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The Upper Limit of Glaciation in the Himalayas*

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Abstract: On the slopes of Himalayan Mountains there is a reduction and culmination of glaciation at 7000–7200 m asl. The presumed cause for this is that the surface temperatures on these slopes are too low for glaciation. This working hypothesis was verified with temperature measurements using collected infra-red radiation. The regression analysis of the measurements taken in the Mt. Everest region during sunny weather conditions of the post-monsoon season resulted in a 0°C line at 7000–7200 m asl. The coincidence of the 0°C line with the upper limit of glaciation is causally definable with the copula between the function of temperature and snow metamorphism: since it is too cold above 7000–7200 m asl, metamorphism into perennial or glacial ice through settling or sintering is absent or simply too slow. High relief and drifting hinder here the processes of ice-formation through pressure compaction of the dry-snow accumulation caused by molecular diffusion and recrystallization. Above 7200 m only continuous leeward accumulations of shifting snow on wall sections with moderate inclination lead to the formation of seracs. However, glaciation generally ceases at this level. This additionally confirms another study. It has been proven that Himalayan glaciers with catchment areas over 7000 m do not extend further downward than those glaciers whose catchment areas just reach this altitude. A break in balance at 7100 m asl is thereby confirmed, and the upper glacial limit is proven. Above the glacial region a rocky zone adjoins with pergelic conditions even in the surface layer. This zone is covered by snow during monsoon season only. Here, the weathering processes take place in an arid environment without thawing and purely by means of temperature variations below 0°C. They could correspond to those occurring on a larger scale on other planets of our solar system.

A lowering of the upper glacial limit by at least 660 or 1200 m respectively, analogous to the Pleistocene snow-line depression reconstructed in S Tibet and the Central Himalayas, is assumed during the Ice Age.

Introduction and Presentation of the Problem

It is generally assumed that the highest stage in the forming of the earth's altitudinal levels is taken up by the glacier region, a fact that seems to be self-evident from the decrease in temperature with altitude. Although it is not uncommon in some places for snow, firn and firn-ice above the snow-line to be interspersed by parent rock, particularly on ridges which are exposed to wind and radiation, as altitude increases above the snow-line there is, in general, evidence of ridges, arêtes and small crags being covered by snow-drifts with an ice-core. Representative examples are offered by the Alps on the ridges of all mountain peaks above 4000 m, and Mt. Blanc, 4808 m high, is completely clad in firn-ice. This observation is confirmed by the examination of other mountain ranges, such as Mt. McKinley, which, at 6193 m, is the highest point of the Alaskan Range. Admittedly, both these peaks – Mt. Blanc as well as the S summit of Mt. McKinley – are but moderately steep, as there are gradients of 32°–46°. The summits of the spurs and ridges below, however, which are pointed by successive steep walls, are only loosely covered by fresh snow. According to these observations, the intensity of firnification and glaciation of high mountains above the snow-line or of level 365 in exclusively favourable climatic conditions depends on *wind exposition* and the *steepness of relief.* As the second climatic parameter, precipitation is the

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necessary prerequisite apart from temperatures below freezing point. This is met everywhere. Though some mountain peaks are situated above the zone of maximum precipitation, none is above the precipitation line, as shown by the seasonal snow cover[1]).

An investigation of mountain ranges like the E Kuen Lun (Animachin), and especially of the tropical Andes, shows that the glaciation of flanks and walls – almost always in the form of riffle-firn, rill-firn or serrated rill-firn– increased towards the top irrespective of the incline of the wall, or at least remains constant. In many cases, not even a section of rock is anywhere kept free. In the Central and Eastern Himalayas, the ice-clad sections of steep walls occur predominantly at altitudes between 6000 and 7000 m asl (Fig 4 and 5). Glaciation of steep faces of this kind confirms the influence of the *adhesive power* on the firn and firn-ice cover in opposition to the power of gravitation. Observations on glaciation at greater altitudes, which are now only possible in the Himalayas, reveal a decrease of the firn-ice cover of the highest summits. The appearance of bare rock on the surface (Fig 1) is noticeably visible from a greater distance. The author was able to collect observations of this kind in the course of four expeditions to the Himalayas[2]): on the N and W flanks of Dhaulagiri I (8172 m), on the S flank of Dhaulagiri V (7617 m), less well defined on the peaks of Dhaulagiri II (7751 m) and Annapurna I (8091 m), clearly on Annapurna II's summit (7936 m), less so on Gyachung Kang (7975 m), faintly on Cho Oyu (8202 m), very markedly on the walls of Makalu (8481 m), Lhotse (8501 m, Fig 6), and on the walls of Mt. Everest (8848 m, Fig 1, 2, 3) with the exception of its E flank. The fact that on all these peaks the firn-ice peters out between 7000 and 7200 m is important in this regard. This shows up as a dividing line between dark and light, which cuts through the flanks of the mountains, often as if drawn with a ruler; at the S face of Lhotse-Nuptse, for instance, it is 7.5 km long (Fig 6 ----). The surface texture of the different parent rocks has no effect on the discontinuation of glaciation on flanks and faces as such, or on the altitude at which it occurs. The samples taken for examination contained limestones (Dhaulagiri I, II and V), crystalline slates and quartzites of varying thickness of strata, as well as migmatites (Lhotse-Nuptse, Fig 6; Annapurna I and II; Everest, Fig 3; Gyachung Kang, Cho Oyu), and even granite (Makalu; Baruntse, Fig 5). The structural conditions do not play a role either: the slopes of outcrops with strata exposures with an upward slant (Mt. Everest, SW Wall, Fig 2; Gyachung Kang, S Wall, Lhotse-Nuptse, S Wall, Fig 6) show an upper glaciation line at the same altitude as diagonally-dipping exposures of strata (Annapurna I, W Wall, Dhaulagiri I, W Wall, and the flanks of Dhaulagiri V), and strata falling away from the face, including slopes of structural surfaces (Mt. Everest,

Fig 1 N main ridge of the High Himalayas, with Mt. Everest (8848 m, 27° 59'N/86° 55'E) as seen from Panga La (5200 m) in the N. Rising to heights of 6800–7200 m (maximum 7550 m), the mountains are covered by glaciers up to the top. On the summit of Mt. Everest, which exceeds the others by 1300 m, dark rock appears on the surface as early as 1 1/2 months after the monsoon. Reaching down to 5500 m, the jet stream blows away the cover of monsoon snow (exposure by M. Kuhle; 5. Nov. 1984).

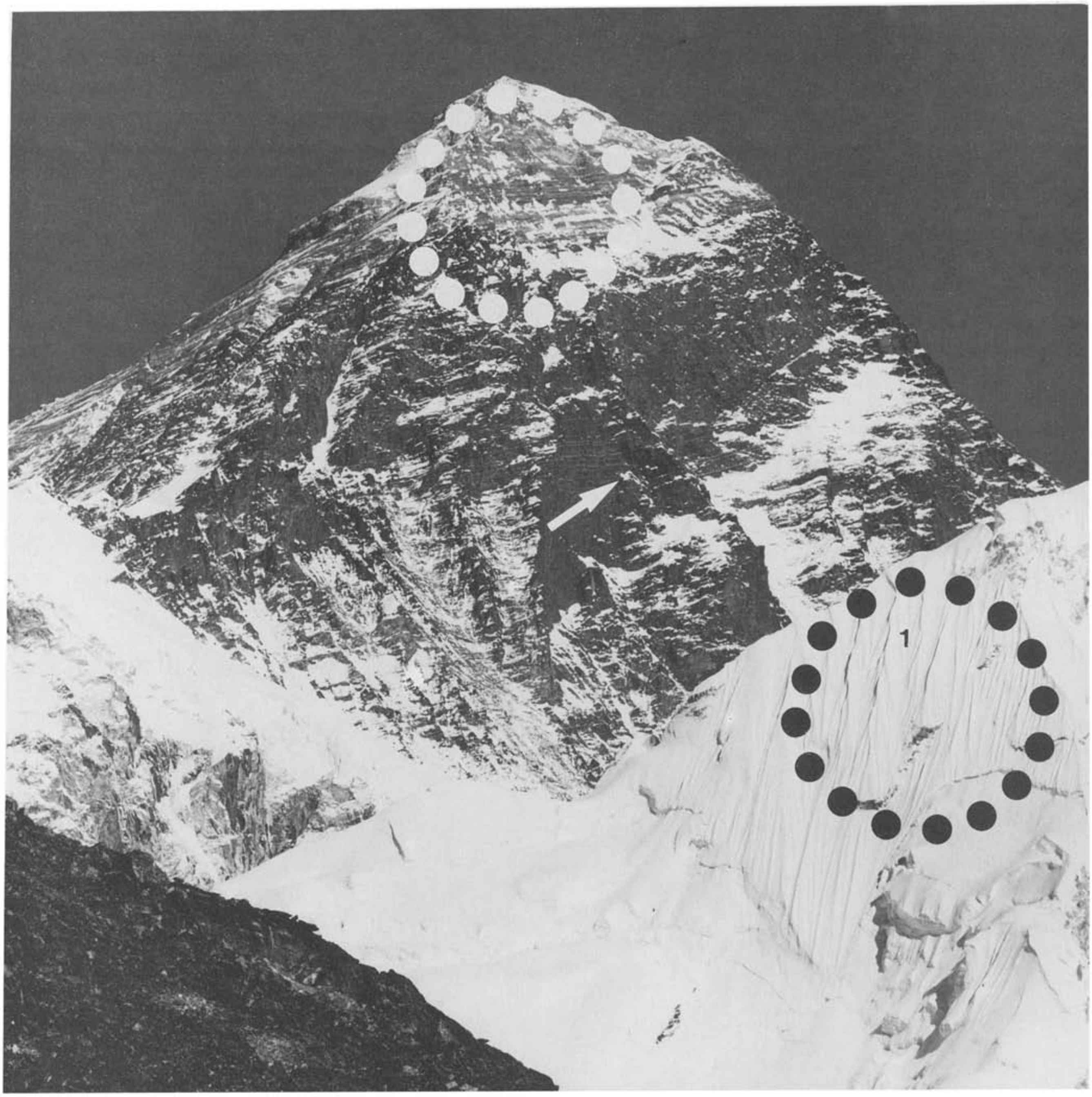


Fig 2 Mt. Everest (Chomolungma), photographed from the W, from an altitude of 5600 m asl. In the foreground on the right, steeply covered with firn- and glacier-ice, the Nuptse NW ridge, merging with the equally ice-covered SSW slope of Mt. Everest's WNW shoulder. Beyond 7000–7200 m the flanks of the outcrop curvatures of the Everest summit pyramid arise without glaciers and with mere patches of monsoon snow. In October, 1982 (11–21. Oct. 82) temperatures from about −7° to 0°C were recorded during full sunshine on measuring fields, with snow between 6530 and 6680 m (dotted circle 1), and −28° to −46°C on those with rock between 8350 and 8848 m dotted circle 2 (exposure by M. Kuhle, 19. Oct. 1982).

NW Wall, Fig 3; Dhaulagiri I, N Wall; the flanks of Annapurna II), and even walls of roughly structured bands of unstratified rock (Makalu).

These uppermost sections of wall above altitudes of 7000–7200 m cannot, of course, always be observed as rock surfaces free of snow. Especially during the monsoon season the peaks in question carry a covering of fresh snow that may well be excess of one metre-deep (remnants of which may be seen in Fig 3). In the course of winter – in the time from October to April, but already tending to occur one or two months after the onset of the jet-stream, i. e. by the middle of November or December – the snow is

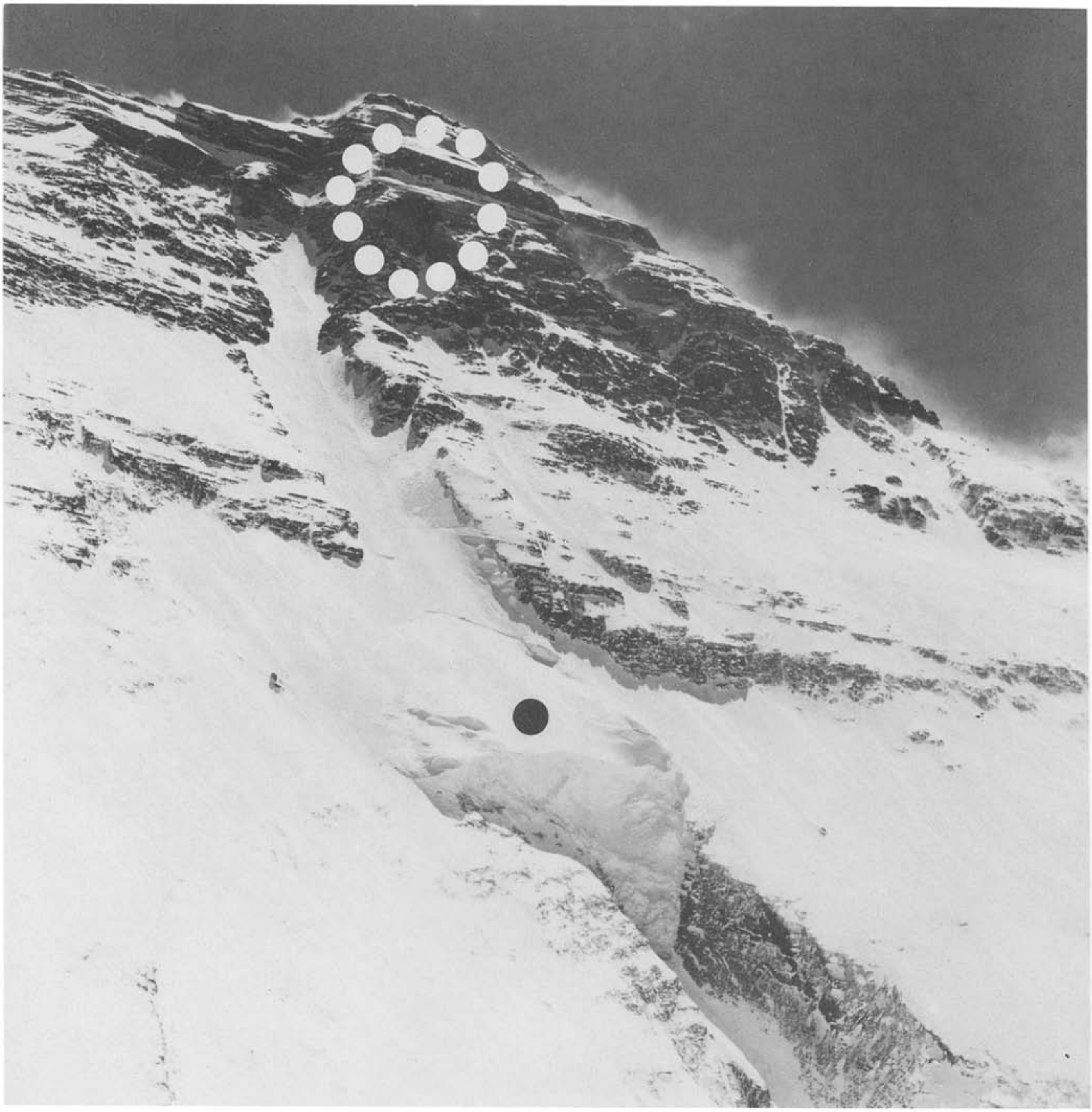


Fig 3 NW wall of Mt. Everest, taken from the SE ridge of Changtse, at an altitude of 7100 m. Notwithstanding its moderate inclination of about 39°C, the flank is not covered by firn-ice, but merely by seasonal monsoon snow. Glacier complexes only form in places with topographically-conditioned and locationally-fixed lee situations. This explains the formation of the ice cornice in the shelter of the Norton Gorge between 7250 and 7660 m (●). In the measurement field on the summit (dotted circle) temperatures from −28° to −36°C were measured in 1984 in radiation weather, though at different times of the day (exposure by M. Kuhle, 21. Oct. 1984 at 14.00 hours).

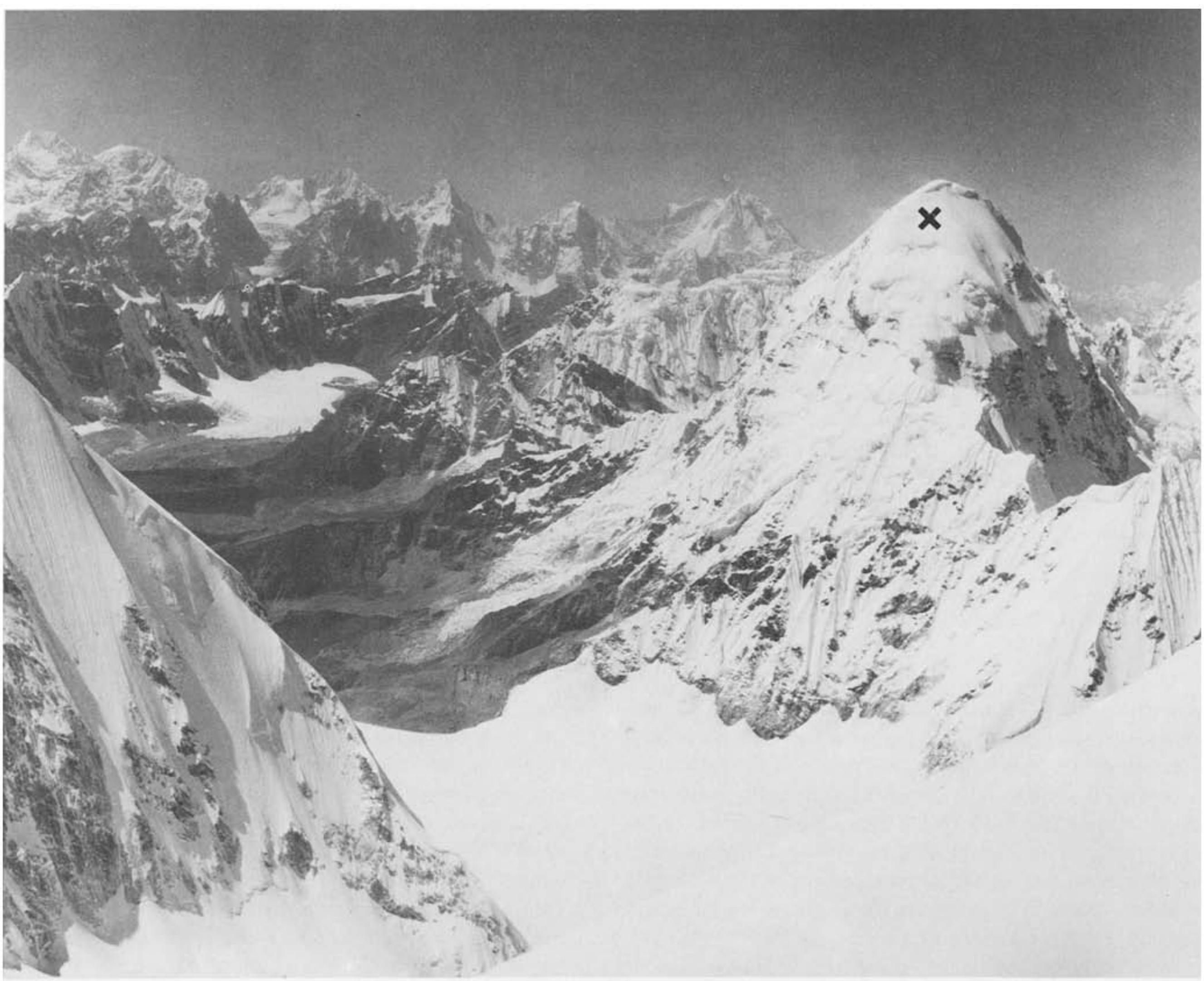
blown off, unless it has previously gone down as an avalanche of solid snow (snow-board), frequently being of the type of a ground avalanche. On Mount Everest the process of snow drifting is visible from afar through the pennant that hangs over towards the E (Fig 1). This process explains the striking armour of snow, firn and ice on the lee-side, the Kangchung or E flank of the mountain.

The locally very clear picture of an immediately visible upper limit to the altitudinal level of glaciers (Fig 6) is disturbed by exceptions. Even at altitudes above 7000 to

7200 m asl the bodies of hanging glaciers may be found in the form of ice-balconies (Fig 3●). Those on the lee-side of the E flank of Mt. Everest have already been attributed to the snowdrift. In general a stable, i.e. a permanent local leeward situation is to be regarded as the best base for the building up of such hanging glaciers (see "Conclusions"). Besides the effect of lee-side whirlwinds which takes snow on to mountain flanks away from winds, topographically favourable factors, such as flat sections and gorges within the walls, determine the locations of ice formations occurring above 7200 m (Fig 3●).

How little of such conditions is required for the formation of glaciers below 7000–7200 m is proved by the strikingly steep and pointed peaks which reach these very heights and are exposed to very high wind velocities when in the wind channel of a valley (jet effect). In spite of this they carry caps of firn-ice which afford a complete cover. Examples are the Pumo Ri (7145 m, Fig 4x) and Baruntse (7220 m, Fig 5). According to observations carried out by Reiter and Heuberger (1960) and by the author himself, the jet-stream, coming from the W, sets in towards the end of September to the middle of October. It affects the relief

Fig 4 Showing the steep, coniform ice peak of the Pumo Ri, which is 7145 m high; the photograph is taken from the SE ridge of Changtse from an altitude of 7100 m, facing W (265°) along the N Wall of Everest below the W shoulder and across the Lho La Glacier pass (6006 m) in the centre of the picture. In spite of its steep flanks and its position in the wind channel of the upper Khumbu Valley, the central source branch of the Imja Drangka, the summit is wrapped in thick firn-ice (x). The same applies to the steep basal section of the N Wall (lefthand side of the picture; for contrast cf. Fig 3). On the higher reaches of these mountain sections surface temperatures were recorded during the post-monsoon times of 1982 and 1984 that almost reached freezing point. The thermal conditions that cause this monsoon snow to settle and sinter together in turn lead to its adhesion to the ranges of steep walls and its gradual metamorphosis to firn-ice (exposure by M. Kuhle 21. Oct. 1984 at 14,00 hours).



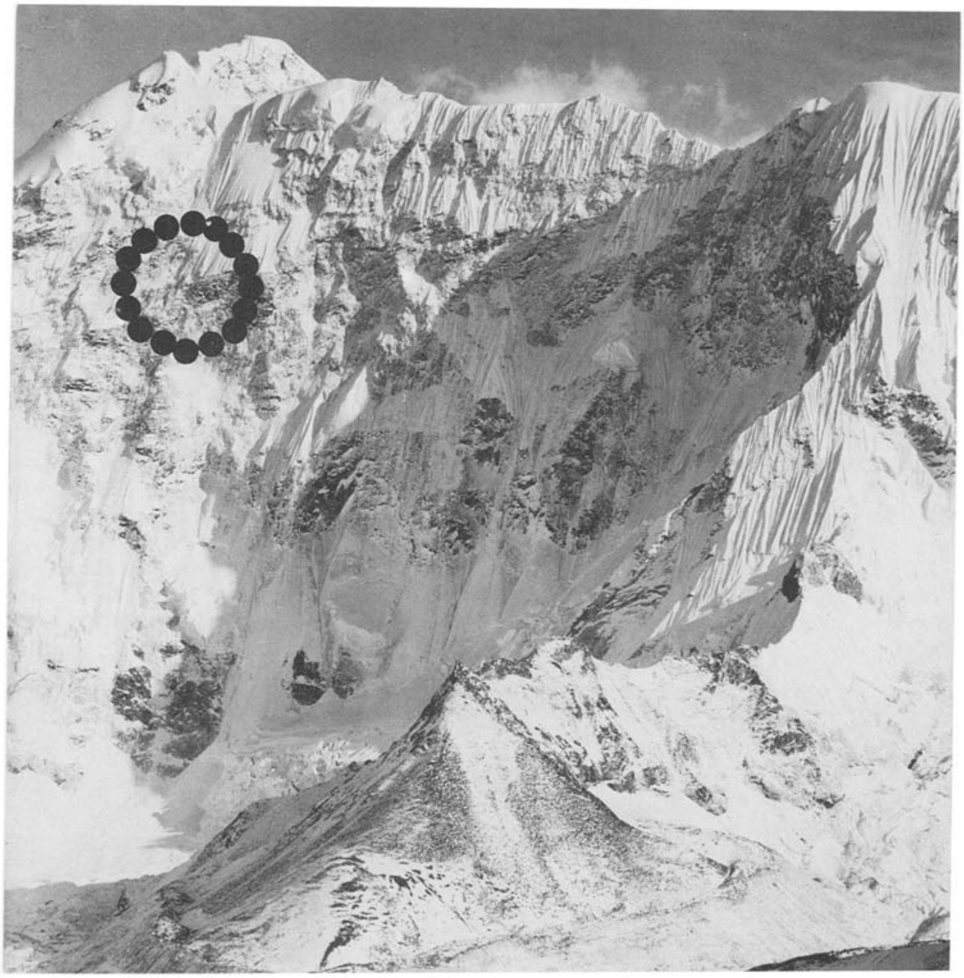


Fig 5 Taken at an altitude of 5500 m asl, facing ESE (110°) towards the NW wall of Baruntse (Mahalangur Himal, 27° 53'N/86° 59'E), which is 7220 m high. Notwithstanding its unusual steepness (up to 65°) and the smooth structure of massive, crystalline rock (Makalu granite, tourmaline granite) this 1700 m high scarp is heavily covered with firn and ice cornices from the glacier at the foot of the wall up to the wind-exposed ridge. In the section selected for measuring, at the altitude of 6450–6650 m (dotted circle), temperatures ranging from 0 to −10°C were registered in the period from October 28th–30th, 1982 (exposure by M. Kuhle, 3. Nov. 1982).

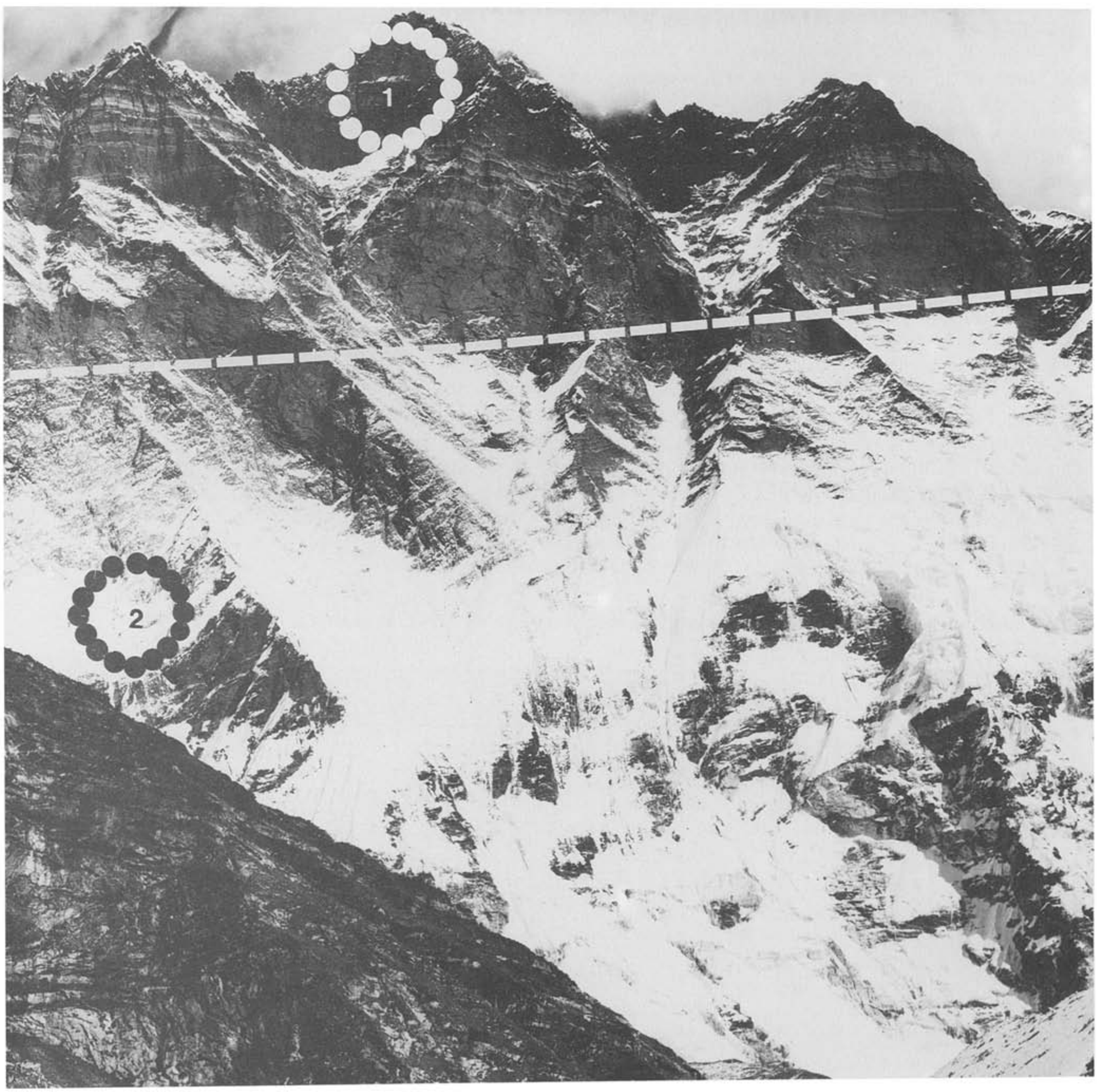
of the Himalayas down to altitudes of about 5500–5000 m asl, so that there is no cause for assuming most extreme wind velocities from only 7200 m upwards, which would, moreover, have to compensate the decreasing atmospheric pressure in respect of transport energy.

The question arising from observations of this kind is the one for *the causal nexus underlying the upper limit of the altitudinal level of glaciers.*

Method of Investigation and Measurement Technique

One supposition and working hypothesis was that it could be the dehydration of the snow as a function of the *low temperatures*, which reduces the adhesion of the snow components, allows it to be blown off more readily, and is the cause of the upper limit to glaciation. Measurements of the air temperature at these altitudes seemed to be too indirect, and thus hardly suited for an investigation. It was important to collect the actual *surface temperatures of the mountain flanks*. For this the technique of telemetric infra-red measurements was chosen. It had previously been applied by Kessler (1971, 1974), and by Jäkel & Dronia (1976) in the desert, on lower altitudinal levels of mountains, and in towns for the measuring of the temperatures of rocks respectively the surface of buildings; Dronia (1978; 1979) had also used it in the context of other problems. The technique is explained in detail chiefly by Lorenz (1973) and to a lesser degree by the other authors mentioned earlier. Few points are therefore in need of particularizing. The measurements were carried out with three different instruments which differ in respect of the focussing of received infra-red radiation, as well as in the precision of their readability. Control measurements with a second instrument kept measurement errors down to a minimum; the same applies to specific repeat measurements of the same section of a mountain flank with the same instrument. At a horizontal distance of 1500 m the type R 380 RVC instrument (Raynger Firm/USA) measures a circular section of a steep face with a diameter of 100 m integral. The Thermopoint 80 (AGA Firm/FR Germany) measures the same diameter from a distance of 3000 m, and the Raynger II HR instrument achieves the same from a distance of 6000 m. The spectral sensitivity lies between 8 and 14 μm. The range of measurement extends as far as −50°C, though the R380 RVC instrument achieves this only after re-adjustment involving loss of accuracy. Both Thermopoint 80 and Raynger II HR are controlled by micro-processors and equipped with a telescopic sight for the exact alignment of sight in the sections that are to be measured. As to precision of measurement, there is agreement within ± 1 % of the measured value. Relative exactness, the separation potential of the devices, lies between 0.25 and 0.1 °C. The instruments work with identical technique in so far as they focus the infra-red radiation of the effective range via a parabolic reflector on a thermopile with an electronic chopper. In the case of the R 380 RVC measurement reading is relayed to an indicator on a scale, whereas the more recent appliances are equipped with a digital indicator.

Fig 6 The Lhotse S wall taken at 5300 m. The visible section of the wall is 3.2 km long and culminates in a main peak at 8501 m (27° 57'N/ 86° 57'E, Mahalangur Himal). Glaciation of the wall ceases above 7100 m (above ---). Whilst surface temperatures in the measurement section at 6000–6125 m asl dotted circle 2 were around 0 to –4°C in radiation weather at noon (12,00–13,00 hours) during the period 24. Oct.– 4. Nov. 1982, temperatures of –16° to –38°C were recorded at the same time of day at altitudes of 8200 to 8500 m (dotted circle 1) (exposure by M. Kuhle, 1. Nov. 1982).



A constituent factor in the quality of evidence of telemetric temperature measurements is the transparency of the atmosphere for infra-red radiation. The filter effect of the air varies with the atmospheric humidity. The specific humidity of the atmosphere falls with increasing altitude asl and with decreasing temperature. In 1984 in order to take this accordingly into account, the author carried out measurements of atmospheric humidity at altitudes between 5000 to 6500 m asl on Mt. Everest in atmospheric conditions of relevant radiation[3)].

To draw a parallel, it can be stated that the lowest values for humidity recorded by A. Kessler in the Sahara (oral communication, January 1985, which is gratefully acknowledged), amounted to 0.5 %, in relation to temperature and altitude asl, this corresponds to a still higher specific humidity than in a relative humidity of 1 % at 6500 m asl. Flohn (1954) reported a daily mean atmospheric humidity of 28 % from the Nanga Parbat area (W Himalayas) from an altitude of 4000 m asl in conditions of atmospheric radiation during July, which decreased to about 15 % at the time of substantial deterioration at noon. At the same time there was excellent long-range visibility above the haze-line – another indication of the great lack of water vapour. On his Cho Oyu expediton (W Khumbu Himalaya, further Mt. Everest Group) H. Heuberger recorded 22–33 % relative humidity, together with a specific humidity (g H_2O/kg air) = 0.4–0.7 mbar in terms of vapour pressure at 14,00 on October 16th and 17th, 1954 at an altitude of 6650 m asl and – also in October – 7–30 % relative humidity at an altitude of 5900 m asl (quoted with kind permission from a letter dated 23.8.1983). These observations are evidence enough that in conditions of perfect radiation the specific humidity is so small as to be negligible.

Using a 300 m -thick air stratum near sea-level (1–300 m asl), Lorenz (1973) showed that the error of measurement of an infra-red thermometer, which is a concomitant of atmospheric humidity, amounts to a positive deviation of the recorded temperatures from the actual temperatures in the range of negative temperatures, whereas in case of positive temperatures the actual values are lying somewhat higher. In the case of the author's investigations of ice and rock flanks, this implies that the recorded negative temperatures correspond to slightly lower actual surface temperatures. A value measured as –30.0°C, for instance, may accordingly be about –30.3°C. Considering the extremely low specific atmospheric humidity at altitudes from 6000 to 8848 m asl (see above), errors in excess of about 0.1–0.5°C are hardly likely to occur in the author's recordings. Possible errors thus remain within the tolerance of the instruments' precision.

The permeability of the atmosphere is proved, moreover, by measurements of global radiation carried out by the author on Shisha Pangma and Mt. Everest from September to November, 1984 at altitudes from 5000 to 6650 m. In conditions of radiation, values from 1000 to 1200 W/m^2 were registered between 13,00 and 14,00[4)], which is equivalent to the extraterrestrial radiation energy of this season.

Exemplary Measurements in the Mt. Everest Region

At the time of least dust in the atmosphere and most intensive solar radiation during the post-monsoon season from early September to the middle of November 1982, exemplary measurements of surface temperatures were carried out at altitudes between 4000 and 8848 m asl. 611 measurement values have been arranged in groups of two diurnal periods from 06,00 to 11,00 and from 11,00 to 15,00. The average time of measurement was about 11,50 local time. The area under investigation is situated in the centre of an

Tab 1 Exemplary humidity values under conditions of atmospheric radiation on the N slope of Mt. Everest

Date	time	altitude	rel. humidity in %			air temperature in °C			air pressure in mbar	actual vapor pressure in mbar			specific humidity			absolute humidity g/m³			
		m	2cm	5cm	200cm	2cm	5cm	200cm		2cm	5cm	200cm	2cm	5cm	200cm	2cm	5cm	200cm	
22.9.84	14.00	5160		14			10		528		1.63			0.00192			1.25		Morainic surface: coarse clasts and gravel; sheltered from wind
24.9.84	14.00	5500	10		7	18		7	505	1.89		0.68	0.00233		0.00084	1.41		0.53	above turf in a depression
29.9.84	7.50	5600	9		8	14		6	497	1.34		0.72	0.00168		0.00090	1.01		0.56	above surface moraine, 10 cm thick, between ice pyramids
29.9.84	21.00	5800	16		8	12		4	484	2.10		0.84	0.00270		0.00082	1.60		0.50	above surface morain, 3 m thick, 20 m from firn fields
8.10.84	21.00	5480	26		27	15		7	507	4.11		2.60	0.00505		0.00319	3.10		2.01	above morainic material, not underlain by ice
10.10.84	14.00	6040	15		20	12		4	470	1.98		1.59	0.00262		0.00211	1.51		1.25	above surface morain, 30 m from ice pyramids
18.10.84	8.00	6500	8		2	-11		-8	440	0.23		0.07	0.0033		0.00001	0.19		0.06	above surface morain, 5 m from firn fields
18.10.84	14.00	6500	4		1	12		1	440	0.53		0.07	0.00075		0.00001	0.40		0.06	- same -
19.10.84	7.55	6500	13		7	-13		-10	440	0.32		0.21	0.00045		0.00030	0.27		0.17	- same -
19.10.84	13.45	6500	4		6	28		8	440	1.33		0.62	0.00188		0.00088	0.96		0.48	- same -
20.10.84	8.30	6500	15		8	-15		-14	440	0.31		0.18	0.00044		0.00025	0.28		0.15	- same -
20.10.84	14.30	5160		13			10		528		1.51			0.00178			1.16		morainic surface: coarse clasts and gravel; sheltered from wind
23.10.84	8.20	6500	12		7	-15		-12	440	0.25		0.18	0.00035		0.00025	0.21		0.15	above surface moraine, 5 m from firn fields
23.10.84	14.15	6500	15		13	2		-6	440	1.05		0.53	0.00149		0.00075	0.83		0.43	- same -
30.10.84	14.30	5160		12			7		528		1.16			0.00137			0.90		morainic surface: coarse clasts and gravel; sheltered by wind
31.10.84	14.30	5160		12			7		526		1.16			0.00137			0.90		- same -

intermediate time zone, so that the times given here agree – but for a few minutes – exactly with the astronomic situation. For the problem under investigation, which attempts the recording of the highest temperatures in specific sections of the wall, values registered between 11,00 and 15,00 are significant. Measurements were carried out in full sunshine, as well as during slightly diffuse radiation on the wall sections (Fig 2, 3, 4, 6), although partial shading by rocky promontories could not be totally avoided (see arrow in Fig 2). The 453 surface temperatures that were registered between 11,00 and 15,00 were arranged in simple linear correlation (Pearson's product moment correlation), with the altitude asl (Fig 7). In the selection of the sample values, the difference arising from the substratum – i.e. whether the section of the wall selected for measuring consists of ice or firn ice, or just of ice with a covering of snow or predominantly of rock – is left out of consideration. There are (in Fig 7) 379 samples of predominantly rocky surfaces. The reason for this is the fact that 83% of the area to be measured was situated above 7000–7200 m, as well as more or less below the snow-line. The rock most frequently met is unstratified crystalline granite and gneiss, and crystalline slate. The mean altitude for measurement was 5867 m. The prevailing expositions were on S to SE aspects. There are few exposures to the W. The reason for this is that only those mountain flanks were measured that receive radiation, and that condensation cloud shades the afternoon sun. The two groups of slopes or surface inclinations of over 40° and below are represented equally frequently. With a correlation coefficient R = –0.835, Fig 7 shows a temperature gradient (y = a + bx) of –1.45°C/100 m, with a probability of error of less than 1%. The mean 0°C limit runs at 5959 m asl (Fig 7 –). Since the measurements that were taken into account had been made within only one year, the predicted accuracy of 68.25% needs to be determined by the standard of error estimation (SEE). This is accepted as ± 10.33 (Fig 7 ---), and shows that, in accordance with the probability mentioned above, all further values from about 6680–6700 m asl no longer rise above temperatures of 0°C. There is a 95% probability that surface temperatures from altitudes of approximately 7400 m remain below freezing-point, as indicated by the double SEE (Fig 7 -.-.-.). The regression analysis thus proves a 0°C line at about 7000–7200 m asl. Above this there commences an almost always frozen pergelide altitudinal level (Fig 8).

Fig 8 represents those 379 samples which have been registered solely on rock wall surfaces. It shows the dependence of the target quantity of surface temperature (y) on the basic quantity of sea-level (x) – largely undisturbed by the influence of the heat loss through melting energy, which in case of the ice and snow surfaces could not be telemetrically considered by way of a correlation coefficient of –0.844, with a probability error of less than 1‰. The rise of –1.51°C/100 m is greater than in Fig 7 – a fact that is attributable to the absence of the melting process. The mean 0°C line is at 5926 m. By means of the single and double standard estimation error of the predictions (SEE) of ± 10.75 or ± 21.5 the freezing-point line at about 6670-7340 m, i.e. bounded at 7000–7200 m, can be arrived at with a probability of between 68.25 and 95%.

Conclusions

In the flanks and walls of Himalayan high peaks the 0°C line of the surface temperature runs at that altitude above sea-level in which the cover of glacier ice, firn ice and firn does not continue any further upwards. The investigation thus results in a coincidence which can be formulated as a *causal effect; as* conditions prevailing above 7000–7200 m are almost exclusively pergelide ones, the glacial level comes to an end at this altitude and has reached its upper line, which implies that it is too cold for the formation of glaciers. In spite of full radiation from the sun, the melting point is not only never reached, but surface temperatures decrease to –20°C and even to –46°C on the summit pyramid of Mt. Everest from 7200–8848 m. In 1982 measurements taken of the uppermost section of the SW wall of the mountain (Fig 2 dotted circle 2) yielded a maximum of –20°C (Fig 7, 8) whereas on the N Wall of Mt. Everest temperatures as low as –28°C were measured during the autumn of 1984 (Fig 3 dotted circle). With temperatures tending to be even lower up there at times of lower positions of the sun, in the shade and during the night, the normal process of the metamorphosis of freshly fallen snow to adhering firn and firn-ice does not take place. According to Benson (1961) and Müller (1962), snow covers subject to temperatures of less than –10°C or an annual mean temperature below –25°C experience a *very slow* settling and sintering, i.e. the formation of ice-bridges between the grains. This is solely a process of sublimation and molecular diffusion, and consequently requires a long time to convert snow into ice of a density of, for instance, 0.83 g/cm³. This necessary *permanence of snow cover* is absent on these high mountains, due to gradients and exposure to wind. The cold and therefore dry snow on them is consequently blown off within a single winter, i.e. the dry period between monsoonal precipitation that falls as snow here. The elucidation of conditions below and above the 7000–7200 m line in the Himalayas can benefit from the observations of the Seward Glacier in the Yukon carried out by Sharp (1951), as well as those by Langway (1968) made on a bore-core from NW Greenland at 77°N. Sharp found that the concentration of snow into ice of 0.83 g/cm³ took place within a mere 3–5 years as a result of high temperatures near freezing point in the zone of wet-snow, which caused it to settle and sinter. This applies to the range in or near the percolation zone. The stratum thus condensed to 0.83 g/cm³ was a mere 13 m below the glacier surface. The bore-core described by Langway had been lifted from the

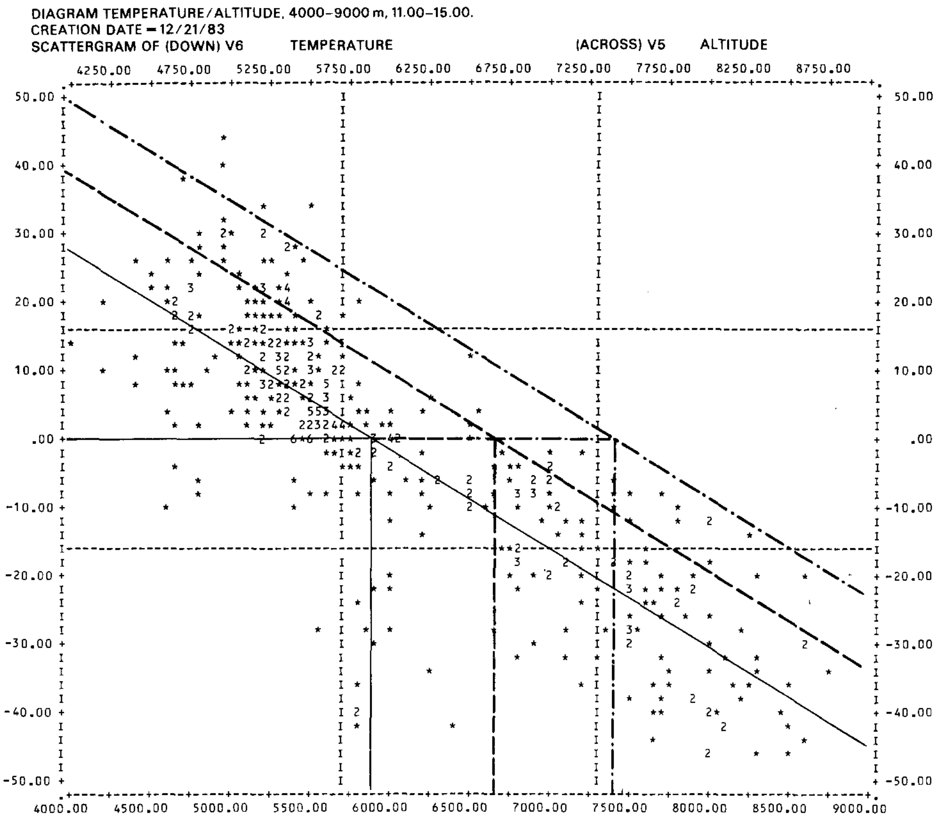


Fig 7 Infrared meassurements: Himalayas/M. Kuhle 1983

0-DEGREE LIMIT: 5959 m a.s.l. TEMRERATURE GRADIENT: 1.45°C/100m

INFRARED MEASUREMENTS: HIMALAYA/M. Kuhle 1983 12/21/83 PAGE 9
DIAGRAM TEMPERATURE/ALTITUDE, 4000-9000 m, 11.00-15.00.
STATISTICS

CORRELATION (R)-	-.83474	R SQUARED -	.69679	SIGNIFICANCE -	.00000
STD ERR OF EST -	10.33038	INTERCEPT (A) -	85.81641	SLOPE (B) -	-.01454
PLOTTED VALUES -	453	EXCLUDED VALUES-	6	MISSING VALUES -	0

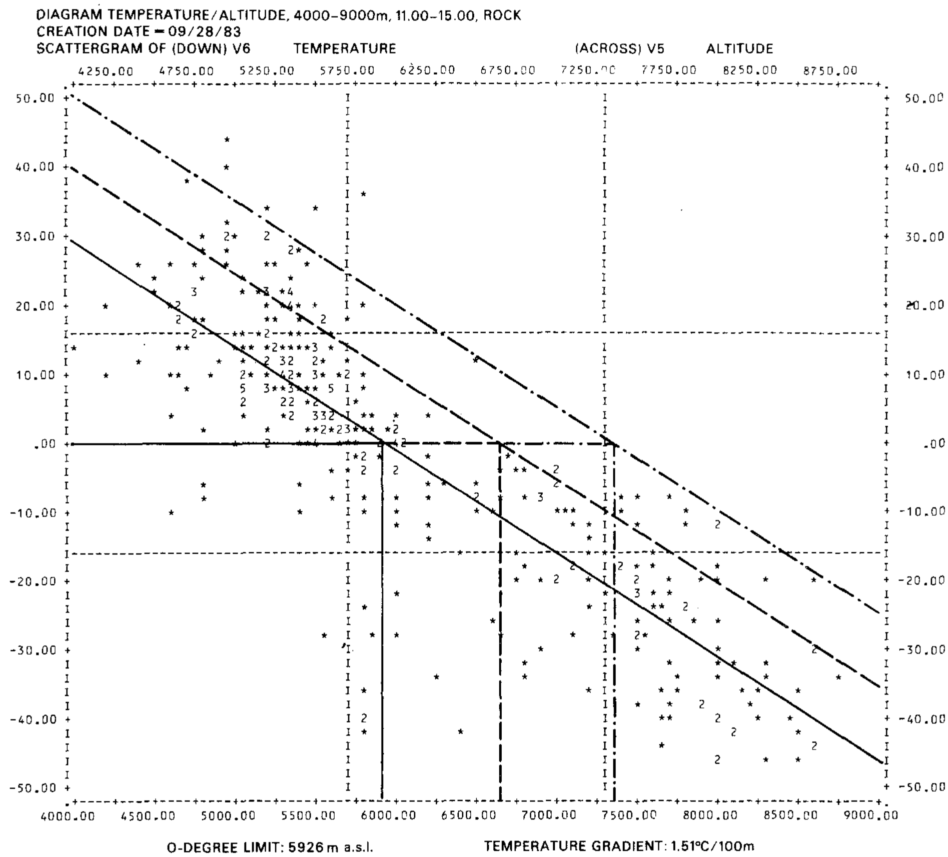
'********' IS PRINTED IF A COEFFICIENT CANNOT BE COMPUTED.

dry-snow line and shows the same density in substratum at a depth of 80 m. Due entirely to the pressure of hanging snow and firn, this compaction and recrystallization (Shumskii 1964) required a period of more than 100 years, and thus at least 20 to 33 times the time needed in high temperatures.

The formation of thick strata, which are the condition for the formation of ice in spite of very low temperatures, is therefore to be attributed to the topographical favourableness of the flat inland ice surfaces of Greenland and Antarctica. This is evidently not affected by the deflation either, which merely causes drifting of the snow but does not blow it away altogether. The absence of ice cover on mountains in Antarctica and Greenland which are exposed to wind has been pointed out by Shumskii (1964, p. 410).

In places above 7000–7200 m asl where deposits of freshly fallen snow are blown off within months down to the parent rock there is not the slightest possibility of periods of more than 100 years, which are required for achieving this density, being available (Fig 3, 1, 2). In other words, the conformity with natural laws indicates that: *at the upper glacier limit in the relief of extremely high mountains the 'blow-off' factor achieves dominance over the 'time' factor in conjunction with compaction as determined by the thickness of snow. At the same time compaction is proportionally related to temperatures varying according to altitude asl and to its accelerating influence on metamorphosis. In this way both blow-off as a cause of the absence of compaction, and the prevention of temperature metamorphosis are attributable to the same cause: the low temperatures above 7000–7200 m.*

Confirmation of these inter-relationships is provided by the formation of separate ice-balconies and hanging glaciers far above the upper glacier limit proper mentioned earlier. Ice of this kind hangs between 7250 and 7660 m asl on the NW Wall of Mt. Everest (Fig 3 •). Corresponding ice balconies may, for instance, be observed in the SW Wall of Cho Oyu (W Everest Group, 8202 m, 28° 05'N/86° 39'E at about 7350 m asl, or at 8300 m in the ENE wall of K2 (Karakoram, 8611 m, 35° 53'N/76° 31'E). Their strata formation indicates ages of about 100 years and more. The firn ice accumulation in the NW Wall of Mt. Everest enjoys the wind protection of the Norton Gorge, which runs down the rockflank from the very top (Fig 3, above •). This is a stable, Fig 8 Infrared measurements: Himalayas/M. Kuhle 1982



INFRARED MEASUREMENTS: HIMALAYAS/M. Kuhle 1982 09/28/83 PAGE 3
DIAGRAM TEMPERATURE/ALTITUDE, 4000–9000 m, 11.00–15.00, ROCK
STATISTICS

CORRELATION (R)-	-.84365	R SQUARED -	.71174	SIGNIFICANCE -	.00000
STD ERR OF EST -	10.75093	INTERCEPT (A) -	89.72289	SLOPE (B) -	-.01514
PLOTTED VALUES -	379	EXCLUDED VALUES-	1	MISSING VALUES -	0

'********' IS PRINTED IF A COEFFICIENT CANNOT BE COMPUTED.

i.e. locationally determined, lee situation with permanent drift-snow surplus, together with a sufficient flattening of the wall. This approximately punctiform favourableness of site enables sufficient pressure to be created by the piling-up of snow, and thus the slow metamorphosis described above to take place. The thickness of such ice balconies governs flow velocity and resistance to strain by breaking-off overhanging ice.

Another approach to the same result: a control test

If the upper glacier line, which has been recorded by the observation of its snow and ice deposits, as well as investigated in regard to its causes by measurements of its surface temperatures, does exist at altitudes of 7000–7200 m in the high mountains of the Himalayas, the *knick* in the *balance* of glaciers of still higher catchment areas, which is to be expected above this altitude, would alter the normal proportion of accumulation area to ablation area. A calculation of the snow-line of a glacier on the basis of v. Höfer's method (1879) or related methods of investigation of the arithmetic relation between mean altitude of catchment areas and the lowest location of the glacier terminus, ought to show a strikingly *large difference from the real snow-line* in poorly-fed glaciers above 7000 m (calculated on the basis of the Lichtenecker method, 1938). The tongue of a glacier that is nourished from an area which extends substantially beyond 7000 m ought therefore not to reach down further than that of an ice-flow from a catchment area extending just up to an altitude of 7000 m. If it were otherwise, it would amount to proof that contributions to the nourishment of glaciers are coming from altitudes in excess of 7000–7200 m.

In another place the author (Kuhle 1986) has shown that the deviation of the mathematical snow-line (S (m)) from the real snow-line, which is determined in the field (S (r)) is functionally dependent on the differences of angle of the average gradients of glacier surfaces above and below the mathematical snow-line (∢ Diff. = $\alpha - \delta$). In order to put the problems which have been raised to the test, by way of comparing glaciers of the same climatic zone with altitudes of catchment areas above and below 7000 m asl, the marginal quantity "difference of angle" (∢ Diff. ($\alpha - \delta$)) needs to be taken into consideration, in so far as it has an influence on the factor of the difference in snow-lines (FSD).

Fig 9 shows the necessary operation, taking as an example the Khumbu Glacier which, after flowing off the S flank of Mt. Everest commands a catchment area that extends far above 7000 m. (1) First of all the basic value is determined by calculating the arithmetic mean between the altitude of the highest peak within the catchment area and that of the glacier terminus (the method used by Partsch 1882; Louis 1955). (2) This is followed by the calculation of the mean altitude of those peaks of the catchment area which lie above the basic value, and by putting them into arithmetic relation to the location of the lowest ice-edge, in order to calculate the mathematical snow-line (S(m))[5]. (3) The average angle of inclination of the ablation area (δ) is assessed throughout its length, from the end of the tongue up to the mathematical snow-line (S(m)), together with the mean angle of inclination of the accumulation area (α) from the mathematical snow-line to the average height of the peaks. (∢ α − ∢ δ) is the difference of angles (∢ Diff.). (4) The real snow-line (S(r)) has been established in the field on the basis of those middle and upper moraine blocks which were found furthest up the glacier. (5) (S(m) − S(r)) is the difference between the mathematical and the real snow-line; on the Khumbu Glacier — to quote just one example — it is 782 m. In order to be able to attribute snow-line difference solely to the difference in angles, the snow-line difference needs to be expressed as a percentage of the vertical distance of the glaciers. (6) The factor of the snow-line difference

$\mathrm{FSD} = \dfrac{(\mathrm{S(m)} - \mathrm{S(r)})\ 100}{\phi \text{ vertical distance}}$ is now to be co-ordinated to the angle distance ($\alpha - \delta$) (Fig 10). Investigations were carried out on 22 glaciers of the Khumbu Himalaya (Mt. Everest Group), 9 of which had catchment areas in altitudes above 7000 m, the remaining ones had theirs below 7000 m.

The result was a significant grouping around two mean value lines of approximately equal gradients, though separated from one another by 14 units of the y-axis (Fig 10). The larger factor of the snow-line difference, the mean value of which exceeded the other one by 14, applies to glaciers with catchment areas above 7000 m. The conclusion is: in cases of equal angle differences, the distance of the mathematical snow-line from the real snow-line increases, because the lowest position of the glacier terminus is entered into the mathematical snow-line as *disproportional too high* in relation to the altitude of catchment areas above 7000 m. This disproportion tends to be bigger, the further the middle catchment area rises above 7000 m (Fig 10). In this way a *knick in the nourishment of the glacier is proved*, which occurs *at about 7000 m asl*. It implies that the catchment area can be called a feeding area only up to approximately this altitude, since it marks the end of the glacier region. The dry-snow precipitation above 7000–7200 m asl is consequently blown out of the topographically -related catchment areas of glaciers. It is deposited in adjacent firn basins and on areas of valley bottoms below the snow-line which are covered with moraines and rubble, thus being — at least in part — lost to the formation of ice.

Confirmation of the 0°C line

through the highest occurrence of lichens on Makalu (E section of the Mt. Everest area) has been obtained by a Yugoslav mountaineering expedition which discovered the highest crust lichens of the genus Lecidea at an altitude of 7100 m (oral communication by J. Poelt, Graz, for which my thanks). According to Kappen and Lange (1970) lichens, though capable of assimilation even in sub-zero temperatures,

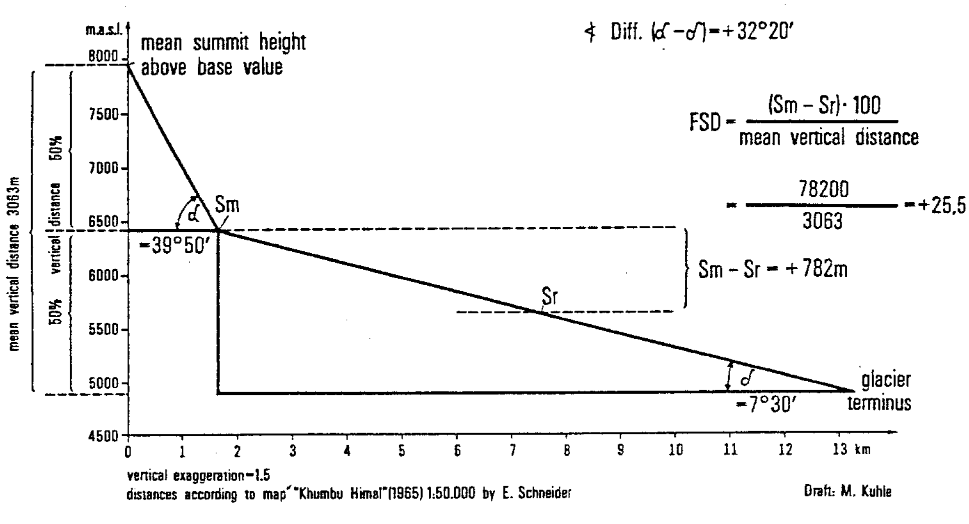
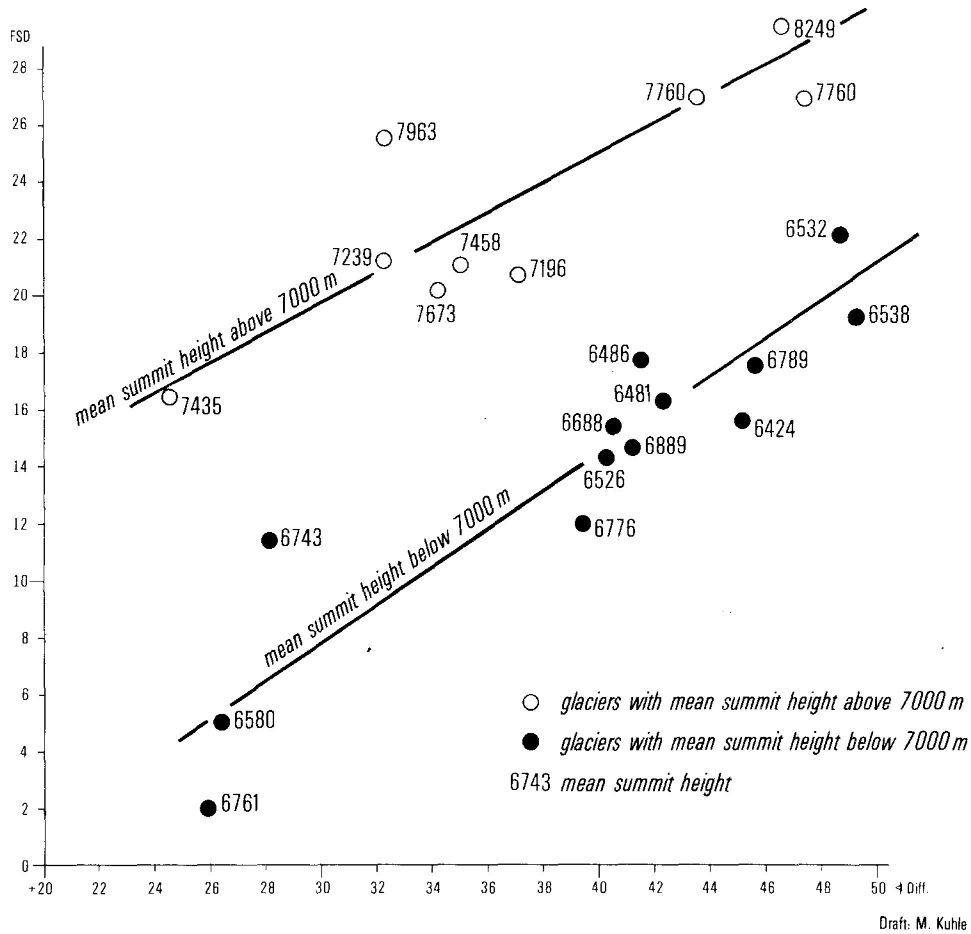


Fig 9 Differences in angles and snow-lines: the example of the Khumbu Glacier (S slope of Mt. Everest)

Fig 10 Gradients of snow-line differences in their dependence upon angular differences in respect of the glaciers of Khumbu Himal



are nonetheless bound to the 0°C line for the absorption of water.

Notes on the problematic of surface temperatures on rock- and ice-walls

Tab 2 contrasts measurements taken on rock- and ice-walls during radiation weather conditions between 11,00 and 15,00 and 6,00 and 11,00 hours. All values were registered on the S slopes of the Mt. Everest area (Khumbu Himalaya) during the post-monsoon season of 1982. As a result of the comparatively low radiation, the early morning values are closer to the nocturnal values than to the maxima at noon, indicated in the 0°C line which drops by 456 m in rock, and by 652 m in ice. The diurnally-related fluctuation of the temperature gradient in ice from 1.11 to 1.47°C/100 m (Tab 2) is characteristic of the thermal behaviour of both the surfaces. The levelling of the gradient by 0.36°C is to be explained by the *heat loss which is caused by the melting process* around noon *that makes itself felt upwards over a further 652 m*. This is absent from rock surfaces. The removal of latent heat below the 0°C line does not permit any warming up of the ice surface to more than 0°C, as happens on rock faces. Thus the gradient for ice surfaces as calculated from all the random samples experiences a *levelling-off*. How much levelling-off takes place depends on the altitude of the 0°C line within the vertical extension of the ice flanks. As shown by the almost corresponding gradient of values recorded between 6,00 and 11,00 hours in rock and ice. The 0°C line is bound to be close to the lower line of glaciation and iced-up flanks at this time, whilst in the case of the higher 0°C line a wider (652 m) altitudinal belt,

Tab 2 Measurements of temperature under sunny weather conditions in rock- and ice walls at 11.00–15.00 and 6.00–11.00 o'clock.

substratum	time	0-degree limit [m.a.s.l.]	temperature gradient [C/100m]	exposure	difference of altitude [m]	difference of temperature gradient [C/100m]
ROCK	11–15 (DAY)	5926	1.51	---	456	0.03
ROCK	6–11 (NIGHT)	5470	1.48	---		
ICE	11–15 (DAY)	5690	1.11	---	652	0.36
ICE	6–11 (NIGHT)	5038	1.47	---		
ROCK	11–15 (DAY)	5920	1.52	SW–SE	214	0.44
ICE	11–15 (DAY)	5706	1.08	SW–SE		

with heat loss through melting processes, is included in the correlation. In the negative range of temperatures, rock and ice surfaces thus behave in a similar way, thermally speaking. Values extracted for the expositions of the strongest solar radiation from SW to SE confirm a difference in gradients from 1.52°C on rock and 1.08°C on ice at noon (Tab 2).

The observation that the 0°C line shows a smaller rise on rock than on ice during a given interval of time (Tab 2) is to be interpreted as follows: the conclusion that rock surfaces due to lesser albedo and heat capacity have been more heated-up by 11,00 than equivalent ice surfaces at altitudes of more than 400 m (432 m) below, suggests itself[6]. Before sunrise, i. e. before the first impact of radiation on the other hand, values recorded on rock are lower than those on ice: as a result of greater emission the 0°C line occurs at a lower altitude in rock surfaces than on ice surfaces. Indication of the same effect is the position of the 0°C line at noon, which is 236 m and 214 m higher on rock than on ice (Tab 2). But in contrast to the difference of 432 m in the early hours of the day, the interval of the 0°C lines on the two materials has narrowed, which points to a catching up in the energy balance of the ice during the middle of the day.

The differentiations referred to in this context do not affect the results fo this investigation, but illustrate modifications conditioned by the substratum. It ought to be said, however, that the differentiation between areas singled out for measurements on rock or on ice amounts to a rough generalization in view of the detailed characteristics of types of rock and degree of metamorphosis of the firn. On the other hand, there are mixed conditions even in this differentiation, for every rock-wall within the altitudinal interval under investigation is covered by fields of snow, firn or ice in places, just as all ice flanks are interspersed with rocky heads and columns as well as dry, over-hanging wall sections (Fig 2, 5, 6).

Some observations on the gradient of surface temperatures between 4000 and 8848 m asl

Registered on the average between 11,00 and 15,00 hours, the gradient amounts to 1.45°C/100 m (Fig 7). Depending on the proportion of ice wall areas *from the 0°C downwards on the measured values*, the temperature gradients fluctuate between 1.45° and 1.52°C/100 m. In general, however, they are 0.6–0.7°C above that of the open air (Tab 2). Whilst the gradient is kept low in the atmosphere by advective energy transport from valley floors, which, being situated at altitudes of up to 5200 m, function as heating panels, the balance of received radiation and refraction in situ decides on the rock surfaces and firn-ice surfaces above 7000–7200 m. The atmospheric transfer of energy might well be much reduced or even suppressed by the cooling areas and melting processes of the ice from the glacier tongues (from above 5200 m) to the upper-line of ice formation, and even further up to the region of snow-rock (max. 8848 m) (Shumskii 1964, p. 408). This effect might be regarded as one of the causes for air masses, which have been pushed up and arrived at the upper limit of glaciation, to be devoid of the energies necessary for metamorphosis of firn, and that the upper limit of glaciation is thus essentially conditioned by the local radiation balance. One has to regard the glaciers, which flow off high mountains, as an advective barrage for air-masses near the surface and their otherwise warming influence.

The pergelid level of rock and scree above the altitudinal level of glaciers

Above 7000–7200 m the upper glacier limit is followed by the truly highest geomorphological stage of a rock and scree region. In a milieu where even the surface of the soil is permanently frozen, the processed scree, which consists of a thin scattering of components ranging in size from that of gravel to coarse blocks, can only be the result of temperature weathering alone. Taking place without a change in the frosty temperature and in physiological aridity, this process decomposes the rock through the differential expansion of its constituent minerals during temperature fluctuations in the minus range between –20° and –40°C, for instance. It has so far not been possible to estimate the speed with which this weathering takes place. It is assumed that the weathering intensity as a function of *increasing brittleness* in decreasing temperatures is greater than the insolation weathering in warm-arid milieu. This weathering, which contributes to the shaping of the very few high peaks on earth, will be of more widespread significance on other planets of our solar system, such as Mars, where surface temperature fluctuations of mean values from –30° to –85°C have been recorded at the landing place of Viking 1 (according to Stanek 1980).

The lowering of the upper limit of glaciation in the Pleistocene

Quaternary-morphological reconstructions of the main ice-age snow-line depressions carried out by the author in the Central and Tibetan Himalayas, as well as in S Tibet (Kuhle 1982b; 1984), resulted in a lowering of the climatic snow-line in the region of the high peaks under consideration here by at least 1100–1200 m. Assuming hygric conditions to have remained unchanged and a gradient of air temperature at 0.8°C/100 m, the drop in temperature during the warmest months is calculated to be 9,6°C. Since thermal conditions constitute the upper glacier-line in the same way as the snow-line, an analagous lowering by 1100–1200 m may be assumed. In accordance with the present state of knowledge, however, it cannot be ruled out that in the matter of Pleistocene *lowering of the upper-glacier limit* the steeper gradient of surface temperatures of about 1.45°C/100 m (Fig 7) must be taken into consideration, this would be approximately equal to half the lowering of the snow-line. For the time being the conclusion will therefore be restricted to a Pleistocene lowering of the upper glacier limit by at least 660–1200 m, i. e. to about 6540–5800 m asl.

The altitudinal level of glaciers as an example of a bilateral altitudinal level

It is commonly assumed in the subject of biology that a biotope, for instance, loses its conditions to adjacent biotopes of unrelated kinds on both sides, and that at a distance from the two liminals the optimal conditions are maintained, which get lost in different ways towards the borders; this approach has its equivalent in geomorphological research into altitudinal levels. The author has shown (1978) that the periglacial altitudinal level possesses such characteristics. It is bilaterally constructed, for it not only fades away downwards from its most marked formation indicators at an intermediate height, but also approximately inversely symetrically upwards to an upper line. Equally *bilateral* is the *altitudinal level of glaciers in the Himalayas*, as proved by the above explanations. Its lower border are the snow-line and the glacier tongues that join on below. On the other hand, the glacier-level also thins out towards an upper-line of the feeding-area between 7000 and 7200 m asl. Between these two in the Mt. Everest group, a few hundred metres above the snow-line, at 6100 m on the S slope and between 6400 and 6600 m on the N one, there is an *optimal zone of glacier-ice formation.*

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Footnotes

1) Paucity of precipitation is no hindrance to potential glacier formation. In the Antarctic inland ice is maintained by an annual precipitation of less than 50–30 mm.
2) All these undertakings were financed by the German Research Society and the Max Planck Society.
3) Jens-Peter Jacobsen, Dip. Geogr. assisted with the collection of measurement data.
4) At 30° North on September 21. (zenith distance 30°)1180 W/m^2 are reached at the upper limit of the atmosphere.
5) The definition of the S(m) is carried out in two steps; the author developed it on the model of v. Höfer's method (1879), using an initial method of approach by adopting a basic value, which is then followed by a calculation of the altitude of the catchment area that considers the heights of the peaks above the basic value. This has the advantage of eliminating v. Höfer's circular argument in the calculation of the average altitude of the ridge course are not indicated on them. In general the results of the approach adopted by the author resemble the values basic value. The method choosing the average altitude of summits in place of the ridge frame, facilitates the definition of the mathematical snow-line on the basis of simple sketches of the course of the ridges which are most likely to be provided with data about altitude of their peaks, whilst the altitudes of the ridge course are not indicated on them. In general the results of the approach adopted by the author resemble the value calculated by v. Höfer.
6) At 13,45 hours on October 21st, 1984, the author measured an albedo of 88.9 % in firn basins on the NE flank of Mt. Everest at an altitude of 6650 m asl.