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INSTITUTE OF GEOGRAPHY AND SPATIAL ORGANIZATION

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SPECIAL ISSUE No. 3

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KAZIMIERZ KRZEMIEŃ

HIGH-MOUNTAIN DENUDATIONAL SYSTEM OF THE POLISH TATRA MOUNTAINS

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Panu Dr. Mieczysławowi Klapie
z wyrazami szacunku i
serdeczności

Adam Kotarba
Kazimierz Krzemień
Ludek Kasz.

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INTRODUCTION

A significant intensification of geomorphological research into high mountain areas has been discernible during recent decades. Many geomorphological monographs providing reviews of mid-latitude Alpine geomorphology and Quaternary history and hillslope development have been published. The morphodynamic monographs contain summaries of the present-day evolution of mountains under the influence of currently active morphogenetic processes (Gardner, Smith and Desloges 1983; Caine 1983), while other papers deal with selected geomorphic processes within the context of the long-term evolution of landforms.

Present-day climatic conditions in the areas under investigation are considered within studies of currently active relief development. The resulting conclusions are organized within the conceptual framework of 'geomorphic thresholds'. Both, slow secular processes and short-term catastrophic events are analysed in a number of papers. Dynamic metastable equilibrium for given natural environmental conditions (*eg* Rapp 1974; Selby 1974; Starkel 1976) are determined for a variety of mountain areas. A different approach to the present-day evolution of mountain environments was presented by C. Troll (1973), P. Höllermann (1973) and other members of IGU Commission on High Altitude Geoecology who characterized the degradation of high altitude environments as being based on geographical rather than geomorphological criteria, *ie* altitudinal limits of soil– vegetation covers, timberline, snow line and landform complexes. This research strategy also includes spatial analysis of the actual mosaic of ecotope complexes within a framework in which landforms are a dominant element.

Present-day glaciated high mountain areas are the object of the greatest interest. This is evidenced, for example, by the problems discussed during the 25th Congress of the International Geographical Union in France, Paris– Alps 1984. One of the goals of that Congress was to identify the main factors, both natural and those related to human impact which are resulting in changes in the Alps. Numerous comparisons of Alpine regions and other mountain areas within the temperate zone have been made, and these include studies of the Tatra Mts.

Although in the Carpathian arc there are several high elevation mountain

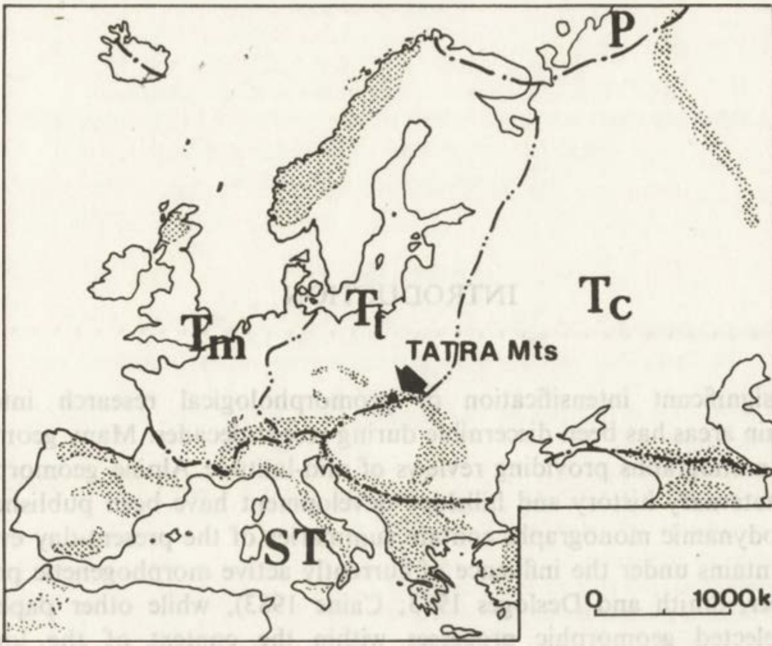

Fig. 1. Location of the Tatra Mts in terms of climatic zones
T_m temperate maritime. T_t temperate transitional. T_c temperate continental. P polar. ST subtropical

groups (Rodna Mts, Fagaras Mts, Retezat Mts in Roumania) the Tatra Mts are the highest mountain massif between the Alps and the Caucasus (Fig. 1). The Tatra Mts are characterized by a high local relief which was affected by Pleistocene glaciation (Fig. 2). The mountain massif is only 53 km long and 18 km wide (Radwańska-Paryska and Paryski 1973). High mountain glacial landscapes have been formed (Fig. 3) and a system of geoecological belts has been developed over this small area. These mountains can be regarded as a model of Alpine relief, since they provide the opportunity to study the relationship between elements of the natural environment, altitudinal differentiation of landforms and their present-day transformation by geomorphic processes.

. Research interest in the Tatra Mts commenced in the 19th century when S. Staszic (1815) carried out geological, geomorphological and climatic observations during an expedition. He noted an altitudinal, climatic and floristic differentiation and provided the first descriptive account concerning the operation of geomorphic processes. In 1843, only a few years after Agassiz's discoveries in the Alps, L. Zejszner (1856) discovered glacial forms in the Tatra Mts, since when the main interest of researchers has concentrated on glaciated landscapes. The limits of mountain glaciations were reconstructed, and problems concerning the number of glaciations, and the morphogenetic role of glacial and interglacial periods (among others Zejszner 1856; Partsch

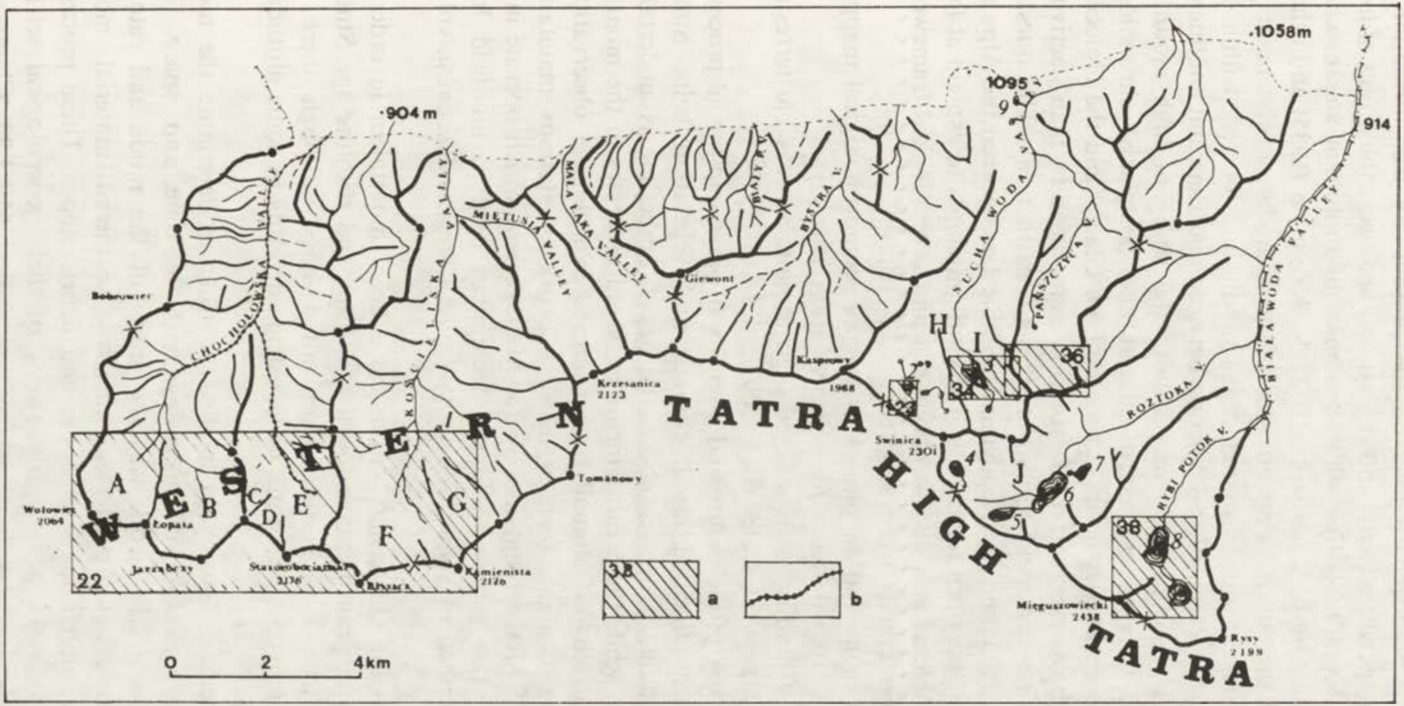

Fig. 2. Main features of the topography of the Polish Tatra Mts

The locations mentioned in the text: A — Wyżnia Chochołowska valley, B — Jarząbcza valley, C — Dudowy cirque, D — Ciełęce Tańce cirque, E — Starorobociański cirque, F — Pyszniański cirque, G — Dolinka cirque, H — Hala Gąsienicowa, I — Czarny Staw Gąsienicowy valley, J — Pięć Stawów Polskich valley; 1 — Smreczyński lake, 2 — Zielony and Litworowy lakes, 3 — Czarny Staw Gąsienicowy lake, 4 — Zadni Staw lake, 5 — Czarny Staw lake, 6 — Długi Staw lake, 7 — Przedni Staw lake, 8 — Morskie Oko lake, 9 — Toporowe Stawki lakes; a — geomorphological maps and figure numbers, b — investigated stream channels

1923; Romer 1929; Halicki 1930; Szaflarski 1937; Klimaszewski 1960; Lukniš 1973) were discussed. Studies on the present-day transformation of the Tatra slopes in specific climatic and vegetation zones were the object of investigations by A. Kotarba (1976) while dynamics of mountain stream channels has been studied by L. Kaszowski (1973), K. Krzemień (1985) and others.

The object of this study is an analysis of the denudation system of the Tatra Mts: at present non-glaciated high mountains, located within the temperate zone but in a transitional position between maritime and continental climates (Fig. 1). Unsolved problems include: the study of high mountain relief conditioned by geology and its glacial and postglacial chronology; evolution of slopes and valley floors during the Late Glacial and the Holocene, present-day morphogenesis, and relief evolution as modified by man's activities (the role of grazing livestock, mining and metal extraction). The question concerning the climatic and morphodynamic similarities between the Alps and Tatra Mts still remains open for debate. The study attempts to explain at least some of these problems in a geomorphological and geoecological framework.

The following methods were adopted in this study:

(i) analysis of maps and air photos as well as geomorphological mapping at 1 : 5,000 and 1 : 10,000 scales for selected areas (Fig. 2);

(ii) mapping of landforms in selected high mountain areas *via* terrestrial photogrametric surveys at scales of 1 : 500 and 1 : 1,000;

(iii) measurement of the nature and intensity of slope and fluvial processes within selected areas located in all climatic and vegetation belts. Simple measurement of slope processes above the upper timberline is undertaken 2–3 times a year, while the measurements of fluvial processes in the montane forest belt are carried out annually according to hydrological observations. Climatic data are available from a dense network of stations maintained by the Institute of Meteorology and Water Management which operate in all altitudinal belts. The geomorphological techniques used in the field have already been described in *Studia Geomorphologica Carpatho-Balcanica*, vol. 4, 1970:

(iv) studies on the Quaternary covers have been undertaken in order to determine their type, grain size composition, sorting, and absolute age. Studies on the age of deposits are still at an initial stage although there are radiocarbon and palynological dates of lacustrine deposits in some altitudinal belts (Kondracki 1984).

The above methods have been applied in order to determine the mode and rate of the operation of processes both in time and space. The primary objectives of this work was the study of the mode and rate of operation of present-day processes whereby weathered material moves downslope on Alpine cliffs, rocky slopes and debris slopes. These processes are examined within the general framework of their geoecological setting.

This study was undertaken as part of the project MR.I.25 'Changes in the natural environment of Poland' coordinated by the Institute of Geography

and Spatial Organization, Polish Academy of Sciences. The field studies were sponsored both by the Polish Academy of Sciences and by the Institute of Geography, Jagellonian University, Kraków. This monograph attempts to summarize the present state of knowledge concerning the dynamic geomorphology of the Tatra Mts dispersed throughout many published papers. It also provides an opportunity to incorporate new field data collected during the last five years.

We wish to thank Doc. dr hab. T. Niedźwiedź for the computer processing of the data related to solar radiation in the Tatra Mts. The figures have been drawn by Mrs M. Klimek. We are grateful to Dr M. Kłapa, Mrs Z. Rączkowska and Mr M. Kot for their assistance in the field. To Dr Alan Werritty we would like to express our appreciation and sincerely thanks for English correction on the manuscript.

Adam Kotarba

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POSITION OF THE TATRA MTS IN THE HIGH-MOUNTAIN SYSTEMS OF THE TEMPERATE ZONE

Large areas within Europe and Asia at the present time located in the temperate climatic zone display major mountain ranges uplifted during the Alpine orogenesis. These mountain ranges within which are outcrops of older crystalline materials are fringed by broad subsidence basins. Within the Carpathian arc (more than 1,300 km in length) mountain ranges are substantially lower in term of absolute relief and narrower than the Alps. Only the highest crystalline massif of the Carpathians enveloped mainly by limestone/dolomitic sedimentary structures exceed 2000 m. These ranges were subjected to glacial and periglacial morphogenesis during the Quaternary. As a result of the morphogenetic activity of glaciers, the preglacial and thus fluvial-denudational forms of valleys and ridges were transformed into high mountain landscapes with Alpine relief elements. Thus a system of rocky crests, rockwalls, glacial cirques and troughs has been formed, while valley bottoms have been infilled with a mantle of glacial and glaciofluvial covers. At the present time only the highest summits are above the snow line (Fig. 3). The upper timberline defines the lowest extent of areas regarded as high mountain areas (Troll 1972, 1973) in which periglacial processes are dominant (active cryonival denudation). C. Troll considers location above the three following geoecological limits:

(i) the Pleistocene snow line controlling the glacial relief,
(ii) the lower limit of active periglacial (cryonival) phenomena,
(iii) the upper timberline

as a criterion to distinguish the high mountain area. Frequently, these limits form an altitudinal belt above which the high mountains rise. In contrast there are many areas located at their foot which are called the middle mountains. In the Tatra Mts we observe these limits. There is not the nival belt, but all other geoecological features typical of high mountains of temperate zone occur (Fig. 4).

The highest part of the Tatra Mts which form the crystalline core rise to 2663 m in Czechoslovakia and up to 2500 m in Poland. The Western Tatra are built of metamorphic rocks and their absolute altitudes do not exceed 2250 m. The northern slopes of the mountains located in the

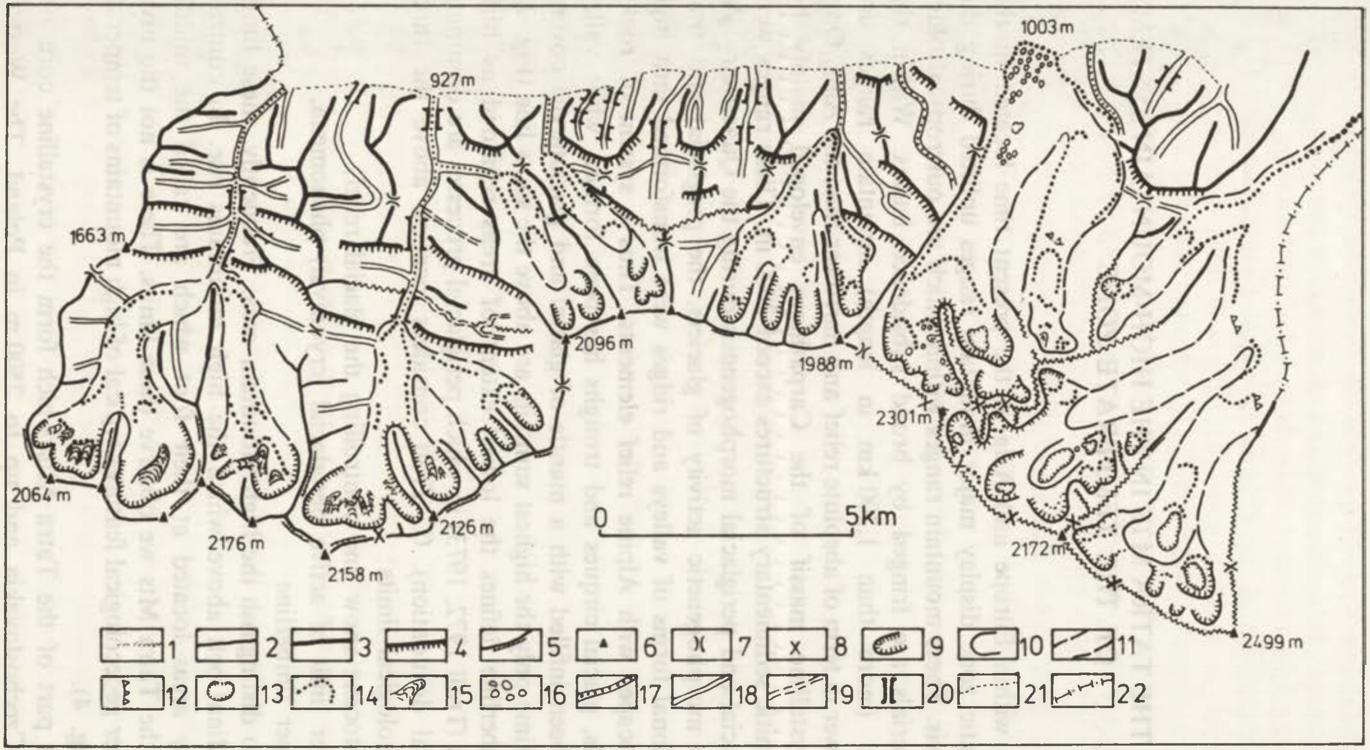


Fig. 3. Geomorphological sketch-map of the Polish Tatra Mts (after M. Klimaszewski 1962, modified)

1 – sharp rocky ridge crest. 2 – narrow rounded ridge crest. 3 – wide rounded ridge crest. 4 – cuesta-like ridge crest. 5 – double ridge. 6 – summit. 7 – pass related to less resistant rocks. 8 – pass related to dislocation zone. 9 – glacial cirque. 10 – trough head. 11 – glacial trough. 12 – valley step. 13 – lake basin of glacial origin. 14 – moraine ridges. 15 – fossil rock glaciers. 16 – dead-ice hollows. 17 – valley floor filled with glaciofluvial deposits. 18 – fluvial (non-glaciated) valleys. 19 – Karst valleys. 20 – structural narrowings of valleys. 21 – northern margin of the Tatra Mts. 22 – state boundary

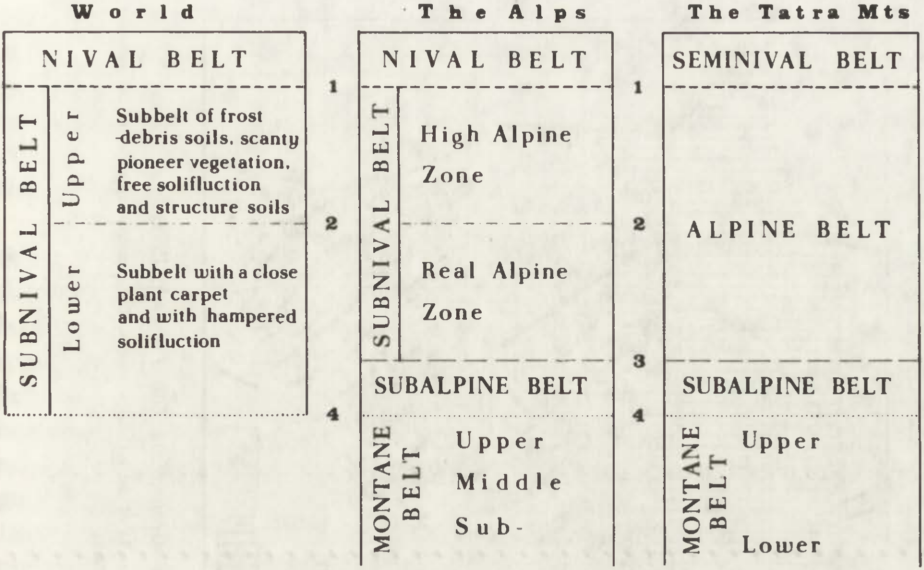


Fig. 4. Tatra Mts in terms of the world geoecological belts (after Löve 1970; Troll 1972, 1973)
1 climatic snow line, 2 – limit of a close soil and plant carpet, 3 – upper tree line, 4 – upper timberline

montane forest zone (up to 1500 m with a typical fluvio-denudational middle mountain relief) are built of Mesosoic sedimentary rocks among which dolomites, limestones, shales and marls are dominant (Fig. 5). The surrounding basins formed in flysch are located at a height of 600–900 m.

In contrast to the high Alps a seminival belt occurs in the Tatra Mts (Hess 1965; Klimaszewski 1961) instead of a nival one. This belt is the area where the mean air temperature of the warmest month (July) is lower than 6° C and the mean annual air temperature falls below –2° C. In addition the number of days with snow cover exceeds 250 *per* year. These factors mean that the climatic snow line is found at a height of 2200 m on the northern slopes of the Tatra Mts whilst on the southern slopes it is at a height of 2350 m (Hess 1965). The summit system rise above this line. It is characterized by narrow to knife-edged crests, and vertical or/and steep rockwall and rocky slopes, which do not favour snow accumulation and subsequent transformation into glacier ice. Thus, the summit area exhibits features which are typical of high, non-glaciated Alpine mountains classified by M. Chardon (1984) as the high rocky mountains (*haute montagne rocheuse et totalment deglacee*). This is the case despite the fact that the climatic conditions for glacier formation exist in the summit area. According to H. Karrasch (1977) who has summarized the data dealing with the altitudinal zoning of active periglacial phenomena in the mid-latitudinal mountains of Europe, the climatic snow line occurs in the High Tatra Mts at a height of 2600 m. Only the highest summits are located above that limit. However, the presence of

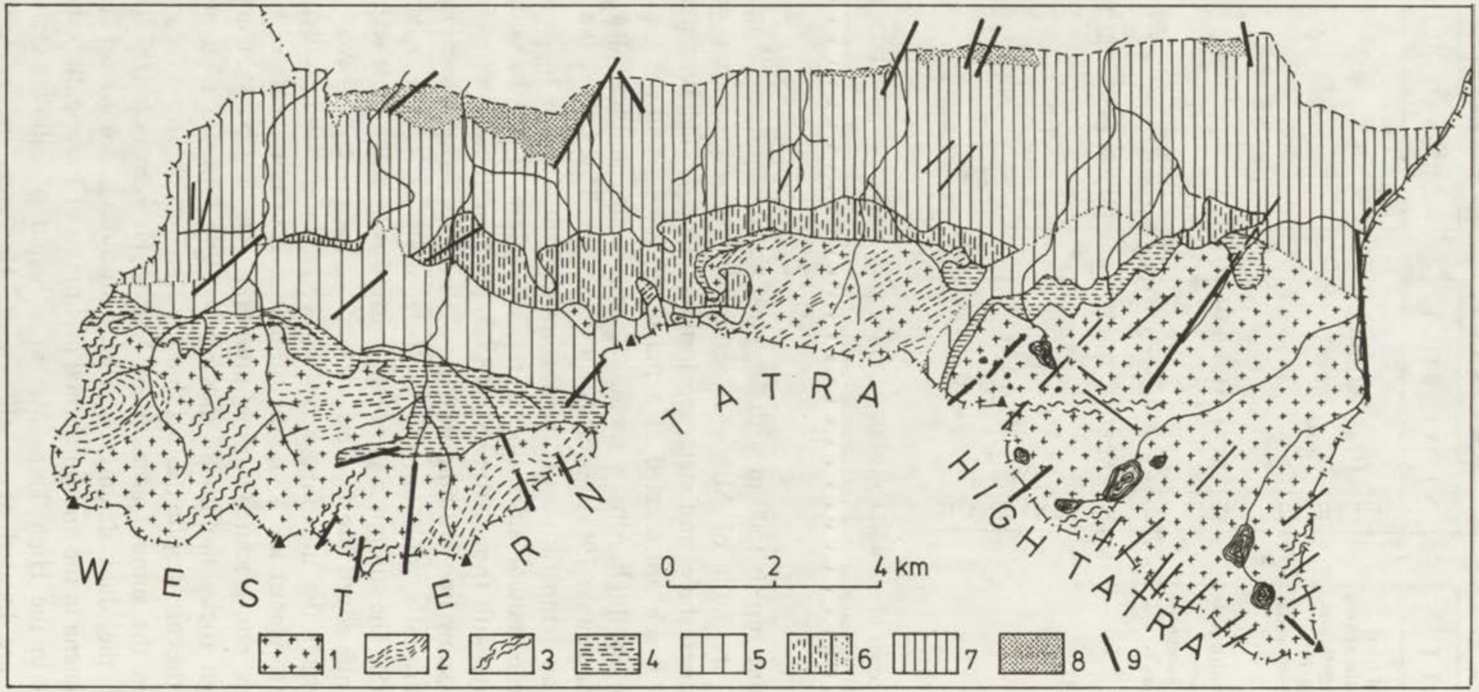


Fig. 5. Geological sketch-map of the Polish Tatra Mts (after *Mapa Geologiczna Tatr Polskich*, 1979)

1 – granitoids, 2 – gneisses, 3 – mylonites, 4 and 5 – sedimentary cover of the crystalline core, 6 – sedimentary rocks of the high-tatric nappes, 7 – sedimentary rocks of the sub-tatric nappes, 8 – Paleogene sediments, 9 – important faults

perennial snow patches (Vitásek 1956; Wdowiak 1961) indicates that already at a height of 2200 m climatic conditions exist which favour the formation of glaciers. Nevertheless, the snow patches are not accompanied by a morphogenetic system which is characteristic of the glaciated and snow-capped high Alps (*haute montagne alpine englacee et enneige*). Process systems typical of the so called high Alpine zone with subnival belt are also lacking.

In the Tatra Mts, as in other high mountains of the Alpine system, there is an altitudinal differentiation of present-day cryonival processes above the upper timberline. Miniature sorted polygons are dominant in the seminival belt and partly in the Alpine one (up to the height of 2000 m). Turf and stone-banked terraces and solifluction lobes are characteristic of the Alpine belt, while in the sub-Alpine belt regelation processes in soils are revealed by bound solifluction and earth hummocks. According to A. Jahn (1970) it is possible to distinguish altitudinal belts for the occurrence of: sorted pattern grounds (above 2000 m), solifluction (1800 – 2000 m), and thufurs (1500 – 1800 m). This scheme of an altitudinal differentiation of diagnostic forms of present-day periglacial environments shows a similarity with the Alpine scheme presented by G. Furrer (1965) and H. Stingl (1969), and with the Appenine scheme of Kelletat (1969). The differences relate to the absolute vertical extent of the occurrence of active forms and the intensity of the cryonival processes forming them. Moreover, A. Jahn (1970) assumes that the upper timberline in the Tatra Mts coincides with the lower limit of cryonival phenomena as is the case for the Maritime Alps (Karrasch 1977). Following Slovak scientists H. Karrasch suggests that the timberline does not coincide with the limit of cryonival phenomena as is the case in the majority of Alpine areas, both limits only coincide in the Maritime Alps. According to H. Karrasch (1977) the lower limit of true solifluction in the Tatra Mts is at a height of 2300 m while the bound solifluction occurs at a height of 1750 m. Only detailed quantitative studies will resolve this problem.

In terms of climate, the Tatra Mts occupy a transitional location between the Alps subjected to the maritime influence of the Atlantic Ocean and the Caucasus which are continental (Fig. 1). A precipitation regime with an autumn maximum typical of a Mediterranean climate is noticeable in the Maritime Alps and Ligurian Alps. Precipitation maxima in summer (July) and minima in winter occur in the Central and Eastern Alps. In contrast north-western slopes of the Alpine arc receive an even precipitation distribution throughout the year, this being characteristic of maritime climates. In these respects the Tatra Mts correspond most closely to the Eastern Alps. That assertion is confirmed by the dendrochronological studies of Z. Bednarz (1975) who investigated the relationship between the growth curves of stone pine from the following distant locations: 1) the Bavarian Alps – Patscherkofel, 2) the Tatra Mts – Morskie Oko valley, and 3) the Gorgan Mts – Eastern Carpathians. There was a definite similarity in mean annual curves from these sites even when the distance between them was

300–950 km. The dendroclimatograms which demonstrate the termal and humidity conditions during the last 213 years display a common mid-Europe chronology for annual tree-rings. The width of the tree-rings show a high positive correlation with mean monthly air temperature and precipitation totals, especially for the months May to August (Bednarz 1981).

CLIMATIC CONDITIONING OF THE PRESENT-DAY MORPHOGENETIC ENVIRONMENT

The climate of the Tatra Mts is determined by a substantial elevation above sea level and a central location within Europe, far removed both from seas as well as the Atlantic Ocean. The climate of the Tatra Mts is a transitional one between maritime and continental influences. Therefore, 65% of the time is controlled by maritime polar air masses derived from the Atlantic Ocean (Orlicz 1962). The air is usually strongly modified (37% of days), sometimes fresh (19% of days), while warm maritime polar air masses constitute only a small part (Niedźwiedź 1981). Maritime air masses occur usually in the autumn. Maritime polar air masses cause warming in winter and cooling in summer. In both these causes, the weather is overcast and accompanied by precipitation. Summer precipitation is substantial and associated with thunderstorms. Continental polar air masses occur during 20% of the year; with a maximum in January and a secondary peak in March. Continental air masses are characterized by low moisture content and low cloudiness. This air also causes major temperature inversions on the intermontane basins during the autumn. Arctic air masses affect the Tatra Mts during 6% of the year while tropical air masses only 3% of the year (Niedźwiedź 1981). Arctic air is always cool and dry providing perfect visibility.

Atmospheric fronts separating incoming air masses are the major control in determining weather patterns. During a year the majority of frontal days occurs in the winter (44%) with a minimum being recorded in summer (37%). Warm fronts prevail in spring and autumn while cold fronts dominate in summer (16.9% of days) (Niedźwiedź 1981). Long duration precipitation and extensive cloudiness are related to warm fronts, while short, high intensity precipitation and strong wind are related to cold fronts (Orlicz 1962).

The spatial climatic differentiation of the Tatra Mts is influenced by the following factors:

(i) Exposure and slope angle (both of which vary with elevation) control the radiation balance, and thus determine the macro-, meso-, and microclimates within particular relief units.

(ii) Particular climatic factors are closely related to annual mean temperature. Their values increase or decrease according to an arithmetic series as height

Table 1. Mean monthly air temperatures for January (coldest month) and July (warmest month) in vertical climatic and vegetation zones (climatic data from Hess 1965)

Vertical climatic zone	Zone of vegetation	Mean monthly temperature [°C] January	July
Cold 2200–2663 m	Alpine summit	−12.0	4.0
Temperate cold 1850–2200 m	Alpine meadow	−10.0	6.0
Very cool 1550–1850 m	Dwarf pine	−8.5	8.2
Cool 1100–1550 m	Upper forest	−6.0	10.5
Temperate cool 700–1100 m	Lower forest	− 5.5	13.0

increases. The limits of climatic belts in principle coincide with the limits of vegetation belts (Table 1) (Hess 1965). Mean annual air temperature decreases by 0.5° C *per* 100 m of height. The upper limit of the Alpine meadow belt corresponds with an annual isotherm of −2° C, and the upper limit of the sub-Alpine belt (*Pinus mughus*) corresponds with an annual isotherm of 0° C. The tree line is characterized by an annual isotherm of 2° C, while the boundary between the upper and lower montane belts is characterized by an annual isotherm of 4° C.

(iii) Generally speaking, precipitation totals increase from the foot of the mountain (1400 mm/yr) up to the summits (1625 mm/yr). However, the largest precipitation totals on the northern slopes of the Tatra Mts (up to 1800 mm/yr) occur at a height of 1500–1900 m (Gieysztor 1962; Hess 1965) and are related to incoming moist maritime polar masses during the summer. The largest daily precipitation totals occur at a height of 1400–1600 m (Cebulak 1983). This zone of the highest precipitation is related to the development of convectional clouds (Weischet, *vide* Starkel 1977).

(iv) As the mean annual temperature decreases by 1° C the number of days with snow cover increases by 20, and reaches 290 days (Hess 1965) in the uppermost parts of the mountains. In contrast to the Alps, however, the Tatra Mts do not lie within the region of frequent and rapid mixing of tropical and polar air masses. Therefore, sudden and abundant snow falls which trigger huge snow avalanches do not occur in the Tatra Mts (Myczkowski 1962). Snow cover depth increases with elevation and at 2000 m can exceed 200 cm in February, March, and April (Hess 1965).

(v) Wind speed increases with elevation with mean wind speed increasing three-fold from the foot of the mountains to the highest summits (*i.e.* from c. 2 $m \cdot s^{-1}$ to 6 $m \cdot s^{-1}$). The number of days with strong wind ($\geq$ 10 $m \cdot s^{-1}$) increases from c. 15 at the foot of the mountains to 200–210 days on the

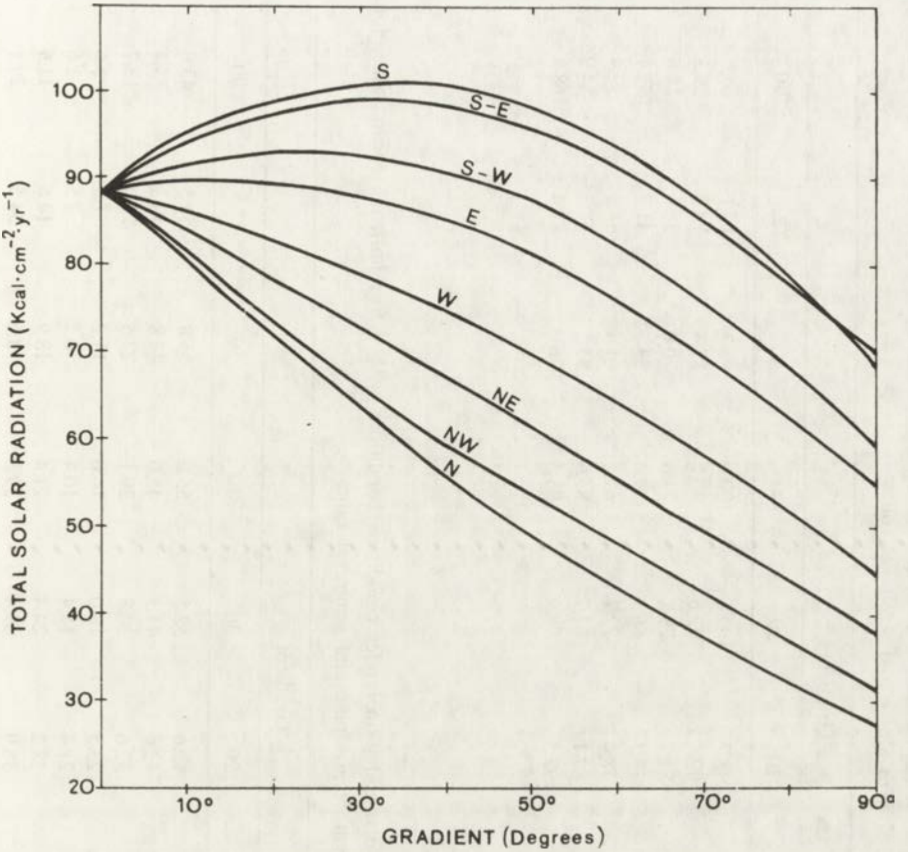


Fig. 6. Dependence of total solar radiation on aspect and gradient (under free horizon conditions) in alpine belt of the Tatra Mts

summits (Hess 1965). The strongest foehn winds attain velocities of the order of 70–80 $m \cdot s^{-1}$ (Otruba, Wiszniewski 1974).

Among the main climatic factors affecting the differentiation of present-day morphogenetic processes radiation, air temperature and precipitation will now be discussed in more detail.

RADIATION CONDITIONS

The termal and moisture conditions on the slopes are mainly caused by variation in the amount of solar radiation reaching sites with varying gradients and exposures. This mainly refers to direct radiation. The total and direct solar radiation for free horizontal surfaces, aking into account the actual cloudiness, were determined at the summit of Kasprowy Wierch (1988 m) in the period 1971–1974. Mean annual values were as follows: total radiation on flat surface $Q = 88.3$ $Kcal \cdot cm^{-2}$, direct radiation $S = 36.6$ $Kcal \cdot cm^{-2}$.

Table 2. Dependence on aspect and gradient of total annual solar radiation (under cloudiness conditions with a free horizon). Mean values for 1971–1974 [$Kcal \cdot cm^{-2}$] in alpine and seminival zones

Aspect	Gradient									
	1°	3.5°	7.5°	14.5°	20°	30°	40°	50°	72.5°	90°
S	89.01	90.7	93.2	96.7	98.7	100.8	100.6	98.3	85.1	68.8
SW	88.6	89.5	90.7	92.2	92.8	92.6	90.7	87.3	73.9	59.3
W	88.1	87.6	86.6	84.6	82.7	79.1	74.8	70.0	56.6	44.6
NW	87.6	86.0	83.2	77.9	73.7	66.3	59.6	53.7	41.2	31.4
NE	87.9	86.8	85.0	81.4	78.4	72.7	67.1	61.4	48.3	37.4
E	88.4	88.8	89.2	89.4	89.1	87.7	85.0	81.1	68.2	54.9
SE	88.9	90.4	92.6	95.6	97.3	98.9	98.5	96.0	83.4	68.3

Table 3. Dependence on aspect and gradient of direct annual solar radiation (under cloudiness conditions with a free horizon). Mean values for 1971 – 1974 [$Kcal \cdot cm^{-2}$] in alpine and seminival zanes

Aspect	Gradient									
	1°	3.5°	7.5°	14.5°	20°	30°	40°	50°	72.5°	90°
S	37.3	39.0	41.6	45.8	48.6	52.5	55.0	55.8	51.5	43.0
SW	36.9	37.8	39.2	41.3	42.6	44.3	45.0	44.8	40.3	33.4
W	36.4	35.9	35.1	33.7	32.6	30.8	29.1	27.5	23.0	18.7
NW	35.9	34.3	31.6	27.0	23.5	18.0	14.0	11.2	7.5	5.6
N	35.8	34.0	30.9	25.4	21.4	15.4	10.4	6.7	2.5	1.6
NE	36.2	35.2	33.5	30.5	28.2	24.4	21.4	18.9	14.6	11.6
E	36.7	37.1	37.7	38.5	38.9	39.4	39.3	38.6	34.5	29.1
SE	37.2	38.7	41.0	44.7	47.2	50.7	52.8	53.5	49.7	42.1

Table 4. Differences in annual totals of direct solar radiation on slopes of opposite aspect [$Kcal \cdot cm^{-2}$] in relation to slope gradient

Aspect	Gradient									
	1°	3.5°	7.5°	14.5°	20°	30°	40°	50°	72.5°	90°
S–N	1.5	5.0	10.7	20.4	27.2	37.1	44.6	49.1	49.0	41.4
E–W	0.3	1.2	2.6	4.8	6.3	8.5	10.2	11.1	11.5	10.4
SE–NW	1.3	4.4	9.4	17.7	23.7	32.7	38.8	42.3	42.2	36.5
SW–NE	0.7	2.6	5.7	10.8	14.4	19.9	23.6	25.9	25.7	21.8

Using these values, the corresponding values for surfaces of varying gradients and exposures were determined according to J. Kondratyev's (1977) formula. These are presented in Tables 2 and 3.

Radiation is very important in heating rockwalls in the uppermost parts of the Tatra Mts and thus strongly affecting the temperature at their surfaces. Accepting the thesis of J. Tricart, A. Cailleux, and P. Raynal (1972), Tables 2 and 3 can be interpreted as follows. Heat reaching the surface of the ground not only affects the temperature of the ground in terms of its diurnal and annual distributions but also in term of evaporation. Providing adequate moisture is present, the changes of temperature in the surface layers of rocks will favour weathering, rockfalls and the transport of debris on the slopes. The total radiation received by slopes of varying gradients and aspects influences the modelling of the type and rate of present-day slope development and seems to affect the climatic asymmetry of the form of slopes of opposing aspects. The largest direct radiation reaches the southern slopes with gradients of 30–50° (Table 3). The smallest radiation is received by the northern slopes and especially vertical walls. Thus, the largest thermal contrasts occur between the N and S facing slopes. Considering absolute values, the largest total solar radiation is received by southern slopes with gradients of 30–40° (100.6–100.8 Kcal$\cdot$cm^{-2}) (Fig. 6). These slopes, along with others, are subjected to strong cooling at nights, thus the largest variation of temperature between day and night occur on such slopes. At an annual scale, the largest differences of 49 Kcal$\cdot$cm^{-2} in direct solar radiation on slopes of opposing aspects occur on N and S facing slopes with gradients of 50–72.5°. The smallest differences relate to more gentle slopes (1–15°) with E–W facing aspects (Table 4: Fig. 7). The above measurements and calculations refer to

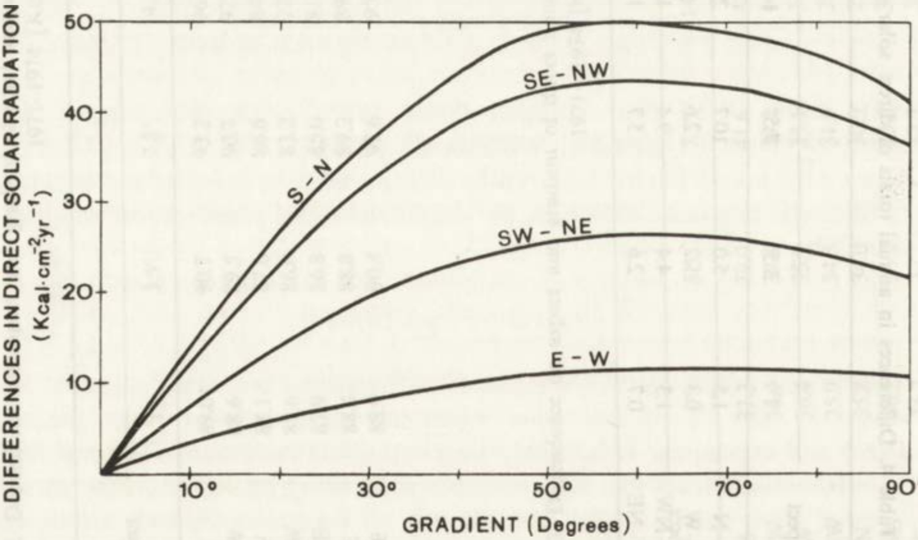


Fig. 7. Variation in direct solar radiation for opposing slope aspects in relation to gradient. Data for an alpine belt of the Tatra Mts

the upper parts of the mountain ridges, located at the height of 2000–2500 m, and constituting the summits of the Tatra Mts, not hidden by other ridges since such hidden horizons were not taken into account in the calculations.

TEMPERATURE CONDITIONS

Alternation of temperature above and below 0° C in the near-surface air layer is an important factor in the physical process of rock weathering. Generally, for each additional 100 m of height the number of days with frost or ground frost increases on average by 7. The largest number of days with temperatures close to 0° C occurs in the 1600–1800 m altitudinal belt where the most favourable conditions exist for frost segregation and slow mass wasting (Kotarba 1976, 1984). Similar values are adopted by M. Klimaszewski (1971) when determining the best thermal and moisture conditions for physical weathering due to frost in the 1700–2050 m altitudinal belt. In the high-mountain part of the Tatra Mts, in the 1800–1850 m altitudinal belt, days with temperature oscillating around 0° C can occur in any month. The foot of the mountains at a height of 1000–1200 m is also characterized by a large number of days with ground frost, but these small oscillations of temperature are of lower morphological importance. The average number of days *per* year with frost ($_{max} < 0°$ C) increases with height at an average rate of 7 days *per* 100 m, and reaches a maximum of 200 days on the highest summits (Hess 1965).

PRECIPITATION

Mean annual precipitation in the Tatra Mts varies from 1400 to 1800 mm, with the maximum values being recorded at 1850 m and only 1625 mm at the highest summits. The belt of maximum precipitation occurs at a height of 1500–1900 m (Hess 1965). A reduction of precipitation on the highest summits is typical of mid-latitude mountains (Flohn 1974). The area of maximum precipitation is related to the influence of moist maritime polar air masses in summer (June–July) moving to the Tatra Mts from the north. Most frequently these air mass movements involve convectional precipitation. Nevertheless, high precipitation totals, mainly in form of snow, are also observed in the highest parts of the Tatra Mts in the winter months (October–April).

Winter precipitation up to a height of 1600 m comprises less than 50% of summer precipitation, whilst winter precipitation on the summits constitutes up to 93% of summer precipitation. In terms of the annual distribution of precipitation, the summits are characterized by an oceanic regime whereas the middle parts of the mountains are more typically continental in regime. Above 1630 m snow fall can occur in any month, and on the highest summits snowfalls can occur even during 9 days in July (Hess 1965).

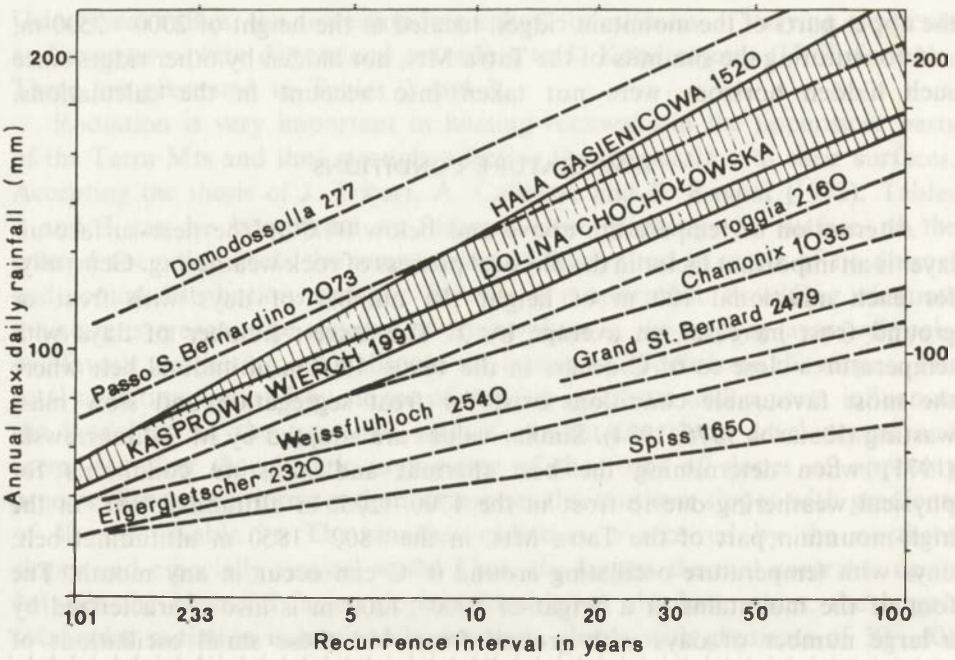


Fig. 8. Recurrence intervals (in years) of annual maximum daily rainfall for high-mountain part of the Tatra Mts compared with selected Alpine stations

Alpine station data from: Zeller, Geiger, and Röthlisberger (1976–1984), Anselmo (1979) and Polish Institute of Meteorology and Water Management. Names of stations are given together with altitude

From a geomorphological point of view daily rainfall is very important as well as the annual and monthly precipitation totals. According to numerous authors such daily rainfall is responsible for catastrophic processes in the high-mountain environment. In this respect the Tatra Mts are comparable to many mountain groups in the Central and Eastern Alps. The annual maximum daily rainfalls on the main ridge of the Tatra Mts (Kasprowy Wierch

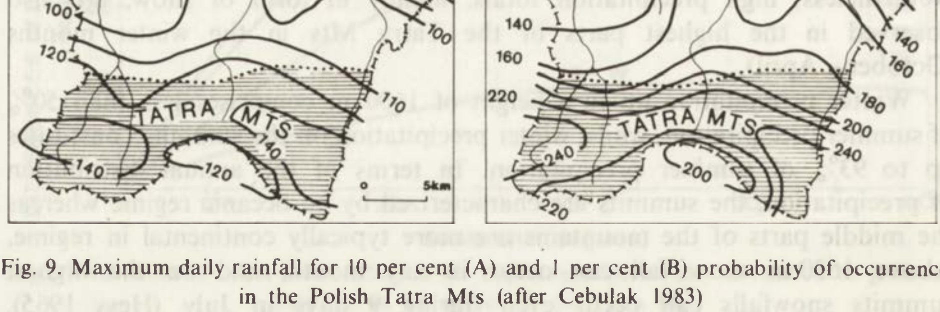


Fig. 9. Maximum daily rainfall for 10 per cent (A) and 1 per cent (B) probability of occurrence in the Polish Tatra Mts (after Cebulak 1983)

1988 m) are of a similar value or even larger than those recorded in the Alpine stations located at heights of 2000–2500 m. The station located in the middle part of the Tatra Mts (Hala Gąsienicowa 1520 m) is characterized by particularly high maximum daily rainfalls which frequently recur (Figs. 8–10). This station is situated in the area of the above-mentioned altitudinal belt of 1400–1600 m where a maximum daily precipitation of the order of 200–300 mm occurs once *per* 50 years (Cebulak 1983). Maximum daily precipitation on the summits is definitely smaller and does not exceed 165 mm once a 100 years (Table 5).

Table 5. Maximum daily rainfall (1951–1980) for chosen stations (after Cebulak, 1983)

Station	Altitude [m]	H_{max}	Maximum daily rainfall [mm] with probability of					
			1%	2%	3%	5%	10%	50%
Kasprowy Wierch	1988	232	165	152	144	133	118	75
Hala Gąsienicowa	1520	300	224	201	187	170	146	85
Zakopane	857	139	177	158	147	133	114	65

However, maximum daily precipitation is not the best measure for determining the precise rainfall threshold for initiating catastrophic slope failures. Such a threshold has to be determined taking into account the total amount of rain, the rainfall duration and the rain intensity (Starkel 1979; Caine 1980). N. Caine suggests that the stability threshold for mountain slopes is exceeded and smaller landslides occur when the rainfall intensity (*I*) expressed in $mm \cdot hr^{-1}$ and rainfall duration (*D*) in hr reach a functional relationship of the form:

$$I = 14.84\ D^{-0.39}.$$

Conditions implying that a given stability value will be exceeded in any year with a probability of 50% for Hala Gąsienicowa (*ie* every second year) can be developed.

* * *

Both the type and the course of present-day geomorphic processes depend on their timing and sequence. Thus four morphogenetic seasons (*ie* periods characterized by different process complexes) have been distinguished in the Tatra Mts above the upper timberline. M. Kłapa (1980) has distinguished nival, niveo-pluvial, pluvial, and pluvial-nival seasons with varying duration. The longest is the nival season (155 days) constituting 42% of a year. However, this season is only slightly active in morphological terms as the ground is protected from the atmosphere by the snow cover. The following niveo-pluvial season, although the shortest one (lasting only 38 days), is

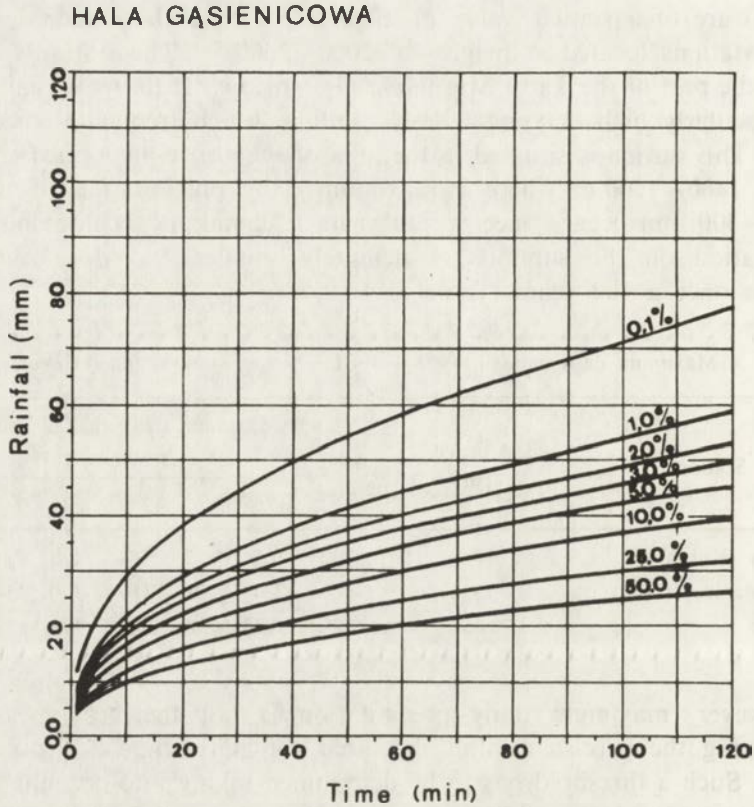


Fig. 10. Gumbel distributions for annual maximum precipitation for various return periods (0.1 – 50 per cent probability) on the upper timberline in the Tatra Mts (after Cebulak 1983)

most important morphologically. This is the period of the spring thaw during which the maximum range of air and ground temperatures occurs. During this period almost all processes are active. The pluvial season lasts a very long period (121 days) according to M. Kłapa (1980). It can be active or very active since the effects of the characteristic geomorphic processes (and especially debris flows) vary between individual years and depend on the occurrence of large daily rainfalls of high intensity. The fairly short pluvial-nival season (being a transition between the warm and cold period of the year) is characterized by the occurrence of many processes typical of the remaining seasons with cryogenic and aeolian processes predominant. The activity of processes during morphogenetic seasons varies from year to year. Although the potential conditions which enable processes to be active can be determined statistically, their actual occurrence is subject to considerable variations between years. This arises because of the transitional nature of the climate of the Tatra Mts which is transitional between a mountain climate of an area close to the Atlantic Ocean and a continental climate of the mountains of Eurasia.

There are significant climatic similarities between the Tatra Mts and other high mountain areas in Europe including the Alps. A comparison of the mean air temperatures for June and July in the Alps and Tatra Mts for 1741– 1965 shows close similarities in terms of temperature conditions (Bednarz 1984). Dendroclimatological reconstructions indicate a similar pattern of temperatures in the summer months. In both mountain areas the lowest temperatures occurred in 1741– 1747, 1770– 1773, and in 1805– 1845 (in the Tatra) and 1790– 1855 (in the Alps). The last period called the Little Ice Age is demonstrated in the Alps by the advance of glaciers. In the Tatra Mts data on weather, snow, and hydrological conditions are absent since regular observations were not started until 1911. However, a comparison of mean air temperatures for June and July with temperatures calculated on the basis of tree-ring analysis of the stone pine for the period of 1911 – 1965 (Bednarz 1984) indicates that the differences do not exceed 0.1- 0.3° C. Therefore, the above author suggests that a climatic affinity exists between the Tatra Mts and the Alps.

POSTGLACIAL EVOLUTION OF RELIEF

The conditions of the natural environment changed during the Holocene and determined the present-day vertical structure of the Tatra Mts and the development of the fluvial-denudation system in the mountains (Fig. 11). The development of relief during the Holocene corresponded to climatic-vegetation belts (Klimaszewski 1962) which in turn changed their altitudinal position and extent. The major transformation of relief occurred during the Last Glacial. Therefore, a reconstruction of the conditions and results of deglaciation is important in understanding the Holocene morphogenesis of the Tatra Mts. In addition, the short period of human impact on the natural environment also appears a very important factor in their postglacial history. Paleogeographical studies of the Tatra Mts allow one to reconstruct only a general incomplete record of postglacial environmental change.

The boundary between the Würm and Holocene within the Tatra Mts has been identified by investigations of lacustrine deposits in the Pięć Stawów Polskich and the Sucha Woda valleys (Wicik 1979, 1984; Krupiński 1983; Szeroczyńska 1984). A series of two-phase deposits exceeding 2 m in depth were obtained using a coring instrument. The amount of organic matter increases steadily from the Würm/Holocene boundary, while two dates C^{14} (10,100 ± 140 BP and 9900 ± 120 BP) for Przedni Staw and Czarny Staw lakes in the Pięć Stawów Polskich valley separate the upper – Holocene part of the profile, from the lower – Late Glacial one. The transition from minerogenic to organic deposits was also marked by the introduction of almost all the species of the invertebrate *Cladocera* (Szeroczyńska 1984).

EVOLUTION OF GEOECOLOGICAL BELTS

The Late Glacial was significant in morphogenetic terms mainly because of climatic oscillations which ultimately tended towards a general warming. Deglaciation and a change in position of the geoecological belts resulted from these climatic changes. During the Alleröd when only 15% of the northern margin of the Tatra Mts (950 – 1000 m) were located within the forest zone, the mean annual temperature did not exceed 3° C (Hess 1968). The cold nival and niveo-pluvial type of high mountain climate during the

Late Glacial is evident both in terms of a scarcity of organic deposits and by exclusively cool-adapted species of *Cladocera* in pre-Holocene lacustrine deposits (Szeroczyńska 1984). During this period the snow line changed its height from c. 1000 to 1850 m as calculated by M. Hess (1968) on the basis of paleobotanical data of W. Koperowa (1962). The Tatra Mts were dominated by a cryonival belt at that time. In the uppermost part of the High Tatra Mts small glaciers could have persisted until the end of the Würm where morphological conditions were favourable. The high mountain landscape of the Tatra Mts resembled present-day mountains which have reached a very advanced stage of deglaciation with cirque glaciers and snow patches preserved in the nival belt. The data of W. Koperowa (1962) as well as those of K. Krupiński (1983) indicate that the Alleröd forest retreated from the Tatra Mts such that cryonival and glacial systems prevailed there during the Younger Dryas.

Using the stratigraphical terminology of L. Starkel (1977) three stages can be distinguished in the Holocene history of the natural environment of the Tatra Mts:

(i) eoholocene stage of warming and associated rapid ascent of the geoecological belts; the upper timberline reached c. 1800 m over a period lasting 2000 years and the associated snow line has been found at the level of the uppermost summits of the Tatra Mts (Fig. 11).

(ii) Mesoholocene stage of forest belt dominance which according to K. Krupiński (1983) should have reached 1850–1950 m.

(iii) Neoholocene stage of a lowering of geoecological belts so the summits of the High Tatra were again above the snow line (annual isotherm of $-2°$ C according to M. Hess 1965).

The hypsometric belt of 1500–1950 m within which important morphological limits shifted during the Holocene was subject to the largest changes of the environment as morphogenetic systems with different dynamics varied here. Most rocky and debris slopes are found within this belt and they are the object of postglacial erosion and aggradation corresponding to the basal part of the step-like preglacial sections of the valleys. Based on the volume of debris covers M. Lukniš (1968) estimates average lowering of ridges during the Holocene of 5 m.

Changes in the upper timberline in the Tatra Mts are also evidenced by dissected, rounded karren (Kotarba 1967) occuring at a height of 1500–1860 m on limestones. The rounded karren which were formed in a forest environment have been dissected after the Atlantic optimum when the upper timberline was lowered. Geomorphological adaptation of the forest belt (Starkel 1977) in the Tatra Mts resulted mainly in a reduction of periglacial forms being modelled by the forest ecosystems, and in an intense chemical weathering, and the dissection of the valley floors and the slopes. Geomorphic processes were only more effective above the upper timberline and on slopes severly affected by man (Klimaszewski 1967b;

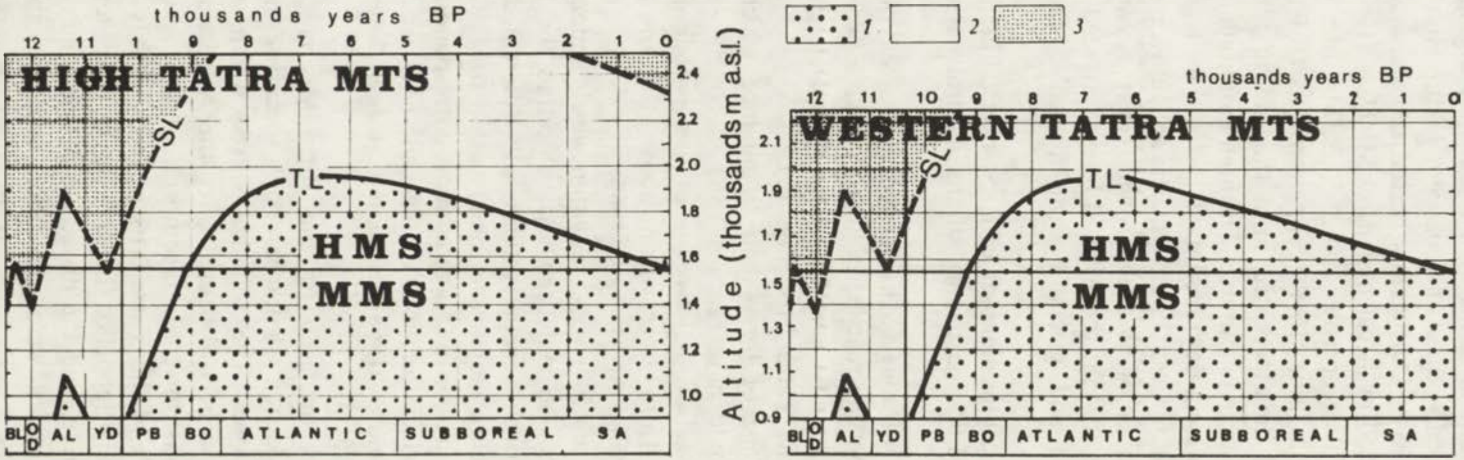


Fig. 11. Late Glacial and Holocene timberline (TL) and snow line (SL) oscillations in the Tatra Mts

Present-day high-mountain (HMS) and middle-mountain (MMS) systems are separated by horizontal line. *1* – forest belt, *2* – subnival belt, *3* – nival and seminival belt (compiled after Koperowa 1962, Hess 1968, Krupiński 1984)

Table 6. Age of the organic and mineral deposits in the Polish Tatra Mts

Research site	Altitude [m]	Beginning of sedimentation	Applied method	Author
Mołkówka (p)	954	Preboreal	palynological	W. Koperowa 1962
Przedni Staw (l) in Pięć Stawów Polskich valley	1668	Oldest Dryas	palynological, sedimentological, ^{14}C	K. Krupiński 1983; Wicik 1979, 1984
Czarny Staw (l) in Pięć Stawów Polskich valley	1722	Oldest Dryas	„	„
Zadni Staw (l) in Pięć Stawów Polskich valley	1890	Subatlantic	sedimentological	B. Wicik 1984
Toporowy Staw Niżni (l)	1087	Atlantic/ /Subboreal	palynological	J. Stasiak 1984
Toporowy Staw Wyżni (l)	1135	late Boreal	„	J. Dyakowska 1932
Wielka Pańszczycka Młaka (p) Pańszczyca Valley	1265	Preboreal	„	A. Obidowicz 1975
Litworowy Staw Gąsienicowy (l)	1618	late Preboreal	„	J. Dyakowska 1932
Zielony Staw Gąsienicowy (l)	1672	Younger Dryas (?)	sedimentological	B. Wicik 1984
Czarny Staw Gąsienicowy (l)	1620	Alleröd (?)	„	B. Wicik 1984
Peat bog near the Smreczyński Staw lake	1226	Boreal/Atlantic	palynological	J. Dyakowska 1932
Smreczyński Staw lake	1226	Boreal/Atlantic	palynological	Z. Skierski 1984

p-peat bog
l-lake

Midriak 1983). According to M. Klimaszewski the scale of Holocene lowering and remodelling of Pleistocene relief in response to climatic-vegetation belts is usually small. Active, small cryogenic forms and non-active and clearly larger ones identified by A. Jahn (1958) above the present-day upper timberline indicate that the cryonival belt was modelled during the Neoholocene period under more temperate climatic conditions than operated at the end of the Late Glacial. The formation of the present-day cryonival system of the Tatra Mts (Kotarba, Starkel 1972) was possibly the result of the post-Atlantic cooling of climate. This is probably denoted by the fifth and sixth *Cladocerophases* distinguished by K. Szeroczyńska (1984) as well as by the fact that warm-adapted species of *Cladocera* fauna: *Alona quandrangularis* and *Chydorus* disappeared from the lakes of the northern slopes of the Tatra Mts at this time. Cooling and an increased humidity in the Subboreal and Subatlantic periods (evidenced by palynological profiles of peats and lacustrine sediments in the Tatra Mts—Table 6), could have caused increased debris flow activity transforming the talus slopes.

The differences in the absolute altitudes between the High and Western Tatra meant that in the final stage of their Neoholocene evolution only the summits of the High Tatra (above 2250–2300 m) were above the snow line (Fig. 11) as in the Eoholocene. The lack of favourable morphological conditions for snow accumulation makes the present-day glaciation impossible.

THE CONDITIONS AND PATTERN OF DEGLACIATION

Deglaciation in the Tatra Mts occurred under conditions of a Late Glacial warming of climate and the development of forest ecosystems. Deglaciation took place almost simultaneously although not at the same rate as evidenced by the ridges of recessional moraines in different valleys. M. Klimaszewski (1967a) identifies 6 recessional stages in the High Tatra and 3 stages in the Western Tatra, and relates them in turn to the recessional stages of the Austrian Alps: Bühl, Schlern, Gschnitz, Daun, Egesen and Fernau. Terminal deglaciation occurred in the lower sections with steep gradients while areal deglaciation occurring in sections with lower gradients. According to M. Klimaszewski (1967b) in the Western Tatra glaciers disappeared in the period which preceded the Oldest Dryas, whilst in the High Tatra it is likely that they were preserved right into the Younger Dryas when cirques above 1800 m could have been filled with ice. M. Lukniš (1973) has distinguished 5 recessional stages in the High Tatra and has identified high elevated protalus ramparts which were still forming during the Preboreal period and continue to develop down to the present day. The pattern of deglaciation reconstructed by M. Klimaszewski (1967b) and by M. Lukniš (1973) was based on the results of palynological studies of peat-bogs in the Nowy Targ basin (Koperowa 1958, 1962). M. Klimaszewski (1967b) relates the upper timberline to the snow line and in turn, the corresponding ridges of recessional moraines. This general model of the pattern of deglaciation has not been based on dates from deposits actually found within the Tatra Mts.

Studies of peat-bogs in the Tatra Mts (Dyakowska 1932; Obidowicz 1975) and especially those of lacustrine deposits (Kondracki 1984; Krupiński 1983; Stasiak 1984; Skierski 1984; Wicik 1979; Więckowski 1984) provide new data on the paleoenvironments of the Tatra Mts during the Late Glacial and the Holocene (Table 6).

By the end of the Boreal and into the Atlantic period lakes located within the zone of terminal moraines (Sucha Woda valley c. 1100 m) started to function as sedimentation reservoirs (Kondracki 1984; Table 6). Deposits in the Przedni Staw and Czarny Staw lakes in the Pięć Stawów Polskich valley have been analysed palynologically and dated by radiocarbon. According to B. Wicik (1979) and K. Krupiński (1983) these sites indicate that ice did not exist in that part of the valley in the Oldest Dryas. The determination of pre-Holocene deposits on the basis of an arbitrary but assumed sedimentation rate is doubtful. It seems more reasonable that deglaciation in the area of the above mentioned lakes occurred during the Bölling or Alleröd as suggested by B. Wicik (1984) on the basis of deposits in Czarny Staw Gąsienicowy lake. The highest lakes of the Polish Tatra Mts (Zadni Staw lake in Pięć Stawów Polskich valley) are relatively young, and have deposits dating from the Subboreal or Subatlantic periods (Table 6). The young age of these lacustrine deposits cannot be attributed to the preservation of dead ice protecting the lake basins until the Neoholocene. Climatic conditions suitable for the preservation of the relict ice, not-buried in the morainic material, were only maintained up to the Preboreal.

As a result of the limnological studies of B. Wicik (1979), Baumgart-Kotarba and Kotarba (1979) were able to relate recessional moraine ridges in Biała Woda, Roztoka, and Rybi Potok valleys to their contemporary counterparts in the Austrian Alps. According to these latter authors, the moraine ridges surrounding Przedni Staw and Wielki Staw lakes in the Pięć Stawów Polskich valley along with the morainic ridges in the vicinity of Włosienica in Rybi Potok valley plus those in the area of the outlet of Żabie Stawy Białczańskie valley are all dated to the Oldest Dryas and correspond to the Gschnitz stage in the Alps (13,000 BP).

Within the most severly glacierized valley systems of the High Tatra (Biała Woda and Sucha Woda valleys) the lower section of the glaciers disappeared areally as a result of the large amount of morainic material supplied within the glaciers from the large source areas. In the lower parts of the glacial valleys of the Polish Western Tatra dead-ice features, which could have provided evidence of the pattern of areal deglaciation, were not preserved. Nevertheless, such deglaciation is confirmed above 1400–1500 m.

Deglaciation in the valleys of the Western Tatra was probably rapid but ceased at a height of 1500 m. The terminal moraine ridges identify the recessional stagnation phase from which areal deglaciation commenced within the cirques basins. Fossil rock glaciers occur above these ridges (Fig. 22), large amounts of debris being supplied from the slopes onto the surface of the cirque glaciers. This, in turn, promoted irregular ablation. Rock glaciers were formed due to the stagnation and disappearance of ice-glaciers. The snow line calculated by the Höffer's method was at a height of 1675 m during the youngest stagnation phase. Thus, almost the entire mass of the glaciers was below that line which, in turn, provided the major climatic cause for their melting. The height of the snow line is compared with its position given during the Younger Dryas by M. Klimaszewski (1967b) and M. Hess (1968). Present studies are not yet adequate to determine definitely the age of the last stagnation phase. Nevertheless, the Younger Dryas should be recognized as the last period with climatic conditions favourable for the existence of active rock glaciers.

Due to a rapid change in the position of the climatic belts at the beginning of the Holocene, rock glaciers started to become inactive and within the forest belt they became entirely fossil. The relict glaciers on slopes with a marked northerly aspect were preserved as part of the final phase of deglaciation at the base of the slopes. The debris supplied from the slopes was transported across the surfaces of snow patches thus generating protalus ramparts which are especially well preserved in the cirques of Pyszniańska valley. On the inner side of these ramparts hollows or accumulation zones, with a more gentle gradient result on account of gravitational and fluvial deposition from the slopes, occur. However, numerous hollows are not buried by these deposits. Stabilization of these slopes by forest during the Boreal and Atlantic periods made the burial of these depressions less likely. The long persistence of the snow patches and thus the extended formation of protalus ramparts is confirmed by the fact that systems of similar ramparts forming the lobate structures of talus-foot rock glaciers occur below them. These are of the same age or younger than the rock glaciers which developed during deglaciation.

RELIEF TRANSFORMATION DURING THE HOLOCENE

Gravitational processes were dominant on rockwalls and rocky slopes during the whole of the Holocene in the High Tatra where the cryonival belt was more extensive both vertically and spatially. The rather moist Neoholocene climate caused an intensification of wash processes on debris slopes.

The role of water on the slopes became significant in the Western Tatra. This is evidenced by interbedded clays, sands, and coarse debris in units at least 3.7 m thick which occur in the troughs between the rock glacier ridges (Kaszowski, Krzemień, Libelt 1987). Palynological dating shows that wash processes eroded debris slopes with the result that the depressions and troughs were infilled with younger sediments until the end of the Boreal period, thus burying the glacial and periglacial forms within the cirques.

Human impact on the natural environment of the Tatra Mts mainly took place during the 19th and 20th centuries as a result of burning and clearing of the forests, overgrazing of livestock, and the extraction of metal ores and mining. The main results of these activities, which occurred with varying intensities, were a lowering of the upper timberline over 67% of its length (Radwańska-Paryska and Paryski 1973) and the destruction or damaging of the turf covering the soils on the slopes. The livestock grazing in the Tatra Mts exceeded the carrying capacity of the Alpine meadows up to the First World War, whilst after the Second World War the number of sheep reached almost 30,000 in 1964, *ie* 5–7 times too many for the area. The environment of the High Tatra became only slightly changed as a result of the impact of overgrazing because of the less favourable conditions for grazing (*ie* rockwalls). Thus, the impact of land management practices mainly affected the Western Tatra.

When compared to the postglacial history of the area, a relatively short period of human impact on the Tatra Mts caused a severe degradation of the slopes, to a large extent caused by debris flow activity on an unprecedential scale. The last 200–300 years are only 1/50–1/30 of the whole Holocene whereas the thickness of sediments removed from slopes and deposited in the troughs and depressions of Starorobociański cirque comprise 50% of their whole postglacial profile, and locally this is probably even a larger percentage. Indices of slope degradation and aggradation on the floor of the Starorobociański cirque have been calculated on the basis of the volume of postglacial deposits. This involves dating the surface deposits by C^{14} method (360 ± 50 BP) and the use of palynological methods both illustrating the scale of the problem being considered (Kaszowski, Krzemień, Libelt 1987). If the slope erosion index (calculated on the basis of the volume of sediments above the ridge of the recessional moraine closing the cirque and related to the area of slopes) was 0.4 $mm \cdot yr^{-1}$ during the last 13,000 years then the same index (calculated on the basis of relating the volume of the upper face sediments accumulated in the troughs and depressions between debris ridges to the area of the slopes) was 2.1–0.7 $mm \cdot yr^{-1}$ for the last 100–300 years. Similarly, the index of aggradation on the floor of the Starorobociański cirque was 0.3 $mm \cdot yr^{-1}$ for the 13,000 years and 5.0–1.7 $mm \cdot yr^{-1}$ for the last 100–300 years. Therefore, during the last few hundred years of intensive human impact, the Holocene tendency of smoothing the glacial and periglacial relief of the cirque floors by infill from slope sediments became strongly accentuated.

GEOECOLOGICAL BELTS

The Tatra Mts– the highest part of the Carpathian system, due to their significant elevation above sea level– are characterized by a vertically differentiated circulation of heat, water, organic, and mineral matters. Differentiation of climatic factors and vegetation associations overimposed on the forms and sediments resulting from the earlier geological periods (especially from the Pleistocene) is a basis to distinguish four geoecological belts, namely: seminival, alpine, subalpine, and forest (montane) ones. The characteristics of the above belts is based on the results of the studies by M. Klimaszewski (1962), S. Pawłowska (1962), M. Hess (1965), K. Wit-Jóźwik (1974), A. Kotarba (1976), T. Komornicki, and S. Skiba (1979) (Table 7; Fig. 12). The upper timberline (1500– 1550 m) separating the forest belt from the subalpine belt is also a limit between two different morphogenetic systems: the forested middle mountains of the marginal part of the Tatra with dominating at present fluvio-denudational morphogenesis and the high mountains of a cryonival morphogenesis. The middle mountain system corresponds to tectonic units built of mesozoic rocks (limestones, dolomites, sandstones, marls, and shales) while the high-mountain system has developed on the crystalline core and mesozoic high-Tatric nappes (Fig. 6). Various rock resistance of the crystalline core-granite rocks in the High Tatra and granitoide-gneiss rocks in the Western Tatra are responsible for 400 m difference in their absolute height. Seminival belt does not occur in the Western Tatra.

Seminival zone includes the area located above 2150– 2300 m. It comprises the system of crests and summits coming down to the valleys by rockwalls or steep rocky slopes. As a result of the significant elevation above sea level (causing a severe climate) and the absence of a thick weathering mantle, the poor polygonal soils only occur sporadically on a slightly inclined slopes or on a few flat elements of the relief. On the granite substratum of the High Tatra the soils are of podzol ranker type. Bare surfaces, and areas only locally colonized by loose pioneer grasses, and of lichens prevail here. The belt is under the influence of a nival (seminival) climate with a mean annual air temperature below –2° C, and is thus situated above the climatic snow line (Hess 1965). The duration of the snow cover

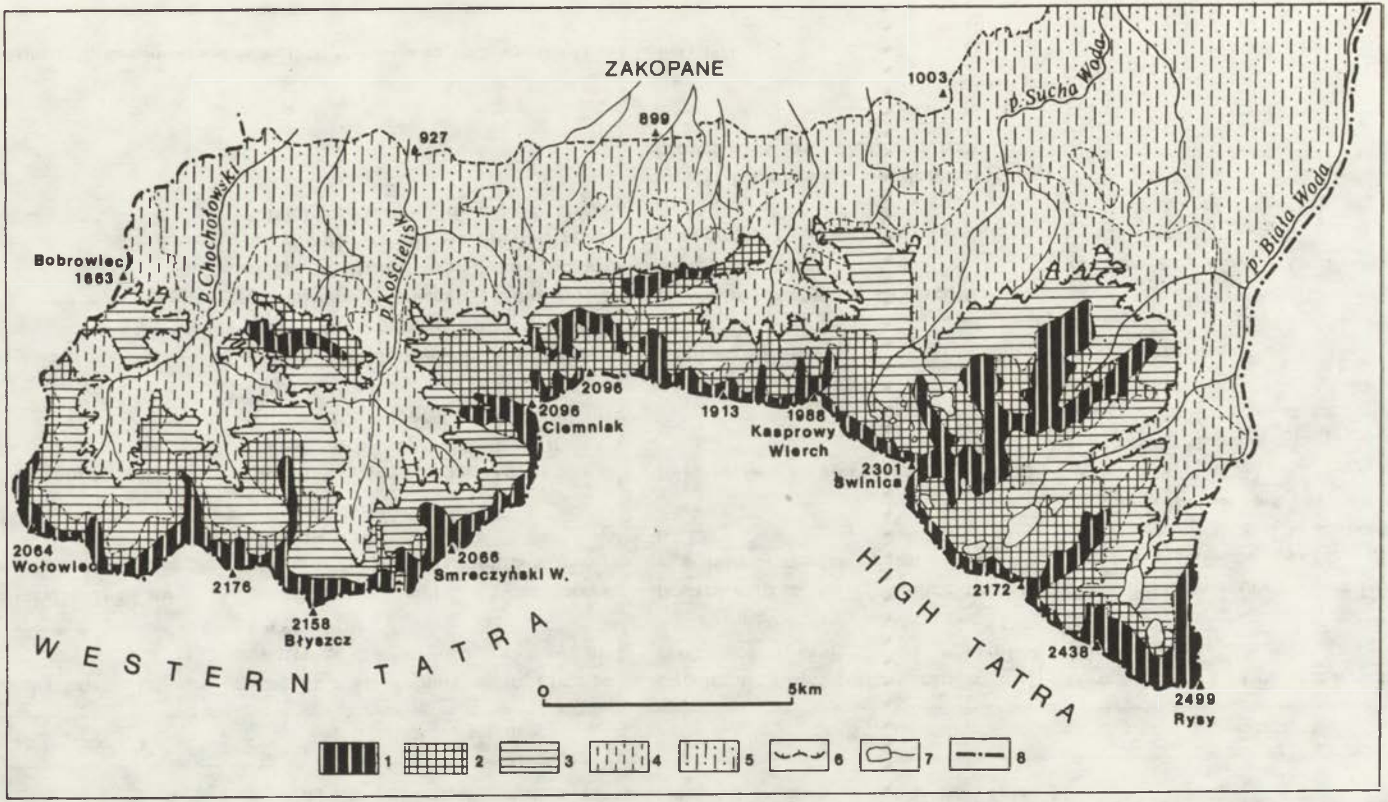


Fig. 12. Geoecological belts in the Polish Tatra Mts (partly after S. Myczkowski *et al* 1974)

1 – seminival belt, 2 – alpine belt, 3 – subalpine belt, 4 – upper forest belt, 5 – lower forest belt, 6 – upper timberline, 7 – lakes and streams, 8 – Polish-Czechoslovak frontier

Table 7. Vertical zonation of geoecological features in the Polish Tatra Mts

Geoecological belt	Landform system	Soils*	Vegetation**
Seminival (above 2150 m)	summit system: narrow to knife-edged ridge crests, steep rockwalls and rocky slopes	polygonal soils, podzol ranker (on crystalline rocks); initial rendzinas (on calcareous rocks)	bare rock or open pioneer vegetation, *Distichetum subnivale*
Alpine (above 1670 m)	interfluve and slope system: broad, rounded ridge crests, debris slopes in glacial corries	podzol ranker (on crystalline rocks); raw humus rendzinas (on calcareous rocks)	grassland association *Trifido-Distichetum* (on crystalline rocks); *Caricetum firmae* (on calcareous rocks), *Versicoloretum tatricum*
Subalpine (above 1500 m)	glacial valley bottom system: glacial floors with slight stream incision within drift deposits, partly debris slopes	humic rendzinas (on calcareous rocks)	dwarf pine: *Mughetum carpathicum silicicolum*, *Mughetum carpathicum calcicolum*
Forest (above 900 m)	erosion-denudational middle mountain system bearing imprint of periglacial morphogenesis: moderately steep to steep slopes, valley floors filled with glacial, glacifluvial, niveo-fluvial sediments, moderate stream incision		
upper		humus-iron podzol (on crystalline rocks); brown rendzinas (on calcareous rocks)	*Piceetum tatricum*
lower		iron-humus podzol (on crystalline rocks); typical brown soils and leached brown soils (on calcareous rocks)	*Fagetum carpathicum Luzulo-Fagetum*

After: * Skiba 1977, Komornicki and Skiba 1979; **-Pawłowska 1962, Myczkowski and Lesiński 1974.

exceeds 290 days *per* year; a period from October to May is hydrologically inactive since the water is stored within the snow cover. In the remaining part of the year only surface runoff occurs whilst the infiltration into a rocky substratum is small. Under such conditions physical weathering (freeze-thaw cycles), rock fall, corrasion of the ground by sliding debris, and cryogenic processes (especially frost sorting) are the dominant processes. Processes acting adjacent to the long persisting snow patches are common in June and July. Snow avalanches occur frequently but are of little morphological significance.

The lower limit of the alpine belt descends down to a height of 1670 m and up to 1800 m (providing the appropriate geomorphological conditions occur). The relief of this belt comprises the systems of slopes and ridges with lower absolute heights (1700– 1900 m) which occur mainly in the Western Tatra. The crests are either narrow or wide and rounded; they are covered with a weathered mantle and are rocky in some parts. They also merge into rocky slopes or mature debris mantled slopes. In the High Tatra the rockwalls and rocky slopes are compact, significant in their vertical extent and dissected by allochtonic chutes which start in the seminival belt. In the Western Tatra the rocky slopes occupy a relatively small area and are strongly dissected by a system of chutes. The small-size rockwalls create a mosaic pattern and sequence of forms. They indicate a mature stage of development of the high-mountain slope system. The gradient of the rocky slopes is over 50°, however the gradient of the mature slopes with numerous rocky forms is 40– 50° or less. Debris-mantled slopes occur in the summit parts of the ridges, especially within the Western Tatra. They accompany dissected rocky slopes. The gradient of the debris-mantled slopes is 30– 36°, and the thickness of the mantle is locally 0.3– 0.5 m. The debris covers exhibit some features of dislocation. The microrelief of the slope surface is usually highly variable. Systems of actually inactive corrasional troughs are well preserved. The turf cover is not complete. Small tongues of material debris stream, irregularities, steps and terracettes, nival niches, gelideflational steps, and block fields provide evidence of a complex evolution of the debris-mantled slopes. The block slopes, lithologically controlled on the quartzite rocks of the Western Tatra and relict forms from the Pleistocene, are particular cases of debris-mantled slopes. At present, they are very stable having become fossil forms stabilized by vegetation. 'Smooth' debris-mantled slopes are rather common. Debris slopes usually stabilized by vegetation occur in the Western Tatra below the rocky slopes and debris-mantled slopes. They have gradients of 26– 35°, and most frequently c. 30°. The upper part of these slopes is dissected by a system of deep (up to 5– 6 m) and wide (up to 20 m) gullies of debris flows. The base of these slopes, with a gradient of 18– 22°, is of colluvial origin and is related to debris flow deposition. It is modified by a large number of gullies up to 0,5 m deep and by accompanying levees, tongues and mud-debris cones. The floors of the cirques are covered by glacial drift deposits and colluvial covers (material originating from sliding and rock fall) while the depressions and troughs are frequently infilled with material derived from slope wash-off debris and glacial covers. The characteristic forms of the lower part of the alpine belt are protalus ramparts and lobate fossil rock glaciers.

The soils of the alpine zone are of the podzols rankers type on crystalline substratum and humus rendzinas on the calcareous substratum (Table 7), where they are covered with rich high mountain meadows. This area is under the influence of a niveo-pluvial climate with a mean annual air temperatures of 0° C to –2° C and a snow cover of 250 days. Rapid overland flow is accompanied by infiltration and interflow. Snow storage lasts a much shorter period and intensified hydrological activity occurs during the thaw period (*ie* in April and May) when storage of water in the regolith is readily observed.

Within the set of dominant morphogenetic processes the following are most significant: gravitational processes on the rockwalls and rocky slopes, frost creep, solifluction, and creep of debris covers on gently inclined surfaces. In addition to the latter processes which cause a slow translocation of the weathering mantle over long periods of time, debris flows occur periodically during the high intensity summer storms. The fresh forms of troughs and cones provide evidence of the significant role played by debris flows in the morphogenesis of the alpine belt. Strong winds, and especially the foehn, cause deflation of soil particles especially in the passes from surfaces which are not stabilized by vegetation. Wind activity is promoted and accelerated by a loosening and dislocation of the weathering mantle as a result of needle ice. In the alpine geoecological belt there are also significant nival processes and the transportation of debris in snow avalanches which occur in rock chutes and on mature debris-mantled slopes.

The subalpine belt occurs above the upper timberline. Here the dominant relief comprises the floors and slopes of the glacial cirques which are mantled by glacial, glaciofluvial, and colluvial deposits. The residual forms of the rocky slopes and debris slopes well visible in the Western Tatra descend down to the timberline. In the High Tatra the cirque floors have been significantly reworked by glacial erosion. They are overdeepened, display numerous lakes, or are mantled by ablation moraine with dead-ice hollows and chaotically distributed moraine hills and ridges. Systems of huge tongue-shaped rock glaciers with highly differentiated topography occur in the Western Tatra in the cirques within the Kościeliska and Chochołowska valleys. They are formed by numerous transverse, oblique- and longitudinal curved debris ridges. The heights of these ridges are usually 3– 5 m although occasionally they reach 7– 10 m. Among the ridges and at the margins of the debris tongues are lateral and longitudinal troughs and depressions. Some of these are infilled with deposits which are fluvial in origin. Systems of dead-ice hollows occur above the rock glacier tongues. In the uppermost parts of the subalpine belt protalus ramparts and fossil lobate rock glaciers are common. In the lower sections of the glacial cirques systems of intermittent stream channels are readily identified.

The typical soils of the subalpine belt are humus podzols on the crystalline substratum and humus rendzinas on the calcareous substratum. The soil is a habitat for the dwarf pine which grows on the ridges of rock glaciers and thus emphasizes their pattern. As in the alpine belt the subalpine belt is under the influence of the niveopluvial climate. However, the mean annual air temperatures are in the range 2° C to 0° C and the duration of the snow cover is up to 215 days. Water penetrates deeply into the glacial and glaciofluvial deposits and there is a large storage of groundwater. A large surface storage in lake basins also occur within this belt, thus surface runoff is relatively small.

Among the dominant morphogenetic processes within the subalpine belt the following are notable: piping, aeolian deposition of material transported from the ridges, and deposition of debris material (mainly silty sands and clays at the limit of debris flow gullies). These deposits have caused the glacial and periglacial landforms within the cirques in the Western Tatra to become fossilized.

The forest belt extends from the foot of the Tatra Mts (c. 900 m) to the upper timberline (1500 – 1550 m). The middle mountain relief is characterized by remnants of the glacial and periglacial morphogenesis in the valleys which were glacierized totally or partially in the Pleistocene. A system of fluvial-denudational forms dominates here. The slopes are steep or fairly--steep and more mantled with covers of periglacial origin (coarse debris inter-bedded with fines originating from chemical weathering in the Holocene). Tors are locally preserved in areas of more resistant bedrock. Valley floors are also thinly covered with glacial, glaciofluvial and niveo-fluvial deposits which are usually coarse and frequently are dominated by the boulder fraction. The valley sides are mainly incised by deep gullies at the outlets of which occur large torrential cones corresponding to former base of the glaciers. In some locations lateral moraine ridges are interrupted by these cones and the main valley is interrupted and blocked, thus providing evidence of the severity of debris flow processes during postglacial time. The glacial and fluvioglacial floors of the valleys are dissected by a system of channels which are well stabilized.

In the forest belt humus-iron podzols have developed on the crystalline substratum whilst brown rendzinas have developed on the calcareous substratum. At higher levels spruce forest grows whilst Carpathian beechwood is found at lower levels, both grow on slopes and valley bottoms. Numerous glades in the forest related to former grazing and mineral extraction are actually occupied by meadow associations. In terms of climate this zone is a transitional one between niveo-pluvial and pluvio-nival climates and is characterized by mean annual air temperatures from 2° C at the upper timberline to c. 6° C at the foot of the Tatra Mts. The duration of the snow cover varies from 140 to 105 days. Although the groundwater flow is dominant, the surface runoff becomes more significant in terms of the whole hydrological system.

The dominant morphogenetic processes are: chemical denudation, piping within the slope and valley covers along with such slow mass wasting processes as creep and slumping. Another important denudational process in this zone is the disruption of the soil which results from tree-fall caused by frequent, catastrophic foehn winds. The transportation of dissolved and suspended material outside the Tatra Mts takes place in the stream channels but the amount of incision is weak.

The account of the geoecological belts presented above is a simplified one in which not all processes, but only the dominant ones, characteristic of these areas, are taken into consideration. The following are the most common processes: chemical denudation, surface wash and piping, fluvial erosion transport and deposition, as well as slow mass wasting. However, the processes characteristic of particular belts are not restricted to these belts only and the limits of the extent of these processes are usually not very distinct. Processes typical of the seminival belt occur in the forest belt provided suitable geomorphic, geological, and floristic conditions are present. Thus, frost weathering and rockfalls occur across the whole profile of the Tatra Mts. Only in their intensity and in the size of the area subjected to modelling by these processes does variation occur. The concept of geoecological belts and corresponding processes are thus simplified and idealized models of the real picture. The environment of the Tatra Mts, like any other high mountain environment, is a mosaic of interlinked natural complexes called ecotopes. Morphogenetic processes, especially gravitational ones, cross and conceal the boundaries of these belts. As well as the microclimatic and mesoclimatic differences related to aspect, lithology, slope gradient, and land use are the major causes of the mosaic-like character of the processes together with their spatial variation and varying intensities. Relative heights in the Tatra Mts of the order of 300–1000 m mean that processes acting in the higher zones affect the environment of the belts located below. These processes, called allochtonous (Klimaszewski 1962) are as follows: rock fall and rock sliding, debris flows, transportation by snow avalanches. Weathered material of the seminival belt is transported down to the alpine or even subalpine belts, and is responsible for a formation of allochtonous debris cones within them. The lack of a seminival belt in the Western Tatra results in a less complicated vertical structure of relief being developed there.

PRESENT-DAY MORPHOGENETIC PROCESSES

HIGH-MOUNTAIN SYSTEM OF SEDIMENT TRANSFER

The structure and operation of the high mountain morphogenetic system of the Tatra Mts demonstrates both vertical and horizontal differentiation. The differences in elevation above sea level are responsible for the development of four vertical belts in the High Tatra and three vertical belts in the Western Tatra. As a result of this a specific system of sediment transfer has developed in both mountain areas: a longer duration, more complex and higher intensity system in the High Tatra, and a shorter, more simplified, and lower intensity system in the Western Tatra. This is related to the operation of two basic morphogenetic subsystems: slope and channel ones.

The slope subsystem consists of various sequences of fundamental slope units (Fig. 13):

1) rockwalls and rocky slopes,
2) debris slopes,
3) debris-mantled slopes.

In the High Tatra the following sequences dominante: 1–2 and 1–3–2. In the Western Tatra, with a more mature relief, the following sequences dominate: 3–1–2; 3–1–3–2; 3–2, and finally 3.

The debris slopes in the Western Tatra are stabilized by vegetation and thus are generally inactive. But at the present time there is some small scale talus development and debris flow activity which is modifying these slopes.

The vertical extent of slopes of a given unit (especially rockwalls and rocky slopes in the High Tatra) is much larger than in the Western Tatra. These slopes in the High Tatra are usually located in two or three geoecological belts. This results in a longer transportation route, the supply of a larger amount of weathered material, and thus gives rise to morphogenetic processes which have a higher erosional potential. The vertical pattern of the glacial cirques hanging one above another and usually overdeepened caused that an independent and unrelated subsystem exist. The absence of direct linkage between the slope subsystem and the channel subsystem on the valley floors is well developed in the High Tatra. The particular topographic form of glacial cirques makes the large scale mechanical transfer of sediment

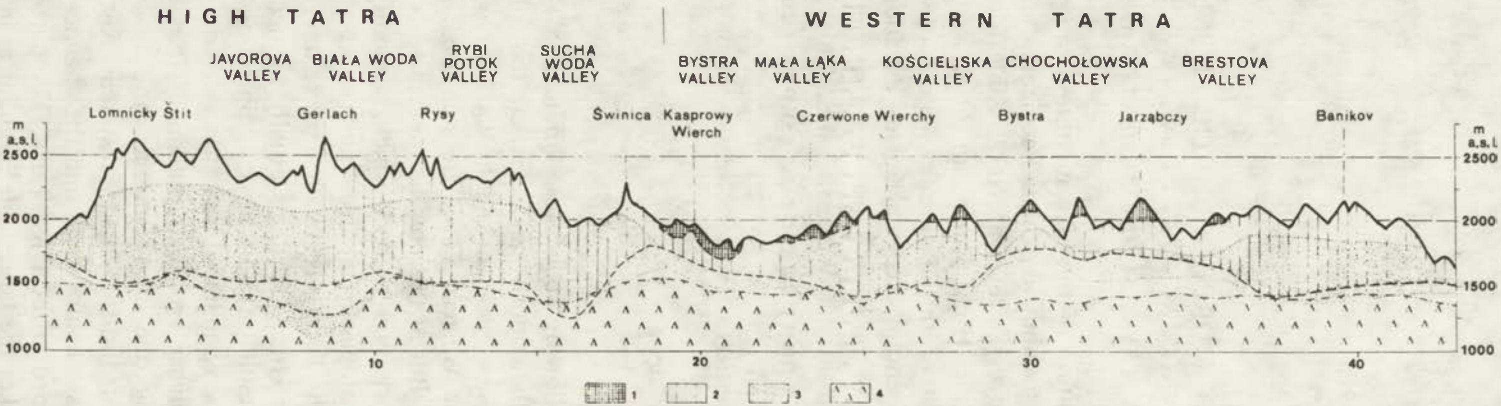


Fig. 13. High-mountain morphogenetic system of the Tatra Mts

1 – debris mantled slopes, 2 – rockwalls and rocky slopes, 3 – debris slopes, 4 – temperate forest morphogenetic system

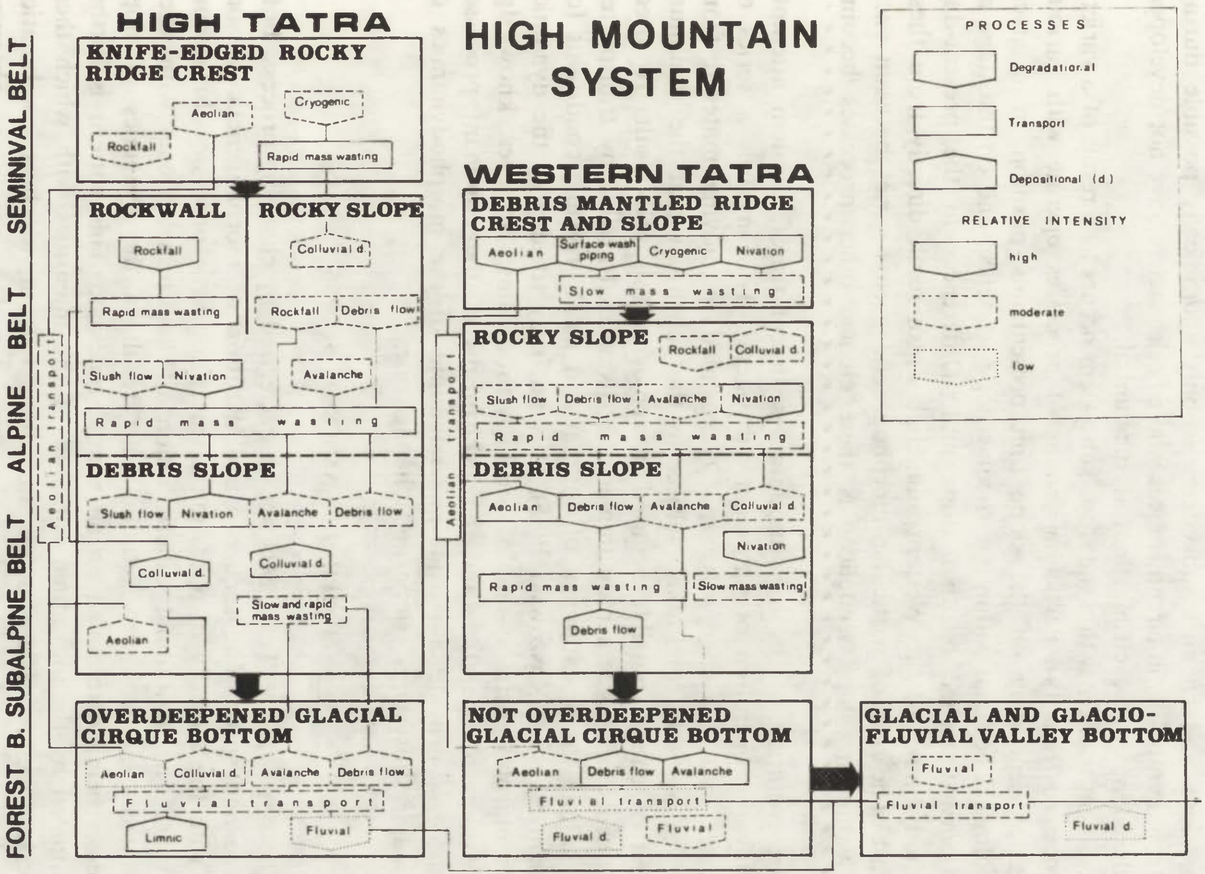


Fig. 14. The operation of the high-mountain morphogenetic system of the Tatra Mts

off the slopes onto the valley floors virtually impossible. In the Western Tatra the slope denudation subsystem also operates independently of the valley floor fluvial subsystem. The morphology of the cirque floors with complicated pattern of rock glaciers prevents a continuous transfer of sediment from the slopes through the cirque to the fluvial subsystem. The linking of the slope and channel subsystems is only sporadically possible during catastrophic rainfalls in circumstances where rock glaciers have not developed and the slope of the cirque floor is substantial.

Sediment transfer within the slope subsystem occurs as a result of a variety of processes (Fig. 14). Particular groups of processes operate with various intensities depending on the slope unit present, its position in a given geoecological belt and within a particular slope sequence, and on the stage of development reached by that slope unit. Differences in the present-day morphodynamics of the Western and High Tatra result directly from these contrast. The role of particular morphogenetic agents varies between each unit within the slope sequence, as these change other processes become dominant.

The evolution of the high mountain sediment transfer system attempts to unify the operation of particular subsystems. This involves a series of processes in which energy is rapidly expended in moving material from unstable to stable conditions, subject to the constraint that the minimum amount of energy possible be used in achieving this end result. A direct consequence of this is a reduction in the distances travelled by transported material and the development of optimal and most probable conditions for sediment transfer (Kaszowski 1985). It is not easy to identify the dynamics of the morphogenetic system of the high mountains. However, knowledge of the rates of operation and the mechanisms of the dominant processes can at least provide a partial understanding of the morphodynamics of two major subsystems: slope and channel ones.

MORPHODYNAMICS OF SLOPES

The slopes in the Tatra Mts are characterized by changing process, both in time and space. The sequence and duration of morphogenetic seasons vary with the intensity of operation of morphogenetic processes throughout the year. The altitudinal differentiation of landform systems with their associated climatic and vegetation belts result in the dynamics of slope processes being determined on an annual timescale. This pattern is further accentuated by the altitudinal zonation of the intensities with which these processes operate. There is an altitudinal belt in which a combination of high magnitude and short recurrence interval processes are observed. These areas undergo especially rapid erosion by events which are able to shape the slopes in such a manner that the resulting changes can be demonstrated without quantitative measurements. Table 8 illustrates the basic

Table 8. Dominant morphogenetic processes on basal debris slope forms in the Polish Tatra Mts

Principal factor	Principal transfer process	Principal slope form	Present-day activity in vertical climatic zones				
			temperate cool 900–1100	cool 1100–1550	very cool 1550–1850	temperate cold 1850–2200	cold 2200–2663
Gravity	falling, rolling, bouncing, sliding	rockfall talus	+	+ +	+ + +	+ + +	?
Snow	snow avalanching	avalanche talus	–	+	+ +	+ + +	+ +
	slow-moving snow				no information		
	sliding on snow surface		–	–	+ +	+ +	+ +
Snow and water	slush avalanching		–	–	+ +	+	–
Snow meltwater	ephemeral stream flow	alluvial talus	+	+	+ +	+ + +	+
Rain storm water	rainwash, debris flow		+	+	+ + +	+	+
Interstitial ice	internally induced mass movement of debris, creep	rock glaciers			–	–	
Freeze-thaw changes	creep	block slope	–	–	+ +	+	+
	frost creep, sliding	debris-mantled slope	+	+ +	+ + +	+ + +	+ +
		Activity index:	4	7	19	17	9

Process: – inactive, + weak, + + strong, + + + very strong.

debris slope forms and their dynamics presented by means of a relative scale. In order to amplify this result more fully the absolute values of the rates of operation of the major morphogenetic processes are presented in the following chapters. The altitudinal belt above the upper timberline (*ie* the area occupied by the dwarf pine and alpine meadows) is especially well-developed. This rates of operation of the dominant slope processes are presented below. These values have been determined on the basis of geomorphological field experiments.

CHEMICAL DENUDATION

Chemical denudation, *ie* dissolving and removal of mineral components by rain water and meltwater, is a common process. This particular process occurs at various rates depending on the lithology of the substratum. The rate at which this process operates is low on the igneous rocks comprising the High Tatra and on the metamorphic rocks occurring in the Western Tatra (Oleksynowa and Komornicki 1965). Mean annual rate of removal of mineral components in the Starorobociańska valley in 1976–1980 was c. 12 $m^3 \cdot km^{-2}$ (Krzemień 1985). Calcareous sedimentary rocks are subject to rapid chemical denudation and karst forms are the direct morphological result of it. The amount of denudation in each geoecological belt depends on the volume of circulating water and solubility conditions determined by the combination of environmental elements typical of a given belt (*eg* soil type, presence of humus acids, and the temperature-moisture regime). In the Western Tatra, built of limestone-dolomite complexes, the surface chemical denudation is differentiated in a vertical profile (Kotarba 1972). The smallest surficial denudation in the Czerwone Wierchy massif occurs in the seminival and alpine zones. Mean annual values for the high mountain system (cool temperate and very cool climatic zones) are of the order of 37 $m^3 \cdot km^{-2} \cdot yr^{-1}$, which when converted into an annual rate of surface lowering is 0.036–0.038 mm. The values of chemical denudation in the forest zone are twice as large (92.9 $m^3 \cdot km^{-2} \cdot yr^{-1}$) so the associated lowering is 0.095–0.085 mm *per* year (Fig. 15). Intensive solution in the forest zone is associated with the presence of a large amount of CO_2 of organic origin which reacts chemically with water and rocks. The surficial deposits in the forest zone are leached more easily than the massive rocks which dominate the uppermost zones. The geomorphological evidence of this process (Głazek and Wójcik 1963; Kotarba 1967, 1972) is provided by the complexes of karren which occur across the whole profile of the northern slopes of the Tatra. Chemical denudation is characterized by a large seasonal variability. The highest rates of chemical denudation occur in summer, during months with the largest rainfalls and during the spring snowmelt. The amount of chemical denudation is directly proportional to the amount of water reacting with the rocks (Kotarba 1972; Pulina 1974).

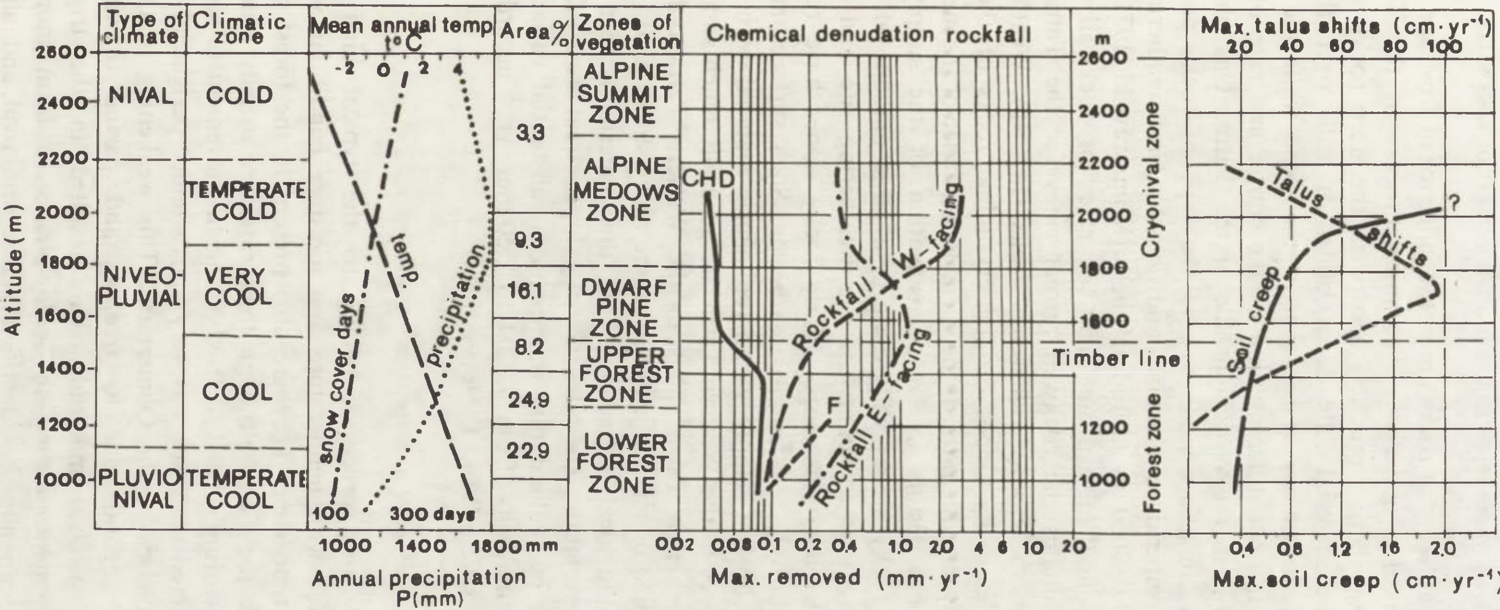


Fig. 15. The elevational zonation of climate and vegetation in the Western Tatra Mts in terms of selected morphogenetic processes (after Kotarba 1984)

PHYSICAL WEATHERING, ROCKFALL, AND THE RETREAT OF ROCKWALLS

Mechanical weathering of rocks under the influence of frost and insolation is, as with chemical denudation, a common process in the Tatra Mts. Removal of the weathering products occurs on the steep rocky slopes and rockwalls as loose material falls down, slides or rolls over the surface thereby enlarging the taluses at the foot of the slope while the rockwalls above retreat. A distinct differentiation of the course and rate of rockwall retreat is noticeable in the vertical profile of the Tatra. This phenomenon depends both on climatic conditions and the type of rocks present. The most favourable temperature-moisture conditions for frost weathering actually occur in the 1700–2050 m altitudinal belt (Klimaszewski 1971). Indeed, against a background of existing data for rates of rockwall retreat in temperate cool climates the rockwall retreat rates for the Tatra Mts are large, especially in the case of calcareous rocks (Fig. 15). Weathering and rockfall within the rocky slopes on the crystalline rocks of the Western Tatra record highly variable rates. If the variable resistance of the metamorphic and granitoid rocks results in an intensification of the supply process, then the mature rocky slopes will result in a large amount of weathered material remaining in the vicinity of the weathering site and not being carried down onto the talus slope. Many rocky slopes have reached the stage of being relatively resistant to weathering. Such surfaces are covered with lichens and do not supply weathered debris material with the result that the rocky slopes in the Western Tatra are generally much smaller in size than those of the High Tatra. Thus in the Western Tatra only a small amount of weathered material is supplied to the rocky slopes (c. 0.17 $mm \cdot yr^{-1}$ on average) causing local aggradation, but this material is then stabilized by the growth of vegetation. The rates of rocky slope retreat are c. 0.2 $mm \cdot yr^{-1}$. However, it must be realised that this includes substantial variation in the subalpine and alpine belts (from 0.0004 to 2.50 $mm \cdot yr^{-1}$ according to the measurements in 1975–1984, Table 9).

SLOW MASS WASTING

Slow mass wasting processes belong to the typical high mountain processes associated with temperature and moisture conditions which exist above the upper timberline. They are also present in the forest belt of the Tatra and outside the Tatra although their rates are usually much smaller there. The mass wasting related to frost action (*ie* movement of soil particles and coarse weathered material) was very active under periglacial conditions during the cold phases of the Quaternary. This weakening of the internal structure of the regolith due to freezing and partial thawing caused mobility of the surficial materials. This resulted in the triggering of gravitational movements during the cold phases to form morphological complexes which can now be identified in the seminival and alpine belts.

Table 9. Annual rate of rockwall retreat in the Tatra Mts in comparison with other mountain areas

Location	Bedrock lithology	Rate of retreat [$mm \cdot yr^{-1}$]	Research period [years]	Author
Pięciu Stawów Polskich valley (High Tatra)	granitoids	0.028–0.26	2	Rączkowski 1981
Starorobociańska valley (Western Tatra)	granitoid bodies within metamorphic rocks	0.0004–0.26 (mean 0.04)	1	Koszyk 1977
Starorobociańska valley (Western Tatra)	– „ –	0.7	7	Kaszowski, Krzemień (unpubl.)
Wyżnia Chochołowska valley (Western Tatra)	alaskites and granitoids	0.4	2	Mamica 1984
Mała Łąka valley (Western Tatra)	limestones, dolomites	0.10–3.0 (mean 0.84)	4	Kotarba 1972
Kärkevagge (Swedish Lapland)	mica schists	0.04–0.15	●	Rapp 1960
Zemmgrund (Austrian Alps)	gneis, metamorphic schists	0.7–1.0	Calculation of scree volume (no measurements)	Poser 1954

The latter is composed of the slope forms found above 1600 m (solifluction lobes, solifluction terraces and earth hummocks). According to A. Jahn (1958), the zone within which active sorted pattern ground (above 2000 m), solifluction (1800–2000 m) and earth hummocks (1500–1800 m) occur in the Tatra. This is also the case for other high mountains within the European temperate zone and especially in the Alps. These processes became much less active during the Holocene and numerous forms have become stabilized. Typical periglacial microrelief continues to develop at lower rates under the influence of cryonival processes. These cryonival processes are identified by active turf-banked terraces (Izmaiłow 1984; Kotarba 1976, 1983) especially well developed on the ridges of the Western Tatra, and by nivation hollows which are continuing to evolve (Kotarba 1976). Slow creep or flow of the weathered material (*ie* frost creep and solifluction) can be demonstrated on the basis of a long period of measurement. Maximum rates of movement due to frost creep are given in Figure 15. These are very small values, the largest value obtained on the northern slope of the Czerwone Wierchy being 1.7 $cm \cdot yr^{-1}$. The rate of movement decreases to 0.4–0.5 $cm \cdot yr^{-1}$ in lower locations in the forest belt. Information on the maximum movement of regoliths on slopes are derived from observations on ploughing boulders. Such boulders can commonly be observed on the slopes of the Western Tatra. They also occur frequently in the High Tatra on slopes covered with morainic material. On the regoliths derived from calcareous rocks the annual movement of the boulders are 0.14–3.25 $cm \cdot yr^{-1}$, with the mean for 30 boulders of various sizes being 1.3 $cm \cdot yr^{-1}$ (Kotarba 1976). On the weathered crystalline and metamorphic rocks the annual values are halved.

The surface layer of the soil in the alpine and subalpine belts is subject to displacement under the influence of needle ice. Diurnal movement during one freeze-thaw cycle can reach 60 mm. On an annual scale, however, this process can cause particles to be transported up to 380 mm (Gerlach 1959).

Creep also occurs on the talus slopes. Ignoring the rapid, short-term transport occurring over long distances as a result of snow avalanches or debris flows, one can observe a slow creep of the debris mass. This occurs at a steady rate although some fragments register larger movements if the slope consists of a series of independent units of debris material subject to processes with contrasting dynamics. Altitudinal differentiation of the maximum rates of the debris transport on the limestone-dolomite slopes is presented in Figure 15.

RAPID MASS MOVEMENTS

Debris flow gullies, levees and tongues, landslide scars, debris slides, and slush avalanches are very common features on the Tatra debris slopes. These types of rapid mass movement differ from each other in term of their magnitude and age. According to Kotarba and Strömquist (1984) 'debris flows and alluvial activity have become more important in recent centuries and now constitute the dominant slope process'. Critical precipitation values responsible for the formation of debris flows are related both to daily precipitation totals and the instantaneous precipitation intensity [$mm \cdot hr^{-1}$] considered as a percentage of the mean annual total. Initial mass movements take the form of shallow translational slides with a variety of critical threshold values depending on the season of the year. Lower thresholds clearly exist in the winter and spring whilst higher ones are characteristic of the summer and autumn. Thus, according to Alpine research (Govi 1984b) precipitation intensities of 5 mm *per* hour which constitute 10% of the mean annual precipitation will trigger shallow mass movements during the winter and spring. Similar precipitation intensities have to reach 16–17% of the mean annual precipitation total in order to cause similar mass movement during the summer or autumn. Soil moisture levels are higher in the winter and spring in comparison with the rest of the year. Thus, the critical precipitation values for generating mass movement are lower during the winter and spring. The susceptibility of the substratum to mass movement also varies according to the properties of different rock types. A general conclusion arising from studies in different mountain areas can be formulated as follows. Providing the amount of precipitation is held constant different amounts of geomorphological work can be done even if the lithological and geomorphological properties are similar (*eg* Govi and Sorzana 1980). Thus, each mountain area possesses its own rainfall threshold which should be determined on the basis of rainfall intensities and their temporal distribution and not solely on the basis of general precipitation totals. For example, on polar slopes in Spitsbergen (Longyear valley) and northern Scandinavia, debris flows are triggered when rainfall intensities exceed 2 mm *per* hour (Larsson 1982) together with antecedent rainfall of the order of 30–50 mm *per* day (Rapp and Nyberg 1981). Polar slopes are active since they are developed on fragile covers which are readily mobilized, moisture only collecting in a thin active layer above the permafrost. However, in the Tatra the slopes are more stable, have higher infiltration capacities and thus the rainfall threshold for slope failure is much higher. A rainfall of intensity of 20 mm *per* hour is required to trigger debris flows and shallow slides. Large-scale catastrophic debris flows observed in the Tatra in 1933 were caused by a summer rainfall intensity of 26 mm *per* hour and a daily total of 62 mm (Lukniš 1973). Statistically, such precipitation can occur every 2–4 years in the Tatra at the upper timberline (see Fig. 10). However, there are occasions when much higher precipitation intensities of the order of 40 mm *per* hour do not result in a large scale mass transport of debris unless they are preceeded by a rainy period, the duration of rainfall exceeds one hour and the daily precipitation total exceeds 50 mm. The magnitude and intensity of the mass movement generated by a given rainfall is also related to the morphometric properties of mountain relief (Kotarba 1976). Mountains with similar hydrometeorological properties can be subject to different types of morphogenetic processes. Geomorphological work depends on the manner in which rainfall and meltwater circulate within the regolith. Local variation in the relief of slopes can be more important than hydrometeorological contrasts. Relative altitudes, slope gradient, type and density of individual slope units (convergent, divergent or parallel planform) plus the granulometry of the weathering mantles, and their infiltration capacities determine a slope's response to a rainfall input. Once within the weathering mantle water can be subject to quick concentration along flow lines, triggering a failure of the weathering mantles either by rapid mass movement or by slow displacement over a short distance. According to Kotarba (1976) rapid mass movements in the Tatra Mts are much smaller in size when compared with corresponding phenomena in the Alps despite the fact that the latter are caused by processes operating with the same intensity. In the Tatra Mts the relatively small potential energy due to relief does not enable the release of a comparable amount of kinetic energy in flowing water as occurs in the High Alps. In the area of the largest debris flow gullies in Starorobociańska valley, during diurnal precipitation of 80– 100 mm, debris material is transported within a range of over 100 m while the volume of debris deposited below gullies is only c. 75– 100 m^3. In 1975– 1984 within the two largest gullies in the Starorobociański cirque flows of this size have occurred three times. Such flows very effectively transform the stabilised debris slopes and at present are the dominant process in glacial cirques especially in Chochołowska valley. Alpine diurnal precipitation values of the order of 50– 150 mm will initiate transportation of weathered material in debris flows with volumes of 400,000 m^3 (Aulitzky 1970) and even 500,000 m^3 (Govi 1984a). Similar precipitation totals in the High Tatra are capable of triggering debris flows in which the maximum transportated debris volume is 22,000– 25,000 m^3 (Midriak 1984). As a contrast, in polar regions the triggering of debris material even at much lower threshold values leads to the formation of debris flows with volumes only of the order of 2,000 m^3 (Larsson 1982).

AEOLIAN PROCESSES

Ridge crests and slopes situated above the upper timberline are modelled as a result of the sculpturing action of wind. Even at a ground surface velocity of 6.4 $m \cdot s^{-1}$, an excessively dry soil not stabilized by vegetation is subject to aeolian erosion. Mineral particles transported in the air or rolled over the surface are deposited far from their sources. Each foehn wind, *ie* wind of minimum speed exceeding 9– 10 $m \cdot s^{-1}$ (Orlicz 1954), can do geomorphological work. Usually, 100 such days are recorded annually at the Kasprowy Wierch summit. Thus, wind plays an important role in the transformation of relief in the subnival and alpine belts, especially when acting in conjunction with cryogenic processes and particularly needle ice activity. Such processes initiated on unconsolidated soil can spread very quickly onto surficial materials on the windward parts of slopes. The so-called turf-banked terraces (gelideflational terraces according to Klimaszewski 1978) are the result of destructive wind action and comprise systems of small terracets (*ie* treads and small risers incised into the soil cover). Soil particles blown from the uncovered surface abrade the surface thereby causing localized slope retreat. Deflation leads to irregular, asymmetric lowering of the ridge surfaces whilst the transported material is deposited on the valley floors (Kłapa 1963, 1980). The largest destruction of soil by aeolian processes occurs during the period of snow cover (October – November) when southerly winds exceeding 5 $m \cdot s^{-1}$ with gusts reaching up to 50–80 $m \cdot s^{-1}$ prevail. Then, aeolian erosion at rates of 10.5 to 130.4 g m^{-2} occurs. On an annual basis the mean for the subnival belt in the area of Hala Gąsienicowa reaches 87.9 $g \cdot m^{-2}$ (Izmaiłow 1984a). Detailed studies by B. Izmaiłow (1984a, b), based on direct measurement of the amount of the deflated soil particles, indicate that deflation in the alpine belt can reach even 163.7 $g \cdot m^{-2} \cdot yr^{-1}$. The deflated material can be deposited on valley floors at the lee side where 70.5 $g \cdot m^{-2} \cdot yr^{-1}$ on average are deposited on the debris slopes equivalent to an annual rate of accumulation of 0.0265 mm of fine material.

Wind is an important morphogenetic agent not only in subalpine and alpine belts but also in the forest belt where wind throw (*ie* the falling of trees with their root systems by wind) is the result of severe foehn winds. The soil is uplifted and retained on the roots. This process, called 'uprooted tree denudation', in exceptional cases (*eg* the catastrophic winds in May 1968) can cause vertical displacement of the soil of the order of 50,000 $m^3 \cdot m$ in one hectare of wind thrown forest (Kotarba 1970).

* * *

The processes discussed above influence slope development in a highly variable manner. In addition to different rates of activity related to geoecological belts there are also considerable contrasts within each belt. Some processes are limited spatially to specific slope surfaces; others occur linearly and are related to surface water circulation systems. Only a few operate on all the relief elements of a slope. All slope processes result in a reduction of mountain relief as a result of the erosion and transportation of weathered material from higher to lower sites. The displaced weathered material accumulates at the foot of the mountain slopes and frequently conceals older landforms related to the deglaciation of the mountain region.

MORPHODYNAMICS OF CHANNELS

The forms and present-day dynamics of stream channels in the Tatra are related to a large extent to glacial morphogenesis. Stream channels in valleys totally or partially glaciated in the Pleistocene were formed by proglacial streams in coarse moraine (average size 0.3– 1.5 m) and fluvioglacial (average size 0.1– 0.5 m) deposits. In contrast, in the non-glaciated sections (Bystra, Mała Łąka, Kościeliska, Chochołowska valleys) stream channels developed exclusively in fluvioglacial deposits (average size 0.1 – 0.5 m). Channels in glaciated valleys were transformed only by the largest floods in the Holocene. They are distinctly different from stream channels in valleys which were never glaciated. Such channels are incised in the solid substratum (dolomites, limestones, conglomerates, shales, marls) or in usually stony (average size 4– 10 cm) and polygenetic drift covers.

The zonation of relief in the Tatra and the climatic-vegetational belts superimposed on it, affect the patterns of water circulation, *ie* the hydrological regime of streams vary with altitude (Wit-Jóźwik 1974). In principal, channel systems are related to the zone of ground water storage and ground water flow which includes glaciated valley floors. Morainic and glaciofluvial covers within the valley floors regulate runoff in a similar way as lake systems in the High Tatra or highly variable underground Karst reservoirs in the calcareous part of the region. This is undoubtedly important in terms of channel morphodynamics as the storage capacity of the groundwater zone cancels the impact of floods. In the region of lower groundwater storage and of higher surface runoff middle mountain streams of the Tatra flow in rocky channels (Wit-Jóźwik 1974) and here floods are morphologically more efficient.

Four types of high mountain channels can be distinguished based on lithological, structural, morphostatic, morphodynamic criteria and on the relation of the channel to the form of valley floor. The last criterion is significant in terms of the channel's mode and rate of adjustment (Table 10). The high-mountain stream channel system includes channels situated in all the zones of the Tatra in valleys totally or partially glaciated in the Pleistocene. A general view of all types of channels in the Tatra enables a comparison to be made of present-day changes operating within the whole fluvial subsystem.

Table 10. High-mountain channel types in the Tatra Mts

Channel type	Relation to valley morphology	Relation to geology	Actual fluvial activity (stability)
1	within the glacial cirque	channel cut into slope deposits and morainic sediments or into solid rock	stable
2	within the glacial valley	channel cut into morainic deposits	stable
3	within the glacial valley	channel cut into morainic deposits reworked by rivers	moderately stable
4	within the unglaciated valley	channel cut into glaciofluvial deposits	moderately stable

CHANNELS IN GLACIAL CIRQUES

Channels in glacial cirques are poorly developed, up to c. 2–3 m in width and up to 0.5 m in depth, and are incised into slopes and morainic deposits. They are sinuous, especially within the cirque floor. Particularly in areas of rock steps in the High Tatra glacial cirques the channels are incised into bedrock and are characterized by gradients up to 40‰. Almost all the channels are intermittent in flow and transport a sandy-gravel bedload fraction. Sediment transport occurs rarely, on average once *per* 4–5 years (Table 11). The mean annual maximum distance of transport is very short (*eg* 0.26 m in the Starorobociański cirque in 1976–1980). The channels incised in the rock steps areas are very stable on account of the coarse debris material stored in the lakes.

Table 11. Bedload transport in the longitudinal profile of the Starorobociański stream channel during the floods of different magnitude

Hydrological year	Number of the morphologically active floods: in the lower channel section within the glacial valley	in the whole channel within the glacial valley	in the whole channel within the glacial valley and cirque
1976	2	–	–
1977	1	1	–
1978	3	1	1
1979	2	–	–
1980	4	2	1
1981	1	–	–
1982	3	–	–
1983	4	–	–
1984	4	–	–
1976–1984	24	4	2

CHANNELS IN GLACIAL VALLEYS

The channels within the glacial valley floors are varied and complex in formation. They consist of medium length (0.6–0.9 km), slightly sinuous, two-branched or anastomosing sections which cut through recessional frontal moraine ridges from the last glaciation. The longitudinal profiles of the channels were formed in stages as the glacier retreated. Postglacial transformation has not changed the longitudinal profile sections formed in this way. The more complex the morphology of the valley floor and the longer the channel system (*eg* Biała Woda valley) the greater the number of distinct channel sections (Fig. 16) that can be distinguished (Baumgart-Kotarba and Kotarba 1979). In areas with terminal moraines the channel gradients are higher, usually in excess of 100–150‰ whilst the maximum size of bed

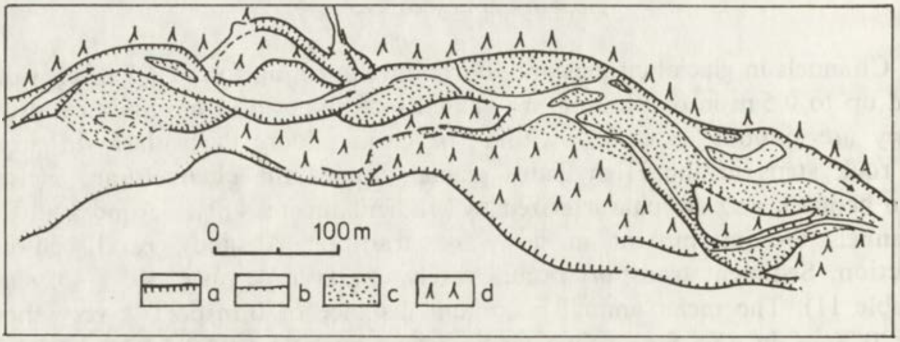


Fig. 16. Biała Woda channel system at the Roztoka valley mouth (after Baumgart-Kotarba and Kotarba 1979)

a – fresh undercutting of terraces, b – fresh undercutting of channel bars, c – alluvial bars, d – afforested alluvial plain

material reaches 1–3 m in the High Tatra and 0.5–1.5 m in the Western Tatra where it successfully armours the channels. Such sections are characterized by a high degree of stability (Fig. 17). The structure of the channels in the glacially less affected Western Tatra corresponds less to the

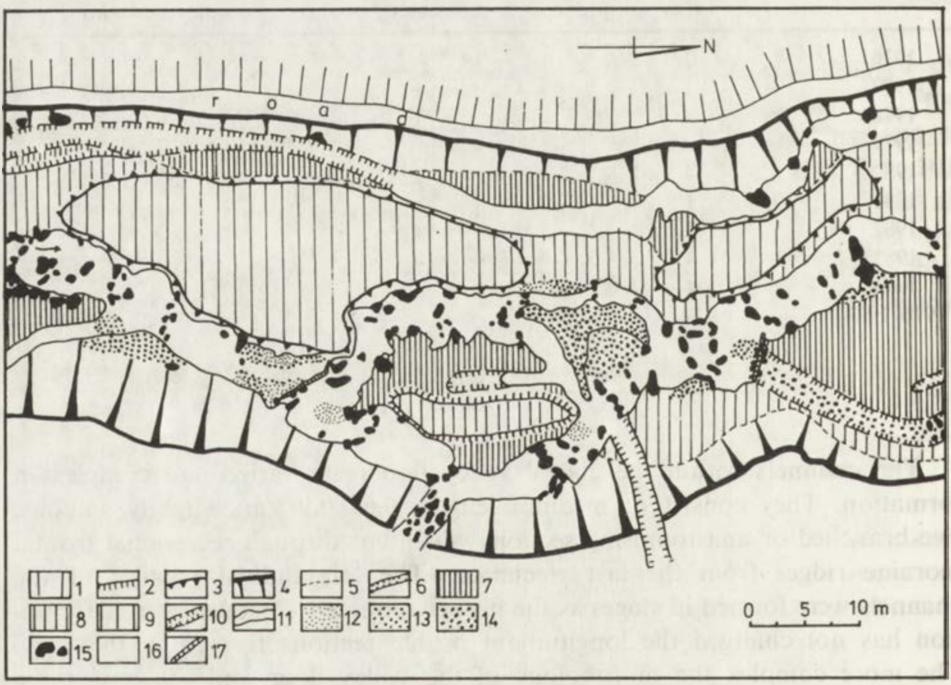


Fig. 17. Section of the Starorobociański stream channel

1 – valley slopes, 2–4– edges of the height up to 1 m (2), 1–2 m (3), 2–5 m (4); 5– concave breaks, 6– slope gullies, 7–9– bottom levels I, II, III, 10– episodically active channels, 11– extent of channel related to bankful stage, 12–14– bars built with material of the size: 1–5 cm (12), 5–15 cm (13), above 15 cm (14); 15-morainic boulders, 16– debris channel-bottom, 17– tree logs

distribution of recessional moraine ridges because of a shift in the limits of morphogenetic sections upward (Fig. 18). It should be noted that the channel systems in glacial valleys incised into the morainic materials are very stable as is confirmed by moraine boulders overgrown with mosses. The fairly high water retention within the morainic materials contributes towards a steady pattern of runoff and thus, decreases the morphogenetic impact of floods.

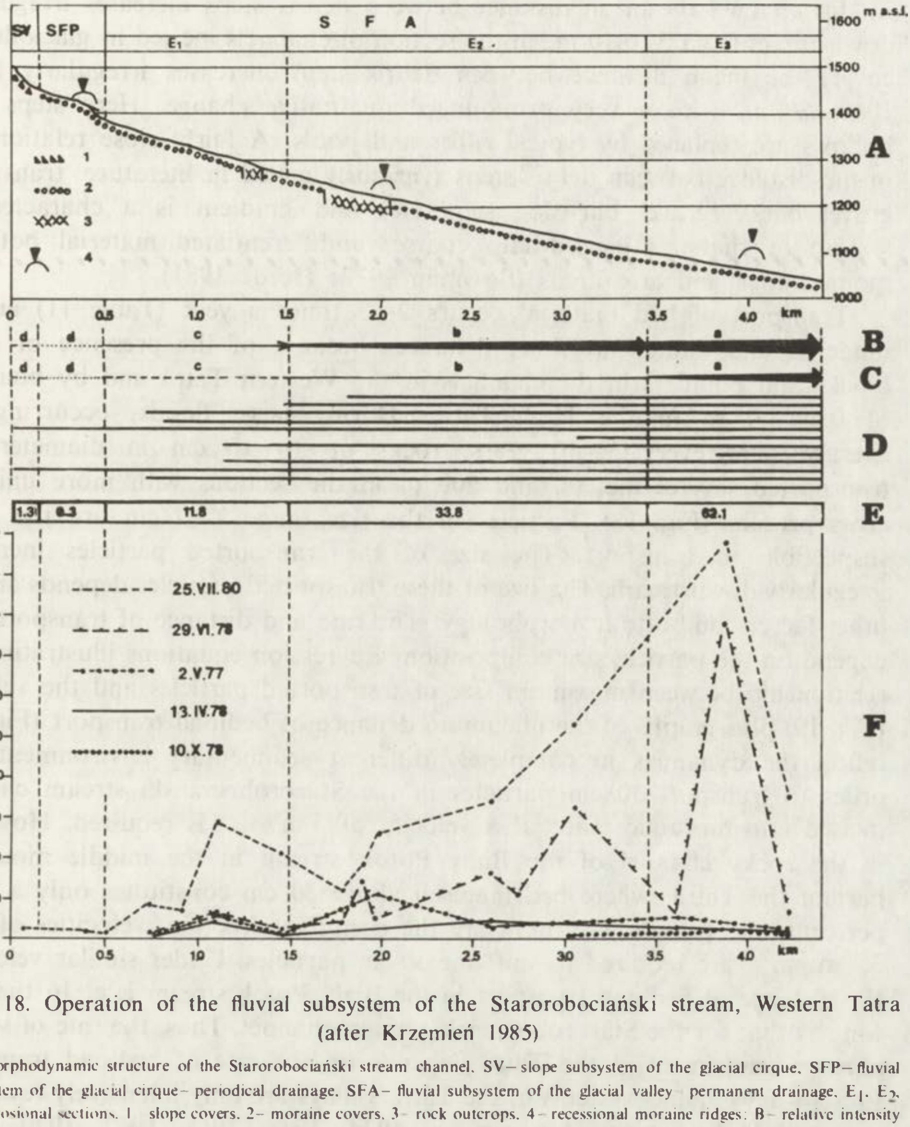


Fig. 18. Operation of the fluvial subsystem of the Starorobociański stream, Western Tatra (after Krzemień 1985)

A – morphodynamic structure of the Starorobociański stream channel. SV – slope subsystem of the glacial cirque. SFP – fluvial subsystem of the glacial cirque – periodical drainage. SFA – fluvial subsystem of the glacial valley permanent drainage. E_1, E_2, E_3 erosional sections. 1 slope covers. 2 – moraine covers. 3 – rock outcrops. 4 – recessional moraine ridges: B – relative intensity of bedload transport: C – relative intensity of bottom erosion (a – moderate, b – small, c – very small, d – extra small); D – the range of conditions for initial motion of bedload during morphologically active floods: E average values of maximum distance of bedload transport in particular morphodynamic sections; F – maximum distance of bedload transport (in metres) during floods of various magnitudes

The large number of steps and hollows in the channels incised into morainic materials probably demonstrates a tendency to creation of a certain balance between the channel form and the pattern of flow. K. Krzemień (1985) has noted a fairly close relationship between the gradient (x) and mean distance between debris steps (y) in the Starorobociański stream expressed by a linear equation:

$$y = -0.013x + 8.5 \text{ with } r = -0.63.$$

In this channel the mean distance between debris steps increases irregularly downstream from 5 to 10 m. In the section of channels incised in glaciofluvial covers the mean distance between debris steps increases irregularly from 10 to 80 m, with a very pronounced qualitative change. Here steps and hollows are replaced by typical riffles and pools. A fairly close relationship in the distance between debris steps (variously called in literature 'transverse gravel bars', 'gravel bar–log step' *etc*) and gradient is a characteristic feature of channels incised into coarse, undifferentiated material both in mountainous and arid areas (Bowman 1977; Heede 1981).

Transport of bed material occurs 2–3 times a year (Table 11) and is hindered and limited to short distances because of the presence of large blocks and boulders in the channels in the Western Tatra and by boulders of 0.6–3.0 m in the High Tatra. During large floods, occurring on average once every 5–10 years, rocks up to 16 cm in diameter are transported several meters, and 200 m in the sections with more uniform cross profiles (Fig. 18). Particles in the size range 2–7 cm are the most suspectible to transport. The size of the transported particles increases irregularly downstream. The size of these transported particles depends among other factors on bottom morphology. The rate and distance of transport also depend on the particle size composition. Regression equations illustrating the relationship between maximum size of transported particles and the velocity (Fig. 19) plus graphs of the maximum distance of bedload transport (Fig. 20) reflect the dynamics in completely different sedimentary environments. In order to transport 30 cm particles in the Starorobociański stream channel incised into morainic material a velocity of $5 \text{ m} \cdot \text{s}^{-1}$ is required. However, in the rocky channel of the Biały Potok stream in the middle mountain part of the Tatra, where bed material above 30 cm constitutes only a small per cent, and gravels 4–10 cm are the dominant fraction, velocities of only $3.7 \text{ m} \cdot \text{m}^{-1}$ are required to shift the 30 cm particles. Under similar velocities the distance of bedload transport in the Biały Potok stream is c. 16 times as long as that for the Starorobociański stream channel. Thus, the rate of stream channel adjustment in the Tatra depends on the rate of bedload transport. Present-day channel systems in the Tatra valleys are only intensively reworked during large catastrophic floods (*eg* 1934, 1948, 1958, 1962, 1970, 1972, 1973, 1980). Investigations into the morphological impact of the 1973 flood (which followed a rainfall with a 50 year return period) indicate that

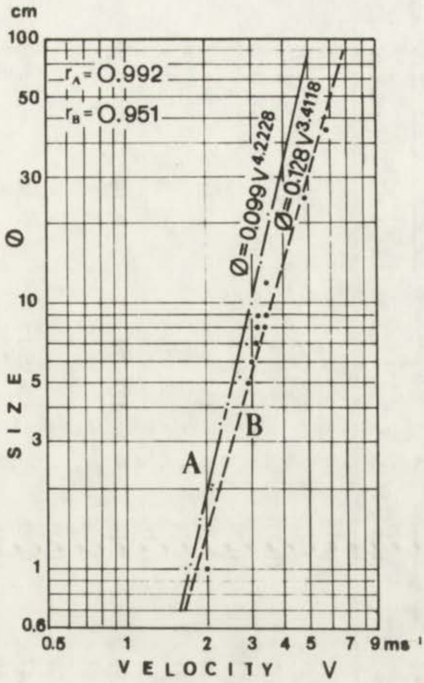


Fig. 19. Relationship between the size of the bedload transported material and maximum water velocity

A Biały stream, B Starorobociański stream

transport in morainic channels was highly site specific and the overall reworking rather limited. Only single bars formed by deposition 10–15 cm gravels were noted in the channel together with local bank erosion caused by the irregular distribution of boulders. In the channel of Sucha Woda in the High Tatra the size of the largest particles transported by the flood increased downstream from 34 cm in vicinity of Hala Gąsienicowa to 60 cm at the outlet from the Tatra. Boulders 70–100 cm in size showed only traces of abrasion on their proximal faces where lichens had been destroyed. During large floods occurring once every years boulders up to 45 cm were moved sporadically in the Starorobociański stream (Krzemień 1985). Mean annual removal of bedload material is twice as large as the supply there. Thus, the dominant process in this type of stream channel is one of very weak incision, fine material usually being completely removed. Similar processes, although on a larger scale, are also dominant in proglacial streams (Fahnestock 1963). The removal of fine material leads to an increased armouring of the stream bed which, in turn, causes a decrease in transport frequency and erosional efficiency. Such a natural mechanism also operates in other mountain crystalline massifs which have undergone glacial morphogenesis (caine 1974; Heede 1981).

Table 12. Balance of transported bedload material, suspension and dissolved material in the crystalline catchment of the Starorobociański stream in 1976–1980

Hydrological year	Kind of transported material	Annual load [t]	Percentage of total annual load	Denudation indices		Total annual load [t]	Total denudation indices	
				$[t \cdot km^{-2}]$	$[m^3 \cdot km^{-2}]$*		$[t \cdot km^{-2}]$	$[m^3 \cdot km^{-2}]$*
	bedload material	1.162	0.44	0.13	0.05			
1976	suspension	30.937	11.81	3.52	1.41	261.935	29.83	11.93
	dissolved material	229.836	87.75	26.17	10.47			
	bedload material	0.980	0.28	0.11	0.04			
1977	suspension	27.983	7.89	3.19	1.28	354.768	40.41	16.16
	dissolved material	325.805	91.83	37.11	14.84			
	bedload material	3.423	0.81	0.39	0.16			
1978	suspension	69.797	16.50	7.95	3.18	423.078	48.19	19.28
	dissolved material	349.858	82.69	39.85	15.94			
	bedload material	0.648	0.20	0.07	0.03			
1979	suspension	21.732	6.65	2.47	0.99	326.990	37.24	14.90
	dissolved material	304.610	93.15	34.69	13.88			
	bedload material	6.891	1.08	0.78	0.31			
1980	suspension	110.535	17.41	12.59	5.04	634.888	72.31	28.92
	dissolved material	517.462	81.51	58.94	23.58			

* Conversion of weight to volume according to the index 2.5 $g \cdot cm^{-3}$.

Table 13. Amount of denudation of crystalline high-mountain areas

Area	Kind of rocks	Author	Denudation indices		Basis of computation
			$m^3 \cdot km^{-2} \cdot yr^{-1}$	$mm \cdot yr^{-1}$	
Alps–Mount Blanc Massif Bossons valley	crystalline schists	J. K. Maizels 1978	630	0.63	$L_S + L_B + L_D$
Upper Rhine Lake Constance	crystalline	Müller, Forstner 1968 *vide* N. Caine 1974	–	0.014	$L_S + L_B$
Colorado Front Range Green Lakes Valley	granites, metamorphic rocks	N. Caine 1974	–	0.0043	$L_S + L_B + L_D$
Front Range Clear Creek	crystalline rocks	USGS 1959 *vide* N. Caine 1974	–	0.015	$L_S + L_B$
Scandinavian Mountains Kärkevagee valley	schists, gneisses	A. Rapp 1960	–	0.01	L_D
Western Tatra Mts Starorobociańska valley	metamorphic rocks, granites	K. Krzemień 1984	14	0.014	$L_S + L_B + L_D$

L_S–load of suspended material, L_B–bedload material, L_D–load of dissolved material.

At present the differences between the morphodynamics of specific channel sections in the glacial valleys are increased by contrasts in the rate and frequency of bedload transport (Fig. 18; Table 11). Bedload transport throughout the whole longitudinal profile in the glacial valley occurs on average once *per* 2–3 years (Table 11) and leads to the irregular but very small scale adjustment of particular sections. Over a short period these changes are not very pronounced in terms of their morphological impact.

Studies carried out in the Starorobociańska valley indicate that in terms of different components of total sediment transport the dissolved load is overwhelmingly the dominant component (81.5–93.2%) followed by the suspended load (6.6–17.4%) and the bed load (0.2–1.1%) components (Table 12). Even during the large floods this overall pattern remains unchanged with only some variation in the proportion of different modes of transported material. In general terms the sediment yield is smal, *ie* 261 tons in 1976 to 634 tons in 1980 were removed from the Starorobociański drainage basin which has an area of 7.8 km^2 (*ie* 29.83 tons · km^{-2}; Table 12). Denudation values for the Starorobociańska valley are c. 35 times smaller than those for the Bossons Alpine valley similar in size, incised in crystalline rocks and at the present time partially glaciated (Maizels 1978; Table 13). However, they are only 3 times greater than the values for the Green Lakes Valley in the Colorado Front Range (Caine 1974).

Downstream of the terminal moraines, where the valleys widen, the gradients locally decrease (Biała Woda 6 ‰, Polana Chochołowska 30 ‰) more particles up to 10 cm in size appear within the bed material and the proportion of large stable morainic boulders covered with mosses decreases (Baumgart-Kotarba and Kotarba 1979; Rączkowska 1983). These are sections

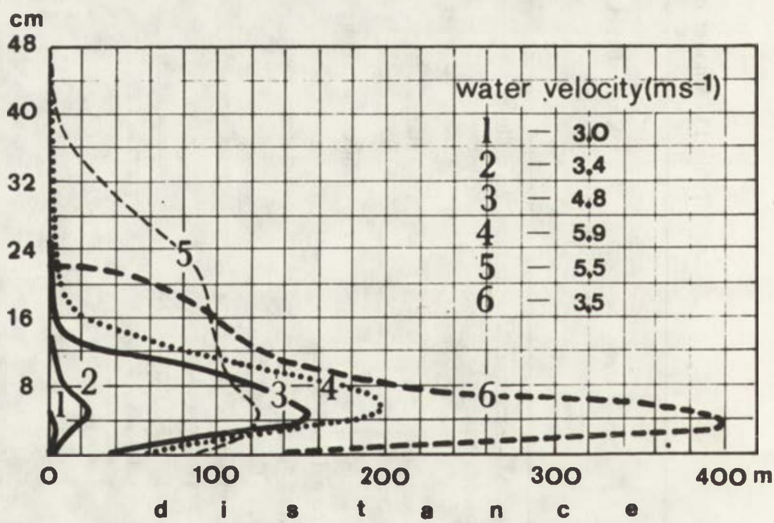


Fig. 20. Maximum distance of bedload transport within the channels of Starorobociański (1–4), Chochołowski (5) and Biały (6) streams

which are locally less stable, they are formed by deposition and redeposition of sediment and associated lateral erosion which, in turn, generates a braided channel pattern. Braided sections are typical of almost all the glacial valleys (Fig. 16). Their lengths vary from 0.15 to 4.8 km and they frequently divide the channel system within the glacial valley into 3 major segments of varying stability. It can thus be presumed that each section is formed in a generally similar way whilst differences between particular sections are at a detailed level and depend on the size of the catchment, valley width, and channel gradient (Fig. 18). Slight reworking can take place mainly during large catastrophic floods which occur on average once *per* 10–50 years. In the channels without morainic boulders bedload transport is very frequent and thus particles even up to 34 cm in size (Fig. 20) can be moved over longer distances. During the last 10 years in the Western Tatra there has been a tendency to abandon spare channels in favour of incision of the main channel. On the other hand in the High Tatra (Biała Woda valley) there is no evidence for such incision (Baumgart-Kotarba and Kotarba 1979).

CHANNELS IN FLUVIAL VALLEYS INCISED IN THE GLACIOFLUVIAL COVER

In valley sections with floors mantled exclusively with glaciofluvial deposits the channels become wider, their gradients decrease to 18–40‰ and they do not exhibit great variation in successive sections. The nature of such channels is mainly determined by the valley morphology (wider and more narrow sections) and the distribution of mouths of the tributary streams supplying sediment. The channels are slightly sinuous and braided in the wider valley sections. The maximum size of bed material decreases to 40–60 cm with boulders (80–100 cm) only occurring below the glacial valley section. The longitudinal profiles of the channels become more graded downstream. The increase in the proportion of finer material and its better sorting result in a more intensive pattern of channel adjustment. However, a trend towards channel stabilization dominates over the tendency for lateral channel migration (Rączkowska 1983). The distance of transport and the size of the transported material are slightly larger than in channels incised in morainic deposits. The material supplied from the undercut bases of terraces and from the side valleys is also larger. Generally, the channels incised into the glaciofluvial deposits seem to be less stable than the channels in the morainic material. Two or three sections with varying relative stabilities can be distinguished in the glaciofluvial sections (Rączkowska 1983).

MORPHODYNAMIC SCHEME OF THE TATRA CHANNEL SYSTEMS

Analysis undertaken on the channels in the valleys of the Tatra enable the construction of a general structural-functional classification (Kaszowski and Krzemień 1979). Channel systems are weakly linked with the slope systems. In the glacial valleys their lengths vary from 100 to 1500 m. The channels in the Tatra are generally characterized by a very high degree of vertical and horizontal stability. During observations over many years, significant channel changes have not been noted and only details in their topography have changed (Kaszowski 1973; Krzemień 1985). The most stable are those in the high-mountain part of the Tatra, incised in the bouldery morainic deposits and also in the coarse glaciofluvial deposits of the middle-mountain part of the region. In the valleys totally or partially glaciated during the Pleistocene the structure and functioning of the stream channels corresponds to the largest floods which formed them in the Holocene (especially in the glaciofluvial sections of the valleys). The glacial character of the boulder deposits is reflected in the channels and in the runoff regime regulated by groundwater (stored within the deposits) and systems of lakes or depressions of glacial and periglacial origin. These influence the rate of dissipation of energy in flowing water and the lower intensity of adjustment of the channel forms. Thus, it can be noted that the channels at the present time incised into the morainic sediments and glaciofluvial deposits create a high-mountain fluvial subsystem. This possesses a much larger but topographically lower areal extent than the slope subsystem. It controls the rate of removal of dissolved, suspended and bedload material to the Tatra foreland. The largest floods are important in terms of the transformation of this subsystem whereas medium and small ones play an important role in the removal of the total load.

The fluvial subsystem of the Tatra is characterized by: a high degree of stability throughout the subsystem and a low overall total sediment load in which the dissolved load is dominant. When compared to the fluvial subsystem in the areas of the middle-mountain parts of the Carpathians, the high-mountain fluvial subsystem of the Tatra is characterized by an unusual degree of stability. In the middle-mountain areas the channels are subject to more frequent and larger changes, and the total sediment load is much larger. Sediment load is somewhat larger in 'dry' years and several times larger in 'wet' ones. The percentage of suspended load within the total load increases to above 80% (Figuła 1966; Froehlich 1975; Froehlich, Kaszowski, and Starkel 1977).

ALPINE ROCKWALL–DEBRIS SLOPE PROCESS–RESPONSE SYSTEMS

The development of a young rock cliff in the high mountains starts at the moment when the inclined surface becomes free of ice. In the Tatra this has lasted at least 12,000–13,000 years and the mean lowering of crests during that period has been 5 m (Lukniš 1968). This measure of denudation of granitoid high-mountain relief of the High Tatra was calculated on the basis of the total volume of debris which had accumulated at the foot of alpine rockwalls. According to the theoretical considerations of Fischer (1866) and Lehmann (1933) the development of rock cliffs occurs as a result of parallel retreat and burying of the base of the slope by the talus. This leads, as a direct result, to the formation of a convex shaped fossil rocky slope. Morphometric analysis of alpine rockwalls in the High Tatra indicates a selective development. The retreat of rockwalls is not uniform and corresponds to fault zones related to mylonites and breccia or dislocation zones with plastic deformations. Therefore, the vertical profiles of recent cliffs formed by glacial over-steepening has been preserved in zones of especially resistant granodiorites. The gradients are always larger than 62–64° and these slopes are completely free of weathered material. Debris can accumulate on rocky slopes with smaller gradients (38–62°). Such slopes are numerous in the High Tatra, and are found above the vertical limits of the valley glaciers during the last glaciation. They are subject to strong degradation by denudational processes (Kalvoda 1974).

Three basic elements are distinguished within high mountain slopes:

(i) the highest, convex summit slope section with gradients of the order of 38–62° which occurs above the vertical limits of Pleistocene valley glaciers. The surfaces forming the ridge crest zone are covered with coarse mountain top debris and are the equivalents of *'arêtes en boulevard'* in the French Alps (Galibert 1960) or 'levelled, veneered rock surface with rock pinacles projecting through the scree veneer' in the New Zealand Alps (McArthur 1975). These surfaces provide evidence of periglacial relief and change abruptly into:

(ii) true alpine cliff (rockwall), vertical or overhanging, which provides material for:

(iii) the basal debris slope being formed at the foot of the rockwall. The intersecting walls of neighbouring glacier cirques have not element (i), and only narrow knife-edged ridge crests called *'arête en lames de sabres'* occur.

The highly varied morphology of the rockwalls is conditioned not only by rock resistance but also by various climatic factors. Well-insolated walls situated above 1800 m are characterized by strong chute dissection, smaller gradients and more vigorous cryogenic processes (Fig. 21). The walls receiving smaller insolation levels are steeper and weakly dissected. Variation in regelation processes has resulted in an asymmetry in the rockwalls. It is commonly

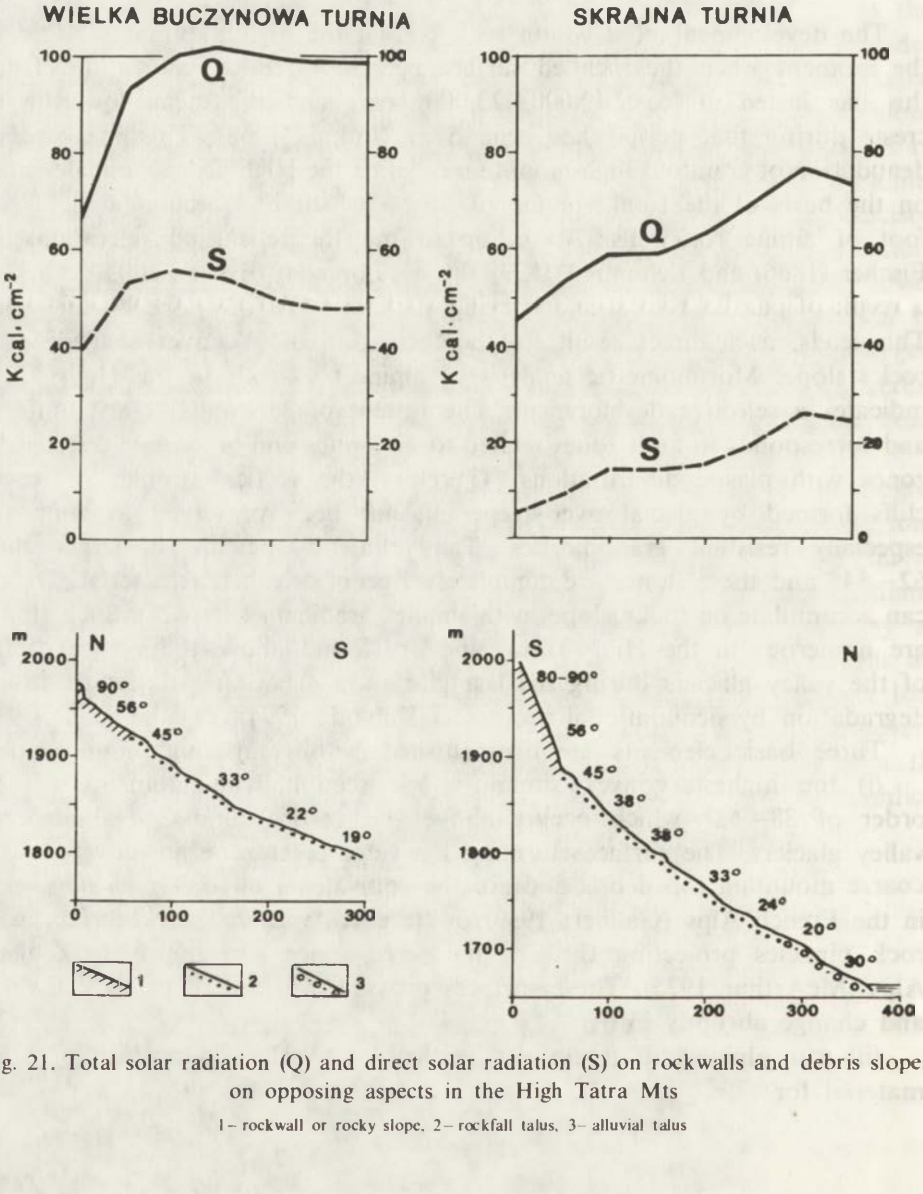


Fig. 21. Total solar radiation (Q) and direct solar radiation (S) on rockwalls and debris slopes on opposing aspects in the High Tatra Mts

1 – rockwall or rocky slope, 2 – rockfall talus, 3 – alluvial talus

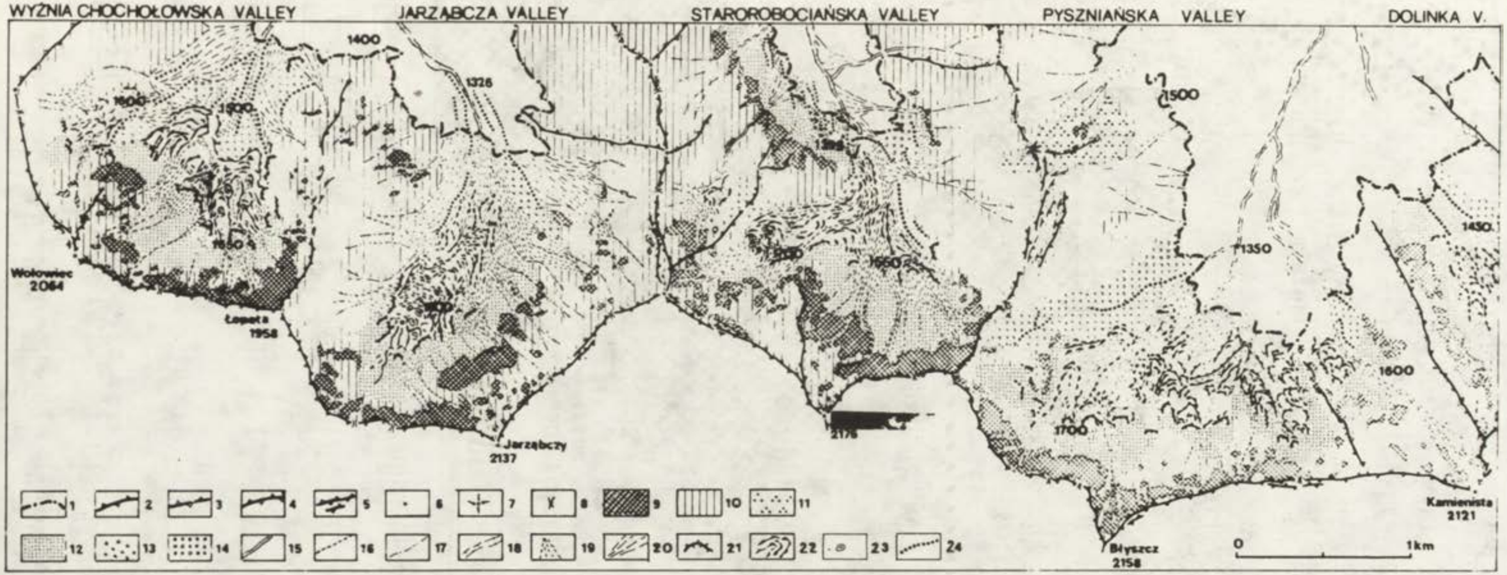
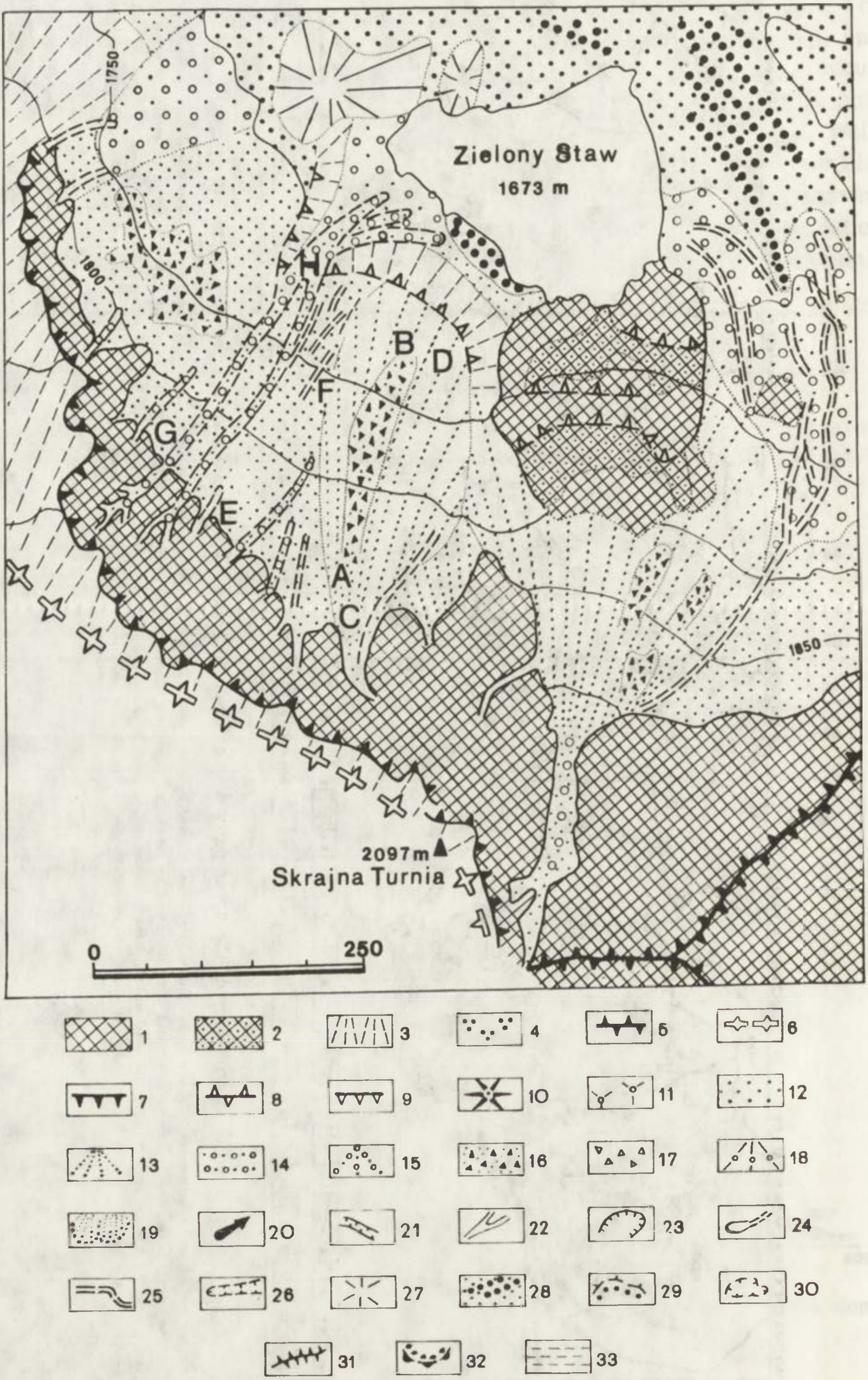


Fig. 22. Morphology of the upper parts of the Kościeliska and Chochołowska valleys, Western Tatra Mts

1– upper timberline, 2– sharp rocky ridge crests, 3– rounded ridge crests covered by debris and vegetation, 4– asymmetrical ridge crests (steep rocky slope from one side and debris-vegetation one from the other), 5– double ridge crests, 6– rounded summits covered by debris and vegetation, 7– sharp rocky summit, 8– passes, 9– rocky slopes and rockwalls, 10– debris mantled slopes, 11– block slopes, 12– debris slopes (rockwall, alluvial taluses) stabilized by vegetation and partly fresh, 13– fresh rockfall talus, 14– debris slope built of morainic deposits, 15– chutes cut in solid rock, 16– debris flow gullies, 17– inactive slope troughs and gullies, 18– stream channels, 19– periodically active debris flow cones, 20– inactive debris flow and stream cones, 21– inactive protalus ramparts, 22– fossil rock glaciers, 23– dead-ice hollows, 24– moraine ridges


Zielony Staw
1673 m
1750
1800
1850
H
B
D
F
G
E
A
C
2097m
Skrajna Turnia
0
250
1
2
3
4
5
6
7
8
9
10
11
12
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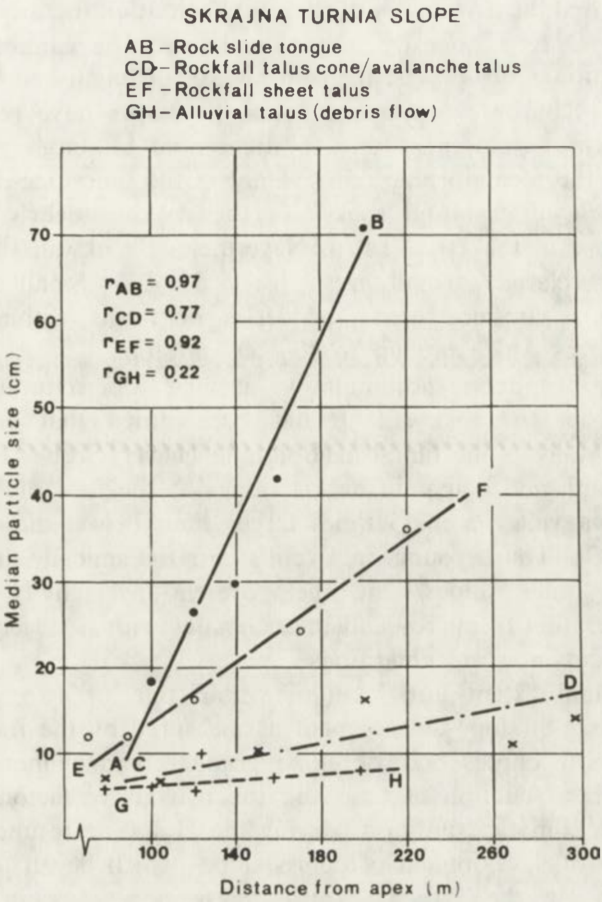

Fig. 23. Geomorphological map of the Skrajna Turnia experimental alpine cliff and debris slope and correlation between particle size and distance from apex (on debris slope) (after Kotarba and Strömquist 1984)

1 – rockwall or rocky slope, 2 – debris-mantled slope, 3 – Richter-denudation slope stabilized by vegetation, 4 – block slope, 5 – sharp, rocky ridge crest, 6 – rounded ridge crest covered by debris and vegetation, 7 – convex break on rockwall or on rocky slope, 8 – narrow rocky ridge crest partly covered by debris, 9 – convex rounded break above rockwall or rocky slope, 10 – sharp, rocky summit, 11 – mountain top detritus with sorted polygons, 12 – rockfall gravity sorted talus slope, 13 – rockfall gravity sorted talus cone, 14 – alluvial talus slope, 15 – alluvial talus cone, 16 – rockslide/rockfall slope (Holocene), 17 – rockslide/rockfall slope related to Late Glacial period, 18 – avalanche accumulation cone, 19 – talus slope with relict sorted stripes, 20 – rockfall routes, 21 – rocky gorge, 22 – chute cut in solid rock, 23 – nivation niche cut in solid rock (deep), 24 – debris flow gullies with levees and tongues cut in debris cover, 25 – high-mountain stream, 26 – slush avalanche gullies, 27 – roche mutonée, 28 – glacial drift deposits forming distinct morainic ridge, 29 – steep toe of recessional moraine ridge, 30 – undrained depression within glacial drift deposits, 31 – protalus rampart, 32 – fossil rock glacier, 3 – fluvioglacial river-buil plain

observed that in the Tatra the N- and E-facing slopes are steeper while the southern and western ones are more gentle (Halicki 1933; Ksandr 1954; Kotarba 1976). Provided the same denudation base of the slopes oppositely faced one can find the asymmetry of their fragmentation (density of channelling of the cliffs). Different density of rockwall dissection cannot be explained by the dislocation zones or systems of cleavages distinguished by Grochocka--Piotrowska (1970). In the Western Tatra the slopes have reached a more mature stage of development due to differential although generally lower resistances in the metamorphic rocks. Only in the limestone-dolomites areas and those built of granitoid rocks are the rockwall-debris slope systems similar to those in the High Tatra. Nevertheless, rockwalls in the Western Tatra occupy relatively small areas (Figs. 13, 22). Small rockwalls and pinacles form a mosaic pattern of series of forms within the strongly dissected surfaces which are up to 40–50° in slope.

Varied rates of debris accumulation at their foot naturally result from fragmentation of the rockwall. It has been noted that given a similar height in the walls in the limestone-dolomite glacier cirques of the Western Tatra, the length and height of debris slopes at the foot of W-facing walls in the Mułowa valley are 2.5 times larger than below the opposite walls (Kotarba 1983). Thus, young rock cliffs are dynamically integrated with debris slopes (talus slopes) and their present-day activity results from:

(i) the operation of micro- and mesoclimates with a variety of radiation, temperature and moisture conditions;

(ii) altitudinal extent and location within one or two climatic belts;

(iii) the stage of slope development as measured by the fragmentation of the rockwalls by chutes and by slope gradient and shape. The stage of development is a function of time and the two above factors.

Changes in climatic conditions during the Holocene resulted in a change in the pattern of development of debris slopes which began to be reworked by alluviation. In the Western Tatra debris slopes are in an advanced stage of development and are usually stabilized by vegetation. These debris slopes are strongly reworked by debris flows (Fig. 22). The lower sections of these slopes are typically aggradational ones and they are built of fine, clayish material derived from wash processes operating on debris materials located higher up the slope. They accumulate in depressions and troughs within cirque floors. Thus, the resulting debris slopes are concave in profile. In the upper section their gradients are c. 35°, in the middle 25° and in the lower section 18–22°. Their transformation by debris flows should mainly be related to the Subatlantic period and especially to the last few centuries when the most intensive exploitation of the high mountain environment by man has taken place. The slopes on the upper part of the Kościeliska valley where the grazing of sheep ended at the earliest in the 1930s, are now totally stabilized and their reworking is very small. Furthermore, the slopes had already been stabilized in the Atlantic period or even in the

Boreal since the climatic-vegetation conditions were more favourable for their stabilization than is the case at the present time.

Postglacial slope deposits were formed under the control of various factors which resulted in the formation of the 6 principal slope units presented in Table 8. A characteristic feature of debris slopes is the simultaneous occurrence of major slope units differing from each other in terms of grain size composition, sorting, and mode of present-day formation (Fig. 23). At present active and non-active or partially active slope surfaces (Fig. 23) occur adjacent to each other.

In terms of the pattern of development of one slope during the Holocene, changes both in its profile and sedimentological preperties can be noticed. Accumulation of rockfall talus by alluviation leads to the elongation of the slope and a decrease in its gradient. The base of alluvial talus is wider and the talus itself is composed of finer material (Fig. 24). In the metamorphic rocks of the Western Tatra particle sizes within debris flow deposits are definitely smaller (Fig. 25).

The active morphogenetic processes on basal debris slopes depend on present-day climatic variables. Thus, the mode and rate of relief transformation varies with altitude above sea level. The largest changes occur in the very cool and temperate cold climatic belts where the highest process rates have been noted (Table 8). Both in the highest cold belt and in the lower cool belt, the processes operate at a lower rate. The low rate of activity of processes in the cold belt is associated with severe climatic conditions which are characterized by a large number of days with snow cover, a small number of freeze-thaw days, and slightly lower precipitation totals. In the two lowermost

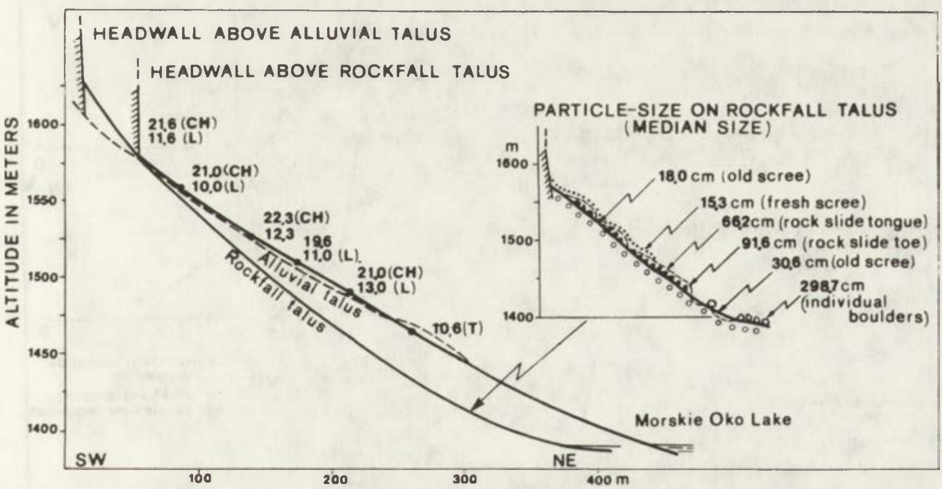

Fig. 24. Profiles through the alluvial talus and rockfall talus south of the Morskie Oko lake

Alluviation of debris slopes give changes of their morphometric and morphographic parameters as well as particle sizes. Median particle sizes along flow channel (CH), the levees (L) and tongue (T) is shown on the profiles (after Kotarba and Strömquist 1984)

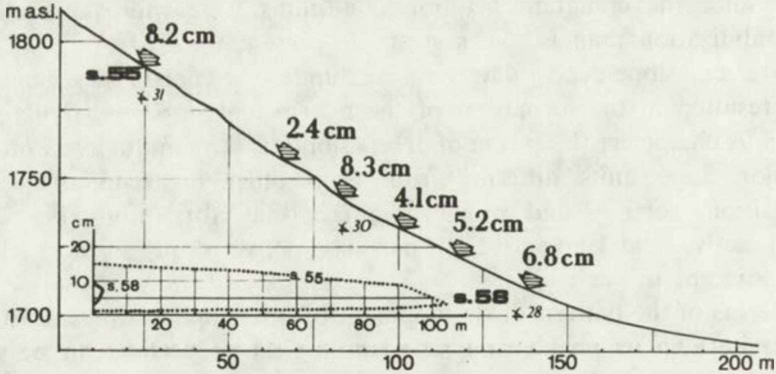

Fig. 25. Longitudinal profile of the debris flow gully in the Dudowy cirque, Western Tatra
Median particle sizes and maximum distance of debris flow transport are shown on the graphs. Period of measurements: 22.08.1983 – 25.08.1984

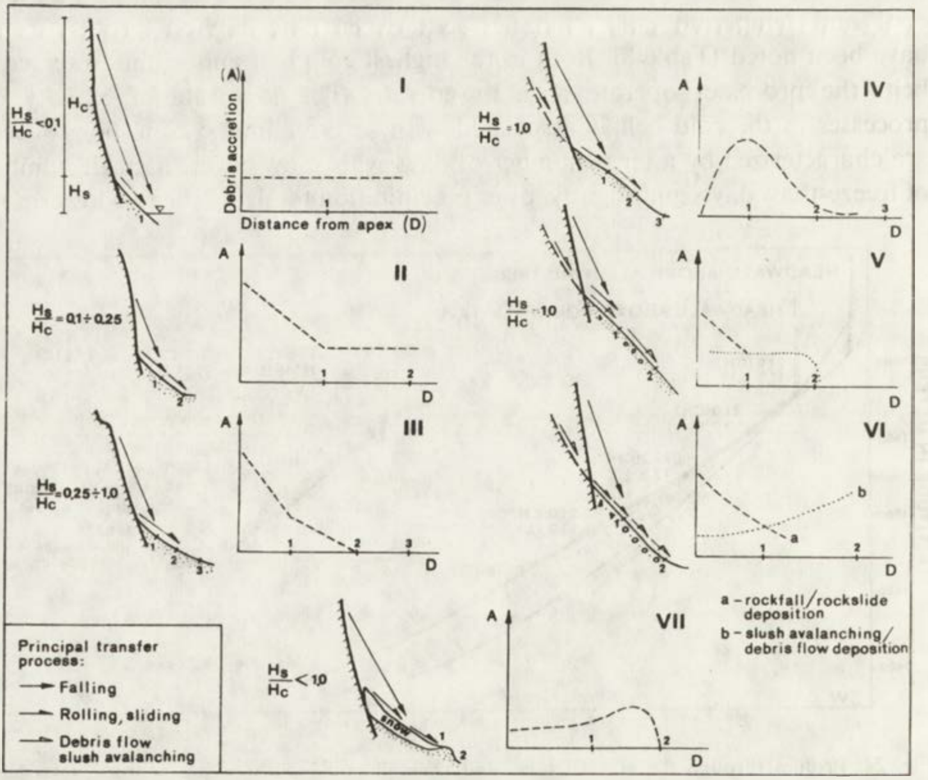

Fig. 26. General dynamics of the debris slope surfaces related to rockwall morphometry in the High Tatra Mts

belts the substratum is stabilized by forest cover. This, taken together with climatic controls, favours a lower rate of operation of morphogenetic processes. The mode and rate of debris slope formation are closely related to the morphological and climatic conditions on the rockwalls. There are 7 types of process-response rockwall-debris slope systems which are typical of the High Tatra (Figs. 26, 27). While in the Western Tatra 5 types of such systems can be distinguished showing the present-day formation of stabilized debris slopes in relation to the formation of rocky slopes and the debris slopes located above (Figs. 28, 29).

Type I. The youngest rockwalls situated above 1700 m comprise elements of the glacier cirque relief system and reach high relative relief of the order of 450–500 m. They were the last areas to became free of ice. When facing to N and E, they are vertical or overhanging. At their bases there are typical talus slopes formed from loose debris which is very mobile and subject to movement even under the pressure of a shoe or a rolling boulder. The ratio of the height of the talus slope (H_s) to that of the rock cliff (H_c) is always less than 0.1. As a rule they are very well sorted (coefficient or sorting $r = 0.98$). Talus slopes, as a whole, are

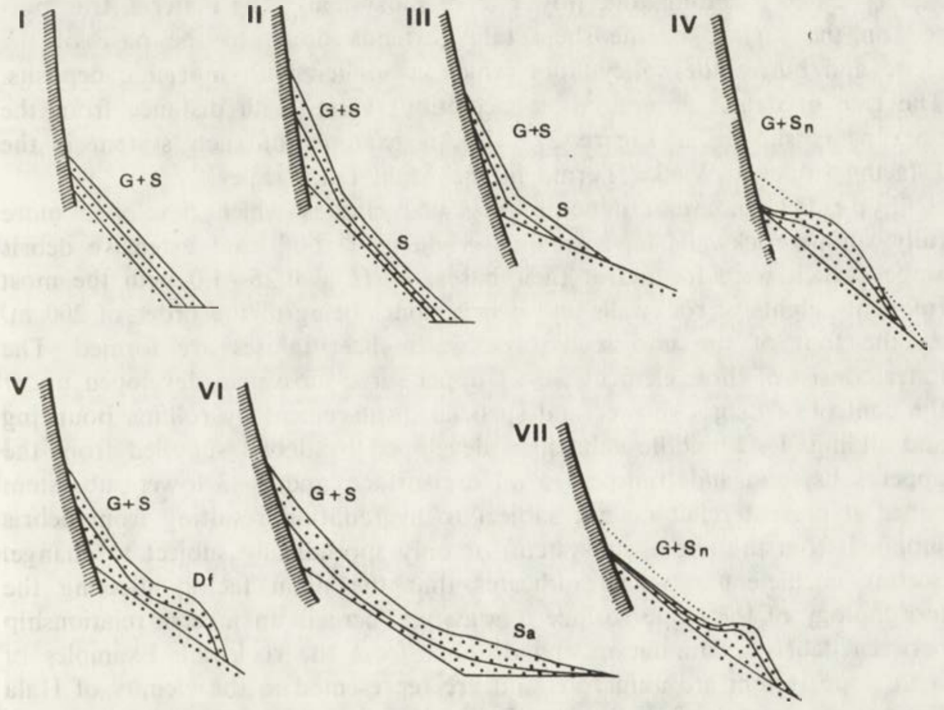

Fig. 27. Most frequent patterns of the accumulation of material on debris slopes in the High Tatra Mts

Supply of material by gravity (G), sliding on debris surface (S), sliding on snow surface (Sn), by debris flow (Df) and slush avalanching (Sa) subject to movement triggered by debris showers (rockfalls). The material supplied from the rockwalls slides over the surface and is only gravitationally sorted (Fig. 27–I). The rate of debris accumulation is uniform along the whole talus slope. If the glacial cirques are filled with lakes, transport of debris material continues under subaquatic conditions which by infilling leads to a decrease in the lake basin's capacity. Talus slopes are developed in a form of regular talus cones or sheets. This type is represented by the E-facing slope of Kazalnica above Czarny Staw pod Rysami lake.

Type II. Alpine rockwalls forming glacial troughs in the main valleys of the Tatra reach relative heights of 350 m. In contrast to type I rockwalls, they possess a profile consisting of at least two elements: the upper one with the smaller gradient is situated above the vertical limits of the valley glaciers, and the lower one has a vertical or nearly vertical profile. The ratio of the height of the talus slope to the height of the rock cliffs is determined by an index within the range 0.1–0.25. Rockwalls are usually undissected and the correlation coefficients measuring sorting on the sheet talus are of the order of $r = 0.90$. The weathered material on the rockwalls which falls onto a talus cone has a limited size range. Thus, two zones are distinguished: 0–1 debris sliding zone also called the upper scree subsystem, and 1–2 debris sliding zone (lower scree subsystem). The material transported on the surface of the sheet talus extends down to the base of the slope and buries the valley floor which is infilled with morainic deposits. The rate of debris accumulation (accretion) varies with distance from the apex as is shown in Figure 27–II. An example of such system is the E-facing slope of Wielka Turnia in the Mała Łąka valley.

Type III. In areas of complex glacial cirques which developed more fully, alpine rockwalls have smaller gradients ($<60°$) and extensive debris slopes which were formed at their bases. $H_s/H_c = 0.25-1.0$ with the most frequent heights of rockwalls and debris slopes being of the order of 200 m. At the foot of the undissected rockwalls sheet taluses are formed. The latter consist of three elements: 0–1 upper scree subsystem developed under the control of debris shower and surficial displacement by rolling, bouncing and sliding, 1–2 middle subsystem developed by debris supplied from the upper subsystem and transported on the surface, and 2–3 lower subsystem either at present relict (*ie* not subject to aggradation resulting from debris supplied from the above subsystem) or only sporadically subject to change. Sorting coefficient $r = 0.90$ indicates that the main factor affecting the morphology of the slope surface is gravity. There is an inverse relationship between debris accumulation and distance from the rockwall. Examples of such slope system are numerous and are represented in the vicinity of Hala Gąsienicowa and in Pięć Stawów Polskich valley (Photo 1 and 2).

Type IV. Alpine rockwalls within glacial cirques or rockwalls of glacial troughs subjected to differential erosion related to systems of cleavages and dislocation zones display a modified relief. These rockwalls are densely

Photo 1. The Skrajna Turnia experimental slope in the Hala Gąsienicowa, High Tatra Mts

From the left to right: rock slide tongue on rockfall talus cone, rockfall talus sheet and alluvial talus channelled by debris flow gullies are to be seen (*cf* Fig. 23)

Photo M. Kot

channelled by chutes which are initiated in the broad rock niches beneath the ridges and supply many weathered products. They form large talus cones at the base of the cliffs and are 200 m or more high. $H_s/H_c = 1.0$. As in the type III cliffs, three elements can be distinguished within the debris slopes. The coefficient of sorting is only $r = 0.80$ because the formation of the surface occurs in response to gravity and dirty snow avalanches. Measurements of fresh debris over a few years have shown (Kotarba 1981) that there is a different relationship between the rate of talus cone accumulation and distance from the rockwall. This relationship can be expressed by a second order polynominal:

$$y = a + bx + cx^2$$

where y – magnitude of accretion and x – distance from rockwall (Fig. 26). This demonstrates that a relatively small amount of debris accumulates at the cone apex. The rate of accumulation reaches a maximum in the middle part of the cone and then frequently decreases to zero at the cone base. This type of relation is well supported by the fact that the main period of material delivery from the rockwall to the debris slope occurs in spring

Photo 2. West-facing alluvial talus slope of Żółta Turnia Mt near Czarny Staw Gąsienicowy lake, High Tatra Mts (*cf* Fig. 34)

Photo A. Kotarba

(May – June). During that period the snow cover starts to disappear. The middle parts of the slope are subject to the most frequent thawing. The rockfall (debris shower), related to spring thawing on the cliffs, falls on the debris slope close to the apex zone. The debris then slides on the firn surface towards the middle part of the slope where it stops on making contact with the rough debris slope surface which has no snow cover. The largest rate of debris accumulation is observed here. Some of the debris rolls further down and is deposited in small volumes at the base of the cone. Quantitative measurements were undertaken on the talus cone below Skrajna Turnia at Hala Gąsienicowa (Figs. 27, 30, Photo 1).

Type V. Alpine rockwalls within glacial cirques or glacial troughs are highly dissected by chutes. These are formed as a result of selective erosion corresponding to narrow zone of cleavages. Sheet taluses have been formed at their base as a result of the interlocking and joining of relatively

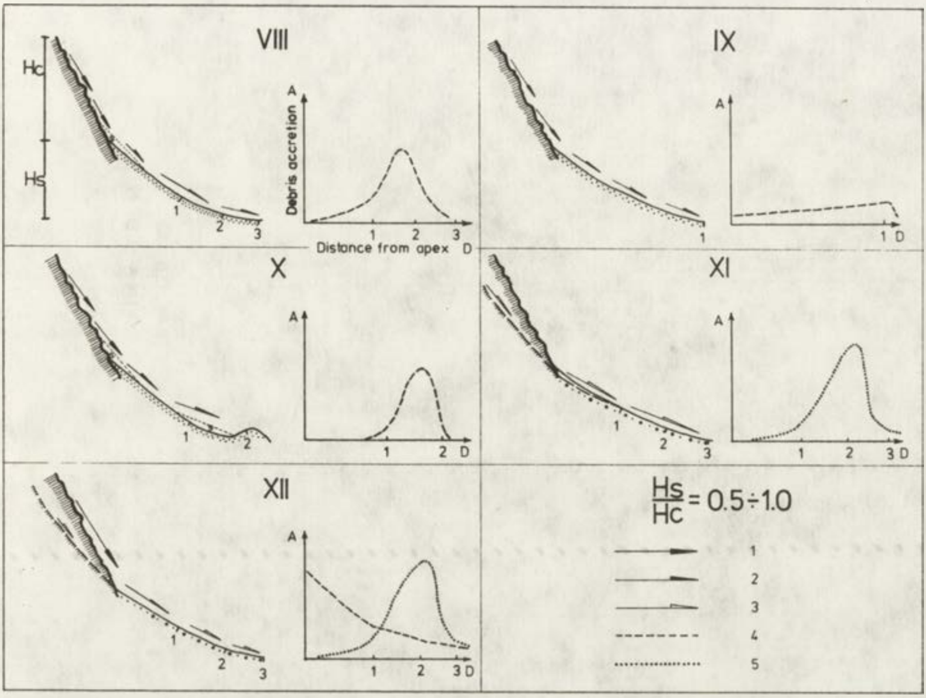

Fig. 28. General dynamics of the debris slope surfaces related to rocky slope morphometry in the Western Tatra Mts

1- falling, 2- rolling, sliding, 3- debris flow, slush avalanching, 4- rockfall/rockslide deposition, 5- debris flow deposition

small adjacent talus cones. At the front of the deep chutes dissecting the rockwalls, the sheet taluses are dissected by debris flow gullies. Such slopes develop both in response to gravitational and debris flow processes. Section 0–1 represents debris accretion caused by deposition of debris flow products. Sorting coefficients are low ($r = 0.20-0.50$). These systems are very common

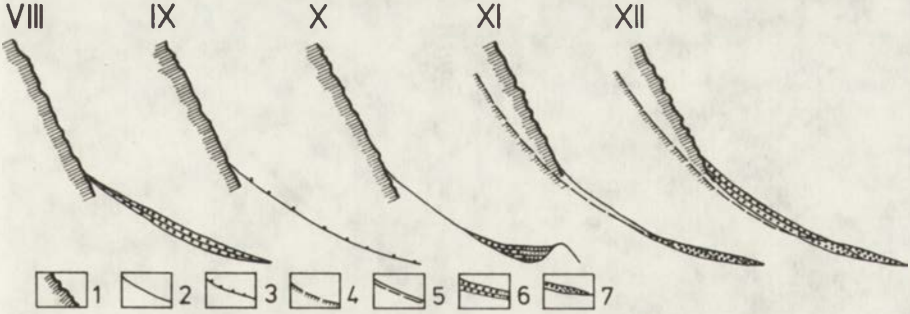

Fig. 29. Most frequent patterns of the accretion of material on debris slopes in the Western Tatra Mts

1 – rocky slope, 2 – debris slope surface most frequently stabilized by vegetation, 3 – sliding of single particles downslope, 4 – rocky chute, 5 – debris flow gully, 6 – deposition of material by gravity, 7 – deposition of material by debris flows

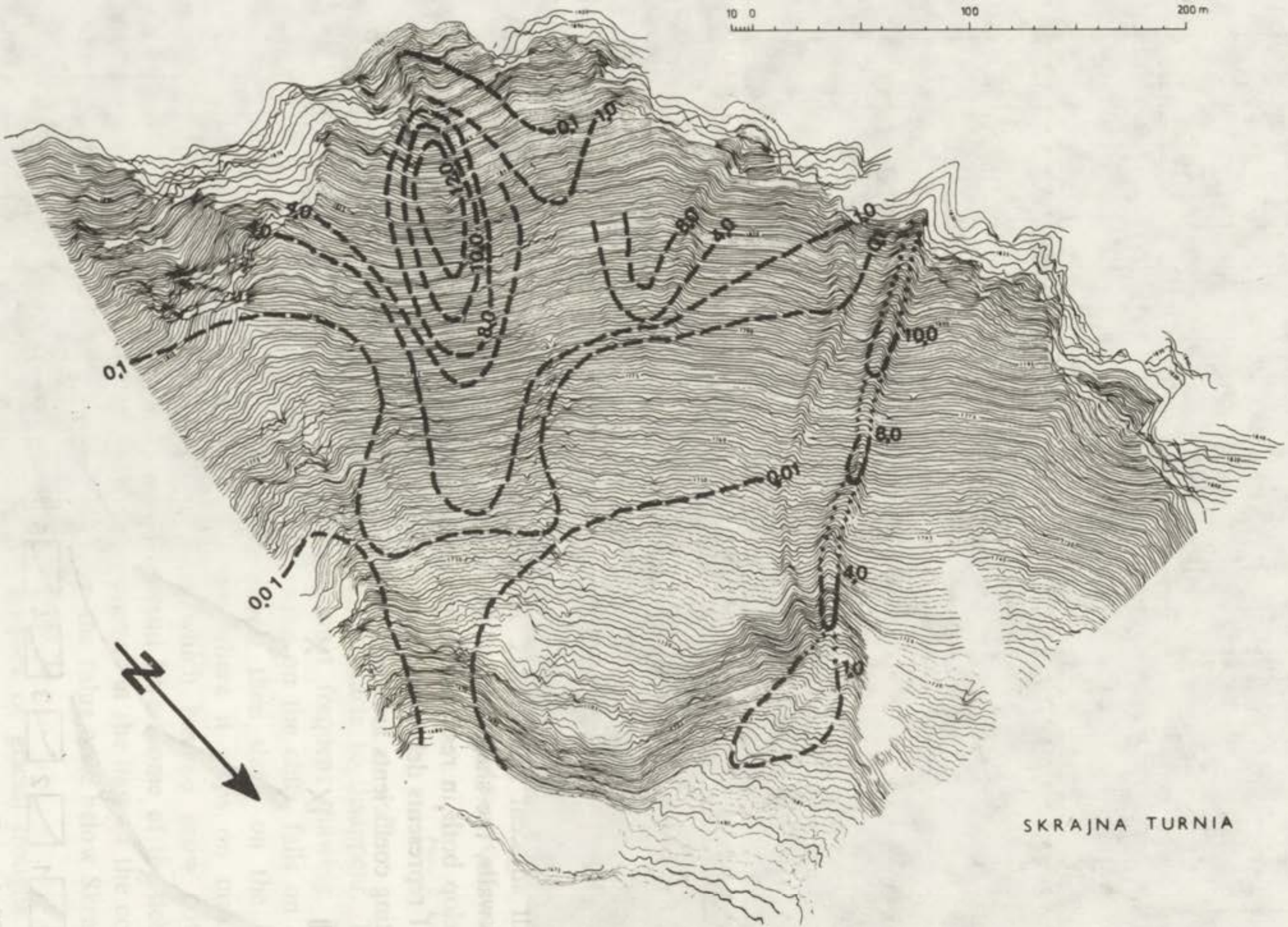

Fig. 30. Pattern of debris accretion (in mm *per* 5 years) on the Skrajna Turnia experimental slope, High Tatra Mts, 1975–1980 (after Kotarba, Kłapa and Rączkowska 1983)

in the Tatra, especially on the S- and W-facing slopes (*eg* slopes of Granaty and Buczynowe Turnie, Photo 2).

Type VI. Long (500–600 m) rockwalls and rocky slopes deeply channelled by chutes corresponding to the resistant zones (gradients of the order of 60°) and located in two climatic zones are subject to a complex pattern of development. Rockfalls and rock slides form section 0–1 within the debris slope, while debris flows and slush avalanches create section 1–2.

Photo 3. West-facing rocky slope of Żółta Turnia Mt near Czarny Staw Gąsienicowy lake, High Tatra Mts (*cf* Fig. 34)

Slopes affected by debris flows and dirty avalanches produce talus cone convering lateral moraine ridge

Photo M. Kot

Debris flows are produced by the selective transport of weathered material of various size across the slope. Therefore, no sorting of material is observed ($r < 0.20$) and the surface of the debris slope is characterized by a complex of secondary forms such as: rockfall debris streams, rock slide tongues, gullies, furrows, and levees. Examples of such slopes can be observed in the glacial cirque of the Czarny Staw Gąsienicowy lake especially under Granaty and in the vicinity of Czarny Staw pod Rysami lake (Photo 3).

Type VII. The N- and E-facing rockwalls located higher than 2000 m in the uppermost part of the Tatra are subject to a seminival climate. Steep perennial snow patches (up to 30°) occur directly at the base of the high cliffs and are subject to impact by weathered products supplied to them. Debris sliding on the snow surface accumulates at the front of the patch to form protalus rampart. This material is poorly sorted ($r < 0.20$). The area of the largest firn fields reaches 5 ha (Lukniš 1973). Protalus ramparts are also formed in basins associated with the highest lakes which are shaded from direct insolation (Photo 4).

Type VIII. Within the rocky slopes at heights up to 200–300 m numerous small rockwalls occur. At their base snow patches frequently last until June or even July. Debris accumulates across the whole length of the debris slope as a result of falling and sliding mainly in the concave

Photo 4. North-facing cliffs and debris slopes near the Morskie Oko lake, High Tatra Mts
Rockfall talus slopes affected by debris flows and avalanches action are related to the allochthonous alpine cliffs dissected by chute systems (cf Fig. 38)

section 1 – 2. The largest singular boulders and blocks gather at the base of the slope in section 2 – 3. Gravitational processes tend to even out the concave profile of the slope, however, their overall impact is relatively small. The index H_s/H_c is in the range 0.5 – 1.0, and frequently close to 1.0.

Type IX. Rocky slopes comprise degraded and inactive, 20 – 30 m high, slope units resulting from glacial oversteepening within cirques and within valleys. The debris slopes are stabilized and inactive and are supported by lateral moraines adjacent to the slope surface. Only individual fragments falling with a low frequency accumulate across the whole length of the debris slope and especially in the lower part. No accumulation can be detected. Index $H_s/H_c = 0.5 – 1.0$.

Type X. Rocky slopes within glacial cirques with heights up to 100 – 150 m are strongly dissected and selectively degradated at a low rate. Debris slopes up to c. 100 m high are stabilized. Inactive protalus ramparts (Fig. 22) occur at their base. In depressions between slope and rampart snow patches are present until June which slows down vegetation colonization. Single boulders slide from the cliffs and accumulate as mounds on the snow patches. Nevertheless, the infilling of the depressions proceeds at a very slow rate.

Type XI. These rocky slopes are dissected by chutes with the height of the slopes most frequently being 150 – 200 m but also below 100 m. The upper part of the talus cone is inactive and dissected by debris flow gullies. The lower part is formed from debris flow tongues (Fig. 22). The slope formation develops once every 3 – 5 years as a result of debris flow processes. The lower section of the slope becomes progressively longer and buries the glacial relief on the cirque floor. The concavity of the slope increases. The index H_s/H_c is within the range 0.5 – 1.0 and frequently has a value of 1.0.

Type XII. Rocky slopes up to 200 m high are formed by material of varying resistance and are dissected by chutes. They develop both by rockfall and debris flow processes. The lower part of the cone aggrades progressively towards the glacial cirque floor. The upper and middle parts are formed by debris falling and sliding from the cliffs. As a result. the shape of the debris slope does not change. The index H_s/H_c is close to 1.0 or within the interval 0.5 – 1.0.

The results from the above analysis demonstrate that the development of debris slopes always occurs at the expense of the rockwall and the rocky slope. This is due to: a) dissection of rockwalls by the relatively rapid deepening and widening of chutes, b) slow retreat of the steep rock surfaces, the rate depending on the relative resistance of the rock and other factors (*ie* aspect). In the advanced phase of development part of the cliff becomes more gentle because of the removal of the inter-chute ridges. The rock surface reaches a slope up to 30° and is veneered with a thin layer of loose debris. A mature slope – called a Richter denudation slope – is created this way.

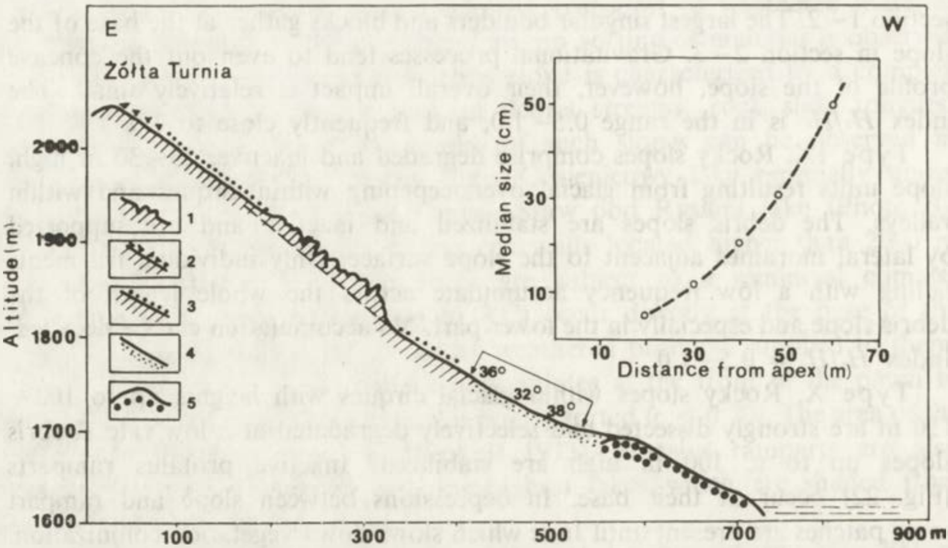

Fig. 31. Profile through the rocky slope near Czarny Staw Gąsienicowy lake (five-unit slope sequence). Top right – the relation between particle size and distance from apex on gravity sorted debris slopes

1 – rocky slope or rockwall. 2 – block slope. 3 – Richter denudation slope. 4 – debris slope. 5 – lateral moraine rampart

According to M. Klimaszewski (1967a, 1978) the whole slope from the ridge summit down to the valley floor consists of:

(i) a rockwall with a gradient greater than 45° dissected by chutes with gradients c. 30°,

(ii) a Richter denudation slope which conveys material supplied for debris transport from the rocky chutes in section (i),

(iii) rockfall talus, alluvial talus and avalanche talus which evolve in response to debris supplied from sections (i) and (ii). The model of 3-sectional development of the rockwall presented above has to be further developed for areas which were very strongly reworked by Pleistocene glaciers.

If the glaciers did not occupy the whole of the valley and near crest fragments of more gentle periglacial slopes projected above the glacier surface (Photo 5), the slope profile is composed of a larger number of units. The following sequence of units can be distinguished:

(i) block slope – commencing at a more gentle ridge crest covered with the products of block disintegration,

(ii) debris mantled slope – a strongly degraded cliff unit mantled with a weathering cover of coarse debris,

(iii) rockwall – formed as a result of erosional action by Pleistocene glaciers, later dissected by chutes and filled with debris,

(iv) Richter denudation slope (lower debris mantled slope) with single tors projecting above a thin debris cover,

Photo 5. Main ridge crest of the Western Tatra Mts with an elevation of the Starorobociański Mt (2176 m)
Debris-mantled slopes dominate the crest zone of the alpine geoecological belt (*cf* Fig. 22)

(v) debris slope developed as a series of interlocking rockfall, alluvial and avalanche taluses frequently superimposed on glacial drift deposits (Fig. 32). The present-day development of 3-unit und 5-unit slopes occurs differently. According to M. Klimaszewski (1967a, 1978), the 3-unit slope develops as a result of retreat of rockwalls and destruction of the inter-chute ridges in the lower part, and a gradual shifting of units upslope so that the middle part of the slope increases at the expense of the upper unit. The inter-chute ridges are subject to erosion leading to the development of tors which initially project above the weathering mantle and are then destroyed to mature smooth slope.

The 5-unit slope develops as a result of retreat of rockwalls. This occurs both from the cliff base in a similar manner as described above, but also with simultaneous lowering from the top downslope. The more gentle relief fragments of mountain crests marking the preglacial relief have been subject to a long period of periglacial and cryonival morphogenesis. Physical weathering in the form of frost shattering of bedrock has caused the development of rocky forms and debris covers. Loose, coarse material is still

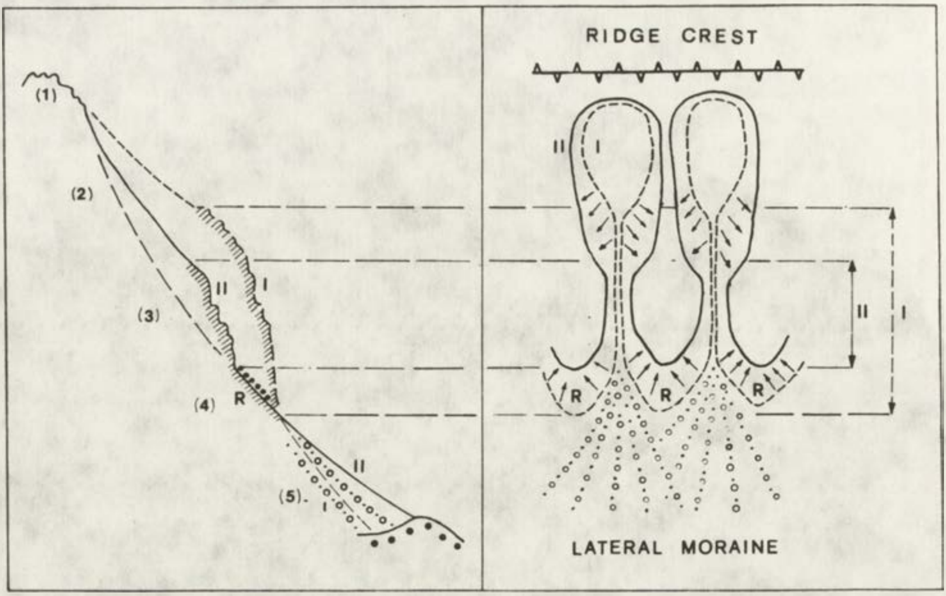

Fig. 32. A model of five-unit slope evolution in a deglaciated valley

R – Richter slope unit covered with a thin veneer of talus

subject to disintegration in the numerous nivation hollows at these heights. It is also subject to avalanche transport in winter and in summer is transported by debris flows to the chutes composing units (iii) and (iv). The released debris is erosive and undercuts the tors, inter-chute ridges and any raised rock surfaces which occur on the slope. This makes the slopes more gentle and widens the nivation hollows. As the gradient of the rocky slopes decreases, the debris is spilled into chutes channelling unit (iii) (Fig. 32). The total effect of both types of development is that the cliffs become more mature

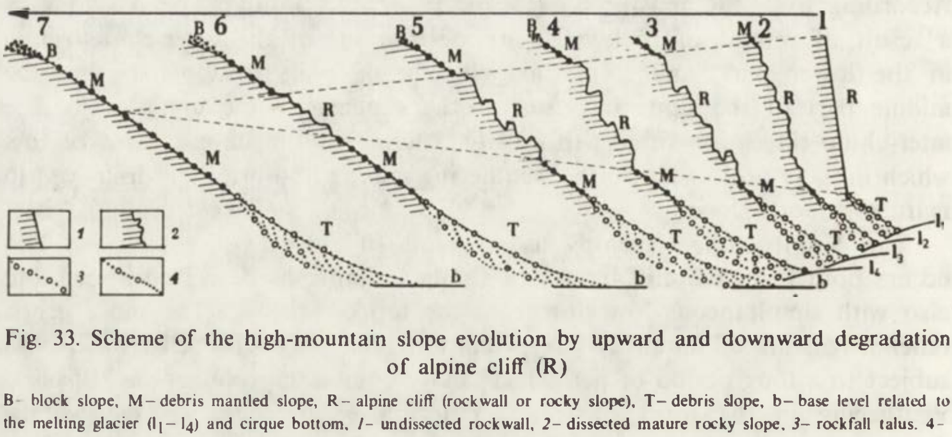

Fig. 33. Scheme of the high-mountain slope evolution by upward and downward degradation of alpine cliff (R)

B – block slope, M – debris mantled slope, R – alpine cliff (rockwall or rocky slope), T – debris slope, b – base level related to the melting glacier ($l_1 - l_4$) and cirque bottom, *1* – undissected rockwall, *2* – dissected mature rocky slope, *3* – rockfall talus, 4 – alluvial talus

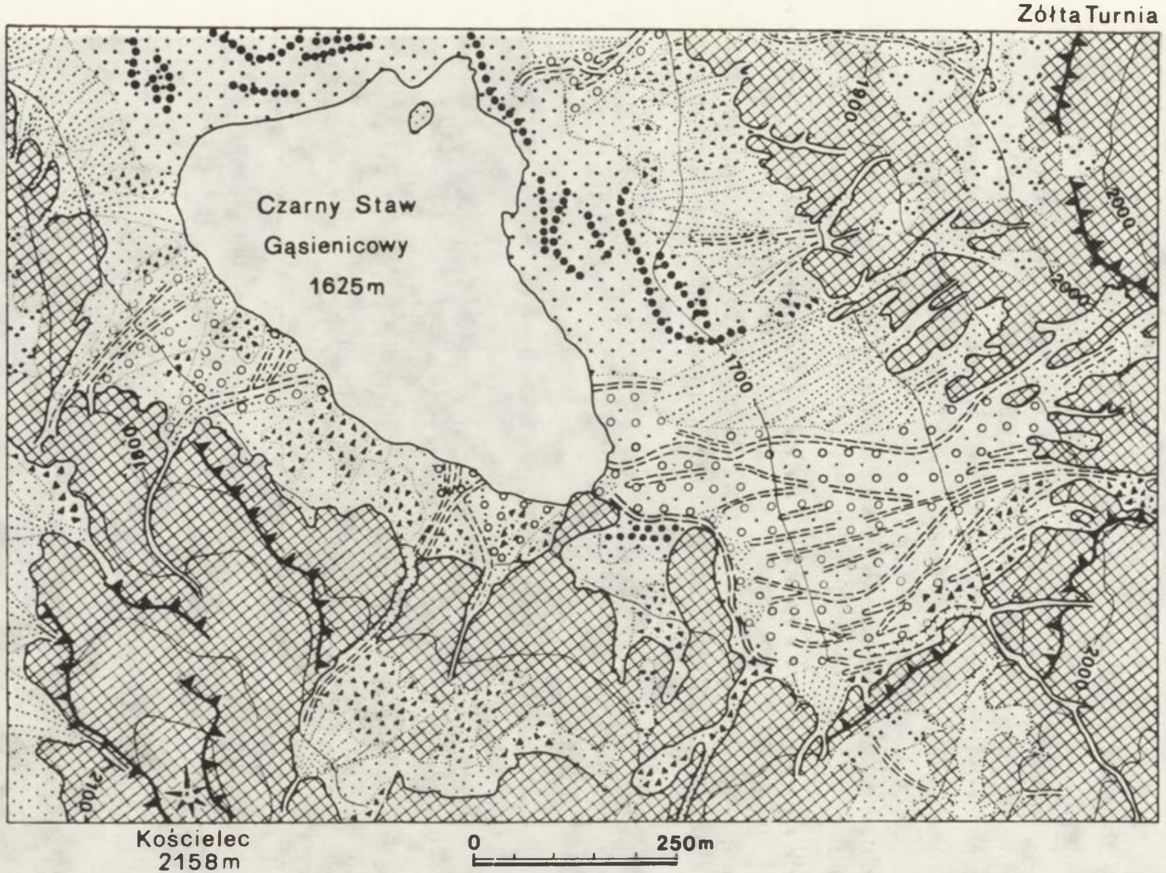

Fig. 34. Geomorphological map of the glacial corrie (Czarny Staw Gąsienicowy lake) in the High Tatra Mts

Note different stages of alpine slope evolution from steep rockwalls on the left side (NE aspect), through rocky slopes and block slopes to most advanced debris slopes channelled by debris flow gullies, levees, and tongues (for explanation see Fig. 23) both in the upslope and downslope directions. Examples of such 5-unit slopes are numerous in all the valleys and have been formed on granite, metamorphic and calcareous rocks. The evolutional sequence of these slopes is presented in Figure 33.

Specific process– response systems frequently occur in close proximity within or at the base of the same complex of rockwalls and differ only with respect to the stage of development during the postglacial period. Careful observations in all climatic belts have shown that the alpine cliffs in the first phase of deglaciation have developed as a result of gravitational processes and have formed extensive rockfall debris taluses (Fig. 34). Their advanced development and the increased supply of weathered material and slope fragmentation can be observed particularly in areas built of less resistant rocks (especially in the zones of cleavage and dislocation).

Alluviation within the glacial drift occurred mainly in the lower parts of the mountain terrain. Because of the absence of vegetation during the Late Glacial and in the early Holocene, extensive alluvial cones were formed under periglacial and perinival conditions with significant contributions from debris flows. Numerous inactive debris flow cones are now to be seen in the forest belt and in the subalpine belt. Alluviation within cirques became important during

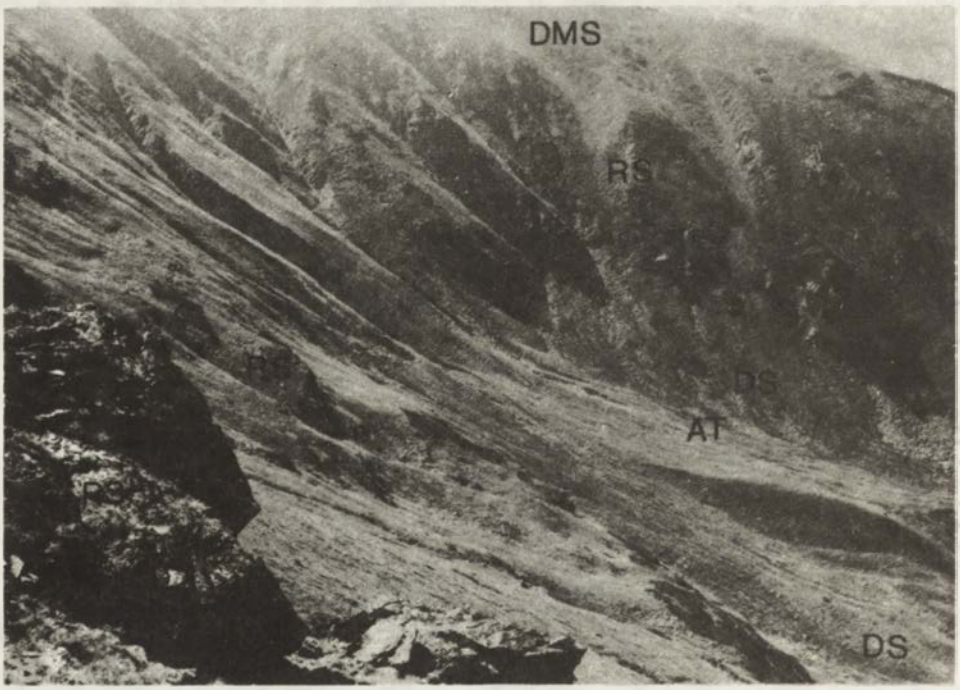

Photo 6. North- and west-facing slopes of Dolinka glacial cirque, Western Tatra Mts
Slopes have reached a more mature stage of development. Small rockwalls and tors form a mosaic pattern of the strongly dissected rocky slopes (RS). Debris-mantled slopes (DMS) occur in the summit parts of ridges. Debris slopes (DS) usually are stabilized by vegetation and changed at their base by debris flow processes (AT)

deglaciation of the higher mountain terrain and during the period when the young cliffs were being dissected. Alluvial talus slopes were superimposed on the rockfall talus slopes changing the profiles of the debris slope. In this manner the Late Glacial phase of rockfall talus slope development was replaced by an alluvial phase. The youngest phase of alluviation can be 'related to a climatic change resulting in a higher influence of convective summer rains since around 1600 AD' (Kotarba and Strömquist 1984). However, intensified alluviation of the last 200 years especially in the Western Tatra is mainly related to an increase in human activity. Therefore, alluvial talus slopes are the principal slope forms across the whole Tatra and in some cirques they produce extensive surface with a dense pattern of debris flow features (Figs. 22, 23, 34, Photo 6). The largest alluvial talus in the Slovak part of the Tatra Mts is 100 m high and the largest rockfall talus is 430 m high (Lukniš 1973). If the rockfall talus slopes are characterized by good sorting of material, the alluvial talus slopes are poorly sorted and the correlation coefficients determining the relationship between the size of material and the distance from the apex are not statistically significant (Fig. 35).

Only the high glacial cirques in the High Tatra, the sites most recently occupied by glaciers, have retained almost primary forms, *ie* vertical cliffs and poorly developed basal debris slopes which at present are developing in response to gravity, falling, rolling, and sliding of scree particles (type I and II in Fig. 26). Such simple rockfall talus slopes are fairly rare in the Tatra. Dissection of rockwalls resulted in the formation of deep chutes and gorges. The less resistant zones were areas of accelerated weathering related to the presence of an active permafrost layer. Chutes, areas of meltwater and rain water concentration, became major routes for debris transport from the rockwalls to their bases.

The complex of processes shaping debris slopes have changed along with shifts in the climatic and vegetation belts during the postglacial period. During the Late Glacial, climatic conditions favoured the supply of the large

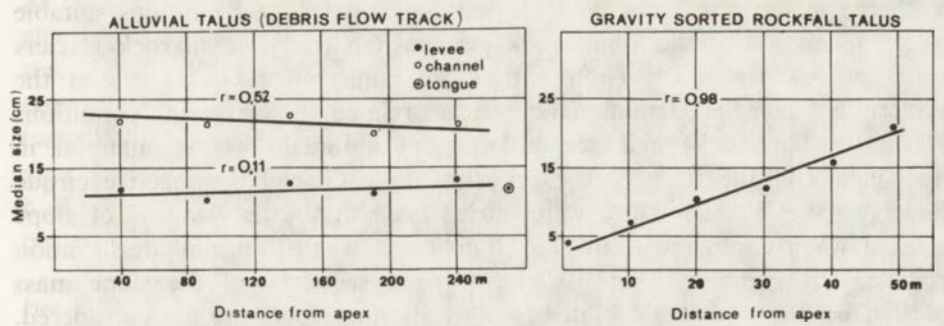

Fig. 35. The relations between block size and distance from apex on an alluvial talus (left side) and on active gravity sorted rockfall talus (right side) (after Kotarba and Strömquist 1984)

N-facing slope close to the Morskie Oko lake and E-facing slope close to the Czarny Staw pod Rysami lake

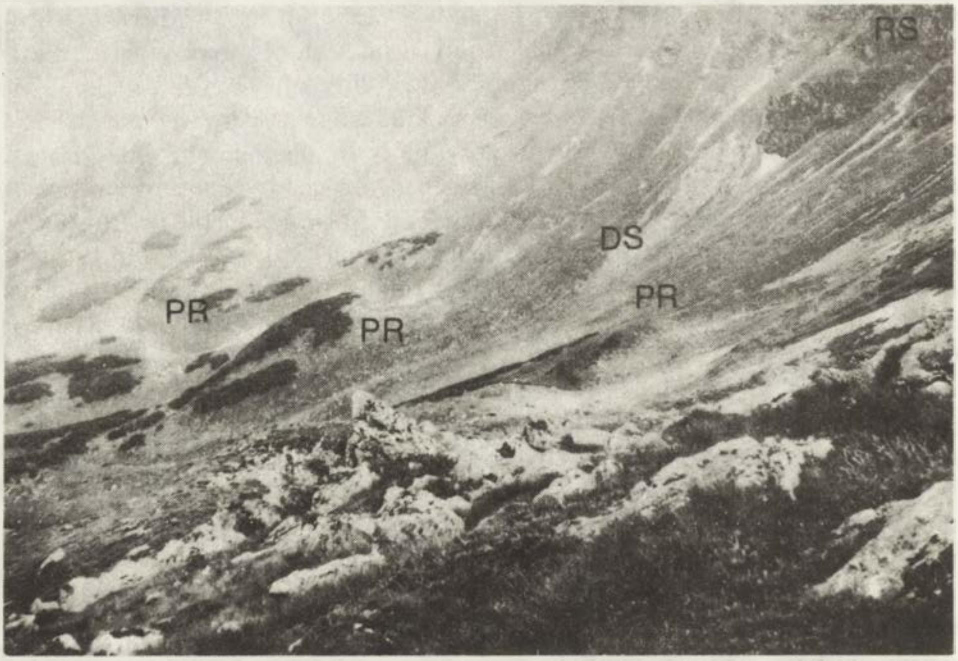

Photo 7. North-facing rocky (RS) and debris (DS) slopes of the Błyszcz Mt (2158 m), Western Tatra Mts

At the base of debris slopes fossil protalus ramparts (PR) occur (*cf* Fig. 22)

amounts of debris and snow from slopes and the formation of rock glaciers in the cryonival belt. As a result of climatic warming, the latter have become transformed into fossil forms but retaining their earlier shapes (Photo 7 and 8). Postglacial gravitational and nival processes and particularly alluviation have partially covered these forms with debris. Figures 36 and 37 show a full sequence of slope mass transfer features in the Pańszczyca valley, within the High Tatra. Only in zones with a significant supply of weathered material (*eg* from Wierch pod Fajki) were the conditions suitable for the formation of the lobate rock glaciers. Generally, fossil rock glaciers occur sporadically in the High Tatra. The numerous rock glaciers in the Western Tatra in the same climatic belt are related to weathering conditions in less resistant rocks and accumulation of abundant coarse material in small glacial cirques (Fig. 22). During the Late Glacial the inactive cirque glaciers in the Western Tatra were buried beneath thicker deposits of slope material when compared with the High Tatra where 'normal' deglaciation prevailed. In Figures 37 and 38 showing the sequence of the slope mass transfer features, the forms related to alluviation of slopes are not considered. The complex of alluvial forms can occur across all slope units. Debris flows formed tracks within rockfall/rockslide slopes, and deposited the transported material onto glacial drift deposits within recessional moraines. Troughs

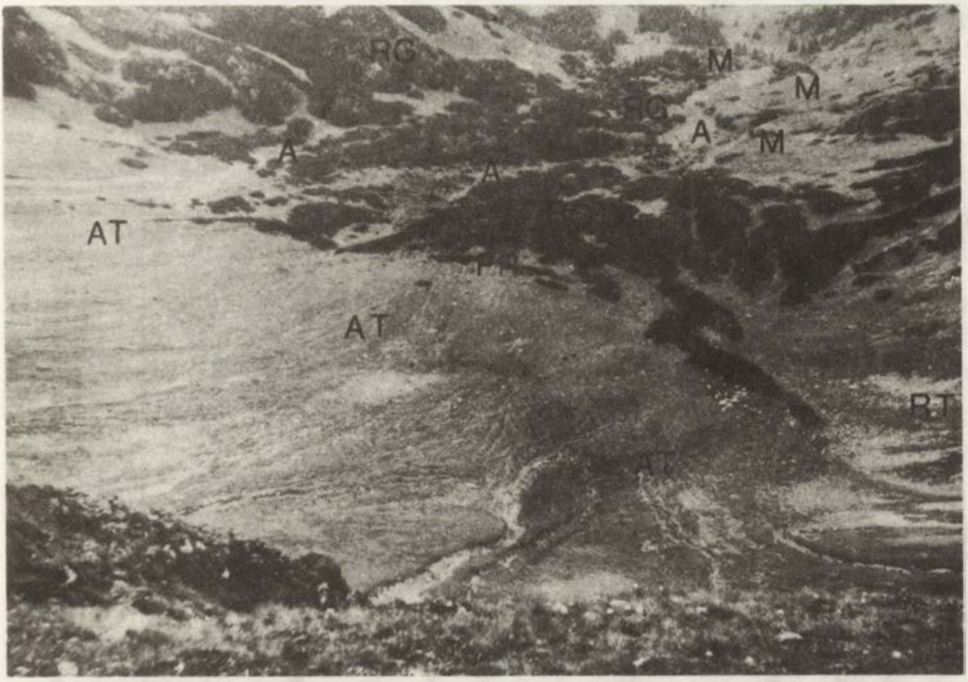

Photo 8. A bottom of the Starorobociański cirque, Western Tatra (view from the top downslope)

AT (alluvial talus) – lower part of the debris slopes transformed by debris flow activity, PR – fossil protalus rampart, RG – fossil rock glaciers, M – frontal and lateral moraine ridges related to the last recessional phase of glacier stagnation, A – marginal troughs filled with alluvial deposits (cf. Fig. 22)

and hollows between the moraine ridges and related to rock glaciers are infilled with slope deposits. Debris flow form generations of cones and tongues descending from the debris slopes and engulfing the underlying form of these depressions (Fig. 37). Thus, they mask the relief associated with the deglaciation of the cirques (Photo 8). Palynological analysis and ^{14}C dating of these sediments in the Starorobociański cirque demonstrate that the rock glaciers were made into fossil forms by slope deposition. This has occurred since the end of the Boreal period as deposits of early Atlantic age lie beneath the latter sediments (Kaszowski, Krzemień, Libelt 1987). At least two-thirds of the deposits infilling the troughs are of Subboreal and younger ages (^{14}C data: 3090 ± 60 BP and 3590 ± 80 BP). In some troughs at least half of sediment accumulation is associated with the period of human impact on the natural environment of the region (^{14}C data: 360 ± 50 BP).

The rockwall – debris slope system is subject to evolution as the relief becomes more mature (Fig. 33). Its development consits of rockwall dissection and then dissection of the rocky slopes; these processes balance each other and sustain a simultaneous decrease of gradient to c. 30°. This development mainly results in a widening of the chutes (Klimaszewski 1971). However,

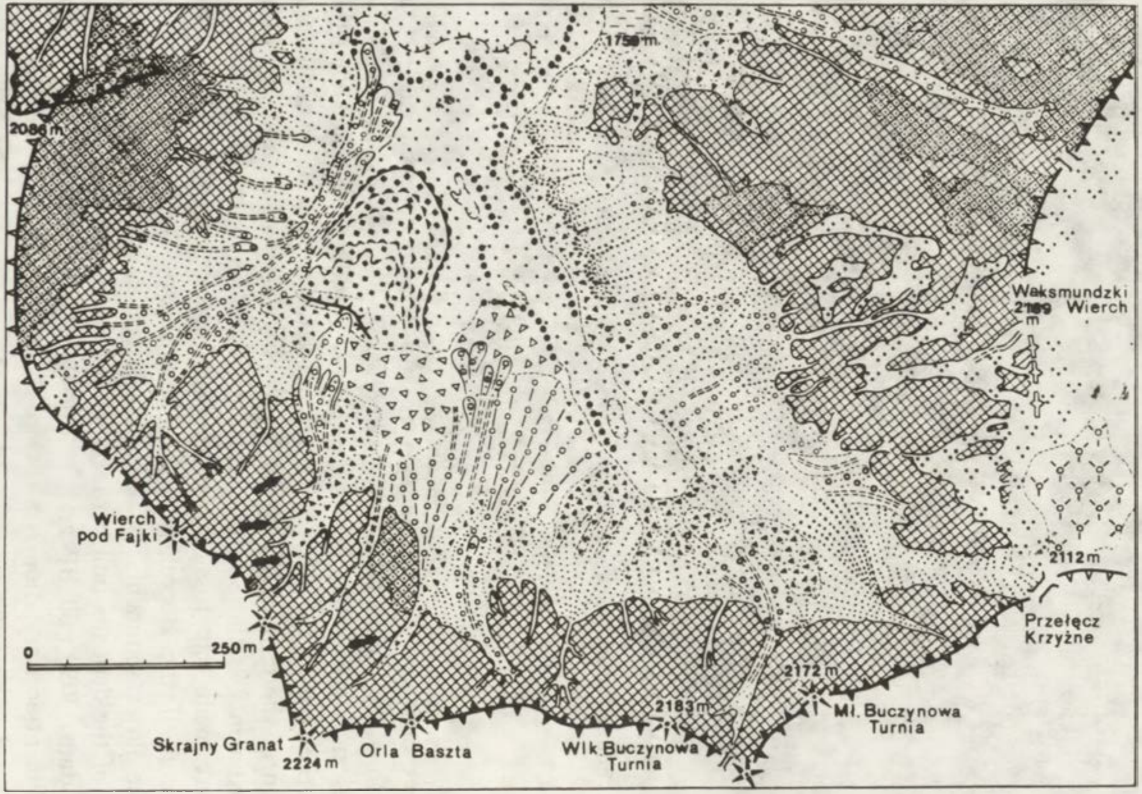

Fig. 36. Geomorphological map of glacial cirque of Pańszczyca valley in the High Tatra Mts

Asymmetry of west- and east-facing rocky slopes is well expressed in morphology. At the base of NE-facing rockwall full sequence of slope mass transfer features is to be seen (for explanations see Fig. 23)

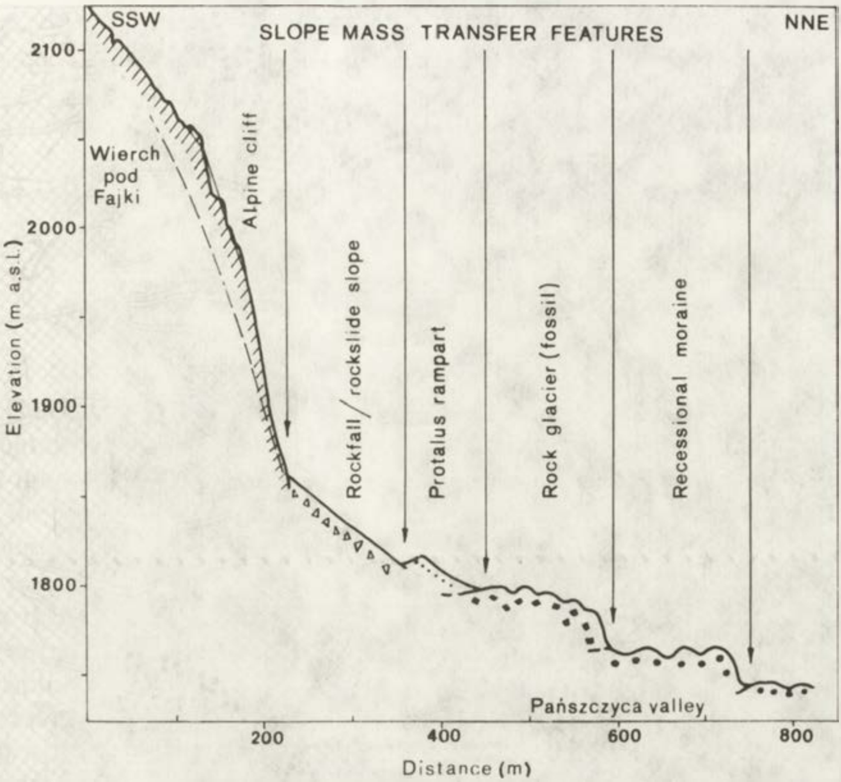

Fig. 37. Full sequence of slope mass transfer features in the High Tatra Mts (for comparison see Fig. 36)

degradation of rockwalls and rocky slopes progresses in both the upslope and downslope directions. Such development means that a more or less graded debris mantled slope is formed. Simultaneously, a permanent transformation of debris slopes occurs. This consists of an increase in alluviation and a progressive elongation of the concave and more gentle base of the debris slope. In this manner debris slopes develop at the expense of the cirque floors which become fossilized beneath slope deposits.

Fig. 38. Geomorphological map of glacial cirque Czarny Staw lake and trough (Morskie Oko lake) in the High Tatra Mts

Very frequent debris flow features have developed at the bases of allochthonous alpine cliffs (for explanations see Fig. 23)

FINAL REMARKS

The development of geoecological belts is a specific feature of high mountains. Particular relationship exists between geomorphological processes and the climatic, geological, and vegetation conditions which are responsible for the mode and rate of the mountain relief transformation. In the high--mountain area of the Tatra there is an altitudinal differentiation of morphogenetic processes which result from these environmental factors. The area most strongly transformed by present-day processes is located between the upper timberline at a height of 1500– 1950 m. *ie* within the alpine and subalpine geoecological belts. Within these belts the most rapid. catastrophic processes which occur are related to extreme hydrometeorological phenomena (debris flows, rockfalls, snow and slush avalanching and shallow landsliding) and to slow mass movements (soil creep, true talus creep). Furthermore, this is the area with the largest accumulations of loose, mobile glacial, glaciofluvial, and gravitational deposits. These deposits are subject to transportation by mass movemens caused by extreme hydrometeorological events.

Below and above this belt, the rate of geomorphic activity decreases and in some cases entirely ceases. The existence of only one altitudinal belt is a specific feature of non-glaciated mountains since in glaciated areas (the Alps, Caucasus, Tian-Shan, and others) two altitudinal belts exist. The upper belt is actively shaped by glacial and periglacial processes while the lower one, corresponding to the belt in the Tatra, is located in the zone of the upper timberline. Above this zone intensive and efficient spring and summer rainfalls operate as the main factors triggering the removal of weathered material.

The high, non-glaciated mountains in temperate latitudes are characterized by two independent morphodynamic subsystems. The slope subsystem is mainly modelled by a complex of gravitational, pluviogravitational, nival, and cryogravitational processes which occur in the seminival, alpine and subalpine belts. The channel subsystem, modelled mainly by fluvial processes, includes the forest belt and partially penetrates the subalpine belt. The two subsystems exist independently of each other because of the lack of a direct linkage between them during average hydrometeorological conditions. However, there are internal processes within these subsystems and the products of weathering and denudation are not transferred from one to the other subsystem. These products remain within each subsystem and are not subject to transfer throughout the geoecological belts. The overdeepened valley floors with lakes located at various heights are natural sediment sinks. The weathered materials transferred from slopes accumulate in the lake basins. Transport of material along the valley floors does not occur since rock steps separating the overhanging valley sections provide ntural barriers to continuous sediment transfer. Only during catastrophic hydrometeorological events, occurring once *per* 100 years are both subsystems linked in some parts of the mountains and then only for a short period. Such conditions occur in the Tatra only under rainfalls of the order of 200–300 mm *per* day. However, in the present-day glaciated high mountains there is a morphodynamic glacial or denudation-glacial system which is linked with the fluvial one. The highly active internal processes of these systems and their vertical connection throughout the geoecological belts result in large amount of weathered material being removed beyond the mountain zone (Kaszowski 1985).

During last few centuries, *ie* since 1600 AD, there has been a tendency for alluviation of the slope subsystem in the Tatra as in the Alps. Alluviation is associated with a change of climatic regime observed in both mountain areas on the basis of dendroclimatic studies and studies of human impact in the mountains. A prospensity towards a change in debris slope profiles and more selective shaping of rockwalls and rocky slopes has been observed. The numerous generations of mesoforms associated with debris flow activity are characteristic features of these slopes subject to alluviation. This trend within the high-mountain denudational system of the Tatra serves to produce links between the slope and channel subsystems. This in turn favours the development of a longitudinal profile providing optimum conditions for the transport of sediment across geoecological belts. In this manner the adaptation of the high mountain system steadily occurs. This system which was formed in the Pleistocene and especially in the Late Glacial is now adjusting to the climatic conditions of the temperate zone.

Although the upper timberline is very important in term of its morphodynamic. it does not produce an abrupt discontinuity. The altitudinal changes of its position during the Holocene created a transitional belt within which a gradual change in the intensity of geomorphic processes can be observed. At the upper timberline there is a distinct change of the rate of chemical denudation, especially on the calcareous bedrock.

Absolute measurements of the rate of operation of morphogenetic processes (especially rockfall and slow mass movements) reach similar values in the Tatra Mts as in other high mountains in the temperate zone (Caine 1974, 1978) while the efficiency of the rapidly operating processes is different in particular mountain massifs despite being caused by similar hydrometeorological events. Geomorphological results of rapid mass movement depend on the relief stability and its morphometric parameters. The forms produced by rapid mass movement in the Tatra Mts are much smaller in size than those in the high massifs of the Alpine arc but larger in size than those in the mountains of Scandinavia. The relatively early termination of glacial activity in the Tatra Mts has meant that the slopes and valley floors became very stable, and independent morphodynamic systems have been formed within them.

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REFERENCES

Anselmo V., 1979, *Il nubifragio del 7 Agosto 1978 nel bacino del Toce*, Boll. Ass. Min. Subalp., 16, 2.

Aulitzky H., 1970, *Der Enterbach (Inzing in Tirol) am 26 Juli 1969: Versuch der Analyse eines Murganges als Grundlage für die Neuerstellung einer zerstörten Wildbachverbauung*, Wildbach- und Lawinenverbau, 34, 1.

Baumgart-Kotarba M., Kotarba A., 1979, *Wpływ rzeźby dna doliny i litologii utworów czwartorzędowych na wykształcenie koryta Białej Wody w Tatrach*, Folia Geogr., Ser. Geogr. Phys., 12.

Bednarz Z., 1975, *Geographical range of similarities of annual growth curves of stone pine (Pinus cembra) in Europe*, Symposium Materials of XII Int. Botanical Cong., 'Bioecological Fundamentals of Dendrochronology', Leningrad, July 1975.

– 1981, *Relationship of tree-ring widths in the Tatra Mountains to variations in monthly temperature and precipitation*, Mitt. Forstl. BundVersAnst., Wien, 142.

– 1984, *The comparison of dendroclimatological reconstructions of summer temperatures from the Alps and Tatra Mountains from 1741–1965*, Dendrochronologia, 2.

Bowman D., 1977, *Stepped-bed morphology in arid gravelly channels*, Bull. Geol. Soc. Am., 88, 2.

Caine N., 1974, *The geomorphic processes of the alpine environment*, [in:] *Arctic and alpine environments*, J. D. Ives, R. G. Barry (eds.), Methuen, London.

– 1978, *Climatic geomorphology in mid-latitude mountains*, [in:] *Landform evolution in Australasia*, J. L. Davies and M. A. J. Williams (eds.), Canberra.

– 1980, *The rainfall intensity-duration control of shallow landslides and debris flows*, Geogr. Ann., 62 A, 1–2.

– 1983, *The mountains of Northeastern Tasmania. A study of alpine geomorphology*, A. A. Balkema, Rotterdam.

Cebulak E., 1983, *Maximum daily rainfalls in the Tatra Mountains and Podhale Basin*, Zesz. Nauk. UJ, Prace Geogr., 57.

Chardon M., 1984, *Montagne et haute montagne alpine, criteres et limites morphologiques remarquables en haute montagne*, Rev. Géogr. Alp., 72, 2–3.

Dyakowska J., 1932, *Analiza pyłkowa kilku torfowisk tatrzańskich*, Acta Soc. Bot. Pol., 9, 3–4.

Fahnestock R. K., 1963, *Morphology and hydrology of a glacial stream – White River, Mount Rainier, Washington*, U.S. Geol. Survey Prof. Paper, 422–A.

Figuła K., 1966, *Badania transportu rumowiska w ciekach górskich i podgórskich o różnej budowie geologicznej i użytkowaniu*, Wiad. IMUZ, 6.

Fisher O., 1866, *On the disintegration of a chalk cliff*, Geol. Mag., 3.

Flohn H., 1974, *Contribution to a comparative meteorology of mountain areas*, [in:] *Arctic and alpine environments*, J. D. Ives and R. G. Barry (eds.), Methuen, London.

Froehlich W., 1975, *Dynamika transportu fluwialnego Kamienicy Nawojowskiej*, Prace Geogr., IGiPZ PAN, 114.

Froehlich W., Kaszowski L., Starkel L., 1977, *Studies of present-day and past river* *activity in the Polish Carpathians*, [in:] *River channel changes*, K. J. Gregory (ed.), John Willey and Sons, Chichester– Brisbane– Toronto.

Furrer G., 1965, *Die Subnivale Höhenstufe und ihre Untergrenze in den Bünder und Walliser Alpen*, Geogr. Helv., 20, 4.

Galibert G., 1960, *L'evolution actuelle des "faces nord" de la haute montagne alpine dans le massif de Zermatt*, Rev. Géogr. Pyrénées S.-Ouest, 31.

Gardner J. S., Smith D. J., Desloges J. R., 1983, *The dynamic geomorphology of the Mt. Rae area: A high mountain region in Southwestern Alberta*, Dept. of Geography Publication Ser. No 19, Univ. of Waterloo.

Gerlach T., 1959, *Lód włóknisty i jego rola w przemieszczaniu pokrywy zwietrzelinowej w Tatrach*, Przegl. Geogr., 31, 3–4.

Gieysztor I., 1962, *Uwagi o opadach w Tatrach Polskich*, Przegl. Geogr., 34, 3.

Głazek J., Wójcik Z., 1963, *Karst phenomena in the eastern part of the Polish Tatra Mts.*, Acta Geol. Pol., 13, 2.

Govi M., 1984a, *Les phenomenes catastrophiques d'origine exogene*, [w:] *Geomorphologie et risques naturels dans les Alpes*, 25^e^ Congrès International de Géographie, Paris–Alpes 1984.

– 1984b, *Le risque lié aux phénomènes exogènes*, [in:] *Geomorphologie et risques naturels dans les Alpes*, 25^e^ Congrès International de Géographie, Paris– Alpes 1984.

Govi M., Sorzana P. F., 1980, *Landslide susceptibility as a function of critical rainfall amount in Piedmont basin (North-Western Italy)*. Studia Geomorph. Carpatho-Balcan., 14.

Grochocka-Piotrowska K., 1970, *Fotointerpretacja i geneza struktur nieciągłych w masywie granitowym polskiej części Tatr Wysokich*. Acta Geol. Pol., 20(2).

Halicki B., 1930, *Dyluwialne zlodowacenie północnych stoków Tatr*, Spraw. Państw. Inst. Geol., 5.

– 1933, *Parę uwag o rozwoju dolin tatrzańskich*, Spraw. Państw. Inst. Geol., 7(2).

Heede B. H., 1981, *Dynamics of selected mountain streams in the western United States of America*, Z. Geomorph., N.F., 25(1).

Hess M., 1965, *Piętra klimatyczne w Polskich Karpatach Zachodnich*, Zesz. Nauk. UJ, Prace Geogr., 11.

– 1968, *Próba rekonstrukcji klimatu w holocenie na terenie Polski Południowej*, Folia Quatern., 29.

Höllermann P., 1973, *Some reflection on the nature of high mountains with special reference to the Western United States*, Arc. Alp. Res., 5, 3.

Izmaiłow B., 1984a, *Eolian processes in Alpine belt of the High Tatra Mountains, Poland*, Earth Surface Processes and Landforms, 9.

– 1984b, *Eolian deposition above the upper timberline in the Gąsienicowa Valley in the Tatra Mts*, Zesz. Nauk. UJ, Prace Geogr., 61.

Jahn A., 1958, *Mikrorelief peryglacjalny Tatr i Babiej Góry*, Biul. Perygl., 8.

– 1970, *Zagadnienia strefy peryglacjalnej*, PWN, Warszawa.

Kalvoda J., 1974, *Geomorfologicky vyvoj hrebenove casti Vysokych Tater*, Rozpr. ČSAV, Řada MPV, 84, 6.

Karrasch H., 1977, *Die klimatischen und aklimatischen Varianzfaktoren der periglazialen Höhenstufe in den Gebirgen West- und Mitteleuropas*, Abh. Akad. Wiss. Göttingen, Math.-Phys. Kl. Dr. Folge, 31.

Kaszowski L., 1973, *Morphological activity of the mountain streams (with Biały Potok in the Tatra Mts as example)*, Zesz. Nauk. UJ, Prace Geogr., 31.

– 1985, *Rzeźba i modelowanie gór wysokich strefy suchej na przykładzie Hindukuszu Munjan*, Rozpr. habil., UJ, 94.

Kaszowski L., Krzemień K., 1979, *Channel subsystems in the Polish Tatra Mts*, Studia Geomorph. Carpatho-Balcan., 13.

Kaszowski L., Krzemień K., Libelt P., 1987, *Postglacjalne modelowanie cyrków lodowcowych w Tatrach Zachodnich*, Zesz. Nauk. UJ, Prace Geogr., (in print).

Kelletat D., 1969, *Verbreitung und Vergesellschaftung rezenter periglazialer Erscheinungen im Apenin*, Göttinger Geogr. Abh., 48.

K l i m a s z e w s k i M., 1960, *On the influence of the pre-glacial relief on the course and magnitude of the glacial erosion in the Tatra Mts.*, Geogr. Pol., 2.

– 1961, *Geomorfologia ogólna*, PWN, Warszawa.

– 1962, *Zarys rozwoju rzeźby Tatr Polskich*, [in:] *Tatrzański Park Narodowy*, Kraków.

– 1967a, *Development of granite slopes in the high mountain areas in the Late Glacial and Holocene times*, [in:] *Guide to excursion of the Symposium of the Commission on the Evolution of Slopes and of the Commission on Periglacial Geomorphology of the IGU, Poland September 1967.*

– 1967b, *Polskie Karpaty Zachodnie w okresie czwartorzędowym*, [in:] *Czwartorzęd Polski*, PWN, Warszawa.

– 1971, *A contribution to the theory of rock-face development*, Studia Geomorph. Carpatho-Balcan., 5.

– 1978, *Geomorfologia*, PWN, Warszawa.

K l a p a M., 1963, *Prace Stacji Badawczej Instytutu Geografii PAN na Hali Gąsienicowej w latach 1960 i 1961*, Przegl. Geogr., 35, 2.

– 1980, *Procesy morfogenetyczne oraz ich związek z sezonowymi zmianami pogody w otoczeniu Hali Gąsienicowej w Tatrach*, Dok. Geogr., 4.

K o n d r a t y e v J., 1977, *Radiation regime of inclined surface*, Technical Note No 152, WMO 467, Geneva.

K o m o r n i c k i T., S k i b a S., 1979, *Soils of the Polish Tatra Mts*, [in:] *Excursion Guide-Book, Field Meeting of the IGU Commission on Field Experiments in Geomorphology, Poland, September 17–25, 1979*, Wrocław Univ. Press.

K o n d r a c k i J., 1984, *Badania paleolimnologiczne w Tatrach w latach 1974–1978*, Prace i Studia Geogr., WGiSR UW, 5.

K o p e r o w a W., 1958, *Późny glacjał u północnych podnóży Tatr w świetle analizy pyłkowej*, Monogr. Bot., 7.

– 1962, *Późnoglacjalna i holoceńska historia roślinności Kotliny Nowotarskiej*, Acta Paleobot., 2–3.

K o s z y k H., 1977, *Zróżnicowanie procesów grawitacyjnych we współczesnym modelowaniu rzeźby krystalicznej części Tatr Zachodnich*, Arch. Inst. Geogr. UJ, Kraków.

K o t a r b a A., 1967, *Żłobki krasowe w Tatrach*, Zesz. Nauk. UJ, Prace Geogr., 16.

– 1970, *The morphogenetic role of foehn wind in the Tatra Mts.*, Studia Geomorph. Carpatho-Balcan., 4.

– 1972, *Comparison of physical weathering and chemical denudation in the Polish Tatra Mts*, [in:] *Processus périglaciaires études sur le terrain, Symposium International de Geomorphologie, Liège, Juillet 1971.*

– 1976, *Współczesne modelowanie węglanowych stoków wysokogórskich*, Prace Geogr. IGiPZ PAN, 120.

– 1981, *Present-day transformation of alpine granite slopes in the Polish Tatra Mountains*, Proc. of the Third Meeting of IGU Commission on Field Experiments in Geomorphology, Japanese Geomorphological Union, Kioto.

– 1983, *Processes on alpine cliffs in the Polish Tatra Mountains*, [in:] *Studies in Quaternary Geomorphology*, D. J. Briggs and R. S. Waters (eds.). Geobook.

– 1984, *Elevational differentiation of slope geomorphic processes in the Polish Tatra Mts*, Studia Geomorph. Carpatho-Balcan., 18.

K o t a r b a A., S t a r k e l L., 1972, *Holocene morphogenetic altitudinal zones in the Carpathians*, Studia Geomorph. Carpatho-Balcan., 6.

K o t a r b a A., K l a p a M., R ą c z k o w s k a Z., 1983, *Procesy morfogenetyczne kształtujące stoki Tatr Wysokich*, Dok. Geogr., 1.

K o t a r b a A., S t r ö m q u i s t L., 1984, *Transport, sorting and deposition processes of alpine debris slope deposits in the Polish Tatra Mountains*, Geogr. Ann., 66 A, 4.

K r u p i ń s k i K., 1983, *Evolution of Late Glacial and Holocene vegetation in the Polish Tatra Mts. based on pollen analysis of sediments of the Przedni Staw Lake*, Bull. Pol. Acad. Sci., Earth Sci., 31, 1–4.

Krzemień K., 1982, *Removal of dissolved material from the Starorobociański stream crystalline catchment basin (The Western Tatra Mts.)*, Studia Geomorph. Carpatho-Balcan., 15.
– 1984, *Fluvial transport balance in high-mountain crystalline catchment basin*, Zesz. Nauk. UJ, Prace Geogr., 61.
– 1985, *Present-day activity of the high-mountain stream in the Western Tatra Mts.*, Quaest. Geogr., Special Issue, 1.
Ksandr J., 1954, *Geomorfologicka studie dolin jizniho svahu Vysokych Tater*, Rozpr. ČSAV, Řada MPV, 64, 5.
Larsson S., 1982, *Geomorphological effects on the slopes of Longyear Valley, Spitsbergen, after a heavy rainstorm in July 1972*, Geogr. Ann. 64 A, 3–4.
Lehmann O., 1933, *Morphologische Theorie der Verwitterung von steinschlag Wänden*, Vjschr. Naturf. Ges. Zürich, 87.
Löve D., 1970, *Subarctic and subalpine. Where and what?* Arct. Alp. Res., 2, 1.
Lukniš M., 1968, *The intensity of lowering of High Tatra ridges since the retreat of glaciers*, Studia Geomorph. Carpatho-Balcan., 2.
– 1973, *Relief Vysokych Tatier a ich predpolia*, SAV, Bratislava.
Maizels J. K., 1978, *Debit des eaux de fonte, charges sedimentaires et taus d'erosion dans le massif du Mount Blanc*, Rev. Geogr. Alp., 66, 1.
Mamica W., 1984, *Współczesna dynamika rzeźby wysokogórskiej na przykładzie doliny Wyżniej Chochołowskiej w Tatrach Zachodnich*, Materiały archiwalne w Inst. Geogr. UJ.
Mapa geologiczna Tatr Polskich 1 : 30 000, 1979.
McArthur J. L., 1975, *Some observations on periglacial morphogenesis in the Southern Alps, New Zealand*, Geogr. Ann., 57 A, 3–4.
Midriak R., 1983, *Morfogeneza povrchu wysokych pohori*, Bratislava.
– 1984, *Debris flows and their occurrence in the Czechoslovak Carpathians*, Studia Geomorph. Carpatho-Balcan., 18.
Myczkowski S., 1962, *Wpływ lawin śnieżnych na lasy Tatrzańskiego Parku Narodowego w dolinach: Rybiego Potoku, Roztoki, Waksmundzkiej i Pańszczycy*, Ochr. Przyr., 28.
Myczkowski S. *et. al.*, 1974, *The map of forest and bushwood communities in the Tatra National Park*, Studia Ośr. Dokum. Fizjogr., 3, Kraków.
Myczkowski S., Lesiński J., 1974, *Rozsiedlenie rodzimych gatunków drzew tatrzańskich*, Studia Ośr. Dokum. Fizjogr., 3, Kraków.
Niedźwiedź T., 1981, *Sytuacje synoptyczne i ich wpływ na zróżnicowanie przestrzenne wybranych elementów klimatu w dorzeczu górnej Wisły*, Rozpr. habil., UJ, 58.
Obidowicz A., 1975, *Entstehung and Alter einiger Moore im nördlichen Teil der Hohen Tatra*, Fragm. Flor. Geobot., 21, 3.
Oleksynowa K., Komornicki T., 1965, *The chemical composition of water in the Tatra Mountains*, [in: *Tatra Mountains*, K. Starmach (ed.), PWN, Cracow.
Orlicz M., 1954, *O stosunkach anemometrycznych na szczytach tatrzańskich*, Wiad. Służby Hydrol.-Meteorol., 3, 4.
– 1962, *Klimat Tatr*, [in:] *Tatrzański Park Narodowy*, Kraków.
Otruba J., Wiszniewski W., 1974, *Veterne pomery*, [w:] *Klimat Tatr*, Veda, Bratislava.
Partsch J., 1923, *Die Hohe Tatra zur Eiszeit*, Leipzig.
Pawłowska S., 1962, *Świat roślinny Tatr*, [in:] *Tatrzański Park Narodowy*, Kraków.
Plesnik P., 1972, *A contribution to the question of the geographical character of European high mountains*, Geogr. Čas., 24, 2.
Poser H., 1954, *Die Periglazial-Erscheinungen in der Umgebung der Gletscher des Zemmgrundes (Zillertaler Alpen)*, Göttinger Geogr. Abh., 15.
Pulina M., 1974, *Denudacja chemiczna na obszarach krasu węglanowego*, Prace Geogr. IG PAN, 105.
Radwańska-Paryska Z., Paryski W. H., 1973, *Encyklopedia tatrzańska*, SiT, Warszawa.
Rapp A., 1960, *Recent development of mountain slopes in Kärkevagge and surroundings, Northern Scandinavia*, Geogr. Ann, 42 A, 2–3.
– 1974, *Slope erosion due to extreme rainfall with examples from tropical and arctic mountains*, Abh. Akad. Wiss. Göttingen, Math.-Physik. Kl., 29.

Rapp A., Nyberg R., 1981, *Alpine debris flows in Northern Scandinavia*, Geogr. Ann., 63 A, 3–4.

Rączkowska Z., 1983, *Types of stream channels in the Chocholowska drainage basin* (*The Polish Western Tatra Mts*), Studia Geomorph. Carpatho-Balcan., 16.

Rączkowski W., 1981, *Zróżnicowanie współczesnych procesów grawitacyjnych w Dolinie Pięciu Stawów Polskich* (*Tatry Wysokie*), Biul. Inst. Geol., 332, *Z badań geologicznych w Karpatach*, 22.

Romer E., 1929, *Tatrzańska epoka lodowa*, Prace Geogr. E. Romera, 11.

Selby M. J., 1974, *Dominant geomorphic events in landform evolution*, Bull. Intern. Ass. Eng. Geol., 9, Krefeld.

Skiba S., 1977, *Studia nad glebami wytworzonymi w różnych piętrach klimatyczno-roślinnych krystalicznej części Tatr Polskich*, Rocz. Glebozn., 28, 1.

Skierski Z., 1984, *Wiek i geneza Smreczyńskiego Stawu*, Prace i Studia Geogr., WGiSR UW, 5.

Starkel L., 1976, *The role of extreme (catastrophic) meteorological events in contemporary evolution of slopes*, [in:] *Geomorphology and climate*, E. Derbyshire (ed.), London.

– 1977, *Paleografia holocenu*, PWN, Warszawa.

– 1979, *The role of extreme meteorological events in the shaping of mountain relief*, Geogr. Pol., 41.

Stasiak J., 1984, *Wiek i ewolucja Szczyrbskiego Jeziora*, Prace i Studia Geogr., WGiSR UW, 5.

Staszic S., 1815, *O ziemiorodztwie Karpatów*, Wyd. Geol., Warszawa 1955.

Stingl H., 1969, *Ein periglazialmorphologisches Nord–Süd Profil durch die Ostalpen*, Göttinger Geogr. Abh., 49.

Szaflarski J., 1937, *Ze studiów nad morfologią i dyluwium południowych stoków Tatr*, Prace Inst. Geogr. UJ, 19.

Szeroczyńska K., 1984, *Analiza Cladocera w osadach niektórych jezior tatrzańskich*, Prace i Studia Geogr., WGiSR UW, 5.

Tricart J., Cailleux A., Raynal P., 1972, *Les particularités de la morphogenèse dans les régions de montagnes*, Centre de Documentations Universitaires, Paris.

Troll C., 1972, *Geoecology and world-wide differentiation of high-mountain ecosystems*, [in:] *Geoecology of the high mountain regions of Eurasia*, Proc. of the Symposium of the IGU Commission of High Altitude Geoecology, November 1969, Mainz, Erdwissenschaftliche Forschung, 4, Steiner, Wiesbaden.

– 1973, *High mountain belts between the polar caps and the equator, their definition and lower limit*, Arc. Alp. Res. 5, 3, part 2.

Vitásek F., 1956, *Snezna cara ve Vysokych Tatrach*, Geogr. Čas., 8, 4.

Wdowiak S., 1961, *Współczesny lodowiec karowy w Wielkim Kotle Mięguszowieckim nad Morskim Okiem w Tatrach*, Biul. Geol. UW, 1.

Wicik B., 1979, *Postglacjalna akumulacja osadów w jeziorach Tatr Wysokich*, Przegl. Geol., 7 (315).

– 1984, *Osady jezior tatrzańskich i etapy ich akumulacji*, Prace i Studia Geogr., WGiSR UW, 5.

Więckowski K., 1984, *Makroskopowa charakterystyka osadów dennych jezior tatrzańskich*, Prace i Studia Geogr., WGiSR UW, 5.

Wit-Jóźwik K., 1974, *Hydrografia Tatr Wysokich, objaśnienia do mapy hydrograficznej „Tatry Wysokie" 1 : 50 000*, Dok. Geogr., 5.

Zejszner L., 1856, *Über eine Längsmoräne im Tal Bialy Dunajec bei den Hochöfen in der Tatra*, Sber. Akad. Wiss. Wien, 21.

Zeller J., Geiger H., Röthlisberger F., 1976–1984, *Starkniederschläge des schweizerischen Alpen- and Alpenrandgebietes*, Birmensdorf.

WYSOKOGÓRSKI SYSTEM DENUDACYJNY TATR POLSKICH

Streszczenie

Cechą szczególną wysokich gór jest wykształcenie się w ich obrębie pięter geoekologicznych. Między procesami geomorfologicznymi a warunkami klimatycznymi, geologicznymi i florystycznymi występują określone relacje, które decydują o sposobie i tempie transformacji rzeźby gór. W Tatrach, reprezentujących wysokie góry niezlodowacone, zaznacza się wysokościowe zróżnicowanie procesów geomorfologicznych, będące skutkiem istnienia owych uwarunkowań środowiskowych. Obszar najsilniej przekształcony przez współczesne procesy znajduje się między górną granicą lasu, położoną na wysokości około 1500 m n.p.m., a wysokością 1950 m n.p.m. Są to subalpejskie i alpejskie piętra geoekologiczne. W tym pasie wysokościowym notuje się największą ilość gwałtownych procesów geomorfologicznych związanych z ekstremalnymi zjawiskami hydrometeorologicznymi (spływy gruzowe, obrywy skalne, lawiny śnieżne i śnieżno-wodne, płytkie ruchy osuwiskowe) oraz powszechnie występujące powolne ruchy masowe (spełzywanie pokryw glebowych i piargowych). Równocześnie jest to obszar, gdzie występuje największe nagromadzenie luźnych, mobilnych osadów glacjalnych, glacifluwialnych i grawitacyjnych. Osady te stosunkowo łatwo podlegają przemieszczeniu przez ruchy masowe wywołane ekstremalnymi zjawiskami hydrometeorologicznymi. Powyżej i poniżej tego pasa intensywność procesów słabnie lub w przypadku niektórych z nich całkowicie wygasa. Występowanie jednego pasa wysokościowego jest cechą szczególną gór niezlodowaconych, gdyż w górach zlodowaconych (Alpy, Kaukaz, Tian-Szan i innych) występują dwa pasy wysokościowe morfogenetycznie aktywne: wyższy – modelowany przez procesy peryglacjalne, i niższy – odpowiadający tatrzańskiemu, położony w strefie górnej granicy lasu i ponad nią, gdzie głównym czynnikiem mobilizacji zwietrzelin jest działalność intensywnych i wydajnych deszczów letnich i wiosennych.

Wysokie góry niezlodowacone średnich szerokości geograficznych charakteryzuje występowanie dwóch niezależnych systemów morfodynamicznych. System stokowy, modelowany głównie przez zespół procesów grawitacyjnych, pluwiograwitacyjnych, niweoablacyjnych i kriograwitacyjnych, występuje w piętrach seminiwalnym, alpejskim i subalpejskim. System korytowy jest kształtowany głównie przez zespół procesów fluwialnych. Obejmuje swym zasięgiem piętra leśne oraz częściowo wnika do piętra subalpejskiego. Niezależność systemów polega na braku łączności między nimi podczas przeciętnych warunków hydrometeorologicznych. Istnieje w ich obrębie wewnętrzna dynamika, lecz produkty wietrzenia i denudacji nie są przekazywane z jednego systemu do drugiego. Pozostają na ich obszarze i nie podlegają ciągłemu przemieszczaniu poprzez piętra geoekologiczne na przedpole gór. Przegłębione dna dolinne w Tatrach Wysokich, wypełnione jeziorami cyrkowymi i położone na różnych wysokościach nad poziomem morza są naturalnymi bazami denudacyjnymi. Zwietrzeliny przemieszczane ze stoków gromadzone są w misach jeziornych, nie są natomiast transportowane wzdłuż den dolin. Dzieje się tak dlatego, że progi skalne oddzielające zawieszone fragmenty dolin w naturalny sposób tamują ten proces. Tylko podczas zdarzeń hydrometeorologicznych o katastrofalnych rozmiarach, powtarzających się raz na 100 lat, dochodzi do łączenia obu systemów – lecz tylko w niektórych częściach gór i na bardzo krótki okres. Warunki takie

występują w Tatrach przy opadach deszczu rzędu 200–300 mm na dobę. Tymczasem w górach wysokich współcześnie zlodowaconych istnieje system morfodynamiczny lodowcowy lub denudacyjno-lodowcowy, powiązany z systemem fluwialnym. Duża dynamika wewnętrzna systemów i ich powiązanie pionowe poprzez piętra geoekologiczne sprawiają, że zachodzi wyprzątanie dużych ilości zwietrzelin poza obręb gór.

W okresie ostatnich kilkuset lat, to jest co najmniej od roku 1600, obserwuje się tendencję do aluwiacji systemu stokowego w Tatrach, podobnie jak ma to miejsce w Alpach. Aluwiacja jest związana ze zmianą reżimu klimatycznego, stwierdzaną w obu łańcuchach górskich na podstawie badań dendroklimatologicznych, oraz z działalnością człowieka w górach. Obserwuje się tendencję do zmiany profilów stoków gruzowych oraz bardziej selektywnego modelowania ścian i stoków skalnych. Cechą stoków poddanych aluwiacji są liczne generacje mezoform związanych z aktywnością spływów gruzowych. Ten trend rozwojowy wysokogórskiego systemu denudacyjnego Tatr zmierza do utworzenia łączności pomiędzy systemem stokowym i korytowym poprzez ukształtowanie profilu podłużnego stwarzającego optymalne warunki do przemieszczania zwietrzelin poprzez piętra geoekologiczne. Następuje więc adaptacja systemu wysokogórskiego utworzonego w plejstocenie, a zwłaszcza w późnym glacjale, do holoceńskich warunków klimatycznych strefy umiarkowanej.

Górna granica lasu, chociaż jest granicą bardzo istotną pod względem morfodynamicznym, nie stanowi ostrej zmiany w dynamice stoków. Wysokościowe zasięgi jej położenia podczas holocenu sprawiły, że istnieje raczej pas przejściowy, a w jego obrębie obserwuje się stopniową zmianę intensywności procesów geomorfologicznych. Na górnej granicy lasu istnieje tylko stopniowe zróżnicowanie intensywności procesów denudacji chemicznej, zwłaszcza na podłożu skał węglanowych.

Bezwzględne wskaźniki intensywności procesów geomorfologicznych, zwłaszcza odpadania i powolnych ruchów masowych, osiągają podobne wartości w Tatrach, jak i w innych górach średnich szerokości geograficznych. Natomiast efektywność gwałtownych procesów, zwłaszcza spływów gruzowych, jest różna w poszczególnych masywach górskich przy podobnej wielkości wywołujących je zjawisk hydrometeorologicznych. Geomorfologiczne skutki działania gwałtownych ruchów masowych zależą bowiem od stabilności rzeźby oraz jej parametrów morfometrycznych. Formy utworzone przez gwałtowne ruchy masowe w Tatrach osiągają rozmiary znacznie mniejsze niż w wysokich masywach łuku alpejskiego, lecz większe niż w górach Skandynawii. Po zaniku lodowców w Tatrach stoki i dna dolin, zwłaszcza ich dojrzałe górne odcinki, uzyskały dużą stabilność. Powstały niezależne systemy morfodynamiczne na stokach i w dnach dolin, nie kontaktujące się z sobą.

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