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The Cold Deserts of High Asia (Tibet and Contiguous Mountains)

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Topographic Characteristics

The block diagram (Fig 1) shows the area; situated between 67° and 105° E and 27° and 44° N, it extends over more than 3.4×10^6 km^2 and attains an average altitude of 4000 m, with Central Tibet rising to 5200 m. The surrounding mountain ranges exceed 7000 m, the crest of the Himalaya Karakorum exceeds 8000 m in 14 different places. Four mountains rise above 8500 m. The most northerly 7000 m-peak is at 43° N, the southernmost is 1750 km further south at 27°42' N. The glacier area is 120×10^3 km^2, i.e. eleven times larger than that of all the European glaciers together, and one fifth more than those of the two Americas combined.

Climatic Characteristics

The core area is subtropical and highly continental, with transitions to maritime climate on the S edge. Whilst 0 to 2 humid months characterize the north as arid, the centre has 3 to 5, the south 6 to 9, and the south east even has 10 to 12 humid months. Moreover, High Asia has a tundra climate (ET), with pockets of permafrost climate (EF). The vegetation formation is to be classified as high mountain arid steppe. Along the 3000-km curve of the Karakorum Himalaya the inner-tropical convergence makes its widest bow to the north on earth and causes monsoon precipitation on the S slopes of the Himalaya in excess of 3500 mm per annum. It is one of the wettest areas, with more than 400 mm precipitation per month. Monsoon precipitation decreases towards the N edge of Tibet and the Tian Shan, as well as from E to W. In the Karakorum precipitation is recorded throughout the year as a result of monsoons and a west wind drift. The zone of maximum precipitation occurs at a greater altitude than in E Tibet and the Himalaya. In the Indus valley between Chilas and Gilgit (1000–1500 m asl) rainfall is 100–200 mm per annum. Above the line of equilibrium (4000–5000 m) accumulations in the firn areas are evidence of precipitation of 1000 mm and more per year (Kuhle 1988: 414). The expanded precipitation between March and August is essential. Though precipitation is reduced in the lee of the Karakorum, and the Shaksgam valley (4000 m) receives less than 100 mm, 1000 to 1490 mm per annum can also be estimated as falling at the ELA level (5400 m) (Ding, Y. 1987: 25). Whilst the climate on the windward side, with variations in the diurnal temperatures of 10–20° C (Srinagar) is to be regarded as temperate, the leeward side clearly manifests a cold desert climate with nocturnal temperatures dropping 20° C and daytime temperatures rising 5° C higher. Precipitation being one third less than that on the S slope, diurnal amplitudes at the Leh station rise to 30–40° C.

In respect of temperature amplitudes between windward and leeward conditions are comparable to those on the S edge of Tibet (between Api-Himal in the W and Khumbakarna-Himal in the E), though the greater differences in precipitation cause a humid adiabatic warming of the leeward airmasses by c. 5° C (föhn). As a result, the tree line in S Tibet, N of the Annapurna-Himal, rises to 4400 m, thus reaching its greatest altitude overall. On the windward side, the forest peters out at 3600 m asl, i.e. at a gradient of 0.6–0.7° C/100 m the 10.5° C July isotherm lies 800 m higher than the leeward side. In addition, the effect of the uplift of the Tibetan mass makes an impact. Evidence of this may be seen in the mean temperatures of Namche Bazar (3440 m; 6.6° C) on the Himalayan S slope towards Lhasa (3730 m; 9.8° C), to name but one example. Absolute precipitation, too, shows a meridional gradient from S to N Tibet and a latitudinally parallel one from E to W. Whilst reaching several 1000 mm per annum locations in the Himalaya, precipitation in the Kuenlun amounts to little more than 500 mm. There is a xeric lower tree line

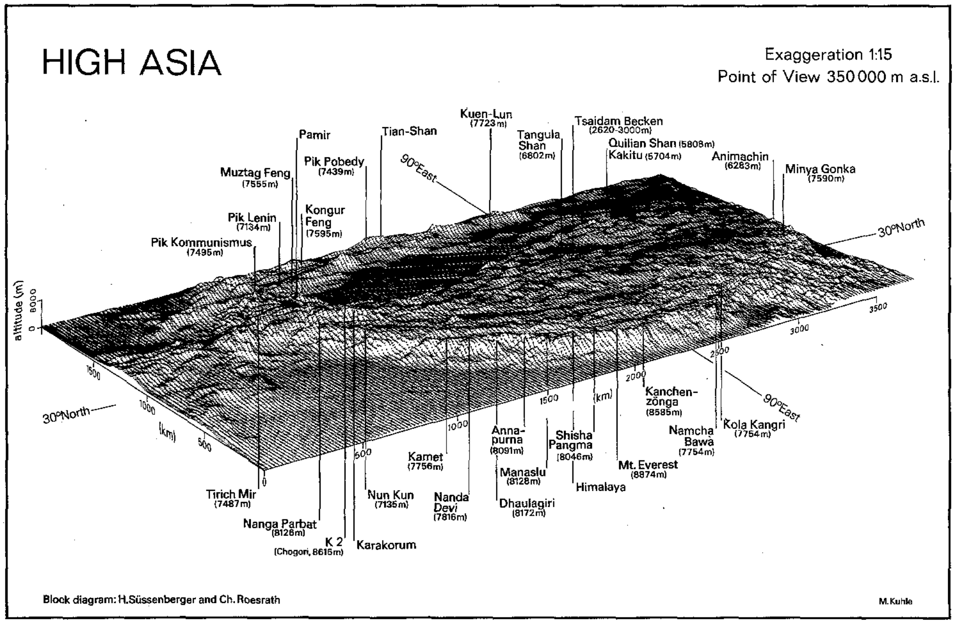

Fig 1
High Asia

in the Hindu Kush and the Karakorum; it is absent in the Bayan Har Shan because precipitation at comparable valley levels is higher. Precipitation decreases towards Central Tibet (peripheral-central), falling to 300 mm (Gyantse 3996 m; 271 mm per annum) towards the Tsangpo. With a mere 200–60 mm per annum, the region from central W Tibet towards the Aghil Mountains, extending as far as the N slope of the W Kuenlun, is driest. The leeward side central mountain areas and plateaux, as the regions of the cold deserts proper, are characterized by semi-arid precipitation and sub-tropically intensive radiation, which is further intensified by the transparency of the thin atmosphere at great altitudes. Such a subtropical highland climate is marked by large thermal day/night differences without an extreme annual amplitude (Leh station in Western Tibet). In monsoon-induced areas it is modified in summer, as shown by the annual amplitude of monthly averages of a mere 17° C at the Lhasa station. In the Central Plateau, the wintry high pressure climate, with its subtropical radiation, results in mild average values for January and caps even the outbreaks of low temperatures on the curve of areas with summer precipitation (average at Lhasa for January is only 0° C).

Glaciological Characteristics

These reflect the climatic characteristics mentioned above at a high level of integration of the line of equilibrium. The rise of the ELA in central W Tibet up to 6000 m reflects the peripheral-central decrease in precipitation. It includes the increase in cloud cover, which is proportional to precipitation, i.e. the decrease in radiation towards the mountain ranges on the fringe. The planetary ELA inclination from S to N is superimposed upon the central-peripheral change. The ELA drops from 5900 m in S Tibet to 4900 m N of the Tsaidam depression and to 4500 m in the N Quilian Shan. In the NW of High Asia, in the Tian Shan S and W of the Issykul, the equilibrium line even drops as far as 3800 m. In the latitudinal profile the influence of aridity and winter precipitation in the W and of monsoonal summer precipitation in the E on the ELA cancel one another out. Between 32° and 36° N the profile shows a similarity of the ELAs in W and E Tibet at about 5000 m asl. Only in the far W, where the mountain ranges fall away and the effect of the mass uplift decreases, does the ELA drop to 4800–4600 m. Though differing in its distribution and precipitation total, in terms of its ELA the W does not differ from the glacier areas in E Tibet because hygric conditions are compensated for by the thermal ones and by the distribution of precipitation. Beyond the region of this comparative analysis there are the findings in the W of the more than 70 km-long – and thus the largest outside the polar regions – valley glaciers in the Karakorum, where 37% of the mountain area is covered by glaciers. This difference from the less-glaciated mountains of E Tibet has its cause in the interference of climate and relief: in the W the mountains are higher and the catchment areas are thus larger; moreover, the glaciated valleys are flat. This results in extreme glacier length, because the tongue – in accordance with the altitude of the catchment area – must reach down far below the ELA in order to come to an end (Kuhle 1986a). The ELA is therefore to be regarded as the constituent climatic dimension of comparison. Assuming constant global radiation, it is possible to establish two climatic types of inverse valency of temperature and humidity factors by systematic comparison:

1) The semi-arid, cold glacier, which – depending on precipitation of 500–1300 mm/per annum, together with average temperatures of –6° to –12.3° C at ELA

level – must be regarded as the High Asiatic type proper. Examples are the Rongbuk glacier and the Yepokangara glacier in S Tibet, the K2 glacier and the Dunde glacier in W and N Tibet respectively (Ding 1987: 9, 10; Kuhle 1988: 414).

2) The humid, warm glacier; dependent upon monsoon precipitation, it occurs on one sixth of the area of High Asia at 91–99° E and 28–33° N in NE Tibet. The catchment area receives more than 1300 mm per annum, and in extreme cases more than 2500 mm per annum. The ELA-level temperatures are warmer than −6° to −4° C. Examples are the glaciers in the 7756 m-high Namche Bawar massif. Additional characteristics for 1) above are a high ELA in relation to temperature; for 2), on the other hand, a relatively low ELA, is the case. In type 1) ice formation predominantly takes place by way of infiltration and congelation, and in type 2) by warm infiltration and recrystallization. The ice temperatures in type 1) are low, down to the ends of the tongues, i.e. their temperatures fall below zero; in type 2) ice temperatures of around −0° C prevail from just under the ELA and further down to the glacier gates. Differing mass turnover results in a lesser rate of flow for type 1) and a higher mean rate of flow for type 2). In long valley glaciers of type 1) it amounts to 40–150 mm per annum but to more than 200 mm per annum in type 2). Higher rates of flow are reflected in the more volatile glacier fluctuations of humid-warm ice flows (2). Significant differences also occur in the ablation rates of the ends of long valley glaciers:

1. 1200–4000 mm per annum compared with 2): 10,000–14,000 mm per annum. Arid glaciers (1) support no life whatsoever, whereas humid ones (2) are inhabited by glacier fleas and algae up to the firn region, and further down by ice worms as well. A specific feature of High Asian cold deserts (cold-arid) is the feeding of Tibetan glaciers ending at high altitudes (such as the Yepokangara glacier (Shisha Pangma), which terminates at 5600 m). During winter and summer precipitation falls in its solid form even at the end of the tongue, since temperatures immediately drop below 0° C when clouds occur together with precipitation screen-off radiation.

However, in the Alps the temporary equilibrium line is so high in summer that rain falls above the ELA. Radiation weather, on the other hand, turns the entire surface of Tibetan ice flows from the tip of the tongue to the peaks far above the ELA into ablation areas which, thanks to evaporation, is more effective above than below. Independent of the seasons, but directed by weather conditions alone, this reversal from a process of accumulation to one of ablation of equal homogeneity, right across the entire vertical extent of the glacier, reflects radiation transparency as well as the lower heat capacity of the high altitude atmosphere. On glacier surfaces near sea-level, on the other hand, where radiation energy is reduced, advective heat transport and heat balance during the course of a sudden change of weather play a substantially modifying role, which in turn has an effect upon altitudinal fluctuations. A seasonally inverse consequence occurs in SE Tibet during the summer periods, i.e. the monsoon supply period of the glacier and its winter ablations (sublimations). In the Chinese literature (Glaciers in China, 1980) the extension of glaciers below the timber line is regarded as a feature of humid glaciers. This, however, applies only to larger glaciers with catchment areas extending far beyond the ELA – mostly for larger valley glaciers. Examples are ice flows in meridional flow furrows which descend to 2400 m asl. On the other hand many of the arid-cold glaciers of Nanga Parbat and the Karakorum also extend up to 1100 m below the timber line at 3700 m asl (Chongra, Bazhin, Minapin and Batura glaciers among others). The latter flows right through the entire tree level down to its xeroitic lower limit. Absence of tree growth in combination with glaciers confirms the aridity and low temperatures of the cold desert. In the literature mentioned above, no attention was paid to the climatically-independent relief altitude, which is the prime fact in determining the extent of the glacier drop. The Ghutumerung glacier with catchment levels of up to 5200 m, for example, only descends to 3800 m and does not reach the timber line, whereas the adjacent Pisan glacier (N flank of Rakaposhi with catchment height at 7700 m) with the same exposition flows down to 2600 m, traversing forest over a descent of 1100 m. This applies for the same ELA, i.e. the same climate here, thus indicating that the altitude of the tip of the tongue in its relation to the timber line does not permit climatic statements to be derived. It is therefore being suggested that one should consider the relationship of ELA to the timber line: in fact there is a greater convergence of the altitudinal lines to be found in the humid areas with cool summers, and a divergence towards the arid-cold ones. Nonetheless, both these boundaries are only conditionally equivalent, although glacier and forest react to summer temperatures in a similar fashion and are less sensitive towards those brought about by the winter. Thanks to its genetic variety of species, the forest is less homogeneous in itself than the simple physical system of glaciers, which knows no adaptation of species. However, by way of generalization it may be said in summing up that the ELA is further depressed by cool, humid summers than the timber line, whereas the latter, capable of adapting to aridity to only a limited degree, cannot follow the rise of ELA there. The coupling of the two boundary lines does not consist of the linear relationship between summer temperature as the absolute limit for tree growth, but in its relative extent for the formation of glaciers and annual mean temperature as the relative factor for the forest, and absolute limit for the glaciers. The adaptation of vegetation to humidity results in a stricter dependence on temperature, though not on mean, but rather on summer temperatures. Though this is contained in the annual mean temperature, it does not determine it. When summer temperatures and mean annual temperatures remain constant, the reaction of the ELA to variations in humidity is not modified by adaptation but is immediate. The forest touches upon the 10.5° C isotherm of the hottest month, unable to go beyond it even during mild winters and hygrically optimal conditions, whereas the ELA undergoes an almost unlimited thermal and xeroitic rise. In the aridity of High Asian cold deserts with their decreasing humidity above a level of maximum precipitation, the rise of the ELA runs way ahead of the rise in the tree line. Nor does the tree line receive a boost through being favoured by increased radiation, as it is correlated with nocturnal radiation and does not raise the mean temperature. By contrast, the monsoon humidity of summer is unable to depress the 10.5° C isoline, which restricts the forest, since humidity reduces incoming and outgoing radiation. The ELA is thus more variable than the tree line. The relationship is therefore to be described as that of a variable advance of the ELA to the tree line. The closer the ELA comes to the tree line, the more humid (i.e. cold desert-line) does the climate become.

Permafrost and Periglacial Boundaries

The periglacial altitudinal level has a bilateral basis (Kuhle 1978). It has an altitudinal zone of optimal formation, and becomes less distinct further down as well as further up until it reaches its lower or upper limit. The zone of optimal formation is determined by the permafrost line, where frost changes and thus, together with it, the sorting of scree material attain their greatest depth. This decrease in the size of structured soil and solifluction forms above the permafrost boundary has some value as an indicator for the demarcation of a cold desert. The interference of aridity and cold manifest themselves in it. In humid, oceanic, mountains like the Alps or the Scandinavian mountains, the glaciers have reached the zone of optimal formation of structured soils, and only a downward decrease in the intensity of formation can be observed.

The decrease in the size of forms above that optimal zone is no longer possible on surfaces covered by ice and firn. The humid climate is the reason for a low ELA, which runs little above the permafrost line. In cold-arid High Asia, however, when the ELA moves up high it leaves a broad and unglaciated band above the permafrost line free for decreasing scree-structuring. Where the depth of thawing decreasing to a very few decimetres, the decrease corresponds closely to the ELA and is within centimetres of level 365. Since diameters of structured soil are proportional to the depth of thawing (1:2 to 1:3), these forms are entirely absent up there. The climatic upper limit of structured soil has been reached. The wider the level of solifluction, the more arid are the mountains. In Tibet it has a vertical extent of 3000 m. A similar extent occurs in the semi-arid Andes at 32–33° S. The bilaterality of periglacial activities is superimposed on by a second, botanical, one (Kuhle 1987: Fig 2). In the arid Thak Khola (N slope of the Himalaya) the periglacial level sets in at about 3000 m asl at the belt of sporadic dwarf shrub vegetation with solifluction below a xeroitic lower tree line. At the altitude of about 3300 m vegetation is denser, resulting in more restrained solifluction. From 4250 m, at the level of condensation, alpine turf completely covers the scree; it is succeeded by a 500 m-wide band of a restricted solifluction level. Below vegetation becomes intermittent as a result of aridity, whereas thermal reasons are responsible for its loss of ground cover character above 4700 m. An upper storey of restrained solifluction thus follows on. The transition from the zone of optimal solifluction formation to the vegetation-free frost-scree level with unrestrained solifluction taking place at the permafrost line (5100 m). Tapering towards the top because of temperatures, and towards the bottom because of decreasing humidity, the double-wedge of receding vegetation (dwarf willow, *chenopodiacea, carex* spp., mosses, lichens) interferes with the characteristic forms of that bilateral periglacial level (see above). In the Central Highland areas (Tibet) with basic altitudes of more than 5000 m, only the upper diminuation of vegetation and solifluction can be developed; they culminate in the upper limit of plants and solifluction in the vicinity of level 365. Situated above the permafrost line, the areas exhibit the high polar permafrost indicators like pingos and ice-wedge networks. In contrast to the Alps, with their solifluction boundaries above the forest, periglacial terracettes in S Tibet (Dhaulagiri Himal) encroach upon avalanche tracts and clearings down to 650 m below the tree line. In the forest itself the forms are suppressed, as the soil is fixed by roots. The reason for the overlapping of the two boundaries is the cold, high pressure climate of the winter half-year, with its intensive inward and outward radiation and little or no snow cover. It leads to frequent (almost daily) frost changes in the top soil, which promote periglacial morphodynamics down to 3100 m. Summer precipitation provides for the necessary supply of humidity. The tree line is not affected by nocturnal winter cold but follows the 10.5° C summer isotherm up to 3700 m. In the Eastern Himalaya, with its winter precipitation peak, and an insulating snow-cover at around 3000 m, there is no solifluction below the tree line. The permafrost line, the course of which can be diagnosed on the basis of optimal forms of macro-solifluction, inclines planetarially from a distance of 1600 km from 5000 m in S Tibet to 3300 m in N Tibet (Quilian Shan) (Kuhle 1985, Fig 2). In W Tibet it runs at a higher level than in the E. Frequently falling as snow, summer precipitation in the high region of E Tibet prevents a comparable deep-soil thaw such as is found in the W with its higher incidence of radiation by way of cloud and snow albedo. In winter cold weather prevails in the E, as precipitation decreases from W to E so that the process of freezing through goes deeper than in the W, particularly so as a wintry snow cover has an insulating effect here.

The Most Extreme Cold Desert on Earth

The most extreme cold desert on earth towers above the altitudinal level of the glaciers: modified by the subtropics and the great altitude, there are characteristics of cold deserts at the glacier level which are familiar from Spitsbergen, Greenland and the Antarctic. Comparable cold aridity can be found in the arctic (NE Greenland), as well as in the marginal areas of the Antarctic ice. A High Asian speciality is the formation of a rocky region above the glacier level, occurring in the Himalaya above 7200–7400 m, and above 6900 m in the Karakorum mountains. This highest rock region is only matched by a corresponding feature in the Vinson massif (Antarctic, 75–80° S/60–85° W) at 3600 m or more. Evidence of regression analyses based on telemetric measurements of surface temperatures, which were carried out in the period 1982–1986 (Kuhle 1986b), has shown that even in summer and under conditions of maximum radiation temperatures up there do not attain, or come near, freezing point. They remain below −20° C to −25° C on average. The result is an extremely cold snow which does not allow settling or sintering processes to take place. Its metamorphosis and absorption into glacier ice is therefore not possible in the short term. Instead the snow is blown off the steep summits like dry blown-sand, and naked rock appears on the surface. Seasonally covered by snow, this rocky region above the upper glacier line is situated in an arid perglacial milieu and exposed to weathering purely by temperature in a minus range. Fluctuations between −6° C and −60° C destroy the rock merely by differing initial amounts of its mineral grains and entirely without water being involved. Corresponding conditions exist in large-scale zones on planets such as Mars.

Summary

1. The 3.4×10^6 km^2 area under consideration is situated in the subtropics; its mean altitude is 4000–5200 m, and its glacier area is 1.2 times larger than those of the two Americas combined.

2. In this area an ET and EF climate with 1 to 10 humid months prevails; monsoon precipitation reaches it from the S from the ITC. In the W precipitation falls throughout the year, in the E in summer only. Mass uplift and föhn effects cause relatively high temperatures.

3. The line of equilibrium (ELA) is evidence of the interference of planetarian and central-peripheral change arising from a gradient from 5900 m in the S to 4500 m in the N of Tibet and, parallel to the latitudes, an up-doming from 5000 m in the W and E to 5500 m in the central area. The smaller degree of glaciation in the E by comparison with that affecting the W, despite the equal ELA, is to be explained by higher catchment areas and shallower highland valleys. Taking the function of temperature and humidity as a basis, two types of glacier emerge: a) the semi-arid, cold glacier (−6° to −13° C at the ELA), with precipitation of 500–1300 mm per annum, a relatively high ELA, ice formation by congelation, low flow velocity and ablation rate; b) the humid-warm glacier, which occurs on the SE edge of High Asia in a region of 2500 mm precipitation, above −6° C at the ELA level – which is low; high flow velocity and ablation rate are further characteristics. Typical for cold deserts is the seasonally independent, i.e. purely weather-dependent feeding of the glacier and the ablation from the ridges to the tip of the tongue, or from there to its highest regions. This is a function of subtropical radiation, of transparency, and of the low heat capacity of the atmosphere at great altitudes. Moreover: the closer the ELA moves to the tree line, the more humid is the climate.

4. Another manifestation of a cold desert is a bilateral periglacial region whose intensity of forms decreases upwards from the line of permafrost to the upper limit of solifluction. This upper limit is characteristic insofar as aridity makes the glaciers become 'upwardly mobile' and exposes scree, which thaws less and less the higher it is – i.e. can be moulded by solifluction. In humid mountains this upper boundary is determined by glacier cover.

5. The most extreme cold desert is the periglacial rock level above the glaciers and in excess of 6800–7400 m asl. Due to too low a temperature, ice metamorphosis as well as glaciation are suspended. The rock surface is loosened through temperature weathering in the minus range.

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