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Altitudinal Levels and Altitudinal Limits in High Mountains

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Abstract: In lowlands climate-specific processes due to weathering and erosion are dominant, whilst the geomorphology of mountains is dependent on the geologic-tectonic structure, i.e., the energy of erosion that increases according to the vertical. The expression "extremely high mountains" has been established as the extreme of a continuous mountain classification. It has to be understood in terms of geomorphology, glaciology and vegetation. Correspondence of the planetary and hypsometric change of forms is of great value as synthetic explanation. It is confirmed with regard to vegetation, periglacial geomorphology and glaciology. Due to the world-wide reconstruction of the snowline its paleoclimatic importance increases, too. Apart from lower limits the periglacial and glacial altitudinal levels also show zones of optimum development and climatic upper limits in the highest mountains of the earth. According to the proportion of the altitudinal levels a classification as to arid, temperate and humid high mountains has been carried out.

Keywords: Altitudinal levels; altitudinal limits; high Mountains

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Introduction

Mountains form the third pole of the earth, because they hypsometrically repeat the climate zones as far as the Arctic within smallest space. The relief energy is the central characteristic of every classification of mountains. Considering this basic size, mountain-typological threshold values and quality leaps have to be newly settled by geomorphologists, glaciologists and geobotanists according to the special question.

1 Relief Energy and Removal of Material

The slope of the mountains accelerates all processes by removal of material and cinetic energy. The relief energy enhances the clearness of forms. Debris flow cones increase and, in steeper catchment areas, they can be more clearly differentiated from alluvial fans. Avalanche ravines become larger at the increase of slope, and gorges in confluence steps are deeper than those in flatly-hanging longitudinal valley profiles.

The idea of increasing processes in the mountains might lead to the impression that the same kind of geomorphological modifications is concerned here than in the lowlands. However, this is not true, because a quantitative continuum between the two extremes of process-intensity does not exist. On the contrary, the cover of residual detritus in the lowland preserves the underground and causes preservation of the lowland. The consideration derived from the removal of detritus accelerated by gravitational force, leads to the polarization of climate-genetic flat relief, and to a steep relief which with increasing incline of slope is more and more controlled by the geological structure (Kuhle 1982a). At the same time, i.e., with increasing steepness, characteristics of the agents water, rock fall, snow- and ice avalanches retreat against their forming kinetic energy (Photos 1 and 2).

Mass movement is only possible on the overhang or the vertical. However, the steep break-offs in the Andes and the Himalaya already show the form of a wall without debris cover at 43° to 50°. This gives evidence of their potential of mass movement. Due to the 2500 to 4500 m-high walls of the Himalaya and the Andes, the kinetic energy of rock fall and avalanches increases. The result of this greater impetus is that the rock faces are free of debris at smaller angles than in the lower Alps. Obviously the slope of the face decreases with increasing height. This is also proved by the large, already flattened flanks in the Alps, e.g., the 3300 m-high Mt. Blanc-S-wall and the 2500 m-high Mt. Rosa-E-wall.

2 Classification of Mountains

Krebs (1928) established for high mountains a difference in height of 1500 m between the summit and the talweg. Louis (1963) tried to determine the relief energy between the non-dissected pedestal and the tangential face of the summits. However, such numerical demarcations against low mountain ranges and lowlands are only descriptive of the Alps, but as to larger mountain systems of the earth they are normative and pure coincidence. Only via quality leaps it is possible to prove that threshold values such as these make mountains to high mountains. Here, geomorphological sequences have to be taken in consideration, starting at a certain vertical distance. Consequently, the definition of a "high mountain nature" according to the glacial forms after Troll (1955) can be accepted. Every relief that has towered above the Pleistocene snowline belongs to it. At the same time it has to rise above the current timberline and to show solifluction. However, a problem is the exclusion of the polar zones according to which the mountains of E-Greenland, W-Svalbard and the Antarctic do not belong to the high mountains because forest is absent. Thus the proposal is a global concept that even today is invalid. A concept valid in the past could not be successful anyway, because at the time when glaciers flowed down into the forelands of mountains like the Alps or Scandes, timberlines did not exist there, too. Furrer et al. (1970) have suggested that the high mountain level have come to an end at the solifluction boundary, i.e., in N-Scandinavia and N-Island according to the sea level. The approach of Jessen (1950) contrasts to that of Troll (1955), who restricts the "high mountain nature" to the highest levels insofar as he interprets the upper levels to be a part of the entire mountain range with a dynamic influence on the valley floors and forelands as basal areas below the timberline.

Apart from these considerations geomorphological classifications of the "Groß- and Größtformen-Gebirge" (Large- and largest forms of mountains) were tried by Maull (1958) and Dongus (1980). Also the classifications of mountains with strong tectonics and steepening due to the resulting erosion are worth discussing. For instance the relief energy of the Himalaya with, e.g., a difference in height of 6800 m at a distance of 17 km was described as "extreme high mountain region" (Kuhle 1982a). A characteristic is the non-modified reworking of the geologic- tectonic structure through all mechanisms of erosion from the glacier level down to the lowest level of forest. The flow of Ice Age glaciers through the periglacial zone down to the colline level, as well as a relief extending as far as the climatic upper limit of glaciers (Kuhle 1986a) (Photos 1 and 2; Figure 1) can be considered to be further criterions (4.2).

The increase of nourishing areas caused the glaciers – also relatively to the decrease of the snowline – to flow lower down than today and to cross the periglacial zone. This restricts the relief energy with regard to the definition of extreme high mountains. It is not the height of the summits above the glacial snowline, which is the reason for the glaciers' flowing far down, but that of the upper limit of the glaciers. This difference in height is based on humidity, because the altitudinal level of the glaciers (between ELA (equilibrium line altitude) and upper limit of the glacier) is the widest where it is more humid than cold, e.g., in Alaska (Figure 1). On the Everest-N-side it is arid at the snowline level, so that the glacier level shows a width of only 900 ~ 1300 m instead of 4000 ~ 4400 m (see Photo 2).

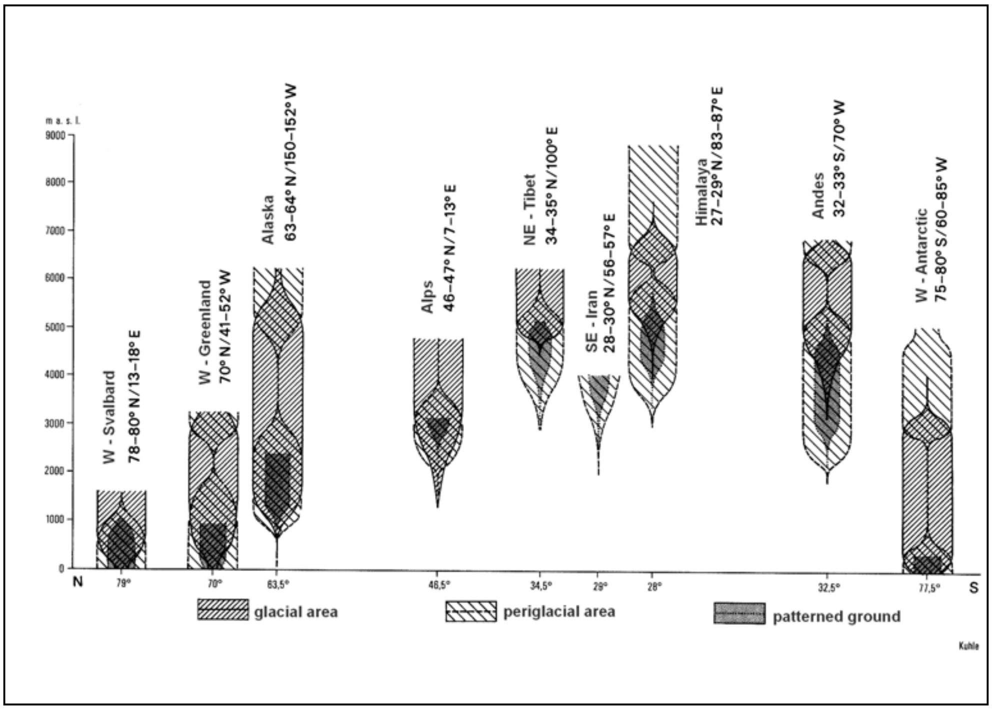

Figure 1 Nine combinations of glacial and periglacial altitudinal levels in mountains of the earth

Increase of humidity as far down as to the foot of the Himalaya reduces frost weathering, which would be more active in an arid-continental area. This is the case in the Karakorum, where the Ice Age glaciers of the N-side did not flow through the periglacial area, so that deep glacier polishings are lacking, whilst the preserved polishings of the humid Himalaya S-side are proof of a discharge up to below the limit of freezing and thawing. Owing to this, at the same altitude the Himalaya is a more extreme mountain range than the more arid Karakorum.

The same is true for vegetation. In the tropical Andes (Lauer 1979) and the monsoonal-humid Himalaya (Schweinfurth 1957) the system with the greatest variety of vegetation levels is the best to define extremely high mountains.

However, the glacier level is the highest in the arid subtropics. In the Andes the snowline exceeds 6000 m, in Tibet and the Himalaya-N-slope it comes to 6000 to 6400 m (Figure 1) (Kuhle 1982a & 1985a) whilst it falls down again to the tropics. The vegetation level is reduced, but the periglacial level reaches its greatest width (Figure 1) and variety (Kuhle 1978b & 1985b; see 4.1). This applies to solifluction types in dependence on the vegetation cover. So, in the N-Karakorum (35°30' ~ 36°10'N/76 ~ 77°E), in the Aghil Mountains (36°10' ~ 30°N/76 ~ 77°E; Kuhle 1988a) and the Tibetan Himalaya (Kuhle 1982a) a 200 ~ 500 m wide belt of solifluction bound by a cover of alpine meadow has been found. Above and below, with the ending of the vegetation, it passes over from retarded solifluction to free solifluction (Figure 2). At the same time, the vegetation cover toward above – due to the decreasing temperature – and toward below – due to the decreasing humidity - thins out and finally stops.

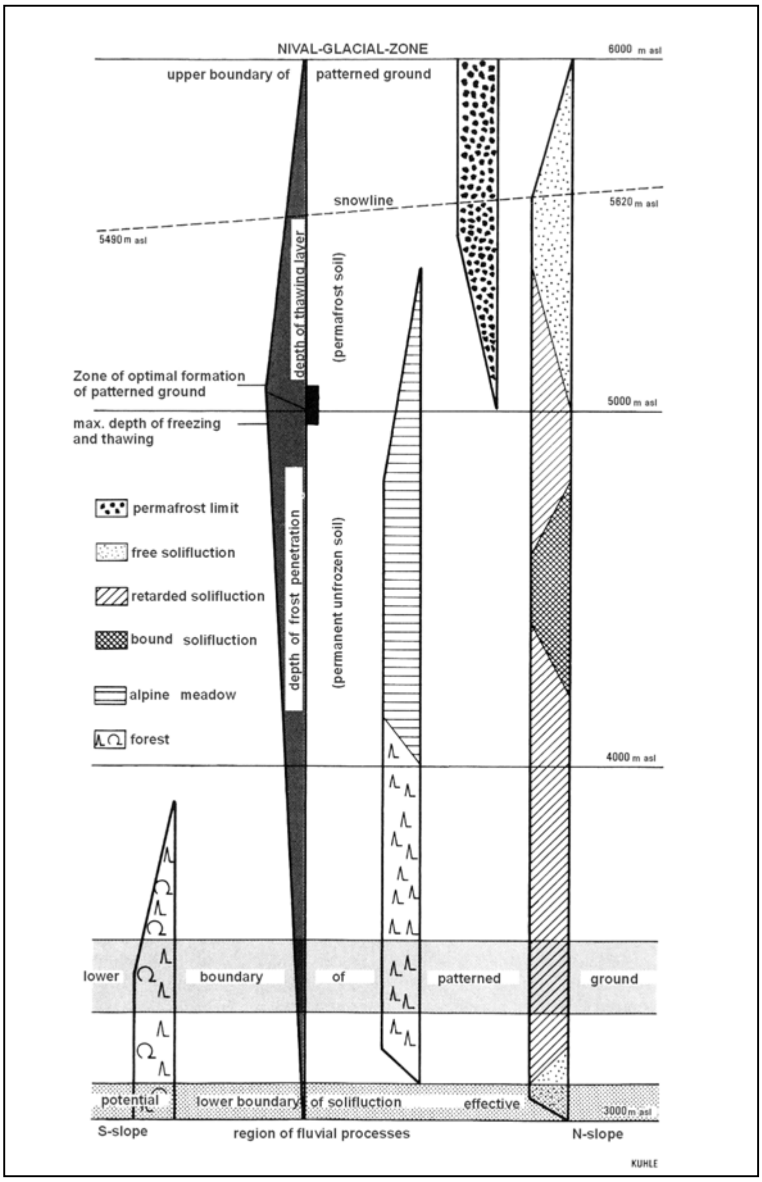

Figure 2 Periglacial area of the Himalaya given with the example of the Dhaulagiri-group (28°20' ~ 28°55'N/83° ~ 84°E)

At the subtropical periglacial level the sorting of material is of great variety, too. Here a mirror-symmetric pattern of a zone of optimal formation of patterned ground (Kuhle 1978b) is shown with activities which above and below come to an end and decrease to form an upper boundary of patterned ground in the area of permafrost and a lower boundary in the area of ever thawing soil (4.1).

Subtropical and tropical mountains are similar as to their great variety. In the first case the periglacial level contributes to it, in the second case the vegetation levels.

A further characteristic of mountain classification is the number of glacier types of a mountain range which are run through during a climate change. This is a function of its relief energy. A glacier typology in dependence on the vertical distance (Kuhle 1988b) makes obvious that all glacier types from the flank glaciation up to the ice stream network are run through, provided that the snowline sinks into a mountain relief. Flank glaciation will occur if the snowline remains under the level of the summit. If it moves downward, cirque- and wall foot glaciers will come into being. Then avalanche cone- (Kuhle 1982a) and avalanche cauldron glaciers will follow (Visser 1934). If the ELA remains under the valley floor, firn cauldron glaciers and then firn stream glaciers will be formed. At an even thicker ice filling of the valley, firn field glaciers will occur and if the ice overflows the valley divides ice stream networks will be made up. The depth of the valley is more important than the height of the summits. At a corresponding ELA-depression the sequence of types is run through faster by high mountains with an insignificant valley depth than by extreme high mountains, the relief infilling of which progresses slower. In the Alps a cupola-shaped ice stream network has been built up during the pleistocene drop of the snowline of 1200, whilst in the Himalaya at a lowering of the snowline of 1500 m, only three-times longer valley glaciers have been developed (Kuhle 1982a), so that the typological distance to the alpine ice stream network has nearly remained.

Accordingly, due to a minor valley depth the Karakorum ranks in this order of priority of high mountains behind the Himalaya. At the same time the height of the valley bottom is the cause of today's more extensive glaciation of the Karakorum.

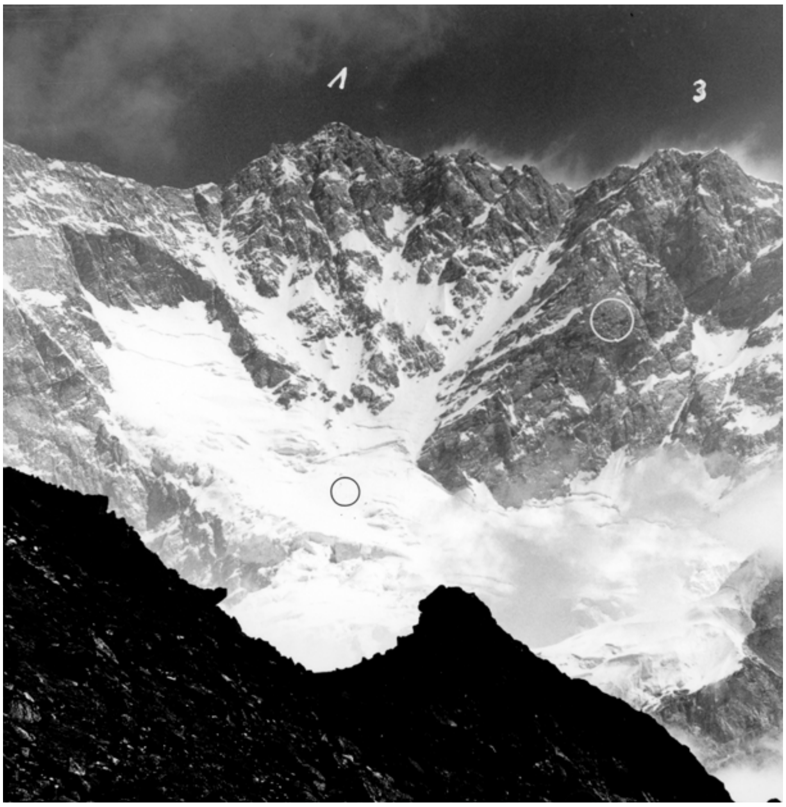

Photo 1 Taken at 5000 m asl from the Yalung glacier valley (upper Simbua Khola; 27°40'58"N/88°05'20"E) seen to the NE into the S-slope of Kangchendz"nga (8586 m, E-Himalaya). No.1 is the 8586 m-high main summit, No.3 the 8474 m-high S-summit. The rock crest on the left stretches up to the 8420 m-high Kangchendz"nga W-summit. The summit-crest of the mountain exceeds 8000 m asl over a distance of c. 6 km. (O black) marks the highest n,v, surfaces of the Yalung glacier between 6600 and c. 7600 m asl. Only in a basin-like and therefore stable leeward position they are able to survive. (O white) marks the rock faces which between 7300 and c. 7900 m are more and more free of glacier ice and n,v,. Between 7900 m and the summit crest the rock is merely temporarily covered with smaller patches of n,v,, annual monsoonal snow and winter snow, but not with a glacier. At the time when this picture was taken the monsoonal snow of the year 1998 was already blown away like dry flying sand. The fresh winter snow, too, will be completely blown away when the monsoon sets in 6 to 8 weeks later. Analogue photo M.Kuhle, 26/4/1999, 11.05 p.m..

Photo 2 At 4500 m asl from the Modi Khola glacier valley (upper Modi Khola; 28°34'N/83°52'35"E) seen to the N into the S-slope of Annapurna (8091 m = No.1, Central Himalaya). (☐ black) marks larger n,v, surfaces of the perennial ice cover of the approx. 4000 m-high Annapurna-S-wall between 5600 and c. 7400 m asl. (☐ white) shows rock faces between 7000 and c. 8000 m asl which are increasingly free of ice. The rock between 7600 m and the summit crest is temporarily covered with comparatively small patches of n,v, and annual monsoonal snow; ice- and snow cover decreases with increasing height. Up there ice balconies are lacking so that tracks of ice avalanches only begin further down (below ☐ white). The rocks marked with (☐ white) are faces of outcropping edges of the strata. The last monsoon snow of the year 2002 fallen some days ago has been deposited on their small-scale ledges so that the rocks show a light-grey colour. This fresh snow will be blown away at the end of October. (▪) is the Modi glacier 1200 m below the ELA and covered with surface moraine. Between the glacier tongue end at 3700 m and the snowline at 5600 m an ablation area of 1900 m is situated; between the upper limit of the glacier about 7600 m and the snowline lies an accumulation area of 2000 m so that the vertical extension of the glacier's altitudinal level on the Himalaya scour-side facing towards the monsoon is approx. 4000 m. Accordingly, it is nearly twice the size of that on the semiarid lee side of the Himalaya (see Figure 1, Himalaya; this example concerns the N-side of Mt. Everest). Analogue photo M.Kuhle, 27/9/2002, 6.25 p.m..

3 Correspondence of Planetary and Hypsometric Change of Forms

The global climate zones with their geographic phenomena reappear in the altitudinal levels of high mountains (Lautensach 1952). So, e.g., in Central Europe a meridional shifting of 100 km corresponds to a hypsometric one of 100 altitude metres. This is equivalent to a difference in temperature of 0.5 ~ 0.6°C. Köppen (1923) developed a climate sytem in which for instance both the tundra areas of Greenland and Svalbard at sea level and the dwarf scrub steppes in Tibet about 5000 m asl belong to the tundra climate. The periglacial-geomorphologic findings fit into this three-dimensional system, too. Forms of patterned ground of arctic dimensions (Furrer 1965, Kuhle 1978b & 1982a & 1983) and ice wedges as well as pingos in High Asia between 27° and 39°N (An Z. 1980, Zho *et al.* 1982, Kuhle 1985b) and between 4200 and 5600 m are proof of a correspondence of subtropical and arctic periglacial phenomena. However, relationship does not mean identity. The effect of solifluction during the time of the day and the time of the year (Troll 1944) with regard to the forms of subtropical mountains and polar latitudes, as well as to the different vertical width of this region of forming (Figure 1) shows differences. Accordingly, comparing the glacier areas close to the sea with those closer to the equator, we have to consider differences caused by the latitude, in spite of the similarity of the metamorphosis of snow into ice and an ice accumulation leading to similar ice streams. It is not the temperature in the true sense of thermal glacier classification (Lagally 1932) which is appropriate for a differentiation, because in the subtropics, too, ice temperatures of -6 to -15°C, 5 m below the surface, have been measured at the level of the snowline (Chijian et al., 1980, Kuhle 1990), i.e., tempered and cold glaciers have been evidenced. In addition, a climatic upper limit of glaciers dependent on extremely low surface temperatures of below -15 to -25°C on average, has been proved in the subtropics (4.2).

A distinctive feature is the span of temperature from extremely cold glacier nourishing areas far above the snowline up to the warm glacier tongues, which flow down to 500 m below the timberline in monsoonal areas. In arid mountains glacier tongues end much higher and are much colder. Due to their snowline near to the sea level, polar glaciers are not in the position to flow down into warmer regions. They already calve into the sea at the subnival level (Figure 1 W-Svalbard, W-Greenland, W-Antarctic). At most in the Subarctic, e.g., in Alaska, glaciers with the highest nourishing areas flow down into the forest and have tempered ice with summer meltwater discharge. This forest grows on permafrost.

In the polar region as well as in the mountains of low latitudes glaciers occur with quasi-laminar flow dynamics (Finsterwalder 1897), i.e., a hill-shaped velocity cross-profile as well as box-shaped profiles induced by the movement of block segments (Blockschollen) (Pillewizer 1958). Due to low temperatures and large ice masses the last ones are characteristic of the Antarctic.

An exclusive feature of the glaciers within the 38th latitude degrees are ice pyramids thawed out up to a height of 25 m (Kuhle 1987). They develop from the 4 to 5 times more intensive insolation (Bernhardt *et al.* 1985) of mountains close to the equator that profit from the transparency of the atmosphere at high altitudes (Kuhle 1986b). At the insignificant heights of polar glaciation, only 60 % of the radiation, which is lesser anyway (Grunow 1961) , reaches down to the ice surfaces.

The planetary-hypsometric correspondence is also evidenced by the decrease in height of the glacier levels according to the latitude.

4 The Structural Pattern of Altitudinal Levels of Mountains Inclusive and Above the Periglacial Level (Figure 1)

4.1 The periglacial area

A comprehensive system of altitudinal levels is the Himalaya (2). The variety of the periglacial level in the subtropical-arid N-slope of the Dhaulagiri- and Annapurna-group (Figure 2) is so great that it contains the conditions of the other mountains. Here it is 3000 m wide. This vertical span is also reached in the Karakorum, Kuen Lun (35°50' ~ 38°N/76 ~ 78°E) and in the subtropical Andes (32 ~ 33°15'S/71 ~ 69°10'W) (Figure 1). Toward above and below it is symmetric. The transition from solifluction bound by meadow vegetation via retarded to free solifluction in the vegetation-free region of frost-debris is also known from the Alps (Furrer 1965), the Apeninnes (Kelletat 1969), the Greek mountains (Hagedorn 1969) and the Scandes (Garleff 1970). The combination of a thermally conditioned upper limit of vegetation with a lower limit dependent on dryness is a speciality of the Himalaya (Figure 2). The vegetation cover of the lower level of retarded solifluction can be recognized by the forest and dwarf scrubs.

Close to the lower boundary of solifluction the forest's boundary of aridity is reached and the vegetation is reduced to a sparse steppe, so that a lower fringe retarded to free solifluction exists. Accordingly, the periglacial level does not come to an end somewhat above the timberline, but intersperses the sparse dry forest with small slope terraces. The Himalaya S-slope, turned towards the monsoon, is covered with dense fog forest, into which patterned ground and solifluction potentially reach down up to 600 ~ 800 m. Evidence of this has been provided by loop-soils in avalanche tracks (Kuhle 1982a). The intersection of the upper limit of the forest and the lower limit of solifluction, which does not occur in the Alps, can be explained by the difference of the two altitude limits: due to the timberline, in oceanic summers the 10.5°-isotherm of the warmest month is low, whilst in the Alps the freezing and thawing during the winter cannot become formative, because of the isolation by snow. Only the freezing and thawing from spring to fall is effective, so that the lower limit of solifluction runs high up. In the Himalaya the winter is continental-cold along with an absent or insignificant snow cover. Its thickness only increases in spring, so that solifluction covers, forced by subtropical-frequent changes of freezing and thawing, reach far down into the forest. At the same time, due to the summer temperature, the forest extends far up.

The second symmetry concerns the thickness of the debris layer of freezing and thawing (active layer) and the diameter of the patterned ground (Kuhle 1978b & 1985b). It is shown above and below of a zone of optimal formation of patterned ground by a decreasing intensity of forming (Figure 2). The extent of patterned ground decreases toward a lower boundary as well as toward an upper boundary realized on the Himalaya N-slope. In each case the depth of the penetration of freezing and thawing diminishes. It starts below the zone of optimal formation of frost penetration into the thawing layer and above the optimal penetration of thawing into the permafrost soil. The lower limit of permafrost runs where the greatest depth of frost penetration in the ever thawing soil coincides with the max. depth of the thawing layer. Accordingly, it can be diagnosed by the most extended patterned grounds, which in the Himalaya reach polar dimensions of 2 ~ 3 m in diameter. It is controlled by the greatest depth of the thawing layer, i.e. frost penetration, because the proportion of 3:2 of diameter to depth of sorting is constant.

In contrast to the lower limit of patterned ground, i.e. the lower limit of solifluction, in the Himalaya the upper limit varies its altitude with the exposition. It fluctuates between 5600 m in N- and 6000 m in S-exposition. The lower limit is dependent on the depth of frost penetration, i.e. night-temperature, the upper limit on the depth of thawing and thus on the geometry of radiation (Kuhle 1982a).

Figure 1 raises the question why the bilateral development of the periglacial altitudinal level has not taken place in all mountains. In the arid Iran the summits are non-glaciated and more and more periglacially formed up to the top. The dimensions of patterned grounds increase up to their highest ocurrence which is restricted by the height of the relief. In the humid Alps where solifluction areas toward above are more intensively sorted, they are limited by glaciers. At most the zone of optimum development of patterned ground is reached, because up to the ice cover the patterned grounds do not get smaller again. The ELA runs so deep in the relief that the glacier tongues pierce the periglacial level. Humidity hinders an upper limit and even a reduction of the patterned ground above an optimum zone. In Alaska, W-Greenland and the W-Antarctic (Miotke 1981) the same conditions as to a glacier-forced upper boundary have been observed, whilst the lower boundary of the periglacial area shows differences to the Alps. In Alaska it reaches lower down than the glaciers. There it must be relatively colder than humid. In Greenland and the Antarctic the periglacial area (Figure 1) reaches up to the sea-level; in the Antarctic it occurs within the optimum zone of patterned ground, whilst in W-Greenland the extent of patterned ground already diminishes from 70°N downwards.

W-Svalbard shows antarctic conditions at the lower boundary, at the upper boundary they are concordant with NE-Tibet. It is so dry that the optimum zone is exceeded and the extent of the patterned ground decreases up to an upper limit dictated by the glaciers. Only in high subtropical mountains the entire periglacial area is cleared of glaciation, so that there is a climatic upper boundary of patterned ground (Figure 1). Up there the necessary thawing process is lacking and – as below of the lower boundary of patterned ground – a structure does not occur. Parts of the Himalaya, Karakorum and the Andes are found to be types of arid high mountains that can be defined in such a way. In spite of the cold which allows an upper boundary of patterned ground, their snowline is so high that debris, free from névé, occurs up to the top.

4.2 The glacial area and the upper limit of glaciers

Along with the decreasing temperature, the periglacial level toward above is followed by the glacier altidudinal level. It has been suggested to be the highest level limited by summits. However, telemetrical temperature measurements on mountains in excess of 7000 m prove that glaciers peter out from 7200 m onwards (Figure 1) in the Himalaya and from 6900 m asl in the Karakorum (Kuhle 1986a). Regression analyses of 2000 surface temperatures have shown that above 7200 m it is too cold for glacier formation. At correlation coefficients of -0.835 the surface temperatures measured on mountain flanks during radiation weather conditions between 11.00 and 15.00 o'clock, lie with an error of probability of <1 ‰, a gradient of -1.45°C/100 m and a 95 %-probability prediction above 7400 m always and above 7000 m nearly always below the freezing point. At this altitude the glaciers come to an end and rock appears on the surface (see Photos 1 and 2). The coincidence of the upper glacier limit with the 0°C-line suggests a causality which reveals the adhesiveness of the névé-ice to be a function of the snow-to-ice metamorphosis. On the slopes of K2, during midday-insolation, -28 to -35°C have been measured up to a summit-height of 8616 m; temperatures close to the melting point have been measured at about 6900 m during calm minutes. Average temperatures below -19 to -25°C prevail above. In an environment of this type the development of ice bridges between the snow grains takes only place by molecular diffusion (Benson 1961). This process needs 20- to 33-times the time which applies to the metamorphosis of temperature close to the melting point. Before the fresh snow – due to this recrystallization by pressure compaction – reaches an ice density of 830 kg/m³, an overlying load of snow of 100 m is needed, lasting more than 100 years. Because the incline of the mountain flanks does not allow snow accumulations such as these and the wind within weeks has blown away the cold dry snow like flying sand, a thermal upper limit of glaciers of more than 6900, i.e. 7200 m asl develops. Only in stable lee positions and due to pressure compaction, hanging glaciers occur above (Photo 1).

Accordingly, in the Himalaya we meet a highest permamently frozen (pergelid) upper altitudinal level of rock (Figure 1) with a summer snow cover. There, weathering without humidity at variations of minus temperatures does occur — just by the extension difference of the minerals.

Also mountains of the Antarctic pierce through the upper limit of glaciers, so e.g. at 3600 m asl in the Vinson-massif (Figure 1). In Alaska, Greenland and the Andes additional surfaces lay above the upper limit of glaciers during the glacial period. They have been lowered in parallel with the snowline. Whilst this has run 1200 m lower than at present, the upper limit of glaciers – which instead of a gradient of 0.7°C/100 m shows that of a surface temperature of 1.45°C/100 m – had dropped by 660 m (Kuhle 1986a). With a lower limit of glaciation that had decreased by an amount of 1.5-to 2-times of the ELA-depression, the picture of a glacier level arises oscillating in the mountain relief, the optimum nourishing zone of which is in the middle between the upper glacier limit and the ELA.

A low upper glacier limit is consistent with a high lower limit of glaciation and the development of an upper limit of patterned ground, i.e. with aridity and minor temperatures (Figure 1 Himalaya). In such a way the sum of precipitation and evaporation is mirrored in the vertical extent of the glacier level.

A wide periglacial area and a narrow glacial one which is not restricted by the relief - as they are in the Himalaya and in some parts of the Andes (Figure 1) – stand for aridity. A far-reaching covering of the periglacial level by an expansive glacier level as in the Alps and Alaska (Figure 1) indicates humid mountains. The Alps are even more humid than the Alaska-range, because they do not tower as far above the ELA and – in spite of that – the larger glaciers pierce through the periglacial area. An upper glacier limit is not realized; not either in W-Svalbard and Tibet. In relation to the periglacial area, here the lower glacier limit runs higher than in the Alps, so that a semi-humid intermediate position exists. In W-Svalbard, Greenland and the Antarctic the glaciers calve into the sea. Accordingly, the sea-level dictates the lower glacier limit. Though the ELA in the Arctic is closest to the sea level, the glacial altitudinal level attains a vertical span of 3000 m up to the upper limit of the glacier. So – in spite of a lesser precipitation – regarding the climate the Antarctic is close to the mountains of Greenland and Alaska.

Supra-regional comparisons of the altitudinal levels of mountains such as these render the understanding of the climate factors temperature, precipitation and radiation possible. They point to the typifying of their product. Analyses of the width of altitudinal levels as well as climate factors of the upper- and lower limits are able to classify arid, temperate and humid mountains – without absolute energy- and mass-balances.

References

An Zhongyuan. 1980. Formation and evolution of permanent ice mound on Quing-Zang Plateau. In: J. Glaciology and Cryopedology 2, No. 2, Lanzhou, Pp. 25 ~ 30.

Benson, C.S. 1961. Stratigraphic studies in the snow and firn of the Greenland Ice Sheet. Folia Geogr. Danica, 9, Pp.13 ~ 37

Bernhardt, F. & Philipps, H. 1958. Die räumliche und zeitliche Verteilung der Einstrahlung, der Ausstrahlung und der Strahlungsbilanz im Meeresniveau. Die Einstrahlung. Abh. Meteor. Hydrol. Dienst DDR Nr. 15, Pp.1 ~ 227.

Chijian Mei & Ren Bing-Hui (Hrsg.) 1980. Glaciers in China. Lanzhou Inst, of Glaciology and Cryopedology, Academia Sinica, Shanghai, Pp. 1 ~ 177.

Dongus, H.J. 1980. Die geomorphologischen Grundstrukturen der Erde. Stuttgart.

Finsterwalder, S. 1897. Der Vernagtferner. Wiss. Erg. H. z.Z. DÖAV. 1/1.

Furrer, G. 1965. Die Höhenlage von subnivalen Bodenformen. Habil., Pfäffikon, Pp. 1 ~ 78.

Furrer, G. & P. Fitze 1970. Die Hochgebirgsstufe - ihre Abgrenzung mit Hilfe der Solifluktionsstufe. GH, 25, Pp. 156 ~ 161.

Grunow, J. 1961. Die relative Globalstrahlung, eine Meßzahl der vergleichenden Strahlungsklimatologie. In: Wetter u. Leben 13, Pp. 47 ~ 56.

Garleff, K. 1970. Verbreitung und Vergesellschaftung rezenter Periglazialerscheinungen in Skandinavien. - Göttinger Geogr. Abh., 51, Pp. 1 ~ 51.

Hagedorn, J. 1969. Beiträge zur Quartärmorphologie griechischer Hochgebirge. Göttinger Geogr. Abh., 50.

Jessen, O. 1950. Fernwirkungen der Alpen. Mit. G. Mü., 35, Pp. 7 ~ 67.

Kelletat, D. 1969. Vergesellschaftung rezenter Periglazialerscheinungen im Apenin.- Göttinger Geogr. Abh., 48, Pp. 1 ~ 114.

Köppen, W. 1923. Die Klimate der Erde. Berlin.

Krebs, N. 1928. Die Ostalpen und das heutige Österreich. Stuttgart.

Kuhle, M. 1978b. Obergrenze von Strukturbodenerscheinungen. In: Z. f. Geomorph. NF. 22,Pp. 350 ~ 356.

Kuhle, M. 1982a. Der Dhaulagiri- und Annapurna-Himalaya. Ein Beitrag zur Geomorphologie extremer Hochgebirge. - Z. Geomorph., Suppl. Bd. 41, Berlin-Stuttgart, Bd. 1 u. 2: Pp.1 ~ 229, Abb. 1 ~ 184.

Kuhle, M. 1983. Zur Geomorphologie von S-Dickson Land (W-Spitzbergen) mit Schwerpunkt auf der quartären Vergletscherungsgeschichte. - Polarforschung **53** (1): Pp.31 ~ 57.

Kuhle, M. 1985a. Glaciation Research in the Himalayas: A New Ice Age Theory. In: Universitas **27** (4): Pp. 281 ~ 294.

Kuhle, M. 1985b. Permafrost and Periglacial Indicators on theTibetan Plateau from the Himalaya Mountains in the south tothe Quilian Shan in the north (28-40°N). In: Z.f. Geomorph. N. F., 29, 2, Berlin-Stuttgart,Pp. 183 ~ 192.

Kuhle, M. 1986a. Die Obergrenze der Gletscherhöhenstufe - Oberflächentemperaturen und Vergletscherung der Himalayaflanken von 5000 ~ 8800 m. In: Z. f. Gletschk. Bd. 22, H. 2, Pp. 149 ~ 162.

Kuhle, M. 1986b. Die Vergletscherung Tibets und die Entstehung von Eiszeiten. In: Spektrum d. Wiss, Sept. 86, Pp. 42 ~ 54.

Kuhle, M. 1987. Glacial, nival and periglacial environments in Northeastern Qinghai-Xizang Plateau. -Journal of Glaciology/Academia Sinica, Lanzhou-lnst, Beijing, Pp. 176 ~ 244.

Kuhle, M. 1988a. Die eiszeitliche Vergletscherung W-Tibets zwischen Karakorum und Tarim-Becken und ihr Einfluß auf die globale Energiebilanz. In: Geographische Zeitschrift **76** (3): Pp. 135 ~ 148.

Kuhle, M. 1988b. Topography as a Fundamental Element of Glacial Systems. A New Approach to ELA-Calculation and Typological Classification of Paleo- and Recent Glaciation. In: GeoJournal 17 (4) Kluwer, Dordrecht/ Boston/ London: Pp. 545 ~ 568.

Kuhle, M. 1990. The Cold Deserts of High Asia (Tibet and Contiguous Mountains). GeoJournal 20 (3), Pp. 319 ~ 323.

Kuhle, M. 1994. Present and Pleistocene Glaciation on the North-Western Margin of Tibet between the Karakorum Main Ridge and the Tarim Basin Supporting the Evidence of a Pleistocene Inland Glaciation in Tibet. GeoJournal 33 (2/3), Pp. 133 ~ 272.

Lagally, M. 1932. Zur Thermodynamik der Gletscher. Z. f. Glkde Bd. 20, Pp. 199 ~ 214.

Lauer, W. 1979. Die hypsometrische Asymmetrie der Päramo-Höhenstufe in den nördlichen Anden. Innsbrucker Geogr. Studien, 5, Pp. 115 ~ 130.

Lautensach, H. 1952. Der geographische Formenwandel. Studien zur Landschaftssystematik. Coll. Geogr., 3 Bonn.

Louis, H. 1963. Über Sockelfläche und Hüllfläche des Reliefs. Z. Geomorph. NF. 7, Pp. 355 ~ 366.

Maull, O. 1958. Handbuch der Geomorphologie. 2. Auflage, Wien.

Miotke, F.-D. 1981. Formungsprozesse in den eisfreien Gebieten des Victorialandes, Antarktis. In: Würzburger Geogr. Arb., 53, Pp. 195 ~ 207.

Pillewitzer, W. 1958. Neue Erkenntnisse über die Blockbewegung der Gletscher. IASH, Publ. Nr. 46,Pp. 429 ~ 436.

Schweinfurth, U. 1957. Die horizontale und vertikale Verbreitung der Vegetation im Himalaya. Bonner Geogr. Abhandl., 20.

Troll, C. 1944. Strukturböden, Solifluktion und Frostklimate der Erde. Geol. Rundschau, 34, Pp. 545 ~ 694.

Troll, C. 1955. Über das Wesen der Hochgebirgsnatur. JDAV, 80, Pp. 142 ~ 157.

Visser, Ph.C., 1934. Benennung von Vergletscherungstypen. Z.f. Glkde, Bd. 21, Pp. 137 ~ 139.

Zho Youwu &Guo Dongyin. 1982. Principal Characteristics of Permafrost in China. In: J. Glaciology and Cryopedology 4,1, Lanzhou, Pp. 1 ~ 19.