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Ice Marginal Ramps: An Indicator of Semi-arid Piedmont Glaciations*

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ABSTRACT: Ice marginal ramps are depositional landforms that often developed over several tens of kilometres along the ice margins of large piedmont glaciers in semi-arid environments. The ramps extend with a 7–15° gradient several kilometres or even tens of kilometres into the forelands. They have front slopes that are tens of metres to several hundred metres high and frame the former terminal basins. The front slopes and the underlying bed consist of till; in the latter case, the till thins out towards the periphery. The overlying beds contain ice contact stratified drift, which, with increasing distance from the former ice margin, is succeeded by clearly sorted glaciofluvial layers of gravel. Under semi-ari d environmental conditions, the syngenetic contribution of *two* agents of transport (glacial, glaciofluvial) in forming *one* accumulation complex produces stratigraphic and phenotypical features that are rare in the glacial morphology of the temperate/humid zones. For this reason, they are often misinterpreted. Being indicators of ice margins, ice marginal ramps permit the accurate reconstruction of extensive piedmont glaciations in the semi-arid highlands of subtropical latitudes. Because of their high radiation values these play a key part in the global energy balance and, thus, in the origin and the evolution of ice ages.

Introduction

What we know about the location and extent of Pleistocene glaciations depends on the discovery and interpretation of their morphological relicts. The pacemakers of Quaternary research are its paradigmata (*sensu* T. Kuhn 1962), and when these paradigmata are established, applied, or rejected specific new perspectives are created in each respective case. Insofar as they channel scientific awareness, paradigmata have always both advanced and impeded the processes of cognition.

The historical controversy between the "iceberg theory" (put forward by leading scientists such as Saussure, Buckland, Lyell, Darwin, von Buch, von Humboldt) and the "glacier theory" (advanced by Kuhn, Hutton, Playfair, Deville, Venetz and others) was first settled with reference to the Alps. On the basis of Lyell's Principle of Uniformitarianism, this currently glaciated mountain region convincingly provided the essential foundations and arguments for the new paradigm that became established with the works of Charpentier (1841) and, in particular, Agassiz (1840). However, the vast areas covered by the Scandinavian ice sheet cannot be linked to present-day relict glaciers via a similarly convincing analogy, and here the iceberg theory was able to stand its ground longer, despite studies by Esmark (1827) and Bernhardi (1832). It was only the description of a recent inland ice sheet (Rink 1852) that helped to propagate the glacier theory here too (Ramsay 1855; Torell 1875; Dana 1863). The interpretation and definition of the characteristics of glacial forms was based on the glaciated regions of the Alps and Scandinavia/Sibiria/Canada (Heim, A. 1885; Hess, H. 1904; Hobbs, W. H. 1911; Drygalski, E. v. & Machatschek, F. 1942; Klebelsberg, R. v. 1948; Tricart & Cailleux 1953; Charlesworth, J. K. 1957; Flint, R. F. 1971). Epistemologically, this paradigm of glacial geology is an anticipatory scheme which channels subsequent research perspectives and, at the same time, delivers an interpretation of the findings obtained from this viewpoint (perceptional cycle after U. Neisser 1976). In this century the paradigm of alpine glacial geology began to be applied to subtropical and tropical mountainous regions, resulting at first in a picture of Pleistocene glacial expansion that was negligibly small in comparison with glaciation in northern and middle latitudes (CLIMAP 1981). Some contradictory evidence pointing to a depression of the equilibrium line of up

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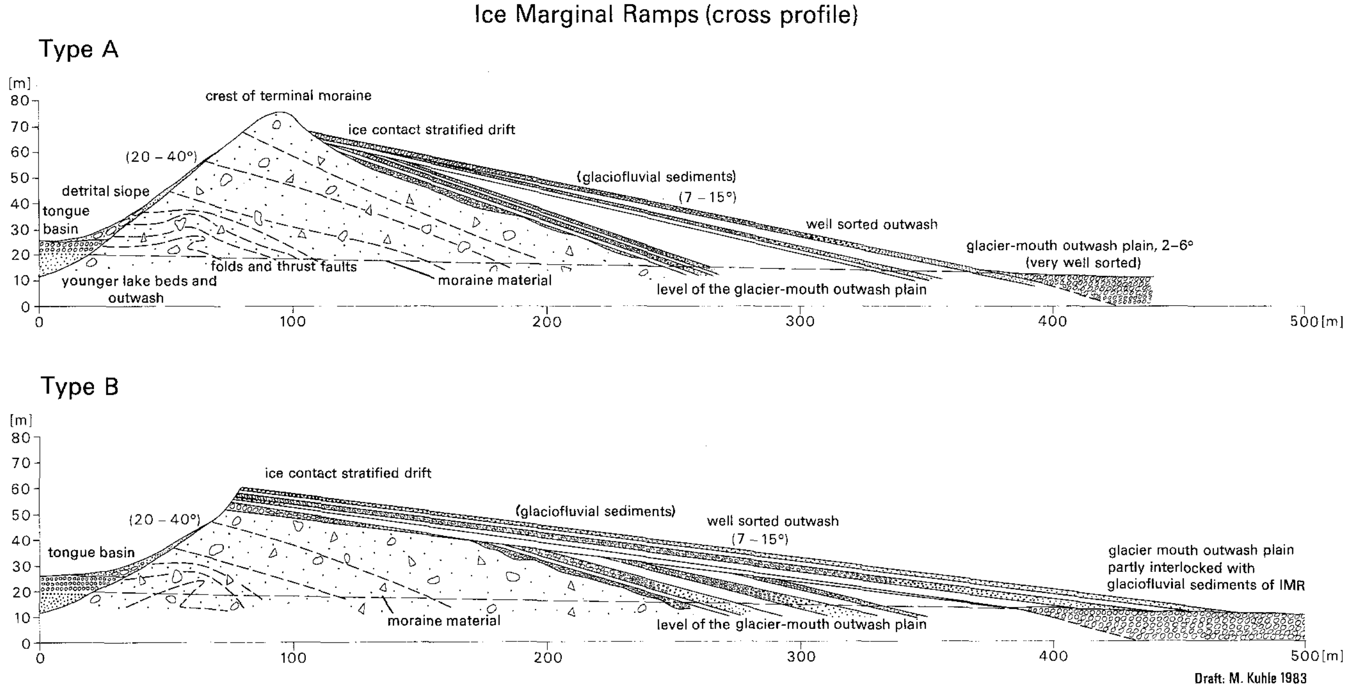


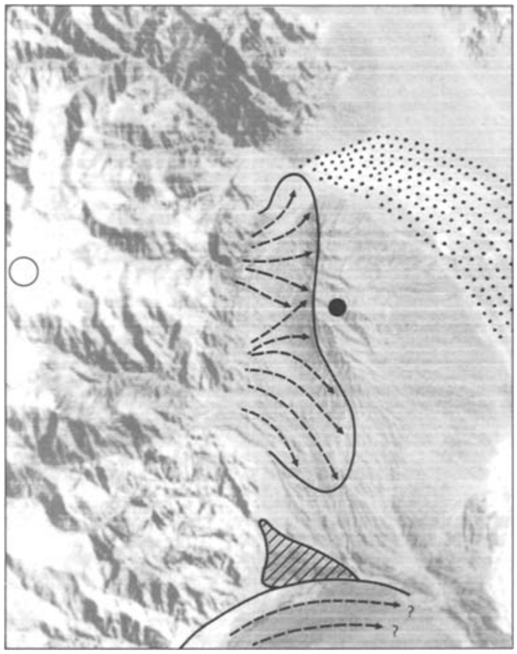
Fig 3 Types of ice marginal ramp representing 2 stages of development. Type B evolves from Type A and, being the last stage, inevitably occurs more frequently.

to 1000 m and more – see, for example, Brüggen (1929, Aconcagua Group, Andes), v. Klebelsberg (1922) and v. Ficker (1925, 1933, Pamir), v. Loczy (1893, E Qilian Shan and SE Tibet), Odell (1925, S Tibet), Norin (1932, W Kuen Lun), Weng & Lee (1946, N Qilian Shan) were either regarded with scepticism, disqualified as "local outliers", or classed as belonging to earlier ice ages (v. Wissman 1959). However, more recent comparative studies in the mountains of Iran, the Andes and High Asia have brought to light general depressions of the equilibrium line altitudes of 1000 to 1500 m for the last glacial stage (Würm). In the case of Tibet, the ice cover amounted to about 2.4 million km^2 (Kuhle 1985–87).

The application of the traditional paradigmata of glacial geology to extra-alpine, subtropical mountains necessarily led to erroneous interpretations. The different nature of landform structures and climatic conditions in these areas had led to a configuration of glacial landscapes that, according to alpine paradigms, had to be interpreted as fluvial or tectogenetic in character. Only the sub-recent and Lateglacial moraines at the termini of the present-day glaciers were generally accepted (they resemble the alpine type owing to their low equilibrium line depression); however, they were classed as High-glacial owing to the lack of evidence of lower ice margins.

Glaciated areas in subtropical/tropical latitudes are necessarily situated at high elevations. The 90% albedo from firn fields here relates to maximum radiation values, far higher (in High Tibet four times higher) than those attained at lower latitudes. Thus, glacier fluctuations at subtropical/tropical latitudes have a far greater effect on the global thermal balance. For this reason, it is of general climatological interest to reconstruct the maximum extent of Pleistocene glaciation in this region. In this connexion, attention should be paid to important ice margin indicators that are considered typical of extensive piedmont glaciers under semi-arid conditions. These deposits, referred to here as ice marginal ramps, comprise a complex of morainic and glaciofluvial elements; however, their external features have previously been considered untypical of both morainic and glaciofluvial deposits. If considered at all, these ice marginal ramps (IMRs) were misinterpreted as tectonically deformed or partly eroded alluvial or torrential fans. The first comprehensive analysis of IMRs was made with reference to the Pleistocene glaciation of Kuh-i-Jupar (Kuhle 1974, 1976), a massif in the Kuh-Rud Mountains, which runs parallel to the Zagros chains in the N (SE Iran). A further example is located in the Uspallata basin in the E of the Aconcagua massif (Argentinian Andes, Kuhle 1984b). The biggest known IMR site is the Qinhai-Xizang Plateau (NE Tibet) where 18 localities were investigated in detail (Kuhle 1982b, 1987a, 1987b). Lateglacial IMRs were found in the forelands of Shisha Pangma and Gang Benchen (S Tibet). Their proximity to present-day glaciers renders the genetic connexion immediately verifiable.

Fig 9 Satellite image (NASA ERTS E-2022-13452-701, Feb. 13, 1975) of the Cos. del Chacay E slope and the western Uspallata basin, where the Ice Age ice margin forked into two lobes and deposited an ice marginal ramp (Fig 8). At the bottom, the lateral or end moraine deposited from the S by the Mendoza glacier (Fig 10 and 11) is visible.



Ice Marginal Ramps of the Kuh-i-Jupar (29° 40'−30° 15' N/56° 50'−57° 35' E)

Two lengthy Pleistocene phases of glaciation were proved to have existed on the Kuh-i-Jupar massif, which rises to more than 4000 m (Kuhle 1974, 1976). The glaciers in question were valley glaciers whose ice masses flowed into the mountain foreland and coalesced to form wide piedmont ice lobes during the earlier glaciation. At the time of the later, less severe glaciations the glacier termini reaching the foreland were generally separated from each other by intervening rock spurs. The outlines of the respective terminal basins are marked by IMRs (Fig 1, 2, 4). These are frontal moraines that were reworked by meltwater from the former glacier surfaces and covered by glaciofluvial material (Fig 5). In the long profile a frontal wedge is visible; it consists of typical non-stratified moraine deposits: polymictic, angular, facetted and rounded boulders of the (here) erratic Jupar limestone (biomicrite), between 6×6×5 m and 2×1.5×1.2 m in size. The boulders are bedded separately from one another in a matrix of fine material comprising all fractions up to silt and clay (Fig 6). Ice contact stratified drift (Fig 5) either follows the culmination of the frontal moraine (Fig 3a) or else unconformably overlies it (Fig 3b; see also Kuhle 1976 Fig 31). In contrast to the moraine, a characteristic feature here is the clearly fluvial grading and bedding of the material. At first, the grain size range is still very wide and variations in grain size are numerous both horizontally and vertically. As the distance from the ice margin increases, typical outwash conditions set in with well-developed fluvial sorting and bedding and decreasing grain size more details concerning sediment fabric of IMRs and their distinction from alluvial fans are given in Kuhle 1988c). The ice marginal ramps are 500−1500 m long in the direction of deposition and their surface inclination is between 7 and 15° (usually 12°), thus distinguishing them clearly from outwash plains originating at the glacier mouths.

The IMRs outline the shapes of former terminal basins. In the case of the earlier piedmont glaciation these are long lines (Fig 1, 4) whereas the IMRs of the later glaciation were often built up in a wedge shape by still separate, tapering glacier tongues (Fig 2, 4). Fluvial dissection, which continued to be active after their formation, created a characteristic converging system of small floodplain valleys (Fig 2, 4, cf. also Fig 9 and Fig 14).

Two types of glaciofluvial cover of the frontal moraines may be distinguished: in type a) the glaciofluvial deposits start immediately after the culmination of the frontal moraine (Fig 3a), in type b) the moraine rampart and glaciofluvial material were truncated and unconformably covered by fresh deposits that outcrop in the direction of the terminal basin (Fig 3b, Fig 5). These internal structures show that the IMRs are the result of longterm, multiphased processes of accumulation reflecting the fluctuations of the ice margins. Frontal moraines and glaciofluvial deposits form a complex shown to be syngenetic by continuous transitions and interlocking. There are no convergent depositional forms associated with this complex. A necessary precondition here is a former glacier surface as a variable initial level for the deposits steeply (7−15°) covering the diamictite. Whilst it sometimes proves difficult to distinguish pure moraines from mudflow or landslide deposits, IMRs can be identified reliably owing to their more complex and distinct structural characteristics.

The glaciogenetic interpretation of IMRs is further confirmed by evidence of classical ice thrust features (cf. Flint 1971: 121–124). For example, 4 km away from the mountain foot, at 200 m asl, there are heavily compressed and folded lake sediments (Fig 7) exposed over the recent gravel floor. They cover a distance of 200 m and are up to 14 m thick. The overlying layers consist of up to 8 m of till, covered by 3 m of glaciofluvial material belonging to an IMR (Kuhle 1976: 90–97). The limnic sediments were formed in an interstadial terminal lake that was impounded by the IMRs of the earlier stage of glaciation.

In the mountain foreland, the older moraines in question are, in places, in contact with dislocated Neogenic sediments. Tectonic dislocations at the ice margins often lead to pushing and diversion of the now only thin ice lobes and form special "crystallization points" of moraine deposition. This resulted in IMRs that covered a solid bedrock core being misinterpreted as Neogenically dislocated sediments. However, it was possible to prove beyond doubt that IMRs were formed independently of tectonic dislocations:

- the undisturbed moraine material and the adjoining glaciofluvial deposits overlie extensively and conformably the erosion surfaces that cut through the Neogenic layers. This means that tectonic processes had ended at the time of glacial accumulation.
- the Neogenic dislocation zone and the IMRs belonging to the earlier glaciation are not congruent.
- the IMRs of the later glaciation with the same structural characteristics as those of the earlier one have no connexion with tectonic dislocations and overlie the old, glacially polished rock surfaces of the mountain base.

Further Indicators of an Ice Cover:

On the evidence of glacial deposits in Iran, a 2-phased mountain glaciation was reconstructed which had a length of 17 km in the direction of flow and a lobe width of 20 km in the foreland at the time of the earlier glaciation (probably Riss). The equilibrium line altitude during the older glaciation was estimated to have been some 1600 m below its present level (1500 m during the younger phase). In the mountains, the former catchment, this glaciation is recognizable over wide areas by landforms of glacial erosion – U-shaped valleys, overdeepened basins, roches moutonnées, glaciated valley flanks, transfluence cols (Fig 1) (Kuhle 1976: 127–179). Rock surfaces with still visible striae and glacial polishing are preserved at 2700 m (Kuhle 1976, Fig 72) and 2800 m (Kuhle 1976, Fig 10). Remains of lateral moraines with proven erratic boulders (Kerman conglomerates on underlying massive biomicrite limestone (Kuhle 1976, Fig 76) and upper limits of polishing at 2950 m (Kuhle 1976, Fig 147) prove that the ice was at least 500–550 m thick in the valleys.

Ice Marginal Ramps in the Uspallata Basin (Central Andes)

Highglacial glaciers flowing down from the 5000 m high Cos. del Chacay (adjoining the Aconcagua Massif in the NE) reached the Uspallata Basin and were diverted northward and southward by a steep, Tertiary rock bar (Fig 9). The bedrock here is covered by frontal moraines up to 420 m high. The polymictic, rounded and facetted boulders embedded in a fine matrix exclude an origin due to landslides. Furthermore, directly on the other side of the culmination there are typical, steep, glaciofluvial gravels that underwent postgenetic dissection by small V-shaped valleys (Fig 8). Their internal structure, external appearance and situation (at the margin of a terminal basin) correspond in all details to the IMRs occurring in Kuh-i-Jupar (type a; cf. Fig 2 and Fig 4). Given an average catchment altitude of 4700 m and an ice margin at 2060 m, the location of the ice age equilibrium line is calculated at 3400 m, corresponding to a depression of 1400 m with regard to the present equilibrium line (Kuhle 1984b).

Further Indicators of an Ice Cover

Immediately to the S of the Cos. del Chacay's eastern glacier tongue are the end moraines of the Highglacial Rio Mendoza glacier (Fig 9). The bases of the hilly moraines are located at 1870 m and they tower up to 190 m above the gravel bed of the Rio Mendoza. The diamictite consists of coarse subangular and rounded polymict boulders (Fig 10). The matrix is clayey to silty (max. grain size: 0.02–0.09). Decimetre-long scratches are preserved on top of the finegrained volcanic rocks (Fig 11). In contrast to the Cos. del Chacay E-glacier, which is only 15 km long, the Rio Mendoza glacier system attained maximum lengths of 112 km and thicknesses of at least 1020 m, its ice coming from the massifs of Co. Aconcagua (7021 m), Co. Juncal (6180 m), Nevado del Plomo (6120 m) and Co. Tupungato (about 6800 m). Proof of such an extensive glaciation was found at several locations (Kuhle 1984b): for example, Rio de las Cuevas, the upper reach of Rio Mendoza, in the valley profile at Cruz de Caña (32° 51' S/69° 50' W; valley floor: 2540 m). At 3450 m and 3560 m below the 3656 m-southern adjunct of Co. Cruz de Caña drift with polymictic striated rocks was found (Fig 12). The subangular and rounded boulders are as long as 0.6 m and document ice thicknesses of at least 910–1020 m. Till also occurs at 3460 m on the corresponding southern flank. It consisted of erratics of pale plutonite, 1.5×2×1.3 m in size, on top of the calcareous and gypseous marl (Fig 13). The shortest distance to corresponding bedrock upvalley is 7 km; between there is the over-900-m-deep valley of the Rio de las Cuevas.

Reconstruction of the equilibrium line of the Rio Mendoza glacier produced a level of 3400 m which would again mean a fall of 1400 m compared to the present-day line. This is clear confirmation of an ice margin at the foot of the Cos. del Chacay (see above).

Ice Marginal Ramps on the Tibetan Plateau (Qinghai-Xizang Plateau)

In a study area measuring 450×820 m in NE Tibet (38°–40° and 34° 50'–36° N/95°–102° E) Highglacial IMRs were investigated in detail at 18 locations in the forelands of the Kuen Lun and Qilian Mountains (Kuhle 1982b, 1987a, 1987b). At the southern edge of the Tibetan plateau there are Lateglacial IMRs immediately in front of the recent glacier on the N slope of Shisha Pangma (8046 m) (see below). Here, the IMRs in the vicinity of the settlement of Chaka (36° 47' N/99° 04' E) will serve as examples of Highglacial IMRs. Up to 200 m in height, with adjoining, 2 km long glaciofluvial deposits, the moraines stretch at a distance of 8–10 km over several tens of kilometres parallel to the mountain base (Fig 15). The moraines are located as far down as 3400 m and contain facetted, rounded and sometimes striated crystalline clasts. In the respective terminal basins clay-rich tills with few boulders are exposed in many places. The tills are located even lower in the forelands than the IMRs, down to 3150 m at least. At the till periphery, 500 m N of the Chaka settlement, a 150 m deep core was taken at 3170 m (Fig 16). The core shows a repeated alternation of till and glaciofluvial deposits; the deepest till was found at 3020 m. An ideal sequence of terminal basins, IMRs and glacier-mouth outwash plains also occurs at 35° 13' N/97° 50' E on the S slope of the Kuen Lun Mts. With a frontal height of more than 100 m, the IMRs stretch over 10 km down into the foreland, where they interlock with the glacier-mouth outwash plains (Fig 14).

One of the rare cases where IMRs occur immediately in front of the recent glaciers is found on the N slope of Shisha Pangma and Gang Benchen (Fig 17, 18, 19). These relict forms of Lateglacial glaciers attain maximum thicknesses of 500–600 m and extend as much as 16–18 km down into the foreland as far as 5050 m. The typical post-genetic dissection of the IMR surfaces is missing here (compare Fig 14 and Fig 18/19), since these Lateglacial landforms are still situated in a periglacial environment, where the movement of several metres thick solifluxion sheets over wide areas impedes the development of linear fluvial erosion.

On the basis of 43 established ice margins, the level of the equilibrium lines in NE Tibet during the last Ice Age may be fixed at about 3750 m (Kuhle 1987b). The depression of the equilibrium line altitude in relation to that today was thus 1100 m (Kuhle 1982b: 74).

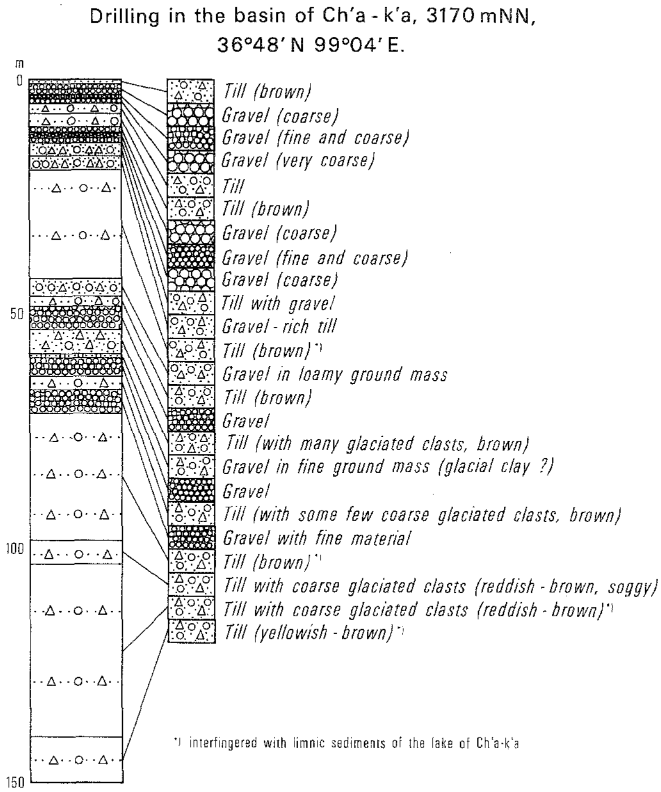


Fig 16 This 150 m deep core, like 7 other cores, was drilled through a maximum of 14 ground moraine layers up to 40 m thick in the former terminal basin S of Quinghai Nan Shan (location see Fig 15). There was a fivefold interstratification with ice advance gravels up to 7 m thick. This shows that during the Pleistocene in NE Tibet piedmont glaciers reached down to an altitude of about 3000 m fourteen times, corresponding to depressions of the equilibrium line of at least 1100–1300 m.

Further Indicators of an Ice Cover:

Assuming that, over wide areas, the equilibrium line altitude fell by 1100 m to about or below 4000 m – as has been reconstructed for the Highglacial period in NE Tibet – and given a mean height of 4500–5000 m for the Tibetan Plateau, an ice sheet must have formed over the entire region (Fig 20).

With respect to the S slope of the Himalayas, on the S edge of the Tibetan Plateau, ELA depressions of between 1100 and 1500 m are documented (Porter 1970; Heuberger 1974; Kuhle 1982a, 1982b, 1985). Glacially polished valley flanks and moraines at depths as low as 1100 m suggest that the Highglacial ELA depression may be reconstructed at 1400 m for the Dhaulagiri and Annapurna S-slope (Kuhle 1982, 1983, with detailed photographic documentation) and at least at 1200 m for the S slope of Mt. Everest (Heuberger 1974, 1986 estimates 900–1100 m). Recently the author has found strong evidence for an Indus valley glacier terminating as low as 980 m. TL-datings of tills and glaciolacustrine

Fig 23
Well-preserved glacial striae, with a ferro-manganese crust, on the right-hand flank of the Surukwat valley (Aghil valley, N of Aghil Pass, Yarkand valley network, 36° 20' N/ 76° 36' E) at 3710 m asl in quartzite rock. Photo: M. Kuhle, Oct. 29, 1986.

sediments immediately upvalley of this terminal moraine and further up in the basin of Chilas (at 1000–1200 m asl) carried out by Schroder and Saqid Khan (1988) indicate (in the context of the author's findings) that this glaciation existed at least up to 37–55 Ka BP. Porter (1970) calculated a Highglacial ELA depression of 1200 m further to the West (Swat Kohistan). The mountain ranges of the Tibetan Himalayas and Trans-Hi malaya, leeward of the main Himalayan ridge, also evidence an ELA depression of 1200–1600 m (Kuhle 1982a, 1982b, 1987b) despite their semi-arid climate (300 m of rainfall compared to 3000 mm and above on the S Himalayan slope).

With rainfall less than 100 mm the mountains of W Tibet and the adjoining Karakoram are of primary importance in proving the existence of inland ice in Tibet. Roches moutonnées and transfluence cols at 4600–5000 m with well-preserved glacial striae high above the valley floor of the Shaksgam valley, as well as glacial polishing very high up on the valley flanks indicate that the ice was at least 1200 m thick here. Glacial polishing with striae at 3500 m (Fig 23, 24, 25) in the Yarkand valley adjoining in the N, and series of moraines, over 30 km long and up to 700 m thick, in the northern Kuen Lun foreland are indicators of ice margins that in turn are evidence of ELA depressions of 1200–1300 m (Kuhle 1988a, 1988b).

However, as erratics on the Chalamba La show, the thickness of the ice cover on the central Tibetan Plateau (Fig 20) amounted to at least 1200 m, but more probably 2600 m (Kuhle 1985, 1987c).

In the context of this evidence, the observations by, for example, Norin (1932), Trinkler (1932), Odell (1925), von Loczy (1893), Tafel (1914), De Terra & Paterson (1939), and Dainelli (1922) gain confirmatory weight, whereas previously they had either been interpreted as "local outliers", regarded with scepticism, or dated back to a pre-Würm ice age (v. Wissman 1959).

Proof of a Tibetan inland ice under conditions of a 1000 to 1500 m lower ELA has been provided by numerous separate finds, comprising the entire canon of the classical glacial landforms. Thus, there is no alternative but to interpret the ice marginal ramps in NE Tibet as being of glacial origin.

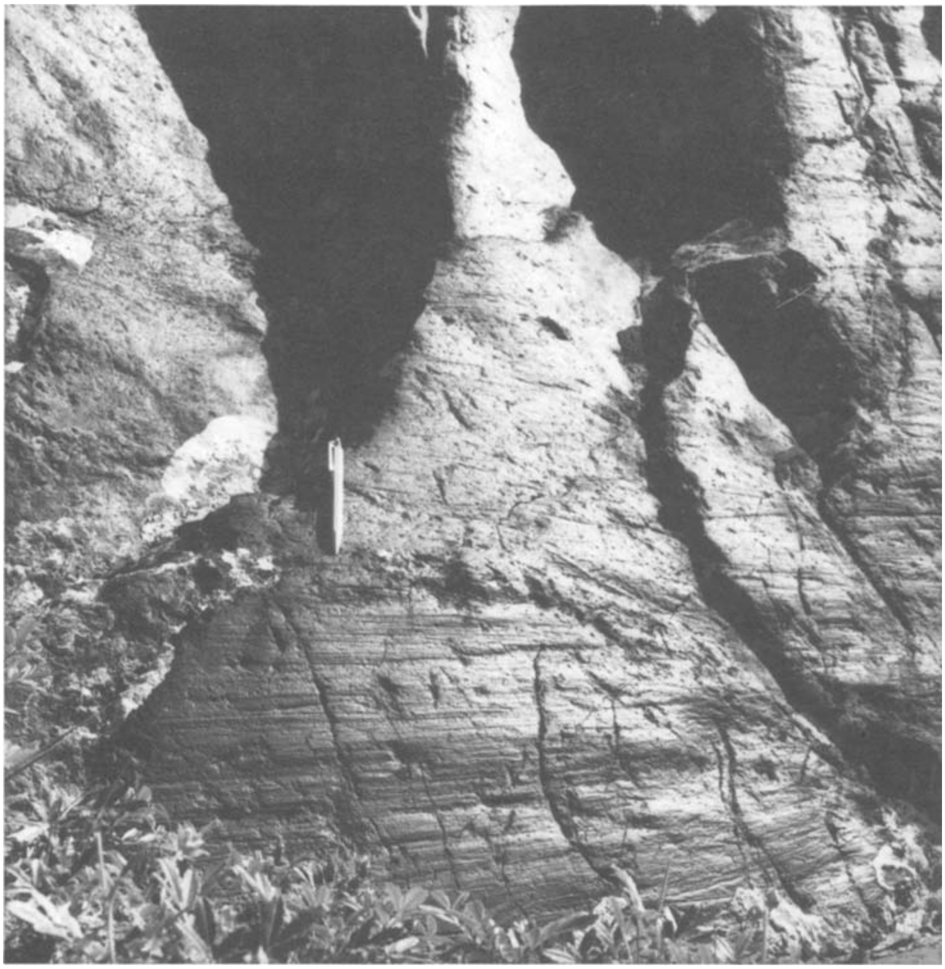


Fig 24
Well-preserved glacial striae, with a ferro-manganese crust, on the right-hand flank of the Surukwat valley (Aghil valley, N of Aghil Pass, Yarkand valley network, 36° 20' N/ 76° 36' E) at 3710 m asl in quartzite rock. Photo: M. Kuhle, Oct. 29, 1986.

Characteristics of Ice Marginal Ramp Formation

On the basis of the evidence to date, the following environmental characteristics of IMRs may be inferred:

a) IMRs are formed only on the margins of large piedmont glacier lobes; barriers, such as valley flanks, along the termini prevent their formation.
b) IMRs are formed only in association with large bodies of ice, whose reaction to short-term climatic fluctuations is slow enough for their ice margins to be sufficiently stable to build up deposits with the dimensions of IMRs.
c) These stationary glaciers react to climatic fluctuations with an increase or reduction in size of the lobe front; the resulting change in the angle of deposition in the tongue marginal area is expressed in the typical erosion and depositional stratigraphy of the IMRs.
d) Retreating glacier snouts not only decrease in thickness, their edges are also wasted and buried and the original lobe sharpens into a wedge-shape; this is often reflected in the preserved, cone-shaped configuration of the terminal basin (Fig 14).
e) The peripheral interlocking of the glaciofluvial deposits of the IMRs with those of the glacier-mouth outwash plains (Fig 14) points to a seasonal alternation between mainly supraglacial and subglacial meltwater flow. For the formation of IMRs, therefore, a location near to or above the permafrost line has to be postulated.
f) The situation of the IMRs in relation to the permafrost line also has a phenotypical significance: High-glacial IMRs now located far below the permafrost line have undergone an intensive fluvial dissection (their characteristically converging V-shaped valley system provides a useful criterion to distinguish them from the discharge pattern on dissected alluvial cones when interpreting satellite images), whilst Late- or Neoglacial landforms that are presently still situated within the permafrost zone have a level surface due to the movement of several metres thick sheets of creeping waste (cf. Fig 14 and Fig 15).
g) The IMRs known at the present time exist in semiarid environments; from this the conclusion may be drawn that, because of the relative aridity, glaciers could not have expanded into the temperate altitu-

Fig 25
Roche moutonnée at the confluence of the Surukwat and Yarkand valleys at 3500 m asl (36° 23' N/76° 41' E). The polish marks are still preserved on this part (foreground) of the some 200 m high landform. In the background, the slightly splintering strata edges of the metamorphic clay schist are already weathered. In this valley cross-section ice scour limits establish a minimum glacier thickness of 700 m. This was the path of the second northwestern Tibetan outlet glacier. Photo: M. Kuhle, Oct. 28, 1986.

dinal zone with year-round subglacial meltwater flow even at the time of the maximum, Highglacial fall in temperature; the glaciers remained within the permafrost limit and were subject to the seasonal alternation between subglacial and supraglacial meltwater debris release.

Results and Consequences

Ice marginal ramps are the result of a complex of factors in which climatic parameters (cold/semi-arid) together with spatial structures (extensive piedmont glaciation near the permafrost line) and temporal factors (persisting ice margins) attain a specific effectiveness. The basic pattern comprises foreland terminal basins which are delimited by the front slopes of ramp-shaped deposits with 7–15° gradients. Their long profiles show a continuous transition from clearly morainic deposits to clearly fluvially bedded and sorted gravels. Two agents (glacial/glaciofluvial) combine under specific climatic, spatial and temporal conditions to build up one complex of deposition. This leads to the emergence of novel phenotypical characteristics unknown to the classical glacial morphology of the temperate/wet zones.

Within the reference limits of this formation pattern each of the three originally independent dimensions possesses a variability that is individually expressed in the phenotype of the IMRs: dissected/planated surface; wedge-shaped/elongated pattern; percentual proportion of glaciofluvial to morainic substratum; thickness and extent etc. . .

Their magnitude, characteristic three-dimensional configuration and link with the sparsely vegetated semi-arid environment mean that IMRs are indicators of former ice margins that are also identifiable on satellite images (Fig 7, 14, 15). This could be a promising approach especially for research on inaccessible areas in High Asia or the Andes of South America. It should be emphasized, however, that IMRs have, up to now, been identified only by means of direct fieldwork. Identification on the basis of satellite imagery should be confirmed by ground checks.

Up to now, IMRs have proved to be characteristic landforms of semi-arid piedmont glaciations in subtropical latitudes. For this reason they should be accorded key importance in Ice Age research.

The reconstruction of the NE and N boundaries of a 2–2.4 million km^2 (Kuhle 1987c) sized ice sheet in Tibet was made possible by the identification of ice marginal ramps. Owing to its situation at a high altitude and a subtropical latitude, the Tibetan Plateau has an unusually high insolation, with values approximately similar to the solar constant. Whereas, today, 80% of solar radiation is absorbed by debris and rock surfaces, during the Highglacial as much as 90% of the energy was reflected back into space by the ice cover and was thus lost to the thermal balance of the atmosphere. The subtropical highlands of Tibet must have played a causal role in the onset of the Ice Ages: even when the initial drop in temperature was only a few degrees (e.g. owing to changes in the earth's orbit) the glaciers reached the plains and their largescale diffluence led to high atmospheric energy losses. The concomitant drop in temperature caused a worldwide depression of the ELA, enabling the previously steep glaciers with small-scale discharge to reach the plateaux in other subtropical mountains too. This resulted in a widescale expansion of the ice cover and a further drop in temperature, which finally caused the development of the Laurentide, Scandinavian and Siberian ice sheets (Kuhle 1987c).

Identifiable by satellite imagery interpretation combined with selective field work, ice marginal ramps provide the possibility of reconstructing the exact demarcation of the extensive Ice Age suptropical highland glaciation of Asia and South America and, thus, decisively advancing knowledge of the energy balance and development of the Ice Ages.

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Fig 1 Glacially moulded valley mouths on the Kuh-i-Jupar N slope (Zagros: 29–30° N/57° 10' E) opening into the foreland at 2300 m asl. The rocks have been polished by outlet glaciers up to 350 m above the valley floor (××). 5–10 km away from the mountain scarp there is a series of end moraines tens of kilometres long (▼▼▼▼). They consist of diamictites with a fine, lean matrix and isolated polymict boulders of chalky limestone and Kerman conglomerates. In the terminal basin, left, a glacially polished rock bar emerges at the surface (●). Photo: M. Kuhle, March 26, 1973.

Fig 2 Type B ice marginal ramps (see Fig 3) are evidence that Würmian piedmont glaciers reached down to 2100 m in the northern mountain base of Kuh-i-Jupar (Zagros/Iran: 29°–30° N/577° 10' E). On the left, these deposits outcrop in the formerly ice-filled terminal basin (×). The other terminal basin (right) is bounded by a ridge-like lateral moraine (●). Photo: M. Kuhle, June 18, 1973 (cf. Fig 4).

Fig 4 One of the Late Ice Age ice marginal ramps (▼) in the N Kuh-i-Jupar foreland. The ridge-like lateral moraine (×) surrounding the terminal basin on the right is a linear landform contrasting with the three-dimensional IMR. Its ridge of glaciofluvial deposits extends into the mountain foreland at right angles to the end moraine ridge and is evidence of a long period of ice margin stability. Photo: M. Kuhle, July 9, 1973.

Fig 5 Relict of a Type B ice marginal ramp (see Fig 3) in the Kuh-i-Jupar foreland (Zagros, SE Iran: 29–30° N/57° E) at 1950 m asl (middle background of Figs 2 and 4). This moraine, assigned by the author to a medial stage of the Riss glaciation (KUHLE 1974, 1976) overlies Neogenic sediments (×). The basal layer consists of morainic diamictites, composed of polymict boulders (light-coloured: limestone; dark: Kerman conglomerates). Towards the top, the diamictite merges into increasingly sorted ice-contact stratified drift. Photo: M. Kuhle, April 2, 1974.

Fig 6 Moraine exposure in the Kuh-i-Jupar foreland (SE Iran) with coarse polymict boulders (light-coloured: limestone; dark: Kerman conglomerate) in a dense till matrix. The viewpoint is located 10 m below the highest moraine ridge surrounding the terminal basin of Fig 2 (background, left). The base height of the inner moraine slope is 2240 m. Photo: M. Kuhle, June 8, 1973.

Fig 7 Pushed glaciolimnic sediments deformed by the overriding Riss-age piedmont glacier of Kuh-i-Jupar (SE Iran, 29–30° N/57° E). This compression at 2040 m asl is unconformably overlain by more or less stratified drift from a younger stadial moraine. Photo: M. Kuhle, March 17, 1974.

Fig 8 Glacial terminal basin (left, cf. Fig 10) with Type A ice marginal ramps (Fig 3) on the E slope of the semi-arid Aconcagua massif (Andes, 32° 34' S/69° 30' W/western Uspallata basin/elevation of glacial deposits 2210–2620 m). These end moraines consist of coarse polymict boulders (▼▼) and were deposited in the last Ice Age. Sloping at 10–12°, they grade into ridges of increasingly glaciofluvially stratified substrate on the right. Further into the mountain foreland (right-hand edge) these ice marginal ridges are overlain by glacier mouth outwash plains (●●●●). The ridges in the terminal basin (×××) at the foot of the frontal moraine slopes are remains of younger glacial and glaciofluvial deposits (outwash cones) of the Late Glacial. Photo: M. Kuhle, April 20, 1980.

Fig 10 Lateral and end moraines comprising polymict boulders with the entire petrographic range of the Rio Mendoza catchment at 1870 m base elevation where the Rio Mendoza valley joins the Uspallata Basin (cf. Fig 9). The moraines rise to a maximum of 190 m above the gravel bed of the Rio Mendoza and comprise well-preserved striated clasts (Fig 11), as well as chemically weathered and wind-corraded boulders. Length of ice-axe: 80 cm. Photo: M. Kuhle, April 22, 1980.

Fig 11 Striated rhyolithic clast (location: Fig 9) at 1980 m asl from the end moraine of the Mendoza glacier, which was at least 112.5 km long (length of the Plomo-Tupungato, the longest sub-stream). – Laboratory photo.

Fig 12 Striated clast at 3560 m asl (Aconcagua group, Andes, 32° 50' S/70° W) on the orogr. left-hand flank of the Rio de las Cuevas valley, 1020 m above the valley floor. This 40 cm long block consists of andesite and belongs to a polymict morainic mass covering calcareous marl. Such findings give evidence of minimum Ice Age glacier thicknesses of 1020 m (Kuhle 1984, Fig 13 is located in the same valley cross-section). Photo: M. Kuhle, Jan. 7, 1980.

Fig 13 Erratic boulder of light-coloured granite on marly talus which was transported at least 7 km down the main valley of Rio de las Cuevas and deposited at 3450 m asl here in the righthand subsidiary valley. This boulder and others up to 1.5×2×1.3 m in size are evidence that the ice was at least 900 m thick in the same valley profile as Fig 12. Photo: M. Kuhle, Jan. 10, 1980.

Fig 14 Ice marginal ramps on the Kuen-Lun S slope (35° 13' N/97° 50' E) mark the margin of a Lateglacial piedmont glacier. The position of the small IMR valleys at right angles to the frontal moraines is characteristic. Unlike relict alluvial fans, which are convex in the centre, these landforms have a concave cross-profile. There the small valleys converge into a central depression. Field investigations in 1981 established the morainic nature of the substratum (Kuhle 1987). Photo: NASA ERTS E-2691-03112-701, Dec. 13, 1976.

Fig 15 Ice marginal ramps in the southern foreland of Quinghai Nan Shan (in the background) (NE Tibet, N of the village of Chaka, 36° 47' N/99° 04' E, 3400 m asl). The southernmost spur of the ice marginal ramp (▼) banked against the 4651 m high mountain is still 8 km away from the solid bedrock. This IMR (here, view of its central depression) forms a wedge between two neighbouring terminal basins (●●). To the W there are adjacent terminal basins separated by further IMR wedges (background, left). These mountainward frontal moraine slopes of these ice marginal ramps (▽▽) consist of polymict boulders in a loamy matrix. Some of these boulders are very large (with cubatures of several m^3). Striated pebbles have also been found (Kuhle 1987b). Photo: M. Kuhle, July 12, 1981.

Fig 17 IMR long profile in the foreland of Shisha Pangma (8046 m), view westward from 5620 m (location see Fig 18 at the edge of the right-hand third of the photo). The Lateglacial glacier – the continuation of the 14 km long, recent Yepokangara glacier (left) – flowed past the IMRs, which fall away like oblique ramps (with a surface gradient of 7–12°) into the N foreland. Alternatively, it flowed through the IMRs in this morainic valley, either upthrusting or aggrading them. For this to take place, the ice must have been 500–600 m thick: since this is the height of the IMR surface above the floor of this valley. Younger lateral moraine ridges were deposited in Neoglacial and historical times and today still widen the lateral extent of this IMR complex. Photo: M. Kuhle, Sept. 14, 1984.

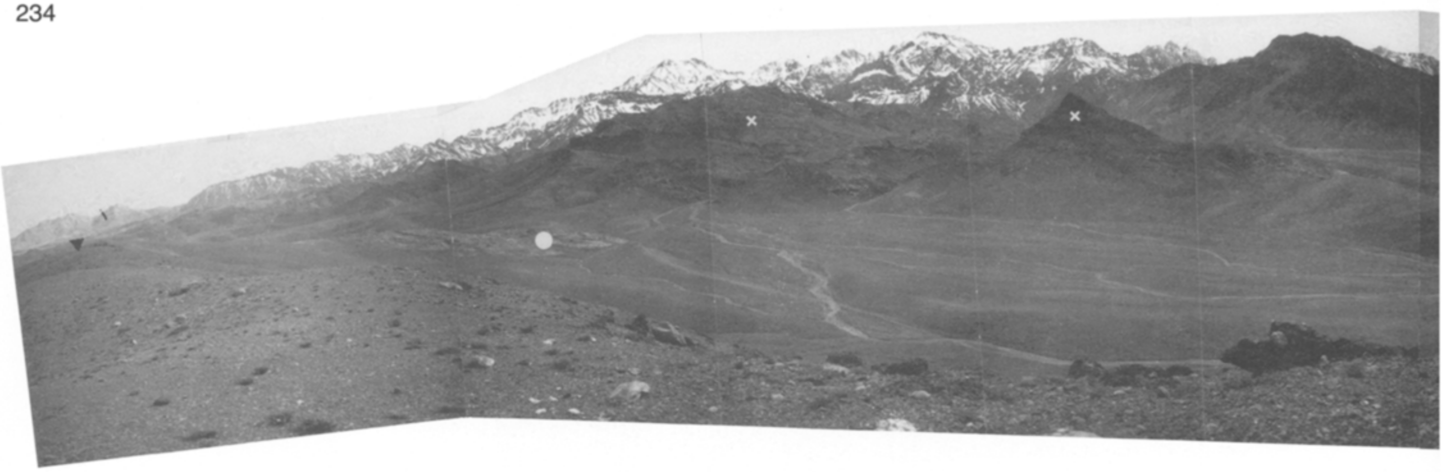

Fig 1

Fig 2

Fig 6

Fig 7

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Fig 10

Fig 11

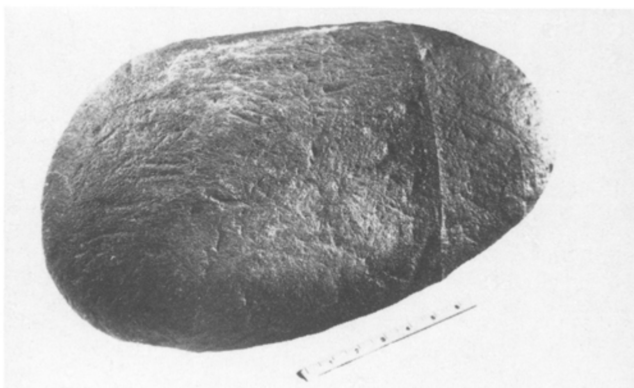


Fig 12

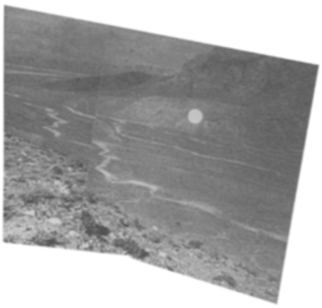
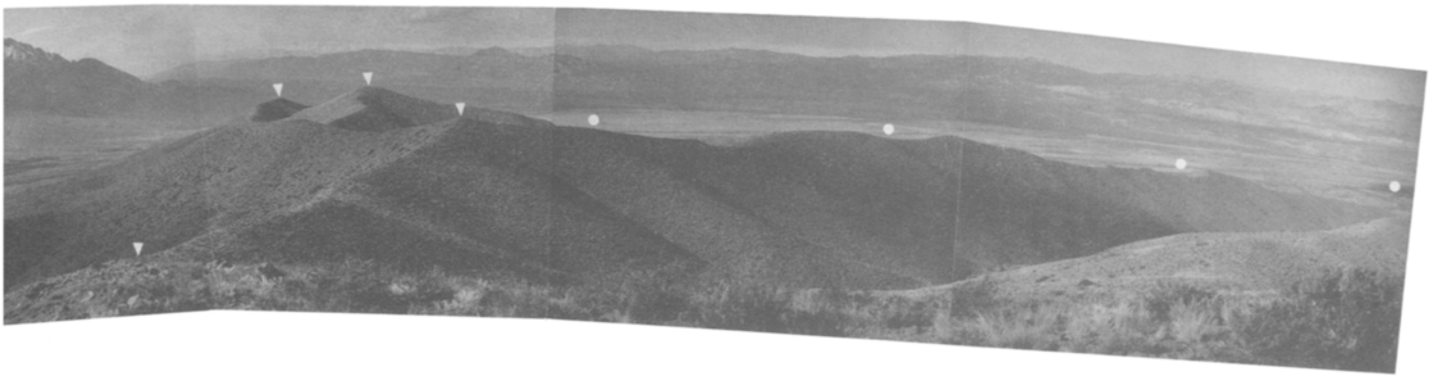
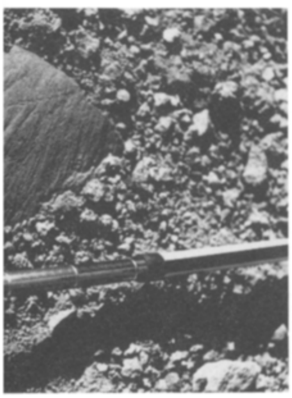


Fig 4

Fig 5





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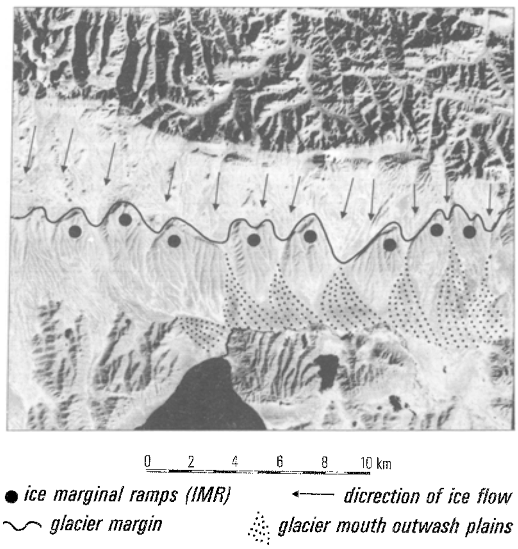

Fig 15

Fig 17

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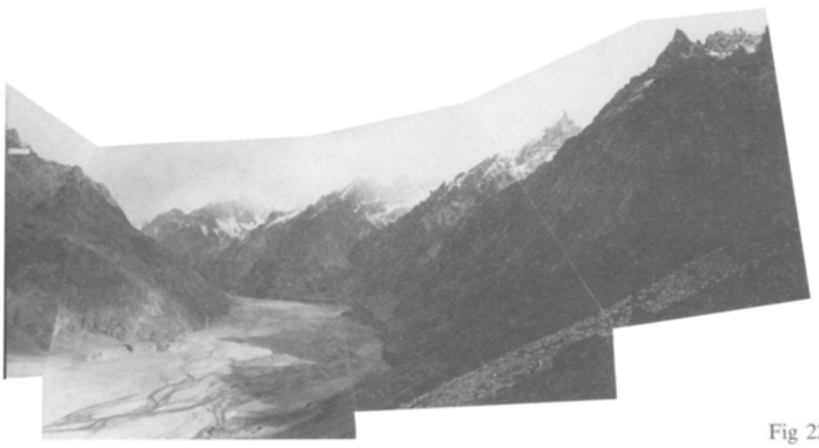

Fig 22

Fig 18

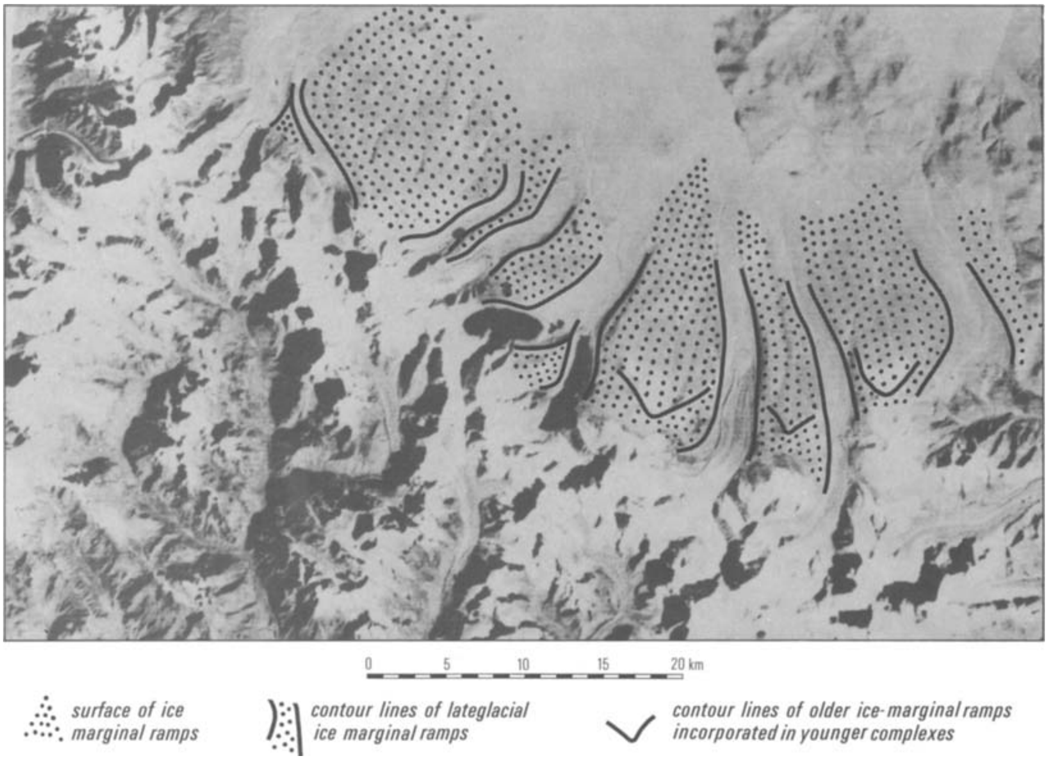

Fig 18 Lateglacial IMRs on the Shisha Pangma N slope (N side of the Himalayas between 8046 and 5020 m asl, 28° 25' N/85° 48' E). Glaciers still reach their base today (see Fig 17 and 19). Owing to Neoglacial and historical glacier oscillations, the original ground-plan form of the IMRs was distorted by moraine deposits. The till and gravel ramps (Fig 14) once stood isolated in the foreland and were subsequently directly joined to the mountain valley mouths by means of the lateral moraines deposited by subsequent, increasingly channelled glacier tongues. Photo: NASA ERTS E-2662-03542-702/ 14 Nov. 1976.

Fig 19 View from a Highglacially rounded roche moutonnée (foreground) in S Tibet at 5250 m down to Lateglacial to Neoglacial generations of IMRs in the N foreland of the Shisha Pangma and Langtang Himal group (the panorama includes the view shown in Fig 18). In the foreland of the left-hand peak, the 8046 m high Shisha Pangma, the recent Yepokangara glacier flows down into the moraine valley between the two IMR units and reaches the mountainward base of the ice marginal ramps (Fig 17). Towards the end of the Lateglacial the glacier tongue still flowed completely through the IMR valley, coming into contact with the IMR units at various levels, as the arcs of the frontal moraines show (▼▼▼). Earlier in the Lateglacial the IMRs – here in rows ten of kilometres long – were deposited by piedmont glacier lobes. The ice marginal ramps may be classed as piedmont, medial to longitudinal moraines. Fan-shaped outwash cones emerge from the IMR transverse valleys (cf. Fig 18). Photo: M. Kuhle, Sept. 16, 1984.

Fig 20 An example of extensive ground moraine transported over a long distance in the central Tibetan highlands. Till with a pelitic matrix and erratic mica-granite boulders in the Lulu valley (central S Tibet, 28° 50' N/87° 20' E, 4950 m asl). The valley is cut in basalt. There are very large, erratic granite blocks located up to several hundred metres above the valley floor (➛➛). On the right (●) the moraine is several tens of metres thick and hummock-shaped. Photo: M. Kuhle, Aug. 31, 1984.

Fig 21 The Shaksgam trough, the biggest longitudinal valley on the Karakoram N slope, viewed from below. During the last Ice Age the ice here was 1200 m thick. This cross-profile is located N of the 8616 m high K2 (Chogori) (36° 06' N/76° 29' E); the valley floor height here is about 3800 m. The lowest Highglacial ice level is documented by distinct scour limits on the calcite and dolomite flanks (– – – –, cf. Fig 22). This is one of the two big outlet glaciers that flowed NW out of W Tibet and were part of an ice stream network in this part of the mountains between the Aghil crest and Karakoram. Present-day rainfall at the valley floor is estimated at less than 50 mm/yr. Photo: M. Kuhle, Oct. 19, 1986.

Fig 22 Roches moutonnées on the transfluence pass between Shaksgam valley and the adjacent Muztagh valley to the S (4520 m asl, 36° 05' N/76° 28' E) in calcite bedrock. In the background, the Muztagh valley floor at 3900 m asl. The small-scale glacial rounding of the roche moutonnée profile is evidence of a quasi-laminate, warm ice dynamic near the compressional melting-point and so proves the existence of a thick, overlying glacier. Ice-scour marks show that it reached a further thickness of about 500 m, i.e. the valley glacier was here 1100 m thick. Boulder-sized mica-granites and dolomitic erratics were found on neighbouring roches moutonnées on this transfluence pass. Photo: M. Kuhle, Oct. 19, 1987.