradients of the long profile, but also to sub-glacial meltwater erosion.

Breaks in the compactness of the glacigenically polished flanks are due to petrographic factors. In places where gneiss covers alternate with soft two-mica phyllites, the ample monsoon precipitation causes recent slate slides and mudflows. These expose the hard strata and destroy the road track every year. This may, for example, be observed on the true left-hand flank down-valley from the Damu border station at 1900 m asl (cf. Fig 50 ▼).

The lowest unambiguous glacigenic polishings reach the location of the Friendship Bridge at about 200 m above the valley bottom-line, so that it was possible to reconstruct a terminal position for the glacier tongue at 1600 m (see above Fig 52 ▲). A c. 100 m high terrace of sorted pebbles and diamictites, which must be interpreted as a kame-like paraglacial formation confirms the altitude of the ice margin positions. The orographic equilibrium line is calculated as being at 4450 m asl, the climatic equilibrium line at about 4310 m. These values are about 260–485 m higher than those in the Dhaulagiri and Annapurna parts of the Himalaya some 200 km further W (Kuhle 1982, p. 152). They are regarded as belonging to the last High Glacial period. There are no preserved glacier locations between this, the lowest ice margin position, and the one near Choksum (Shū Shang) which are considered as belonging to the Late Glacial stages II and III (Taglung or Dhampu Stadium). The Taglung or Ghasa Stadium (II or I) and the two pre-Ghasa stagnations (●^{1/2}) are likely to have been cleared out completely on account of the steep and narrow contours of the gorge (Fig 50, 51). In the cross-section of the valley at Damu (27° 59' N/85° 58' E) the Sun Kosi flanks are well marked by steep "bottleneck" corries with small and well-rounded barrier mountains (glaciated knobs), which occur as far down as 3500 to 3300 m. They are evidence of a depression of the ELA on the Himalayan S side to well below 4000 m (to c. 3700 m). Due to the inaccessibility of the very steep, wall-like and densely wooded (*Tsuga dumosa*, *Abies spectabilis*, *Rhododendron lepidotum*) valley flanks it proved impossible to establish whether all the remnants of loose rocks preserved in the runnels of the walls are ground moraines or lateral moraines which have been infilled, or scree from the flanks that got stuck in the long profile. On the true right-hand at about 3300 m asl, c. 300 m above the valley bottom-line of the Bote Chu gorge there is polymictic moraine material which plugged a tributary valley (28° 02' N/85° 59' E).

An estimation of the minimum age of the glacigenic forms that have come down to us requires the forced dynamics in the recent mudflow gullies to be taken into account. The humid monsoon climate, together with the melting of the snow, the vertical distance and the absolute altitude of the catchment areas with their prehistoric moraine deposits which can be moved down, all combine to form ideal conditions for these dynamics. Hanging valleys and high altitude basins of prehistoric tarns and nivation niches give way to runnels and gullies. Their steep floors contain mudflow tunnels which are like bottlenecks, U-shaped with cross-sections of 5×20 m. Fresh abrasion levels on the rocks are evidence of mudflow passages up to a depth of 4–5 m in the cross-section, and evidently resulted in considerable corrosion effects. If a body of mudflow suspension comes to a halt in such a tunnel, i.e. does not reach the lowest level of the main valley, the fine material is washed out by the stream. This includes gravel and coarse valley



Fig 54 The Dzakar Chu seen looking upwards from the same valley cross profile as Fig 53. The overwhelming maximum glacial ice cover of this area has led to a considerable valley-wards extension of this roche moutonnée landscape (●●). The comparatively meagre remains of moraines (▼▼) in relation to the appreciable glacial erosion recalls the relationships in the Scandinavian mountains where erosive forms also dominate. Photo: M. Kuhle.

train. However, very large blocks stay where they are and block the landslide runnel. They dam up finer materials or are worked free from beneath by the stream. The tunnel exits of particularly active mudflow passages retreat from the main valley back into the flanks. This is evidence of post-glacial erosion amounting to several tens of metres.

The former Glaciation of the North Slope of Mount Everest from the Rongbuk Valley down to the Middle Dzakar Chu (upper Bhong Chu or Arun Valley; $28^{\circ} - 28^{\circ} 26' N / 86^{\circ} 38' - 87^{\circ} 15' E$)

The following comments continue the discussion above (p. 11) on the area N of the Pang La massif. In 1984 the author had had the opportunity of proceeding with investigations up-valley of the Dzakar Chu, starting at the Tushidzom settlement. In the valley cross-section of the settlement, which crosses the 4200 m isoline there are no glacial deposits whatsoever. It is a "box" profile with a kilometre-wide valley train floor, and a flat alluvial fan poured out from the Nyomdo valley provides security from floods for the dwellings and fields of Tushidzom. Here the valley is not dissimilar to the Rhone valley profile at Siders (Valais, Alps), though much more arid. Glacial formations on the valley flanks are not clearly identifiable. The metamorphic bedrock and the heavily folded, flysch-like sedimentary rocks have been broken up into crags, i.e. rock heads and ribs. Post-glacial talus slopes follow on below. Some remnants of lateral moraines have been preserved on the true left hand, about 18 km up-valley at an altitude of about 4350 m and 100 m above the valley train floor (Fig 2, no. 36). The corresponding position of the ice margin may have been situated tens of kilometres outside the valley and has so far escaped reconstruction. (Investi-

gations in the lower Dzakar Chu and Arun valleys are planned for 1989). The extrapolation of a hypothetical Dzakar glacier from these moraine findings, which stretched down to 4180 m, would produce an equilibrium line depression of about 350 (345) m, as opposed to the recent orographic equilibrium line of the Rongbuk glacier system at 6235 m. Taking into account the thermo-dynamic upper limit of the glacier, the ELA runs at 5910 m (Kuhle 1986a, b) (Tab 1).

The estimation of the absolute level of the equilibrium line is difficult here in the Everest region, due to the adjacent catchment areas at very high altitudes – the mean being 7300 m asl – with their special conditions. The definitions of ELA depressions, however, are subject to consistent conditions and are, in consequence, reliable.

An ELA depression of merely 350 m is evidence of the at least recent Late Glacial age of the extrapolated ice margin position outside the Tushidzom valley. (Being too uncertain, i.e. only an extrapolation, it was not included in Tab 1). By comparison with conditions N of Shisha Pangma, a significant post-glacial uplift of the Dzakar Chu, the part of S Tibet that borders on the High Himalayas, must also be assumed here N of Mt. Everest.

There is no change in principle in the geomorphology further up the Dzakar Chu until the valley floor reaches an altitude of about 4500 m. Once again (Fig 53; $28^{\circ} 23' N / 86^{\circ} 57' E$) the very wide valley floor between two steep valley slopes strikes the observer as characteristic, and not unlike those in the formerly ice-covered areas in the interior of West Spitzbergen (i.e. Sauriedalen and Lyckholmdalen in Dicksonland; $78^{\circ} 27' - 78^{\circ} 45' N / 15^{\circ} - 16^{\circ} E$). There, in the Arctic, these valley floor fillings have been deposited near sea-level, i.e. close to the absolute erosion base. Here, in the subtropics, conditions for ample valley floor sediment-



Fig 55 The upper Dzakar Chu seen towards the SW from 4600 m (Fig 2, No. 37). The glacial scouring of the strata (▲) is remarkable insofar as under unglaciated conditions the layer-edges would have been decomposed to rock towers and gendarmes. The present day snow line (equilibrium line altitude) runs some 1000 m above the valley floor in this semi-arid part of S Tibet (outwash plain in foreground). Photo: M. Kuhle.

ation in a thawing-out (apering) region are comparable, thanks to the small gradient of the valley floor and the sedimentary rocks. From the confluence of the Gyachung Kang N valley with the Rongbuk valley ($28^{\circ} 17' N/86^{\circ} 48' E$) to the Tushidzom settlement ($28^{\circ} 24' N/87^{\circ} 08' E$) over a distance of 45 km, the valley floor drops by only 400 m. One ought therefore to imagine an ice-filling during the glacial period which stayed put in this S Tibetan valley grid and, due to a great deal of friction, displayed no impulses of discharge worth mentioning. It merely distributed the Pleistocene inter-glacial debris on the valley floors, but had not scoured it out in the way which is characteristic for steep relief.

One difference between this profile and other valley cross-sections of the Dzakar Chu further down the valley, however, is the much better state of preservation

of glacial fluvial polishings. The outcrops of more or less metamorphic sedimentary rock series have been preserved in their clearly polished, rounded form (Fig 53 ▲▲). They are evidence of a glaciation that included the entire valley. In some depressions remnants of moraines and para-glacial terrace deposits (glacigenic outwash alongside the lateral moraines) have been preserved (Fig 54 ▼▼). Established by lateral moraines or kame terrace remnants on both sides, the nearest ice margin was about 15 km down-valley of the Zambu settlement (Fig 53 ▼) at 4380 m asl (Tab 1). It corresponds to an equilibrium line depression of c. 250 (245) m. (Fig 2, no. 36).

Further up-valley there is a glaciated-knob landscape of very large dimensions; its metamorphics with a concordant layer of congelifractions produce soft forms, which are evidence of a large-scale overwhelming of the relief by glaciation during the High Glacial period (Fig 54, 55 ▲). Now only 60 m high, a true right-hand terrace of lateral moraine at 4600 m basic height asl (Fig 54 ▼ left; Fig 56 ▼▼; Fig 2 no. 37) marks another, Late Glacial position of an ice-margin up-valley from the Zambu and Chöbuk settlements at 4550 m, and thus an ELA depression of c. 225 m (Tab 1). A corresponding moraine remnant is preserved on the true left in the exit of the Gyachung valley. Here again the glacier catchment area had been established by the N slopes of Gyachung Kang (7975 m) and Mt. Everest (8874 m). Its altitude is, rather comprehensively, given as 7300 m. Further up-valley it was only possible to reconstruct the glacier arm of the Rongbuk valley.

The next and more distinct end moraine is to be found at about 4780 m asl (Tab 1) in the uppermost Rongbuk valley chamber, up-valley from the bottleneck of the valley in the area of the confluence with the Gyachung-Kang N valley, 400 m lower than the recent terminus of the glacier tongue (5180 m). Evidence for it is provided by boulder clays with very large granite components. Considerable glaciofluvial remoulding has taken place, i.e. it has been integrated into an outwash



Fig 56 True right bank lateral moraine terrace (▼▼) in the upper Dzakar Chu (Fig 2, No. 37, 4600 m) seen looking E. These deposits were formed in the late-glacial around the end of a glacial tongue at 4550 m (equilibrium line depression c. 225 m). Post-glacial solifluction has smoothed the edges of the moraine and softened their bulges. Measurements show that periglacial debris movement on slopes such as this reaches 4–8 cm per year. Photo: M. Kuhle.

terrace on the true right-hand side (Fig 57). The calculated equilibrium line depression amounts to only 200 m. Apart from a likely post-Late Glacial uplift (see above), which is bound to lead to faulty calculation (to too small a value of ELA depressions) this position of the ice margin is about 12 km away from the recent glacier (Fig 2, no. 38). This extension of the glacier tongue can scarcely have been achieved by that small ELA depression of a mere 200 m. The de facto value may have been accordingly distinctly greater. As far as the Rongbuk settlement (monastery) no further lowest ice margin positions can be distinguished. Higher up, however, the valley slopes show accumulation ledges and traces of exarations which are evidence of relief-filling glaciation during the High Glacial like the polishing and scouring facets (glacigenic triangular slopes between the junctions of tributary valleys) which run high up on the valley flanks (Fig 58 ● ●). Post-Late Glacial morphodynamics are extremely intensive in this section of the valley. Quite apart from a nivally-induced periglacial formation, mudflows in tributary valley junctions are very active (Fig 58 ◆ ◆). Large mudflow cones within which extremely large blocks of polymict lateral moraine material was shifted, block the valley from the true right-hand, i.e. produce a chamber. Examples of this may be seen in the three mudflow cones 1, 3 and 6 km down-valley to the N of Rongbuk. All three of them at the exits of recently glaciated tributary valleys with catchment areas at altitudes of 6060–6260 m.

Before going into the question of the most recent prehistoric and probably neo-Glacial and historic states of the glaciers of the Rongbuk glacier system, the author would once more like to draw attention to the considerable glacio-fluvial transformation of the middle Rongbuk valley. At the same time though, the transformed end moraines concerned were not at all distinctly formed – if they existed at all. The cause of this was the direction of glacier discharge which changed during the Late Glacial stage from a prehistoric southerly to an at present northerly flow. The prehistoric ice-filling of the relief from an altitudinal ice-level of more than 6000 to 6500 m asl had consequently to spill over the passes W and E of Mt. Everest on to the S slope of the Himalaya (see below).

Neo-Glacial and historic positions of the Rongbuk glacier (28° 09' – 12' N/86° 50' E)

Fig 58 shows the succession of moraines from the most recent dumped end moraine (X–VIII) of the Rongbuk glacier tongue to 1.4 km up-valley from the Rongbuk monastery, where the upper neo-glacial end moraine forms a distinct arch (VI). Whilst stages X–VIII continue to hold dead ice and meltwater ponds and remain in contact with moving ice (◆), stage VI is 2.2 km away from the recent glacier terminus. The zone of dead ice and dumped end moraines (stages X–VIII)



Fig 57 The blocky-loam of the end moraines at 4780 m in the lower Rongbuk valley (Fig 2, No. 38) lies 400 m lower than the recent end of the Rongbuk glacier terminal tongue (Fig 58). The accumulation of granite blocks in the foreground is glaciofluvially modified and included in a younger glacial outwash plain (the terrace forms ◆). These outwash terraces are of neoglacial age (c. 4500–2000 years BP; cf. Fig 58, V–VI). The recent Rongbuk river flows 20 m below the terrace edge. The glacially rounded bosses in the background were maintained in shape by freeze-thaw action (periglacial) and covered with frost debris with a maximum thickness of one metre. View down-valley: M. Kuhle.

is about 1 km wide. The discharge of the meltwater from the recent glacier snout occurs about 1 km upstream of the dead ice tongue, thus dividing it from the glacier ice still on the move (Fig 58 ×). Between the outer slope of stage VIII (after Kuhle 1982: a more recent Dhaulagiri Stadium) and the inner slope of stage VI (older Dhaulagiri stage) a classic kame field (● ● ●) has formed. Having been built up from surface moraine which broke through the former basal ice, the conical kames are 4–10 m high. Even an esker-like subglacially formed remnant of a meltwater terrace, a 6 m high drift platform (Fig 58 ■), which extends in the direction of the valley, is part of the inventory of this tongue basin. The tongue basin is filled with more recent outwash material of the stages VIII–X (–4, –5), and of recent outwash deposits (–6), as well as with recent fans of mudflow and alluvial drift (◆). This resulted in the almost complete obliteration of rudimentary end moraines of an intermediate middle Dhaulagiri stage 'VII or an oldest more recent Dhaulagiri Stadium VII. 1.9 km away from recently flowing ice, this stage can be extrapolated by way of lateral moraines on the true left (Fig 58, 'VII or VII). They are submerged below that more recent drift deposit. The difference in basal altitudes from the recent

Fig 58

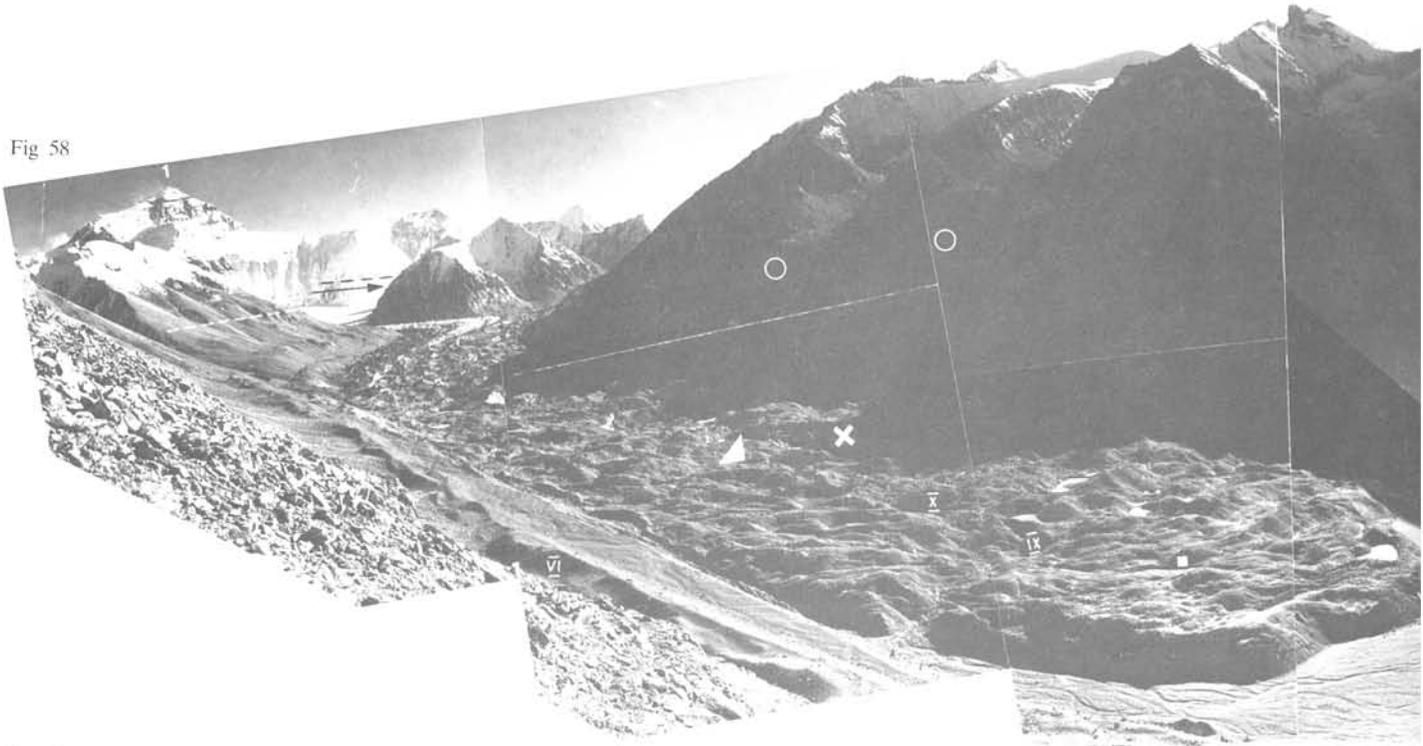


Fig 59

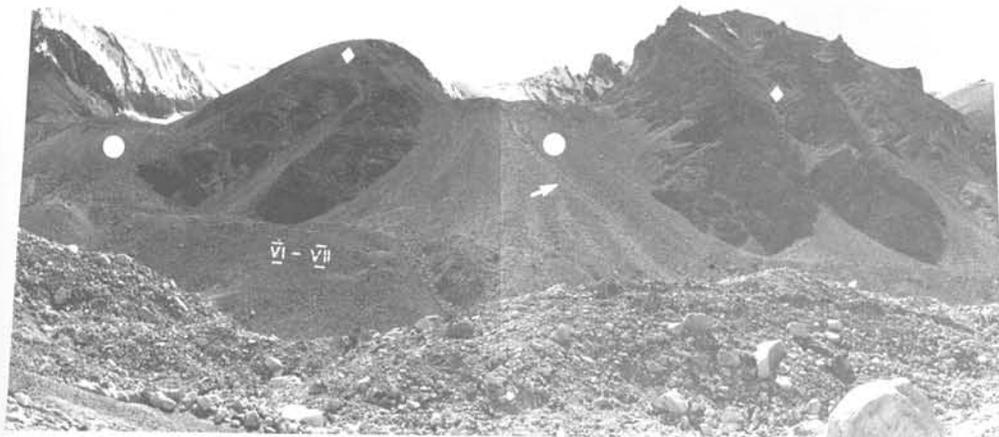


Fig 60



Fig 58 Panorama of the Rongbuk valley towards the S showing Mt. Everest 8874 m (No. 1; Fig 2, No. 40) above the almost 18 km long Rongbuk glacier. In front of it towards the right it is joined to its historical to neoglacial glacier tongue basin (VIII–VI). Towards the N down-valley lie the late-glacial moraines (III) (Fig 2, No. 38). Far right (▼▼) there stretches the mountainous area of S Tibet which was completely glaciated at the maximum. The glacier flow direction in this section of the Rongbuk valley has been completely reversed to the N today as compared with that to the S through the main crest of the Himalaya in maximum and late-glacial times (→). Photo from 5530 m; M. Kuhle.

Fig 59 Looking S from 5650 m from the confluence area of the Rongbuk valley with that of the Rongtö over the Rongbuk glacier to the left and the Rongtö glacier to the right. Mt. Everest 8874 m is at 27°59'N 86°56'E (No. 1). Both ice streams, with their considerable cover of debris, are in retreat. The significance of debris accession from the slopes above the glacier surface is clear. In a few thousand years the intense freeze-thaw climate of S Tibet is able to destroy the lateral striations of the valley sides (cf. Fig 60). Photo: M. Kuhle.

Fig 60 True left slopes of the upper Rongbuk valley near the central Rongbuk glacier (in the foreground; Fig 2, between No. 38 and No. 39). Pedestal moraines (●●) project from two parallel hanging valleys above the neoglacial to historical lateral moraines (VI–VIII; cf. Fig 58, ○○). The recent glacial ice traverses the upper 30 m of the pedestal moraine slopes as far as their edges. Thence the meltwater runs down in thin channels (↗) and forms a kind of sandr skirt or transitional cone (or very steep ice marginal ramp). The quantity of weathering debris on the intervening rock slopes (◆◆) is significant. Periglacial weathering is thus shown to be important even in the morainic deposits surrounding these relatively small slope glaciers. Photo: M. Kuhle.

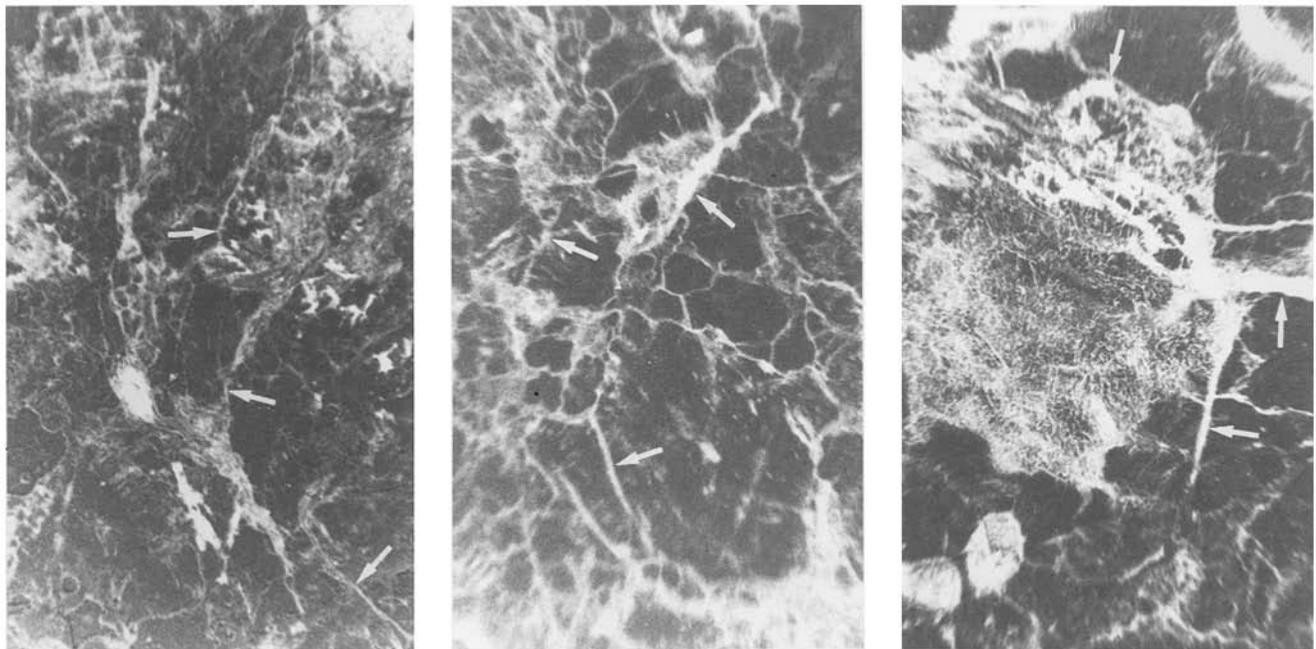


Fig 61 Thin section of a granite morainic block at 5200 m on the upper moraine of the Rongbuk glacier (Fig 58 ■ left; Fig 59 ■ left). The illustration is 4.1 mm long. The iron hydroxide veins on the bounding surfaces of the crystals (oriented along the cross-wires) are only detected in the microscope (—→). They are not visible macroscopically on the debris particles of the upper moraine which is only a few hundred years old; (cf. in contrast Fig 62 and 63). Laboratory photo: A. Heydemann and M. Kuhle.

Fig 62 and 63 Thin sections (1 mm long portions) of the same granite variety of moraine components as Fig 61, but from the 5650 m lateral moraine terrace of the Dhampu stadium (Fig 58 ■ right III; Fig 59 ■ right). The weathering period of at least 7000 years more in comparison to Fig 61 resulted in a macroscopically visible red colouration of the whole surface of the section. The veins of iron hydroxide on the crystal surfaces are appreciably broader (—→). Laboratory photograph: A. Heydemann and M. Kuhle.

glacier tongue to the position of the ice margin of stage VI is about 50 m (5180–5130 m asl). This allows a maximum equilibrium line depression of c. 25 m to be calculated for these historic to Neo-Glacial ice margin deposits. These strikingly small equilibrium line depressions are contrasted by ELA depressions of 280–470 m on the Mt. Everest S slope (Cho Oyu S slope, Kuhle 1987b, p. 205, Tab 2) during the past 2000 to 2500 years, or 150 m during the past 440 years. It is highly likely that this difference can be explained by the post-Late Glacial uplift of the Himalaya N side and S Tibet (cf. above). The glaciated Rongtö valley, which joins on the true left-hand, contains the corresponding moraine sequence to the main valley (Fig 58 VIII–X, VI). The recent glacier tongue of this c. 10 km long Rongtö glacier, with a maximum altitude of 7516 m for its catchment area terminates at 5500 m (28° 09' N/86° 49' E, Fig 59). The basis of the moraine surrounding the prehistoric tongue basin is at about 5410 m, so that the ELA depression is the same (Fig 59, VIII–X, VI). Terminating at a greater altitude in accordance with the lesser altitude of its catchment area (mean: c. 6550 m) the glacier of the tributary valley, like numerous other tributary glaciers of the Rongbuk valley system (Fig 60 ● ●), has aggregated a large pedestal or platform moraine with steep ice marginal ramps (IMR; continuous and transition cones;

Kuhle 1988) into the mouth of the main valley (Fig 58 ××). This platform moraine is situated 350 m above the main valley floor.

In comparison with the mass of glacier ice, the amount of debris collected for these platform moraines is uncommonly large; it is the effect of the increasing debris deposits of regressing glaciers (with a negative mass balance), which is a corollary of the increasing surface covering of such ice streams with debris. There is a cumulative concentration of debris in the valley glacier as it retreats in the course of centuries or even thousands of years (cf. below).

1.2 km down-valley of the Rongbuk glacier stage VI on the true right-hand lateral moraines of a stage V (Nauri Stadium) are preserved over a distance of 800–900 m. The Rongbuk monastery has been erected upon them. Corresponding lateral moraines have also been preserved on the true left-hand; both terminate in the same valley cross-section. Descending much more steeply, their upper edge (Fig 58 ↙ V) contrasts with Late Glacial lateral moraines of stages IV, III and possibly even II (Sirkung-, Dhampu- and Taglung stages), which continue further down the valley. The ice margin of the Nauri stage (V) lay at about 4950 m, c. 230 m below the recent altitude of the ice margin; this points to an ELA depression of 115 m.

Some notes on the problem of dating moraines in the area of the Rongbuk glacier system

There are no absolute datings. Due to the recent root penetration all the C-14 analyses carried out on the shallow soil formations beneath the meadow vegetation in altitudes of up to 5500 m failed, or indicated ages of only a few decades. In most cases it was possible to relativize these moraine ages geomorphologically or glaciologically as being too recent. Weathering permits a relative dating beyond the geomorphological method of dating moraines, which takes its bearings from succession, altitude above and distance from the actual body of ice. It was possible to establish at least two intensities of iron hydroxide formations with the aid of granite components on moraines (Fig 61–63). The Recent to sub-Recent, i.e. at most a few hundred years-old upper moraine of the Rongbuk valley glacier at 5200 m (Fig 58 ■ on the left) does not yet show macroscopically any iron hydroxide precipitation, and only very little microscopically (Fig 61 →). The moraine terrace at 5650 m, i.e. 450 m above the recent glacier surface in the same valley chamber (Fig 58 ■ right-hand), already shows distinct iron hydroxide colourings (Fig 62 and 63 →) and not only along the larger hairline fractures. The components of this moraine generation present a complete rust colouring, which is optically clearly visible when under the microscope. The location of these findings on the corresponding moraine terrace (Dhampu Stadium III) (Fig 58 ■ right-hand) is evidence that this weathering intensity is to be attributed to the middle Late Glacial Dhampu stage (III). A corresponding weathering intensity has been diagnosed with the help of thin sections of the very large eye-gneisses in the ice marginal ramps of the N slope of the Shisha Pangma (locality: Fig 40 III or IV), where the iron hydroxide formations continue from the grain boundaries far into the large albite-oligoclase and K-feldspar crystals (Fig 43 →). Near the surface the period of weathering may also have been at most about 9,000–15,000 years here.

Besides related or corresponding massive crystalline rocks (granite, gneiss), comparable edaphic conditions, which stand for the cold-arid climate of S Tibet, can be regarded as fulfilling the conditions for comparison of that sort. Comparisons were made between vegetation-free locations especially where there is no iron mobilisation by humic acids.

The geomorphological conformity of the entire moraine sequence from the neoglacial stage to the historic moraine formations on the S slope of the Mt. Everest group a mere 20–30 km away (Kuhle 1986d, 1987b) points to a greater age for moraine terraces IV and III than the 4000–4500 BP attributed to the moraines of the Nauri stage (V). These absolute age marks gained from the rich organogenic substratum of the S slope were ‘parallelized’ by means of the degree of corresponding geomorphological preservation and on the basis of the sequence of the moraine succession on the N



Fig 64 East Rongbuk glacier seen looking N down-valley from 6030 m. (Fig 2, halfway between No. 38 and No. 41). The viewpoint is on the medial moraine (●) between the confluent Changtse glacier and the main component of the East Rongbuk glacier from the NE slope of Mt. Everest (Fig 67, 68). The layer of debris protects the glacier ice from ablation so that the protected mounds (●) stand above the white unprotected ice (right); (cf. Fig 71, in contrast Fig 70). In the background (■) is an advancing corrie glacier. Photo: M. Kuhle.

slope. A study of the forelands of comparable glaciers proved to be useful for this. In terms of size and type the Rongbuk glacier is very similar to the Ngozumpa glacier, and therefore suitable for a comparison. The result is the following age sequence:

- Sirkung stage (IV) and older (Late Glacial) = older than 4000–4500 years
- Nauri stage (V) = 4000–4500 years
- older Dhaulagiri stage (IV) = 2000–2400 years
- middle Dhaulagiri stage (‘VII) = 2000 years
- recent Dhaulagiri stage (VII) = c. 440 years
- stage VIII = c. 320 years
- stages IX and X = less than 320 and more than 30 years old
- sub-Recent to Recent = less than 30 years old

Evidence of these stages is provided in Fig 58 and 59. A comparison of levels permits them to be traced in many places in the area of the Rongbuk glacier system, as for instance in the area of the E Rongbuk and the far E Rongbuk glacier confluence S of the 7071 m summit. During the deposition of the moraine in the Nauri and the older Dhaulagiri stages (V and VI), probably even well into the historic stages (‘VII–IX), the tongue of the small glacier in the hanging valley extended as far as the E Rongbuk glacier (Fig 64 V and VI). Downstream the

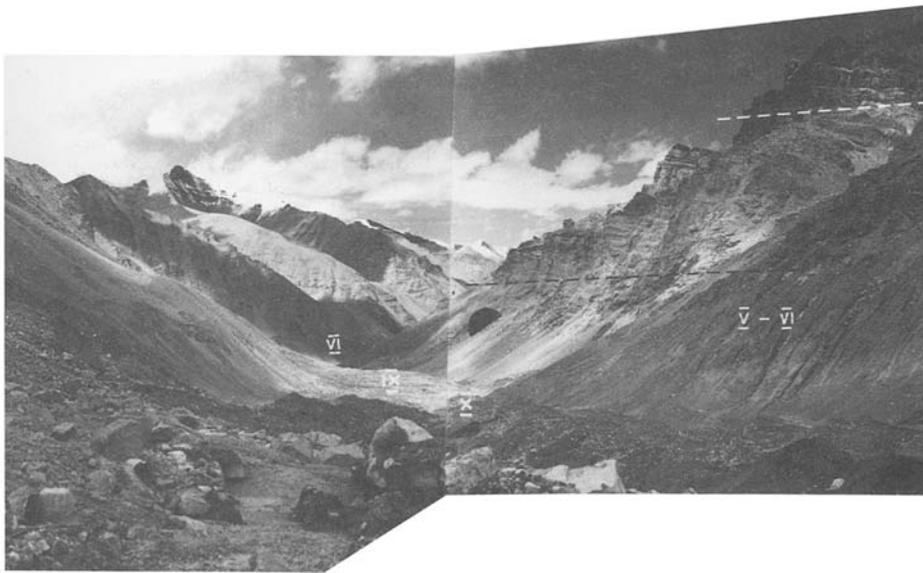


Fig 65
Debris covered lower East Rongbuk glacier (X) seen from 5750 m valley outwards towards the WNW at the confluence with the central Rongbuk valley (Fig 2, central between No. 38 and No. 41). Trim lines of the neoglacial (--- black) and late-glacial (--- white) on the migmatite rocks of the true right valley slope which is strongly weathered and joint eroded. Photo: M. Kuhle.

E Rongbuk glacier follows a narrow stretch in the valley, so that no lateral moraines, but polished areas have been preserved at the corresponding level on the true right-hand (Fig 65 ●). At 5500 m asl the E Rongbuk glacier tongue is now 2.2 km away from the confluence with the central Rongbuk glacier (Fig 65 X). During stages VI or VII it still reached it by way of a platform- (dam-) moraine. A historical tongue basin, filled with ice even 30 years ago (X – sub-Recent) is 300–500 m long (Fig 65 X-IX).

The relative heights of the neoglacial to historic remnants of lateral moraines decrease in direction of the upper glacier areas. A few hundred metres up-valley, they get closer and closer to the recent glacier surface until they come together in their source area in just a few, in two, or even a single level. Moraine levels are accordingly best differentiated in the vicinity of the present glacier tongue end (Fig 58, 59).

Some notes on the subdivision and geomorphology of the Recent and sub-recent Rongbuk glacier surface

By comparison with alpine glaciers, glacier morphology on Mt. Everest above the equilibrium line only differs in respect of the steepness of the catchment area. At its source the central Rongbuk glacier is surrounded by the 5 km-wide and up to 2500 m high N wall of Mt. Everest, the 3 km-wide and 1300 m-high S wall of Changtse (Fig 66), as well as secondary head-walls like the Khumbutse (6636 m) and Lingtren (6714 or 6697 m) walls. Relief conditions are similar above the ELA on the E Rongbuk glacier, where a 3.5 km-long and up to 1800 m high headwall between the E shoulder of Mt. Everest (8390 m) and of Changtse (7580 m) closes off the valley (Fig 67). N and S of the glacier basin the

Everest spurs and the ENE spurs of Changtse follow on over a 3 km distance with 600 m to more than 1500 m high ice flanks. Though not in contact with Mt. Everest, even the W Rongbuk glacier has very steep catchment areas, like the 1500 m high Gyachung Kang ESE flank, and the walls of other high peaks like the Pumori (7145 or 7170 m) with its N wall (Fig 69). This topography causes a considerable proportion of avalanche feeding of the glaciers. Avalanche erosion supplies the ice on the head-walls with debris. This explains the considerable surface moraine cover of the glacier tongues. Since head-wall areas tend to be only a few hundred or thousand metres away from the equilibrium line in up-valley direction, the surface moraine cover already begins just below the equilibrium line in *down-valley direction*.

This is a factual situation which contradicts the method of calculating the equilibrium line in accordance with Lichtenecker (1938), as it assumes the course of the equilibrium line to be generally 50 m above the place where the highest internal moraine appears (recovers by thawing) and becomes a surface moraine. According to observations conducted by the author, however, the internal moraine emerges to the glacier surface below the equilibrium line the sooner the fewer days the debris from the headwall is covered by snow from primary precipitation. How deeply it is covered by fresh snow depends on the time it takes to reach the equilibrium line, and thus on the distance to the foot of the wall (the point of impact of avalanche debris).

Surface moraines increasingly covering areas further down the glacier not only have the effect of changing the energy flow at the glacier surface, but also of a qualitative leap, geomorphologically speaking. A surface moraine which continuously increased its thickness also becomes a protection against ablation. Whilst thicknesses of debris deposits are not more than a few centi-

metres, the surface moraine still forms a concave depression let into the glacier surface; the cryoconitic effect of the debris – heated by insolation – has the effect of melting and moulding a box-shaped moraine channel down to 10 m deep in the surrounding, debris-free reflecting ice (Fig 70 and 69 ●). This form of melting away can be observed in the middle moraine of the E Rongbuk glacier, starting at an altitude of about 6300 m (Fig 70 ●) and descending over 3.3 km to 6000 m asl (Fig 71 ↓). From that point onwards the surface moraine becomes so thick (at least 20–30 cm, and exceeding 50 cm in some places) that it isolates the underlying ice from radiation. After this profile has been reached, the surface of the middle moraine thus returns to the level of the detritus-free ice (Fig 64 and 71 ●). For the next 3–4 km, down to 5800 m, the moraine-covered ice overlaps the open ice on both sides as a ridge that is gradually increasing in breadth (Fig 71, foreground).

At the same time it ought to be borne in mind that the other side of this melting balance is presented by the forced thawing down (and evaporating) of a glacier surface already broken down into ice pyramids (Fig 69). This multiplication of surfaces (and of the extend of the overall ice – surface) increases ablation as a result of the enlargement of areas open to very dry air passing through them. Further upwards the compact ice surface of the glacier remains intact (Fig 66). On a sunny day a specific humidity of 4.2 g/kg was measured directly on ice pyramid surfaces at 5650 m, and of merely 1.0 g/kg at the same time at a distance of 2 m. This humidity gradient is evidence of the great ablation efficiency of the air passing between the pyramids. At 6500 m and a somewhat greater distance from the glacier the dryness of the air temporarily fell to 0.2 g/kg, and even to a minimum of 0.13 g/kg.

This points to a bilateral change in the balance of the longitudinal profile of the glacier. Ablation in the vicinity of the equilibrium line is intensified by the thin cover of debris, and kept low by the still intact surface of the open ice (Fig 70 ■ ■). Further down the thickening cover of debris exerts an increasingly isolating effect, whilst areas of the glacier surface roughened up by ice pyramids dwindle with increasing speed (Fig 64).

The phenomenon of ice pyramid formation is almost independent of the moraine content of the ice. Ice pyramid fields on the tongue of the far-E Rongbuk glacier (Fig 72 ●), containing little surface moraine, are evidence of this. This ice stream arises in a saddle glacier with comparatively little relief as the W component. Although its catchment area reaches 7227 m asl, there is a complete absence of avalanche feeding. This is a case of the relief-specific, glacier-typological leap (Kuhle 1986b) from the ice streams of the High Himalayas to those of the Tibetan Himalaya (Inner Himalaya), which is marked by the near absence of steep walls. The feeding of the Tibetan glaciers is almost exclusively due to primary precipitation.

The formation of ice pyramids is interpreted as specific to subtropical climates, and attributed to the steep angle of radiation at 28° N. Radiation-intensifying factors are the great altitude above sea-level with its extremely radiation-transparent atmosphere (Kuhle 1988a, b), and the position in the monsoon-lee on the N side of the Himalayas, which reduces precipitation and clouding. The Khumbu and the Ngozumpa glaciers on the S slopes of Mt. Everest and Cho Oyu also carry ice pyramids, though to a lesser extent. They, too, find themselves already in the precipitation shadow of the Himalayan chains extending in front of them (with the 6000 to almost 7000 m high peaks of the Kangtaiga, Amai Dablang, Tramserku and others) and receive less than 600 mm of precipitation a year. In general the development of ice pyramids can be observed on all the glaciers of the Himalayan N slope, on the massifs of the Shisha Pangma, as well as on those of Mt. Everest. They appear even in places where small hanging valleys or corries lack breaks in slope which would cause the disintegration of the glacier surfaces (Fig 71 ■ ▼); Fig 64 ■). However, the author regards the vertical discontinuity surfaces of joints and fissures which develop to wide crevasses in the area of large ice falls as the structural protoform inside the glacier for the subsequent development of ice pyramids. This would imply an absence of ice pyramids in small joint-free hanging glaciers, and they are indeed much more markedly represented on large glaciers like the Rongbuk, which are detritically composed of numerous components, with many ice-falls (Fig 69, 72 foreground). But less deep crevasses, of smaller extension, are evidently also sufficient for the formation of ice pyramids. Thanks to their very low viscosity, they occur in almost all cold hanging glaciers, and cause the formation of ice balconies (Fig 71 ■ ▼).

A longitudinal profile of the central Rongbuk glacier provides a clear outline of the genetic series, and thus of the formation of ice pyramids: ice crevasses occur at the level of the equilibrium line and above (Fig 66 ■; below the N wall of Mt. Everest, the W wall of Changtse, as well as E of the Khumbutse and Lingtren walls). Below the scarp section or other unevennesses in the glacier bed they mend again and section by section gets covered by snow (precipitation). As they are not entirely frozen together, the progressive melting process below the equilibrium line prefers and thus emphasizes the fault cracks as they move down-valley (Fig 73 ↓). Radiation follows the intra-glacial network of crevasses and makes the unbroken blocks stand out residually (Fig 73 ◆). Besides tension crevasses running across the direction of movement, there are also longitudinal fractures in the ice resulting from differing flow velocities. This explains the rectangular to almost square ground-plan of the pyramids (being parallelogram-like distorted in case of shear-stress). But before this is revealed, ice walls begin to form across the glacier at first (Fig 70 ◆), which are separated by increasingly deep furrows frequently filled

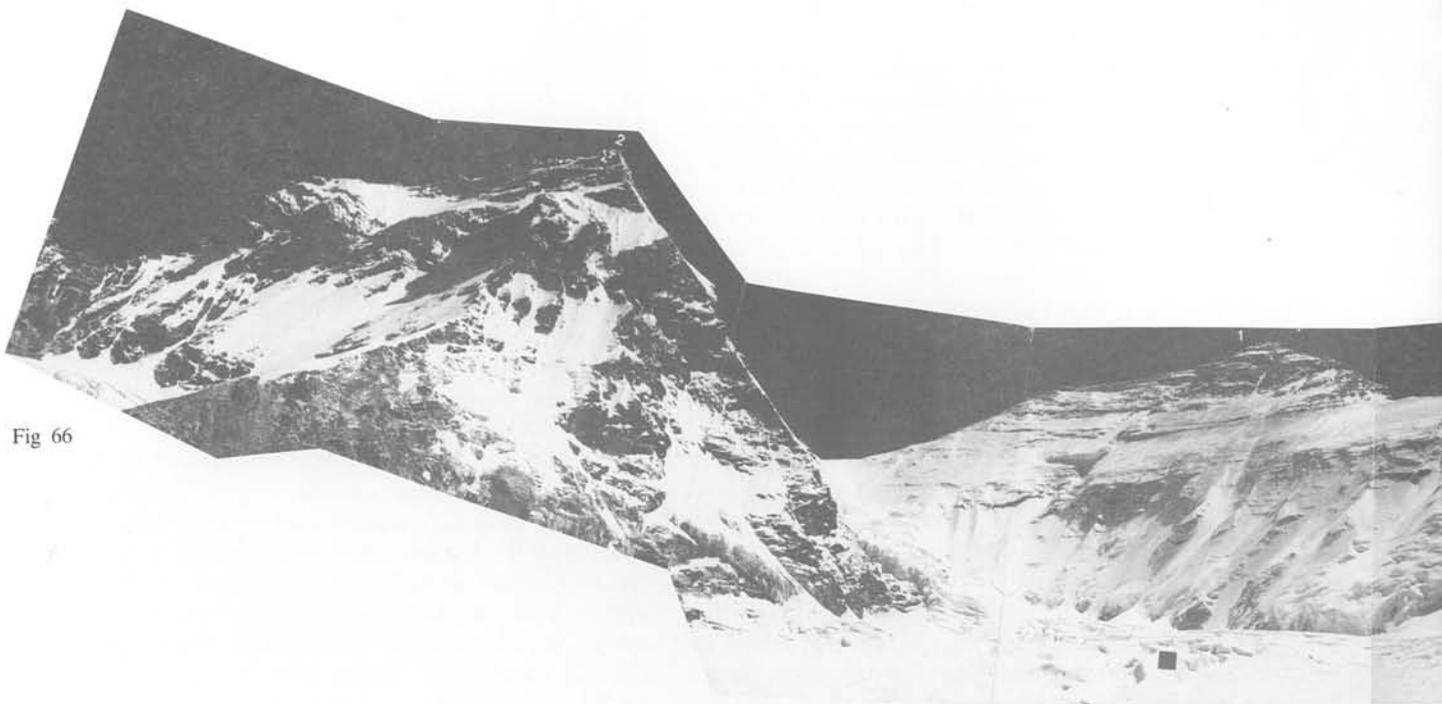


Fig 66

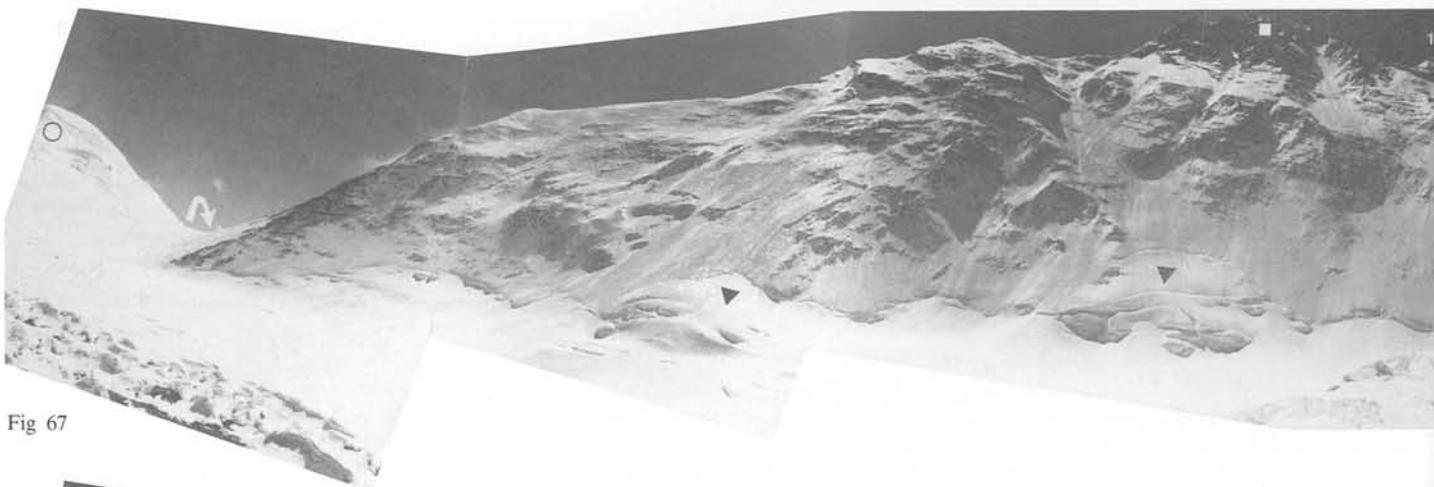


Fig 67

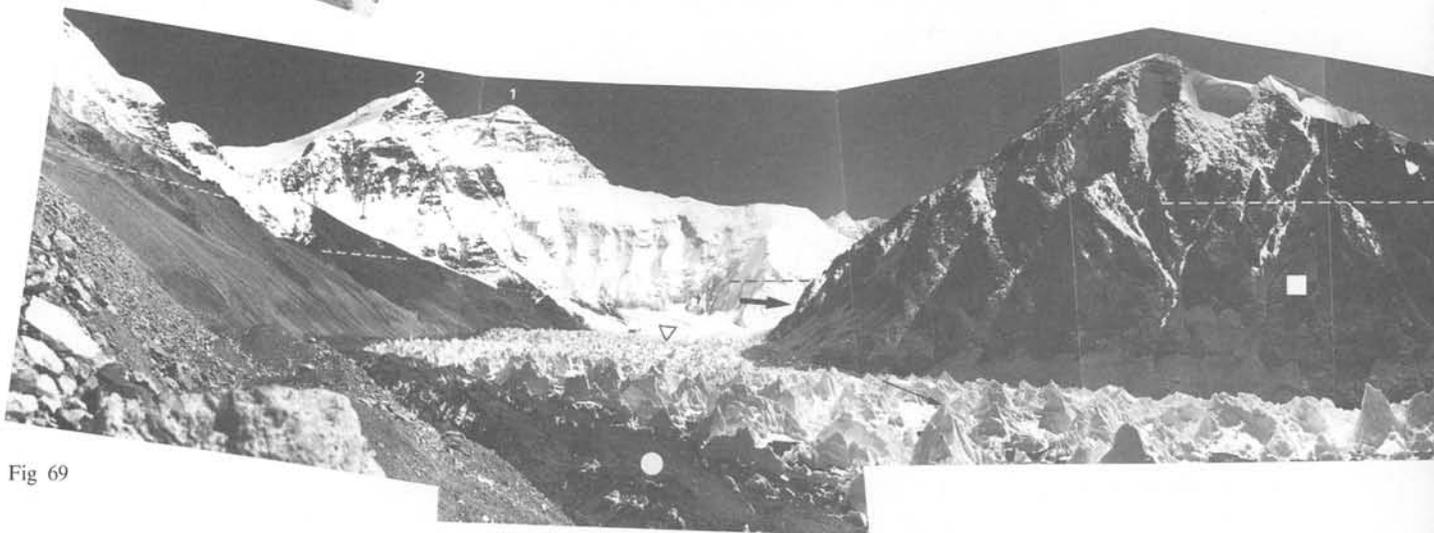


Fig 69

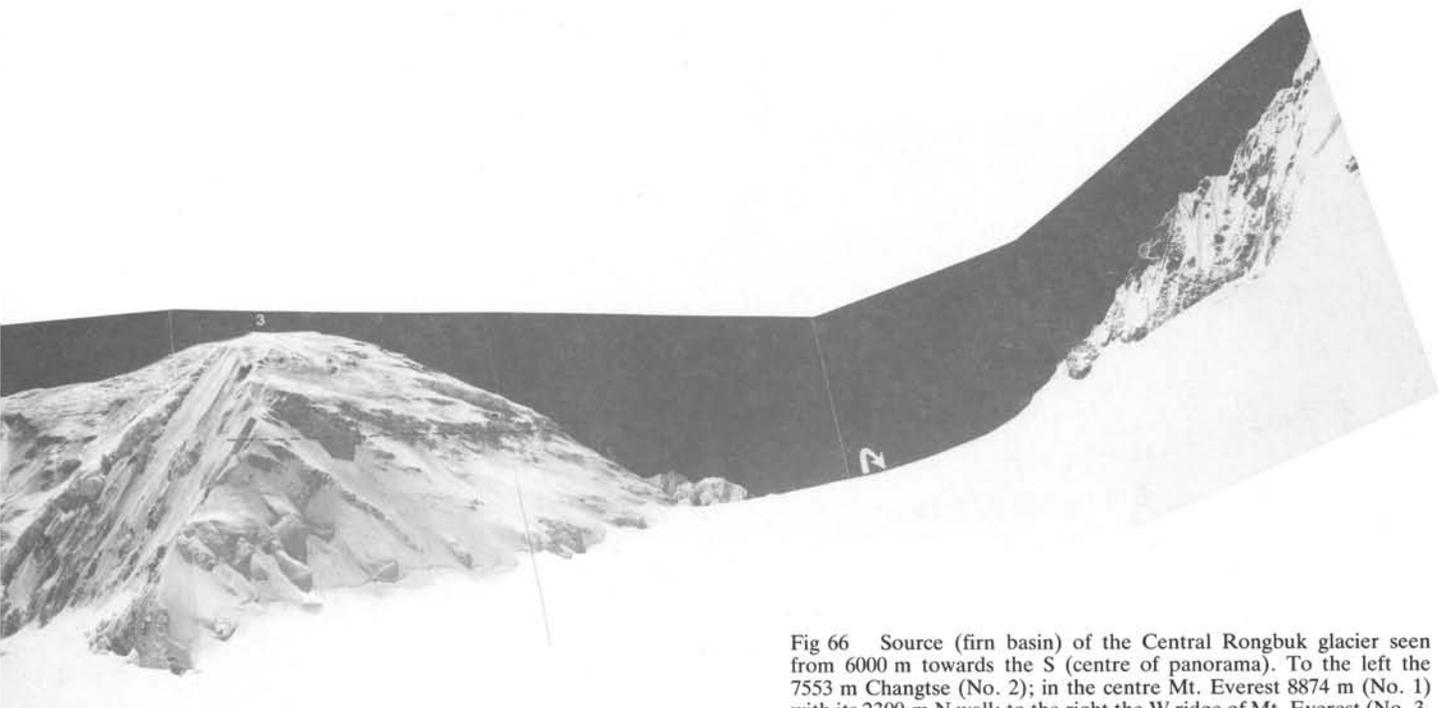


Fig 66 Source (firn basin) of the Central Rongbuk glacier seen from 6000 m towards the S (centre of panorama). To the left the 7553 m Changtse (No. 2); in the centre Mt. Everest 8874 m (No. 1) with its 2300 m N wall; to the right the W ridge of Mt. Everest (No. 3, 7205 m) which falls to the 6009 m Lho La (▲) (Fig 2, No. 40). Here the Rongbuk glacier overflows on to the Khumbu glacier on the S slope of the Himalaya. (---) marks the level of maximum glacial surface in this area of glacial pass breaching (cf. Fig 74). Avalanches are responsible for at least 50% of the supply to this source basin. Photo: M. Kuhle.

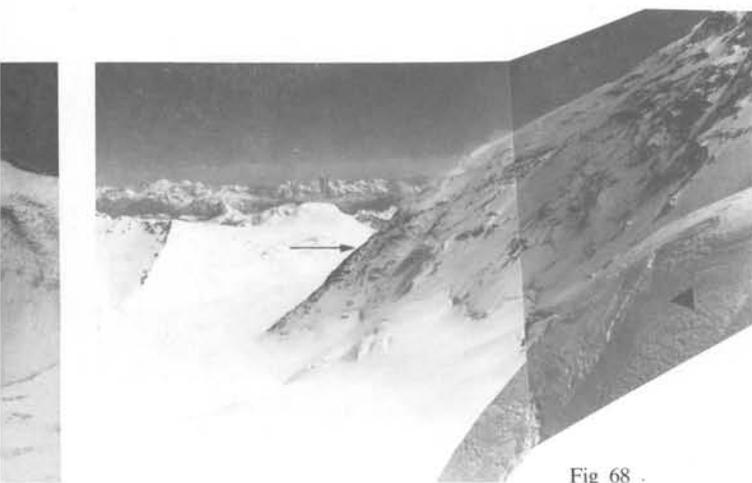


Fig 68 .

Fig 67 Panorama of the East Rongbuk source basin with the 6500 m glacial breach Rapiu La (Raphü La ▲; Fig 2, No. 41; Fig 68) leading to the S slope of the Himalaya; view from the true left edge of the glacier at 6510 m towards the NE (centre of panorama). The avalanche supply is largely from the up to 1800 m high E part of the Mt. Everest N wall, coming from the 8395 m high NE ridge of Mt. Everest (■). Typical avalanche cones are formed at the foot of the wall (▼▼). To the right above is the main peak of Mt. Everest at 8874 m (No. 1). Although the 6833 m peak NE of Rapiu La (○) is covered with glacier ice some decametres thick the ice covering the slopes of the Mt. Everest N wall decreases steadily above 7200 m where rock comes to the surface. The only partial snow cover is that blown into gullies (left and right of ■). Photo: M. Kuhle.

Fig 68 The central source region of the East Rongbuk glacier seen from 7030 m on the N col of Mt. Everest. (→) shows the transfluence of the present glacier ice (much thicker during the pleistocene) across the Rapiu La (Raphü La; Fig 2, No. 41) in the S slope of the Himalaya. This same overflow limited the maximum glacial accumulation of ice on the edge of the S Tibetan plateau to only a few hundred metres more than at present. In the foreground (♣) are the persistent lee side cornices (with ice cores) of the Changtse SSE ridge. In the middle distance (right) the eastern N wall of Mt. Everest falls from its 8395 m NE-spur into the firn basin. Behind are the glacier mountains of the Tibetan Himalaya. Photo: M. Kuhle.



Fig 69 Panorama showing the relationship of the central (left) and W Rongbuk glaciers, one of the major recent valley glacier systems of the Himalaya N slope, seen from 5520 m towards the SW (centre of panorama) (Fig 2, between No. 38 and No. 40). The valley of the Central Rongbuk glacier ends at the summits of Mt. Everest (No. 1) and Changtse (No. 2, 7553 m) 11 km away. The pleistocene surface (---) of the glacier system was controlled by the Lho La transfluence pass (→, Fig 2, No. 40) as is shown by the lateral moraines and the undercutting in the country rock. In the area of the at least 12 km long W source glacier (W Rongbuk glacier ▲) further transfluence passes led to the reversal of the outflow direction of this portion of the flow during the glacial period (Fig 2, No. 39) The central mountain spur at the confluence (■) shows the great post-glacial modification of the lateral scouring (below ---) by frost weathering and gullying. Photo: M. Kuhle.



Fig 70

Medial and surface moraines (●) of the E Rongbuk glacier viewed down-valley from 6300 m towards the NNE (Fig 2, halfway between No. 38 and No. 41). To the left (■) a footwall and slope glacier component from the E slope of Changtse; to the right (■) the intact surface of the E Rongbuk glacier not so far ablated into ice walls and pyramids. Above here the medial moraine (●) is only a few centimetres or decimetres thick so that the dark debris particles experience daily temperature increases and thus promote ablation. Therefore the dark bands of medial moraine (●) are more deeply melted into the glacier surface than the bounding light ice surface (■■; contrast Fig 64 and 71). Where patches of snow and firn persist and slope glaciers occur the strata-etched slopes (background ▽▽) are polished and smoothed. Where no snow-shed activity or glacial bottom scouring occur steep arrowhead slopes and rock steps are typical (□). Photo: M. Kuhle.

with internal moraines. It is true, too, that irregularities (due to shear-stress) in the form of diagonal bridges have created connections between the alignments of the walls. They are attributable to small-scale fluctuations of velocity in the glacier. Some kilometres down the glacier longitudinal furrows have been added and separate the walls into pyramids. At first the pyramids are truncated (Fig 73 ◆), i.e. standing out, they still conform to the glacier surface (Fig 69 ▽). As they melt down the pyramid flanks consume more and more of these small remnant areas until they meet at one point (Fig 72 ▲; 69 ▼). Yet further down the glacier, as a function of the width of the pyramid base, the ice pyramid peaks are

lowered at different speeds. From this point onwards the former level of the glacier surface in the sense of an “upper denudation level”, which is lowered simultaneously by melting, is subject to an irregular dissolution i.e. the speed of melting differs from pyramid to pyramid (Fig 69, foreground, right-hand side ooo; Fig 71, right and left ○○○).

Ice pyramids occupy an intermediate position in the transition from an unbroken snow surface above the equilibrium line to the complete covering with surface moraine in the lowest section of the glacier tongue. In this the surface moraine dovetails with the white areas of ice pyramids in a linkage pattern of counter-flow. The

Fig 71 View from the ridge of the medial moraine of the E Rongbuk glacier (●, Fig 2, between No. 38 and No. 41) at 5800 m upward to the SE to the Rapiu La (↓) No. 41. To the right (No. 1) the rocky upper 800 m of the N slopes of Mt. Everest in front of which is the firn-covered NE ridge of the 7553 m Changtse (No. 2); left (No. 3) a nameless peak of the Tibetan Himalaya at 7050 m. This carries relatively gently inclined slope glaciers typical of S Tibet (■). The true right bank walls of the trough valley of the E Rongbuk glacier (left on both sides of ▼) are broken up by frost weathering into abrupt steps, ridges and gullies. The medial moraine components (●) with their considerable debris thickness stand above the mean level of the white ice with its ice pyramids (○○) because they protect the ice below from ablation. Photo: M. Kuhle.





Fig 72 The geomorphological and glaciological phenomena of S Tibet provide a picture of cold-arid climatic conditions. The frost smoothed slopes (Glatthänge ■■) are the result of the great freeze-thaw frequency of subtropical highlands and mountains. Snow patches are only to be seen in summer after monsoon snowfalls. These lie high above the ends of the glaciers, generally only above the equilibrium line (above ■ right). The up to 30 m high ice pyramids of the ablation region of the glaciers (↑) are the result of a maximum global radiation of over 1200 W/m² together with an absolute humidity of only 0.15 to 2.01 g/m³ (at 200 cm above the surface) as measured during the 1984 expedition. The low temperature results in low fluidity of the ice, resulting in the strange forms of the pyramids and their persistence. The ice flows as discrete blocks. View from the medial moraine of the E Rongbuk glacier at 5800 m (Fig 2, between No. 38 and No. 41) towards the E to the far east Rongbuk glacier (●). Photo: M. Kuhle.



Fig 73 View WNW from 6060 m from the E Rongbuk glacier (Fig 2, between No. 38 and No. 41) over the Changtse glacier to the 6855 m peak (No. 6). The development of the compact ice surface (↓) into ice walls (◆) is here obvious; further down-glacier this allows the formation of ice pyramids (Fig 71 ○○○ to the right). Photo: M. Kuhle.

ice fields decrease as they move down, whilst the areas with debris increase in size (Fig 64, 69, 71) until (from 5400 m asl on the central and W Rongbuk glacier) the ice pyramids thin out completely and the glacier tongue is entirely covered by surface moraine (for another 5 km on the central Rongbuk glacier) (Fig 58, 59, 65). The author would wish to make the point that the ends of the ice pyramid bands on the glacier fluctuate more readily than does the debris-covered glacier snout. This allows smaller climatic fluctuations to be registered than those of the larger valley glacier permit. Consequently they provide a less disturbed picture of the specific climatic conditions. Since the 1:50,000 mapping of the central Rongbuk glacier in 1966 (Academia Sinica 1979), the lowest ice pyramid spurs have retreated about 1.1 km, whilst the glacier terminus at most has witnessed a minor up-valley shift of the dead ice line.

In general, even on the Yepokangara glacier on Shisha Pangma the retreat of ice pyramids shows a negative balance for recent valley glaciers of more than

10 km in length (Fig 59). The same is indicated by the increasing deposition of debris (Fig 59, 65) of even small tributary glaciers (cf. above, Fig 60).

Reconstruction of ice levels and ice overspill into the southern slope of the Himalayas: some observations on the overall picture of prehistoric glaciation in South Tibet

The valleys of the Himalayan S slope, like the Rongbuk valley and the Dzakar Chu, its continuation, as well as the valley which continues the Kyetrak glacier on the Cho Oyu, slope gently and in conformity with the gradient their recent valley glaciers flow N, just as they did in historic and neoglacial states and back in the Late Ice Age. The northerly direction of the flow was coupled with equilibrium line depressions of just a few hundred metres. Due to the postglacial glacio-isostatic uplift it is not possible to determine the ELA depression more precisely by way of moraine remnants in the Dzakar Chu, up to which the glacier discharge in a northerly direction took place. In any case the equilibrium line depression at which a glacier discharge towards the N was possible, may well have been greater at that time than at present, since glacio-isostatically the relief N of the Himalaya must have been situated at an even lower altitude.

Previously, during the older Late Glacial period and probably still during the glacial state of the Dhampu-



Fig 74

View at 5300 m to the NE to the Lho La (▲) from the S slope of the Himalaya (S of the watershed) and the Khumbu glacier (Nepal in the foreground; Fig 2, No. 40) (cf. Fig 66). To the left (No. 5) is the 6640 m Khumbutse, in the centre (No. 2) Changtse (7553 m) and to the right (----) the undercutting below the W spur of Mt. Everest showing the maximum glacial surface. The S flowing overflow glacier was thus 400 m thicker. Even today part of the Rongbuk glacier ice still flows across the Lho La (-Pass ▲) at 6010 m to the Khumbu glacier breaking as avalanches from the overhanging ice ledges (■). Photo: M. Kuhle.

(III) or Taglung stage (II), but certainly during the Ghasa stage (I), the entire relief had been infilled with ice. At that time the glacier discharge must have taken a course to the S. Ice-fillings in the valleys N of the High Himalaya had reversed their flow direction once they had reached a certain thickness of many hundreds of metres. In the course of this process, which extended from the Early to the High and Late Glacial period, their surface came to slope southwards instead of northwards. This shift of the ice-shed from the Himalaya to the N can be compared with the N European glacial ice-shed shift from the Scandinavian Alps to the inland ice-dome in the SE. At present it is merely a matter of hanging glaciers on the Himalayan flanks, whereas the ice ages saw massive glaciers flow through (and leave) the High Himalayas. Present glacier feeding is restricted to the slopes of the High Himalayas, whereas it then took place to the N in the area of the S Tibetan ice-stream network, which filled the relief of the Tibetan Himalaya (Kuhle 1982, 1983). This shift in feeding areas even went so far that the then outlet glaciers received almost no ice from the walls of the High Himalayas, at least less than is now available. At that time the thermal upper limit of glaciation (cf. Kuhle 1986a) running parallel to the equilibrium line, had been depressed by more than 1000 m to an altitude of about 6000–6200 m. The surface of the outlet glaciers tended to be at least at this altitude, and usually even higher, so that dry, i.e. glacier-free, walls rose above these glacier streams. On the Cho Oyu it was the now also glaciated Nangpa La (5717 m) to the W, and in the Mt. Everest area, as the more closely defined area under investigation, the passes

of the Rapui La (6500 m; Fig 67 ▲ ; 68 →) to the E, and Lho La (6009 m; Fig 58, 69 →, 66, 74 ▲) to the W of the mountain, which opened the way for the discharge of those outlet glaciers on to the S slope of the Himalayas. Reference must also be made to the 5860 m high Nup La and the 6147 m high pass between Mt. Everest and Cho Oyu (Fig 69 ▲). All these glacier passes are in one line with the Himalayan main ridge and together they form the present-day ice-shed (Fig 2, no. 39, 40, 41), so that one can no longer speak of an overflow from the N slope to the S slope. Nonetheless, they do give the impression of such a prehistoric overflow through the fact that the glacier surfaces, especially those of the central Rongbuk glacier on the Lho La (Fig 66 ▲) and the E Rongbuk glacier on the Rapui La (Fig 68 →, Fig 2, no. 41) are almost horizontal, and that the ice in the pass zone accumulates increasingly before discharging. The process of an overflow from the N slope is thus suggested by these passes, the Himalayan S slope of which is characterized by a steep drop with ice balconies tens of metres thick. The thickness of the glacier at the time of the prehistoric overflow across the Lho La has a witness in the glacial undercutting on the W shoulder of Mt. Everest (Fig 58 -----; Fig 69 -----; Fig 66 -----; Fig 74 -----). The upper limit of glacial scouring runs at about 450 m above the present Lho La. The level at 6400 m is about 300 m above the highest right-hand side E moraine terrace on the upper central Rongbuk glacier (Fig 69 -----; Fig 59 ······; Fig 58 -----).

The moraine at about 6100 m is evidence of the most recent Late Glacial stage, during which the glacier over-

flow into the S slope persisted. It is about 5.3 km from the fully 100 m lower overflow of the Lho La (Fig 2, no. 40). On the one hand it depended on the altitudes of these four overflows how thick the glacier above was and how much ice flowed across them and down on to the S slope. On the other hand, it also depended upon their location in relation to the general direction of main valley axes and main valley discharge to the S. Although the Nup La (Fig 2, no. 39) is 200 m lower than the Lho La, the latter, being an extension of the main valley line of the Rongbuk valley, presented a much better discharge route for ice than the Nup La, situated at right angles to the extension of the W Rongbuk valley. It follows that the main discharge must have been by way of the Lho La (Fig 2, no. 40). Highest of all is the Rapui La (Raphü La, 6510 m) at the end of the E Rongbuk valley (Fig 2, no. 41). Its location is only marginally less favourable than that of the Nup La. However, overflow was least here due to its height.

Large-scale evidence of an almost relief-filling ice-stream network glaciation N of the High Himalayas, i.e. in the Tibetan Himalaya, has been preserved in the mountain ridges rounded by glacial forms of polishing. A good example is the terrain N of and down-valley of the Rongbuk valley (Fig 53–55, 57). The glacier discharge through the Rongbuk valley southwards to the High Himalayas and through it, is shown by Late Ice Age lines of ice scouring on the true left-hand in the central Rongbuk valley ($28^{\circ} 16' \text{ N}/86^{\circ} 48' \text{ E}$), which are inclined to the S in contrast to the valley gradient (Fig 58 ▼▼). On the Himalayan main ridge the level of the High Glacial ice surface was at an altitude of 6200 m and 6400 m. Towards the N, the Dzakar Chu region in the Tibetan Himalaya, it rose to approximately 6500 or 6600 m.

This interpretation of the reversal of the glacier discharge as a result of the Ice Age filling of the relief with glaciers explains the scarcity of moraines in the Rongbuk valley, outside the neoglacial glacier locations. The absence of moraines here is in marked contrast to the large-scale, substantial moraine cover in the N foreland of the Shisha Pangma group a mere 120 km further W (cf. above). The floor of the Rongbuk valley was accordingly covered by several hundred metres of near-motionless ice. There were therefore little or no debris deposits. Only the upper ice strata, being at a higher level, were able to follow the small gradient to the transfluence passes into the S slope. The tongue levels of the remnant glaciers only followed the actual relief of the valley landscape after far-reaching deglaciation had taken place. Only then was the aggregation of moraines intensified. This development set in with the more recent Late Glacial period (stages III and IV) and continued during neoglacial time (V–VII).

The prehistoric ice overflow into the S slope also explains the relatively low height of the oldest lateral moraine terraces above the recent Rongbuk glacier. These moraines are only fully 600 m above the present

surface moraine (Fig 58, 59, 69 —). The small prehistoric ice thickness of about 600 m below the recent ELA is out of the question for such a shallow valley floor gradient without overflow towards the S. Without overflow an enormous build up of ice would have occurred in this valley grid even in the Late Ice Age, for this valley cross-section at the time of the moraine deposits was at or above the level of the ELA. This is the climatically optimal altitudinal area for glacier build-up. This build-up of ice was consequently prevented by the overflow passes leading to the S slope of the Himalaya.

The collective findings concerning the topography explain the reasons for the striking paucity of traces of glacial accumulation on the N slope of Mt. Everest and the contrast they present to an Ice Age equilibrium line about 1200 m lower than today (see Tab 1).

Summary Review of Equilibrium Line Depressions in South Tibet and Climatological Conclusions

In the area under investigation, the recent orographic equilibrium line fluctuates between 5700 and 6325 m. This applies to the Transhimalaya at $29^{\circ} 43' \text{ N}$ and the N side of the High Himalayas in the precipitation shadow of the main ridge at $28^{\circ} 26' \text{ N}$. The mean average value to be cited as the integral, i.e. the macro-regional climatic equilibrium line, is around 5900 m (see Tab 1). The mean value of the High Glacial equilibrium line depression is 1100–1200 m, so that the integral climatic equilibrium line must have been at an altitude of 4700 m. This depression value very likely still is too small in comparison with the ELA depressions 300 km further W, where, N of the Dhaulagiri and Annapurna Himalaya values of 1530 m, orographically even 1630 m, were found (Kuhle 1982). In places where glaciers were able to penetrate the Himalayan main ridge and descend steeply on the S slope, the detailed analyses show an orographic ELA depression of up to 1900 m. This value applies to the Bo Chu in the Shisha Pangma massif. This is the only location where it was possible to survey the terminus of an outlet glacier in the area of the 1984 investigation. But even the ELA depression for the Ice Age Khumbu glacier (Dudh Kosi glacier on the S slope of Mt. Everest), which flowed down to at least 1800 m, to the Lumding – Drangka confluence ($27^{\circ} 38' \text{ N}/86^{\circ} 42' \text{ E}$) (Kuhle 1988b; Heuberger 1986, p. 30) was about 1400–1500 m. This additional information supports the findings that an ELA depression of 1200 m is likely to be on the low side of operational values for S Tibet. It is confirmed in detail by local and orographic values of the Transhimalaya at Lhasa as well as in the Tibetan Himalaya (see Tab 1). It must, moreover, be taken into account that values like those E of the Man-ko-pa basin and from the Latzu massif have been added to those of the High Glacial period, in order to be on the safe side, but probably belong to the Late Glacial

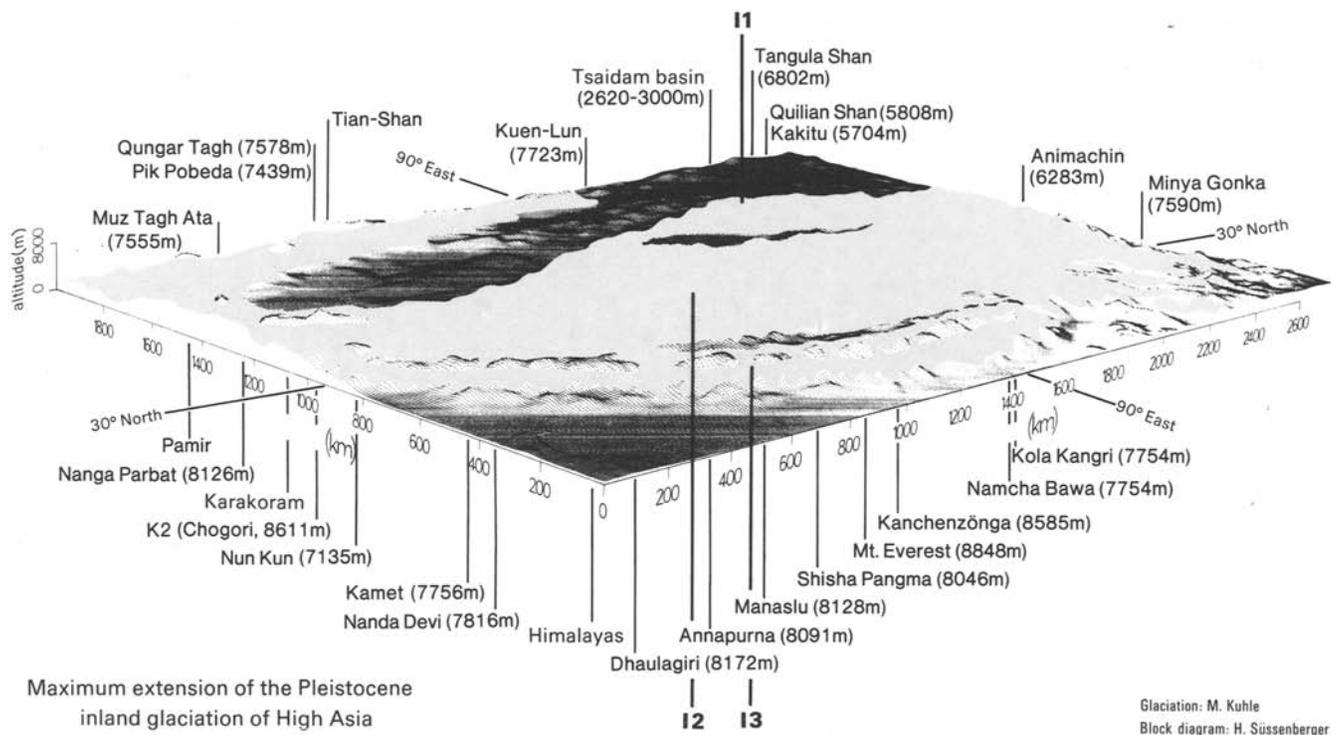


Fig 75 Fig 1 (which includes all the areas investigated so far) shows that the glacier surface in Tibet (with its surrounding mountain systems such as the High Himalayas in the S) covered 2.4 million square kilometres. In addition a network of ice streams occurred towards the NW in the adjacent Tien Shan range. The glacier area reconstructed in this paper lies in the area of I2 and I3 N of Shisha Pangma and Mt. Everest and also N of the ice-free Tsangpo valley (cf. Fig 76).

period. In the basin of Man-ko-pa an ice margin position on the basin floor at 4400 m was taken into account, though there is a near certainty that the entire basin was filled with ice during the Middle Ice Age (see above). A distinctly lower equilibrium line would accordingly have to be assumed than in a case where a glacier merely reaches the basin floor. Such a lower equilibrium line can, however, not be reconstructed. This touches upon one of the major problems of palaeo-climatology concerning equilibrium line reconstructions in the regions of the Tibetan Plateau: all the equilibrium line values are merely minimum figures. From the stage at which glaciation reached the level of the plateau, even a complete inland ice cover was able to establish itself without having a *single* lower moraine as evidence.

Assuming a mean gradient of $0.7^{\circ}\text{C}/100\text{ m}$, which, according to the author's measurements (Kuhle 1988a) applies to the K 2 glacier between 4100 and 5300 m asl, and is characteristic of cold-arid high mountain climates, the ELA depression indicates a High Glacial lowering of temperature of 8.4°C for the warmest month. This estimate is based on recent conditions of radiation and of hygric relationships (Kuhle 1983, p. 90). This degree of cooling is, if anything, too low a value when taking into consideration that it must have been drier during the High Glacial period than now, thanks to global cooling and the atmosphere's lower humidity capacity. Long-

term means of annual precipitation, which can now be measured in the lowest valley locations of the area under investigation fluctuate between 270 mm (Gyantse, 4000 m) and 440 mm (Lhasa, 3730 m). As a result of the Tibetan ice sheets reducing or completely preventing the monsoon, they must have been lower then. A prevailing shallow cold high pressure area with katabatic winds must be assumed to have been stationary over the Tibetan ice. On the 1986 expedition to K 2 (Karakoram) from the N annual temperature of -9° to -10°C were ascertained at the level of the equilibrium line, and the same values are known to occur on the Shisha Pangma N slope (cf. Ding Yongjian 1987, p. 10, Tab 3). (For comparison: the Alpine values are around -3° to -4°C). Thanks to increased aridity the Ice Age Tibetan glacier may, if anything, have been even colder. These considerations indicate a summer cooling down by a likely 10°C or more.

In the area under investigation the annual 0°C isotherm runs between 5000 and 5500 m, which is the same as the mean altitude of the termini of valley glacier tongues. (At the altitude of 4000 m the annual temperature is 5° – 6°C). According to the measured data mentioned above, the -10°C isotherm must have been at about 4700 m or still somewhat lower during the Ice Age, with the 0°C isotherm running at an altitude of at most 3800–4300 m. It is likely that the 0°C line had

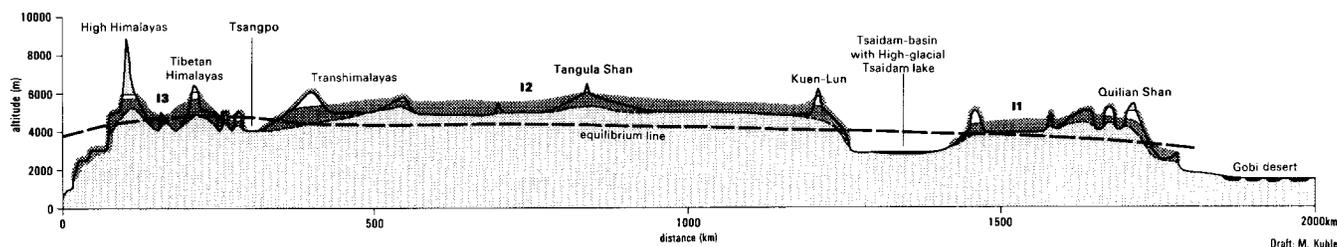


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been shifted markedly lower still by the glacial self-intensifying cooling process.

Late and neoglacial glacier development, successive equilibrium line uplifts since the High Glacial period, and the corresponding warming up in S Tibet may be gathered from Tab 1 by analogy with these arguments.

The Ice Age Glacier Cover in South Tibet and Its Indicator Value for the Reconstruction of the Tibetan Inland Ice as a Whole

High Glacial equilibrium line depressions of around at least 1200 m defined by ice margin positions of smaller prehistoric glaciers, together with glaciated forms of valleys and polishings, are evidence of an almost total glacial cover for the Transhimalaya, the Tibetan Himalaya, and the High Himalaya (Fig 75, between Shisha Pangma and Kanchenzönga). In terms of typology this glaciation is to be classified as an ice-stream network. Individual peaks and ridges rose above the glacier surfaces. This is a function of the extreme relief energy, especially in the High Himalaya. In the region of I 3 (Fig 75, 76) the proximity to the S escarpment of Tibet exerts its influence. Discharge was greatly accelerated here. The ice level inclined southwards towards the overflow passes of the High Himalayas. For this reason extreme build-up of ice was impossible. Circumstances in the Transhimalaya were similar. It formed the escarpment of I 2 (Fig 76) to the Tsangpo furrow, which was probably free of ice in its E section. From the central inland ice sheet in Tibet, the basis of which was more than 5000 m asl (Kuhle 1985, 1986c, 1987a, 1988b) outlet glaciers flowed through the transverse valleys of the Transhimalaya, almost reaching the Tsangpo down at 3950 m (Fig 75, 76, I 2). Evidence of the glacier level dropping to these ice margins is supplied here, as in the High Himalayas, by nunatak-like pointed ridges. They rise several hundred metres above the ice level. In the Transhimalaya there was consequently an ice-stream network as well. The finding of erratic blocks along its constituent streams is evidence of glacier thicknesses of at least 1200 m. These statements on the Transhimalaya

apply only to the section E of 86° – 87° E, however, which was investigated in 1984, and in particular to the Nientschen-tang La. Further W the valley floor of the Tsangpo rises to 4600 m asl, and must have been completely filled with ice according to the reconstructed equilibrium line depression. It is likely that the greatest ice thicknesses in S Tibet formed here. Ice thicknesses building up in the course of this development through back-damming assisted the progressive integration of the Transhimalayan ice-stream network into a continuous inland ice I 2/I 3. There must have been a transitional zone between 87° and 86° E where ice just about filled the Tsangpo valley. It was, however, still part of the ablation area, i.e. it contained glacier areas below the equilibrium line. In this section of the E–W profile of this transitional region, the Tsangpo valley reaches an altitude of about 4300 m and extends only 400 m below the Ice Age ELA.

It has not hitherto been possible definitively to exclude the possibility that some of the S parallel valleys between 86° and 88° E, like the Tingri and Lulu basins, were free from ice during the High Glacial period, although their floors are situated around 4300–4400 m asl. With an ELA around 4700 m, this would imply that 300–400 m lower valleys in the vicinity of the 6000 m-high mountains, at present still glaciated, – more than 8000 m high mountains in the case of the Tingri basin – would have remained unglaciated. This would be highly unlikely, and indeed difficult to explain, especially since the glacier tongue fronts would have undergone approximately twice the amount of depression undergone by the ELA. A comparison: in the W Alps, as for example in the case of the Rhone valley, a very large longitudinal valley with mountains rising 2000 m above the ELA and a floor that was found to have been 1600 m below the Ice Age equilibrium line, was filled completely with ice (1600–2000 m thick). How could it have happened that this filling should not have taken place in the lowest valleys 300–500 m below the ELA, and with the catchment areas rising to at least 1300 m above the former? Even greater aridity fails to explain this.

The significance of the investigations in S Tibet for the reconstruction of the entire Tibetan ice sheet is to be

found in the southernmost, and thus warmest, position in the highland (Fig 1, no. 4). This selective experimental arrangement of nature permits statements to be made which go beyond the area under investigation in the narrow sense. In the S, the equilibrium line today attains its greatest altitude in Tibet altogether – by rising to 6000 m or beyond. If it was possible to obtain evidence of large-scale prehistoric glacier cover here, it is simultaneously proof of a Tibetan glaciation further N. This deduction is supported by the following facts: 1) The recent ELA inclines towards the N – as it must have done in prehistoric times as well, due to the planetarian decrease in temperature. 2) The mean altitude of the central plateau rises from the area under investigation in S Tibet to large continuous plateaux at altitudes between 5000 and even 6000 m in the N (on the Mayer Kangri 33° N/86° E, for instance). 3) These plateaux are up to 7000 m high mountains, some of which have considerable recent glaciations including plateau glaciers, on top. These functioned as “crystallization centres” for a large-scale build-up of glaciers already at a time when the ELA depression was only 300–500 m. 4) Precipitation increases from 200 mm on the leeward side of the High Himalaya, i.e. from the Tibetan Himalaya, to 400 mm/year on the Transhimalaya and in the central high plateau in the N.

Based on the findings made in S Tibet, these four points were the confirmation and completion of the results of a prehistoric inland ice-cover of Tibet of approximately 2.4×10^6 km², which were obtained in W and N Tibet (Kuhle 1987a, 1988b).

Summary

The last Ice Age (Würm) glacier cover was reconstructed on the basis of standard geomorphological

indicators in S Tibet between the S slope and N slope of the Himalaya by way of the Tibetan Himalaya to the Transhimalaya (28°–29° 50' N/85° 40'–91° 10' E). At the same time, though subject to varying density of data, the process of Late and Post-Glacial deglaciation to Neo-Glacial and Recent glacier cover was considered. Evidence of an almost total glaciation of S Tibet was found in indicators like glaciated knobs, trough valleys with pronounced flank polishings and limits of glacial scouring on nunataks, as well as in findings of erratics, lateral moraines, end moraines, and terraces of outwash plains. This total glaciation took the form of an ice-stream network and attained a thickness of at least 1200 m. Ice-free to about 87°–86° E, the Tsangpo valley with its sander deposits occupied the gap between the glacier areas of the Tibetan and High Himalayas in the S (I 3) and those of the Transhimalaya in the N (I 2). In the light of recently glaciated Late Glacial terminal moraines and ice marginal ramps it has been possible to estimate a glacio-isostatic uplift of c. 400 m during 10×10^3 years (an average of 40 mm/year) following deglaciation. It is about 3 to 8 times greater than the tectonic uplift of the High Himalaya. The post-glacially intensified uplift of the S Tibetan Plateau by comparison with the High Himalaya is attributed to the much greater glacier burden during the Ice Age.

In the area under investigation a High Glacial ELA depression (equilibrium line altitude depr.) of at least 1200 (1180) m was reconstructed for a mean altitude of about 4700 (4716) m asl. Assuming constant hygric conditions and a gradient of 0.7° C/100 m, the temperature drop at the time would have been 8.4° C. Since precipitation during the Ice Age must, if anything, have been less, a drop in summer temperature of about 10° C may be regarded as probable.

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