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The Pleistocene Glaciation of Tibet and the Onset of Ice Ages – An Autocycle Hypothesis

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ABSTRACT: During seven expeditions new data were obtained on the maximum extent of glaciation in Tibet and the surrounding mountains. Evidence was found of moraines at altitudes as low as 980 m on the S flank of the Himalayas and 2300 m on the N slope of the Tibetan Plateau, in the Qilian Shan. On the N slopes of the Karakoram, Aghil and Kuen Lun moraines occur as far down as 1900 m. In S Tibet radiographic analyses of erratics document former ice thicknesses of at least 1200 m. Glacial polishing and knobs in the Himalayas, Karakoram etc. are proof of glaciers as thick as 1200–2000 m. On the basis of this evidence, a 1100–1600 m lower equilibrium line altitude (ELA) was reconstructed for the Ice Age, which would mean 2.4 million km^2 of ice covering almost all of Tibet, since the ELA was far below the average altitude of Tibet. On Mt. Everest and K2 radiation was measured up to 6650 m, yielding values of 1200–1300 W/m^2. Because of the subtropical latitude and the high altitude solar radiation in Tibet is 4 times greater than the energy intercepted between 60 and 70° N or S. With an area of 2.4 million km^2 and an albedo of 90% the Tibetan ice sheet caused the same heat loss to the earth as a 9.6 million km^2 sized ice sheet at 60–70° N. Because of its proximity to the present-day ELA, Tibet must have undergone large-scale glaciation earlier than other areas. Being subject to intensive radiation, the Tibetan ice must have performed an amplifying function during the onset of the Ice Age. At the maximum stage of the last ice age the cooling effect of the newly formed, about 26 million km^2 sized ice sheets of the higher latitudes was about 3 times that of the Tibetan ice. Nevertheless, without the initial impulse of the Tibetan ice such an extensive glaciation would never have occurred. The end of the Ice Age was triggered by the return to preglacial radiation conditions of the Nordic lowland ice. Whilst the rise of the ELA by several hundred metres can only have reduced the steep marginal outlet glaciers, it diminished the area of the lowland ice considerably.

'Very likely the future will see greater changes in the glacial map of eastern Central Asia than in that of any other part of the world.' Richard Foster Flint (1967, p. 421)

The State of Research up to 1975, together with Investigations Subsequently Carried Out

A synopsis of older results and views on the Pleistocene glacier cover in Tibet has been provided by v. Wissmann's compilation (1959; the author himself has, however, never set foot in the high regions of Asia). It is echoed in the recent Chinese literature in Shi Yafeng et al. (1979), and has also been reproduced by Climap (1981 Map entitled "Last Glacial Maximum"). These authors speak of a from 10% to maximally 20% ice covering of the mountains and plateaux of Tibet. But time and time again, from as long ago as the turn of the century, there have been single researchers like v. Loczy (1893), Dainelli & Marinelli (1928), Norin (1932), De Terra (1932), v. Handel-Mazzetti (1927) and others (cf. Kuhle 1987a), who described ancient ice-margin sites scattered throughout the high regions of Asia. According to the author's calculations, they presented ELA (equilibrium line altitude) depressions of more than 1000 m, and thus indicate locally much more significant glacier formations than the v. Wissmann scheme had acknowledged. However, these authors neither drew nor gave voice to such conclusions. Other early researchers like Tafel (1914), Prinz (1927), Trinkler (1932), Zabirov (1955) (cf. Kuhle 1987a), making more or less direct use of the data they obtained by observation, reconstructed larger glacier areas which, depending on the great altitude of mountains or plateaux respectively, had built up at only a few hundred metres of ELA depression.

The author has been fortunate in being able to carry out nine expeditions and research visits since 1973, some

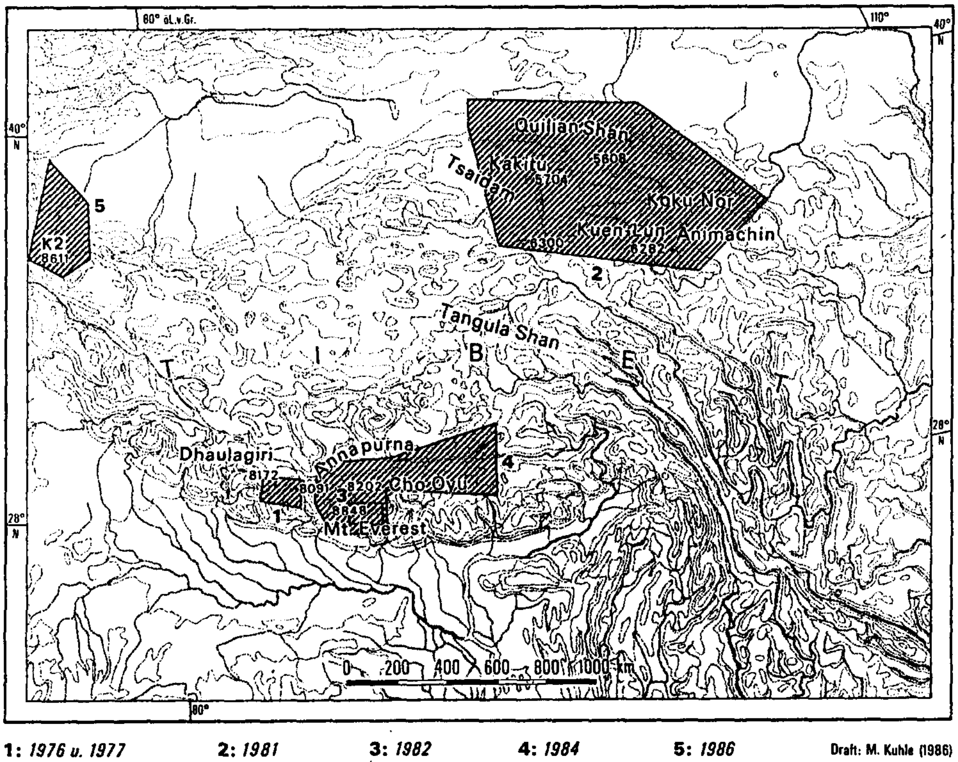


Fig 1
Research areas in High Asia visited in the course of seven expeditions

of which extended to seven months, with the purpose of reconstructing the extent of glaciers in Asia during glacial periods[1]. Two of these were to the arid East Zagros Mountains, the others to Tibet and its flanking mountain systems (Fig 1). By now the location and number of areas under investigation permit representative data on the glacier areas to be made for the whole of Tibet. They are supported by data from some earlier authors, and are in glaring contrast to the negligible ice-cover published by Climap as late as 1981 (see above). Apart from the Tienshan the glaciation of Tibet during the last Ice Age is given as approximately 2.4×10^6 km^2 (Fig 2), and is estimated to include central thicknesses of about 2.7 km (Fig 3 and Fig 4). There was thus inland ice with a central dome of about 7000 m asl in Tibet, the details of which are to be demonstrated below. Breaking up on the edges, it discharged through the surrounding mountains as steep outlet glaciers.

Evidence of a Large-Scale Glacier Cover on the Tibetan Plateau

Three areas covered by boulder-clay and erratics, or by erratics alone, may be adduced:

In S Tibet, at 28° 50' N / 87° 20' E (Fig 1, No. 4), there is the Lulu Valley; it cuts deeply into hydrothermally decomposed basalt (50% pyroxenes, pseudomorphically replaced by dolomite and chlorite). Between 4400 and 4950 m the valley floor is filled with boulder clays, which extend far (170 m) up the sides in some places. In its very fine intermediate material there are some isolated deposits of very coarse 2-mica granite components (Fig 5). Transported over long distances from the N, tens of metres thick, and spread over tens of kilometres, these glacigenic diamiktites should be regarded as ground moraines. A convergence with a mud-flow must be ruled out for sedimentological, petrographic and topographic reasons (Kuhle 1987b).

The second example is also provided from a finding in S Tibet. That is so because throughout Tibet the ELA attained its highest level here during the Ice Age, and still does so now. The evidence of a glacial cover that extends over entire areas thus receives the greatest possible extrapability, particularly to the N towards which the ELA in any case inclines for planetarian reasons. At 29° 41' N / 90° 12' E, N of the Tsangpo Valley, there is a 5300 m high pass known as the Chalamba La. It is, in other words, situated at the point at which the valley network of the Transhimalaya leads out of the Central Plateau of Tibet. Up to at least 200 m above the depression that forms the pass, lying on dark rhyolite bedrock with chlorite but without potassium feldspar, there are superimposed, light-coloured, tectonically-marked granite erratics with potassium feldspar components, but lacking chlorite. On both sides of the Chalamba La steep valleys lead down immediately about 1000 m (to 4300 m asl); according to these moraine finds the valleys must have been filled with glacier ice well above the pass and up to 1200 m at least. This proves the

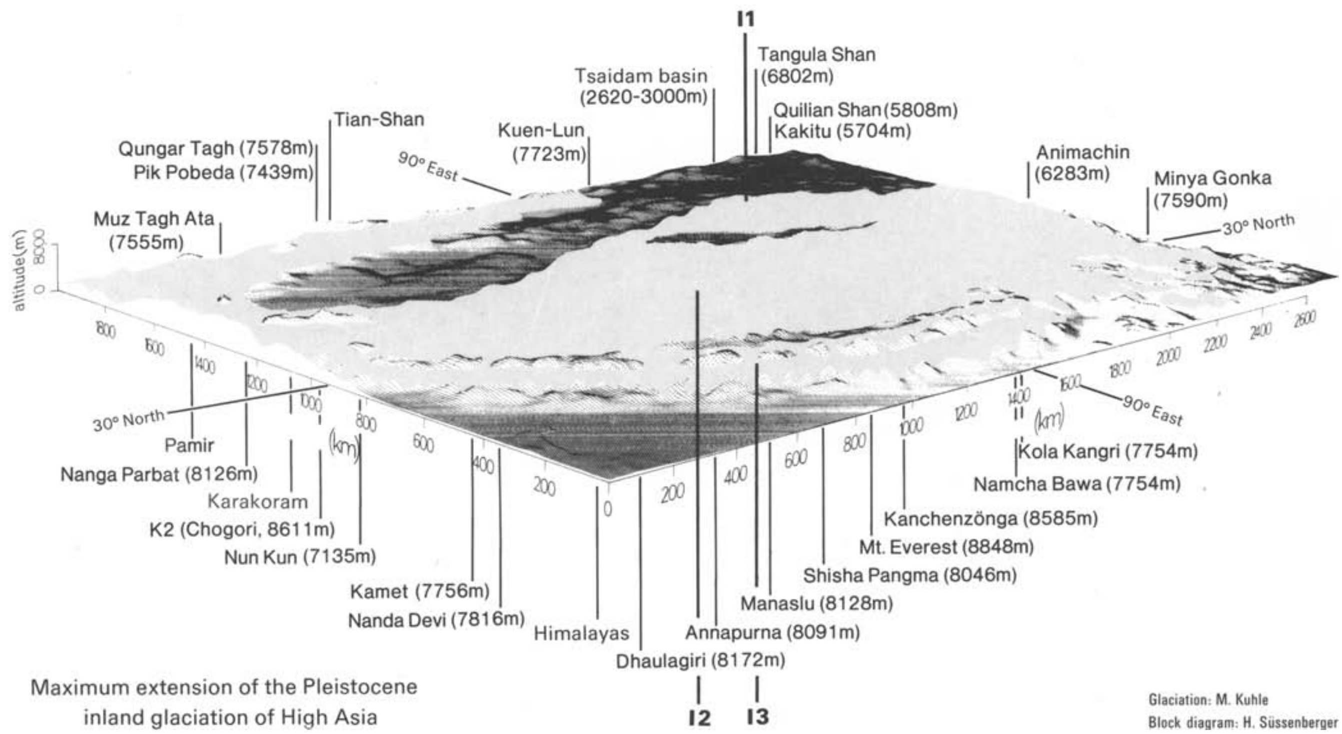


Fig 2 The $2{,}4 \times 10^6$ km^2 continental ice-sheet on the Tibetan highland with its centres I1, I2, I3. Only peaks reaching more than 6000 m to 6500 m project above the glacier surface (exaggeration 15-times)

existence of a ramified network of ice streams, subordinate stream surfaces of which have moved over and across the counter gradients of divides with their saddles. The nearest bedrock granite is known to occur 80–100 km further E, and is part of the Lhasa Pluton (Gansser 1964). The direction of transportation and the glacier direction given by the valley system point to numerous, quasi-parallel outlet glaciers of substantial thickness. This is confirmed by roches moutonnées, which rise to 5600 m together with rough crest-lines and peaks which begin only above the latter. The outlet glaciers have left the central Tibetan inland ice by way of an interposed section of the ice-stream network (Fig 2, I 2) and at this longitude (about 89° E) just missed the

Fig 3 The cross section of Tibet

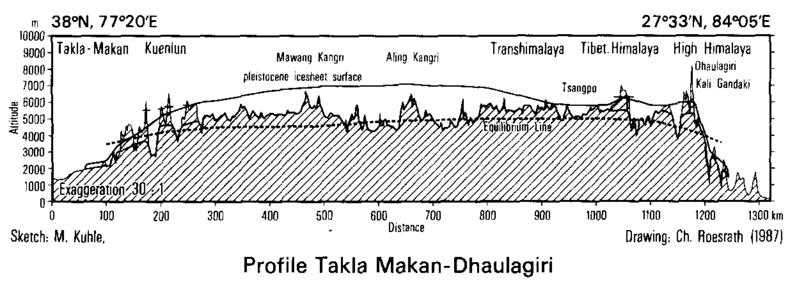


Fig 4 The cross section of Tibet

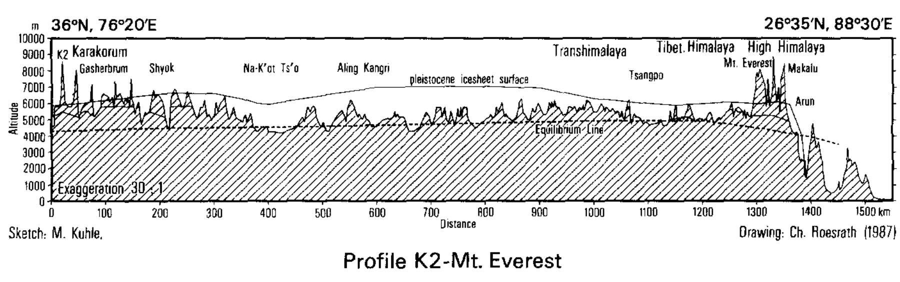


Fig 5 Erratic 2-mica granites on basaltic bedrocks in S Tibet (28° 51' N/87° 21' E, 4900 m asl). The Lulu Valley is set in bedrock basalts. The floor of this valley is covered by ground-moraines containing granite blocks. The largest ones measure up to 4,3×2,6×2,2 m. Photograph by M. Kuhle 31. 8. 1984

Tsangpo isobath at 3800 m (Kuhle 1987b). The final extremity of the outlet glacier concerned came to its end at 3900 m asl inwards of a side valley, a mere 6–10 km from the Tsangpo, exactly like that of the westerly, parallel valley of the Orio Matschu (29° 23' N / 89° 37' E). Evidence of this may be found in the over 120 m high lateral moraines which coalesce in terminal moraine walls.

From these a sander root or outwash cone emerges. The deposits in the side-valley that joins the main valley near Lhasa (29° 43' N / 91° 04' E) must also be seen in this context of moraine findings as evidence of ELA depressions of at least 1075–1200 m in S Tibet. They are situated at 4250 and 3950 m asl.

Another finding of erratics in the Shaksgam Valley on the W edge of Tibet (36° 06' N / 76° 28' E) (Fig 1, No. 5) is evidence of a relief-filling glacier cover rarely presented in such an unmistakable way. This is the most arid part of High Asia, with less than 40 mm of precipitation annually at 4000 m asl. The area in question is a cross-section of the Muztagh Valley in the area of its confluence in the Karakoram N slope to the N of the 8616 m high K2, which has been ground down to a glacial trough up to an altitude of 1200 m above the valley floor (Fig 6). At altitudes between 4400 m and 4700 m gneiss and granite as well as dolomite erratics (90% Do, 5% Ca, micritic and sparitic) have been found on roches moutonnées of 90% pure calcite (Fig 8), standing 600 m above on a transfluence pass that leads from the Shaksgam to the Muztagh Valley. These erratic blocks have been transported over a long distance following the Shaksgam Valley. More than 1.5 m long, they occur singly as well as in the context of bands of lateral moraine material. Gneiss and granite appear on the inner side of the Shaksgam Valley. These blocks required transport along the valley at a high level in order to be deposited here.

Fig 6
Glacier polishing on the right flank of the Shaksgam Valley in dolomite bedrocks (36°09'N/76°36'E, 4100 m asl, Karakoram-N-slope, Aghil-S-slope). It proves a minimum thickness of the Shaksgam Glacier of 1200 m (-----). It was one of the big pleistocene W-Tibetan outlet-glaciers.
Photograph by M. Kuhle 30. 8. 1986

On the outward side of the valley in the W the transfluence pass is bounded by a 4730 m high polished (i.e. to its top) glacial horn – a massive rock of micritic calcite. Up-valley and to the E of the saddle, the glacially-polished calcareous rock flanks of the intermediate divide rise to more than 500 m above the floor of the transfluence pass. Clearly situated above the polished band, its uppermost section is a roughened and earlier crest-line which rose above the high, or even only late, glacial ice stream network like a nunatak. The great thicknesses of ice are confirmed by very small-scale relief and at the same time by softly polished roches moutonnées on the pass. They are only brought about through ground polishing in the subglacial region where the melting point is reached as a result of ice-pressure. Now, as well as during the glacial period, the mean annual temperature at the glacier surface was and is about −10° C at the equilibrium line. At the time of the main Ice Age the corresponding glacier surface of this locality was in the glacier supply area, c. 1000 m above the equilibrium line. This is an indication that the forms of the roches moutonnées transfluence pass are most likely testimony of the late Ice Age. Corresponding grooves can be found on the orographically right-hand side of the Shaksgam Trough and on the 5466 m high 'Shaksgam Horn' 25 km up-valley and up to 1200 m above the rocky floor of this large W Tibetan longitudinal valley (Fig 6). This is the area in which a distributary stream of the Shaksgam Glacier system buried the 4863 m high Aghil Pass (36° 11' N / 76° 36' E) under a 500 m thick cover, thus communicating with the Yarkand ice-stream network system. Glacier polishings in the massive limestones on the orographic left-hand in the Aghil Valley and in the granite on the right are evidence of this. Below 3700 m well preserved striae in the quartzite are encrusted with iron-manganese (Fig 7). Others can be found on sandstone outcrops at about 3600 m, and roches moutonnées with polishings on metamorphosed schist outcrops were found at 3400–3600 m near Illik (36° 23' N / 76° 42' E). That is the trough-shaped confluence area of the upper Yarkand Valley. Besides many other forms of polishing (Kuhle 1987a, c) these exemplary data are evidence of a W Tibetan Karakoram – Aghil – Kuen Lun ice-stream network at the time of the main Ice Age. The ice fillings of the central longitudinal valleys (Shaksgam and Yarkand Valley) (Fig 6) were large outlet glaciers – attaining thicknesses like Alpine glaciers during the Pleistocene – (cf. Fig 3 and 4), which had flowed down from the Central Plateau ice.

In addition to the findings of erratics mentioned above, roches mountonnées fields in Central Tibet are evidence of the ice-cover: in north-central Tibet (Kuhle 1987c) S of the Kuen Lun Pass (35° 33' N / 93° 57' E) there are roches moutonnées in metamorphic sandstones and crystalline schists at 4800–5350 m; in the S Tibetan Latzu Massif (28° 55' N / 87° 20' E) in basalts at 5000–5500 m asl, and 100–120 km to the W, N of the Menlungtse Group at 28° 32' N / 86° 09' –25' E in meta-

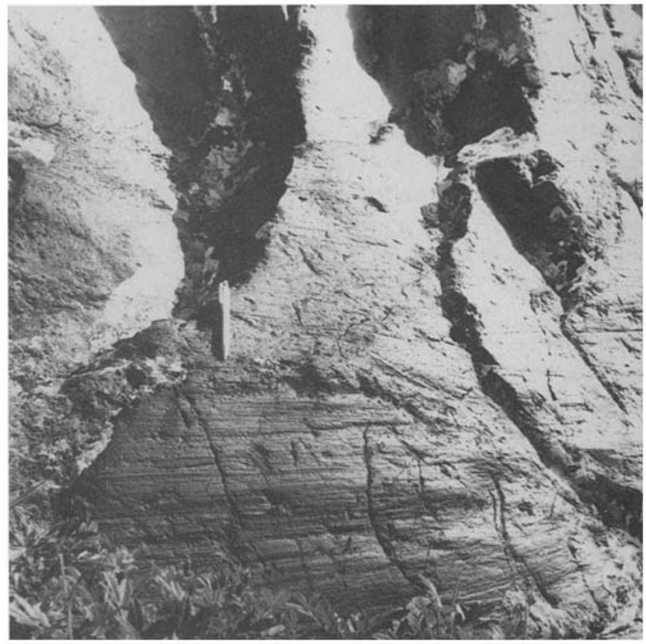
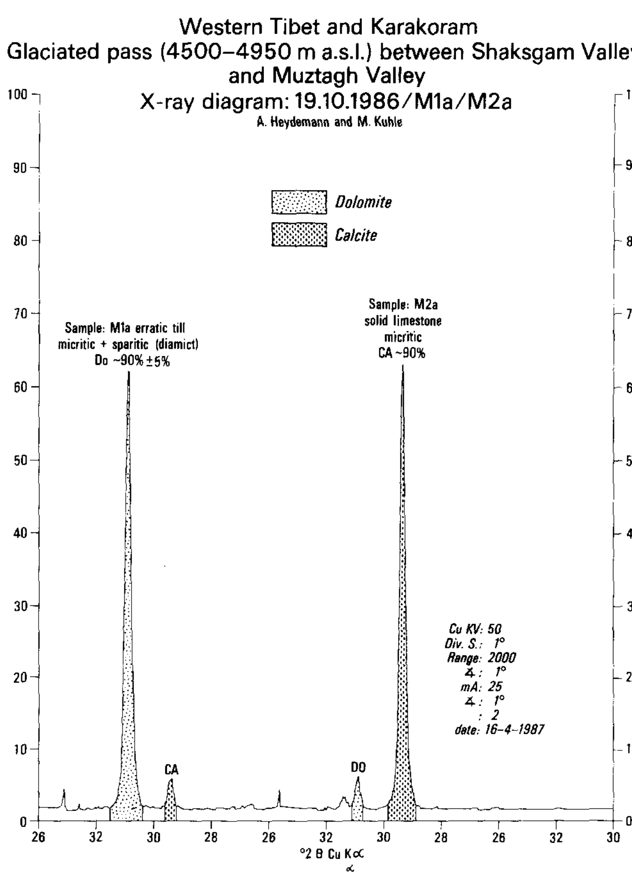


Fig 7 Glacier polishing with striae and polish marks on the right flank of Aghil Valley (36°13'N/76°38'E, 3700 m asl, NW Tibet, Aghil-Kuenlun) in metamorphic bedrocks. Photograph by M. Kuhle 27. 8. 1986

Fig 8 Erratic till and solid rock



morphic sediments between 4400 m and 5100 m, to name another example, or – another 50 km away – roches moutonnées in the Shisha Pangma foreland (28° 37' N / 85° 49' E). Although pebble fillings now frequently give them the appearance of box profiles (Fig 6), the predominantly trough-shaped valleys of N and S Tibet are of an appropriate character to prove the existence of what was once probably a more than 2000 m thick inland ice cover (Fig 3 and 4).

The Lowest Positions of Marginal Ice around the Tibetan Plateau and Depression of the Equilibrium Line

Determined by the position of 46 glaciers, the present climatic line of equilibrium (ELA) on the S edge of Tibet, S and N of the Dhaulagiri and Annapurna Himalaya (Fig 1, No. 1) lies at about 5550 m asl. It was calculated in accordance with v. Höfers method (1879) and that of Louis (1955, which has a mathematical equivalent in the 'maximal method' developed by Kuhle, 1982), as well as in accordance with the Lichtenecker method (1938) (Kuhle 1982, 1986a). Going into detail, the climatic equilibrium line N of the main crest of the Himalaya is at 5620 m, and S of it at 5490 m, i.e. only 130 m lower in the humid monsoon position (6000–2000 mm p.a. at 1600–3000 m) than on the dry N side in the rain shadow (270 mm p.a. at 2700 m asl). Moraines of the *main glacial* period are to be found on the S slope down to 1100 m in the Mayangdi Khola (28° 23' N / 83° 23' E) and the Thak Khola (Kali Gandaki; 28° 24' N/ 83° 36' E) (Kuhle 1979/80, 1982). Using five valley glaciers, two of which were large outlet glaciers flowing simultaneously from the N slope of the Himalaya, the equilibrium line was calculated as being at 4060 m asl. In this

$$ELA_{Depr.} = \frac{tp-ti}{\quad} \text{ (m asl) and}$$

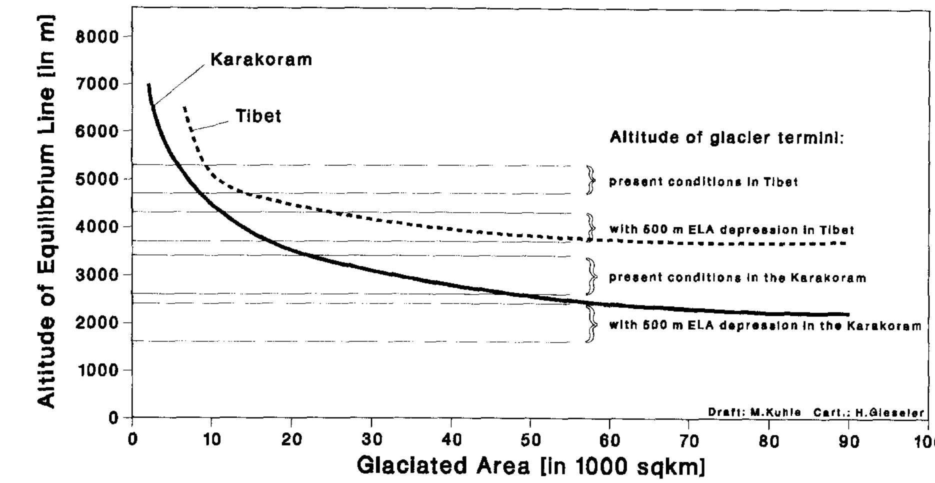
$$Si = Sp - S_{Depr.} \text{ (m asl)}^{2)}.$$

By means of glacial positions of ice margins, still evident in 31 terminal moraines, it was possible to establish the equilibrium line north of the main crest as being at about 3980 m asl. The large number of glacier margins is explained by the very deeply incised former transverse valley of the Thak Khola (Kali Gandaki), together with a catchment area of the S Tibetan Mountains that is only a little higher than 6000 m. The transverse valley cuts through at almost 2000 m asl, and N of the main crest it has only reached the 2700 m line. It follows that here the hanging glaciers, as well as those from longitudinal valleys, having been outlet glaciers of the S Tibetan inland ice and the ice-stream network as well (Fig 2, I 3), though reaching the main valley, no longer coalesced into a single tongue (Kuhle 1982, Fig 8). An integral equilibrium line for this section of the S Tibetan Plateau edge would be 4200 m. This is the equivalent of a depression of the equilibrium line of 1530 m. This calculation even includes the Late-Glacial position of the ice margin of the Jhong Khola Glacier (Muktinath Basin) at 3250 m (near Kingar: 28° 49' N / 83° 50' E; see map in Kuhle 1982, Fig 184 and Fig 38). It is situated in a very dry valley N of the main Himalayan crest, and is evidence of an equilibrium line at about 4450 m, the equivalent of a local equilibrium line depression of only 1210 m (Kuhle 1979/80, 1982). Since then this moraine has been confirmed by Iwata et al. (1982, p. 87). Yamanaka (1982) has dated the minimum age of the corresponding glacier basin as being 8670 ± 200/210 (C 14). This Late Glacial Age and the same equilibrium line depression brings it into line with other moraines like the terminal moraine at Ghasa ('Ghasa Stadium' I, after Kuhle 1979/80, 1982, Fig 91, 92) in the Thak Khola (Kali Gandaki) transverse valley.

The positions of marginal ice and the extent of sometimes large – almost up to 60 km (58 km) long – outlet glaciers flowing down from the S Tibetan ice which have been reconstructed here, are based on terraces of riverside moraines that may rise as high as 570 m and lie up to 1060 m above the valley bottoms, as well as on well-marked terminal moraines with erratics. Moreover, there are trough profiles with glacier striae descending as far as 1700 m to the evergreen forest (with *Quercus semecarpifolia, inter alia)* and grooves as evidence of glacier thicknesses of 1600 m in the central Mayangdi Khola (28° 08' N / 83° 23' E) to give just one example (Kuhle 1979/80, 1982). It must be emphasized that in its glacial drop to 3980 m asl the equilibrium line N of the main crest ran more than 1000 m below the average level of valley floors of the S Tibetan mountains (Tibetan part of the Himalaya) (Kuhle 1982, Fig 184). Even when the Late Glacial equilibrium line ran at an altitude of 4450 m, there was still a difference of 600 m. This had to lead to a glacial filling of the entire Tibetan Himalaya, which overwhelmed the relief – as is evident from the ubiquitous smoothly polished intermediate valley divides – and extended to an ice-stream network and further N to the formation of inland ice (Fig 2, I3).

340 km further W on the N side of Mt. Everest, and again on the dry leeward side, conditions are the same (Fig 1, No. 4). In this area (28°–29° 50' N / 85° 24' –91° 13' E) investigations established a recent macroclimatic equilibrium line at almost 5900 m by means of 15 values that are based on fieldwork. The Chinese map for snow lines and glacier and snow equilibrium lines (Xie Zichu, 1:2 000 000) indicates values between 5500 m and 5900 m for this area, so that in consequence a level of 5700 m is to be assumed. Evidence of the maximum equilibrium line depression of the area puts this at 1180 m (or, if taken in relation to the map, at only 980 m); it was arrived at by reconstruction of eight ice margin positions. The climatic equilibrium line accordingly ran at 4700 m asl (Kuhle 1984/85). It is, however, probably a matter of ice-margin positions which had already receded, i.e. Late Glacial ice margin

Fig 9
Correlation of equilibrium line altitude and glacier area



positions which were the only ones to be found in this area, consisting of vast areas altitudes of 5000 m to 6500 m and never dropped below 4200 m. It is likely that glacial margin positions from the High Glacial period can only be found near the terminal points of outlet glaciers which formerly flowed through the High Himalaya. The Bo-Chu or Sun Kosi glaciers of the High Glacial period (28° 50' N / 86° 09' E) may serve as an example. Located on the Tibetan Plateau, which emerges here from the High Himalaya in 5000-5500 m high, continuous and large remnants, it was supplied by the N, E and SW flank of the Shisha Pangma, as well as by the Menlungtse group. It descended to at least 1600 m (near Kodari: 27° 56' N / 85° 56' E) and is evidence of a climatic equilibrium line at 4300 m, a level that is still 300 m too high by comparison with the Dhaulagiri and Annapurna area (see above). Evidence of the minimal glacier end exists in the form of perfectly preserved polishings on outcrops of the metamorphosed sheets of Kathmandu and Nawakot.

As had been discovered in 1982, the ice-streams of the Cho Oyu-, Everest- and Lhotse S slopes flowed down similarly far and coalesced to form the Dudh Kosi Glacier. The glacier tongue smoothed the gorge between Nangbug and the mouth of the Surke Drangka, and may well have passed by the Lumding Drangka confluence as far down as 1800 m (27° 38' N / 86° 42' E). There is evidence (Fig 1, No. 3) of glacigenic bank formations with disturbed glaciolimnic sands at Namche Bazar, Nyambua Thyang, Chu Chhutawa and Julming (27° 44' – 50' N / 86° 42' E) (cf. Heuberger 1986, p. 30). Just like the Mayangdi, Thak Khola and Bo Chu glaciers, the Dudh Kosi Glacier, too, was a large inland-ice glacier or ice-stream network outlet-glacier, which to some extend was fed by tributaries from the Tibetan ice. It was brought about by the confluence of the Nangpa La (28° 06' N' / 86° 35' E; 5700 m) W of the Cho Oyu, and the later re-routed Rongbuk Glacier, which had couse down from the Lho La (28° 0' N / 86° 53' E; 6010 m) W of Mt. Everest.

There is evidence of overflows from the present N to the Himalaya S side, thanks to attrition from below, supra-glacial weathering, below the W shoulder of Mt. Everest at an altitude of 6500 m (500 m above the transfluence pass; Fig 4). In the N forefield of the present northward-draining Rongbuk Glacier, the gradient of the lateral moraines accordingly tilt so that at an ice-level on 900 m above the valley floor the gradient begins to face S. This 180° re-orientation of the Rongbuk Glacier explains the total absence of older frontal moraines as little as 8 km down valley from the present glacier tongue from the Neo-Glacial ice-marginal deposits at the Rongbuk monastery. They had never been formed. Even during the Late-Glacial period the glacier still drained to the S (Fig 4). These continuing over-spills (Fig 2, I3) into the steep S ramp of the Himalaya also explain the slight thickness of the Ice-Age Rongbuk Glacier. Its highest traceable strip of lateral moraines runs a mere 600 m above the recent glacier surface. A further heightening of the ice became impossible through an overspill into the steep S side that was only 5 km away.

In the S part of the highland as well, with a mean altitude of 4800–5000 m and valley floors of 4200 m at the lowest, the reconstructed equilibrium line depression to at least 4720–4300 m asl (see above) had to lead to relief-filling glaciation. Its level was determined by the topographic proximity to the steep S edge of Tibet alone. The result was an approximately 900–1200 m thick ice-stream network.

How unproblematic or cogent a glacier-filling of this mountain landscape is, whose main valley floors fail to reach the equilibrium line by perhaps 600 m, is shown by a comparison with the Ice Age Alps: the Rhone Valley, for example, having been infilled with 2000 m of ice, finds its valley floor to be 1600 m below the main Ice Age equilibrium line (cf. Fig 9).

In 1987 the investigations on the S fringe of Tibet were extended to the extreme SW edge of the Highland, to Nanga Parbat and to the S slope of the Karakoram (Fig 1, No. 6).

The present almost 60 km long glaciers end here in the catchment area of the Indus Valley at altitudes of at least 2500–2600 m (Hunza-Karakoram). They are thus the ones to descend most deeply in the whole of High Asia. On Nanga Parbat they reach 2900–3600 m asl. During the last Ice Age, however, all the tributary glaciers from the Karakoram S slope and the Nanga Parbat group united in a 12×10^4 to 18×10^4 km^2 ice-stream network – the Indus ice-stream network. At 980 m asl its largest outlet-glacier reached the lowest common ice-marginal position at Sazin at the mouth of the rivers Daret and Tangir, a little above the bend of the Indus (35° 34' N / 73° 28' E). There are two graded series of lateral moraines more than 100 m in height, with a final bend of the terminal moraine. These series of moraines occur at a distance of about 10 km from one another, each flanking the Indus valley over several kilometres. The moraine ramps consist of typical glacial diamictites of polymict composition. Down-valley the valley cross-section becomes a V-shaped valley gorge. Up-valley from the position of the ice-margin a concave, worn-down U-profile takes over and eventually expands to a trough profile with well-preserved striae on the flanks. Sixty km up-valley, at about 1100 m asl, there is a basin-shaped opening – the 'Chilas Chamber'. Here remnants of lateral moraines on the orographic left bank provide evidence of the prehistoric in-filling with ice. Exactly opposite the Chilas settlement glacial polishing on the orographic righthand side provides evidence of a minimum glacier thickness of 450–550 m. Between Chilas and the lowest moraines, the subsidence of the valley glacier edge during the late High Ice Age is retraced by stable lines of light coloured glacio-limnic bank sediments. In some places the limnites were dammed back into the side-valleys.

The next 60 km along the NE slope of Nanga Parbat are characterised by ground moraine covers, which coat the roches moutonnées fields, and by several hundred-metre-deep deposits of platform and lateral moraines. This wealth of moraines may be explained by the immense supply of scree from the steeply descending Nanga Parbat glaciers. For the SE flank of Nanga Parbat (Rupal Valley) which (by way of the Astor Valley) had also connection to the Indus Valley the author was able to provide evidence of an ice-stream network of a glacier thickness of 900–1200 m. Apart from polishing on the flanks, which in many places reaches up hundreds of metres, the Bunji moraines are an important feature of the following valley section and up to the mouth of the Ice Age Gilgit Glacier where it debouches into the Indus Glacier (1228 m asl). Thrust into this side valley of Bunji, this orographic moraine on the left bank corresponds to those in the opening of the Astor Valley. They are 400 m high. Subsequently in 1987 the Gilgit-Hunza part of the glacier stream was reconstructed up to the 58 km long Batura Glacier with the help of diamictites and glacial polishings.

Below Gilgit the tributary glaciers of Batkor and Bagrot joined up. Their Late-Glacial advances are provisionally classified by the author as belonging to the Ghasa Stadium (I) (cf. Kuhle 1986b, c); they reached the Gilgit Valley outside the equally old terminal basin of Gilgit at an altitude of 1300–1380 m. The former is caused by the steepness of the valley, and the latter by the high altitude of the catchment area (Rakaposhi 7788 m). Thanks to glacial polishings (together with striations extending as far down as 1900 m asl at Rakaposhi (36° 15' N / 74° 24' E) the Hunza Glacier component can be reconstructed up to 1600–1800 m above the floor of the valley. It is an exemplary demonstration of glaciation of Pleistocene-Alpine dimensions, with a central-montane thickness of at least 2000 m for this part of the area under investigation. The approaching Shishal Glacier provided a link with the N side of the Karakoram and presented a connection with the Shishal Glacier system (see above). The Pleistocene predecessors of the Hispar-Biafo Glacier produced the S transverse connection with the Muztagh-Karakoram ice. Consisting of acute and wide angles, the ice-stream network was subject to a great deal of friction, and therefore tended to dome-shaped prominence. On the E edge of the Karakoram, approximately at the Pangong Tso (33° 45' N / 79° E) and further N on the Depsang Plateau (35° 25' N / 78° 20' E), it gradually merged with the compact inland ice cover of Central Tibet (Fig 2).

By contrast with the recent glaciers in the Indus Valley catchment, which tend to terminate at about 3400 m, the equilibrium line depression for the lowest prehistoric ice-margin at about 1000 (980 m) is calculated to be 1200 m[3]). Compared with a contemporary equilibrium line at 4600–5000 m, this implies an Ice Age line at 3400–3800 m in the area under investigation. The exponential increase in the area under glaciation, together with the depression of the equilibrium line, is shown in Fig 9; it is established for the first 500 m of the depression of the equilibrium line altitude (ELA).

This implies that here also, on the extremely arid W edge of Tibet, where the mean annual precipitation of stations (Misijiaer, Gilgit, Chilas; 1961–70) in valley floor locations is a mere 142.5 mm p. a., there is evidence of relief-filling glaciation which completely masked it all. Its equilibrium line ran 1200–1600 m below the average altitude of W Tibet.

TL-datings of tills and glaciolacustrine sediments immediately upvalley of the terminal moraine at the village Sazin near the mouth of the rivers Daret and Tangir at 980 m asl (see above) and further up in the basin of Chilas (at 1000–1200 m asl) carried out by Schroder and Saqid Khan (1988) indicate (in the context of the authors findings) that this glaciation existed at least up to 37–55 Ka BP.

During the 1986 expedition to the N slope of the Karakoram and to K2 (Fig 1, No. 5) the author investigated the even more arid part of NW Tibet as far as the N ramp of the Aghil and Kuen Luen ridges and down to the desert-like Tarim Basin (cf. above). The long-term mean annual precipitation at the stations was 67.3 mm p. a., and at 2500 m asl it is 4–5° C colder than on the Karakoram S side. At an altitude of 3090 m the station of Tashikuergan measured 3.1° C during the period 1957–80. The lowest moraines of the last glacial maximum are at 2000–1900 m asl. Tens of kilometres long, these chains of median moraines extend the transverse valleys of the Kuen Lun (37° 20' N / 77° 05'–35' E) S of the Yeh Cheng far out into the mountain foreland, (Fig 3, left). The very extensive and thick moraines are 400–700 m high. Made up of in parts very large, polymictic, facetted blocks with rounded edges which are embedded in a loamy ground mass, they also include in many places stratified glacio-fluvial drift and rythmitic limnites that have been thrust in as well (Kuhle 1987a). 30–40% of the rough blocks are adjusted to the direction of movement. The extreme thickness and the flexures of these diamictites exclude convergent mudflow deposits, as does their granite, phyllite and limestone composition. Their ground-plan morphology is one of elongated ramps, which bend to form frontal moraines around terminal basins. Even exaration striations at the foot of the inner slopes of the moraines have been handed to us. These are forms which have been deposited by the large Tibetan outlet glaciers in the course of the repeated glaciations of the Pleistocene (see below). During inter-glacial periods – as again now – renewed processing of continental frost scree took place in the highland. Outside the recent moraines described above there are more extensive, wide-ranging terminal glacier basins. They have remained untouched by the superimpositions of the last Ice Age and reach another 200 m further down.

The glacier cover of the last glacial period of the area is shown in Fig 3, left. The depression of the equilibrium line during the last glacial period amounted to 1300 m, so that the equilibrium line in the area under investigation (Fig 1, No. 5) ran at about 3900 m asl (3700–4100 m).

Like the area treated above, the more humid NE area (Fig 1, No. 2), too, stands for the N fringe of Tibet (Kuhle 1981). It does not have a key function for the reconstruction of the ice cover to the same degree as on the S fringe (Fig 1, Nos. 1, 3, 4). The altogether highest levels of the equilibrium line (see above) are reached in the S, where they are in most marked contrast to a formerly very large glacier area. Due to present aridity, and thus insufficient likelihood of an extensive Ice Age glaciation, the W edge of the plateau is also more brittle. The 450×820 km area under investigation in the NE, however, completes the Tibetan profiles from S and W. During an expedition in the year 1981, the Ice Age glaciation was reconstructed on the basis of 35 representative positions of ice margins (end moraine complexes) (Kuhle 1987c). In the extreme N, on the edge of the Gobi Desert, as for instance at 39° 50' N / 97° 33' E, as well as at 39° 49' N / 97° 49' E, the glaciers flowed down to 2150 m. This corresponds to a depression of the equilibrium line by 1225 m to 3375 m asl. The equilibrium line depression by 1450 m to 3250 m which also occurred during the last Ice Age (at 39° 23' N / 98° 49' E) presents an extreme value[4].

On the basis of these values the equilibrium line depression from its highest altitudes in S Tibet to its N edge here amounted to 1470 m (4720–3250 m). The S to N distance being 1500 km, this makes for a good accordance with the temperature equivalence, in which 100 m vertical distances is equal to 100 km of horizontal distance.

A mapping of the reconstructed equilibrium line heights in NE Tibet during the last Ice Age (Kuhle 1987c, p. 302) showed one isochion running S of the Tsaidam Depression at 4100 m, i. e. 100–400 m below the mean altitude of the plateau there. This must have led to a self-accumulating formation of inland ice (Fig 2). Polishing limits are evidence of a thickness of at least 500–700 m of ice in the area concerned (Kuhle 1987c). The existence of a total ice cover has been proved by lodgement till covers with locally prominent rough granite blocks; such granite block loams occur for instance in a valley on the N edge of the plateau, extend over the 4500 m high Oh La Pass and rise another 200 m along the slopes, dropping to 3950 m in the S in the plateau area proper (35°25'N/99°25'E) (Kuhle 1987c, p. 256).

North of the Tsaidam Depression, approximately at the latitude of the Kukunor (36°30'–50'N/98°–100°E), the equilibrium line has descended to 3750–3650 m. Towards the Qilian Shan's N slope it fell to 3400 m and further (see above). At a median altitude of this most northerly complex of plateaux and mountains of still more than 4000 m, the equilibrium line depression alone – apart from findings of ground moraines – allows conclusions to be made concerning the most northerly inland ice-complex I1 (Fig 2). In the W and E it was linked to I2 by bridges of glaciated mountain chains. At its centre was the 5704 m high Kakitu mountain range (38°09'N/96°29'E), the forelands of which are covered by a lodgement till measuring tens of metres in depth with five varieties of granite, sandstones and metamorphics. A 150 m deep core boring has been carried out in the foreland of the Kukunor Shan (Qinghai Nanshan; Chaka Basin, 3170 m asl, 36°48'N/99°04'E). It manifests a multitude of alternating deposits of stratified advance scree, ground moraine and limnic sediments of terminal basins, thus rendering likely the passage of 8 to 11 Pleistocene glaciations of the foreland and, in consequence, of the interior of Tibet (Kuhle 1987c, p. 261, Fig 6).

Concerning the dating: the Würm Age date of the last complete glacier cover described above has been established through the interlocking of recent outwash plains with the Ice Age limnites of the Tsaidam

Depression. These sediments were secured in 1981 by boring 10–180 m below the present sediment surface (36°48'N/96°27'E; 2806–2706 m asl) and found to date from 35 120–47 270 years BP (C 14 analyses by M. A. Geyh, Hannover, FR Germany). According to further radio-carbon datings in the Kakitu Massif (38°02'N/96°24'E) (Kuhle 1986b, pp. 456–7: samples taken by J. Hövermann) and in the Animachin (34°43'N/100°12'E) (Kuhle 1987c, p. 300, Fig 12/13), the more wide-ranging end of High to Late Ice Age glacier cover in Tibet had been completed by approximately 9400 to 8600 BP. This is also supported by the date – older than 8670 C14 years BP by Yamanaka (1982) (mentioned above) for the age determination of the moraine located by the author in S Tibet (Jhong Khola, 28°48'N/83°51'E) at 3250 m (Kuhle 1979/80, 1982).

A Holocene, Neo-Glacial glacier advance was established for two stages, in accordance with the 14–16-step stadial scale developed in the Dhaulagiri-Himal (Kuhle 1982): the Nauri Stadium (V) 4165±150 C 14 years ago, and for the Dhaulagiri Stadium (VI) 2050±105 to 2400±140 years ago (Kuhle 1986 b, c). In the Khumbu Himalaya (N of Cho Oyu; 27°52'N/86°42'E), the Nauri Stadium (V), averaged for seven glaciers, reached an equilibrium line depression of 560 m. N of the Dhaulagiri and Annapurna Himalaya (28°43'N/83°45'E) Stadium V was found to have experienced an equilibrium line depression of 570 m on 17 glaciers, whilst descending to not quite 400 m on the S side (28°35'N/83°45'E) (calculated on the basis of eight glaciers; Kuhle 1982). A reconstruction of the equilibrium line depression on the Animachin Massif in NE Tibet (34°48'N/94°33'E) during the Nauri Stadium (V) worked out at only 240 m. This is an indication of the equilibrium line reactions in response to short-term climatic changes being less significant in N Tibet than in S Tibet.

Especially the younger, Late-Glacial positions of the ice margins, but possibly even the Neo-Glacial advances, may provide a conception of the extent of glaciation on the Tibetan Plateau during the Early-Würm ice age. According to the author's hypothesis, it was the cause of the global triggering of the Ice Age proper (see below).

The Overall Picture of Glaciation in Tibet down to the Lowest High Glacial Positions of the Ice Margin

Fig 2 shows the reconstruction of the maximum glaciation in Tibet, with an area of about 2.4×10^6 km^2. In the central part it formed a compact inland ice, the outflows of which descended through the surrounding mountains for tens of kilometres. They terminated at the steep edges of the high plateau.

To be on the safe side, the 50–70 m high, glacio-fluvial gravel terraces E of 84°–85° E, which lie on the valley floor at 3800–3900 m asl, will be classified as High Glacial deposits. They thus indicate a glacier-free Tsangpo section, which separated the glacier complexes I3 and I2 as far W as 84° E. Still further W they merged once again (cf. Fig 3 and 4). According to the reconstruction of an equilibrium line depression as low as 600 m below the average plateau altitude and the resulting self-elevation of the inland ice (Fig 3 and 4), a glacial filling seems likely, even in the case of this more easterly Tsangpo section. This would assign these terraces – like the varve clays in the deepest parallel valleys to the Late Ice Age. Nevertheless this valley section is to be regarded as free of ice until there is firm evidence to the contrary as a result of determining the age of terraces (Fig 2, N of Annapurna to Namcha Bawa). The second ice-free area is the Tsaidam Depression. In the block diagram perspective and exaggeration it is made to seem a merely narrow strip below I1.

Running in a NW direction from Mt. Everest to K2, and from Dhaulagiri to the W Kuen Lun (Fig 3 and 4), the profile shows that the Ice Age ELA had been upvaulted parallel to the present one – in the sense of a modified 'principle of uniformitarianism'. Over S Central Tibet it attained an altitude of fully 4700 m (see above). Nonetheless, an equilibrium line depression of at least 1200 m led to a drop below 83–86% of the plateau surface. The accumulating ice necessarily led to the infilling of the in-set valleys, which account for the remaining 14–16%. This resulted in the formation of ice, which grew higher through positive feedback, attaining an approximate thickness of 2700 m. Glacier thicknesses, ascertained by means of polishings and erratics, reach 1600 m in the Himalaya, and 700–1200 m in Central and N Tibet. In the W Karakoram even thicknesses of 2000 m have been observed (see above). However, these are minimum values, which may well have risen to 2500–3000 m in Central Tibet thanks to the compact ground-plan extending over 1500–3000 km. The high viscosity of cold, continental glacier ice with annual temperatures of around −10° C at ELA-altitude (Kuhle 1987a) is bound to have been conducive to the build-up of ice. An average thickness of c. 1000 m would imply that 2.2×10^6 km^3 of water had been bound up in the ice-sheet of Tibet. This corresponds to a lowering of sea-level by about 5.4 m (calculated on the basis of data provided by Flint 1971).

Fig 9 shows how the glacier area in Tibet relates to an equilibrium line depression of only 500 m. At the same time it permits an estimation of conditions if the equilibrium line drops by 1200 m, and makes plausible the cupola-shaped build-up of the inland ice to a considerable thickness. With the ice held back by mountain barriers, build-up had been assisted by the slow spread of its run-off, and its freezing to the subsurface. At a later stage the pressure-induced melting point was passed. Run-off from the central inland ice-sheet via a zone of ice-stream networks in the highland rims and down to the tongues of the outlet glaciers gradually increased until an equilibrium had been achieved. This was the end of the build-up of ice.

Glacial isostacy in Tibet – an indirect proof: this must result in a glacio-isostatic drop of about 700 m (730 m) (density difference of ice versus material of the earth's mantle (Δ g = 1:3.7 at an ice thickness of 2700 m). Since deglaciation occurred about 9000 years ago (see above) remnants of glacio-isostatic uplift with a falling phase should still be recoverable – as in Central Scandinavia (Mörner 1978) or the Pleistocene glaciation of North America (Andrews 1970) – the latest tectonic map to be published in China (Beijing 1987) does indeed show rates of uplift of more than 10 mm/year for Central Tibet. This value is two to three times higher than what had been established for the much younger High Himalayas[5]), which had actually been uplifted much more rapidly, as the evidence of antecedent valleys demonstrates (Kuhle 1982a). Epirogenetically slow rather than fast, the tectonic uplift of Tibet can accordingly attain more substantial rates of uplift than the Himalayas[6]) *only due to the reduction of glacio-isostatic pressure.* This is further, albeit indirect, evidence for the ice in Tibet. This is the point at which to have recourse to the discussion of the 1932 Swedish triangulation by Norin (1982). This showed that since the Survey of India in 1861 the profile by way of Leh and into SW Tibet had undergone an uplift of 37 m. This implies a rate of uplift of 521 mm/year (37 m/71 years). In the Scandinavian centre of uplift such extreme rates of uplift only took place around 10–8 Ka (Mörner 1978, Fig 1) and decreased to the present vestigial uplift by about 4 Ka. This permits the analgous conclusion that in the region of the Na-K'ot Ts'o (Fig 4), which is represented by that profile of uplifting, the Tibetan ice persisted several thousand years longer than the Scandinavian inland ice.

Measurements Regarding the Radiation Balance in Tibet in the Light of the Energy Balance of the Ice Age

From August until November 1984 and 1986 climatic parameters were measured on Mt. Everest and Shisha Pangma in S Tibet (28°N; Fig 1, No. 4), as well as on K2 in NW Tibet (36°N; Fig 1, No. 5). Eight climatic stations were installed for that purpose at altitudes varying from 3800 m to 6650 m asl. At the same time, portable, hand-operated instruments allowed comparative measurements to be carried out in other places[7]). In this context measurements of radiation and radiation balance on rock or scree, as well as on glaciers, are of interest. Approximately 25000 representative data on global radiation, return-radiation and albedo were obtained. When weather conditions were such that radiation was not impeded by any cloud, the values of incoming radiation were between 1000 and 1300 W/m^2, which is approximately the solar constant at the upper limit of the atmosphere in relation to the corresponding position of the sun at the time (Fig 10 and 11). Theoretical incoming radiation on September 21st as a mean value, is about

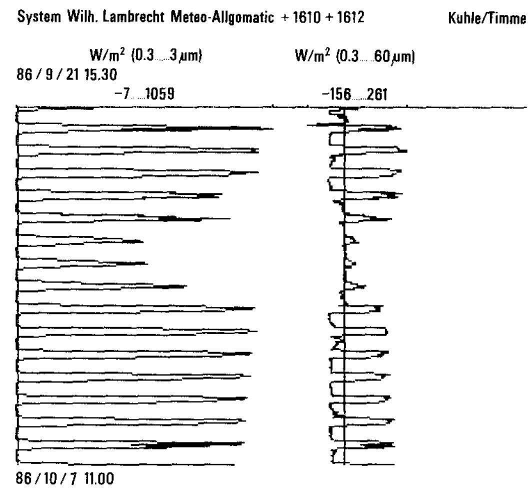


Fig 10 Radiation and radiation balance on K2 (Karakoram 36°03'N/76°32'E) above firn (rough)

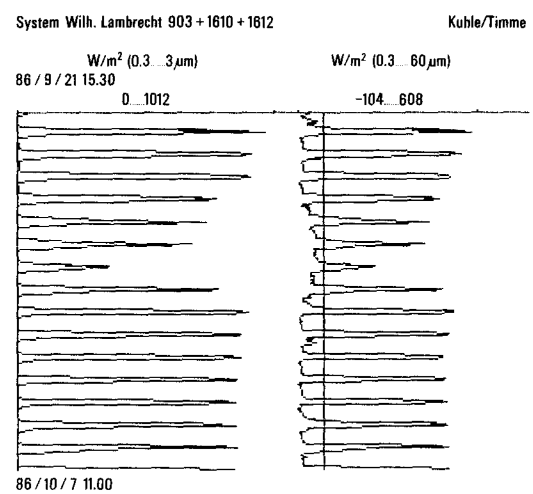
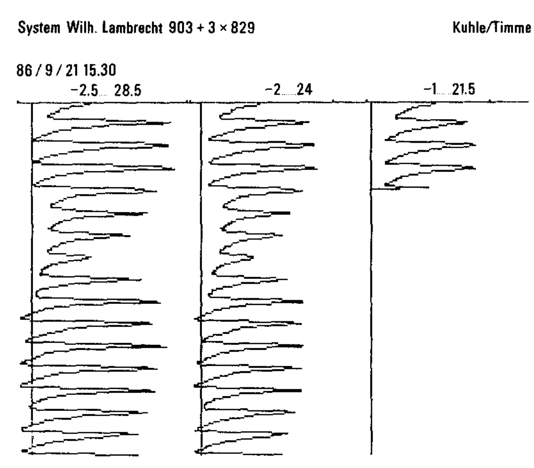


Fig 11 Radiation and radiation balance on K2 (Karakoram 36°06'N/76°32'E) above light scree (rough)

Fig 12 Soil temperature on K2 (Karakoram 36°06'N/76°32'E) light scree (see Fig 10) (depth: 1, 5, 10 cm).



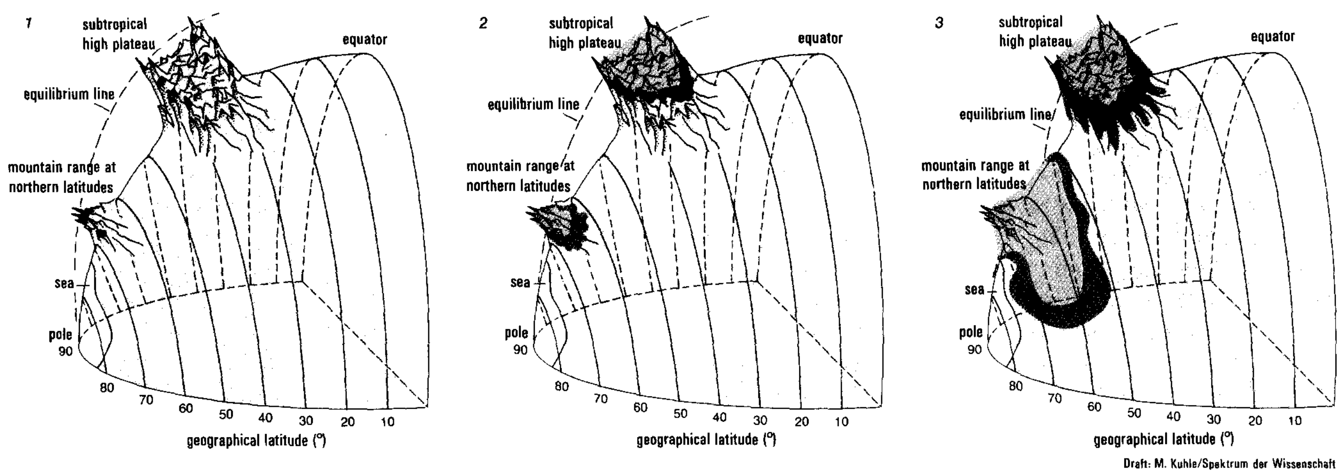


Fig 13 Strongly schematized presentation of the principle of the relief related origin and termination of ice ages. The causative intensification of the cooling-down process emanates from a subtropical high plateau (such as the Tibetan High Plateau) due to the fact that the initial lowering of the equilibrium line by c. 500 m leads to the glaciers descend by 1000 m from the mountains, thus suddenly glaciating large plateau areas (step 1 to 2). Mountain chains at higher latitudes experience the same amount of equilibrium line lowering as a result of the cooling-down by 3,5° C during that change in the parameters of the earth's orbit. But, since the altitudinal distance of the present glaciation to the height of the foreland areas is too great, glaciation has as yet had little effect on area and thus on reflection (2). As the subtropical high plateau has undergone large-scale glaciation and the transformation of a formerly very effective 'heating panel' into an area of reflection, the further cooling down of the atmosphere caused in this way leads to a renewed lowering of the equilibrium line. The consequence is a chain reaction-like, worldwide enlargement of glacier areas. This particularly advanced very fast in all those places where the lowering of the glaciation line reaches the flat mountain forelands (step 2 to 3). The sequence of additions of mountain foreland glaciation depends on the particular altitudinal distance of pre-Ice Age hanging glacier ends from the altitudinal level of the foreland. Although, due to the conditions of radiation, the cooling effect per glacier area is greatest in the sub-tropics, the areal gain of glaciers increases significantly with the decreasing equilibrium line at higher latitudes (3). The reason for this is the fact, that the equilibrium line dips towards the polar regions, and that starting point of equilibrium line heights comes progressively lower towards the lowlands. In the end the ice areas of the high latitudes outnumber those of subtropical high plateaus and mountains by approximately 8:1, by which time their cooling effect has increased around twofold.
Nonetheless, such far-reaching glaciation would not have occurred without the impact of the subtropical inland ice. The cooling, which reacts upon the subtropical plateau ice as well can hardly result in any further increase of the area of ice there because the glaciers cannot reach the lowlands when flowing over the edge of the plateau (step 2 to 3). In a reverse process (3 ⟶ 1) the end of the Ice Age begins in the N and S lowland plains: on the return to normal values of solar radiation and a rise in temperatures by those initial 3,5° C the corresponding rise in the equilibrium line by 500 m and the rise in the glacier ends thus by 1000 m becomes particularly effective for areas of flat lowland glaciation (step 3 to 2). Whilst lowland ice areas experience extreme reductions, thus forcing a global warming-up, the surface areas of the subtropical highland ice will remain almost constant, because only the steeply descending outlet of glacier tongues on the margins will become shorter on the initial upward move of the equilibrium line, whereas the reduction in glaciation is far from reaching the flat plateau ice proper (step 3 to 2). Only when the further warming-up of the earth has been initiated and progressed through the disappearance of lowland ice, will the subtropical highland areas also be freed from ice (step 2 to 1).

1180 W/m^2 at the latitude of the area under investigation (30°N). With an atmosphere approximately transparent to radiation, incoming radiation at Tibetan altitudes produces an energy input at least four times higher than that obtained between 60° and 70° of northerly latitude of the Pleistocene North European inland ice centre (cf. Bernhardt & Philipps 1958). Apart from the much lower angle at which the sun strikes the earth's surface, radiation losses through diffuse reflection of the atmosphere down to sea-level (Lauscher 1956) are about 7%. A radiation loss of this magnitude is in part brought about, and indeed increased, by the low angle of incidence in so far as it prolongs the distance in denser atmospheric layers as compared with a vertical incidence. On the expanses of rock and scree, which now make up 99% of the area of Tibet, albedo values of 15–20% of the global radiation were recorded (Fig 11). The increase in the temperature of the scree (transformation into long wave radiation, Fig 12) is accordingly high. Thus Tibet is now the most effective heat surface of the earth for the hot season. On glacier surfaces, especially on the snow surfaces of the feeding areas, 85–90% of the short wave radiation (0.3–3 μm) is reflected (Fig 10). As incoming radiation increases, so does the reflection caused by the transparence of the high altitude atmosphere. At 6600–7000 m asl, i.e. at the altitude of Tibet's Pleistocene ice dome, the 'greenhouse effect' no longer applies (Fig 3 and 4). About 97% of the highland area was covered by ice, thus transforming the heat surface into a cooling surface with a 70% energy loss of that extremely high subtropical incoming radiation (Figs 10, 11). With an inland ice surface of 2.4×10^6 km^2 it implies a global cooling effect, equal to that of an at least 9.6×10^6 km^2 ($4 \times 2.4 \times 10^6$) Nordic inland ice – i.e. an ice-sheet of more than twice the size of the N European Weichsel Ice (Fig 13).

The Cooling Effect of the Tibetan Ice

In line with the author's theory (Kuhle 1981, 1987b) that the Tibetan Ice must have exerted a considerable global cooling influence, the hypothesis was mathematically examined by Lautenschlager et al. (1987, pp. 8–40) by using the above data of glacier expansion and radiation balance in conjunction with the T21 Model. According to the Hotelling T^2Test of surface temperature, the cooling influence of the Tibetan Ice starts off very high in the test hierarchy. This test hierarchy consists of 10 parts, and assesses all the other globally reconstructed factors, such as ice-sheets on land and sea as well as on ice-free land and sea surfaces, along with their albedo and their role in the system of atmospheric circulation at about 18 Ka. In this test hierarchy the Tibetan Ice starts off with a significance of approximately 10 to the second power above the 99.9% confidence interval. It dominates the remaining nine influences to such an extent that together they were the cause of only the remaining scarcely three powers of ten (i.e. little more than half) of the total cooling down of the main Ice Age period.

A Relief-Specific Ice Age Theory Based on Global Radiation Geometry (Fig 13)

The specific geometry of the earth and the position of the earth's axis in relation to the sun, which has been stabilized by rotation, together result in a radiation balance that provides the sub-tropics with many more times the energy than occurs at high latitudes. It causes the inclination of the equilibrium line from the sub-tropics towards the poles. From this there followed the formation of large expanses of ice in flat or lowland places where the equilibrium line approached sea-level during periods of prehistoric equilibrium line depressions. This was the case in high latitudes unfavoured by radiation. On the other hand, locations on the globe which are favoured by radiation, i.e. the lower latitudes, require even higher elevations the greater their proximity to the equator, in order to permit the formation of glaciers. This led to the mountain and highland ice which was unable to reach the sub-tropically warm plains and lowlands there, thereby terminating with their ablation areas on the steep highland margins.

This relationship of shallow-ended lowland ice in areas not favoured by radiation to steep-edged ice on mountains and highlands in the sub-tropics is the foundation on which the possibility of a relief-specific Ice Age cycle is founded. The comparatively small spatial expanse of ice in the sub-tropics as a function of the necessary high altitude above sea-level, which at the same time is coupled with those locations on steep edges, is compensated for from the same root: a global radiation several times greater than that hitting the ice regions of the lowlands. The pre-condition for the triggering of an ice age is the adequate – i.e. relative to the necessary altitude above sea-level – and considerable size of a mountain highland favoured by sub-tropical radiation, which thus makes it a rare feature in earth history. Such a case, and with it the readiness of the earth for an ice age, occurred with the uplifting of the 2.4×10^6 km^2 area of the Tibetan Highland. Proven findings of moraines from that time, plus the reconstructions of their positions, must permit a later ratification of the kind of orogeny that extended up into the vicinity of the equilibrium line during the Permo-Carboniferous glaciation, for instance.

It is characteristic of the structure of this mechanism of Ice Age auto-cycles that the degree of cooling and thus of global radiation, depends on the extent of the sub-tropical highland glaciation. The fixed proportion of sub-tropical ice to lowland ice also guarantees that the cooling process, which had built up as a result of self-intensification, is to be reversed from the earlier shrinking end of the lowland ice in the manner described above.

The approach of the sub-tropical plateau uplift does not require the principle of the Milankoviĉ Cycle for the triggering of the Ice Age. Continuing uplift by another 500 m, which corresponds to that of the Milankoviĉ cooling by c. 3.5° C, would ensure the glaciation of the Tibetan Plateau even without this extra-terrestrially induced cooling. Admittedly a 500 m uplift, and the consequence of one third of Tibet being under cover of ice, would not be sufficient to trigger the Ice Age. It is likely that an uplift of 900–1000 m is required, so that the amount of global cooling may also be obtained through the Milankoviĉ effect by way of a very much larger initial Tibetan ice. It is, however, indispensable for the re-warming. Only the rising of the equilibrium line by these 500 m can lead to the loss of area occupied by lowland ice necessary for this.

This demonstrates that an extra-terrestrial influence over a short period merely caused the short-cycle to be followed by the Pleistocene cold periods, but not by the Ice Age itself, and it must be stressed that the occurrence of these short cycles of about 10×10^4 years will not come to an end through the loss of global readiness for an ice age. This will be the case when the uplift phase, primarily of the Tibetan Plateau and other mountains and highlands favoured by radiation, has been completed, and regions distant from the equilibrium line have been aggraded.

Footnotes

1) Six expeditions were financed by the German Research Society (DFG), and in part by the Max Planck Society and the Academia Sinica, two by the State of Lower Saxony and the University of Göttingen, and one privately.

2) $ELA_{Depr.}$ = depression of the equilibrium line; tp = recent terminus of the glacier-tongue; ti = prehistoric terminus of the glacier-tongue; Si = prehistoric equilibrium line; Sp = recent equilibrium line; S = equilibrium line = ELA

3) Porter (1970) also found the depression of the equilibrium line in the Swat Kohistan, 100 km farther W, to have been about 1200 m, by contrast with a present ELA of 4200 m at approximately 3000 m asl.

4) An equilibrium line depression of 1430 m (with a maximum value of 1575 m) in NE Tibet was established for an older glaciation (Riss) on the basis of five margins (Kuhle 1987c).

5) Schneider (1957, pp. 468 & 475) puts the uplift of the NW Karakoram at 12000 m since the end of the Late Tertiary, which corresponds to 3–4 mm/y for Tibet and the Himalaya. Gansser (1983, p. 19) gives an integral value of 10–15 mm/y for Tibet and the Himalaya and separately for the Himalaya alone of 4–8 mm/y (oral communication, 1982).

6) As Fig 3 shows, the ice burden decreased in the direction of the outlet glaciers on the rim of the Tibetan Highland as a function of its steep gradient curves. On the one hand this serves to lessen the uplift, thanks to the reduction in pressure in contrast to Central Tibet, and on the other hand mountains like the Himalaya were merely affected by ice filling the valleys. This led to a comparatively linear ice burden, whereas the Central Plateau carried an extensive ice-sheet.

7) The author wishes to thank J.-P. Jacobsen, Diplomgeographer, for help with obtaining data during the 1984 expedition, and H. Diedrich, J.-P. Jacobsen and A. Schulze, Diplomgeographers, for such assistance during the 1986 expedition.

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