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Edited by

ALFRED JAHN

CONTENTS

PREFACE

The present volume represents a collection of papers by Polish geographers, who while cooperating with the International Geographical Union, have carried on their research in different climato--geomorphological zones. The majority of these papers refer to the temperate zone since Poland belongs to this climatic region. A considerable amount of research was done in high-mountainous zones such as the Peruvian Andes, Tibet (Kunlun) and the Tatra Mts.

Polish geographers have also been active in the polar zone, chiefly in Spitsbergen where a permanent Polish research station is located. Polish research penetration has been intensified in the warm climatic zone of America, Asia and Africa.

The papers included in this collection are to be regarded as selected examples of current Polish research in the zones mentioned above. The present volume has been prepared for the 27th International Congress to be held in Washington, USA, in 1992.

The Editor

GEOGRAPHIA POLONICA 60, 1992

SLOW SOIL MOVEMENT AS A GLOBAL PHENOMENON

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ABSTRACT. Slow mass movement of the soil is a form of transporting of slope material. It can be classified according to: the dynamics of the process, the structure of the movement (particles, layers) and the relation to the morphological surface (shallow or deep movement). In this way it is possible to distinguish soil creep as the important process within this group of mass wasting factors. Seasonal creep is first of all a function of climate whereas continuous or rheological creep is a function of geological structure.

The slow movement of the soil depends on local factors (slope angle, moisture of the soil, vegetation cover) and on latitudinal and altitudinal climatic zonal factors (temperature, precipitation). The rate of soil movement in the polar (periglacial) zone is the highest on the globe. It is, as mass movement of the soil, three times higher than in the alpine zone. The soil creep processes in temperate climates are relatively small but greater than in the tropic zone. The smallest soil movement is in the arid and semi-arid zone.

1. INTRODUCTION

Slow mass movement as a process forming the slopes of the earth's surface was known in the 19th century, and it was mentioned by Thomson as early as 1877 (Young 1972). It is Davison (1889), however, who deserves the greatest credit for his research in this field. He was the first to carry out experiments on changes in the position of rock blocks under the influence of temperature changes, thus providing a quantitative evaluation of the process.

Götzinger (1907) recognized the morphological significance of slow mass movement. He considered these movements, which he called *Gekriech*, as an essential factor in the modelling of mountain ridges. Also Gilbert (1909) associated the upper convexity of hillslopes with the action of slow mass movement. W. Penck (1924) believed that these movements occurred in all climates and environments. According to him, forests do not protect the soil from this movement as the root system acts like a sieve.

The modern view on the problem began with the classical definition of Sharpe (1938) who using the term "creep of soil" described the process as a "slow downslope movement of superficial soil or rock debris usually imperceptible except for observations of long duration". The functioning of this process from the viewpoint of soil mechanics was described by Terzaghi (1950) who realized the difference between "seasonal" and "continuous" creep.

Soil creep is thus the subject of research in both geomorphology and soil mechanics. Its study necessitates quantitative investigations and measurements, hence extensive field and laboratory experiments. The purpose of this paper is to present information collected about these investigations from various scientific centres in the world. To this end (1984) a questionnaire was distributed through the IGU Commission on Field Experiments in Geomorphology. Fifteen replies were received (see Table 1), but many other stations report

TABLE 1. Sites of slow mass movement measurements of the soil — results of questionnaire (1984)

Location	Climatic zone	Duration of measurements	Method	Researcher
Canada British Columbia	temperate semi-arid	long-term	inclinometer stakes	M.J. Bovis
USA Tennessee	temperate	long-term	deformation of artificial structures	G.M. Clark
N. England	temperate maritime	long-term	young pit	A. Young
Poland Sudetes	temperate, high mountains	long-term	buried pegs	A. Jahn
Luxembourg	temperate maritime	long-term	vertical velocity column	P.D. Jungerius
Belgium Ardennes	temperate maritime	long-term	inclinometer	A. Pissart J.L. Schepers
France Vosges	temperate maritime	long-term	extensiometer	A.V. Auzet
Romania E. Carpathians	temperate	long-term	young pit	N. Bacaintan
Rwanda Central Africa	savanna	long-term	young pit	J. Moeyersons
Zaire	tropical	long-term	T-shape plates pegs, plastic rings	J. Soyer
Japan Mt. Shirouma	high mountain	long-term	plastic tubes	Shuji Jvata
Japan Konto Plain	temperate	short-term	trap method	Yago Ono
Japan Usu Volcano	temperate	short-term	inclinometer	Hiromitsu Yamagishi
Japan Niigata	temperate	short-term	geomorphological mapping	Yukinari Fujita
SW Japan	temperate	long-term short-term	seismological electric methods	Takashi Fujita

systematic soil movement measurements in the literature. Data from the questionnaire have been supplemented with published material available from earlier investigations (Tables 2–5). However, the data available are still insufficient for a satisfactory synthesis. The precision of the investigations which makes it possible to detect not only seasonal but even daily soil movements allows us to understand the mechanism of the process but not its long-term, geomorphological effects. On the other hand, it is clear that long-term investigations lack precision for mechanism interpretation. They reveal the cumulative effects of the process but offer no basis for a more accurate determination of its mechanism.[1]

[1] The source and base of this elaboration are main materials completed by the Commission on Field Experiments in Geomorphology, International Geographical Union, in 1980–1983. It is in that time that

2. CLASSIFICATION OF MASS MOVEMENT PROCESSES

Carson and Kirkby (1972) recognize three types of soil movement: heave, slide and flow of the soil. A triangular diagram presents these processes in relation to one another (Fig. 1). Seasonal soil creep as a slow movement (expansion and contraction of the soil) is chiefly connected with the heaving of the soil. These authors do not regard the gravitational "fall" of rock material as a mass movement.

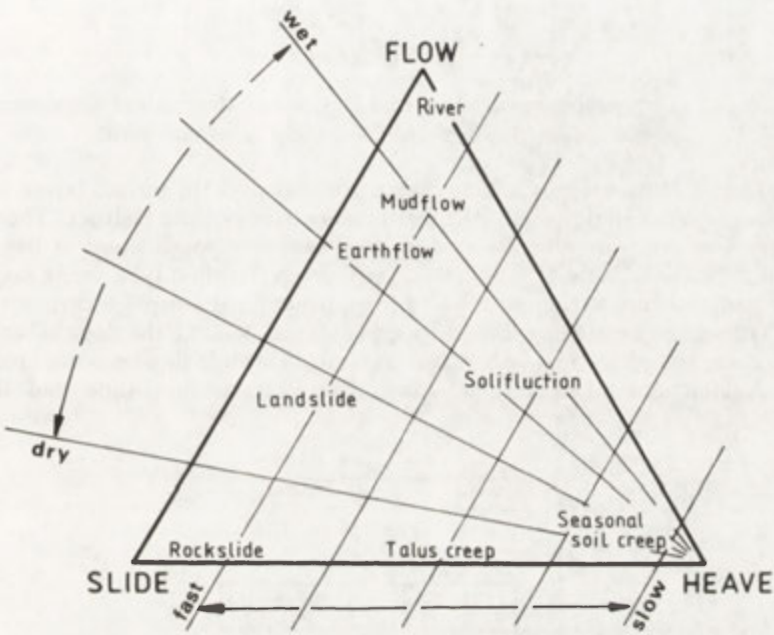

Fig. 1. Classification of mass movement processes — after Carson and Kirkby (1972)

An attempt could be made to present the position of soil creep in the form of a model with its sides constituting a coordinate system which determines the type and dynamics of the movement (Fig. 2). There are three main types of movement: (a) fall, (b) flow and slide and (c) creep. Fall is a purely gravitational movement, dry, i.e. water does not participate in it, directed at different angles to the slope surface. Flows and slides are movements in which water has a share as a transporting agent, directed parallel to the surface of the ground. Creep is a movement in which water may or may not participate, acting perpendicularly or parallel to the slope. Fall may only be rapid and creep may only be slow. Flowing and sliding movements may be of varying velocity, hence their directional connection with both rapid and slow movement. In the diagram, "rapid mass movement" (RMM) and "slow mass movement" (SMM) are distinguished as two areas which border upon each other in the place denoted as flowing and sliding movement.

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the Chairman of the Commission, Professor O. Slaymaker (University of British Columbia, Vancouver, Canada) undertook the task to publish the paper what had been accepted not only by the author but also by numerous groups of scholars in different countries, participating in collecting materials (cf. Questionnaire). I do not know the reasons why this paper has not been printed in the announced publishers of the Commission.

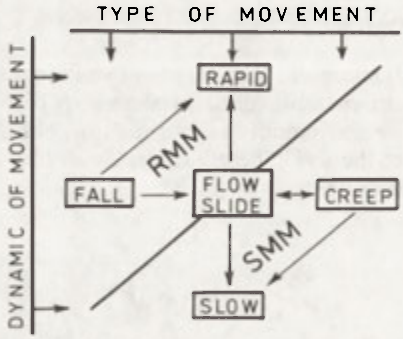

Fig. 2. Classification of mass movements with regard to type and dynamics of the movement: RMM – rapid mass movement, SMM – slow mass movement

Slow movement of the creep type may extend not only over the surface layers, e.g. the soil, but it may also reach deep down into the weathering cover and the bedrock. Thus a shallow and a deep type of creep may be distinguished, which corresponds more or less with what Terzaghi called "seasonal" and "continuous (mass)" creep. The first type, being mainly due to temperature and moisture variations (also to frost, hence "frostcreep") is certainly a shallow movement. The second type being caused by gravitational action ("rheological" creep) affects deeper soil masses and rocks (Fig. 3). Shallow movements include slow seasonal creep and also solifluction (gelifluction) when vertical creep (heave) turns into slide and flow. Deep

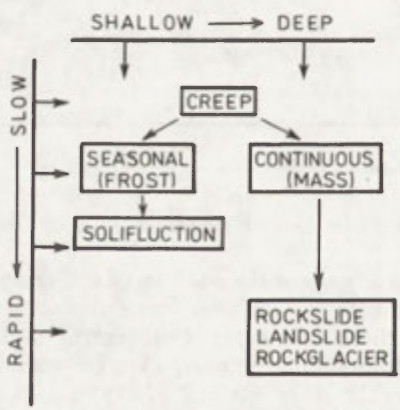

Fig. 3. Classification of mass movements with regard to velocity and depth of the movement

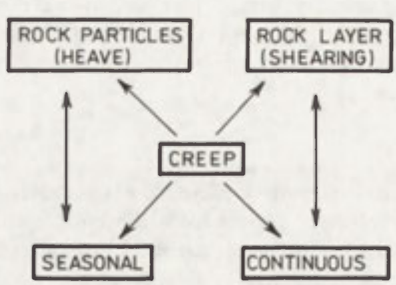

Fig. 4. Classification of mass movements with regard to the material structure

movements include rheological creep (continuous) which, with growing dynamism, turns into landslide, rock glacier, and the like.

Finally, creep-type movements may be interpreted with regard to the nature of the rock debris participating in them (Fig. 4). The movement may affect individual soil particles or aggregates. Heave is then the actual process of the movement, which is the essence of seasonal creep. Slow creep may take place with the participation of whole soil and rock layers separated from one another by boundary surfaces. This creep results from shearing stress and may affect both shallow and deep layers of the slope.

3. MORPHOLOGICAL SIGNIFICANCE OF SLOW MASS MOVEMENT

Creep movement is a factor of mass transportation on a slope and its denudational effects are more difficult to assess. The movement rate is high on the soil surface and decreases with depth. If we assume the mean velocity on the surface to be 1 cm year^{-1}, it is still low as compared with the depth range, i.e. 20–50 cm. Hence the first conclusion: the mass of material moved in this way depends on the depth of the movement, that is on its component perpendicular to the slope, rather than on the component parallel to the slope surface. Therefore even a high rate of surface movement, amounting to 2–3 cm year^{-1}, is of only slight importance for mass transportation, if it concerns only a thin soil layer.

In shearing movement the whole layer is affected, thus its rate is determined by the thickness of the layer. If the value of this movement is expressed as 1, then the rate of a normal movement (presented as a concave curve) will be only 1/4, which can be seen in the appended diagram (Fig. 5). Thus taking into consideration the cross-section surface of the mass moved on the slope, the following formula is applicable:

$$S = d\,h\,0.25$$

where

d = annual creep rate on the surface,
h = depth of movement in the soil.

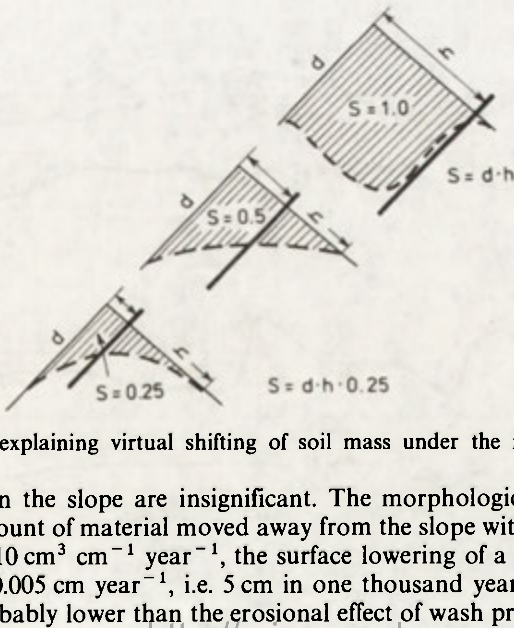

Fig. 5. Diagram explaining virtual shifting of soil mass under the influence of creep

The creep effects on the slope are insignificant. The morphological degradation can be calculated from the amount of material moved away from the slope within a year. If the rate of soil creep amounts to 10 cm^3 cm^{-1} year^{-1}, the surface lowering of a slope being 100 m long can be determined as 0.005 cm year^{-1}, i.e. 5 cm in one thousand years or 50 Bubnovs (Jahn 1981). This value is probably lower than the erosional effect of wash processes and gullying on the slope. Thus in temperate climatic conditions where the surface creep movement is 2–3 mm year^{-1}, no great morphological importance should be attached to creep processes. (Caine 1979, agrees, though see Kirkby 1967, for a different interpretation). This process has a different meaning in the periglacial zone where the maximum frost creep effect is 1 m in the depth, with an annual rate of 4 cm, which gives a significant degradation of 1 m in one thousand years (1000 Bubnovs).

The analysis of soil creep as a denudational agent indicates that its effects are more dependent on the depth of movement than on its rate at the surface. The effects of the movement are relatively greater on short slopes than on long ones. This contrasts with other erosion processes in which the effects increase with slope length.

By defining the morphological significance of creep in this way, it becomes possible to explain the known effects that this process has on the shaping of mountain divides and ridges, as it was understood by Götzinger (1907) and Gilbert (1909). For divides and ridges are the only elements of relief where the carrying away of material over very short distances produces visible morphological effects. It is on the divides only that the vertical component of movement is of importance, producing the greatest effect in places from which the material is carried away in two opposite directions. Owing to creep action therefore, the denudation balance of divides is always positive, since the removal of the material progresses faster than weathering (Jahn 1968). A noticeable effect is the subsoil rock uncovered on the divide when the slopes are covered with detritus.

4. SLOW SOIL MOVEMENT IN DIFFERENT CLIMATIC ZONES

As in all geomorphological processes the principal question about slow soil movements is to what degree they are dependent on local factors and to what degree on global ones. There is general agreement that slope inclination is of particular importance among local factors, hence slow soil movements are considered to be linked with the sine of the slope angle. Washburn (1967) having discovered on Greenland an immense difference between the mean solifluction movement in dry places (0.9 cm year^{-1}) and that in moist ones (3.7 cm year^{-1}) found soil

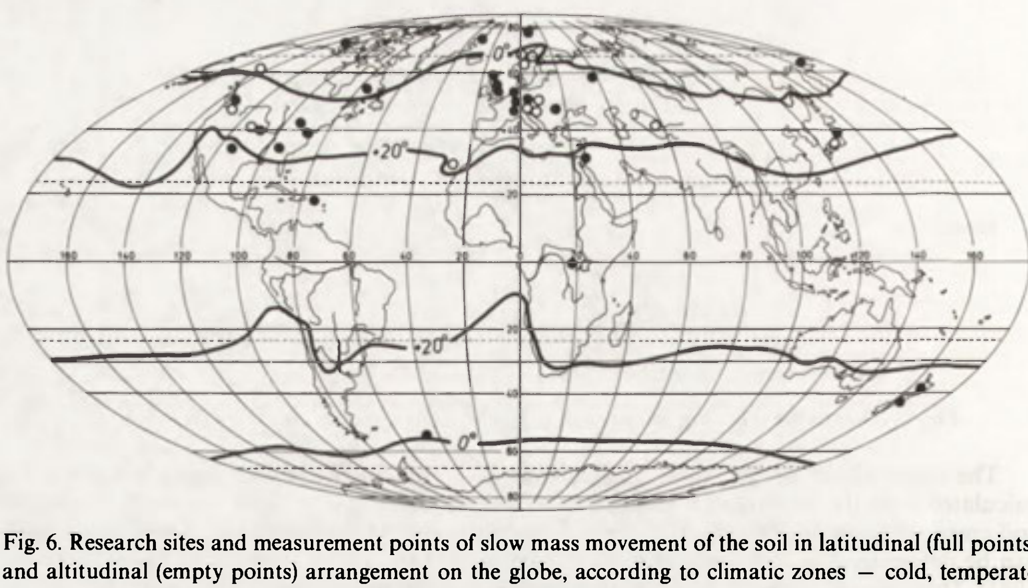

Fig. 6. Research sites and measurement points of slow mass movement of the soil in latitudinal (full points) and altitudinal (empty points) arrangement on the globe, according to climatic zones — cold, temperate and warm (isotherms of annual mean temperature)

TABLE 2. Slow mass movement of the soil in the polar (periglacial) zone

Location	Method	Years	Altitude m	Slope angle	Surface cover	Movement depth cm	Movement rate mm/yr	Movement volumetric $cm^3/cm/yr$	Source
Banks Island	tinfoil markers	3	130	2–4°	tundra	50	20	50	French (1974)
Schefferville Quebec Labrador	plastic tubes	5	550	6–8°	scare vegetation	70	100	175	Williams P.J. (1966)
Mesters Vig, NE Greenland	targets	5	50–250	10–14°	tundra, bare slope	100	9–37	45–185	Washburn (1967)
Smith Georgia	pegs	2	120	21°	bare slope	25	50	87	Smith (1960)
Spitsbergen	pegs	2	30	7°	tundra, bare slope	40	40	80	Jahn (1961)
Anadyrskiy Kray Siberia	linear motion transducer	1			tundra, soli-fluction	100	22	110	Zhigarev (1960)

TABLE 3. Slow mass movement of the soil in the alpine zone

Location	Method	Years	Altitude m	Slope angle	Surface cover	Movement			Source
						depth cm	rate mm/yr	volumetric $cm^3/cm/yr$	
Garibaldi Park, B.C. Canada	tubes, strings, sand column	10	1600 – 1900	10 – 15°	bare slope sorted stripes	10	250	125	Mackay, Mathews (1974)
Ruby Range Yukon Territory Canada	plastic tubes	2 – 5	1675 – 1980	14 – 18°	tundra	47	19	44.6	Price (1973)
Front Range, Colorado, USA	pegs	5	3500 – 3700	6 – 17°	turf (turf banked terraces)	50	30	75	Benedict (1970)
Karkevagge Sweden	pegs	13	700 – 1000	15°	tundra vegetation	25	20	25	Rapp (1960); Rapp, Stromquist (1979)
Karkevagge Sweden	pegs	8	700 – 1000	30°	bare slope	20	10	10	Rapp (1960)

cont. Table 3

Norra Storja, Lappland Sweden	pegs	7	1300	10–15°	bare slope	50	5	12.5	Rudberg (1964)
Norra Storja, Lappland Sweden	pegs	7	600–900	10–15°	tundra vegetation	50	18	45	Rudberg (1964)
Karkonosze Sudetes Poland	pegs	17	1200	8–39°	alpine meadow	20	15	20	Jahn (1981)
Śnieżnik Sudetes Poland	plastic tubes	6	1400	13–24°	alpine meadow	10	5	10	Jahn (1981)
Canary Islands	pegs	1.5	2140–2370	10–12°	bare slope	7	125	62.5	Höllermann (1979)
Munt Chavagi Alps Switzerland	pegs	1	2400	3–10°	grass land	30	40	30	Furrer (1972)
Carinthia E Alps Austria	pegs	6	1900–2100	27–41°	alpine meadow	50–75	12.4	19	Stocker (1979)

TABLE 4. Slow mass movement of the soil in the temperate zone

Location	Method	Years	Altitude m	Slope angle	Surface cover	Movement			Source
						depth cm	rate mm/yr	volumetric $cm^3/cm/yr$	
South Pennines, England	young pits	12	350	25°	grass land	60	1	0.61	A. Young (1978)
S.W. Scotland	young pits t-shaped pegs	2	300–500	13°	grass land sheep grazing	20	2	2.1	Kirkby (1967)
Wales	young pits	3	250–600	26°	grass land sheep grazing	30	2.1	2.7	Slaymaker (1972)
North Pennines England	different methods	1.5	367–487	11°	grass land	50	1.3	1.6	Anderson and Cox (1978)
Baltimore Maryland, USA	different methods	3		17°		25	4	2.5	Carson and Kirkby (1972)

cont. Table 4

Kletno Sudetes Poland	plastic tubes	5	600	15–24°	grass land pasture	10	10	2.5	Jahn (1979)
Taunus Germany	plastic tubes	5	260–800	4–23°	forest	15	4	2.0	Göbel (1977)
Belgium Liege	pegs with nylon thread	1.5	100	27°		15	1–5	1.1	Schepers (1977)
Vosges France	extensio-meter	2	750–940	17–22°	grass land forest	15	10	7.5	Auzet (1982)
Carpathians	young pits	4	600–625	8–17°	grass land pasture	32–45	4	0.64	Bacaintan (1983)
Kazan Solviet Union	young pits	3–4	200–300	22°	grass land	75	2	7.26	Dedkov and Duglav (1967)
Central North Island New Zealand	buried cones	3.5		7–33°	grass land pasture	75	7.1	28	Selby (1974)
Chilton Vailey New Zealand	plastic tubes						11	3.2	Owens (1969)

TABLE 5. Slow mass movement of the soil in the warm zone

Location	Method	Years	Altitude m	Slope angle	Surface cover	Movement			Source
						depth cm	rate mm/yr	volumetric $cm^3/cm/yr$	
Western Colorado, USA	different markers	3	1500–1800	15–18°	sparce vegetation	5	6–12	2.5	Schumm (1964)
Santa Fe N. Mexico, USA	pins			45°	bare slope		5	4.9	Leopold et al. (1964)
Puerto Rico	young pits	2–3	260–740	17–19°	forest	50	4	10	Lewis (1974)
S. Rwanda Africa	young pits	1–5	1800	0–35°	savanna vegetation	100	10	50	Moeyersons (1981)

moisture to be uppermost in the agents of movement. He attached greater importance to soil moisture than to vegetation cover. Vegetation cover is favoured by other workers as the most important local factor (Dedkov and Duglav 1967; Young 1972; Price 1973; Selby 1974; Lewis 1974). There can be no doubt that the type of material, the local activity of animals as well as human activity exert a considerable influence on soil movement. Moreover, the research methods themselves, the different techniques applied may sometimes be decisive for the results, as was pointed out by Kirkby (1967), Benedict (1970), Young (1972), Anderson and Cox (1978). So the variety of local factors as well as the diversity of investigation methods influence and impede a synthetic evaluation of slow soil movements in connection with more general causes.

It is a matter of common knowledge, however, that seasonal creep depends on such essential climatic agents as temperature course and distribution and also on the sum of atmospheric precipitation. On this assumption the IGU Commission of Field Experiments in Geomorphology and, earlier, the IGU Commission on Present-Day Geomorphological Processes have attempted to collect material from various climatic zones of the earth to answer the question of the role of climate in controlling slow mass movement rates.

All known measurement sites were taken into consideration. The map appended (Fig. 6) shows the points where the sites are situated. The isotherms of the annual mean air temperature 0° and +20°C make it possible to approximately delimit large zones of cold and temperate climates in both hemispheres and zones of warm (arid and tropical) climates between both tropics. Of these 34 sites, 18 (53%) are situated in zones of cold climates, including 6 in the periglacial zone (Table 2) and 12 in the alpine zone (Table 3), that is above the upper timberline. There are 13 (35%) measurement sites in the temperate zone (Table 4) and only 4 (12%) in the zones of warm climates (Table 5). Quantitative investigations of slow mass movements are most common from the zone where these movements are relatively vigorous, that is in cold climatic zones. It should be added that the present author resolved to treat both forms of soil movement, frost creep and solifluction, in this zone as one, since they are practically impossible to differentiate in terms of effects. Together they make up slow mass movement.

The tabulation comprises slope inclination, the method and time of measurement as well as the presence of vegetation. In this way the evaluation of the movement rate may be corrected to a certain degree from the viewpoint of local influences. The most important data can be found in the column showing annual mean rate of movement. These refer to the mean values of several years and from different profiles, in so far as these values could be obtained at a given site. Unfortunately, there is no way of avoiding error derived from the use of different profiles. The median of the movement might have been a better choice. It should be emphasized that here the movement in question is the largest in each vertical profile regardless of the level on which it occurs. In the majority of cases the largest movement takes place at the surface of the soil.

Measurements of boulder movement on the soil surface were relinquished when it turned out that the movement showed colossal differences — from several centimetres to several metres a year (see Michaud and Cailleux 1950, Caine 1968). Pissart (1964) maintained that small stones moved faster than big ones, and Benedict (1970) wrote that the movement of these stones "is rapid and usually random because of the tendency of small stones to roll and slide downslope after they are dislodged by surficial frost creep, needle ice and raindrop impact" (p. 211). Surface stones can therefore not be taken as indicators of soil movement, particulary when we are trying to detect climatic causes in this movement. An important item is the volume of the shifting soil material at 1 cm cross-section in the course of a year (in cm^3). This value was calculated only when the depth reached by the movement was known. The calculations were performed according to the principle expressed above, that is a different multiplier was used in dependence on the type of the movement. With a concave movement curve this multiplier was 0.25, with a convex one 0.50 and with a full shearing movement (displacement of the whole layer) it was 1.0.

5. ZONE OF COLD CLIMATES (POLAR AND ALPINE)

The basic form of slow soil movement on slopes is frost creep being strictly dependent on frost heave. The mechanism of process with regard to its activity in the periglacial polar zone is particularly well known from Washburn's investigations (1967, 1979), and its activity in the alpine zone from Benedict's work (1970, 1976). Washburn gives the following definition of the process: "the ratchetlike downslope movement of particles as the result of frost heaving of the ground and subsequent settling upon thawing, the heaving being predominantly normal to the slope and the settling more nearly vertical". The settling takes place on the principle of retrograde displacement which is an "upslope component of movement" (Benedict 1976).

The other process in the slow mass movement group of polar regions is gelifluction, the working mechanism of which belongs to movement of flow and slide type. It is a movement of the whole soil layer, which often functions on the principle of shearing (Williams 1957). Washburn (1979) is of the opinion that: "...contrary to Sharpe (1938), solifluction is excluded from creep and regarded as a separate process". He based his views on investigations carried out in Greenland (Washburn 1967); but where the effects of slow slope movement of the soil are concerned he acknowledges a fairly equal share of both processes, even with a frequent predominance of frost creep over gelifluction. Schepers (1977) estimated the share of frost creep in the soil movement of the Belgian climate at 40–85%. Rapp (1960) distinguished "frost creep" and "solifluction" in the Scandinavian mountains, but linked the effects of both processes together. The differences in their velocity are very small anyway, and so the mean rate of solifluction is 2 cm year^{-1} and that of "talus creep" 1 cm year^{-1}. Benedict (1976) wrote: "Acting in various combinations frost creep and gelifluction produce distinctive lobate and terrace-like landforms — they commonly operate together".

The same conclusions were reached by the present author, although he had earlier tried to distinguish these processes from each other in Spitsbergen (Jahn 1961). After all, these two are the only slow movement processes in the environmental conditions of polar and alpine regions. If they do not differ in velocity, their morphological effects are hard to distinguish, and this is the chief argument for treating both processes as one. In the present report account was taken of 18 selected measurement series from cold climates — 6 polar (periglacial) and 12 alpine ones (Table 2 and 3). In the polar series the predominating region is the Arctic, whereas from the Antarctic region only single observation results were obtained, with the exception of a two-year measurement series carried out by Smith (1960) on South Georgia. Of the 12 alpine series of measurements one part comes from the Scandinavian peninsula where both cold zones are very close to each other. The Alps have numerous measuring sites, though not all of them have been included in this study. The farthest south-reaching mountains of the northern hemisphere, where soil movements were studied high above the timber line, were the Canary Islands and Central Japan (incomplete data), the highest places on the earth where frost creep and gelifluction soil movement measurements in cold climate environments was carried out by the peg method (12 sites).

As to the periglacial environment the mean velocity of the soil surface layer amounts there to 4 cm (42.5 mm) per year, whereas the whole layer involved in the movement is 60–70 cm thick. This soil movement rate was found in the classic periglacial areas of Spitsbergen and the wet tundra covers of Greenland.

In this respect the alpine zone of the northern hemisphere is known by its 10 well developed sites. The mean value of the soil surface movement is there almost 1.7 cm year^{-1}, the mean depth of the movement being 36 cm. Two sites (shown in Table 3) with extreme movement values — Garibaldi Park near Vancouver, Canada, and the Canary Islands — have not been included in this list. Both sites reach close to 2000 m of absolute altitude, the soil movement at both is very shallow (to 10 cm deep) and chiefly involves needle ice. The latter fact explains the tremendous rapidity of the movement, which is in the range of 2–3 m year^{-1}. A similar phenomenon of rapid soil movement under the influence of needle ice was observed in the Tatra mountains (Gerlach 1959).

Thus soil movement velocity is nearly three times higher in the polar than in the alpine zone, and the layer involved is twice as thick.

It follows that the volume of the soil mass transported on high latitude slopes must also be larger than of the temperate alpine zone. The relevant data are: 103 $cm^3 cm^{-1} year^{-1}$ and 30 $cm^3 cm^{-1} year^{-1}$. This means that the soil mass transported on the polar slopes is at least three times that of alpine slopes. The polar zone is therefore a privileged environment as far as transport and, naturally, denudational activity of slow mass movement are concerned, and that even in relation to the climatically similar alpine zone.

Although I wrote before (Jahn 1978): "In cold climate environments the features of mass wasting on slopes are usually fairly identical, irrespective of the presence of permafrost" and further ..."permafrost does not control the mechanism of soil movement but exerts a control on the depth of the movement", I think now that this view needs to be completed. The mechanism of this movement is doubtless the same in both environments (the same type of movement curves) but its effects differ not only with regard to depth but also to velocity.

In periglacial conditions there are two predominant motive forces of movement: frost heave and soil moisture. The first agent is expressed by the amplitude of soil surface heave which in extreme cases may even reach 0.5 m $year^{-1}$. The other factor is the impermeability of the permafrost subsoil (substratum) to soil water, which creates conditions for a large water retention. Washburn (1967) rightly regarded moisture as an agent of prime importance among the factors that facilitate slow soil movement in the polar environment. Moisture contributes to the formation of segregation ice structures in the soil, which is also important to the movement in the time when the soil melts in the active layer of permafrost. It is very likely that permafrost itself, that is soil in frozen state, might shift in summer. This fact was pointed out by J.R. Mackay (1980) who detected expansion movements of the frozen soil in positive temperatures.

Finally a short remark on these exceptional alpine sites in the Vancouver region and on the Canary Islands. When the soil movement is shallow and rapid, the amount of transported mass can approach that in polar zones. Such is the mechanism of frost processes in the high mountains of the tropics, which was recognized by C. Troll as early as in 1944. He called the process *Kammeis Solifluktion.*

6. TEMPERATE ZONE

Of the 13 selected slow movement measuring sites in this zone, ten are in Europe, one in the USA, and two in New Zealand (see Table 4). These are series lasting mostly 2—5 years, the longest, however, being a 12 year series conducted in the South Pennines in England. Half of the measurements was carried out by means of "young pits." Contrary to those in alpine and particularly polar regions, the measurements in temperate zones took place in man-controlled areas (cattle and sheep pasturing, skiing, tourism). This fact depreciates the value of the material as an indicator of the natural process of soil movement. This remark concerns those sites where maximum movement values of up to 10 mm $year^{-1}$ were recorded (Vosges, Sudetes). If these sites are not included in the statistics, the mean of the surficial soil layer movement is 2.5 mm $year^{-1}$, which seems a conservative value. Actually, the natural soil movement in European conditions has long been supposed to amount to about 1–2 mm $year^{-1}$. It should be emphasized that this movement occurs in grass-covered, meadowy regions. Less is known about what takes place in forests. According to the 5-year measurement series in the forests of the Taunus Mts. (Gõbel 1977) the movement there reaches almost 0.5 cm $year^{-1}$. Although this value seems high, it is possible that the movement is in some way accelerated by human activity. The depth of the movement reaches an average of 40 cm, but the thickness of the layer involved in the transport is not greater than 20 cm. It is thus a shallow movement, and that is why the mass of material removed by it is

$3.0\ cm^3cm^{-1}year^{-1}$. This mean value does not include the New Zealand measurement (Selby 1974) which relates to soil of very specific material, that is pumiceous soil.

It follows from these data that slow soil movements, i.e. creep, are relatively small in temperate climates. Their morphological meaning is negligible, and even if there exist any perceptible effects of the process, they are visible only in the zone of divides, which has been mentioned above. In the European environment, creep action is responsible for the upper convexity of hillslopes. Unfortunately, the measurements of this process have so far been rather random and the local, environmental changes have not been taken into account. The problem posed by W. Penck referring to slow mass movement in forest areas with a strongly knit root network still remains unsolved.[2] In temperate climates there is practically no solifluction in the sense in which it is known from polar zones. Yet there exist slow soil movements in which pulsation (expansion-contraction) is associated with flow or slide movements, i.e. transition from creep to earthflow and landslide. Theoretical assumptions indicate that this movement ought to be faster and more effective in a temperate climate of marine rather than of continental variety. Meanwhile the measurement results from England (where the best and most reliable material has been obtained), that is the data provided by Kirkby (1967), Anderson and Cox (1978), Young (1978), are almost identical with those of Dedkov and Duglav (1967) from the Volga region, the continental area of European Soviet Union.

Now a final remark on the zone under discussion. Slow mass movements in temperate zones are being investigated at sites situated in highly civilized countries. These sites, being equipped with the best possible technical instruments, can therefore provide a very close insight into the actual mechanism of the creep movement. The widely developed method of short-term (daily) measurements has its advantages and disadvantages. There still exists the danger of neglecting the essential goal of these investigations, that is the long-term geomorphological effects.

7. ZONE OF WARM CLIMATES (ARID AND TROPICAL)

This zone comprises arid, semi-arid, savanna and rainy tropic climates. Unfortunately, the measurement data obtained from this zone are extremely scarce and are limited to only several measurement series (Table 5). Some had to be discarded because of the unreliability of the results. Much discussed were, for instance, S.A. Schumm's (1964) measurements of several years in the semi-arid regions of Colorado, where the author found a creep movement rate which on the surface amounted to $6-12$ mm $year^{-1}$. The movement was very shallow (5 cm) and – according to Schumm – much affected by frost action. Schumm compares the soil movement rate in Colorado to that on Greenland (Washburn) and northern Scandinavia (Rapp). Schumm's measurements were later called in question on account of their interpretation rather than because of their results. And so Young (1972), referring to Kirkby's opinion, maintains: "... the markers used in this work were subject to considerable disturbance by livestock" (p. 52), and Carson and Kirkby (1972) write: "Schumm's data for the movement of surface markers (1964) has been included, although it is a particular rather than a mass movement, and probably includes some movement due to wash processes" (p. 289).

The same objection may be raised to the measurements of Leopold et al. (1966) in the semi-arid regions of New Mexico where this team recorded a peg displacement of even 8.4 mm $year^{-1}$ with a mean value of 5 mm and a large slope inclination (45°).

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[2] After I had already written this paper, I published the results of my many-years' measurements of soil creep movement in the Sudetes. Pegs columns, set into the soil of the old forest on steep slopes, do not show any displacement after two years. It is quite unexpected fact, even surprising. It can not be attributed to a measurement error. It was found in many places. The thesis about the existence of soil movement in the forests should be deeply reconsidered.

Should then the possibility of creep action in arid conditions be doubted altogether? This problem has puzzled Yair and de Ploey (1979) who, however, found traces of block movement on desert slopes ("dry block creep"). The laboratory tests carried out by these authors have proved that the changes in the block volume caused by temperature variations should – apart from other influences – lead to a steady block movement in the range of 1 mm year^{-1}. This value seems plausible and thus acceptable as a minimal movement rate.

More than any other zone, the arid and semi-arid zones require systematic field investigations of many years. Random observations are quite insufficient for a quantitative evaluation of the soil movement process. It should be kept in mind that certain chemical and physical processes accuring in the soil and the slope debris, such as salt crystallization, hydration and efflorescence, contribute to the sorting of the soil and its downslope displacement. Knetsch (1950), basing on investigations in the Libyan deserts and Kaiser (1970) who carried out investigations in the Tibesti mountains went so far as to write about a solifluidal texture of slope covers, about a "dry" and "salty" solifluction resembling gelifluction of the periglacial zone. If therefore the debris movement caused by expansion and contraction reaches 1 mm year^{-1}, then additional impulses may increase the movement several times. Such an impetus may also be given by frost action which is a most significant morphological factor in the higher areas of desert mountains (Troll 1944). Proof is the enormous soil movement in the peak parts of the Canary Islands mountains (Höllermann 1979). Even in the mountains of the Negev Desert in Israel, where 30–40 frost cycles have been recorded in the course of a year, frost is an important motive force of soil movement (Yair and De Ploey 1979). In some measure justice has also to be done to Schumm and his interpretation of soil movement in the semi-arid Colorado environment. Schumm (1964), comparing the exceptionally strong soil movement and erosion effects in a dry climate with those in a temperate climate, writes: "The sparse vegetational cover allows frost action and rainbeat to become very effective in arid and semi-arid regions, where they are less effective under the dense vegetational cover of humid temperate regions" (p. 235).

Fewer doubts arise in connection with the problem of slow soil movements in tropical climates of either variety – the temporary, savanna conditions as well as the conditions of rain tropics. These movements are furthered by the nature of the soil which is a fine-grained mass with a large content of clay particles. Seasonal creep results from a soil volume change developed under the influence of wetting and drying. This change is certain to be the chief factor but the role of termites is still unknown. The problem of "stone lines" is still unclear. No one has yet measured by way of field experiments whether these stones have been buried because termites contributed to the dislocation of the upper soil horizons on the slope. "Stone lines" also exist in the semi-arid environment (Yair and De Ploey 1979) where any soil moving activity of living organisms could hardly be suspected.

Interesting but not very copious material has been obtained from the measurements in Puerto Rico (Lewis 1974), Rwanda (Moeyersons 1981) as well as Kuala Lumpur (Eyles and Ho 1970) and northern Australia (Williams 1968). The first two experiments show movement curves differing from "standard" curves in that they direct attention to the role played by viscous mass flow in the slow mass movement of the tropics. The presence of soils with large clay particle content is a major factor. The double maximum of the curve noted by Lewis (1974) may be accounted for by the shearing plane which separates the surficial soil layer with its strongly knit root system from the layer below. The seasonal creep of the rainy tropics tends to pass downwards into continuous creep, i.e. rheological creep, which makes the whole phenomenon resemble a slow landslide.

The movement rate of tropical creep is lower than in cold climates but probably higher than the rate of slow mass movements in temperate and arid zones. It is therefore likely that the surface movement ranges between 0.5 and 1.0 cm year^{-1}, its volumetric value being about 20 cm^3 cm^{-1} year^{-1}.

8. RESULTS

Slow mass movement is a form of transporting slope material. It can be classified according to: the dynamics of the process, the structure of the movement (particles, layers) and the relation to the morphological surface (shallow or deep movement). Seasonal creep is first of all a function of climate whereas continuous or rheological creep is a function of geological structure.

The mechanism of slow mass movement is studied experimentally by means of precise instruments with only a small measurement error. Such investigations can be carried out effectively in short-term measurement series (hourly, daily, monthly observations).

The morphological meaning of slow mass movement can be evaluated in long-term measurements (seasons, years) and by using methods that do not disturb the natural equilibrium of the soil. It is only after a measurement period of at least 5 years that we can speak of any morphological effects. Slow mass movement is of morphological significance only in the upper parts of slopes. The morphological effects of slow mass movement seem to be inversely proportional to the lengths of the slopes, which makes this factor differ from the erosive action of water, i.e. slope wash and gullying.

Being the most important form of slow soil movements, seasonal creep is a pulsating movement dependent above all on soil temperature and humidity. This is why this process can be viewed from the standpoint of climatic-geomorphological regional divisions of the earth, provided that a sufficient number of data have been collected by a uniform measurement method eliminating local influences. The existing data has revealed differences in the functioning of this process in the latitudinal and altitudinal zonations (Table 6). These figures should be regarded as values which have not yet been sufficiently tested, but which should be a stimulus for further geomorphological field experiments.

TABLE 6. Summary rates of movement by slow mass movement in climatic zones

Location	Rate mm yr^{-1}	Volumetric $cm^3 cm^{-1}$ yr^{-1}
High latitude zone	42.5	102.8
High altitude zone	17.4	29.1
Temperate zone	2.6	3.1
Arid (semi-arid)	1–2 (?)	?
Tropical zone	7.0	19.2

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I would like to thank all my co-workers whose names are included in Table 1 and who sent me the results of their measurements referring to slow mass movement of the soil. At the same time I have to apologize for the significant delay in their publication which was not due to my neglect.

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ALTITUDINAL ALTERATION OF THE MORPHOLOGICAL SYSTEMS IN THE RIO CHECRAS BASIN OF THE PERUVIAN ANDES

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ABSTRACT. The vertical alteration of geomorphic systems results from Quaternary evolution. The huge stabilized landslides are the main feature of the slope morphology. They were produced in a periglacial zone of the older Rio Blanco glaciation. The present glacial and cryonival systems were affected by the Punrun glaciation phase and postglacial climatic conditions. The valley bottoms and the lowermost parts of the slopes were transformed by mass movements and mud/debris flows in the postglacial period.

INTRODUCTION

This article is the effect of the author's participation in the archeological research project in the Rio Checras drainage basin in 1985. It contains preliminary results of the field studies on the altitudinal alteration of geomorphological features and processes as a reflection of the Quaternary events. Some conclusions were correlated with those obtained by Wright (1983) in the Junin Plain located east of the Checras basin. Topographical maps, 1:100000 in scale, their enlargement to 1:50000 scale, aerial photographs (scale ca 1:21000), and geological maps (scale 1:100000) were used in the work. Although it was possible to perform research in long-distanced parts of the area, some regions could be studied only by means of aerial photo analyzes. Apart from the Rio Checras drainage basin, with special attention being paid to the Cayash valley, a reach of the Huaura valley near Churin, was also investigated (Fig. 1). Travertine samples were dated by Prof. M.F. Pazdur at the Radiocarbon Laboratory in Gliwice.

AREA OF FIELD STUDY

The Rio Checras is a big tributary of the Huaura River, flowing down from the west-facing slopes of the Western Cordillera to the Pacific Ocean. It is situated north of Lima, at about 10°S (Fig. 1). The drainage basin, 825 km^2 in size, lies at the altitude of 2150–5300 m a.s.l., thus within five vertical belts: yunga (2150–2500 m), quechua (2500–3500 m), suni (3500–4000 m), puna (4000–4900 m), janca (4900–5300 m) (Pulgar Vidal 1988). Janca is related to the nival and puna to the subnival geoecological belts. There is no forest zone and different varieties of the mountain steppe dominate there. The Pacific-facing slopes of the Andes are arid and the maximum precipitation, of about 600–800 mm, is due to the quechua belt. The research area is built of stratified rocks of the Cretaceous age dipping almost vertically. Amongst the geological complexes of differentiated resistance, the limestone and

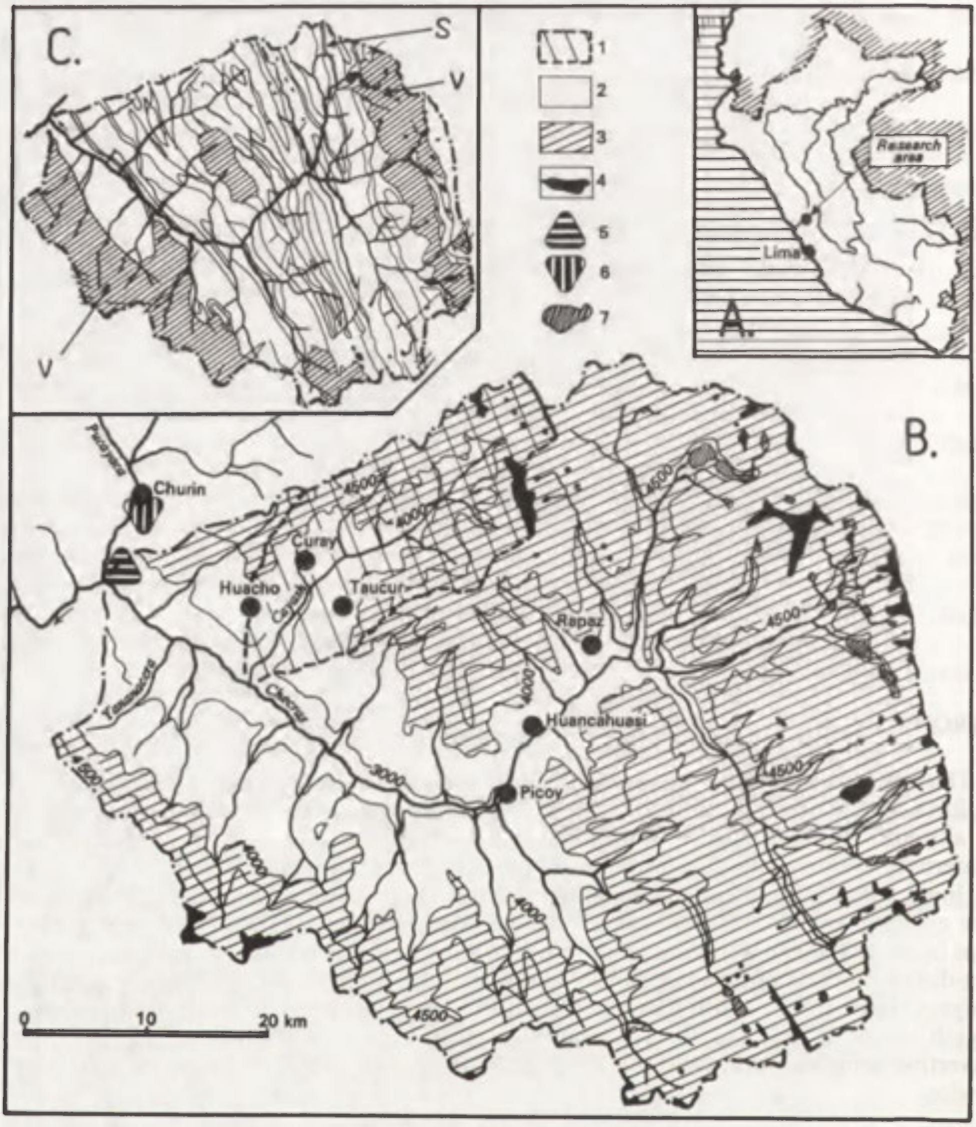

Fig. 1. Major elements of research area: A. Situation sketch, B. Map of the Checras drainage basin: 1 — Cayash drainage basin, 2 — area located below 4000 m, 3 — area situated above 4000 m (puna), 4 — area exceeding 5000 m (janca), 5 — site with older travertine (Checras junction), 6 — site with younger travertine (Churin), 7 — high mountain lakes. C. Geological sketch: S — bedded sedimentary rocks of Cretaceous age, V — volcanic rocks of the same age

quarzite beds are the hardest. There are also volcanic rocks in the south-western and north-eastern parts of the drainage basin (Fig. 1). The relief of the Peruvian Andes was developed in the Tertiary and Quaternary periods in conditions of changing climate and tectonic activity (Dollfus 1965; Megard 1978).

PUNA SYSTEM

The Miocene surface of planation called "the puna level" is the main feature of the high parts of the Andes. It was uplifted and deformed during Pliocene and Quaternary by tectonic movements (Megard 1978). In the Checras basin the puna level is wide-spread at the altitude of 4400–4800 m. More than 50% of the area is situated above 4400 m, but only ca 3% above 5000 m (Fig. 2). It is the reason why nival vertical belt occupies a rather limited space. Only a few ridges elevating more than 5000 m – Yaruhuayno, Parahuayna, San Camilo, Perurayoc, Chururuyo and Cordillera Callejon – are covered with small ice caps (Fig. 1). There are common traces of glacial transformation. The lowermost features of glacial erosion and

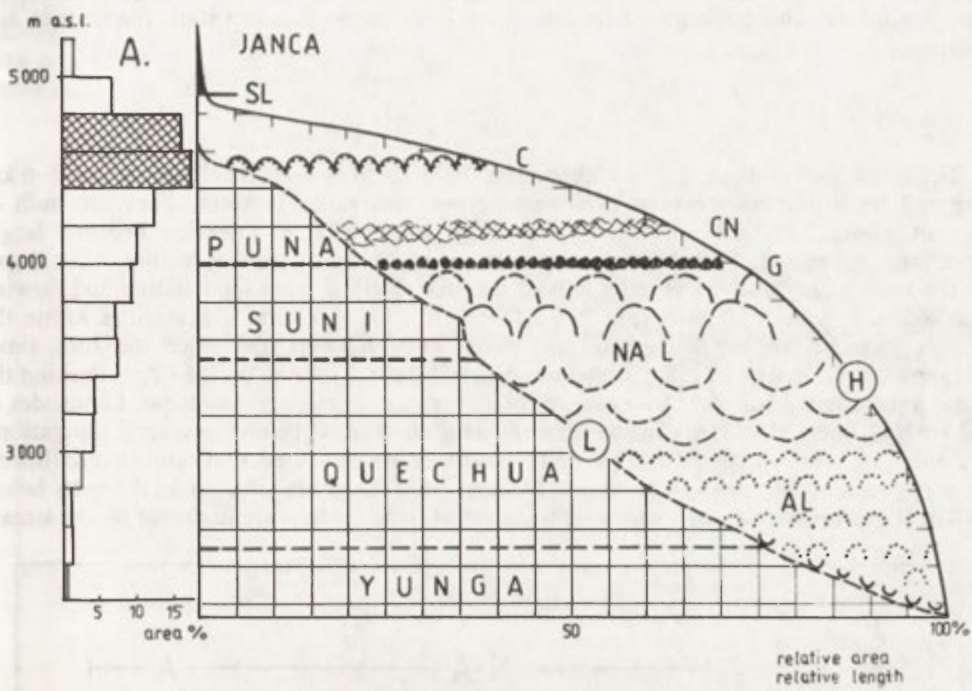

Fig. 2. Vertical alteration of geomorphological systems in the Checras basin: H – relative hypsographic curve, L – relative long-profile of the Rio Checras, SL – present-day snowline, C – lower limit of glacial cirques, CN – lower limit of cryogenic features, G – lowermost limit of Pleistocene glaciations, NA L – non-active landslides, AL – active landslides, A – distribution of catchment area in relation to altitude a.s.l.

deposition were found at the height of ca 4000 m. This is the same height as observed by Dollfus (1965) in the Central Peruvian Andes, east of Lima. The lowermost position of glacial cirques was read from topographical maps and aerial photographs. It ranges from 4500 to 4700 m. The mean altitude is close to 4500 m but there is a difference between the SW and NE faced slopes, 4500 and 4625 m respectively. The above mentioned data confirm Wright's (1983) observations and support his conclusion about the ca 300 m Pleistocene lowering of the snowline in the Western Cordillera. About 42% of the Checras catchment basin lies above 4500 m and moreover 35% lies above 4600 m a.s.l. In the headwaters of the Cayash valley there are systems of 4–6 recessional morain ridges. They are very well-preserved and comparable with the morains in the Junin Plain related to the Late Glacial and Holocene deglaciation phases (Wright 1983). Thus the puna level conditioned an extensive Pleistocene glaciation of most part of the Checras basin (Fig. 1). Wide spread ice covers, especially at the phase of deglaciation, controlled the hydrological and geomorphic processes in the lower belts. The large quantity of meltwaters was an important agent of the slope evolution there. Within the puna belt, covers of different origin — gravitational, glacial, glaciofluvial, fluvial, limnic, organic — serve as important reservoirs supplying water to the lower located suni and quechua zones.

Despite glacially affected topography, the puna level is closely related to the geological structures and the differentiated resistance of the sedimentary rocks. Thus the long profiles of the glacial valleys are stepped. Between the numerous steps there are basins with lakes and pit-bogs. The narrow and sharp ridges, 500–1000 m high, rise above the base. The peaks are higher by 200–400 m the present snowline being at ca 4900 m. At the foot of the rock walls, there are taluses and huge rock falls spread for from the wall base within the cirques and valleys.

SLOPE SYSTEM

The puna level is dissected by main valleys 1000–2000 m deep. Their slopes are 2–6 km long and become more steep or even vertical near the valley bottoms. They are built of different covers, but colluvial ones are especially common. Their thickness becomes larger downslope and values of 40–50 m are rather frequent. The most striking feature of the slopes are the huge niches and stabilized colluvial tongues resulting from land sliding and flowing processes of very high intensity (Fig. 2–3). There are a lot of natural irregularities within the tongues adapted for use as agricultural plots. They have existed since the Inca times (Krzanowska, Krzanowski 1987). Pre-columbian villages are located on the ridges limiting the niches and colluvial tongues. Everywhere solid rocks are the village basement. Landslides of 1–3 km long and 100–500 m wide which were mentioned above belong to several generations. The oldest ones are stabilized and the colluvial covers are cemented. Old cemented sediments are very common in the Central Andes (Dollfus 1965). They are situated in the belts below 4000 m. Their lower sections, as for example in the Cayash valley, are undercut by the stream

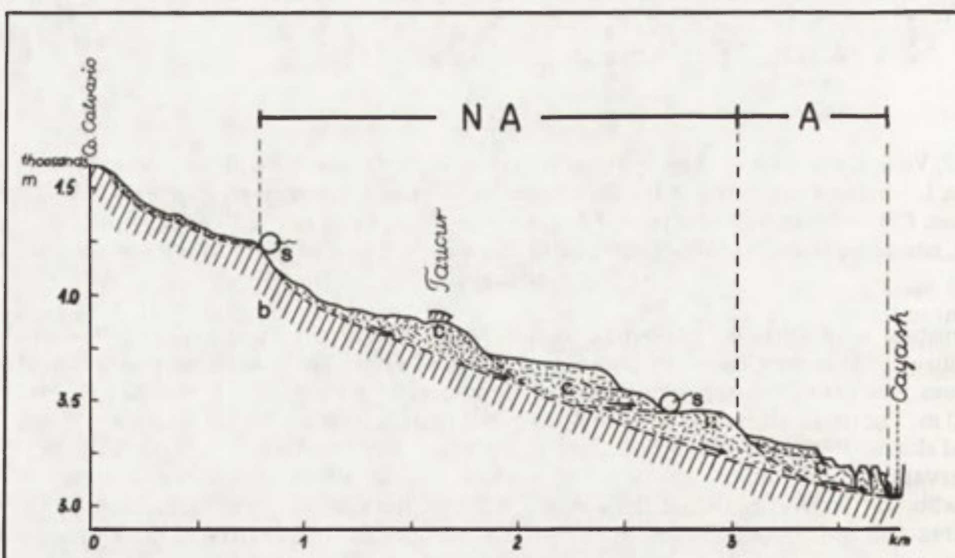

Fig. 3. Landslide in the Cayash valley: NA — zone of non-active landslide, A — zone of active landslide, c — colluvial covers, s — spring, b — bedrock

(Fig. 3). Erosional impulse wanders upslope. At the 2150–3000 m belt old landslides are rejuvenated. Near the Checras junction a lacustrine series ca 200 m thick is deposited on the colluvial material. This confirms that huge landslides dammed the valleys in the past. There are also quite young and active landslides like that in the Qebrada Yanaraccra. A number of landslides damming the valleys in historical times was observed in various sectors of Rio Checras (below Picoy, above Huancahuasi), or in Qebrada Pucayaca near Churin (Fig. 1). It seems, that mass movements were the most important processes transforming the slopes during the whole Quaternary period but their intensity was not the same just as climate changes occurred. Another striking feature of the slope system is that there are shallow and wide dry valleys produced mainly by solifluction. In the middle and lower sections they are intersected by gullies and V-shaped valleys. The exposures along the road from Churin to Huacho and from Huacho to Curay within the Cayash drainage basin, provide a lot of information on slope covers and fossil valleys (Fig. 4). The are at least 2–3 fossil valleys filled with a number of covers of different origin. Amongst them – clayey-silty covers with single rocks and debris inclusions and debris covers produced by violent flows. Alternate coarse and

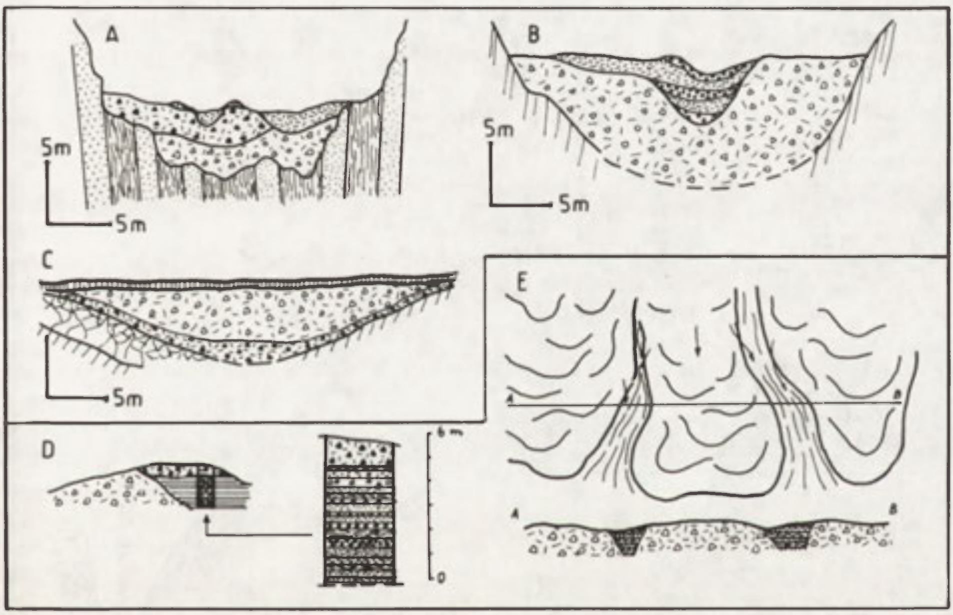

Fig. 4. Examples of slope cross-sections between Churin, Huacho and Curay: A, B, C – examples of fossil slope valleys with generations of debris covers, D – example of bedded slope sediments above Churin and their situation side by side to landslide deposits, E – sketch explaining origin of layered sediments

finer sediments evidence the changing conditions of erosion and deposition. In general, the layered and non-layered slope sediments are located side by side and they form the complicated pattern of the slope system. Within the fossil valleys it is possible to determine 1–3 covers produced by solifluction and the same number of debris flow fills. But above Churin, was found a 5 m high exposure of layered deposits. They were composed of 18 beds of fine and coarse material alternatively (Fig. 4D). A general conclusion that can be drawn is that slope sediments were produced by more cold and wet (solifluction, landslides) or warm and dry (mud- or debris flows – huaycos, wash) conditions related to a periglacial or arid environment respectively.

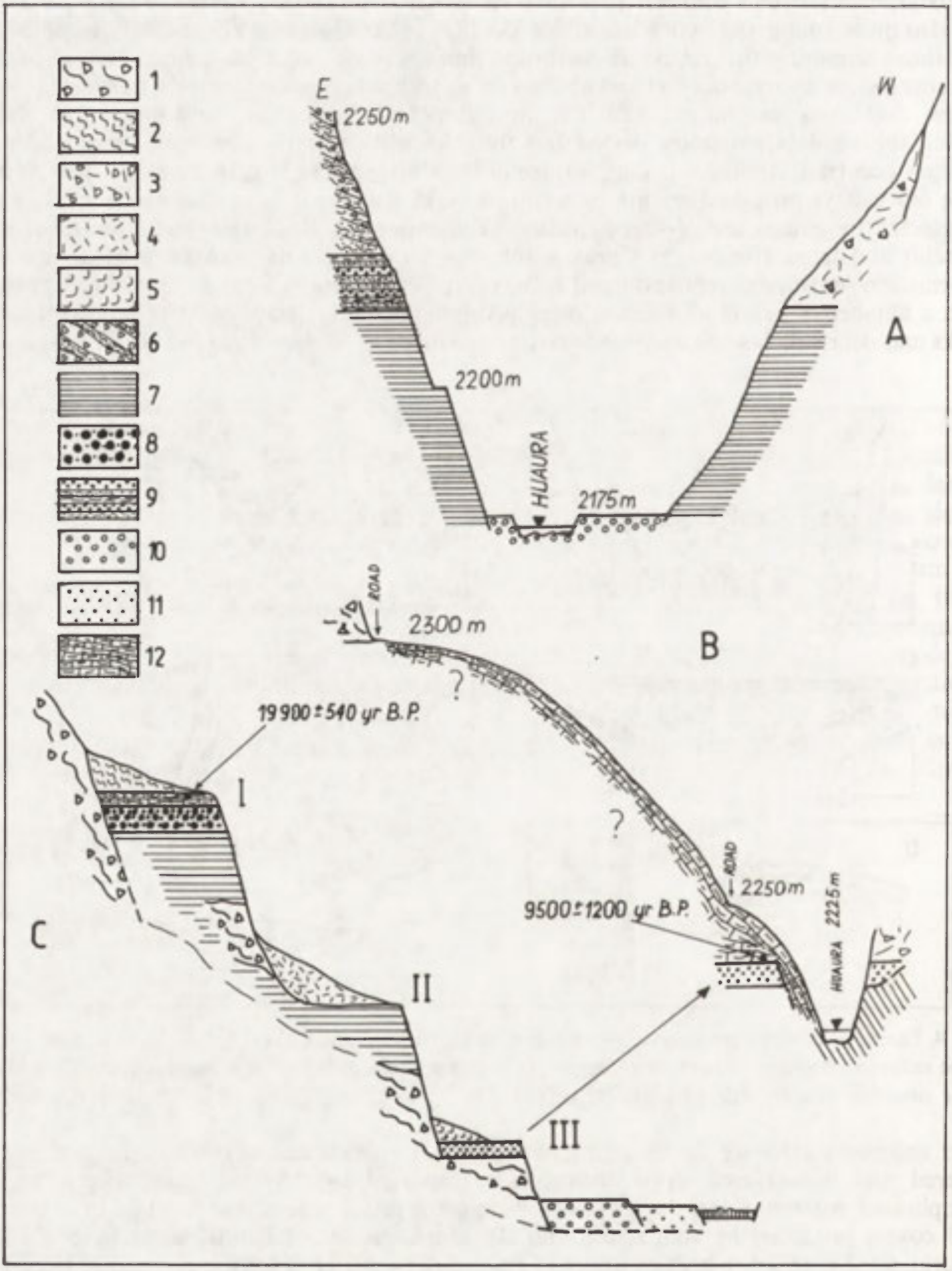

Fig. 5. Terrace systems in the Huaura-Checras valleys. Cross-sections: A — above the Huaura junction, B — in Churin, C — synthetic, I, II, III — number of levels, 1 — landslidecover, 2–5 colluvial covers of different age, 6 — layered slope covers, 7 — lacustrine sediments, 8–11 river sediments of different age, 12 — travertine

VALLEY FLOOR SYSTEM

The bottoms of the main valleys are more wide and more narrow alternatively. Both sides of them are steep or even vertical and built of solid rocks or cemented and very hard slope deposits. A large amount of loose material is deposited at the foot of the cliffs. Tributary fans extend into the central part of the Huaura and Checras valley bottoms. Braiding rivers are very active during heavy rains and then they undercut slopes at whole distance. It is the reason why the slopes situated below 3000 m are episodically remodelled by landslides and debris flows termed here as *huaycos*. Thus more narrow sections of a river valley are dammed by slope and fluvial material. It was found that in the past very high damming had occurred near the Checras and Huaura junction. Limnic, silty-sandy, well-beded sediments have been preserved there. The surprisingly large thickness of that series, ca 200 m, lying on the colluvial material could have resulted from exceptionally high damming. Terrace systems corresponding to the phases of dissection of the sediments are preserved above the junction in the Huaura and Checras valley. At the top of limnic series a 10 m thick layer of gravel bed-material and a 1 m thick travertine layer were deposited (Fig. 5). Thus it became possible to date some sediments and erosion episodes by means of the radiocarbon method. Terrace levels are numbered from 1 to 6:

1st level, about 2400 m a.s.l., ca 250 m above the Checras channel, lacustrine sediments with travertine on the top, radiocarbon date 19 900 $\pm$ 540 yr BP (Gd 4201),

2nd level, about 2200 m a.s.l., ca 50 m above the Checras channel, cut in the lacustrine sediments with river gravels on the top,

3rd level, about 10–30 m above the channels, river gravels with the travertine on the top, radiocarbon date 9500 $\pm$ 1200 yr BP (Gd 4216),

4th–6th levels, about 0.5–5.0 m above the channels, recent river deposits (Fig. 5).

The 1st level is built of yellow silts and sands horizontally bedded. In the lower part of the series there is ca 15 m thick clayey-debris cover of slope origin. The level is seriously dissected and eroded. Nevertheless, its residuals appear in few places in the Huaura and Checras valleys and lacustrine sediments are included to the younger morphological and sedimentation units. The travertine top of the level is upbuilt with a 10–30 m thick colluvial cover. The 2nd and 3rd levels are cut within the lacustrine series and upbuilt with fluvial bed material. The travertines of the 3rd level form a very extensive, oblique cover continuing from the height of 2300 m a.s.l. to the Huaura channel at the height of ca 2225 m. Probably also the 2nd level is hidden under the cover. It is easy to observe the process of travertine precipitation at the top of the thick cover and in a few places in the vicinity of Churin. If the sample of travertine is from the real floor of the cover, then the date shows the beginning of its formation and the beginning of a warm and dryer period – Holocene. And then the gravels lying below the travertine are related to the last cold phase of the Late Glacial. The lowest accumulation terraces are situated in the valley bottoms and are the result of the most violent flood events in the Holocene period.

RESULTS OF ANALYSIS AND CONCLUSIONS

The two above cited dates are the basis for recommending the different field observations. It is clear that the damming of the Huaura-Checras valley by huge landslides and the sedimentation of the lacustrine sediments and the 15 m thick river cover took place before ca 20 000 yr BP. Those events can be correlated with the older glaciation phase called Rio Blanco by Wright (1983) and dated by him as earlier than 42 000 yr BP. The huge stabilized landslides, so clearly-visible in the belt below 4000 m could be related to that cold, unknown in detail, period. Also the oldest fossil valleys and cemented periglacial covers within the slopes probably derive from that phase.

The date 19 900 ± 540 yr BP of the top of travertine layer, which is upbuilt with 10–30 m thick colluvial deposits (cold?), means the end of warmer and dryer phase previous to the younger phase of glaciation named Punrun (Writht 1983). There is no very precised agreement on the probable date of the Punrun phase beginning (23 980 ± 320 yr BP). It is possible that the Punrun glaciation phase in the Junin Plain, situated about 2000 m higher than the outlet of the Checras valley, influenced the far-distanced lower areas later.

The deep dissection of the 1st level, cut out the 2nd level, the incison of the last one and finally forming the 3rd level, should be connected with the Punrun glaciation phase placed between ca 24 000 and ca 12 000 yr BP (Wright 1983). The two levels mentioned above are built of the coarse river material and are the basis to differentiate a number of more or less cold climatic subphases (Cardich 1981). The system of 2–6 well-preserved recessional morain ridges in the puna belt support this idea. Probably also most of the dry valleys on the slopes were developed under the periglacial conditions of the Punrun phase.

The date of the travertine base – 9500 ± 1200 yr BP determines the end of the last cold period of the Late Glacial and the beginning of the warmer and dryer Holocene. The accumulation of travertine has been continued from that time. The still lasting deglaciation, rejuvenation of the lowermost sectors of old landslides and the episodic activity of mud/debris flows are the most important features of the geomorphological dynamics in the Holocene period. The ice caps, existing in the janca belt, have now a limited geomorphic significance. Thus the altitudinal alteration of the main geomorphological features was shaped in the Pleistocene during cold and wet climatic phases separated by warm and dry ones. Torrential flows of high activity which occur now from time to time produce significant morphological changes within the valley floors.

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DEGLACIATION OF THE NORTHERN FOOTHILLS IN THE EAST KUNLUN MTS.

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ABSTRACT. Results of field investigations at the northern foot of East Kunlun Mts. as well as the known so far literature data allowed a reconstruction to be made of the history and type of glaciation and deglaciation of this area during the Late Pleistocene and Early Holocene. The former glaciers had developed as piedmont glaciers debouching into the high-glacial Qaidam Lake. The largest glaciers at their maximum stage reached about 45 km in length and at least 20 km in width, producing great quantities of glacial and glaciofluvial material which have been fixed in thick sequences within the area of the former Qaidam Lake. At the decline of the Pleistocene, the piedmont glaciers underwent a large-scale stagnation probably brought about by a rapid decrease in moist air flux. In consequence, kame terraces abutting on the mountain ridges were formed. They are interstratified and covered with angular debris of supraglacial derivation, protecting them against post-depositional erosion and denudation. The latter feature contributed to the fact, that the kame terraces have remained in almost unshaped form up to the present. The thermoluminescence dating of kame sediments yielded a date which suggests that local deglaciation of the investigated foot of the East Kunlun Mts. took place early in the Holocene, around 8000 years BP.

INTRODUCTION

A number of papers has recently been published on the present and former Quaternary glaciations of the Qinghai-Xizang (Tibet) Plateau. Worthy of note are particularly the papers presenting the results of the joint Sino-German and Sino-Japanese expeditions edited by Hövermann and Wang (1987), Kuhle and Wang (1988) and Higuchi and Xie (1989). However, our knowledge of the nature and course of glaciation and deglaciation processes in these areas is still incomplete and on some points inadequate. In recent investigations, the glaciologists and glacial geomorphologists have dealt mainly with active ice masses, concentrating their field works on modern glaciers and the high-mountainous portions of the former ice covers. They describe mainly such types of forms or sediments which are clearly connected with flowing ice, like glacial cirques and troughs, terminal moraines, polished and striated rocks or ground moraine. These forms and sediments have long been researched as far as the glaciation of the Tibet Plateau and surrounding mountains are concerned (cf. Sobolewski 1919; Hedin 1922; Trinkler 1930, von Wissmann 1959; Shi and Xie 1964; Kuhle 1987; Zhang et al. 1989). However, in these Central-Asian areas, under the severe continental conditions of the Late Pleistocene, as yet insufficiently known, features of ice-contact morphology associated with the melting of stagnant ice had formed, too. Such a type of landforms, represented by kame

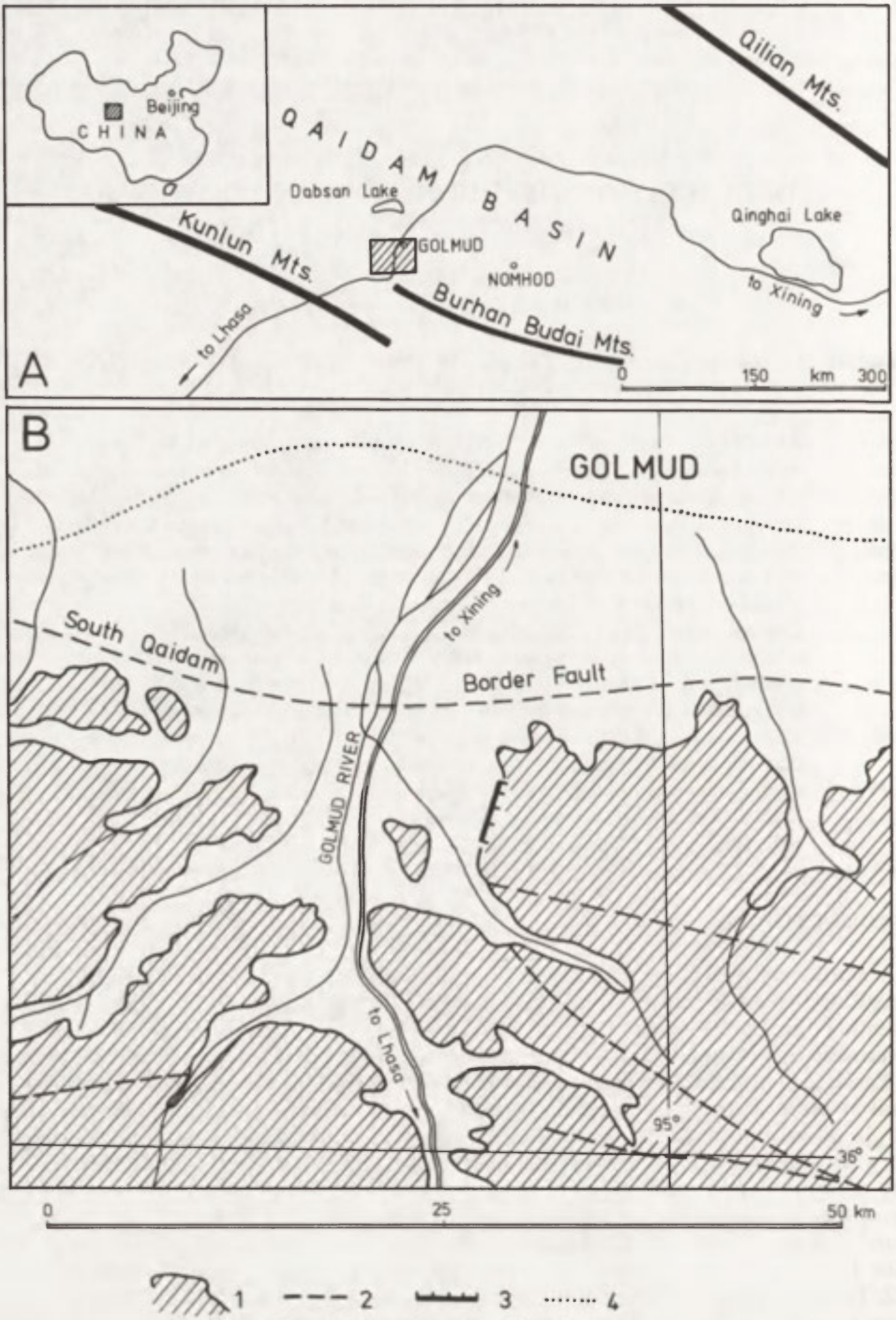

Fig. 1. Schematic map of study area

1 — rock surfaces, 2 — main fault lines (from *Geological Map...*, 1987), 3 — kame terraces, 4 — ancient shoreline, denoting southern border of high-glacial Qaidam Lake (from Halimov and Fezer. 1986)

terraces of considerable size, was encountered by the author at the northern foot of the East Kunlun Mts. in 1989*. Inasmuch as kame terraces are indicative of stagnant ice and to date no such features have been reported from Tibet or its surroundings, it appears to be interesting to present them in more detail and to discuss their implications for the nature and course of glaciation of the Central-Asian region investigated.

LOCATION AND PHYSIOGRAPHY OF THE STUDY AREA

The study area is located at the foot of the Burhan Budai Mts. (Fig. 1), being the outermost northeastern range of the East Kunlun mountain system which has in its main ridges elevations of over 6000 m a.s.l. (the highest summit reaches 6224 m a.s.l.). A characteristic feature of these mountains, as similary the West Kunlun Mts. (Sobolevski 1919; Zhang et al. 1989), is the steepness of their northern side, unlike the southern side, adjacent to the vast Tibet Plateau, which is relatively flat and wide. Therefore, characteristic glacier types on the northern side are valley and dendritic glaciers, whereas on the southern side, the glaciers tend to develop as ice caps and outlet valley glaciers spreading out on the adjacent plateau.

Photo 1. General view on the lateral branch of the Burhan Budai Mts., as seen from the south toward the Qaidam Basin. In the first plan — kame terraces reshaped by wind-blown processes, on the opposite side of the mountain branch, an irregular patch of eolized glaciofluvial sands, is visible

*During the summer season of 1989, the author had the possibility of participating in the Chinese expedition leading from Lanzhou to Lhasa, for which his sincere thanks are extended to Professor Xie Zichu, the head of the Lanzhou Institute of Glaciology and Geocryology, Academia Sinica.

The present equilibrium line on the northern slope of the mountains, according to Shi et al. (1988), is situated at 5200 m a.s.l., and the modern glaciers descend here as far as 4300 m a.s.l. (Kuhle 1987). The lower limit of the permafrost zone is situated at elevations ranging from 3300 to 3500 m a.s.l. (Shi et al. 1988). The Golmud River, being the main stream draining this area northwards, flows to the relic Dabsan Lake in the centre of the closed Qaidam Basin. The Golmud-Lhasa highway runs through the study area in the southwest direction, along the Golmud River (see Fig. 1).

The main geomorphological feature of the terrain investigated is a broad depositional piedmont plain (Photo 1), resulting from coalescing alluvial fans and opening towards the north between lateral branches of the Kunlun mountain system. This plain is traversed here by a prominent foothill fault – the South Qaidam Border Fault – running E-W and denoting a tectonically active line at the northern foothills of the East Kunlun Mts. The piedmont plain extends across it and spreads out before the front of the mountains, debouching finally into the former high-glacial Qaidam Lake. The latter developed at an elevation of approximately 2900 m a.s.l. and is marked today by a belt of dunes along its southern shore (Halimov and Fezer 1986).

The investigated fragment of the piedmont plain, named here the Golmud valley, is bordered from the east by a mountain ridge, displaying steep slopes, inclined frequently at an angle of 35–40°, whereas the summits of the ridge, rising 350–450 m above the plain, have mostly rounded shapes, lowered by denudation. The rocks exposed in this area consist of batholic granite of Permian-Triassic and Devonian age, additionally, of limestones and arenite interspersed with shaly beds of Devonian age (*Geological Map*... 1987). The majority of the land surface (piedmont plain and lower parts of the hillslopes) consist of clastics, chiefly of alluvial fan sand and gravel.

KAME TERRACES

The kame terraces are already visible from afar due to the light colour of their sandy material, abutting on the grey lateral ridge of the Burhan Budai Mts., approximately 30 km south of Golmud city (see Fig. 1). They occur in a series of four steps sloping downstream and overlying one another in a stairway fashion at roughly the following elevations: 10–15, 30–35, 50–55 and 70–80 m above the adjacent piedmont plain (Photo 2). The descending, actual ice-contact slopes of the terraces vary in their inclination. The higher and lower terraces have slope angles of between 8° and 28°, whereas the middle terrace, situated at an elevation of 50–55 m, has a slope inclined between 25° and 30°.

The largest and, simultaneously, the best-developed kame terrace is the middle one, bounded by a spectacular descending slope with a sharply expressed straight edge. It extends over a distance of more than 800 m, is up to 300 m wide and possesses a flat-topped surface with a relatively gently inclined longitudinal profile compared to the higher and lower terraces, which at their outer parts have been more denudated and also somewhat undulated by wind erosion.

The well-preserved form of the terraces seems to be explained by the undisturbed geological structure and the presence of angular debris of supraglacial derivation (Photo 3). The debris had been produced by rock fracturing above the glacier which had fallen from the flanking rock-slopes onto the glacier surface. Subsequently, being transported in a high position in the glacier or on surface of the glacier, it was dumped and slid down onto the sides of the kame terraces. The loose, angular rocky material, composed chiefly of granite akin in particle size from those of silt and clay to heavy fragments larger than 20 cm in diameter, have intertongued with the glaciofluvial sediments concurrently being deposited, thereby constructing a strong, "armoured" slope, resistant to any post-depositional erosion.

The amount of supraglacial material incorporated into the kame terraces was controlled by the volume of debris transported supraglacially and englacially and its distribution within

Photo 2. General view on the kame terraces, seen from the south toward the Kunlun Mts. The distinct ice-contact face of the middle terrace, rising 50–55 m above the adjacent piedmont plain, is evident

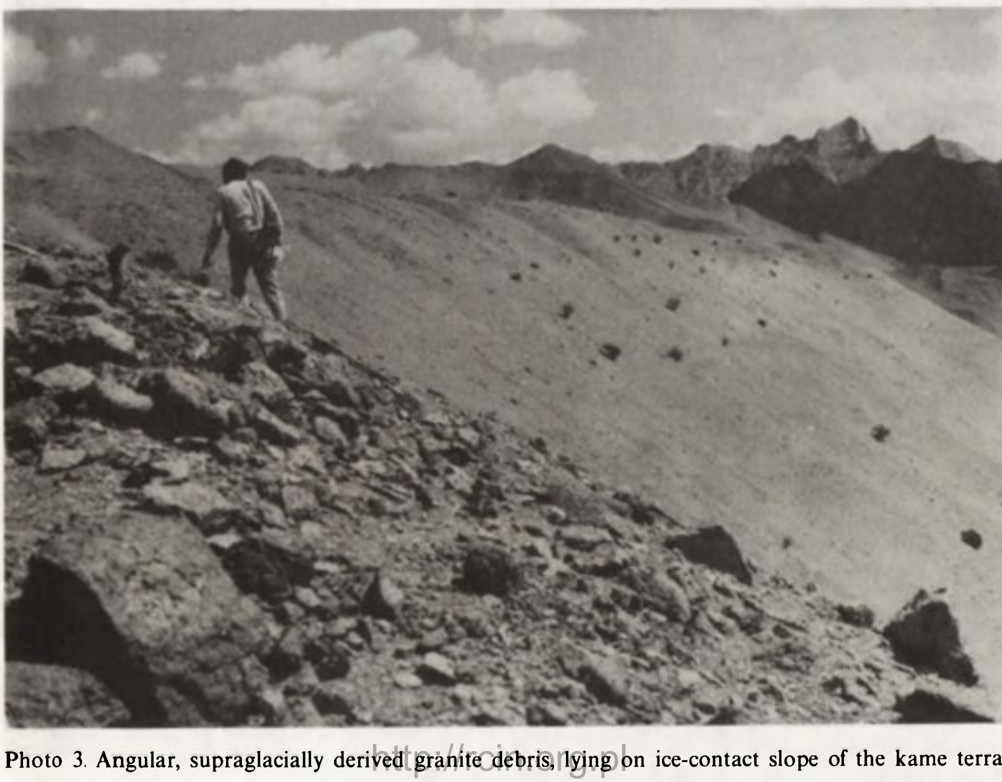

Photo 3. Angular, supraglacially derived granite debris, lying on ice-contact slope of the kame terrace

the glacier. Where a concentration of debris occurred, as for example, along the former ice gullies or in the ice crevasses, there was a tendency for thick localized debris accumulations at the lateral glacier margin. It is likely that such an enhanced debris supply had accompanied the formation of the kame terraces, particularly at the end of the formation of the middle terrace, producing an exceptionally thick "armoured" slope. Also, for this reason and for the lack of dead ice blocks buried by kame sediments, the kame terraces did not experience the deformations caused by the collapsing of the sediments after the disappearance of the ice that had supported them. Therefore, they have preserved their original constructional forms up to the present, despite the destructive high erosional energy, which characterizes the tectonically active desert environment in which they occur.

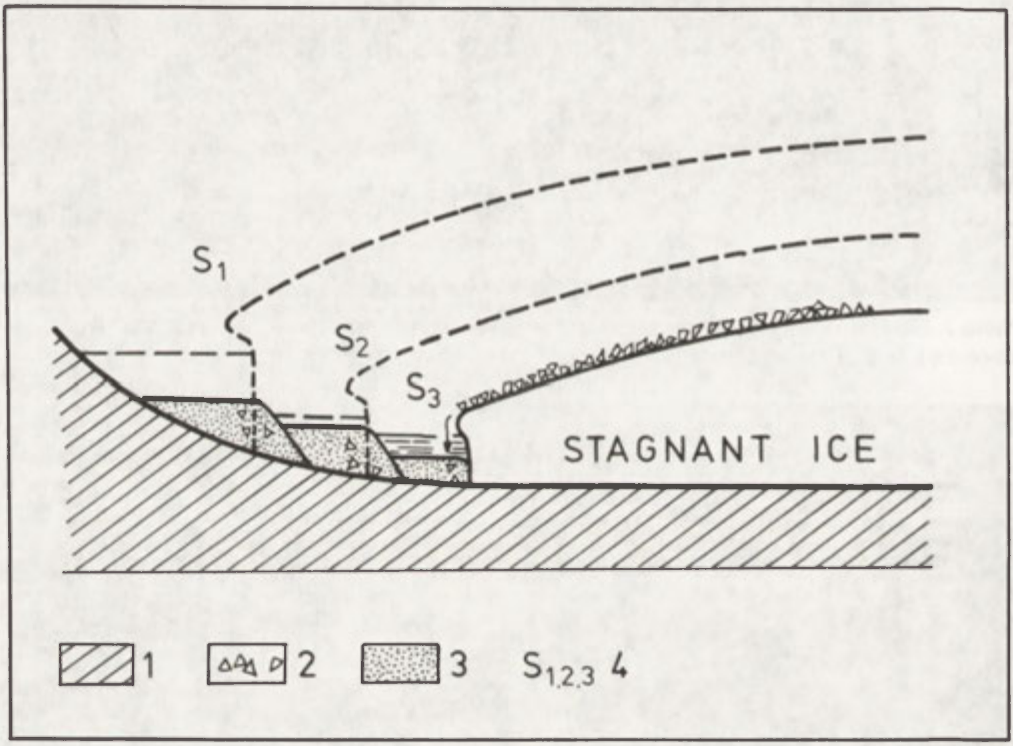

Fig. 2. Diagram showing successive stages of the formation of kame terraces
1 — rockslope, 2 — angular supraglacial debris, 3 — glaciofluvial sediments, 4 — successive stages of the formation of kame terraces

The stepped pattern of the kame terraces is believed to denote successive stages in the process of the thinning and undercutting of the last glacier, as illustrated in Fig. 2. This explanation implies the successive lowering of the water level in the ice-lateral stream or temporary lake and the concurrent increase of the terrace width as the supporting glacier is melting. These processes proceeded also in the neighbouring foothill area of the Burhan Budai Mts., occupied by a glacier abutting on the opposite side of the mountain ridge and divided further downglacier by a low col from the Golmud piedmont glacier. Between these glaciers, probably in a later stage of deglaciation, there existed an interrelated meltwater system which affected the rate of ice melting and, in consequence, the distribution and formation of ice-contact features.

However, on the opposite side, no comparable kame terraces developed, instead, irregular, more or less discrete patches of sandy sediments are to be seen (Photo 1), which might have originally been deposited in the ice-lateral stream as kame sediments and, later on, redeposited by wind. The reason, which could here explain the lack of well-formed kame terraces, may be the insufficiency or the lack of coarse heavy debris of supraglacial derivation, unfavourable slope inclination or the deposition of sediments mainly on the ice that led to the collapse of the sediments after the ice had melted away. Similar landforms in the Sudetes Mts. produced in association with various stages of deglaciation are described by A. Jahn (1963).

A sample of sediments from the middle kame terrace was submitted for thermoluminescence (TL) dating in the TL Laboratory at the Gdańsk University. The result obtained is 6700 $\pm$ 1200 yr BP. This result, even if we take into account the broad measurement error, which is assessed to be 17%, seems to be too young when compared to the timing of final deglaciation in the north-eastern part of the Qinghai-Xizang Plateau which – as interpreted from a series of radiocarbon dates (Kuhle 1988) – occurred between 9400 and 8600 yr BP.

CONCLUSIONS

The described ice-contact features substantiate the drawing of a general conclusion about the occurrence in the area investigated of an assemblage of sediments and landforms, defined by Boulton and Eyles (1979) as glaciated valley sediment and landform system. The principal feature, closely associated with this system and differing this from those produced by a non-valley depositional system, is the supraglacially-derived debris (resulting from rock breakage on the sides of valleys or nunatacks) that had not undergone a phase of tractional transport at the glacier bed. Fortunately, this significant element of the glaciated valley system, producing supraglacial material preserved in the kame terraces described is shown to have occurred here. Other constituent elements of the system, such as terminal or lateral moraines, have been erased by subsequent Holocene fluvial erosion.

The next important conclusion, which might be inferred from the occurrence of the kame terraces themselves, concerns the transformation of the former active valley glacier into a stagnant ice mass. Most probably, this phenomenon occurred within a relatively short time span as the consequence of a rapid decrease in moist air flux. Leaving aside the atmospheric circulation changes as a separate problem, demanding special studies, here questions arise, pertaining to the general conditions of the glacier before this phenomenon, a reconstruction of its morphology and dynamics. It is likely that the valley glaciers which had debouched from steep mountain valleys into the Qaidam Basin, broadened out and formed large piedmont glaciers of the Malaspina Glacier type in Alaska. In the Golmud valley, such a type of glacier during its maximum extent might have reached 45–50 km in length and at least 20 km in width, counting from the mountain front to the shoreline of the former Qaidam Lake. This glacier as well as a row of similar piedmont glaciers at the foot of the Kunlun Mts. presumably coalesced with one another. They were drained by large subglacial and englacial streams which carried great volumes of bed loads and suspended loads, depositing them finally north of the South-Qaidam Border Fault, in the area of relative subsidence, corresponding to the Qaidam Lake. This is indicated stratigraphically in the deep bore-hole at Nomhon (see Fig. 1) by thick glaciofluvial sequences covered and interlocked with lacustrine sediments (Hövermann 1987; Kuhle 1987).

The Qaidam Lake and the glacio-hydrological system connected with it – in the light of the available radiocarbon dates – existed approximately between 35000 and 23000 yr BP. The kame terraces described should already be related to the close of the last glacial period in the foothill area of the East Kunlun Mts. The buried dead ice supporting the kame terraces remained here as long as the depth of freezing in winter exceeded the depth of thawing in summer. Under suitable climatic and ground conditions, the dead ice masses might have persisted for a considerable span of time, possibly for hundreds of years. The TL date obtained for the sediments in the kame terraces suggests that they were deposited late in the early

Holocene, that is 7900 yr BP (if the date is corrected in plus according to the maximum measurement error). Hence, the estimation of the age for the local deglaciation, which could have occurred a little earlier, is around 8000 yr BP. However, this dating figure for the deglaciation chronology should be regarded with caution, not only because it is younger than the related radiocarbon dates, as mentioned earlier, but also because it is based on a single, stratigraphically isolated sediment. Further corroborating dating is needed to substantiate this interpretation.

Information on the glaciation pattern of the East Kunlun Mts. at the close of the Central-Asian Ice Age is still vague. Based on the presented and discussed here evidence, it is more likely that the glaciers were then largely controlled by the underlying topography. In other words, they drained several icefields situated higher in the heavily dissected mountains, resembling in this respect, for example, the present-day glaciation of the St. Elias Mountains. Nevertheless, it cannot be excluded that the preceding piedmont glaciers might have had the form of outflow glaciers radiating from more extensive areas of ice and snow developed in the form of ice caps or an ice sheet, as suggested by Kuhle (1988). This supposition is, as yet, inadequately proved, although sound in the light of some geophysical and paleoclimatological evidence.

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HISTORIC SLOPE DEGRADATION ABOVE TIMBERLINE IN THE BALKAN MTS., BULGARIA

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ABSTRACT. Climatic amelioration in the Holocene resulted in a rise of the upper timberline in the Balkan Mts. (Stara Planina) up to 1800–1950 m a.s.l. Incorporation of the Balkan Peninsula in the Turkish Empire in the 14/15 centuries has opened the Muslim markets for mutton, skin and wool. As a result of grazing pressure, the upper timberline has been lowered by several hundred metres over the last 4–5 centuries. This exposed Pleistocene slope deposits to contemporary geomorphic processes. The results of these processes occur over the entire area above the timberline, but are in particular well developed between 1500–1700 m a.s.l. on slopes which were formerly afforested.

In these areas small slumps or landslides are very common on fine grained slope deposits. On the debris-loamy slope deposits semi-circular niches, 5–8 m long, are common. As a result of upslope enlarging and capturing of adjacent niches, the grass ridges separating them become isolated "monadnocks". In the areas underlain by shales, on slopes with gradients exceeding 30°, soil slides are particularly common.

Processes of slope degradation caused by the overgrazing exhibit various stages of development. Some areas are characterized by fresh relief forms, other by more subdued forms which are progressively colonized by vegetation.

Balkan Mts. (Stara Planina) is a mountain arc of the alpides, located in the northern part of the Balkan Peninsula. The present-day form of the Balkans reflects the differenial uplift of a Miocene land surface of a low relief (Vaptsarov 1982). Fragments of this Miocene surface are well developed in the summit plateau of the Zlatensko-Tetevenska Planina and Troyan Planina, especially, in the zone of crystalline rocks that form the core of the massif (Fig. 1). The Subbalkan fault cuts the southern flank of the Balkans and has resulted in the formation of steep slopes of relative altitude of 1500–1700 m (Fig. 2). Northern slopes are more gentle and formed of sedimentary rocks of various age.

The Balkan ridge, over 600 km long, runs in an east-west direction between latitudes 42°40′–43°40′ N and forms a barrier separating temperate and subtropical climatic zones. Consequently there are significant climatic contrasts between its northern and southern slopes, and large differences in relative heights result in the vertical differentiation of climatic and hydrological conditions, soils, vegetation and also present-day morphogenetic processes (Klimek 1989).

The 2°C annual isotherm lies at a height of 2100 m a.s.l. on the northern slope of the Balkans (Brzeźniak 1987) and 2200 m a.s.l. contour separates the zone ol the very cool and cool climates. The coldest month is January (mean temperature −7°C), the warmest July

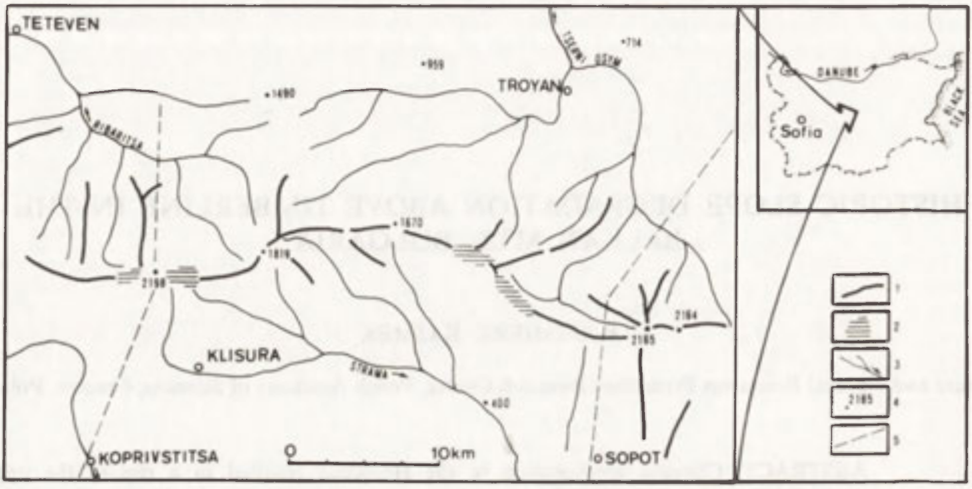

Fig. 1. Location map of the central part of Balkan Mts.: 1 — main mountain ridges, 2 — fragments of Miocene planation surface, 3 — main rivers, 4 — heights, 5 — cross-profiles (cf. Fig. 2)

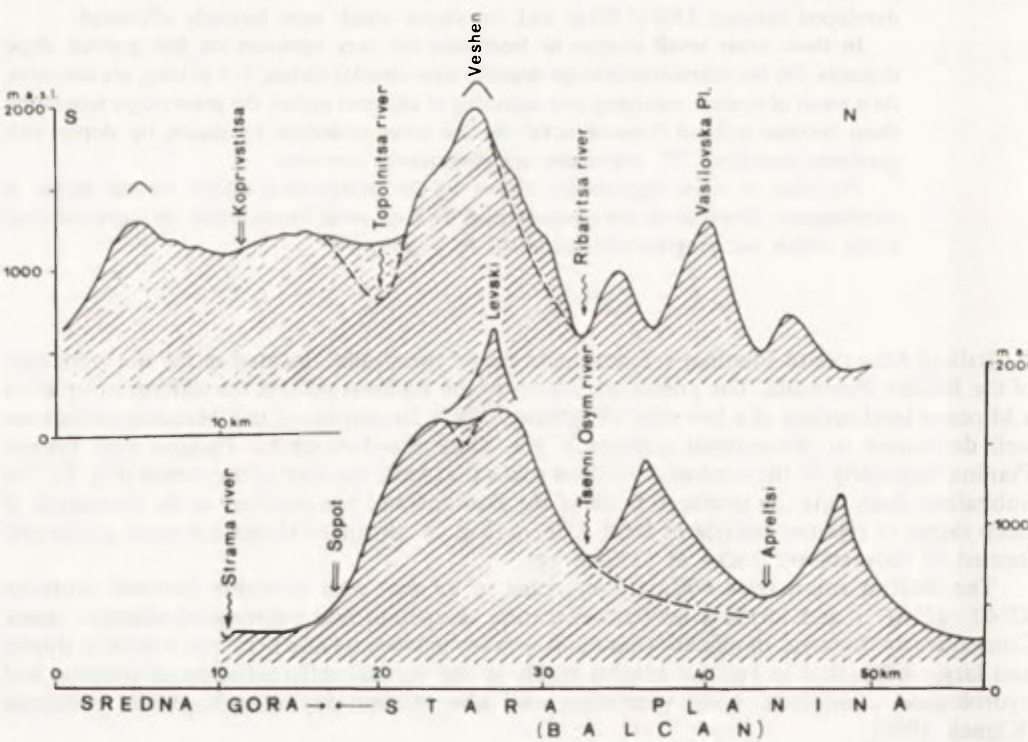

Fig. 2. Cross-profiles of the central part of the Balcan Mts. (Stara Planina)

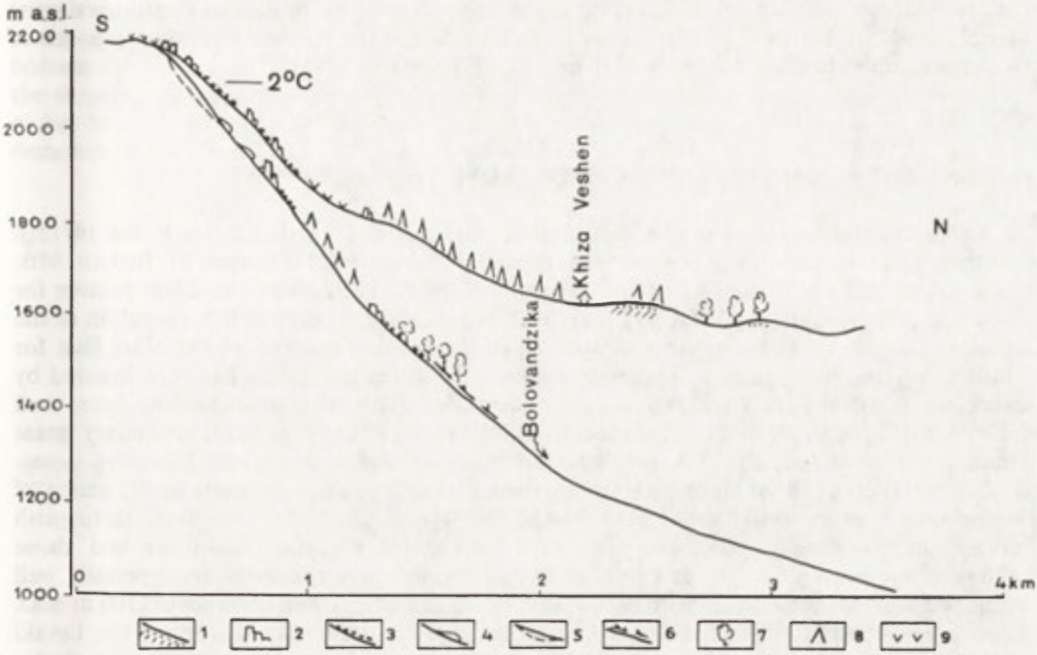

Fig. 3. Cross-profile of the northern slope of Balkan Mts. (in the neighbourhood of Veshen massif): 1 — Tertiary regoliths, 2 — rocky slope, 3 — block slope, 4 — debris-solifluction lobes, 5 — nival niches, 6 — avalanche levee, 7 — deciduous forest upper timberline, 8 — coniferous slope upper timberline, 9 — juniper scrubs, +2°C annual isotherm position

(mean temperature up to 9°C). Annual precipitation ranges between 1200–1300 mm. Vegetation cover on northern slopes of the Balkan corresponds to a climatic vertical zonality (Michalik 1987). Latifolious forests, mainly beech predominante up to the height of 1600 m a.s.l. Above, (Fig. 3) there is a zone of the coniferous forests of subarctic type with *Pinus peuce* and *Picea excelsa*, however, its upper limit has been lowered artificially to 1800 m a.s.l. as the result of grazing pressure (Filipovitch 1988). Subalpine plant communities dominated by secondary juniper (*Juniperetum sibirici*) reach up to 2100 m a.s.l. with the highest summits (Veshen, Levski, Botev) covered with grassland of *Festuca Agrostis*, *Carex*, *Vaccinium* type. According to Gylybov (1966) above 2300 m a.s.l. a cryonival zone occurs, however, this includes only parts of the Botev massif (2376 m a.s.l.).

PLEISTOCENE AND HOLOCENE BACKGROUND

Global climatic cooling during the Pleistocene caused a lowering of climatic zones, and as a result, the higher part of the Balkan experienced a periglacial climate in cold stages of the Quaternary. Degradation of slopes resulted in a creation of a range of periglacial forms and deposits. For example, on northern slopes above 1700 m a.s.l. formed of granite, vertical rock faces are common, below which extensive block fields occur. There are also numerous inactive debris-solifluction lobes, that generally are found on slopes with gradients below 25–30° and occur down to 1400 m a.s.l. Relict periglacial forms and deposits suggest that the lower limit of the cryonival zone during Pleistocene was at least 900 m lower than at present. This also confirms current opinion on the lowering of the upper timberline in Balkan Peninsula during glacial times (Starkel 1977, 1984). Climatic amelioration in the Holocene resulted in a rise of the upper timberline up to 1800–1950 m a.s.l. (Filipovitch 1981).

PRESENT-DAY GEOMORPHIC PROCESSES ABOVE THE TIMBERLINE

Before the incorporation of the Balkan Peninsula in the Turkish Empire in the 14/15th centuries there is little evidence of man related environmental changes in Balkan Mts. However, the flat and dome-like summit plateaus of the Balkans offered excellent pasture for sheep and higher biomass production than in the Asia minor. Therefore incorporation of the Balkans into the Turkish Empire and opening of the Muslim markets of the Near East for mutton, skin and wool, increased grazing pressure; the upper timberline has been lowered by several hundred metres over the last 4–5 centuries. This phenomenon has been well documented by the paleobotanical studies (Filipovitch 1981, 1988), with secondary grass communities with *Juniperus Vaccinium* and *Bruckenthalia* developing on deforested areas.

Deforestation of the summit plateaus exposed Pleistocene slope deposits to the action of contemporary geomorphic processes. (Mikhailov 1982). One can, however, distinguish between morphogenetic processes controlled by natural climatic conditions and those influenced by sheep grazing. The effects of natural geomorphic processes are especially well developed on the northern slopes of the mountain massifs at heights above 2100 m a.s.l. (Levski, Veshen). Below inactive Pleistocene rock faces on the northern slope of the Levski massif (at the height 2050–2100 m a.s.l.), on slopes with gradients above 30°, active chutes originate. These are 1–2 m wide and up to 1 m deep and in their upper sections are cut in rock. In their lower parts they are incised into debris filled gullies. Coarse material up to 10–20 cm in diameter is transported down these chutes. Needle ice appears to be important for sediment movement in day periods, as observed during high air pressure conditions in October 1985. Debris is transported down the chute to ca 1900 m a.s.l., where it is deposited in the form of a small (30 m long) talus cone. At numerous sites actively sedimenting cones are inserted into inactive, older cones deposited in the last few hundred years (Little Ice Age?).

On the northern slope of the massif Veshen between 2050–2100 m a.s.l., on debris mantled slopes with gradients above 20° nival niches are frequently formed up to 50 m wide and up to 100 m long. Usually, they are located on the lee side of passes through which snow is blown by southern winds. Snow patches survive in niches to the end of May and hinder the growth of summer grasses. Consequently, the exposed bottom of the niches is susceptible to erosion by either surface runoff (from rain or meltwater) or deflation. Runoff from snow patches frequently creates chutes below niches.

Large thicknesses of winter snow on the higher northern slopes of the Balkans (in February up to 160 cm at 1300 m a.s.l., Gerasimov, Galabov 1984), initiate snow-debris avalanches. Avalanches usually form at valley heads and move downslope in slush-avalanche gullies, widening and undercutting slopes and depositing debris levees on the valley sites. Debris levees, with boulders 1–2 m in diameter, were observed in the forest zone at 1400–1500 m a.s.l. (Fig. 4). In the Desen Tsaritsin gully debris levees are overgrown by trees, the oldest of which (> 120 years old) are frequently broken or demaged. This suggests greater avalanche activity in previous centuries probably during the second part of the Little Age (Grove 1988; Orombelli, Porter 1982).

Results of the morphogenetic processes induced by sheep grazing occur over the entire area above the timberline, but are particulary well developed between 1500–1700 m a.s.l. on slopes which were formerly of forested. Deforested undulating summit plateaus, or south facing slopes of smaller gradients, are usually covered by up to several metres of loamy-debris. In this areas small slumps or landslides are very common on poorly resistant shales covered with fine grain slope deposits. Mass movement appears to be triggered by sheep paths traversing slopes of gradients exceeding 15° (Fig. 5). Head walls of landslide niches or larger slumps provide wind-breaks for sheep whose trampling destroys plant cover. As a result the bottom of the niche is subjected to erosion by rain or overland flow from the upper sections of the slope. Eroded material is deposited in the form of a small alluvial fan directly below the niche. Sections in the head wall of the niche (Fig. 5) reveal similar alluvial deposits and demonstrate the importance of surface erosion in recent times.

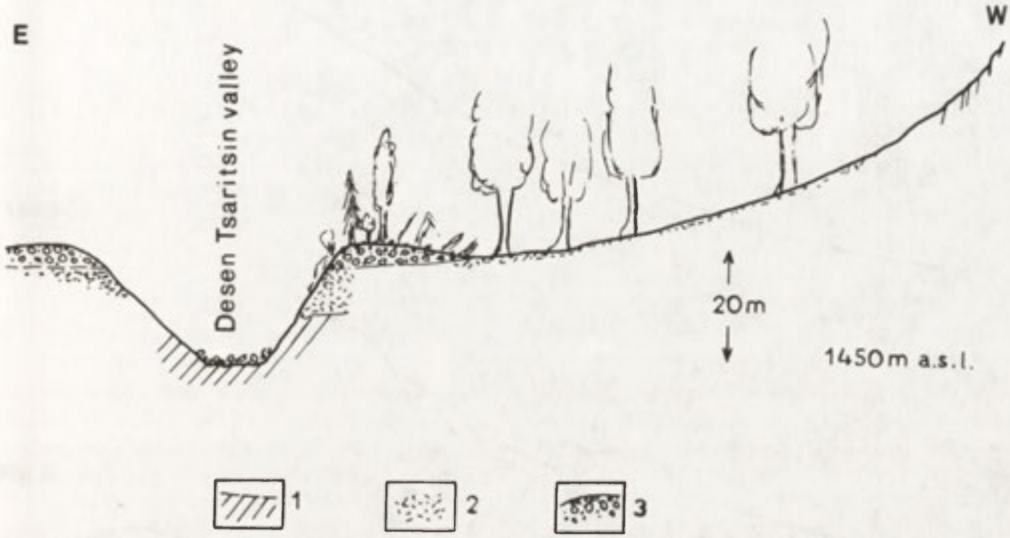

Fig. 4. Cross-section of a snow-debris avalanche gully on the northern slope of Veshen massif: 1 — bedrock, 2 — debris cover, 3 — snow-debris avalanche block

Debris-loamy deposits (0.5–1.5 m thick) occur on slopes with a gradient of 20–25°, usually on the gneiss or shist substratum. Here, sheep trampling results in the removal of grass, soil and roots, and the formation of shallow depressions. With further overgrazing these depressions develop into semicircular niches, 5–8 m long, which tend to enlarge upslope (Fig. 6), and capture adjacent niches. As a result of this process, grass ridges separating niches become isolated "monadnocks" (Fig. 6). Removal of fine material from the floor of the niche creates pavement or armour of coarse debris which restricts further degradation. Water draining system of niches is able to erode loamy-debris slope deposits and to form gullies up to 3 m deep.

In areas underlain by shales, on slopes with gradients exceeding 30°C, soil slides are particularly common. They frequently result in the exposure of the weathered regolith or bedrock in belts 20–40 m wide and 60–80 m long.

Processes of slope degradation caused by the overgrazing typically exhibit various stages of development. Some areas are characterized by fresh relief forms, others by more subdued forms which are progressively colonized by vegetation, confirming observations of Jahn (1979) in the Tatra Mts. that degradation of slopes can be effectively reduced by plant succession after grazing has ended.

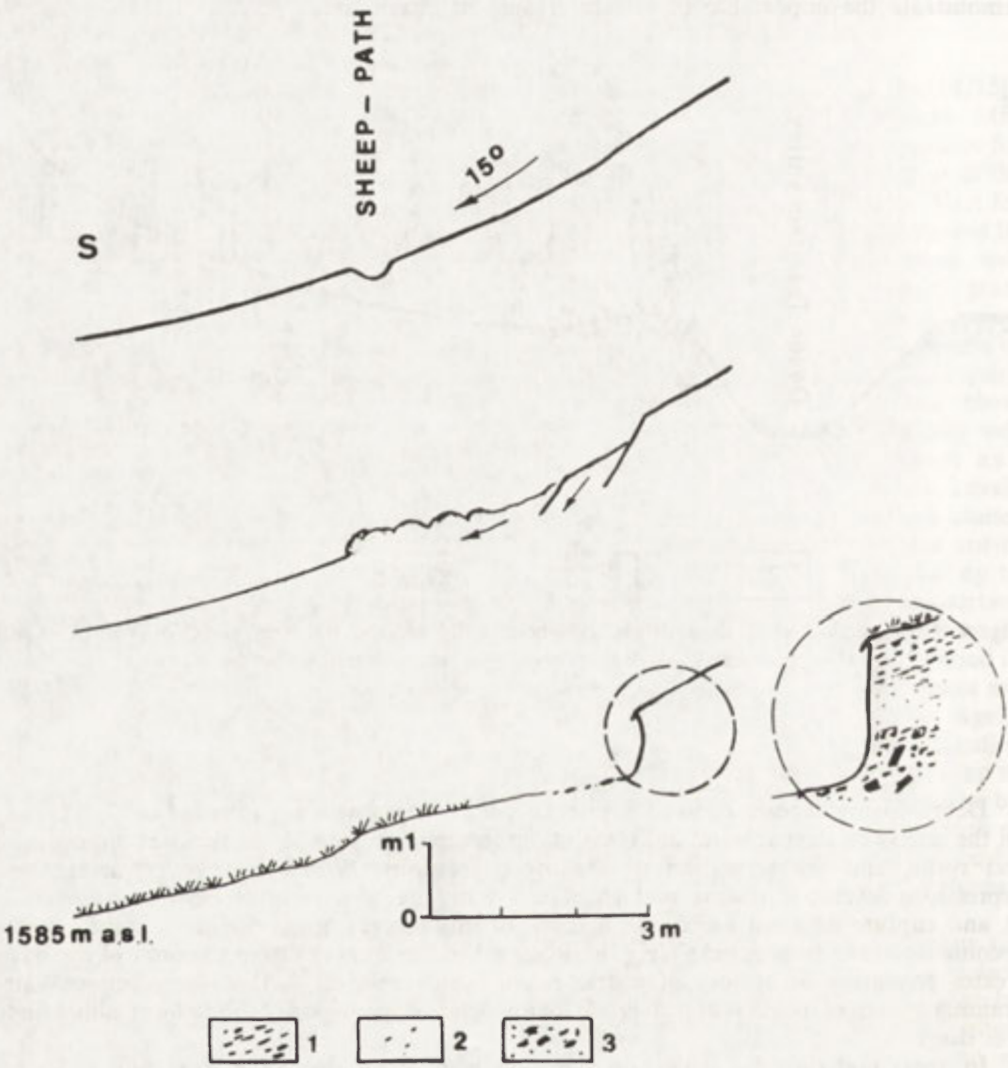

Fig. 5. Slope degradation caused by small landslides at the edge of sheep-path: 1 — alluvial debris, 2 — soil horizon, 3 — slope debris cover

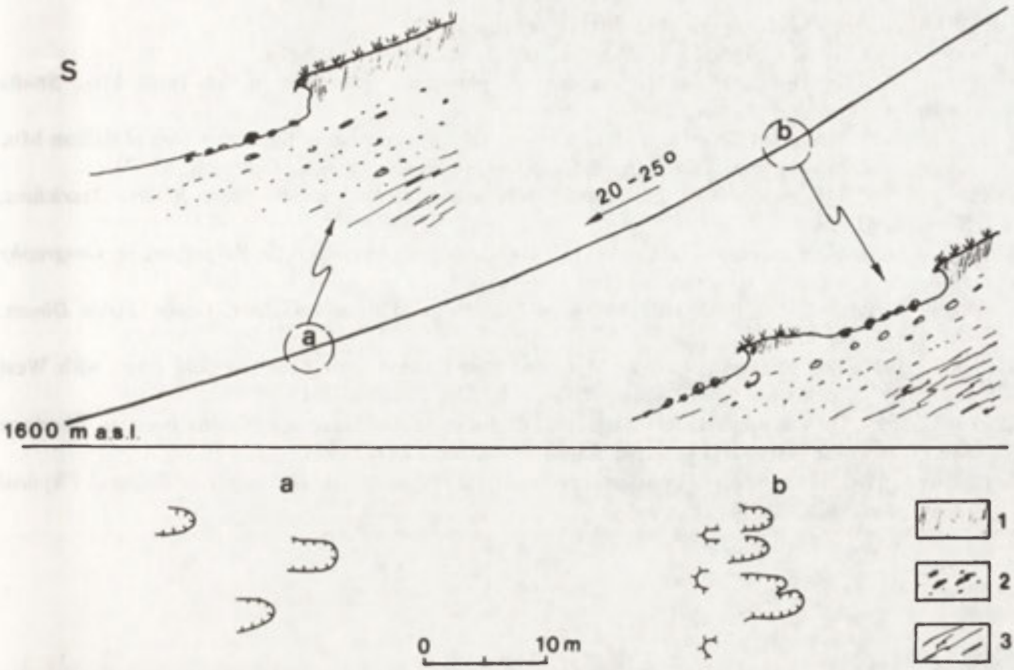

Fig. 6. Mode of slope degradation by sheep niches: 1 — soil cover, 2 — slope debris cover, 3 — bedrock (gneiss), a–b — stage of degradation

RESULTS AND CONCLUSIONS

Historic deforestation in the higher parts of the Balkan Mts. has resulted in significant slope degradation during the last few centuries. Anthropogenic factors, in particular overgrazing, appear to have been most important, although natural processes have accelerated slope erosion.

Acknowledgements

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THE HIGH MOUNTAIN FLUVIAL SYSTEM THE WESTERN TATRA PERSPECTIVE

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ABSTRACT. On the basis of the 12 years of field studies (1975–1987) regularities in the dynamics of high mountain fluvial system in an annual cycle have been determined. The results have been compared with the results obtained for other high mountain regions of the temperate zone.

1. INTRODUCTION

In recent years the interest in fluvial problems in high mountain areas has rapidly increased. Up to now the major interest has been focused on the glaciated high mountain areas, where seasonal, usually short term expedition-type studies have been carried out. Such studies are focused on channels in a close foreland of glaciers due to their high dynamics (Maizels 1978; Gurnell 1982; Hewitt 1982; Hammer, Smith 1983; Gurnell, Fenn 1984; Richards 1984).

The first monographs summarizing the knowledge on the dynamics of material transportation by proglacial rivers and transformation of their channels in a short distance from glaciers are issued (Gurnell, Clark 1987).

Although fluvial literature is extensive, quantitative approach to regularities in reshaping of the entire high mountain massifs from ridges to their foothills is not well recognized. According to Tricart et al. (1962) regular transit of debris from ridges down to the base level is just a myth and results mainly from imagination rather than from detailed field studies. On the basis of the analyses of alluvia originating from numerous Alpine valleys he stated that bedload (coarse sand, gravel and boulders) was transported over a short distance. Deposited material during transportation was replaced by other material which has not been moved over a long period, even during the entire postglacial period.

The explanation of the mechanism of bedload transportation in the high mountain areas in a quantitative approach has usually been attempted in a single small catchments and took into account one, two and exceptionally three types of transportation (Caine 1974; Maizels 1978; Hammer, Smith 1983; Richards 1984). The same applies to the Tatras, the only high mountain massif in Poland (Gieysztor 1971; Kaszowski 1973; Krzemień 1984).

Such a state of the investigations has induced the author of this paper to initiate the studies in the Western Tatras representing high mountains glaciarized in Pleistocene.

This work aims at:

1) determination of regularities in the dynamics of the fluvial system in the high mountain region of the Western Tatras,

2) comparison of regularities in morfodynamics of fluvial systems in various high mountain areas in the temperate zone.

According to J.H. Mackin (1948), R.J. Chorley and B.A. Kennedy (1971) and N. Caine (1974), the term “fluvial system” is understood as a sequence of dynamic subsystems interrelated by transfer of mass and energy. Therefore, stream reaches form the subsystems. The transfered mass is water, dissolved and suspended loads as well as bedload. All of them are carriers of energy.

Determination of vertical extent of fluvial system is an important problem of fluvial studies in high mountain areas. Action of high mountain streams is not only limited to the high mountain zone defined by C. Troll (1973). According to C. Rathjens (1982) mountains in which high mountain zone is present should be geographer’s field of interest as a whole down to the foothills instead of the high mountain zone exclusively. The high mountain zone affects the lower parts of mountains – down to the valley bottoms and foothills.

2. STUDY METHODS

The paper is based on the 12 years (1976–1987) field studies carried out mainly in Chochołowska valley which dissects the northern slope of the Western Tatras (Fig. 1).

In order to determine regularities in the dynamics of the morphogenetic fluvial system it was essential to investigate the following aspects: hydrodynamic properties of streams, dynamics of delivery and transportation of dissolved, suspended loads and bedload as well as dynamics of stream channels themselves.

Studies on fluvial transportation were performed in six major gauging profiles in Chochołowski stream (CH, L, D, H, S) and in Starorobociański stream (St) (Fig. 2). Those profiles delimit subcatchments whose areas vary from 8.78 to 34.78 km^2. Moreover, records were taken in eight other profiles (Fig. 2) on permanent or periodical basis.

Mechanism and intensity of bedload transportation were investigated using painted material (Miller, Leopold 1963) and 23 catching boxes of dimensions 100 × 50 × 50 cm which were placed at the level of the channel bottom. During the 12 years period, there were established ca 420 sites. Circa 50–400 pebbels and cobbels have been painted in each site. As a total ca 47 000 pieces of different sizes were marked. The load of bed material transported via channel cross-section was determined after each flood. Volume weight of 1574 t/m^3 was adopted for calculations. The accuracy of measurements of bedload is estimated for $\pm$5–12%.

In order to determine magnitude of dissolved and suspended loads 1 dm^3 water samples were taken every 3 days in 10 gauging profiles, in other profiles less often. The samples were taken more often in 4 gauging profiles, even every 15 and every 5 minutes.

The concentration of the suspended material was determined by centrifuging of water samples in the MPW-6 centrifuge and evaporation of water in temperature of 105°. The suspended load was determined by multiplying the suspended material concentration by water discharge and time intervals between them. The accuracy of calculated load based on suspended load records taken in smaller time intervals is estimated to be ca $\pm$8%.

The concentration of the dissolved load was determined by water evaporation in temperature of 105° and drying in temperature of 180°. As the relation between the concentration of the dissolved load and water discharge in a given gauging profile had been determined it was possible to calculate loads in particular days, months and years. The accuracy of load calculated on the basis of the records taken in smaller time intervals is estimated for $\pm$7%.

In order to determine regularities in modelling of stream channels cross-section and plans of stream reaches were prepared as well as stream loads and channel forms were recorded (Fig. 2). Periodical studies have been carried out in stream channels of Kościelski and Bystra (Fig. 1).

Parallely to fluvial studies, investigations on dynamics of debris flows in Starorobciański glacial cirque were done (Fig. 2).

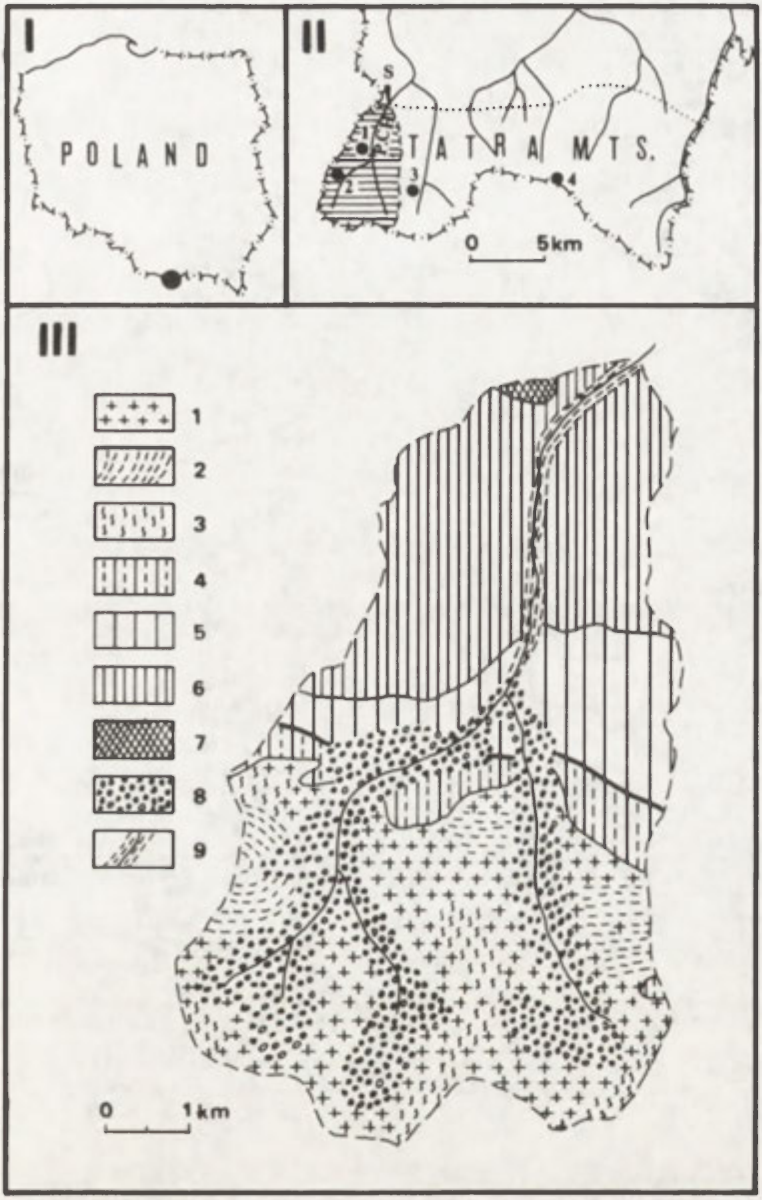

Fig. 1. Geological structure of the study area

I — Location of the study area in Poland; II — Location of the study area in the Tatra Mts.: P.Ch. — Chochołowski Stream; meteorological stations: 1 — forester's lodge (1028 m a.s.l.), 2 — Polana Chochołowska (1147 m a.s.l.), 3 — Hala Ornak (1110 m a.s.l.), 4 — Kasprowy Wierch (1991 m a.s.l.), S — water gauge in Siwa Polana; III — Study area: crystalline core: 1 — granitoids, 2 — gneisses, 3 — mylonites; high tatric nappes: 4 — quartzitic sand-stones, 5 — limestones, dolomites, shales; 6 — sub-tatric nappes series (mainly limestones, dolomites, shales and marls), 7 — conglomerates, eocene limestones, 8 — moraine and glaciofluvial covers, 9 — glaciofluvial and fluvial covers

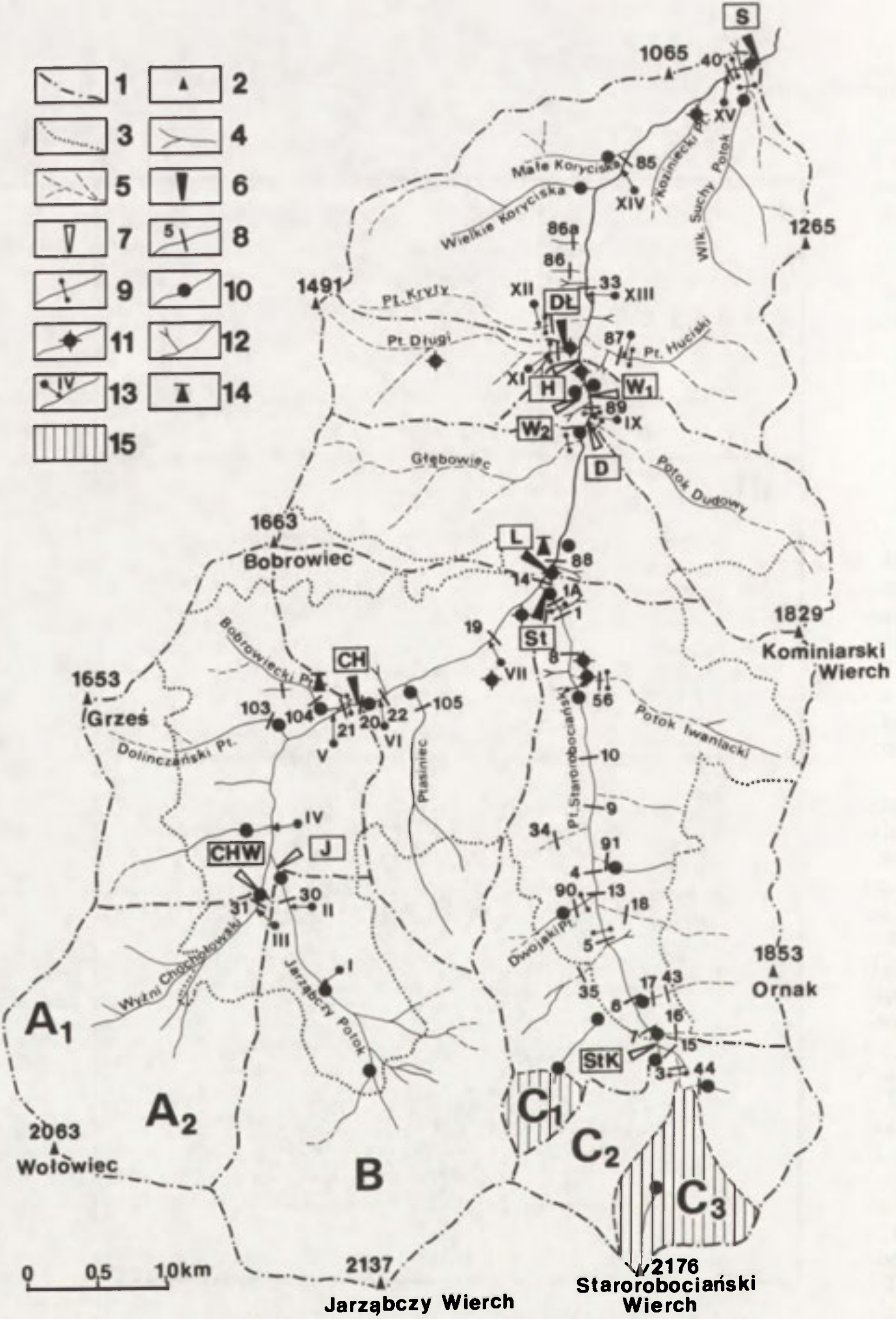

Fig. 2. Catchment of Chochołowski Stream

A_1, A_2, B, C_1, C_2, C_3 — glacial cirques, C_3 — Starorobociański cirque, 1 — water divide, 2 — major summits, 3 — upper timber line, 4 — permanent streams, 5 — intermittent streams, 6 — water level recorders, 7 — water gauges, 8 — sites with painted material, 9 — traps, 10 — dissolved material sampling sites, 11 — suspended material sampling sites, 12 — downcutting and lateral erosion measurements, 13 — particle size sampling sites, 14 — meteorological stations, 15 — study area for debris flow dynamics, StK — Starorobociański Stream — beginning of permanent stream, St — Starorobociański Stream — mouth, CH, L, D, H, S — gauging sites along Chochołowski Stream: CH — Polana Chochołowska, L — forester's lodge, D — outlet of Dudowa valley, H — Polana Huciska, S — Siwa Polana, CHW, J, W_1, W_2, DŁ — other gauging stations

3. STUDY AREA

The Western Tatras are an example of high mountains of the temperate zone, glaciarized in Pleistocene.

Due to limited possibilities, the studies were carried out mainly in Chochołowska valley which represents the main valleys of the northern slope of the Western Tatras. Those valleys dissect crystalline core and sedimentary series (Fig. 1) and are quite similar.

The valleys in their crystalline parts possess features of glacier troughs whose bottoms are lined with moraine covers. There are distinct ridges of end moraines as well as end-lateral moraines. In northern middle mountain part, there are V-shaped deep valleys with narrow bottoms and ungraded longitudinal profiles. The Chochołowska valley in this part of the Tatras has steep slopes and an almost flat bottom.

In the study period (1976–1987) mean annual precipitation totals did not differ significantly from the multi-year average and were 1378 mm in Polana Chochołowska (1147 m a.s.l.) and 1791 mm at Kasprowy Wierch (1991 m a.s.l.). A significant precipitation deficit took place in the recent 5 years only (1983–1987). The largest diurnal precipitation, up to 164 mm in Chochołowska valley, may occur from June till August. Snow cover appears in October and disappears in May and June.

The study area is found at the height of 925 to 2176 m a.s.l. within alpine, subalpine and forest zones.

4. STAGE OF DEVELOPMENT OF THE CHANNELS OF HIGH MOUNTAIN STREAMS OF THE WESTERN TATRAS

In the Starorobociańska — Chochołowska as well as in Jarząbcza — Chochołowska valleys, there has been developed a complex fluvial system, controlled to a large extend by glacial morphogenesis. Three basic types of channels can be distinguished in channel fluvial system in longitudinal profiles of the valleys in question, namely, channels in a glacial cirque, glacial trough and fluvial valley (Rączkowska 1983; Krzemień 1985). The high mountain fluvial system of the Western Tatras includes channels in subalpine and forest zones in valleys which had been entirely or partially glaciarized in Pleistocene. The channels of high mountain streams are characterized by similar properties. Their bottoms and banks are armored with boulders and blocks which originate from washed out moraine and glacio-fluvial covers. Numerous steps founded on large boulders occur in the channels in question. The mean heights of those steps are ca 0.5 m while the distance between them is 2.5–15.0 m. The relations between the distances separating the steps and the stream gradients are expressed by similar line equations: $y = -ax + b$. The high correlation coefficients from 0.87 to 0.89 are characteristic of those relations. Similar forms were studied in other channels of high mountain streams (Heede 1972; Day 1972; Kellerhals 1972; Gardner et al. 1983).

In the case of the Chochołowski Stream, in non-glaciarized part of the valley, the distance between the steps increases irregularly to the threshold value of 80 m and thus a qualitative change takes place. The steps and kettles being typical of the upper sections are replaced by riffles and pools. The threshold value occurred upstream of the Polana Huciska glade (gauging site H). Downstream, to the border of the Tatras the distances between riffles are differentiated (up to 233 m) while at the foreland of the Tatras those distances increase (even up to 300 m).

5. HYDROLOGICAL REGIME OF STREAMS

In the high mountain streams of the Western Tatras the runoff is concentrated in the summer season. Mean monthly discharges of the streams reach the highest values in May and the smallest values in February and March. In 1983–1987 mean annual discharge of the

Starorobociański Stream was 0.25 m^3s^{-1}. The discharge of the Chochołowski Stream was 0.40 m^3s^{-1} in the upper reach and 1.27 m^3s^{-1} at the Tatras border. The maximum discharges of the Chochołowski Stream at that time were 2.8 m^3s^{-1} in the upper reach and 8.4 m^3s^{-1} at the Tatras border. The maximum runoff in The Western Tatras takes place in May and therefore the regime of streams in that area is called temperate nival regime.

6. BEDLOAD TRANSPORTATION

In the channel system of the Starorobociański and Chochołowski streams bedload transport occurs rarely, 1–8 times per year at average in the lower section of the Chochołowski Stream and only 1–4 times per year on average in its upper reach as well as in Starorobociański Stream. The entire fluvial system in the drainage basin of Chochołowski Stream is remodelled very rarely, avery 2.5–5 years.

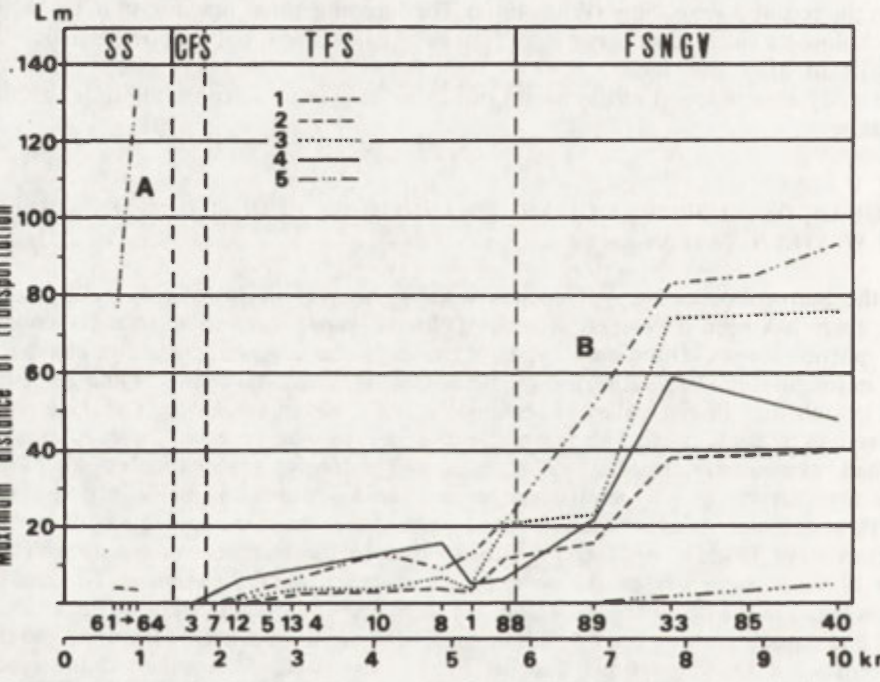

Fig. 3. Maximum transportation distance of painted material in longitudinal profile in debris flows troughs in Starorobociański Cirque and in channels of Starorobociański and Chochołowski streams during floods of various sizes

1 — 23 May 1987, 2 — 2 Jun. 1986, 3 — 8 Aug. 1985, 4 — 17–18 May 1985, 5 — 27 Sep. 1987; A — debris flows troughs, B — stream channels; SS — slope system, CFS — fluvial system of cirque, TFS — fluvial system of glacial trough, FSNGV — fluvial system of never glaciarized valley

In the fluvial system of the Starorobociański and Chochołowski streams the route of stream load transportation increases unevenly downstream (Fig. 3). In the channels of glacial cirques the distance over which the bedload is transported is particularly short — up to ca 1 m. Due to the presence of boulders and blocks as well as of numerous kettles and steps transportation of bedload the channels of glacial troughs is hindered and can occur over a short distance up to a few metres; in dry years to several metres, in moist years at average (Fig. 3). The mass transport does not occur in those channels even during catastrophic floods.

Transportation in slope chutes accompanied by debris flows takes place rarely, every 4 years, but over long distances (Fig. 3). In the study period, that transportation was founded to reach 133 m. Mean transfer of the material in debris flow troughs was 3 times larger than in stream channels of glacial troughs.

In that section of the valley which had not been glaciarized in Pleistocene (from the site 88), transportation in Chochołowski Stream increases rapidly to the region of the Polana Huciska glade (water gauge H) to several metres on average during floods (Fig. 3). In the lower section, up to the Tatras border the increase of bedload transportation distance is small or it does not occur at all. The presented regularity has been repeated during floods of various magnitudes (Fig. 3).

In the studied fluvial system bedload is subjected to multiple transportation and deposition before it becomes transfered to the Tatras foreland. Provided the most favourable conditions appear, the material delivered to the channel in the Starorobociański cirque can be transfered out of the Tatras after 3–4 thousand years.

Transportation model. Despite large differentiation in bedload transportation in longitudinal profiles of the stream channels of the Western Tatras, certain regularities are noticeable. They are presented in Fig. 4. Each curve represents combined results from some or several sites. Therefore, one curve shows maximum transfer of bedload in a given channel type during one flood.

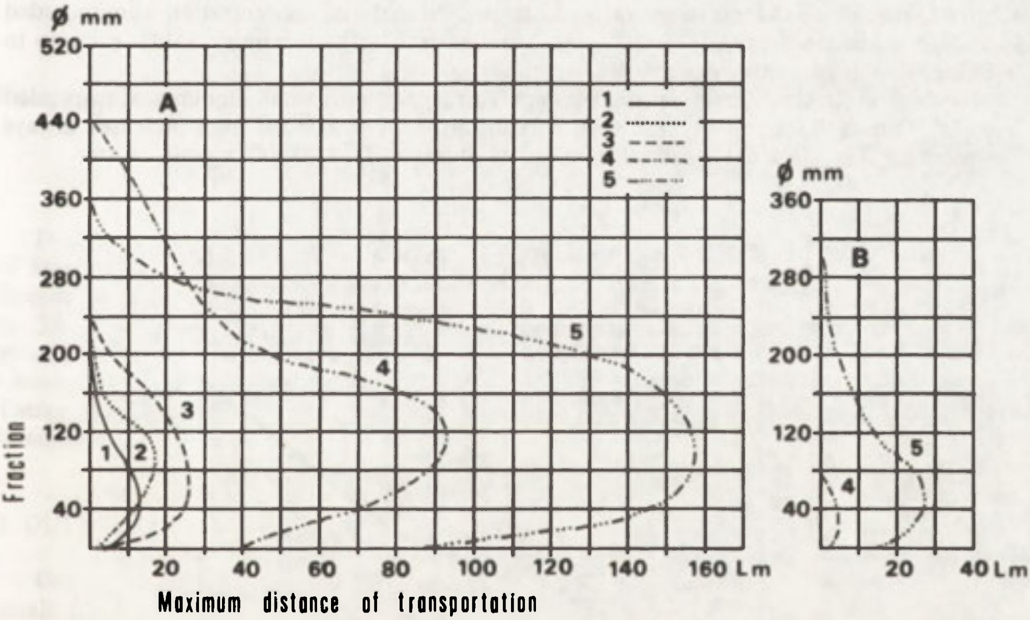

Fig. 4. Curves of maximum distance of bedload transportation in channels of Chochołowski Stream and its tributaries during floods of various sizes

A – flood of 23 May 1987, B – flood of 27 Sep. 1987, 1 – Starorobociański Stream, 2 – tributaries to upper reach of Starorobociański and Chochołowski streams, 3 – upper reach of Chochołowski Stream, 4 – Chochołowski Stream from forester's lodge (L) down to Siwa Polana (S), 5 – middle-mountain tributaries

In the stream channels in glacial troughs bedload was transfered over similar distances, up to several metres during medium and large floods (Fig. 4A, curve 1). Only in the case of Chochołowski Stream channel, in section with even out bottom, cobbles up to 20 cm could have been transported over a slightly longer distance – to 26 m (curve 3). In the channels of tributaries in glacial troughs, bedload of the size of 12 cm mainly was transfered over several meters. So, in all types of channels in glacial valleys bedload is transported over similar short distances (Fig. 4).

In the case of streams in valleys which were non-glaciarized in Pleistocene and which are cut in glacio-fluvial covers, the distance of transportation of bedload was 4–6 times longer than in the streams in glacial troughs.

In the channels of middle mountain area streams cut in mesozoic sedimentary rocks and fluvial covers, bedload was always transported over the longest distances. During medium and large floods it reached up 160 m. The transportation route in those channels was usually 8–10 times larger than in the channels in glacial troughs. During small floods in channels of medium-high mountain streams bedload was transported over 30 m whereas in the case of the channel of Chochołowski Stream cut in glacio-fluvial covers bedload was transported over 30 m the same time and in the channels in glacial troughs transportation did not occur.

7. SUSPENDED LOAD TRANSPORTATION

Concentration of suspended material in channels of the streams in glacial valleys varies during a year from 0.7 to 250 mg · dm^{-1} (Fig. 5). In the lower reach of Chochołowski Stream, in the channel cut in glacio-fluvial material, amplitudes of concentration were larger, i.e. from 1.2 to 965 mg · dm^{-1}. Much larger values of suspended material concentration were recorded in middle mountain streams — from 4.0 to 150 mg · dm^{-1}. The maximum values even up to 2150 mg · dm^{-1} were always recorded on roads.

The high mountain streams of the Western Tatras transport small amounts of suspended material. During floods of various sizes concentration of suspended load increases always downstream (Fig. 5). That material consists of at least 20–25% of organic matter.

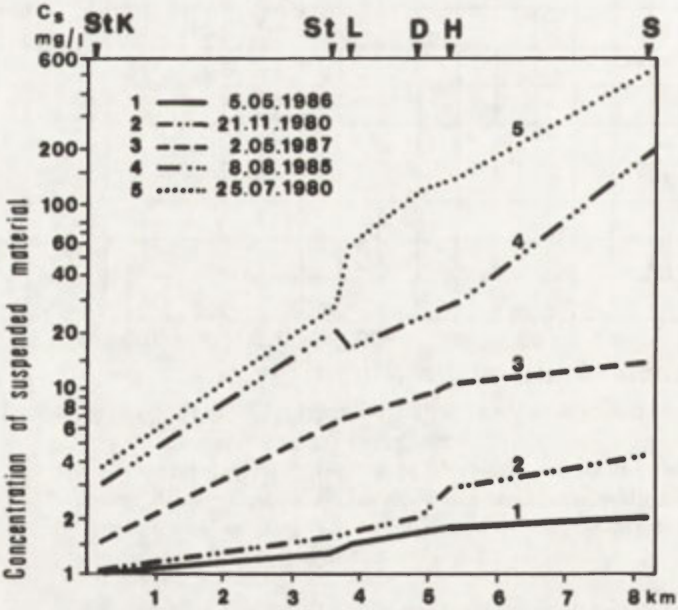

Fig. 5. Concentration of suspended load along Starorobociański and Chochołowski streams during low water stages (2), floods due to thawing (1, 3), medium floods induced by rains (4), large floods induced by rains (5); StK, St, L, D, H, S — gauging sites — cf. Fig. 2

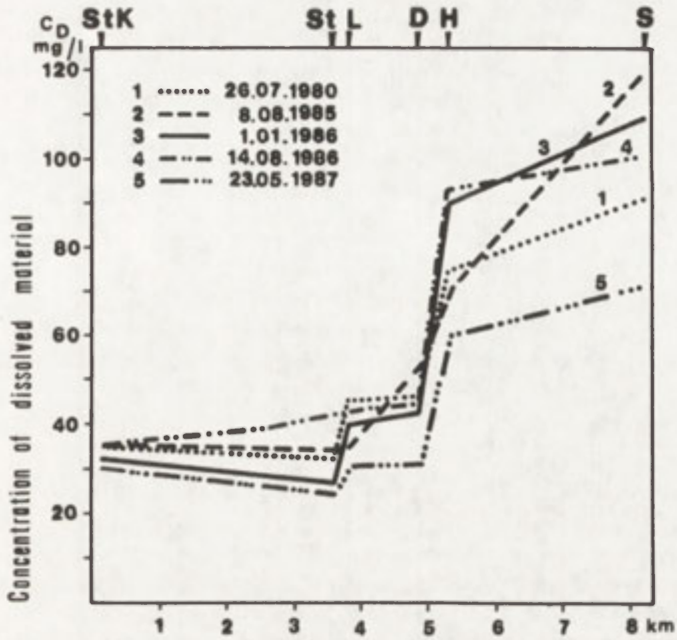

Fig. 6. Concentration of dissolved load along Staroroboclański and Chochołowski streams during: low water stages (3, 4), floods due to thawing (5), large floods induced by raine (1, 2). Other explanations as in Fig. 2

Dissolved load transportation. Concentration of dissolved load is differentiated downstream of Starorobociański and Chochołowski Streams (Fig. 6). Downstream of Starorobociański Stream to the outlet (water gauge S) concentration decreased usually from 30–35 $mg \cdot dm^{-1}$ to 25–32 $mg \cdot dm^{-1}$. That is related to the relatively high concentration of salts in groundwater filling the deposites occurring in glacial cirques. In the section of the Chochołowski Stream located slighlty lower, salt concentration increases to the border of the Tatras to the value of 58–114 $mg \cdot dm^{-1}$ (Fig. 6). Particularly large increase in concentration is recorded in a spot where water from karst springs is discharged to the stream.

8. OUTPUT OF MATERIAL FROM CATCHMENTS OF VARIOUS SIZE

Output of dissolved, suspended and bedload material from crystalline catchments was small: from 247–406 tons in the Starorobociański Stream catchment (8.8 km^2) to 876–1181 tons in the upper part of Chochołowski Stream catchment (23.3 km^2 – to the water gauge near forester's lodge – *L*). Output to the foreland of the Western Tatras from the entire catchment of Chochołowski Stream (34.8 km^2) was from 2853 to 4206 tons of the material annually. Load of material carried out to the Tatras foreland originates generally from the middle mountain region built of mesozoic sedimentary rocks. On the whole the dissolved load dominates over the suspended one and bedload (Table 1).

Material carried out from particular catchments cannot be related to the entire area of the catchment. Only dissolved material can originate from the entire catchment area. I was able to discover that clastic material (suspended solids and bedload) however, can originate from maximum 3.8–14.5% of the catchment area.

TABLE 1. Percentage of bed load, suspended load and dissolved load in total annual load in various high-mountain regions

Region	Catchment area km^2	Bedload %	Suspended load %	Dissolved load %	Author
Alps – Mont Blanc massif Bossons valley	10.5	29.0	70.0	1.0	J.K. Maizels (1978)
Rocky Mountains (Canada) Hilda valley	2.2	54.0–57.1	39.9–44.5	1.5–3.0	K.M. Hammer N.D. Smith (1983)
Western Tatras Starorobociańska valley (St)	8.8	0.1–0.4	6.5–15.5	79.8–93.4	K. Krzemień
Western Tatras upper part of Chochołowska valley (L)	23.3	0.1–0.4	8.0–14.8	84.8–91.8	K. Krzemień
Western Tatras Chochołowska valley – border of Tatras (S)	34.8	0.1–0.3	4.3–8.0	91.7–95.6	K. Krzemień

9. DYNAMICS OF HIGH MOUNTAIN FLUVIAL SYSTEM OF THE WESTERN TATRAS

The high mountain fluvial system of the Western Tatras is characterized by a complex internal structure. There is a weak transfer of the material between its subsystems. Therefore, the entire fluvial system is reshaped only slightly. During the 12 years field studies the large changes in the system were not detected. Application of various study methods made possible to determine only a tendency in the development of the high mountain fluvial system of the Western Tatras (Fig. 7).

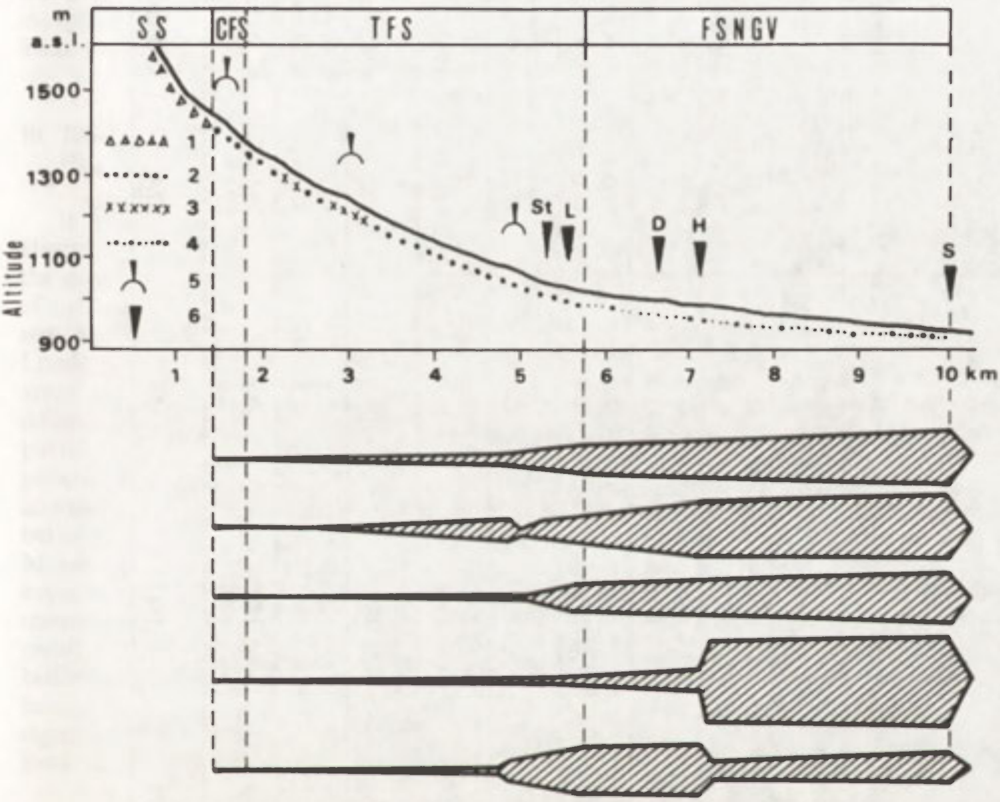

Fig. 7. Morphodynamic structure of channel system of Starorobociański and Chochołowski streams 1 — slope covers, 2 — moraine covers, 3 — rock outcrops, 4 — glacio-fluvial covers, 5 — ridges of end moraines and marginal-lateral moraines, 6 — water gauges, A — frequency of morphologically active floods, B — average maximum distance of bedload transportation, C — output of bedload, D — delivery of bedload from tributaries and bottom and banks of the main channel, E — erosion of bottom. Other explanations as in Figs 2 and 3

The frequency of modelling of particular reaches is significant in the whole longitudinal profiles of the channels of Starorobociański and Chochołowski Stream. In 1976–1987 the lower section of Chochołowski Stream was modelled 58 times, as so many times morphologically active floods took place. During those floods bedload was transported. Middle reach was modelled 31–39 times while the upper section in the glacial cirque only 3 times. Three fundamental sections which differ with respect to the frequency and intensity of their modelling (Fig. 7) can be distinguished in the longitudinal profile of the discussed fluvial system. The lower section, extending from the Polana Huciska glade (H) to the Siwa Polana

TABLE 2. Percentage of bedload and suspended load in total annual clastic load in various high–mountain regions

Region	Catchment area km²	Years	Bedload %	Suspended load %	Authors
Tsidjiore Nouve	4.80	1981	40.0	60.0	I.R. Beecroft (1983) vide
Sitzerland		1982	33.0	67.0	A.M. Gurnell, M. Clark (1987)
Bandhusbreen	12.60	1979	56.0	44.0	O. Kjeldsen (1981)
Norway		1980	42.0	58.0	O. Kjeldsen, G. Østrem (1980)
Engabreen	50.0	1979	37.0	63.0	O. Kjeldsen (1981)
Norway		1980	36.0	64.0	O. Kjeldsen, G. Østrem (1980)
Nigardsbreen	65.0	1979	43.0	57.0	O. Kjeldsen (1981)
Norway		1980	30.0	70.0	O. Kjeldsen, G. Østrem (1980)
Hilda valley	2.24	1977	59.0	41.0	K.M. Hammer, N.D. Smith
Canada		1978	55.0	45.0	(1983)
Bossons valley France	10.50	1971	30.0	70.0	J.K. Maizels (1978)
Starorobociańska valley	8.78	1980	5.9	94.1	K. Krzemień
		1983	2.4	97.6	K. Krzemień
Chochołowska valley	23.30	1980	5.9	94.1	K. Krzemień
upper part		1983	2.6	97.4	
Chochołowska valley Siwa Polana glade	34.78	1983	2.6	97.4	K. Krzemień

glade (S) is cut in glacio-fluvial material. The channels of Starorobociański and Chocholowski streams down to the Polana Huciska glade (H) form the middle section while the uppermost part of the Starorobociański Stream forms the upper section. In the middle of the profile presented in Fig. 7, there is the largest contrast in the frequency of morphologically active floods and in the intensity of various fluvial processes. As the result of long-lasting processes, the largest concave fluxion in the longitudinal profile should be formed in that spot due to relatively intensive downcutting (Fig. 7).

The present-day fluvial system is reshaped very slowly. That reshaping consists in washing away of glacio-fluvial and moraine covers, which results in slow elongation of the morphodynamic section located downward at the expense of the sections located upward. That process is hindered due to armored channels and numerous steps.

10. TRANSPORT OF MATERIAL OUTSIDE THE GLACIATED AND NON-GLACIATED HIGH-MOUNTAIN REGIONS

It follows from the studies of J. Maizels (1978) in Bossons valley and those of K.M. Hammer and N.D. Smith (1983) in Hilda valley, that glacial streams carry large amounts of the material as a total. In comparison with the amount of water flowing out of the catchment of similar size, proglacial streams carry 10 to 100 times more material than high-mountain streams of the Western Tatras. In the case of suspended load the differences are even larger. Loads of suspended material carried out of the catchment of comparable size in glaciated areas are 10–1000 times larger than in areas glaciarized exclusively in Pleistocene. Those differences are associated with a possible delivery of clastic material in given catchments and particularly with a stage of channel modelling during deglaciation and in the entire postglacial period. Material is carried out of the areas glaciarized exclusively in Pleistocene and of the currently glaciated regions in different manners. There are also differences in proportions between particular types of transportation (Table 1). Proglacial streams in the Alps and Rocky Mountains transport more suspended material, depending on the area, and small amounts of suspended load (1–3%). Streams in areas glaciarized exclusively in Pleistocene, however, transport mainly dissolved load and only small amounts of bedload and suspended load. As it results from Table 2, alpine proglacial streams transport much more suspended material than bedload. In the case of proglacial streams in Rocky Mountains and Scandinavian Mountains, bedload and suspended load can similarly contribute to the total clastic load. Suspended load significantly predominates over the bedload (in more than 94%) when considering the total load of the material carried out from the catchments by Tatra streams.

11. CONCLUSIONS

High mountains of the temperate zone consist of the present-day glaciated mountains and mountains glaciarized exclusively in Pleistocene. That fundamental differentiation of the environment affects the stage of development and dynamics of the fluvial system.

In the present-day non-glaciated mountains, the Tatras for example, the long-lasting transportation of the material (fine material mainly) in the entire postglacial period resulted in the formation of a mature fluvial system. The structure of the system corresponds to pattern of marginal moraines ridges. The present-day fluvial processes only slightly remodel the channel hence its natural structure.

The structure of the transported material in all gauging profiles in the high mountain fluvial system is characterized by the predominance of dissolved material over suspended load and bedload.

The glaciated high mountains of the temperate zone are characterized by a more complex fluvial system. There are: young active subsystem in the subalpine and alpine zones as well as mature, relatively stable subsystems in the forest zone. When considering the stage of their development, all high mountain fluvial systems in valleys partially glaciated and glaciarized exclusively in Pleistocene may be compared only within the forest zone.

In the present-day glaciated mountains, in a close distance from glaciers, the structure of the transported material is characterized by the predominance of suspended load over the bedload and dissolved load. Clastic material is deposited downstream in the area where valley becomes flat in lake basins and in valley bottoms. Part of suspended material is carried out to the mountain foreland via geoecological zone. Coarse clastic material remains in the mountain area.

Transport of weathered material outside the present-day nonglaciated high mountains occurs mainly due to large delivery of material originating from middle mountain area. The high mountain fluvial system is less important in that transportation.

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POLISH PEDOLOGICAL STUDIES ON SPITSBERGEN. A REVIEW

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ABSTRACT. The article is a review of Polish pedological studies carried out on Spitsbergen since 1957. The first part deals with the important papers going through the sections of soil science. Special attention was focused on arctic soil forming processes, physical properties (changes in soil volume, soil movements, thermal features), chemical properties and soil biology. Another part gives the lines for future pedological studies.

INTRODUCTION

The Polish studies on Spitsbergen started after World War II (in the 1930s there were three expeditions of a recognition character). The initiator of polar pedological investigations was a well known Polish geographer, professor A. Jahn who took part in the expedition to Greenland in 1937. The most active part in soil studies on Spitsbergen was undertaken by the University and Agricultural Academy in Wroclaw and moreover by Cracow, Toruń, Lublin and Warsaw Universities.

The pedological studies in the polar region, particularly those concerning soil forming processes are important for the studies carried out in Poland and other countries of Northern and Central Europe or Syberia. These areas underwent glaciation several times. They had a significant effect on relief as well as on quaternary geology and as a result on a soil cover. Knowledge of contemporary soil processes in the polar countries enables reconstruction of the processes which took place in Poland and other countries thousands years ago.

1. DIRECTIONS AND STATE OF INVESTIGATIONS

Hitherto existing studies have been converged in the Western Spitsbergen in Hornsund region, Bellsund and Oscar II Land. They cover both mineral and organogenic (peat) soils. The studies cover the main problems of contemporary soil science which are: 1) polar soil forming processes, 2) soil morphology, 3) mineralogical composition, 4) physical properties, 5) chemical and physicochemical properties, 6) biological activity, 7) soil fertility, 8) soil property dynamics, 9) soil classification, 10) soil cartography, 11) soil contamination, 12) paleopedology.

The state of investigations is quite differentiated. It will be presented while discussing the above mentioned sections.

There are relatively many papers dealing with the processes forming polar soils, morphological features and classification, physical properties especially grain size distribution and chemical characteristics. There can be found some papers about soil biology. Very few papers, sometimes even single ones presenting the other problems have been published.

It should be emphasized that not only pedologists but also geographers (geomorphologists) study soil and polar ground forming processes. The precise differentiation between soil and ground — usually a soil parent rock is not always easy. However, some physical features like soil movements or thermal properties were studied by geographers.

1.1. SOIL FORMING PROCESSES

Soil forming in the arctic area is influenced by the three groups of factors: cryogenic processes, soil forming processes and parent rocks. According to some authors the cryogenic processes predominate over those soil forming and usually act in a parallel way to them. They cause structural soil formation.

"Structural soils" (introduced into literature by Meinardus in 1912) were studied only by geographers-geomorphologists whose object were periglacial problems. There should be mentioned the papers by Jahn (1946, 1948, 1958, 1961, 1968), Czeppe (1960a, 1961, 1966), Klimaszewski (1960), Baranowski (1968) as well as Dżułyński (1963), Jersak (1968), Szczepankiewicz (1968), Cegła (1973), Klatka (1958).

Cryogenic processes and structural soils are affected by climatic conditions particulary low temperature and a small amount of rainfall. There were many hypotheses (over 20) concerning structural soils. Almost all of them assume that soil structures were caused by soil mass movements within the permafrost active layer which can be included in 4 groups: circulatoty, pulsatory, fissure-expansive and hydratation-dehydratation (Klimaszewski 1978). The above mentioned Poles contributed much to these studies.

In the structural soil genesis according to Jahn (1970), the most important are 4 groups of processes: 1) frost segregation, 2) swelleng and cryostatic pressure, 3) shrinkage and fissure, 4) slope frost and gravitational movements. As a result there are formed structural soils of different types: circles, polygons with stone circles or frost fissures, loamy and clayey outflows, stripes and others. Despite much research tundra microrelief forming processes have not been fully explained. The water and climatic conditions particularly those concerning soil thermal conditions play a significant role.

The basic soil forming processes are an initial developing stage, gley, peat-formation, browning, alluvial and deluvial. These problems were studied mainly by soil scientists and partly by botanists (peat soils): Kowaliński, Szerszeń (1963), Szerszeń (1960, 1965, 1968, 1974), Klimowicz, Uziak (1988), Melke, Chodorowski, Uziak (1990), Klimowicz, Melke (1991), Fabiszewski (1976), Klementowski (1977).

Initial developing stage soils are closely connected with rock formations and loose deposits mainly of stony and gravel character. They take up a large area in the studied region and are often soil free in practice.

The processes and alluvial soils are connected with river valleys and deposits but the deluvial soils are formed at the foot of the hill and sometimes on its slopes. Both groups of soils do not take up large areas in the soil cover.

Gley and particularly peat soils are formed under the strong influence of water. The essence of a gley process is reduction observed in the colour of a soil profile. This is shown in the studies on morphological features in some soil science papers and redox potential (Eh) as well as oxygen diffusion rate (ODR) in very few papers. Gley soils are very common and related to more compact formations as well as moist tundra.

Peat soil moistening can come from the stagnant water (in local depression) and from that flowing down the slopes. In this case moistening comes from thawed water and summer rainfalls. In this way there are formed so called slope or close to the slope morasses. Peat bogs are of low character and seldom transitional. Mosses and gramineous plants constitute main vegetations but there are no sphagnum mosses. Peat thickness is very differentiated (10–180 cm). Bird colonies have a great effect on eutrophy of waters nourishing peatbogs.

The browning process in morphology is relatively poor. To establish its course in the arctic climate, the mechanism of horizon (B) formation must be explained. It may be of a relict or weathering character. In the present stage of investigations, the view by Uggolini and Sletten (1988) about a weathering character of horizon (B) may be justified. It should be noted that brown soils were created on light formations and under dry tundra.

1.2. SOIL MORPHOLOGY

The investigations carried out on soil morphology refer mainly to macromorphology i.e. the observations of outer characteristics of soil profiles in the terrain. Micromorphology based on the studies of soil thin section under microscope is only a part of the paper by Szerszeń (1974).

Morphological features of Spitsbergen soils are poorly marked which is stressed by the authors studying soil forming processes in this area. The soils are usually of a shallow profile and genetic horizons are poorly formed because frost phenomena disturb soil forming processes. Many of them have features of initial developing stage soils.

1.3. MINERALOGICAL COMPOSITION

The problem of mineralogical composition was studied only by Szerszeń (1974), Szerszeń and Chodak (1983) and Chodak (1988). They stated that primary rocks were evenly distributed in the soil profile as was the case with secondary minerals. Of the primary minerals: quartz, grains of biotite, muscovite and feldspars as well as cacilte were found in the studied soils.

Chlorite, sericite, quartz, feldspars and illite as well as minerals transitional between primary and secondary were found in a clay fraction (< 2 μm). There are no smectites which is characteristic for cold zone soils (mainly physical weathering). Clay minerals are poorly hydrated, of a small specific area and little sorptive capacities. Illite occurring in the soil of this zone is different from that in other zones. It is homogeneous in shape and size.

1.4. PHYSICAL PROPERTIES

They cover such characteristics of soils as: grain-size distribution, so called primary physical properties (density, porosity, compactness, permeability, viscosity, plasticity, swelleng, shrinkage, structure) and secondary (water-air and thermal) properties.

The data concerning grain-size distribution i.e. granulometric composition are presented in some soil science papers about Spitsbergen. The arctic soils are characterized by share of all mechanical fractions, though in different proportions (from sands to clay loam) similar to some post-glacial formations in Poland. Participation of a skeleton fraction is particularly great. The content of a colloidal fraction is differentiated and sometimes very large (to 60%). In some soils the content of a silt fraction — a result of frost weathering, is significant. This phenomenon also occurs in other areas e.g. in the Tatra or Mongolia.

The studies of basic physical properties such as density, porosity, water capacity are limited (Szerszeń 1965; Klimowicz, Uziak 1988). There are no data about water-air properties. The soil moisture content and level of ground water were studied by Czeppe but the results have not been published. As for air properties there is known only one paper about CO_2 in soil air (discussed in the section about dynamics). The water-air properties can be determined from morphological features and the values of redox potential and ODR.

Relatively a great number of papers is devoted to soil freezing, thawing and vertical movements as well as thermal conditions namely soil temperature. These problems were studied by periglacialists and meteorologists (in the field of temperatures). The soil scientists did not participate in them though their object were soils and their parent rocks.

The above mentioned studies were carried out in different regions of Spitsbergen by various scientific centers: mainly in the stationary center of Polish Academy of Sciences in Hornsund (the Institutes of Polish Academy of Sciences and Universities), in the base close to the glacier Warenskiold – the Wrocław centre (University, Agricultural Academy), at the station in Sörkappie – the Cracow centre (University), in the area of Kaffiöyra – the Toruń centre (University and Departments of the Polish Academy of Sciences), lately in Bellsund – the Lublin center (Maria Curie-Skłodowska University). Only very few studies of these problems were made at the station in Hornsund throughout the year and therefore they are of great importance (Czeppe, 1961; Baranowski, 1968). The others were made in the summer periods.

As stressed by Jahn (1987), large scale experiments and measurements of quantitative determination of periglacial process dynamics were a great achievement. Some soil movement measuring device (according to Bac) to determine volumetric changes in freezing soils were applied for the first time in the world.

On the account of many papers (about 50) devoted to ground freezing, thawing and vertical movements as well as soil thermal conditions and limited space of the article there will be mentioned only some of them and only the most important problems will be presented. The papers by Czeppe, Baranowski as well as Jahn, Grześ, Marciniak and Szczepanik, Miętus, Glowicki should be mentioned.

According to Czeppe (1960b, 1961, 1966) the course of soil movements i.e. ground freezing and swelling as well as its thawing and sedimentation is closely dependent on air and ground temperature. He distinguished a few successive phases in a yearly cycle of frost movements. The strongest soil movements resulting in structural soil formation taken place in autumn while they are much weaker in spring. Summer and winter do not affect them significantly.

Studying a whole year's temperature course in soil Baranowski (1963, 1968) showed that tundra in Hornsund was characterized by frequent ground freezing and thawing cycles. Their number depends on depth (on the soil surface – 50 cycles, at the depth of 10 cm – 13 cycles). It follows from the ground vertical mean temperature distribution that the deepest layers are the warmest and the ground does not freeze throughout the year at the depth of 6 m (which is the effect of fiord waters).

Marciniak and Szczepanik (1983) stated that summer thawing was spatially differentiated and changed with time (it was fast at the beginning of summer and slow in the end). Moisture content in the ground a vegetation cover and types of rock are also of great effect.

From many measurements of Grześ (1984, 1985) determined thawing curves for individual kinds of ground (totally 8).

The investigations by Glowicki (1985) are very interesting. On the basis of direct field measurements he analyzed yearly and 24 hrs course of heat flux conduction in the soil on the background of its main physical properties. However, Miętus (1988) analyzed the results of many years' soil temperature measurements at different depths made throughout the year.

The investigations on Spitsbergen and in the polar regions by Jahn (1946, 1948, 1961, 1968, 1982, 1983) are of particular interest. He presented a model of summer ground thawing. The rate of ground thawing drops with the depth in a geometrical progression.

The problems of ground thawing and temperature courses in soils were also studied by: Chmal et al. (1988), Grześ, Babiński (1979), Gluza et al. (1988), Gluza (1990), Kejna (1990), Kejna et al. (1991), Marciniak et al. (1981, 1990, 1991), Miętus (1988a, 1988c, 1988d), Migała (1988), Repelewska-Pękalowa et al. (1987, 1988a, 1988b, 1988c) Wójcik et al. (1984, 1987, 1988, 1990a, 1990b).

1.5. CHEMICAL AND PHYSICOCHEMICAL PROPERTIES

These properties were seldom discussed in separate papers. They usually constitute a chapter of the paper dealing with many other problems. The results of the simplest determinations: reaction $CaCO_3$, K_2O and P_2O_5 as well as humus and total nitrogen contents are most frequently given. Only some papers present the results of the studies on humus fractional composition, cations in a sorptive complex, total chemical composition, mobile iron and others.

The soil reaction varies from alkaline to acid depending on the character of a parent rock and presence of $CaCO_3$. The organic soil reaction is usually acid or slightly acid. The amount of $CaCO_3$ also varies, it does not occur in many soils (usually in organic) and in others it may reach up to 50%. It should be said that besides $CaCO_3$ there can be found $MgCO_3$ in some soils.

The organic substance occurs in the studied soils in different amounts; in mineral soils – from 1.5 to several per cent (usually 1.5–5.0%), in organic soils up to 90%. In spite of small and slow biomass increase, a great amount of organic substance is accumulated because its decomposition is slow. It is also poorly distributed. The organic substance is present in the whole profile though in the largest amount in horizon A_1 which is a result of organic substance dislocation because of cryogenic processes.

Inactive compounds are most abundant in the arctic soil organic substance of Spitsbergen (over 2/3 rds of the total). As a rule fulvic acids are two or three times as humic acids (Boratyński et al. 1968; Szerszeń 1974).

Nitrogen content is dependent on the organic substance content and therefore it is differentiated. The ratio C:N also varies but most often it is 1:8–14.

Sorptive properties depend on the contents of organic substance and clay fraction. In many soils exchangeable cations are ranged: $Ca < H < Mg < Na < K$ and sometimes Mg comes before H. Generally cation content in organic soils is much higher (even several times) compared with the mineral ones. There is also a great contrast in sorptive complex saturation by cations. The degree of base saturation of soils both mineral and organic is differentiated (50–100%; Szerszeń 1964, 1974; Melke et al. 1990).

The data concerning the total chemical composition of a clay fraction ($< 2\,\mu m$) are included in the paper by Szerszeń (1974). The SiO_2, Al_2O_3 and Fe_2O_3 contents are often greater in horizon A_1 than in the bottom horizons though morphological differentiation is not observed. Szerszeń also stated that large amounts of mobile iron forms equally distributed throughout the profile could be found in the arctic soils of Spitsbergen.

It is worth mentioning about the paper on the contents of heavy metals (Pb, Zn, Cu, Mn) in soils as well as in vascular plants, mosses and lichens in the Bellsund region (Jóźwik, Magierski 1991) which shows that Zn and Pb contents in the studied soils are high as in the case of plants where high content of Mn was also found.

1.6. BIOLOGICAL ACTIVITY

Microbiological studies in the polar region soils began 90 years ago (rather in the Antarctic) and are well developed. The Polish investigations began on Spitsbergen within the International Geophysical Year in 1957. They were carried out occasionally due to the initiative and help of the Wroclaw centre geographers. Their specific character and problems resulting from it should be stressed. Though there were no stationary bases, significant achievements are noted.

The papers mainly by the Wroclaw center researchers covered the groups of fungi, algae, actinomycetes and bacteria in soils mostly in the Hornsund region (Zabawski, Żurawska 1975a, 1975b, 1977; Zabawski 1976, 1978, 1982, 1984; Krzywy et al. 1961; Wieczorek et al. 1964; Matula 1982; Plichta, Lusińska 1988). It was shown that in all studied soil habitats bacterial and fungal microflora and also in most of them, actinomycetes were found. The biotopes close to glaciers were characterized by poor microorganism content as far as number and species are concerned. The most abundant was the bog microflora with fungi and bacteria due to a large amount of poorly decomposed plant fragments and low reaction.

The qualitative analysis of soil material showed a great variety of forms and fungus species which is similar to the results coming from other latitudes. It should be noted that some fungi and algae (Matula 1982) have not been found in the polar regions so far. It is quite interesting to find a great similarity between bog soil microflora in Hornsund and mountain bog microflora in the Karkonosze and Tatra due to similar habitat conditions (Zabawski 1976, 1978, 1982).

Intensity of tundra soil metabolism in the Hornsund region was investigated by Fischer and Bieńkowski (1987) and Fischer (1988). Intensity of respiration processes was measured in soils by the oxygen consumption and the amount of emitted CO_2. Organic soils were characterized by the activity ten times as great as that of structural clayey soils. Soil activity is affected by temperature and humidity. In structural soils psychrophile microorganisms are responsible for oxygen metabolism but in organic soils fungi.

1.7. SOIL FERTILITY

There are a few papers dealing with tundra soil fertility. Apart from various chemical determinations, there were made those of available phosphorus and potassium as well as total nitrogen contents. They show that the contents of available phosphorus and potassium are low as a rule particularly in mineral soils. It is justified by the fact that chemical processes especially biochemical are dependent on biological activity which is very small in the tundra soils. However, this differentiated and sometimes even high in organic soils. It constitutes another problem as phosphorus and potassium in peat soils are determined by different methods. Thus, comparison of the results obtained by different methods must be made cautiously.

The papers by Czajkowska (1984) and Stępniewicz (1984) deal with the problem of plant fertilization and as a result of tundra soil by excrements from the birds' colony. Czajkowska points out to a significant role of NO_3 in plant fertilization through aqueous solutions. Stępniewicz calculated the amount of bird excrements (in tons of dry mass/km^2). The effects of such fertilization can be seen with the naked eye (even from a distance) in some places in a form of intensive green colour of the tundra vegetation.

As for the tundra soil fertility not only availability of components but also the water-air relations, particularly thermal should be taken into consideration. Taking the above into account, the estimation of soil fertility in this area connot be satisfactory.

1.8. SOIL PROPERTY DYNAMICS

Studies on property dynamics provide full knowledge about the processes in soils. Therefore they are of great value. However, they require a great expenditure of work because of measurements and determinations made every several days and even every day for a definite period of time. It refers largely to Spitsbergen soils.

There are three papers published on dynamics — Dziadowiec (1983), Melke and Uziak (1989, 1991).

Dziadowiec studied CO_2 release (every 3 days) from different soils in Kaffiöyra. The results point to very small respiration (CO_2 release) in the studied soils. They are much lower than in the soils of other latitudes. In its essence, the paper belongs to those devoted to soil biology.

The investigations by Melke and Uziak (1989) on redox potential (Eh) as well as oxygen diffusion rate (ODR) in 5 soils of Calypsostranda in the Bellsund region i.e. water-air relation of the arctic soils are precious and quite new. They showed that in the studied soils particularly in clayey soils reductive conditions and sometimes even strong anaerobiosis were found.

Another paper (1991) presents the results of preliminary studies on dynamics of the available components (Ca, Mg, Na, K, Fe, Mn, Cu, P, NO_3, SO_4 and Cl). There were found some regularities in their dynamics, some of them (Ca, Na, Fe, Mn, Cu) have a similar course and their contents are correlated with soil moisture (higher moisture was accompanied by higher contents of some elements and the other way round).

It should be emphasized that soil property dynamics covers the studies of soil moisture (carried out in a limited range) and soil temperature (at different depths) of freezing, thawing, vertical movements which were discussed in the section on physical properties.

1.9. SOIL CLASSIFICATION

Classification was discussed by Szerszeń (1965, 1974), Klimowicz and Uziak (1988) as well as Melke, Chodorowski and Uziak (1990) in the papers on general characteristics of Spitsbergen soils. Only Plichta (1977) devoted his paper entirely to polar soil classification.

Szerszeń based his classification on the genesis of the materials (of marine, deluvial, alluvial and organic origin) as well as on soil forming processes (gley, brown, structural soils and fossil soils).

Plichta made his classification partly from the literature data and partly from his own observations. He singled out the soil types according to Tedrow et al. from 1958 and the World Soil Map, FAO from 1974. Subtypes were dependent on microrelief which is an indicator of the processes taking part.

Classification adopted temporarily in the Soil Science Department, M. Curie-Sklodowska University is that by Tedrow and Plichta and partly own results of the investigations.

1.10. SOIL CARTOGRAPHY

Precisely cartographic studies were not carried out on Spitsbergen. There were only made schematic maps of soils from some terrain fragments by the way of general soil studies. Three papers should be mentioned.

The paper by Szerszeń (1965) includes the scheme of soil distribution in the Hornsund region (in the scale 1:300 000). Klimowicz and Uziak (1988) as well as Melke, Chodorowski and Uziak (1990) included maps of soils from some parts of the Bellsund region (the former in the scale 1:50 000 and the latter – 1:100 000). The number of shown soils in each map is different depending on terrain differentiation, lithology and even a snow cover free period.

It should be noted that independent of the map scale, they present soil complexes but not individual units. This results from a great variety of soils, difficult terrain, large labour consumption of cartographic works, small personnel and a long period of time to determine soil boundaries. However, this kind of investigations can be carried out for only three summer months.

1.11. SOIL CONTAMINATION

Soil contamination was studied only by Reszke and Szczypa (1991). They present the results of the investigations on soil and plant contamination by radioactive isotopes (in the energy range Cs-137) in the Bellsund region. It is surprising that the Spitsbergen environment proves to be greatly contaminated. According to the authors it is not only a result of the Czernobyl explosion but also vicinity of military firing grounds on New Land where nuclear explosions are carried out.

1.12. PALEOPEDOLOGY

Paleopedology was the object of investigations by Baranowski and Szerszeń (1968) which allowed for determination of climatic conditions in the period before the last progressive glaciation. The paper confirms the suggestion made by Baranowski that so called "interstadial" period was longer and warmer than the present warm period.

It should be said that fossil soils were discussed in several pedological papers (Szerszeń 1965; Melke et al. 1990) but they were not used for paleopedological interpretations. The soil age was estimated according to C^{14} to be 3–11 thousand years (Szerszeń, 1965) and in the opinion of others 9–12 thousand (Mann et al. 1986).

2. FIELDS OF FUTURE STUDIES

The state of hitherto existing Polish pedological investigations on Spitsbergen is an initial point for future lines.

The evaluation of the works done so far is not simple. The studies were not made systematically. It is not accidental that the Wrocław centre is leading.

Others centers participated rather occasionally. There was no all-Polish programme of studies which could be carried out systematically by various pedological departments. Close cooperation between soil scientists and geographers was necessary as only such a form was justified in its essence and organization.

Despite these difficulties great contribution to the studies of Spitsbergen soils has been made. Pedological studies in this area were also made by the Norwegians, Germans, Russians, Americans and others. Taking into account the achievements and possibilities of other coutries the Polish results of the investigations are significant.

The future programme should include two groups of activities: organizational and essential. The first should cover 1) coordination of research through a programme worked out for the centres taking part in the pedological studies; 2) closer organizational and scientific cooperation with geographers and biologists; 3) cooperation with scientists from other countries particularly with Norwegians and Russians and others who are interested in such a cooperation.

The second group of activities refers to research subjects which can include the following problems:

1. Polar soil forming processes – structural, gley, browning, evolution of these soils.
2. Physical properties, particularly those concerning volumetric changes in soil, its vertical movements, water-air and thermal relations.
3. Chemical and physicochemical properties: composition of a sorptive complex (organic compounds, clay minerals).
4. Expansion of the studies already begun, especially based on stationary measurements in longer cycles (from spring to autumn if possible). First of all it concerns dynamics of chemical, water and biological (changes in species composition of microorganisms and their biological activity) properties.
5. Studies of ecological character i.e. in the soil-plant system.
6. Preparation of a soil map at least of the regions in which the investigations have been made for many years.
7. Writing monographs about the important problems concerning the soils on Spitsbergen and generally about polar countries.

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PERIGLACIAL STRUCTURES IN SVALBARD AS INDICATORS OF A CENTRAL EUROPEAN CLIMATE IN THE LAST GLACIATION

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ABSTRACT. Contrary to widely held opinions, the archipelago of the Svalbard Islands does not belong to the oceanic part of the periglacial zone but has its own diversified climate being humid-maritime on the coast and dry-continental in the interior. These differences in the periglacial enviroments manifest themselves in contrasting structures of the active layer. The periglacial zone of the last glaciation in Europe was characterized by a climatic gradient more oceanic in the west and more continental in the east of Europe. The transitional section of this zone lay in its narrowest part, between the Scandinavian ice border and the northern timber line, delimited approximately by the arch of the Alps and Carpathian Mountains. In this part periglacial structures are abundant and, what is more variable in vertical profile.

This evidence reflects great variations in the glaciation period, of changes of the oceanic and continental phases. There are at least 4 frost wedge horizons, often of "sand wedge" type, that are indicators of a continental climate. The Svalbard archipelago where oceanic as well as continental facies of periglacial structures can be found at present provides a good representation of the spatial diversity of periglacial phenomena which occurred at different times in Central Europe.

INTRODUCTION

The principal purpose of the periglacial investigations to which I devoted 50 years of my life was to find an answer to the question whether the structures in th Pleistocene periglacial sediments known to me mainly from the regions of Poland, i.e. from Central Europe, have their counterparts in the contemporaneous periglacial zone. I searched for them everywhere — from Alaska and the Mackenzie river delta as well as Labrador through Greenland, Iceland, the Svalbard islands to northern Siberia. In this full circle of the global periglacial zone I learned to distinguish the periglacial forms and structures related to a humid, maritime climate as e.g. that of Iceland, as well as those occurring in the dry, periglacial climate of northern Alaska and Siberia. It may have been a matter of accident that the periglacial forms and structures I was able to study most closely were those of Svalbard, Spitsbergen — and it was in those structures that I found most analogies to what is represented by the group of fossil structures of Central Europe. The creator of the study on periglacial phenomena was Walery Łoziński "the father of periglacial studies", as he was called by A.L. Washburn (1979–1980). At the beginning of this century he put forward the concept of the so-called "periglacial facies" as a particular modification of the climate, accompanying the Pleistocene glaciation of Europe. Are the countries of Central Europe, including Poland, which were investigated by Łoziński, really classical periglacial areas of the Pleistocene? Or is this a matter of priority – that is, of the historical rather than the cognitive moment? The same applies to Spitsbergen which was explored earliest as a region of classical, actual periglacial phenomena, Was it by accident that Högbom's, Meinardus's and Nansen's investigations on frost as a factor of geological processes (*Über die geologische Bedeutung des Frostes*, the famous work by Högbom of 1914) were carried out mainly on Spitsbergen and made this

Fig. 1. Svalbard Islands situated in the transitional climatic zone. Limit of annual Arctic sea-ice cover in summer (1) and of floating pack ice in winter (2), Arctic (3) and Atlantic (4) sea currents. Direction of climatic influence of Arctic (5) and Atlantic (6). Investigated areas with periglacial maritime (7) and inland-continental (8) features

archipelago of islands, lying halfway between northern Europe and the Pole, a classical periglacial region? Although it was by chance that those two regions were to become regarded as "classical", it may be worth comparing them with each other in view of the concept best introduced by A.L. Washburn — that is, from the viewpoint of "periglacial environment." What it meant here is, of course, the contemporary and the Pleistocene environment, strictly speaking the environment of the last (Würm) glaciation. The following principle formulated by this author should, however, be taken into consideration: "The reconstruction of past climates through periglacial evidence requires extreme caution, but also holds promising rewards. To aid interpretations, high priority should be given to critical paleoenvironmental studies comparing the periglacial evidence with that derived from other types of proxy data" (Washburn 1985).

THE CLIMATIC CONDITIONS OF SPITSBERGEN

The Svalbard archipelago lies on the borderline of two oceans — The Atlantic and the Arctic ocean (Fig. 1). It would, however, be a mistake to assume that Svalbard-Spitsbergen is an area of the maritime variant of the periglacial climate. The maritime regions of the Arctic are covered with ice — their effect on the climate does not differ in any way from that of cold land areas. The cold air mass cap, the arctic anticyclone, affects the Spitsbergen climate from the north and east. The warm water of the Atlantic and the Icelandic Low related to it is the source of cyclonic travelling air masses reaching to the south of Spitsbergen, over the Barents Sea and even farther to the Kara Sea. These agents have a warming effect on Spitsbergen. This is why the Svalbard archipelago is a transitional area between the maritime and the continental variety of climate. This was also found by (among other researchers) G. Stäblein (1982) who — while investigating the aridity of the Arctic — included the Svalbard archipelago in the "cryo-arid geomorphodynamic system" of semi-arid areas. On Spitsbergen, the prevailing winds are of easterly and north-easterly directions. The mean annual air temperature varies around −6°C, which makes Spitsbergen an area of permafrost. The Spitsbergen summer lasts for three months — June, July, August — with a mean temperature of 3°C to 5°C. This is a sufficient heat source for the summer thawing layer to reach down to a depth of 0.5 to 2.0 m (Baranowski 1968; Pereyma 1983).

The conditions of climatic oceanity and continentality change in dependence on local factors. The glaciers covering the whole interior of Spitsbergen create a specific, climatic variety. In the internal part of the fiords the annual precipitation sum (about 200 mm) is twice as low as that on the external coast (over 400 mm). The climate of the interior of the islands shows continental features, whereas at the outlet of the fiords facing the Atlantic ocean the climatic conditions are maritime, which means more precipitation, cloudiness, fog and stronger winds.

ACTIVE LAYER OF PERMAFROST

The summer thawing of the soil is best characterized by the climate of the season. Generally speaking, the depth of thawing corresponds to the sum of heat directly derived from solar radiation or air mass advection. The rate of thawing down into the soil decreases together with the increase in thickness of the thawed layer. The process is thus spread out in time and takes its course in a function of decreasing geometrical progression (Jahn 1982). There exists a correlation between the depth of thawing and the square root of time (Jahn, Walker 1983). The warmer the summer the thicker the active permafrost layer. By and large, this factor expresses the global, zonal effect of the climatic. The thawing becomes deeper and more rapid if the soil is devoid of natural thermal insolation, the insolator being mostly vegetation cover or its transformations as e.g. peat. Other local agents affecting summer thawing in a periglacial environment are soil moisture, and slope inclination and direction. A combination of these four factors — that is, warm summer (W), cold summer (C), good (E) or bad (V) local susceptibility to thawing — permits the building of the following model (Fig. 2). The length of the line of the model corresponds to the depth of thawing and the inclination to the rate of thawing. The polar summer may last for two months (Franz Joseph Land) or for five months (some parts of Alaska or of north-east Siberia), thus the thawing depth may reach from 0.5 to even 3.0 m.

Investigations into the progress of summer thawing on Spitsbergen (see Grzes, 1984) in dependence on the local, environmental factors have provided the following model (Fig. 3). The fact that these results are so strongly differentiated on the background of the general model proves that the influence of azonal, local factors upon the thickness of the active

In this model, Spitsbergen takes a kind of intermediate position since the summer there lasts three months and the summer thawing depth is an average of 1 m — this being the thickness of the structures produced by frost action sorting which reflect the many-year action of the climate.

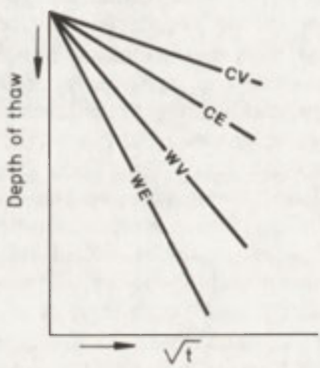

Fig. 2. Dependence of soil thawing on environmental factors such as summer temperature (C — cool, W — warm) and vegetation (V — vegetation cover, E — lack of vegetation cover)

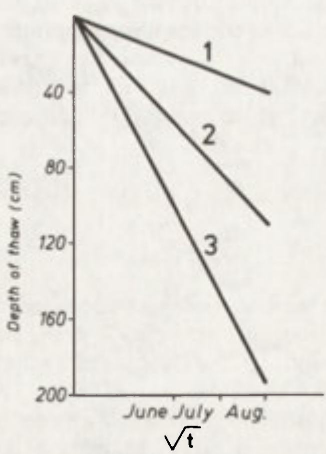

Fig. 3. Diagrammatic model of three-month soil thawing on Spitsbergen in different environments: 1 — muddy-silty soil with over 20 cm thick plant-peat cover, 2 — sandy soil with poor vegetation cover, 3 — sandy-gravelly soil (on raised beaches) without vegetation cover

permafrost layer equals or maybe even exceeds that of the zonal (global) factors. These findings which agree with Washburn's opinion (1979) may be of great significance to the interpretation of Pleistocene structures of the periglacial environment.

PERIGLACIAL SOIL STRUCTURES

Taking into account the climatic characteristics of those structures, I distinguish, two groups which are furthered in their development by: a – maritime, b – continental features of the climate (Fig. 1).

In the category of maritime forms and structures I include sorted circles and polygonal nets. These are a result of very intensive frost sorting, i.e. strong frost heave and segregation. The processes may develop in strongly water-saturated soil, in zones of heavy and more frequent precipitation and limited evaporation. Their classical forms are found on the coastal terraces at the outlets of the fiords (Isfiord Radio, Hornsund, Polish Research Station), but they also occur – though rarely – in the arid part of Spitsbergen, in the interior of the fiords (Longyearbyen). They can be found in the tundras as well as on surfaces devoid of vegetation, in places flooded with periodical lake water. There are also the familiar subaquatic forms. The vertical extension of the structures corresponds to the thickness of the active permafrost layer, being mostly 0.7–1.2 m. I have never noticed any forms deeper than that.

In this coastal zone of Spitsbergen there are ice wedges *in statu nascendi*. There also exist traces of older ice wedges which developed in the colder, postglacial climatic phase of that region. The former can be distinguished by the form of fresh ice veins, visible in spring, and the

Photo 1. Gelifluction cover (sorted stripes) in Hornsund fiord. Surface angle 7°

Photo 2. Displacement of wooden pegs in five-year period (1979–1984) on Jens Erikfjellet slope (Hornsund region). Concave down-slope curve of movement involved the entire active layer. Photo H. Chmal

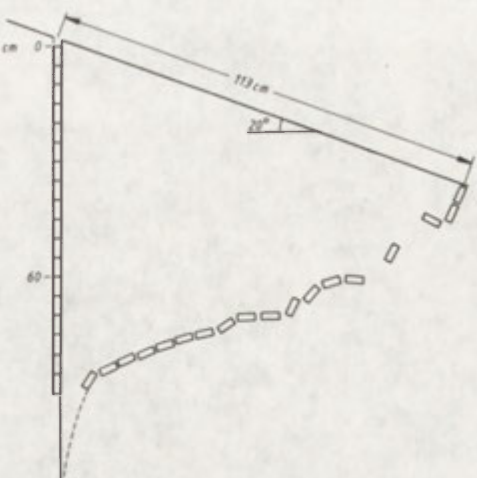

Fig. 4. Displacement of pegs in five-year period (1979–1984) on Gulliksenfjellet gelifluction slope (Hornsund region). Convex downslope curve of movement

latter by furrows overgrown with grass. They are polygonal nets of a diameter of several to several dozen meters – remnants of melted ice wedges.

Of dominant importance to the coastal zone is gelifluction. All slopes of an inclination smaller than 25° and all surfaces at the foot of slopes are smoothed out and made even by the soil masses flowing down (Photo 1). It is gelifluction of sorted stripes and steps with turf-and stone-banked lobes. In the years 1957–1960 I measured the rate of surface movement which was on the average 4 cm per year, with 12 cm being at its highest (Jahn 1961). In the five-year period of 1979–1984 we conducted new measurements of the movement by means of the peg method at the depth of the whole active permafrost layer. The investigations were carried out on slopes of angles from 8° to 29° in the vicinity of the Polish Research Station near Hornsund. In some of the profiles the mean annual rate reached unusually high values. With a slope angle of 20° the mean annual movement rate was 27 cm. The movement involved the whole layer of summer thawing down to a depth of 1.0 and 1.2 m. The downslope movement curve was convex (Photo 2) or concave (Fig. 4).

The examples given indicate that gelifluction is the most active and effective periglacial process on Spitsbergen. They also prove that in this region there exist both types of the hitherto known mass wasting movements of cold climates, i.e. frost creep (concave downslope) and mass flow (convex downslope curve). The first occurs as a movement of the particular soil particles ("free solifluction"), and the second as a mass flow resembling rather a landslide or earth flow. This fact deserves attention, as both types of structures are known from the European Pleistocene.

Now I am going to give a characteristic of those types of Spitsbergen periglacial structures in which I perceive an influence of continental climate. As I have already mentioned, these structures can be found in the more arid part of the archipelago, nearer the internal ice cap, in the interior of the fiords.

As most significant I regard the soil wedges discovered in the interior of the Van Mijen fiord on alluvial fans covering the low coastal terraces (Jahn 1983). The whole area is a typical "polar desert", almost devoid of vegetation and covered with fine-grained debris of dark

Photo 3. Exposed soil wedges on undercut slope of alluvial fan (coastal terrace) in Van Mijen fiord

Eocene shales originating from frost weathering on the slope (congelifraction — frost shattering and splitting) and subsequent transport onto the coastal terrace (Photo 3). The most striking surface features are the regular polygons of prevailingly pentagonal and hexagonal shapes. The soil wedges of a width from a dozen or so to several dozens of centimeters are on the borders of the polygons.

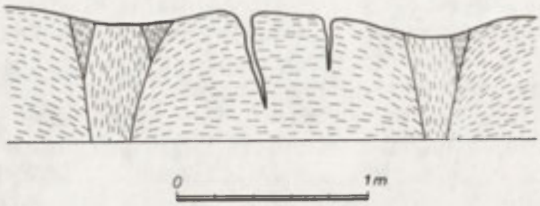

Fig. 5. Cross-section of polygonal net and adjoining soil wedges — on the alluvial fan in Van Mijen fiord, Spitsbergen

The arrangement of the particles gives an exact indication of the direction taken by the forces forming the soil structures (Fig. 5). No upturned structures produced by frost thrusting were noticed anywhere. The soil wedges of this area constitute a passive element developed from ice melting and not from frost heaving or thrusting. The regularity of the polygon patterns as well as the depth of the soil wedges leave no doubt as to their frost origin. Soil wedges are started by repeated frost cracking in the winter season. The cracks then trap debris slipping from the polygon surface into the furrow or accumulate it directly by wind action.

These soil wedges resemble "sand wedges" of the Antarctica. Thery are identical with the Siberian structures known as "ground wedges or veins." Analysing those structures I have

Photo 4. Overall view of slope with rhythmically bedded deposits in Van Mijen fiord

Photo 5. Outcrop revealing details of rhythmically bedded deposits in Van Mijen fiord. Length of metal stick – 120 cm (above ice-pick)

drawn the following inference: Structures of this type are known from areas of the periglacial Pleistocene zone of Europe and have generally been interpreted as evidence of a climate closer to that of Siberia than to that of Spitsbergen. The Van Mijen fiord wedges point to a possibly different climatic interpretation of Pleistocene wedge structures.

Another phenomenon related to the continental climate of the Spitsbergen interior are rhythmically bedded slope deposits of *grezes litees* type. They have also been found on the above mentioned, strongly weathered, dark Eocene shales in the Van Mijen fiord. The slope angle varies about 30° and the loose slope material slips down gravitationally (Photo 4 and 5). The whole process of movement and slip sorting takes place within the thawing permafrost layer, the depth of which being dependent on the slope inclination and amounting to 40–60 cm at the end of the summer (Jahn 1982). The smaller the inclination the thicker the active permafrost layer. The layer is divided into two parts (Fig. 6). The lower part is moist

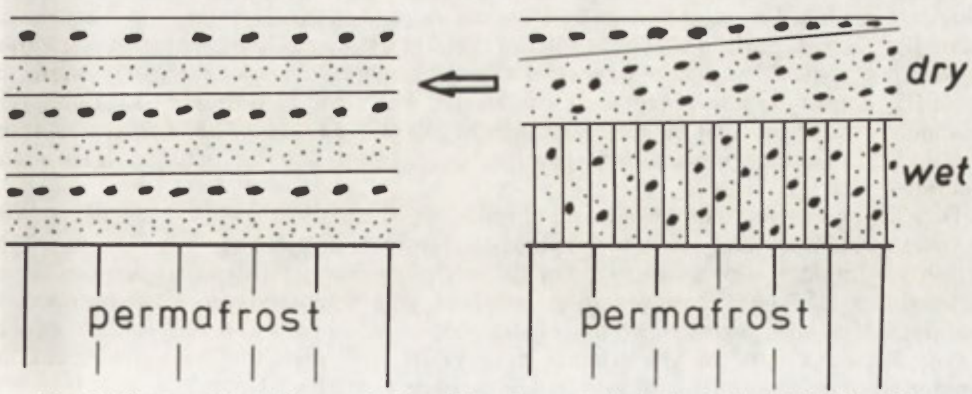

Fig. 6. Diagram illustrating gravitational development of rhythmically bedded deposits

and stable, the moisture being derived from permafrost. The upper part is well desiccated and loose. Its thickness amounts to 10–15 cm. It should be added that J. Büdel (1960) noticed this layer of summer desiccation on Spitsbergen (*Untergrenze hochsommerlicher Austrocknung*) with a thickness of 8–12 cm. When the gravitational stress exceeds the angle of internal friction, the desiccated layer begins to slip. In dry, fine weather the characteristic, subtle noise of the slipping debris can be heard. During the movement a selection of coarser elements takes place. This is how those rhythmically bedded, sedimentary structures, resembling *grezés litées*, are built. They develop in the gravitational part of the slope on the principle of dynamic selection, without the contribution of an additional agent. In the lower part of the slope the selected material encroaches upon the surface of snow patches. It is also washed down onto the surface of alluvial fans, retaining the same bedding rhythm. The essential cause of selection seem to be the gravitational movements of the loose debris in the full, polar summer season. The snow facies in the lower section of the slope and the water facies of the alluvial fans only reflect the slope processes mentioned. Thus great summer temperature variations, summer drought periods, i.e. elements of continental climate, further the development of rhythmically bedded deposits. It stands to reason however, that these sediments can evolve only where there are rocks that easily disintegrate from frost action – that is, congelifraction, such as shales or limestone producing fine debris. Such places are in the oceanic areas of the Pleistocene periglacial zone. The well-known occurrence of *grezés litées* (Guillien 1951) in France (Charente) should – in my opinion – be linked with the climatic summer conditions to be found at present in the interior of the Spitsbergen fiords.

The washing activity of the slope water on Spitsbergen has long been matter of attention. In the neighbourhood of the Polish Research Station at Hornsund, sediment traps were installed to assess the amount of material washed down the slopes in the course of the year (Jahn 1961). The experiment was only a partial success, because the region at the fiord outlet, with its maritime climate and compact tundra vegetation, was not a fortunate choice. The slopewash process is much more intensive in a barren continental region of the interior as e.g. in the Van Mijen fiord. It should be added that in the Pleistocene periglacial areas of Central Europe washing deposits area is a common feature.

Another phenomenon related to the Spitsbergen summer temperatures are the commonly occurring desiccation cracks. Special credit must be given to A. Washburn who found that "desiccation cracking is probably one of the most common and important patterned ground processes" (Washburn 1979). There is an essential difference between desiccation cracks and frost cracks. The former are tighter and more shallow and do not develop thrust structures. There exist, however, some points of similarity between those desiccation forms and the above described soil wedges, although the latter are much wider and deeper. To serve the purpose of the present analysis, it should be stressed that the problem of desiccation cracking has not been fully accounted for in the interpretation of Pleistocene structures. There exists a quite unfounded tendency to regard almost all Pleistocene wedge structures as frost phenomena. Yet on Spitsbergen, particularly in the interior areas of the islands, the number of desiccation cracks is not smaller than that of frost cracks. In the height of the polar summer, towards the end of July almost the whole loam surface is covered with small, polygonal desiccation forms. It cannot be ruled out that at least part of them follow the traces of older frost structures, nevertheless the summer impulse to desiccation does not seem to be less active than the winter impulse to freezing.

Desiccation processes are linked with chemical weathering and chemical sediments. With low relative humidity and intensive evaporation, brown or rust-brown crusts of desert-type varnish are formed on porous rocks. On the coastal terraces of Hornsund one can notice brick-red gravels which, however, form only the very external layer of the surface. In conditions of a high degree of aridity of the periglacial climate, iron compounds, easy to become displaced, settle on the external, evaporating rock surface. Those phenomena are common and well-known in the whole, contemporary periglacial zone.

Photo 6. Thermal (frost) cracking of block at Isfiord Radio

Photo 7. Thermal (frost) cracking of granite block in the gravels of Pleistocene (Late Vistulian) terrace, Karkonosze Mts., Poland

Another feature frequently encountered is the thermal weathering (contraction) of stones and blocks (Photo 6 and 7). These characteristic frost cracks in the form of three axes have a strong resemblance to frost cracking of the ground (polygonal system of wedges). Chemical and physical weathering phenomena like these, being regarded as indicators of the climate, are worth mentioning if only because of identical characteristic found in the Pleistocene deposits of Central Europe. J. Akerman (1980) considers cavernous weathering to be also typical of Spitsbergen.

On Spitsbergen there is evidence everywhere of intensive wind action (abrasion). It can be found in the maritime zone where winds are strong but the soil is protected by vegetation, as well as in the continental part where the winds are weaker but the soil and rocks are well exposed to their action. Thus there are deflation (stone) pavements as well as stones and rocks faceted and polished by the wind, sometimes they are typical ventifacts (Photo 8). As is well

Photo 8. Block from deflation surface (deflation pavement) of Isfiord Radio area. A case of "surface facets", according to J. Akerman (1980)

known, it is not only sand but even more so snow and ice particles that have an abrasive effect. With the ground being usually not covered with snow, the easterly and north-easterly winds prevalent in winter produce deflation on loose rocks and the polishing of solid rocks. However, the amount of ventifacts is not large and cannot be compared to what is found in the Pleistocene periglacial desert of Central Europe.

A separate problem concerning Spitsbergen are the niveo-eolian deposits (Photo 9). They are common everywhere in the area, both the mineral and the organic kind (Jahn 1961). Material that has settled on the snow in winter or has been deposited together with snow has a specific structure. The dust is blown out of the outwash plain, with dust-storms occurring particularly in the time between the warm and the cold season. As yet a good, comparative analysis is lacking of the contemporaneous, Spitsbergen and the fossil, Pleistocene niveoeolian deposits of Europe.

Photo 9. Niveo-eolian deposits on Griegfjellet slopes, Isfiord Radio

Photo 10. *Naledi* at Tempelfiord. Stage of vanishing ice cover with esker forms in the foreground

Another specific element to be found in the forefields of nearly all glaciers whose fronts end on land as well as in valley bottoms are ice-sheets *naledi* (icing, *Aufeis*). These forms have received the special attention of J. Akerman (1980) and S. Baranowski (1982). They suggested the use of the Russian name, i.e. *naled* in the singular and *naledi* in the plural number, as international terms. In most cases these forms are the effect of water circulating below or inside the permafrost. The outwash plains and wide, alluvial valley bottoms are covered with ice sheets of several to a dozen or so meters in spring. These sheets disappear towards the end of summer. In their place there remain numerous forms of subglacial ridges and esker-type cones (Photo 10) as well as crusts of chemical deposits, limestone and gypsum crusts, since the permafrost water depositing this material reflects the chemical composition of the rocks from which it comes. J. Akerman (1980) attaches great importance to this phenomenon as an indicator of climate when he writes: "The *naled* is dependent upon the climatic conditions during its entire lifespan and the effect of various climatological factors is different during different stages of development and disintegration." My personal opinion is that *naledi* may be a good indicator of the climatic conditions of the summer season — that is, of the period in which those ice sheets disappear as a result of melting, sublimation and evaporation. They are of lesser importance to the evaluation of the thermal conditions of winter, as there is no evident relation between the air temperature and subglacial hydrology or water circulation in deep permafrost layers.

Large forms of ground ice of pingo type are not characteristic of Spitsbergen, although they can sometimes be found there e.g. in the Advent fiord (Ahman 1973). It is rather those relatively small hydrolaccoliths that occur at the foot of slopes. In the tundras of the maritime part of the archipelago small frost mounds can be seen which resemble palsas.

Thermokarst topography is less significant on Spitsbergen than e.g. in Siberia. This process is typical of the high-continental varieties of the periglacial climate. Nevertheless melting pits (hollows) and first of all troughs, being remains of melted ice wedges, can be found there. In old, partly fossilized polygonal nets there are troughs which are partly scoured by meltwater action and filled with sand or mud. A certain analogy may be drawn between those forms and similar, Pleistocene structures (see Photo 16).

INTERPRETATION OF PERIGLACIAL FORMS AND STRUCTURES OF THE CENTRAL EUROPEAN AREA

The periglacial zone of the last Würm-Vistulian glaciation extends from Western Europe over France, Germany, Poland and the Soviet Union as far as to the Ural region. H. Poser (1948) was one of the first to distinguish the maritime from the continental part of this climatic landstrip. The areas to the north of the Alps and Carpathians (their Ib province) are to represent the intermediate glacial part. I expressed a similar view when I noticed in the Central Europe of the Pleistocene a "moderate periglacial climate" (Jahn 1975). I am still holding this view, although — taking into account the map of vegetation of the period of the last glaciation maximum by J. Büdel (1948), B. Frenzel (1967) and L. Eissmann (1981) — I am now slightly changing the borders of those provinces (Fig. 7). The southern border of this zone was marked by the slopes of the Alps and Carpathians as well as by their timberline. Between the limit of glaciation and the line of those mountains, the periglacial zone narrowed to a mere 500 km wide strip. The zone widened considerably to the west and east of this corridor. Close by the inland ice margin we can trace the tundra rubble zone which may be identified with the *Frostschuttzone* of J. Büdel (1948), resembling a "polar desert." From the south this zone was bordered by a loess zone, a dry tundra steppe, containing plant elements of a warmer climate. This transitional area of the periglacial climate of Europe stretched between the Elbe and Dneper rivers. To the west of the Elbe was the maritime, to the east of the Dneper the continental, periglacial province. In both of these areas the sharp borders of tundra, steppe

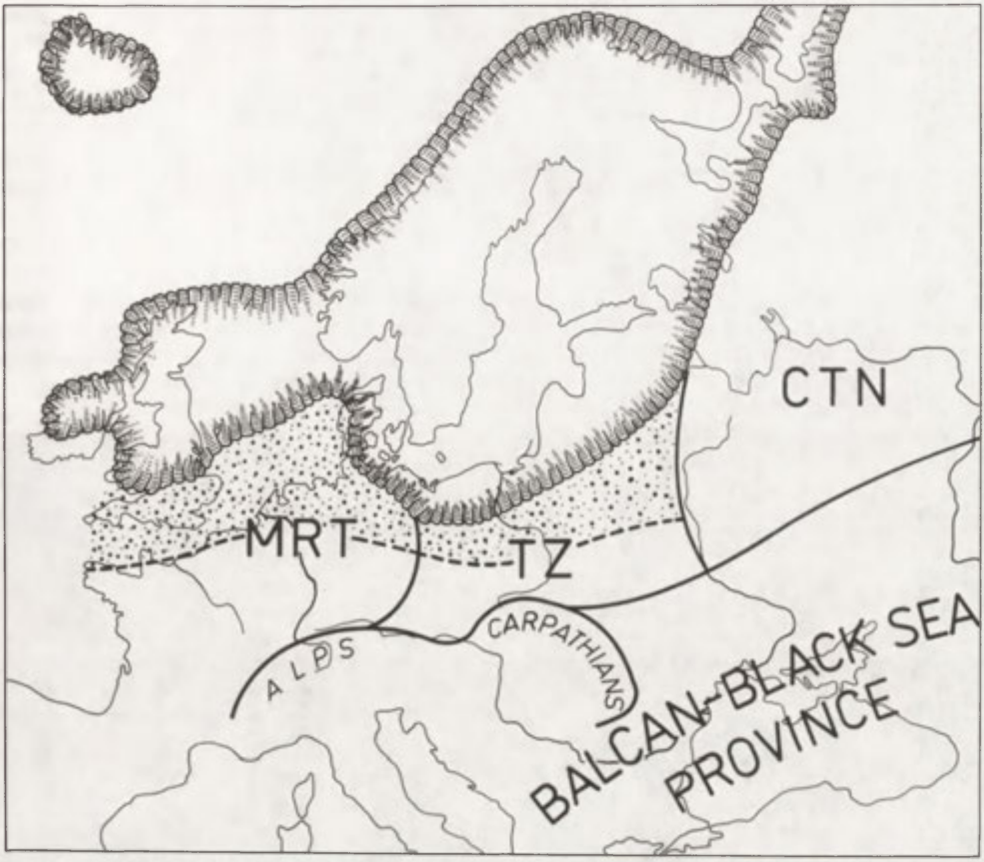

Fig. 7. Periglacial zone during maximum range of last glaciation (Vistulian) in Europe. Maritime part (MRT), continental part (CTN) and transitional part (TZ). Dotted stripe — frost debris zone

and forest were vanishing, especially in the continental part. This happened not only because the parallel mountain barrier (Alps, Carpathians) was lacking, but also because of the strong impact of the Atlantic ocean on one side and of the huge continent of Asia on the other. These effects were weakened by the Black Sea and the then much larger Caspian Sea and reached the Balkan Peninsula. This is why I call that climatic area, which is largely steppe, the Balkan-Black Sea Province.

The periglacial, transitional zone denoted on the map by the letters TZ is a matter of particular interest, especially the region of the present Vistula and Odra basin. This area is the subject of an excellent study by K. Kaiser (1960) in which an attempt is made to reconstruct the paleogeographic (climatic) conditions on the basis of detailed material. J. Dylik's synthesis (1956) concerns forms of degradation rather than soil structures. A reconstruction can thus be made of the active permafrost layer by basing on the structures of fossil ice and frost wedges as well as on frost segregation structures (Fig. 8, Photo 11). In the area of Poland the active permafrost layer was changing from 1 m in the north to 3 m in the south in the Sudetes Mts. forefield as well as in the east, in the Lublin Upland (Jahn 1977). The structures that have survived to this day make it possible to draw some cautious inferences. These structures have now taken on a modified form in comparison with their previous one, due to the general permafrost degradation in the postglacial period and the melting of the ground ice. I do not think it is at all possible to recognize in them the local effects of the periglacial environment, as

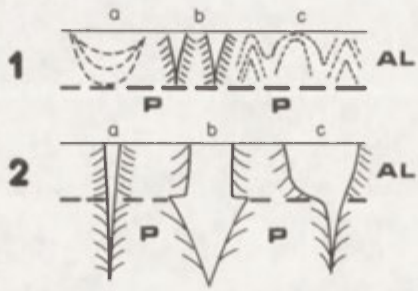

Fig. 8. Methods of determining the lower limit of active layer (AL) on horizontal surfaces: 1 — sorted structures (a), frost wedges (b), cryoturbate structures, involutions (c); 2 — fissures and deep wedges (a), ice-frost and soil wedges where soil part is narrower (b) or wider (c), AL — active layer, P — permafrost

Photo 11. Asymmetrical ice wedge in loess, permitting the reconstruction of the summer thawing depth of permafrost, Komarów, Lublin Upland

has been attempted in some regions of Germany or England. At present we have come to realize that many of those apparent frost involutions are not to be regarded as cryoturbation structures but as pseudofrost phenomena, as load casts. They have arisen in places where unconsolidated sediments of different densities and specific gravities were contiguous to one another, e.g. sand and silt. Under the influence of triggering impulses, such as tremors (tixotropy) or rapid temperature changes there may occur perturbations in the static equilibrium of the material and the development of involution structures which, however, have nothing in common with frost structures. Nevertheless it should be remembered that it is in the periglacial zone that triggering impulses take place both in flat areas and — even more so — on slopes, and that these occurrences are facilitated by the lack of stable equilibrium in the active permafrost layer. The impermeability of the permafrost surface is the reason why the lower, strongly water-saturated parts of the active layer produce a lack of equilibrium, especially in a wet autumn period. A general attempt at interpreting formations resembling frost structures in South-West Poland has been made by K. Brodzikowski and A.L. Van Loon (1983).

Gelifluction structures have most commonly been noted in the deposits of the last glaciation on Polish territory. As those on Spitsbergen, they are of two types – frost creep structures (concave, downslope) as well as structures of earth flow and landslide type. A classical profile of gelifluction found in the vicinity of Łódź (St. Małgorzata Hill) was described by J. Dylik (1963). These sediments, which he called "congelifluction", are of amorphic character, "lobate in shape." They may be compared to the Spitsbergen tongues, i.e. to the type of soil-rubble mass flow that has been demonstrated above with the example of the peg profile of the Hornsund region. In gelifluction sediments there occur, as a rule, wind-faceted stones (ventifacts).

Slopewash deposits are nearly as common as gelifluction. In all periglacial valleys the principal mass that fills them are stratified sand and silt washed down from the slopes. They are absolute proof of climatic continentality, to a much higher degree even than to what we now see on Spitsbergen.

Rhythmically bedded (stratified) sediments are common in the Pleistocene Upland regions of Central Poland (Dylik 1960). Their origin is for the most part related to water and snow action as well as to nivation processes and even to niveo-eolian deposits. The example taken from the Van Mijen fiord, where I described rhythmically bedded deposits developed from gravitational slip sorting, indicates a similar genesis of the deposits in the mountains of Central Europe. The flysch shales of the Carpathian Mts. created good conditions for frost-gravitational processes of the same type as those of Spitsbergen. In the Sudetes Mts. I saw such deposits on steep slopes built of marly-sandstone rocks (Photo 12).

Photo 12. Rhythmically bedded deposits on marl-sandstone substratum in Kudowa, Sudetes Mts.

Ice wedges, frost fissures and soil wedges have so far been the best determinant of the periglacial climate of Europe. They are evidence of the existence of permafrost, i.e. of a mean annual temperature of at least −2°C. The basis of the copious classification of the fossil. periglacial wedge structures is the principle of their infilling (cast). If an open frost crack has already been filled in winter with mineral material, i.e. a primary, seasonal infilling has taken place, then a soil wedge ("sand wedge") develops. If an ice vein (or wedge) existing in winter becomes filled – within the framework of the active layer – with mineral material in summer, i.e. a secondary, seasonal infilling takes place, then a partial soil wedge develops (Romanowski 1985). If, however, a deep ice wedge does not fill with mineral material until the climate has changed, i.e. after the ice has melted, then the ice wedge cast or the pseudomorphosis of the ice wedge is preserved. The first two cases are a sign of the aridity of the climate. The last example points to a humid, oceanic climate in which the thick vegetation cover keeps the frost fissure from becoming filled with silt, sand or debris.

The Central European region analysed here has been called the transitional sector of the periglacial zone (T Z). It possesses all the wedge structures mentioned earlier in this paper. J. Goździk (1973) is absolutely convinced that primary ice wedge casts, specifically eolian sand

Photo 13. Sand wedge in gravel terrace, Lubań, SW Poland. Photo J. Czerwiński

Photo 14. Fissure structure with secondary seasonal infilling. Generation of medium and small polygons at Hrubieszów, Lublin Upland. Photo J. Jersak (1975)

wedges, are a common feature in Central Poland, in the Łódź Upland. The examples of this kind of wedges we find in western (Photo 13) and eastern part of Poland (Photo 14). The examples from the Van Mijen fiord on Spitsbergen seem to confirm that such a situation was possible and that it might be linked with certain features of aridity of the Pleistocene climate of Central Europe. It can also assumed that this periglacial environment underwent local changes. The presence of dry and barren as well as of moist and fully overgrown areas does not necessarily depend on the zonal features of the climate only. On account of those soil and sand wedges we tend to believe that the mean, annual air temperature in the pleniglacial period of the last glaciation in Poland was $-6°C$.

Another example of the interrelated oceanic and continental periglacial structures on Polish territory is the famous loess profile in the village of Łopatka (Fig. 9) in the Lublin

Fig. 9. Loess structure in brickyard at Łopatki, Lublin Upland 1 — red loess, 2 — fossil soil, 3 — grey-yellow loess, 4 — solifluction loess, 5 — sand, 6 — loess deluvia, 7 — soil, a–e — ice wedge casts

Upland (Jahn 1956, 1975). In the loess of the last glaciation there are distinctly visible structures of soil wedges and gelifluction. These appear together, which shows that the changing climate furthered the development of both structures. With the developing profile and the accumulating loess, the initially cold, continental climate became milder. Four generations of wedges are proof of those climatic oscillations. The upper part of the loess is pure gelifluction having developed at a ground thawing depth of down to 3 m.

The profile gave rise to analyses of numerous other loess cross-sections in Poland and Germany. They confirmed the thesis of climatic oscillations in the last glaciation in the area of the "corridor" between the border of the ice cap and the Carpathian mountains (Jahn 1969; Jersak 1985). From early glaciation to Bölling there occurred warmer and more humid phases (gelifluction), followed by colder and drier periods, revealed by a whole generation of wedge structures. In order to get to know the environment of the paleogeographic Pleistocene a very efficient method of reconstructing periglacial structures has been applied. At least four or five of those climatic oscillations were recorded to have occurred in the time between 60 000 and 10 000 years BP. These changes therefore were in accordance with our thesis that the effects of maritime and continental climate were of alternating character. The last frost wedges (incluging even soil wedges) in the Lublin Upland are recorded to belong to the Younger Dryas (Maruszczak 1956), and those in northern Poland can quite certainly be attributed to the Older Dryas (Kozarski 1974). According to K. Kaiser (1960) the mean annual air temperature was by 11–12° lower in the Younger Dryas than it is at present. Fossil ice wedges and related phenomena are also to be found in Denmark, Holland and Sweden, which shows that there still climatic conditions favourable enough for the permafrost to retain even after the inland ice had withdrawn from Central Europe, from the moraines of the Pomeranian stage. Even at the time when – at the beginning of the Holocene – permafrost disappeared

Photo 15. Sorted structural soil (polygonal net) originating from the decline of the Pleistocene. On surface there are peat and bleached Holocene soil. Karkonosze Mts., 1400 m a.s.l.

from the whole Lowland area of Central Europe, its remnants survived in the mountains. In the Sudetes, on morphological planation surfaces at a height of 1400 m, a polygonal net was found which was covered with peat of an age determined as being 5000 years BP (Photo 15). Shortly before the postglacial, climatic optimum, permafrost existed in that place, and the depth of summer thawing reached not deeper than 1 m.

I am now coming to the problem of *naledi* – that is, ice sheets related to the freezing of water at valley bottoms and outwash plain, this water being of various origin but mostly derived from the depth of permafrost. The common occurrence of this phenomenon on Spitsbergen, which was also noticed by J. Akerman (1980), proves to be true. Of the abundance of Pleistocene periglacial forms in Europe the *naledi* structures have remained almost unnoticed. The reason seems to be rather methodological shortcomings that the actual state of the matter. Researchers dealing with the Pleistocene are too fascinated with glacial forms to notice similar forms (e.g. eskers) related to *naledi* ice. Yet they belong to the group of periglacial forms and as such they ought to be distinguished in the areas of the former polar desert and tundra of Europe. Their presence may be of considerable help in the reconstruction of the climatic conditions of the summer season.

Photo 16. Thermoerosional sand-filled trough developed from an ice-wedge, Mieścisko, Central Poland. Photo T. Krzemiński

As I have already mentioned, in spite of the classical investigations by A. Cailleux (1942) we have as yet not been able to form a clear concept of the significance of wind faceted stones as indicators of climate. There are more often ventifacts to be found in Europe than on Spitsbergen. In Europe the abrasive material was sand which was naturally abundant in a *pradolina* environment with its huge amount of river and fluvioglacial water. The abrasive action of snow and ice particles on Spitsbergen is certainly less effective than that of sand. What develops there nowadays are rather "surface facets" as J. Akerman (1980) calls them, i.e. "large facets developed upon rounded blocks normally of glacifluvial origin" and not "edge facets" which are typical ventifacts (*Dreikanter*). The different material of the substratum and not climatic reasons would then account for the differentiated shaping of eolized forms.

The Pleistocene thermokarst of Central Europe is well known, but it corresponds only partly to the thermokarst phenomena on Spitsbergen. In Poland there are very few remnants of melted pingo but a frequent occurrence are somewhat enigmatic furrows (Fig. 16) which I tend to regard as modified ice wedge furrows. Absolutely incomparable are thermoerosion phenomena little known of Spitsbergen, but constituting classical forms of *pradolina* in Europe. It is not the climate that determines the differences in the phenomena but the extension of the area as well as the inflow of large masses of warm water from the south in the European periglacial zone.

I would like to complete this brief comparison between the European periglacial structures of the last glaciation and the contemporaneous Spitsbergen structures with the following conclusions:

1. In both areas a tremendous variety of those structures can be found, which is evidence of the great spatial variability and temporary instability of the climate in either of the periglacial environments.

2. The variety of the periglacial forms and structures of Svalbard, including Spitsbergen, results from the following factors:

a) The archipelago is situated in a transitional zone between the interior of the Arctic and the Atlantic zone of influence of the Icelandic Low.

b) The climatic parameters of the interior (dry) differ from those of the coastal areas (humid) of the archipelago.

c) The local and therefore azonal environmental factors (soil, vegetation, moisture a.o.) exert at least the same influence on the differences in the structures of the active permafrost layer as the zonal climatic agents.

d) The variability of forms and structures takes its course in horizontal and vertical extension. Only the periglacial structures of the low, coastal terraces bear comparison with one another. The permafrost structures of mountain areas are different.

e) Not all the Spitsbergen periglacial structures visible at present reflect the prevailing climatic conditions. Part of them can be regarded as fossilized structures originating from colder, warmer drier or moister phases of the climate. Thus there are decayed structures as well as such in different stages of development. This fact has hitherto received very little attention.

f) The ice structures of Spitsbergen are of specific character. As a rule, large ground ice mounds, e.g. pingo, are lacking, while supra permafrost ice phenomena, such as ice phenomena, such as ice covers of *naledi* type are of great importance.

3. The variety of periglacial forms and structures of the last glaciation in Europe is a result of the zonal and azonal change of climatic conditions.

a) In the transitional zone of Central Europe ("corridor" T Z) there are structures of the periglacial maritime and continental climates. In this way they are of a character comparable with the periglacial structures of Spitsbergen.

b) It is difficult to precisely discern the dependence of those structures on the local, environmental factors.

c) The active permafrost layer can be reconstructed in general outline. Its thickness grows from the north towards the south, from the west to the east and reaches maximum values of up to 3 m. The summer of the Pleistocene periglacial zone is likey to have been warmer than the Spitsbergen summer. There exist certain reasons to assume that the summer season in the glaciation period lasted even up to 5 months. (On the basis of the diurnal amounts of varves in the varved clay of Sudetes Mts. — Schwarzbach 1940; Jahn 1976).

d) Periglacial structures provide numerous data useful to the understanding and identification of climatic changes having occurred during the glaciation period, and specifically the changes of the degree of oceanity and continentality of the climate. This result is of importance to the assessment of the stratigraphy of the Pleistocene deposits.

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SOIL SAND (SOILARENITE) AND RELATED COHESIVE DETRITAL DEPOSITS: EXAMPLES FROM A WET TEMPERATE CLIMATIC ZONE

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ABSTRACT. Detrital cohesive deposits belong to the most common present-day sediments known from various terrestrial environments, some shallow-marine ones as well as tidal flats. Lithologically, they include a very wide range of clastic materials varying in size from rudites (angular or rounded), through arenites, to lutaceous materials. Detrital cohesive deposits constitute a variety of landforms from small anthills to extensive floodplain sheets or prominent clay dunes that accompany some coastal zones. To the most common lithologic varietes belong clay rubble, clay gravel (pebbles, balls), soil pebbles (sand) clay sand (clayarenite) soil sand (soilarenite) and soil mud (soillutite). Genetically, the deposits may be classified as autoclastic, hydroclastic, anemoclastic, bioclastic or hybrid. All transitions may be encountered from pure cohesive deposits, through mixed ones, to ordinary (epi)clastic materials. Mixed deposits (e.g. the majority of intraformational conglomerates) have perhaps the best chance to be preserved in a fossil record and have been reported since the Late Pre-Cambrian till the Holocene.

> ...fine-grained sediments may become aggregated into particles of sand, gravel, and pebble dimension, following which the particles behave as sands and may be built into dunes.
>
> W. H. Twenhofel, 1950, p. 342

INTRODUCTION

In the geological sciences, terms such as "clastic or detrital deposits" refer traditionally to materials composed of fragments or grains derived from older lithified or solid rocks. Such grains are more or less hard and abrasion-resistant. Both the grains and their matrix-free accumulations are cohesionless. Such materials constitute the bulk of detrital sedimentary deposits, and extensive literature has covered all the geologic aspects concerning them, including origin, terminology, and classification (cf. Grabau 1904, 1913, 1932; Norton 1917, 1920; Tyrrell 1921; Wentworth 1922; Reynolds 1928; Krumbein and Pettijohn 1938; Shrock 1948a, b; Lombard 1949; Rodgers 1950; Folk 1954, 1968; Shepard 1954; Pettijohn 1957, 1975; Crook 1960; Boswell 1961; Klein 1963; Dott 1964; Williams 1966; Picard 1971; Pettijohn et al., 1973; Perriaux 1974; and Blatt et al., 1980).

On the other hand, wet soils and clays behave as a plastic or quasi-plastic mass, but their dried varieties may behave as a rigid elastic solid displaying fractures along shear and tension zones and may be easily disintegrated into a granular material. Upon rewetting, such materials not necessarily break down at one and may be facilely transported as bed-load and/or saltation load being subjected to abrasion and hydraulic sorting or another type of sorting (e.g., owing to Bagnold dispersive pressure). Finally, they may be accumulated as distinct individual deposits composed of grains that are actually or potentially cohesive. The properties of these deposits are changeable and reversible with all possible transitions from a quasi-solid state when dry to a plastic one on rewetting, and *vice versa.*

Despite of a long history of investigations into such deposits, known at least since the end of the 19th century, they lack well-established terminology and classification. This article is an attempt to summarize the actual state of our knowledge concerning the cohesive detrital deposits and to elaborate a more comprehensive terminology and classification. Examples of cohesive detrital deposits described and illustrated in this paper come from southern Poland including the Sudeten Mts. and their Foreland as well as central Poland (some fossil cohesive deposits collected from the Bełchatów Browncoal Mine). All the present-day deposits are characteristic of a wet temperate climate and some of them at least have been described for the first time from this climatic zone (Teisseyre 1990a, 1991, in press).

CLASSIFICATION OF COHESIVE DETRITAL DEPOSITS

In classifying detrital (clastic) sediments and rocks one must take into account the most fundamental principles of sedimentary rocks classification (see references listed above). These principles may be summarized as follows.

1. A detrital (clastic) sediment or rock comprises fragments of pre-existing materials called clasts or grains. With few exceptions, these owe their origin to weathering and erosion of a source rock or material, transportation of the weathering products, and deposition either under the direct influence of gravity (e.g., talus or scree cones) or by gravity-controlled mechanical agencies (wind, running water, moving ice, mass wasting). Some local accumulations of detrital material are the products of the activity of plants, animals, and man. Many clay-rich materials become a source of new sediment as a result of the process of "soil ripening" (Pons and Zonneveld 1965; better: "clay ripening") that involves some irreversible modifications taking part in a water-laid cohesive deposit when subjected to the air for the first time.

2. Under favourable conditions, any material may yield clasts or grains irrespective of its mineral and/or chemical composition. Such grains may accumulate to form a "monomineralic" or "ologomictic" deposits or be mingled with another clasts to constitute "polymictic" or hybrid sediments. Clasts derived from semi-consolidated muddy or limy deposits of the same formation, i.e., those which are intrabasinal and (pene) contemporaneous in origin, are spoken of as "intraclasts" (Folk 1959, 1968; Pettijohn 1975). Recently, the term is applicable to all clasts caused by intrabasinal erosion, irrespective of their lithology. Thus, sandstone slabs may also act as intraclasts, though rarely.

3. Modern classifications of detrital sedimentary rocks are based primarily on grain size (Wentworth, 1922, grade-scale is preferred in geology, see also Shepard 1954) and, secondarily, on mineral composition (Folk 1954, 1968; Pettijohn 1957, 1975; Pettijohn et al., 1973). It is generally accepted that detrital deposits with grain-size over 2 mm be termed rubble (detritus, debris, rock waste, grus, clastogene) if angular, or gravel if rounded (lithified counterparts: breccia and conglomerate); those of medium-texture and grain-size from 1/16 to 2 mm be named sand (sandstone), and those with fine-texture be called silt (siltstone) if composed of grains 1/256 to 1/16 mm in diameter, or less, respectively. Mud is a mixture of silt and clay (eventually plus water). This classification is independent of mineral and/or chemical

composition of grains. However, if the mineral composition of a particular deposit departs aubstantially from an ordinary silicic one, the fact should be emphasized by an appropriate adjectival connotation e.g., olivine sand, monazite sand, black sand, calcarenite (Grabau 1904), mud-pebble conglomerate, oolite sand, calcareous sand (Illing 1954), calcite sand, mudstone conglomerate and so on (cf. Pettijohn 1957, 1975; Pettijohn et al. 1973).

Consequently, following these fundamental principles, a sandy deposit composed of sand-sized soil aggregates or peds, eroded on the hillslope or floodplain, transported and redeposited in any positon and form is simply *soil sand* (Teisseyre 1990a, 1991, in press; Photos 1, 2). Similarly, terms such as *soil gravel* (soil pebbles, soil balls, cf. Bell 1940), *soil mud*, *clay sand*, *clay gravel*, *clay rubble* and so forth may be created and applied, if necessary. Soil mud is composed of soil aggregates and products of their abrasion that are less than 1/16 mm in diameter, but which are still aggregates of finer individual particles. Such grains are different in origin and character from ordinary silt and clay (cf. Price and Kornicker 1961; Pryor and Vanwie 1971).

In order to develop a more detailed classification of these deposits, a modified Grabau (1904, 1913) taxonomy is adapted here. In his classification, Grabau included several prefixes to denote the bulk composition of detrital rocks, namely: silic-, calc-, and argil-. However, the prefix "argil-" refers to all ordinary fine-grained sediments including lutites and pelites. Consequently, this prefix cannot be used to create proper terms for soil sand, soil gravel, clay sand and related cohesive detrital deposits. Thus, I suggest that the latter should be simply spoken of as *soilrudite*, *soilarenite*, *soillutite*, *clayrudite*, and *clayarenite*, because the terms "rudite", "arenite", and "lutite" are size notions devoid of any compositional denotation. From the genetic point of view they may be subdivided into the following subclasses.

1. Autoclasts (e.g., autoclayrudite or autoclayarenite originated owing to physical weathering of clays, including salt weathering, "soil ripening", exfoliation, and frost weathering);
2. Hydroclasts (e.g., hydrosoilarenite, hydrosoillutite, hydrosoilrudite, hydroclayarenite etc., developed by the activity of running water on a slope, within a channel, or over a floodplain);
3. Anemoclasts (e.g., anemosoilarenite, anemosoilrudite, or anemoclayarenite produced by niveo-eolian processes);
4. Bioclasts (e.g., biosoilrudite or biosoilarenite generated by the activity of plants or soil-living animals) and
5. Hybrid detrital cohesive deposits being transitional between the above-listed categories or between them and ordinary clastic or biochemical sediments.
6. To this list one should add cataclastic cohesive rocks (e.g., cataclayrudites or clay breccias originated due to tectonic movements and till breccias attributable to the glacial activity).

Following the East- and Central European usage, one may suggest another genetic terms including:

1. Deluvial cohesive deposits corresponding to cohesive slopewash or deluvial sediments (e.g., deluvial soilarenite, deluvial soilrudite, deluvial clayarenite and so forth; cf. Teisseyre, in press);
2. Alluvial cohesive deposits (e.g., alluvial soilarenite, alluvial soilrudite, alluvial clayrudite etc.);
3. Niveo-eolian cohesive deposits (e.g., niveo-eolian soilarenite, niveo-eolian soillutite, niveo-eolian clayarenite etc.).

However, for the field practice simple half-descriptive terms are indinspensable involving such notions as: soil sand, clay sand, soil gravel (soil pebbles) and so on. In fact, these terms may be thought to be abbreviations of the full triple connotations such as: soil-aggregate sand, clay-aggregate sand, soil-pebble gravel, and so forth. The term "pellet", sometimes used to designate soil aggregates (peds), is not favoured, because it refers traditionally to some components of chemical or biochemical rocks (cf. James 1954) or faecal pellets (cf. Moore 1939).

Pryor and Vanwie (1971) described peculiar sands composed of cohesive sand-sized aggregates, the so-called Eocene "Sawdust Sand" of the Wilcox Group of Tennessee and Kentucky. However, these aggregates are more likely of floccule origin and are distinct from the detrital cohesive grains. According to the Grabau's classification, the Sawdust Sand falls into the group of endogenic deposits and may be termed hydrosilicgranulite.

It is obvious that any soil sand or clay sand may grade imperceptibly into an ordinary siliceous one. Thus there is the need for sharpening the end-member definitions. Soil sand or clay sand is an arenaceous deposit containing, it is suggested, 75% by volume or more of cohesive grains composed of soil or clay. Mixed sand contains 15–75% grains of soil or clay and (ordinary) sand (arenite, orthoarenite) comprises 0 to 15 per cent grains of cohesive material (Fig. 1).

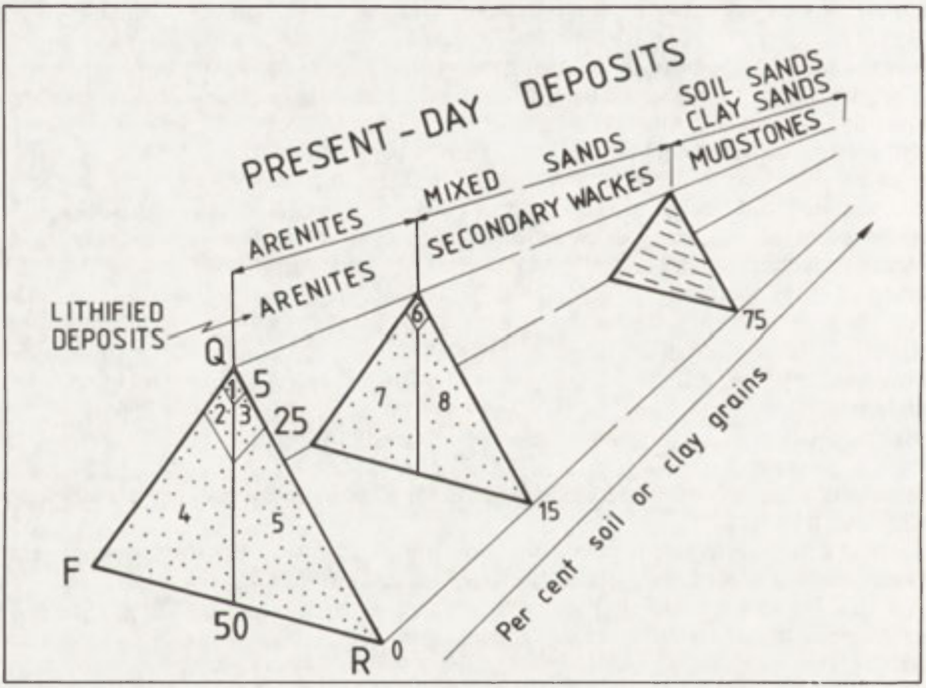

Fig. 1. Classification of cohesive detrital deposits (modified from Dott 1964): Q — quartz, F — feldspar, R — rock fragments, micas etc. 1 — quartz arenite, 2 — subarkose, 3 — sublitharenite, 4 — arkose, 5 — lithic arenite (subgreywacke), 6 — quartzwacke, 7 — feldspatic wacke, 8 — lithic wacke. See text for more details

It may be intuitively assumed that diagenetic processes will tend to obliterate the detrital character of the cohesive deposits. Owing to the process of welding, detrital grains of soil, clay, or mud merge with one another losing their identity; the result will be a sort of secondary matrix or mudstone. Consequently, diagenetic alteration of an originally clayey or soil sand will lead to the formation of sandy or gritty mudstone; secondary wacke, or greywacke; and (ortho)arenite or secondary subgreywacke (lithic arenite) depending on the original content of detrital cohesive material (Fig. 1). Such secondary wackes, greywackes, or subgreywackes may be hardly distinguishable from primary ones, if ever (cf. Teller 1972). It is also very likely that some diamictites (Flint et al. 1960), interpreted commonly as mudflow deposits, might have

TABLE 1. Grain size analyses of cohesive detrital deposits (gentle hand sieving)

Parameter		Average genetic types		
		autoclastic clayarenite	alluvial soilarenite	slopewash (deluvial) soilarenite
Grain size (cumulated percentage)				
−4 ø	16 mm	0.2	0.9	0.3
−3 ø	8 mm	2.5	5.1	1.4
−2 ø	4 mm	15.9	19.0	5.2
−1 ø	2 mm	46.2	46.8	18.8
0 ø	1 mm	74.3	71.3	48.2
1 ø	0.5 mm	88.1	85.2	76.8
2 ø	0.25 mm	92.7	93.9	92.1
3 ø	0.125 mm	94.8	96.4	95.7
4 ø	0.0625 mm	96.1	97.2	97.4
Per cent fines (by weight)		3.9	2.8	2.6
Statistical parameters (range)				
	Md_{ϕ}	−2.90 ÷ +0.85	−1.15 ÷ +0.55	−2.40 ÷ +2.30
	M_z	−2.82 ÷ +1.50	−1.05 ÷ +0.58	−2.37 ÷ +2.17
	σ_I	1.10 ÷ 1.79	1.41 ÷ 1.72	1.05 ÷ 1.45
	Sk_I	+0.20 ÷ +0.76	−0.13 ÷ +0.23	−0.10 ÷ +0.11
	K_G	1.27 ÷ 1.79	0.09 ÷ 2.0	1.16 ÷ 2.20
Number of analyses		18	15	37

Note: Autoclastic deposits collected from Biały Kościół, Gębice, Henryków and Ziębice study reaches (Holocene floodplain clays, Miocene clays); alluvial soilarenites sampled in Biały Kościół, Gębice, Kazanów, Wadochowice, Brukalice, and Henryków study reaches (present-day alluvia of the River Oława); and slopewash (deluvial) soilarenites taken from hillslopes of Wadochowice, Brukalice, and Henryków experimental fields (all locations situated within the upper River Oława catchment basin, Sudeten Foreland, southwest Poland; samples collected in a period 1984–1990).

Statistical parameters according to Folk and Ward (1957): Md_{ϕ} — median grain diameter; M_z — mean grain size, σ_I — standard deviation; Sk_I — skewness, and K_G — kurtosis.

originated as mixed soil gravel (soilrudite) or mixed clay gravel (clay-rudite). Thus, a wide occurrence of the present-day cohesive detrital deposits sheds new light on the problem of the origin of matrix in terrestrial paraarenites and paraconglomerates at least (cf Pettijohn 1957, 1975; Flint et al., 1960; Dott 1964; Harland et al., 1966; Whetten and Hawkins 1970).

McCrone (1978, p. 493) stressed that: "Some 'mudstones' and 'claystones' are actually breccias and conglomerates composed of fragments of mudstone, claystone, or shale in a matrix of mud and clay". Such rocks, the origin of which is usually debatable, may be lithified and partly altered clayrudites or soilrudites or may be of tectonic origin, as the so-called "Black Cube Clays" known from the Miocene browncoal formations of central Poland.

The results of grain-size analysis of 70 samples of present-day cohesive detrital deposits from the upper River Olawa catchment, Sudetes Foreland, are summarized in Table 1 (note small amount of fines in the deposits investigated).

SOILS AS A SOURCE OF DETRITAL COHESIVE MATERIALS

Any soil may be a source of detrital material (Twenhofel 1950). However, clay-rich soils distinguish themselves in having peddy textures (Photos 3, 4) and *ipso facto* are granular discontinuous media (Pons and Zonneveld 1965; DeVos and Virgo 1969; Ollier 1969; Fitzpatrick 1983, 1984). In fact, textural characteristics of such soils changes with depth within the zone of aeration including a crumb-, platey-, blocky-, and prismatic-textured layer (Fig. 2). Thus, it is obvious that upon erosion such soils will break down into cohesive granular materials ranging in size from fine sand, through granules, to pebble gravel. Columnar jointing, found in many such soils (Teisseyre 1980, 1984, 1990a), gives rise even to the formation of quite large blocks of soil, particularly along river cutbanks or exposure walls (so-called "vertisols"). Additionally, ploughing of frozen soils or deeply dried ones may result in the development of large plates and soil clods.

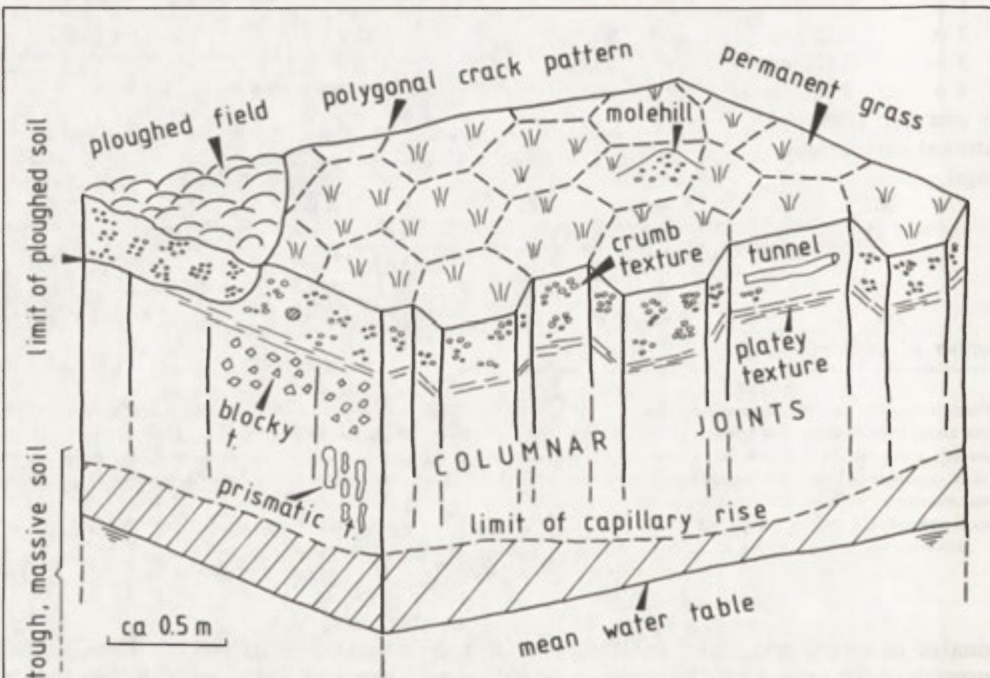

Fig. 2. Textures of clay-rich soil. Modified from Teisseyre 1990a

The tendency to formation of soil peds (i.e., crumbs, block, etc.) is enhanced if the soil contains montmorillonite, smectite, or related expanding clays; organic matter and calcium carbonate are also of importance. Kaolinite, on the other hand, seems to generate weaker peds.

As it has been emphasized by many authors, peds may be transported by running water or wind and then accumulated as distinct, granular cohesive deposits. Nanson et al., (1986, 1988) and Rust and Nanson (1986, 1989) were perhaps the first to recognize such deposits in the fluvial environment. Recently, Teisseyre (1990a, 1991, in press) described analogous deposits from wet temperate climate giving them terms "soil sand" and "soil gravel", depending on their mean grain size. In fact, slopes underlain by crumb-textured soils are the most widespread source of redeposited detrital soil materials (Alberts et al., 1980; Bryan and De Ploey 1983; De Ploey and Poesen 1985). Such deposits were recently described by Teisseyre (in press) as deluvial soil sands and gravels.

AUTOCLASTIC COHESIVE DEPOSITS

SOILRUDITES AND CLAYRUDITES

Autoclastic cohesive deposits were the first to be recognized, though they were described originally under a variety of terms quite distinct from the genetic ones suggested by Grabau (1904, 1913, 1932). Among them, lithified rudaceous deposits were the first clearly distinguished at the end of the 19th century; these were described as intraformational conglomerates and breccias or sharpstone conglomerates (Walcott 1894; Mansfield 1907; Hyde 1908; Field 1916; Foerste 1917; Norton 1917, 1920; Shrock 1948a, b; McKee 1954; Pettijohn 1957, 1975; Williams 1966; see also Barrell 1906).

Among intraformational conglomerates and breccias only some may be classified as autoclastic (cf. Barrell 1906; Mansfield 1907; Hyde 1908; Field 1916; Shrock 1948a), while others are hydroclastic-, anemoclastic-, or even bioclastic in origin. Those varieties that are evidently autoclastic in derivation have been described from ancient formations including fluvial-, lacustrine-, desert-, deltaic-, tidal fiat, and shallow marine sediments. Similar present-day deposits have been found in analogous sediments (cf Haas 1927; Longwell 1928; Trusheim 1929; Burt 1930; Grabau 1932; Young 1935; Shrock 1948a; Price and Kornicker 1961; Karcz 1969; Glennie 1970; Pryor and Vanwie 1971; Bigarella 1972; Cooke and Warren 1973; Tucker 1974; Jahn 1975; Scott 1978; Teisseyre 1979, 1980, 1984, 1986, 1991, in press; Rachocki 1981; Zachar 1982 and many others). Recently, Teisseyre (in press, in preparation) has found them among present-day slopewash (deluvial) deposits.

Synonymous terms being in a common usage include: clay-pebble conglomerate and breccia, flat-pebble conglomerate and breccia, shale (slate)-pebble conglomerate and breccia (Williams 1966), clay-pellet conglomerate (Allen and Nichols 1945; Weaver 1959), mudstone conglomerate (Pettijohn 1957), and mud-pebble conglomerate (McKee 1954). The origin of some clay-pellet conglomerates is unknown or uncertain, however. Similar textures described by the author from the Lower Carboniferous tonsteins of the Intra-Sudetian Basin, Sudetes Mts., appear to be early diagenetic in origin (Teisseyre 1970a, b, 1971). Good examples of clay-pebble gravels are known from the Neogene terrestrial deposits of the Sudetes Foreland (Photo 10) and central Poland (Bełchatów Browncoal Mine).

Autoclastic varieties of intraformational conglomerates and breccias correspond primarily to indurated autoclastic clay- or mud rubble, composed of clay or mud fragments embedded in a muddy-, clayey-, sandy-, or limy matrix. Important are also intraformational conglomerates and breccias composed of limestone fragments, while those containing sandstone slabs seem to be less frequent; these will not be considered here, however. An intraformational conglomerate and breccia may be classified as an autoclastic deposit if there is conclusive evidence of its formation *in situ*, with but a minor displacement of the original fragments. The fragmentation and rearrangement of the original fragments may be thought to be syndepositional or penesyndepositional (penecontemporaneous).

There are at least three main processes that may be responsible for the production of the original fragments of cohesive materials. These are the following.

1. Soil ripening (Pons and Zonneveld 1965) and sediment ripening, which includes some irreversible processes taking part in a soil or newly deposited sediment exposed to the air for the first time.

2. Ripping, which depends on tearing off clay fragments from cohesive deposits overlying slightly cohesive fine sands and muds susceptible to liquefaction and intrastratal flow. Shale

fragments of such an origin have been termed rip-down clasts by Chough and Chun (1988; cf. also McKee and Goldberg 1969). However, following much earlier suggestions by Shrock (1948a, p. 208–209 and 176–277), it may be argued that intrastratal flow may also generate rip-up clasts, i.e., those torn from underlying clays or limestones. Fairbridge (1946) ascribed rip-up clasts to subaqueous slides.

3. Physical weathering of newly exposed, originally tough and compact clays, muds, and soils (Photos 5, 6, 8). Such parent materials (with the exception of soils) may represent a wide series of formations ranging in age from Holocene till Late Pre-Cambrian. The mechanism of fragmentation involves several distinct processes including thermal and volume shocks offered by repeated wetting and drying as well as freezing and thawing (Williams 1967; Harrison 1970; Jahn 1975; Van Vliet-Lanoë and Coutard 1984; Van Vliet-Lanoë 1985), and the mechanical activity of ground ice, needle ice, salt crystals, plant roots, or a combination thereof. These processes may be operative anywhere, but are particularly effective along river cutbanks, on walls of exposures, bluffs, cliffs, and related rather steep surfaces that as a rule are poor in vegetative covering. However, it should be stressed that the processes affect also vegetated soils and grounds, both inclined and level, to a depth limited by the extent of desiccation and/or winter freezing (in southern Poland usually 0.3–0.6 m except of high mountains). In general, desiccation and shrinkage processes are limited by a mean position of the ground-water table (here usually 1–3 m). In particular, repeated drying and wetting may affect the whole zone of aeration leading to the formation of columnar jointing (Teisseyre 1984, 1990a; Fig. 2). It is the zone of capillarity which is transitional between the upper portion of the weathering mantle, that is fragmented, and its lower part, which is tough and more or less compact.

This fragmental layer of the weathering mantle is susceptible to gravity movements (mass wasting) which affect primarily steep unprotected surfaces. Earth falls accompanied by sliding and creep give rise to the formation of talus (micro)cones at the base of such scarps (Photo 7). These commonly coalesce to form more or less continuous aprons. The deposits are composed of cohesive rubble or detritus (debris) comprising angular fragments and chips of dry clay, mud, or muddy sediment, while soil fragments may be either angular or roundish (the latter feature is not the result of abrasion, but inherited from the original soil texture). Such clay or soil rubble may reveal relatively a good sorting with crude form-concordant lamination and the coarsest fragments concentrated at the base of the talus cones. Fine-grained autoclastic clayrudites composed of chips of granule-size have been also found to form owing to frost weathering of an originally water-saturated tough clay (Photos 5, 6) or due to salt weathering of brine-saturated clay (Price and Kornicker 1961).

Intraformational conglomerates resulted from the three above listed processes differ from one another in the origin and character of matrix. With deposits attributable to sediment ripening and physical weathering it is not uncommon that the matrix is a secondary component and, in so being, is not a diagnostic feature. On the other hand, ripping leads to mixing of several contemporaneous deposits, part of which at least are in a semi-consolidated condition. In such a case, the matrix is a primary component and is important in reconstructing the origin of the deposits investigated.

Autoclastic soilrudites develop commonly owing to disintegration of a blocky-, or prismatic-textured clay-rich soil or due to frost weathering of tough, compact soil (Photo 8). The latter case is a common phenomenon on ploughed fields not protected by snow drift.

Summarizing, one can easily demonstrate that clay (or shale)-pebble conglomerate may originate not only at the expense of (pene)contemporaneous cohesive deposits, but also owing to physical weathering of any pre-existing cohesive material. In the former case, an intraformational conglomerate or breccia will develop, while in the latter – a true *interformational* one. In fact, some basal sharpstone conglomerates (weathering breccias) did originate as interformational cohesive clayrudites or mudrudites.

SOILARENITES AND CLAYARENITES

Under continental conditions, these materials may originate anywhere as a result of physical weathering of either pre-existing or contemporaneous cohesive deposits and/or soils. They accumulate both on talus cones and at the surface of a level, clay-rich ground, both vegetated and bare. The deposits have been rarily described as yet.

HYDROCLASTIC COHESIVE DEPOSITS

Actually, these deposits are rather common including shallow marine environments, tidal plains, lake-, delta-, and fluvial milieu involving both channel- and extra-channel environments. Also, many slope-wash deposits belong to this category. Moreover, analogous sediments have been reported from deep-marine turbidite sequences; these, however, will not be considered here. Not all detrital cohesive materials described below are strictly hydroclastic in origin, though all of them were transported and deposited by running water.

CLAYRUDITES AND SOILRUDITES

To this category belong the well-known "edgewise" conglomerates and breccias. This is a variety of intraformational conglomerate composed of transported and imbricated mud, clay, or limestone crusts, known from many shallow marine and terrestrial formations since the Pre-Cambrian (Photos 9, 10; cf. Shrock 1948a; Lindholm 1980; Sepkoski 1982).

Another variety may be mud-pebble or mud-crust rubble (breccia) composed of platey, circular to polygonal, or irregular fragments of partly dried to dried mud, transported as floating load by streams or lake currents. A good example of such deposits is known to the author from Turawskie Lake (a reservoir on the River Mala Panew, Silesia Lowland northeast of Opole, southern Poland). Here, owing to large fluctuations in water-level, extensive mud flats developed within an avant-reservoir as well as on the lake delta are subjected periodically to desiccation and cracking. A subsequent rise in water level causes some of the partly dried mud polygons to float being redistributed either by flows operative within distributary channels or by alongshore lake currents. These floating mud crusts may even be swept into the central part of the lake and finally deposited within the littoral zone as a peculiar deposit composed of littoral sand mixed with apparently allochtonic mud fragments. In Turawskie Lake, mud fragments float mostly owing to numerous gas bubbles giving to the mud a bread-like appearance. However, convex-up mud crusts may float over some distance even if gas content in the deposit is low. Similar processes of floating were reported by Twenhofel (1950) and Fagerstrom (1967) from shallow marine environments.

Teisseyre (1975) described floating transport of turf sheets up to 4 m^2 in area, occasioned by a heavy flood in July 1971 in the upper River Bóbr drainage basin, Central Sudetes.

Hydroclastic clay-pebble gravels and conglomerates originate commonly as a product of fluvial erosion of cohesive bank- and bed materials. Examples of present-day deposits attributable to the failure of cohesive bank material were described, among others, by Bell (1940), Baluk and Radwański (1962), Głazek and Radwański (1962), Nordin and Curtis (1962), Nordin (1964), McKee et al., (1967), Jońca (1968, 1981), Karcz (1969), Laury (1971), Klimek (1974a, b), Teisseyre (1975, 1980, 1984, 1986, 1988, 1989), Florek (1978), Rachocki (1981) and Krzysztoń (1984).

It is long known that fragments of wet mud are relatively resistant (Twenhofel 1950) and may be transported over quite a large distance, particularly if armoured with gravel (Bell 1940). Smith (1972) confirmed the opinion experimentally demonstrating that fragments of wet mud may be transported by running water as bed load over a distance up to several tens or even several hundreds of metres. Field observations by the present author on the River Oława at Biały Kościół, Sudetes Foreland, indicated that clay pebbles originated owing to massive bank failure of wet Holocene clays were transported as bed load on a distance up to 1 km without being destroyed. These pebbles were accumulated, in part, as peculiar lateral bars composed of clay pebbles and balls with some admixture of silicic and quartzose pebbles. Earlier still, Bell (1940) reported fluvial transport of armoured mud balls ("pudding balls") on a distance up to 4–5 km least.

It has been found that clay pebbles deposited as a permanently waterlogged sediment are stable features, whereas those exposed to the air on bars' surfaces underwent soon destruction owing to physical weathering including, in particular, repeated wetting and drying as well as freezing and thawing (Teisseyre 1979, 1980, 1984). In the Sudetian gravel-bed rivers, large mud balls (up to 0.5 m in diameter) are easily transported thanks to low specific gravity of mud or loam, even by sheet flows that are less deep than the balls, diameter (Teisseyre 1986).

Analogous, usually flattened bits of clay are known from many present-day and fossil deposits under a variety of names including: Tongallen (Trusheim 1929), clay-galls Burt 1930), clay flakes, clay chips, and clay shavings. Clay-, mud-, and till fragments are also reported from many Pleistocene glaciofluvial and glaciolacustrine deposits; those composed of till are spoken of as till balls (Photo 11; Leney and Leney 1957).

Lime-mud balls ("sea balls") were reported from marine deposits (Croneis and Grubbs 1939). Another examples of mud balls (armoured or not), clay and peat pebbles and boulders as well as limestone intraclasts were described from shallow marine deposits by Shrock (1948a), Twenhofel (1950), Van Straaten (1954), Trefethen and Dow (1960), Nessin (1961), Tanner (1961), Stanley (1969), Lindholm (1980), Sepkoski (1982), and Whisonant (1987). Clay pebbles (in part armoured) and peat boulders are known to the author from the Baltic coastal zone as well as from coasts of many reservoirs situated in south and central Poland (cf. Dickas and Lunking, 1968, for another examples).

Hydroclastic intraformational conglomerates and breccias may also originate owing to the process of ripping. In this case, this depends on tearing off clay fragments from a cohesive bottom sediment by powerful currents (Black et al. 1980; Mutti 1981). The fragments of such a derivation have been termed "rip-up clasts". It should be mentioned, however, that "ripped up" clasts were described as early as in 1948 by Shrock (1948a, p. 69–71 and 276). Also, Shrock ascribed some edgewise conglomerates to the process of ripping (1948a, p. 71). The process was also known to Twenhofel (1950), who reported it from tidal flats (p. 593). Ripping may be thought to be a process analogous to quarrying known from bedrock channels.

Some mudflow deposits, composed of clay- or mud rubble embedded in a clayey or muddy matrix can also be classified as hydroclastic cohesive deposits, whereas others are hybrid (auto-hydroclastic) in origin. These are rather common in alluvial fan and play ervironments (Blackwelder 1938; Sharp and Nobles 1953; Bull 1977).

Erosion by flowing water of deeper horizons of clay-rich soils i.e., those having platey-, blocky-, or prismatic texture, leads to the formation of soil pebbles and balls (Photo 12; Jońca 1980). These may locally be accumulated to form soil gravels. Present-day deposits of this kind were recently described by the present author from the extra-channel (floodplain) subenvironment (Teisseyre 1990a, 1991) as well as from slopewash (deluvial) deposits (Teisseyre, in press, in preparation).

CLAYARENITES AND SOILARENITES

These deposits may originate within river channels, as a result of erosion of weathered clays, soils, and muds; on floodplains, owing to the extra-channel erosion of soils or, rarely, clays; and on hillslopes due to heavy slopewash or overland flow. Soilarenites, represented by soil sands and mixed sands, may be predominant slopewash deposits, particularly in regions characterized by clay-rich soils and loess-derived soils (Teisseyre 1990a, in press, in preparation).

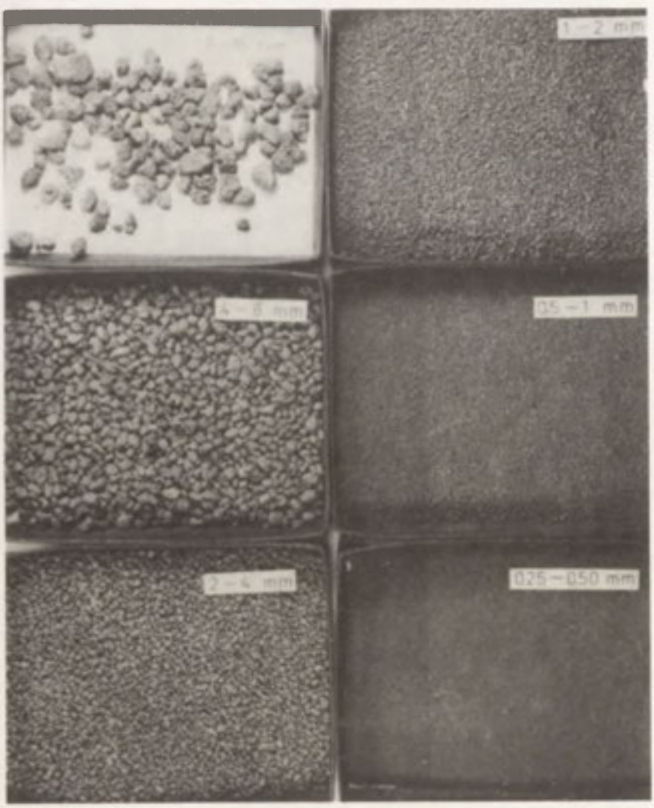

Photo 1. Alluvial soil sand grading to soil gravel. Kazanów, Sudetes Foreland, May 1984
Photo 2. As above, after delicate hand sieving. Fractions finer than 0.25 mm not shown

Photo 3. Crumb (left) and blocky (right) textured soil. Gębice, Sudetes Foreland, June 1986
Photo 4. Prismatic-textured soil. Location as above.

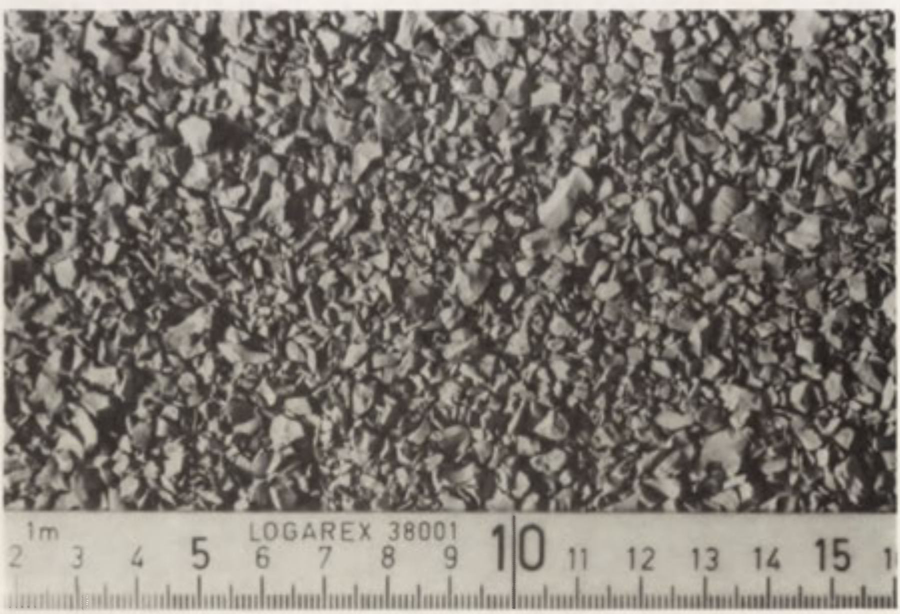

Photo 5. Frost weathering of water-saturated silty clay. Henryków, Sudetes Foreland, January 1991
Photo 6. Frost-riven debris originated from wet massive Holocene clay. Henryków, Sudetes Foreland, January 1991

Photo 7. Talus cone composed of fragments of cohesive deposit (Pleistocene varved clay). Bełchatów Browncoal Mine, May 1991

Photo 8. Weathering of tough Holocene clay. Note talus microcones, prismatic texture and columnar joints. Gębice, June 1986

Photo 9. Clay pebbles and clay sand on bed of small stream, upper River Olawa catchment, Sudetes Foreland, July 1989

Photo 10. Fragments of silty clay (in part indurated) at bottom of sand bed grading upwards into silty sand. Neogene. Nowy Dwor, Sudetes Foreland

Photo 11. Till balls in fluvioglacial deposit (Riss). Jankowa Żagańska, Żagań Hills. May 1991
Photo 12. Alluvial soil gravel. Flow from right to left. Gębice, June 1989

Photo 13.. Alluvial soil sands deposited as large lobate embankments. Kazanów, Sudetes Foreland, May 1984

Photo 14. Alluvial soil sand on upper River Bóbr floodplain. View is upcurrent. Błażkowa, Sudetes, March 1985

Photo 15. Small anthill. Note crusted soil. Wadochowice, Sudetes Foreland, July 1990
Photo 16. Soil sand from anthill. Wadochowice, June 1991
All photos by the author

Soilarenites of fluvial (floodplain) origin were first recognized in arid areas of central Australia by Nanson et al., (1986, 1988) and Rust and Nanson (1986, 1989). Actually, the deposits originate on the Cooper Creek floodplain, Lake Eyre Basin, where they form channel bars, braid bars, and sheets. According to the authors cited, the eposits are composed of "sand-sized pedogenic aggregates" or "sand-sized mud aggregates", but it is evident from the descriptions that they are derived from deeply-cracked, floodplain clay-rich soils. Consequently, the deposits may well be called soil sands according to the present writer's terminology (Teisseyre 1990a, 1991). The deposits constitute the so-called "mud braids" that are active during floods coexisting with typical anastomosing channels (Nanson et al., 1986, 1988; see also Rust 1981, and Pickup et al., 1988).

However, alluvial soil sands do originate and occur also under wet temperate climate and examples of such deposits were described by the author from south Poland (Photos 13, 14; Teisseyre 1990a, 1991, in press). In southern Poland, floodplain soil sands have been found to occur as sheets, irregular patches of sediment, sediment shadows, bars, and lobate microdelta--type embankments. Top surfaces of these deposits may be smooth, flat, dissected (terraced), or rippled. Internal sedimentary structures include parallel subhorizontal lamination, small-scale cross-lamination (attributable to ripples), graded bedding, and large-scale microdelta-type cross-lamination. External (top) sedimentary structures include lag soil pebbles (balls), small sediment shadows, current crescents, current lineation, mud cracks, raindrop imprints, mud ripples, tracks, and trails. Water-level marks have been found on the lobate embankments (Teisseyre 1990a, 1991, in press).

Soilarenites, represented by soil sands and mixed sands, are very common slopewash (deluvial) deposits, too. These originate from erosion, transportation, and (re)deposition of crumb-textured clay-rich soils including in particular brown soils and loess-derived soils (Alberts et al., 1980; Bryan and De Ploey 1983; De Ploey and Poesen 1985; Teisseyre, in press). The deposits may be laid down in a within-slope position or at the base of the slope, on deluvial fans, or bajada-like deluvial sheets. Internally, the deposits are well stratified revealing parallel, subhorizontal (parallel to the slope) lamination, graded bedding (normal, inverse, or pensymmetrical), or a combination thereof. Small-scale cross-lamination attributable mostly to sediment shadows and microdelta-type cross-lamination occur sporadically, mostly at the base of the slope or within deposits of slope micro-depressions. Small-scale cross-lamination connected with ripples occur in some deluvial fans and bajada-like deluvial accumulations. The assemblage of external (top) structures is very impressive including mud ripples, sediment and erosional shadows, current crescents, soil balls and pebbles, current lineation, flutes, grooves, raindrop imprints, mud cracks and curls, ice crystal imprints, tracks, and trails (Teisseyre in press, in preparation).

In the Polish literature, deluvial soil sands were sometimes described as "soil deluvium." The term is not favoured, however, for it tells us nothing as to the petrographic character of such deposits.

ANEMOCLASTIC COHESIVE DEPOSITS

Among these deposits, clayarenites and soilarenites were the first to be recognized. Clay dunes and accompanying wind-generated landforms are known from warm semi-arid to arid areas since the pioneer work by Coffey (1909), who was one of the first to recognize them from the delta of the Rio Grande. These dunes of detrital clay are up to several kilometres long, 60–90 m wide, and 6–9 m high. Sand-sized grains of clay that constitute these landforms are derived from disintegration (salt weathering) of dried lagoonal clay. Later, similar deposits and clay dunes were described from many parts of the world characterized by a warm to hot, semi-arid to arid climate (Price 1933, 1963; Young 1935; Hills 1940; Ives 1946; Stephens and Crocker 1946; Shrock 1948a; Huffman and Price 1949; Twenhofel 1950; Boulaine 1956; Butler 1956, 1974; Roth 1960; Price and Kornicker 1961; Ollier 1969; Bowler 1973; Wasson 1983).

Moreover, red to brown lateritic dunes were reported from tropical regions (Twenhofel 1950). Some varieties of intraformational conglomerate may also fall into this category of deposits, particularly those composed of mud curls embedded in wind-driven sand (Price and Kornicker 1961; Glennie 1970; Bigarella 1972; Tucker 1974; Teisseyre, in press). It is very likely that soillutites may be present among deposits of dust storms, but this needs further examination.

To the category of anemoclastic cohesive deposits one should include niveo-eolian soil-derived deposits called also NIV (Cailleux 1978). These are composed of soil aggregates blown out from frozen grounds, particularly if bare or poorly vegetated, transported by wind, with or without snow, and (re)deposited on an earlier snow drift, on bare or snow-covered ground, or as snow drifts composed of a snow-sediment mixture. The deposits, including both soil sand (soilarenite) and soil mud (soillutite), were described from Polar regions, North America, Europe, Asia, Antarctica, and perhaps also from other continents (cf. Witek 1956; Strzemski 1957; Niewiadomski and Poradowski 1959; Jahn 1961, 1969, 1972; Michel 1964; Gerlach and Koszarski 1967, 1968a, b, 1969; Janiga 1971, 1975; Cailleux 1972, 1974, 1978; Gerlach 1967, 1977; Teisseyre 1979, 1990b). Soil sands of niveo-eolian origin were described from southern Manitoba, Canada, by Teller (1972), who gave them a misleading term of "clay sand." In fact, being composed of soil aggregates, the deposits should be classified as soil sands sensu Teisseyre (1990a, 1991).

Niveo-eolian soil sands and soil muds occur in the form of sheets, covers, irregular patches, sediment shadows, ridges, and wedge-shaped accumulations. Usually they are accumulated upon snow drift, river or lake ice, or simply on frozen ground. Not uncommonly the deposits are interbedded with snow, either clean or dirty ("black, yellow, brown, or reddish snow", "blackened snow drifts"). The top surface of these deposits is originally featureless or rippled (Jahn 1969, 1972; Cailleux 1972, 1974, 1978; Teller 1972). However, after snow melting, the deposits become more or less disturbed or even homogenized revealing a very irregular, cracked hummocky microrelief. Internally, they are more or less homogeneous. Under wet temperate climate, the deposits may soon be incorporated within the soil profile, so that their preservation potential seems to be fairly good. However, many authors suggest that the deposits may lose their original detrital character being converted to a more or less massive soil, clay, or mud.

It should be added that many niveo-eolian deposits are composed of ordinary epiclastic materials, in particular quartz sand and silt. These may be accumulated to form dune fields (Cailleux 1972, 1974, 1978).

Similar nievo-eolian deposits were also reported from the Pleistocene of central Poland and elsewhere (cf. Klatkowa 1985).

Clayrudites and soilrudites may also be produced by salt weathering, desiccation and cracking as well as exfoliation (warm, arid to semi-arid climate; Blackwelder 1931; Price and Kornicker 1961; Glennie 1970; Bigarella 1972; Cooke and Warren 1973; Tucker 1974; Cooke 1979, 1981; and others) or frost weathering (temperate, cold, or periglacial climate). However, with the exception of mud curls and granule-sized fragments, these materials are hardly transportable by wind, though examples of such deposits have been reported from various terrestrial environments (cf. Grove and Sparks 1952).

BIOCLASTIC COHESIVE DEPOSITS

Bioclastic cohesive deposits known to the author include primarily soilrudites and soilarenites. A molehill composed of peds is, in fact, a small biogenic accumulation of bioclastic cohesive deposit. Anthills consisting of rather well-sorted soil sand piled up by ants may be quoted as another example (Photos 15, 16). In some cases, bioclastic mudrudites or clayrudites owe their origin to young shoots penetrating through a thin layer of freshly deposited mud or clay. Individual fragments of cohesive deposits originated in this way are usually platey, more or less equidimensional in the maximum projection plane ranging from several mm to several cm in diameter. Such fragments are commonly lifted by the growing shoots. These mushroom-looking structures seem rather intriguing when seen from some distance for the first time. Certainly, one can mention another examples of bioclastic cohesive deposits, but to compile a more complete list of them has not been the aim of this subsection.

HYBRID (TRANSITIONAL) COHESIVE CLASTIC DEPOSITS

It may be easily demonstrated that the Grabau's taxonomy fails or is difficult to apply when classifying some cohesive detrital deposits. Consider, for instance, the systematic position of a deposit originated as mud curls, which before burial were subject to destruction by rainbeat and then transported for some distance by wind. The original detrital material is autoclastic in origin, or auto-hydroclastic, and the final product will be auto-hydro-anemo--clastic or transitional in character. Despite the complex origin of the detrital material, it may be finally accumulated as intraformational breccia. In fact, many desiccation breccias are hybrid cohesive deposits being the result of the following combinations of processes: desiccation and rainbeat, desiccation plus wind activity, or desiccation accompanied by salt weathering (and eventually also by wind activity; cf. Longwell 1928; Glennie 1970).

It is clear, therefore, that traditional half-descriptive terms like intraformational conglomerate for instance tell us nothing about the complex origin of the detrital material as do many fundamental geologic terms (e.g., sand, gravel, arkose, diamictite etc.). Grabau's genetic classification (and many other genetic systems) has the virtue of being precise and makes the overall organization of our knowledge logical, though slightly academic in character. However, it has at least two major drawbacks: it may be hardly adaptable to consolidated deposits the origin of which is not easy to be established with the sufficient accuracy and it generates long complex terms that may sometimes be inconvenient in practice. An ideal situation is to create a universal set of terms, with descriptive or half-descriptive connotations applicable to analytical stages of the research process supplemented by proper genetic terms necessary for synthetic stages of the scientific work.

SUMMARY OF FIELD AND LABORATORY DATA

It is generally accepted in the pedologic literature that peds of clay-rich soils are relatively stable soil aggregates. Their stability is substantially improved if the soil contains smectite or related expanding clays, calcium carbonate and organic matter. Preliminary field and laboratory experiments made by the present author (1983–1991) on clay-rich brown soils of southwestern Poland have indicated that: 1° Under both natural (forested landscape) and man-controlled conditions (cultivated fields), peds are resistant to repeated wetting and drying. The only exception is rainbeat, which destroys them at the ground surface giving rise to the formation of surface crusts composed of more or less disintegrated soil material. Upon drying the crusts become resistant enough to protect the underlying crumb-textured soil from the direct influence of rainbeat, slopewash, and deflation. 2° Spring melting of snow cover causes partial breaking down of the larger peds, which are subject to disintegration into sand-sized crumbs being the most stable under such conditions. 3° Prolonged permanent submergence of well-ripened peds (laboratory conditions, till 3 months) leads to a similar effect causing the larger peds to break down into sand-sized and silt-sized soil aggregates. In any case a crumb-textured soil became slaked to give a slurry-like mixture of individual silt and clay particles, as intuitively might be expected. 4° During moist days following periods of good weather, the cohesive detrital deposits take up some moisture and become temporarily plastic. In general, this process does not destroy the porous granular texture of the deposits, but some changes are easily perceptible, particularly within the surface layer. With soil sands, these depend upon a tendency to formation of friable surface crusts and/or polygonal platey mega-aggregates composed of soil grains and crumbs that are delicately welded with one another. 5° Slopewash (deluvial) soil sands, deposited in the area of Henryków, Sudetes Foreland, owing to the May 12, 1990 storm, have survived the winter of 1990–91 with no change in both texture and structure. It is suggested that at the earth's surface soil sands are stable deposits at least at the short time scale. 6° Slopewash (deluvial) deposits accumulated as soil muds free from megascopically discernible peds may regain its crumb texture within several weeks after deposition. Crumb texture may also develop within overbank muds and clays in no more than several weeks to several months after the flood. 7° Well-ripened peds are resistant mechanically enough to withstand aqueous transport on a distance up to several hundreds of metres at least (Teisseyre 1990a, 1991, in press, in preparation). (Field and laboratory studies by Nanson et al., 1986, and Rust and Nanson, 1986, 1989, led to similar conclusions.) 8° Peds may also be transported by wind on comparable or even greater distances leading to the formation of cohesive niveo-eolian deposits (see references quoted under the subsection on anemoclastic deposits). However, even small amounts of moisture makes the soil resistant to deflation and may stabilize deposits of soil sand. Thus, formation of niveo-eolian soil sands requires frozen ground, windy weather, and a thin or discontinuous snow cover. In the warm half of the year and wet temperate climate, deflation of crumb-textured soils is negligible or non-existent. Under such conditions, it is only slopewash which may erode, transport, and (re)deposit crumb-textured clay-rich soils. In fact, this happens commonly on bare ground, during early and late phenologic seasons, and on fields occupied by such plants as dent corn, potato, sugar beat, tobacco etc. On grass-covered and forested slopes only catastrophic rainfalls (over 60 or even 100 mm/day) may initiate these processes (Carson and Kirkby 1972; Teisseyre, in press).

Detrital clay materials are the product of sediment ripening involving in particular dewatering, which may be realized in several ways. Dewatering may be occasioned by drying under the direct influence of solar radiation, the activity of wind, or simply moisture-deficient air. These processes result in cracking and exfoliation of the clay surface as well as swelling and curling of superficial laminae of a drying clayey or muddy sediment. However, freezing of wet clay leads to comparable results, because in clays pore water freezes up in the form of segregation ice. This occurs as a network of ice needles or platey crystals as well as veins of segregation ice that in general are parallel to the surface of cooling (Teisseyre 1979, 1984, 1990b). Such a frozen clay or soil is in fact dry and is much lighter-coloured than the original wet deposit. Ice crystals are susceptible to sublimation (Teller 1972), even on vertical cliffs or cuttings sheltered by overhanging turf sheets. The result is frost exfoliation and disintegration of clayey deposits leading to the formation of granule-, pebble-, and sand-sized fragments of dry clay. Such fragments are resistant to subsequent wetting and drying and to some degree also to abrasion. It is thus commonly found that frost-riven clay fragments are transported by wind or running water on a distance up to several hundreds of metres at least, as it happens in cool and periglacial climates. In wetter and warmer climates, larger fragments of clay may be transported by streams on a distance up to several kilometres, as reported by Bell (1940) and others.

CONCLUSIONS

1. Actually, detrital cohesive deposits are rather common in shallow marine and terrestrial environments and their lithified counterparts are known at least since Late Pre-Cambrian. Original cohesive detrital materials are the products of either soil ripening and sediment ripening including in particular desiccation, frost-, and salt weathering, or erosion by running water or moving air.

2. Detrital cohesive materials may accumulate to form individual depositional landforms that range in size from local microforms to large dune fields. Deposition may be accomplished

under the direct influence of gravity (talus cones) or by running water or air (fluvial-, slopewash-, eolian-, and niveo-eolian cohesive detrital deposits).

3. Unconsolidated detrital cohesive deposits may be described using the following terms: soil sand, soil gravel, soil mud, clay sand, clay gravel, soilarenite, soilrudite, soillutite, clayarenite, and clayrudite. The prefix "mud-" may be also applied, if necessary. The deposits can be grouped into genetic categories applied in the Grabau's genetic classification, i.e., autoclastic-, hydroclastic-, anemoclastic-, bioclastic-, and hybrid cohesive detrital deposits (the family of exogenic deposits). Cataclastic cohesive deposits have also been recognized and described. Endogenic granular cohesive deposits have been found among shallow marine sediments. It is suggested that soil sand should contain 75% or more detrital soil grains (peds) and mixed sands 15–75% detrital peds. Deposits containing less than 15% detrital peds are ordinary arenites, if sand-sized.

4. At the short time-scale, detrital cohesive deposits are stable and may be preserved in present-day deposits. Upon compaction and diagenesis pure cohesive detrital deposits will be transformed into mudstones or gritty mudstones, whereas mixed deposits have much better chances to be preserved in a fossil record and, in fact, are known from many formations under a variety of traditional terms (e.g., intraformational conglomerate, if contains shale pebbles, clay-pebble conglomerate etc.).

5. Processes of soil- and sediment ripening are aclimatic and this is why detrital cohesive deposits may be expected under any climatic conditions. However, arid to semi-arid warm or hot climates are perhaps the most favourable for the development of more extensive deposits of detrital clay or soil.

6. Finally, it may be added that scarcity of the literature concerning present-day detrital cohesive deposits grows in part from the fact that they commonly were (and are!) described as soils or clays — an apparently incorrect practice (examples of such misconceptions are quoted, among others, by Boulaine, 1956, and Price and Kornicker, 1961). Another delusion stems from that geologists are not accustomed to distinguish between soils and cohesive deposits such as muds and clays.

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INHERITED LANDFORMS IN THE CRYSTALLINE AREAS OF THE SUDETES MTS. A CASE STUDY FROM THE JELENIA GÓRA BASIN, SW POLAND

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ABSTRACT. Inselbergs are the most striking feature of the granite scenery of the Jelenia Góra Basin. They are either lithologically or fracture controlled. The former are built up of more resistant types of the granite (aplogranite, equigranular granite), while the latter reflect mostly the occurrence of domical structures. Therefore they could be considered as bornhardts. The properties of granite, which forms bornhardts indicate, that the selective deep weathering of Tertiary age following or accompanying by stripping was the most important morphogenetic process leading to the origin and exposure of the residual hills. After the exposure bornhardts were developed in changing morphoclimatic conditions. The formerly steep-sided, bare slopes were remodelled to the form of the boulder-controlled slopes due to strong physical disintegration of domes in the dry environments. The age of this process is probably Late Miocene. Both Pliocene and Pleistocene morphogenesis, which took place respectively in temperate and cold conditions, did not change the inselberg landscape considerably.

INTRODUCTION

Inherited landforms occur practically in all regions of the world and they often play a dominant part of landscape. The presence of several generations of relief, which are of different ages was especially noticed in Central Europe (Büdel 1963, 1977). During Cainozoic climates have been changing so often, that only 5% of nowadays observed landforms can be considered as connected with the present-day temperate morphoclimatic zone (Büdel 1977, p. 197).

The sediments coming into being under definite environmental conditions and disharmonious landforms are the proof of the variability of climatic zones. To these old saprolites, duricrusts, laterites and kaolins and also remnants of planation surfaces, inselbergs, pediments, tors and weathering phenomena are included (Linton 1955; Wilhelmy 1958; Dury 1971; Büdel 1977; Thomas 1978a; Battiau-Queney 1987; Summerfield, Thomas 1987; Bremer 1989; Godard 1989). The elements of ancient relief are still being transformed and they are adjusted to the present-day conditions. Thanks to this, within macroforms smaller forms can be identified, which are connected with variability of morphoclimatic zones (Bremer 1965).

During last years a considerable increase of interests in long-term landform evolution can be observed. The problems of the morphogenetic role of climatic changes in Cainozoic and the presence of inherited landforms play the most important part in these topics.

INHERITED LANDFORMS IN THE SUDETES MTS.

The presence of disharmonious, pre-Quaternary landforms was mentioned in the beginning of the 20th century. Remnants of planation surfaces and single, isolated inselberg-like hills in the Sudetic Foreland were included among these forms (Berg 1927; Anders 1939). On the ground of their relation to the Tertiary kaoline mantles their age was defined as Early Tertiary. Particularly inselbergs were considered as indicators of tropical morphogenesis (Gellert 1931). Afterwards, some differences in appearance between inselbergs of the tropical zone and hills of the Sudetic Foreland were stressed. They were considered as a proof of creation in cooler climate of Mediterranean type (Gellert 1970). J. F. Gellert's conclusions were based mainly on visual resemblances and differences.

Then, the granite tors of the Karkonosze Mts. became the subject of detailed study (Jahn 1962). Their origin is connected with selective, deep weathering in pre-Quaternary period following by stripping of the weathering mantles. Cooling was probably a main reason of exposing of tors. It caused that processes of deep weathering were getting less and less intensive, while denudation was getting more and more active.

The presence of the elements of Tertiary relief and their palaeoclimatic significance was discussed with regard to the crystalline areas of the adjacent Bohemian Massif. Tors and microforms on rock surfaces were the subject of detailed investigations (Demek 1964a; Demek et al. 1964). The tors were considered as forms of Tertiary origin, partly remodelled during Pleistocene. Most of microforms as tafoni, pseudokarren and hollows were thought to be fossil, dated from Neogene. Also granite inselbergs were described as Tertiary landforms (Demek 1964b; Ivan 1983).

A. Jahn's work (1980) gave general comprehensive ideas of the evolution of the Sudetes Mts. relief during Tertiary. According to him macroforms as planation surfaces, inverted forms and intramontane basins have clear traces of Palaeogene tropical morphogenesis. Later periods of arid and periglacial morphogenesis gave only minor elements of relief and did not change previous forms in significant way.

This review shows there is no work about the Sudetes Mts., describing results of variability of the environmental conditions on examples of particular landforms. The most suitable area for such investigations seems to be the Jelenia Góra Basin (the Western Sudetes Mts.), built up of granite. The processes of sub-aerial degradation have been lasting here since Permian (Mierzejewski 1985). The striking feature of the Jelenia Góra Basin landscape is the occurrence of inselbergs. They are elements of at least pre-Pliocene relief. It is evidenced by Pliocene and Quaternary sediments occurring at the foot of these hills (Zimmermann 1937; Berg 1941; Grodzicki 1967). So, it seems that climatically induced geomorphological events of at least the last several million years could be recorded in relief of inselbergs.

GEOLOGICAL SETTING AND RELIEF OF THE JELENIA GÓRA BASIN

The Jelenia Góra Basin is the north-eastern part of the Karkonosze — Izera granite block of Late Carboniferous age. It occupies an area of approximately 270 km^2. Although there is a general agreement, that the basin came into being during Late Tertiary (Berg 1927; Jahn 1980), its origin is not conclusively explained. The rectilinear southern margin is a typical example of fault scarp (Berg 1927) and displays features of Quaternary rejuvenation (Migoń in press). The nothern slope is also related to fault line and is considered as a fault scarp too (Fig. 1). The western margin is lithologically controlled and separates granite of the Jelenia Góra Basin and gneisses of the Izera Foothills. The differences in altitude reaching up to 100 m reflect probably different intensity of deep weathering during Tertiary. The Late Tertiary faulting, which strongly influenced relief of the adjacent Karkonosze Mts. did not disturbe the relief of the bottom of the basin. The long, rectilinear escarpments of endogenetic origin, dislocated remnants of planation surfaces and deeply incised river valleys are almost

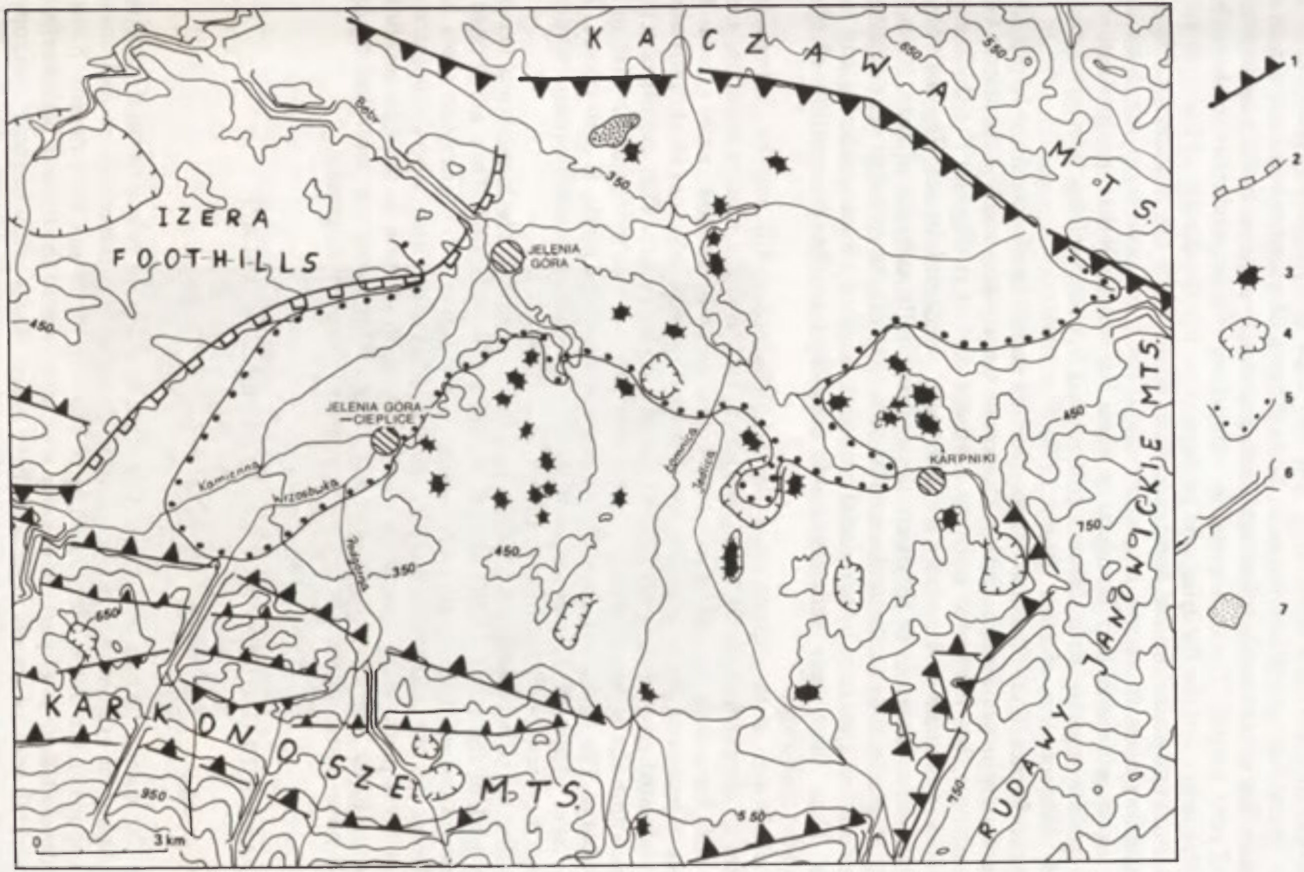

Fig. 1. Main features of relief of the Jelenia Góra Basin
1 — fault scarps, probably of Late Tertiary age, 2 — structural escarpments originated due to differential etching, 3 — inselberg-like hills, 4 — closed depressions and minor basins, 5 — southern limit of the Scandinavian ice-sheet, 6 — deeply incised valleys, 7 — Pliocene gravel sediments

quite absent. The variety of landforms and distribution of hills, ridges, basins and valleys derives from structural characteristics of granite masses.

The porphyritic granite containing large crystals of orthoclase is the most widespread and occupies about 80% of the studied area. The outcrops of aplogranite, granophyre and fine-grained granite are much less common. Several types of younger veins cut across the granite masses. There are mostly aplites and microgranites. Granite is strongly differentiated in respect of joint pattern. The orthogonal set comprising two perpendicular to themselves, nearly vertical joints and the flat-lying one predominates, but the density of fractures varies greatly. In numerous places distinct dome structures occur, which are characterized by well developed sheeting joints. Moreover, large parts almost free from joints occur, which manifest themselves as huge, monolithic boulders. It must be stressed, that predomination of porphyritic granite together with different pattern and frequency of fractures gives rise to considerable differentiation of relief (Brook 1978; Pye 1986).

The direct influence of the Scandinavian ice-sheet on granite landforms in the Jelenia Góra Basin was generally inconsiderable. During Pleistocene the ice-sheet reached the Jelenia Góra Basin at least once, but it covered only the northern part of it (Fig. 1). The thickness of decayed glacier was small and so only few hills were overlained by ice. The majority of inselbergs was either nunataks or they were situated within the extraglacial zone. If yes, the remodelling of granite hills to the roche-moutonnėes form can be excluded. The inselbergs might have been reshaped only by periglacial processes, mostly by frost weathering and mass movements. Thus, there were very favourable conditions to the preservation of older, pre-Quaternary landscape.

In the whole area of the Jelenia Góra Basin 120 individual hills can be distinguished. Among them 25 display features of "true" inselbergs. They stand in relative isolation, are at least 25 m high, have steep and sometimes bare slopes, which inclination locally exceeds 45°. The presence of piedmont angle is more or less distinct, so some inselbergs rise abruptly above the gently undulated ground. Another 21 hills are of complex nature. They consist of several small knobs within the wide top surface. Their slopes are as a rule more smooth than in the case of inselbergs. The change of slope at the piedmont angle is gentle, if even occurs. The other hills have shape of low convex knolls or block- and boulder-strewn residuals. Respectively, they can be classified as ruwares and nubbins.

The distribution of inselbergs and high complex hills is irregular. The highest ones occur in the eastern part of the basin, in the vicinity of the village Karpniki. They are formed of resistant fine-grained aplogranite and their height reaches up to 125 m (Krzyżna Góra hill). These inselbergs are obviously petrologically controlled. The next area, where concentration of inselbergs takes place is the central part of the basin. In relation to the hills mentioned above they are lower and their height does not exceed 60 m. They are controlled by joint density and have developed in less fractured parts of porphyritic granite.

CLIMATICALLY INDUCED ELEMENTS OF INSELBERG LANDSCAPE

DOMICAL STRUCTURES

Among 120 granite hills in the Jelenia Góra Basin about 40 reflect domical structures. Their morphological expression depends on some structural characteristics of rock masses. The most important are dome size, dip angle of sheeting joints and their density. Some of these hills, developed on large parts of massive rock, gave rise to the form of high inselbergs. They are several tens of meters high and have steep slopes with frequent outcrops of spheroidal surfaces of sheeting. These bare surfaces are common mostly in the upper slope sections. The slope changes at the foot of the domed inselbergs are rather abrupt and distinct. Domes are constructed of massive granite, which is characterized by the presence of few open joints. The morphological and structural features allow to regard these hills as bornhardts in sense of Thomas (1978b) and Twidale (1982). Although bornhardts have been strongly remodelled, their characteristic features described above are in some places still well recognizable. Hills like Witosza (484 m), Czop (458 m), Gołębnik (459 m), Radlica (427 m) and Koziniec (462 m) offer the best examples (Fig. 1). At the same time, at least several hills undoubtely controlled by dome structures have been only partly exposed and display features of half-domes. They are asymmetrical in shape and are situated mostly within the flanks of depressions or on valley sides.

The primary sheet structures crop out on the bornhardt slopes in few places. However, the significance of structural factor can be well recognized in numerous quarries. The elongated half-dome near by Jelenia Góra – Cieplice cut across by the old quarry offer probably the best opportunity to study the internal structure of the hill. It is of 20–30 m height and is built up of porphyritic granite. The inner parts of this bornhardt were excavated up to depth of 20 m. The orthogonal joint system with increasing number of open joints towards the topographic surface is absent. Near the surface only irregular fractures originated probably due to offloading can be in places observed. On the contrary curvilinear sheeting planes dipping radially from the axis of the hill at low angles (15–25°) are good visible on the quarry walls. Jointing planes are widely spaced (1–3 m), tight and without signs of weathering (Photo 1). In the bornhardt morphology they manifest themselves as smooth spheroidal surfaces in the upper sections of slopes. The absence of vertical and sub-horizontal joints is of the greatest morphogenetic significance. Infiltration of percolating water and its later stagnation along subhorizontal planes is then impossible.

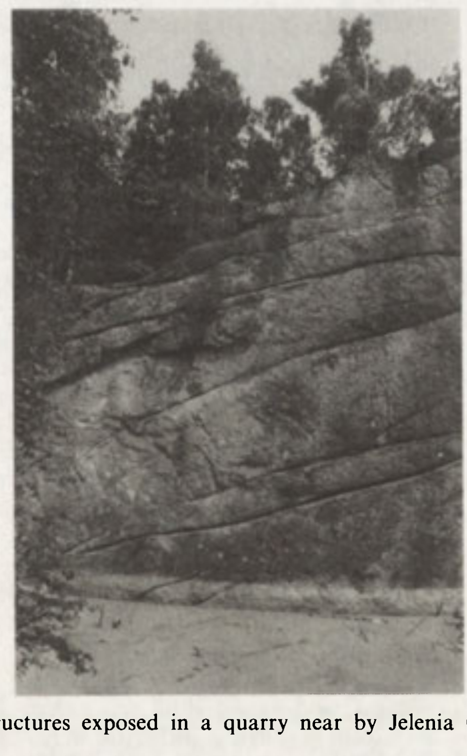

Photo 1. Sheet structures exposed in a quarry near by Jelenia Góra Cieplice town

The similar absence of orthogonal joints was stated in the case of Witosza hill. However, sheeting joints are very well developed, as regular circular shape in plan and a considerable number of exposed, smooth spheroidal surfaces indicate. Thickness of particular exfoliation slabs is 1–3 m, or even sometimes more. Joints separating following slabs are rather tight.

Photo 2. Sheeting planes on the western slope of the Witosza hill

Sheeting planes are steeply inclined, from 15–20° in the summit parts up to 45–60° in the lower slope sections (Photo 2).

The common feature of bornhardts is a conspicious accordance between their outline and slope morphology and geometry of the dome structure. Therefore apart from the symmetrical hills there occur also the asymmetrical ones. In the case of the second group inclinations of the opposite elopes may differ 2–3 times (Czubek 388 m). Some of the complex hills consist of several minor domes. They are separated by 15–20 m deep linear depressions reflecting lines of dome junction (Radlica 427 m).

Both the shape and the plan of the inselbergs, strongly influenced by the geometry of domical form suggest that geological structure plays an important part in their origin and further development. Features of granite masses in which bornhardts are best developed including solidity, a lack of petrographical differentiation, the presence of sheeting joints and limited number of open vertical joints indicate, that they are particularly resistant to chemical attack. Then they must have weathered more slowly than the adjacent dense fractured and permeable parts. Progressive etching following and/or accompanying by stripping of weathered mantles had to bring to the emergence of these solid parts, which gave rise to the bornhardt form. It seems that their occurrence indicates developement of landscape in warm, seasonal humid climates of savanna type, due to etchplanation. Many scientists assume, that the most favourable conditions to the creating of bornhardts take place in the tropical and subtropical zone (Thomas 1965, 1978b; Wilhelmy 1977; Bremer 1981).

The wide occurrence of residual, well rounded boulders seems to confirm the hypothesis, that the landscape here was created through continuous etchplanation. Their size strongly differs reaching dimensions 10 × 6 × 4 m, though the modal diameter is between 1.5–2 m. They are situated in various morphological positions. Boulders occur on flanks of inselbergs, surmount small hills and ridges and emergence from flat bottoms of small circular basins and elongated depressions. They stand either in isolation or in groups, which may be considered as

Photo 3. Single huge corestone exposed on gently inclined slope. Height of this boulder is about 4 m

nubbins. Both their shape and distribution exclude the possibility they are derived from mechanical weathering of tower-like tors. On the contrary, similarly to the bornhardts and other inselbergs, they had to have been excavated from deep weathering profiles too (Photo 3).

BLOCKS AND BLOCK-FIELDS

The characteristic feature of bornhardts in the present-day tropical zone is often the absence of talus accumulation below them. The convex profiles of hillslopes are well marked up to the foot of inselbergs. Sometimes large boulders occur near the hill/plain transition, but they were also exposed due to stripping (Thomas 1978b). In the case of bornhardts of the Jelenia Góra Basin the lower sections of hillslopes are covered by extensive block accumulation. It causes, that slope profiles are block-controlled and rather straight. Moreover, in the upper parts of domes there are relatively few outcrops of spheroidal surfaces related to sheeting planes. It proves that bornhardts have been strongly destroyed after their exposure.

The majority of blocks consisting block fields has similar shape. On average they are of 3–4 m length and width, while their height is smaller and is 1.5–2 m. Both upper and lower surfaces, being parallel to themselves, are slightly curved, therefore blocks look like large slabs. The edges of blocks are often sharp or only partly rounded. Particular blocks are inconsiderably or not at all jointed. It gives them the monolithic form. The largest blocks were observed on the slopes of the Witosza hill, where they are of 15 m length and of 3–4 m thick. Granite blocks on the slopes lie chaotically, so empty spaces between them often occur. The dimensions of some of them exceed several meters length and height, which allows to regard them as small boulder caves (Photo 4). The extent of block fields is limited and restricted to the steep slopes of bornhardts. Because they do not encroach on adjacent parts of plains, their strong connection with development of inselbergs is suggested.

Photo 4. Extensive block fields on the Witosza hill. Due to irregular accumulation of blocks small caves have been originated

The described blocks differ strongly in terms of shape and morphological positions from corestones mentioned earlier. They resulted probably from the rapid disintegration of sheet structure. After the breakdown of following slabs blocks were transported to the lower parts or even to the base of the slope. More steep were sheeting planes, more effective was this process. The slabs were subdivided into blocks along crossing them vertical joints or radial fan-joints. The latter were created in places of strong tension and are disposed to each other at the angles 90–120°. The infrequent outcrops of lower slabs below already subdivided and moved blocks indicate that block fields can not be derived from differential weathering of bedrock.

The residual character of summit parts of the hills is an another evidence of considerable disintegration of bornhardts. On the exposed sheet surfaces often occur isolated, monolithic pedestals of 1–3 m height. They are in fact remnants of upper, already completely degraded slabs. On the Witosza and Skalista hills the uppermost sections of slopes make elongated crests reaching up to 20 m height and 100–200 m length. In the case of the Witosza hill the top surface of that crest is inclined at low angle, according to the inclination of the upper slab.

The occurrence of block fields on the slopes of formerly created bald, steep-sided bornhardts indicates the rapid change in the morphogenetic system. Till then predominated deep weathering and following excavation of resistant compartments was replaced by mechanical disintegration of outcrops. This change led to the transformation of bornhardts into the form of block-strewn inselbergs. It seems evident that this replacing had to have been connected with severe climatic change, probably with the shift to more arid conditions. Factors influencing rapid physical breakdown of the rock as salt, insolation or even frost weathering operate in this zone very effectively. It may be assumed that remodelling of bornhardts to the form of boulder inselbergs took place in semi-arid and arid climatic conditions. The boulder inselbergs are the characteristic element of this morphoclimatic zone

(Bremer 1981). That period was rather short, and therefore some features of bornhardts' morphology preserved.

The piedmont surfaces cut in bedrock, relatively smooth, gently concave and inclined at angles of several degrees seem to have been also reshaped in the arid conditions. Their smooth profiles are sometimes broken through the occurrence of ruwares and boulders. These features suggest they are rock wash prediments in sense of Büdel (1957) and Twidale (1982).

DEEP WEATHERING PROFILES

Deep weathering profiles in the Jelenia Góra Basin have been evidenced in several places. Their common feature is the same morphological position. They occur in the lower parts of slopes, where the slopes contact with the piedmont surface.

Weathering profiles are about 5–7 m thick, but they can reach even 12–15 m. Their characteristic feature is almost uniform rotting of granite in the whole profile and often lack of corestones. Corestones can be observed only in few profiles. The length of them is about 2 m, but they have also distinct traces of disintegration. They are uncomparable with the big granite boulders described in chapter concerning dome structures, which are rather unweathered. The parts of less disintegrated granite, occuring in some profiles, are connected with more equigranular granite or with veins of microgranite. Porphyritic granite is almost completely disintegrated into the grus. The crystals of quartz and orthoclase are quite good preserved, but plagioclases have been remarkably alterated. In these profiles clay minerals, chlorite, epidote and iron oxides can be observed. Generally it is impossible to reconstruct the original joint pattern.

Some of the Jelenia Góra Basin profiles were the subject of detailed study (Borkowska, Czerwiński 1973). The domination of the physical disintegration over the chemical decomposition and general progress of weathering from a surface downwards was stated as a result of the investigations. The grus was classified as a weathering mantle of growan type. According to the quoted authors this weathering took place in climate not so much different from the present one. It was moderately warm and humid.

The features of regoliths show, that they can not be the relics of the subtropical morphogenesis when bornhardts were progressively exposed and it was accompanied by subsequent deep weathering on the surrounding plains. They also can not be connected with the period of semi-arid morphogenesis. Because periglacial covers overlie these regoliths and they partly disturbed their upper parts, regoliths must have been of pre-glacial age. They are indicators of morphogenesis taking place under conditions of temperate climate, in pre-Pleistocene period.

SLOPE COVERS

The slope covers are the youngest sediment, which registers the evolution of the inselbergs. They occur both on the hillslopes and on the piedmont plains. They cover the grus described above. Their thickness is from 0.2–0.4 m up to 3 m in the vicinity of the main valleys. On the inselbergs they are not more than 1 m thick, so they do not play an important role in the slope morphology. Sections in which slope covers overlie formerly exposed corestones or blocks coming from disintegration of sheet structures have not been recorded.

Structural and textural features of these sediments indicate that they originated under periglacial conditions (Jahn 1960; Traczyk in press). They include material resulting from mechanical weathering, give evidence of rhythmicity and bedding. Structures derived from solifluction were also noted. A common feature of these sediments is their fine-grain structure. They are mostly sandy and silty, although they contain fine debris too. The presence of the sharp-edged pieces is connected with outcrops of vein rocks or fine-grained granite. The slope covers include material, which has been incorporated from the upper parts of the regoliths and products of porphyritic granite microgelivation. Typical periglacial block covers, made of sharp-edged blocks have not been noted within inselbergs built up of porphyritic granite. These are characteristic for outcrops of the equigranular granite in the Karkonosze Mts. (Dumanowski 1961). The character of the periglacial covers suggests that the Pleistocene cooling caused only inconsiderable changes in the relief of the inselbergs.

MICROFORMS OF ROCK-SURFACES

Surfaces formed in porphyritic granite are characterized by the variety of microforms, which are mostly related to different weathering processes and response of rock. Their distribution and features suggest that they are of different age and came into being in following stages of bornhardt development.

The most rare and rather fossil forms are pseudokarren and shallow gutters. The former occur on vertical, sometimes overhanging walls, while the latter on gently inclined surfaces do. The gutters often exploit joint lines and seem to be partly fracture-controlled. The distribution of pseudokarren is limited, because they are observed only in the upper parts of the inselbergs, on pedestals or surfaces of exfoliation slabs in situ. These parts probably survived from a subtropical phase of inselberg development (Photo 5). The origin of karren is often related to the warm humid tropics (Wilhelmy 1958, 1977; Bremer 1965; Twidale 1982), therefore they were considered as indicators of former subtropical climate (Demek et al. 1964). Although it was stressed that they might originate also in other climatic conditions (Thomas 1976), it is salient, that pseudokarren did not develop on rock surfaces exposed later. The other elements, which may be connected with subtropical conditions are remnants of flared slopes. They were also recognized only in the upper slope sections.

Photo 5. Karren on residual boulder in the upper part of slope of the Witosza hill

Basis-tafoni observed along sub-horizontal joints are more frequent in granite inselbergs of the Jelenia Góra Basin. These forms still develop, as it is evidenced by the occurrence of fresh grus in niches and hollows. However, it is likely, that large hollows and niches up to 2 m depth originated in semi-arid climates, when the most favourable conditions for their development took place. One of the main factors controlling the origin of tafonis is salt weathering, which causes both flaking and granular disintegration (Bradley et al. 1978; Twidale 1982; Smith, McAlister 1986). The rapid development of tafonis located on the undersides of sheet structures may help to explain a considerable disintegration of bornhardts during period of semi-arid conditions.

Weathering pits and pans are regarded as the present-day phenomena (Wilhelmy 1958; Demek et al. 1964), although in the Karkonosze Mts. their initiation dates surely from cool periods of Pleistocene (Chmal 1974). On the contrary to the karren and flared slopes, pits occur in various morphological positions. They were recorded both on outcrops and loose blocks and boulders at the foot of the hills. The contemporaneous weathering process is mainly physical granular breakdown. It operates especially rapid in hollows, niches and on shaded walls, producing quartz-feldspathic grus in and below them and alveoling of rock surfaces.

STAGES OF INSELBERGS' DEVELOPMENT

Features of relief of inselbergs and the characteristics of sediments, which occur in the Jelenia Góra Basin suggest following variability of morphoclimatic phases. The oldest phase was connected with the subtropical climates, warm and seasonally humid, probably of savanna type. The dominant processes were differential etching accompanied by progressive stripping, leading to the general lowering of the surface. At the same time both bornhardts and single corestones have been exposed. Later, predominant chemical weathering was replaced by the physical one, as an effect of increasing aridity. Bornhardts being till then strongly resistant became a subject of rapid disintegration. As a result of it, extensive block fields containing fragments of following exfoliation slabs developed around inselbergs. The repeated increase of humidity accompanied by gradual cooling causes that the Jelenia Góra Basin became under the temperate climatic conditions. During that period regoliths of growan type developed at the foot of the hills, whereas the form of bornhardts and boulder inselbergs inherited from past morphoclimatic phases was remodelled inconsiderably. The periglacial processes creating fine-grained slope covers did also not change the older relief.

It seems interesting to compare this sequence of morphoclimatic changes with palaeoclimatic reconstructions of Cainozoic in SW Poland (Głazek, Szynkiewicz 1987; Sadowska 1987). It is obvious that the inselbergs had to have been exposed not later as in Early Pliocene and it is indicated by gravel sediments of Pliocene age preserved in the northern part of the basin.

The warm and humid subtropical climate, which favoured the development of inselbergs, prevailed until the end of Middle Miocene (about 12 My ago). At the turn Middle/Late Miocene a severe increase of aridity took place. Until the end of Late Miocene (5.5 My ago) the Jelenia Góra Basin remained in the arid zone. This shift towards aridity was connected with the Messinian Crisis (Głazek, Szynkiewicz 1987). This period was favourable to considerable disintegration of inselbergs in the dry, arid environment. Some domical structures, especially the smaller ones might have been then completely degraded. Throughout Pliocene the temperate climate prevailed with tendency to gradual cooling. The origin of regoliths of growan type could be correlated with this period. Some of these profiles might have been truncated during cold stages of Pleistocene. During these stages an action of periglacial processes (frost weathering, solifluction) took place, while during interglacials growan regoliths were probably further evolved. The Pliocene and Pleistocene morphogenesis did not change the inselberg landscape of the Jelenia Góra Basin sufficiently. At that time inherited contrasts of relief were rather softened by processes of fluvial (Pliocene, Pleistocene) as well as glacial and slope accumulation (Pleistocene).

CONCLUSIONS

The investigations carried out in the granite area of the Jelenia Góra Basin have shown that within residual inselberg-like hills one can indicate forms belonging to different generations of relief in J. Büdel's sense (1977) (Fig. 2). They confirm that morphoclimatic phases during Neogene and Pleistocene were changing. Both the macroforms and the microforms on rock surfaces and sediments occuring on the slopes of the inselbergs are the proof of this variability. The sequence of morphoclimatic changes was gained from the analysis of forms, which evidences the evolution from the subtropical morphogenesis through the semi-arid and the temperate to the periglacial and then back to the temperate one. This sequence has been confirmed by palaeoenvironmental reconstructions.

The granite hills of the Jelenia Góra Basin are generally the result of the subtropical morphogenesis and their age goes back at least to Middle Miocene. So the relief of the basin is old and the last significant changes were connected with Late Miocene. The fact that this relief has persisted to nowadays can be explained by two factors. Later stages of morphogenesis were not effective, it refers especially to the periglacial morphogenesis. It was mainly caused by strong resistance of porphyritic granite to destructive factors operated during this period. So even the microforms from the earlier stages of morphogenesis were able to persist. The second factor was insignificant influence of the Pliocene tectonic movements on the older relief. During these movements only surrounding massifs of the Karkonosze Mts. and the Izera and Kaczawa Mts. were uplifted.

The results of the investigations in the Jelenia Góra Basin can become a base for the morphoclimatic reconstructions in surrounding areas, where the possibility of persisting of the old Tertiary relief was less favourable and where only relics of the ancient landscapes occur.

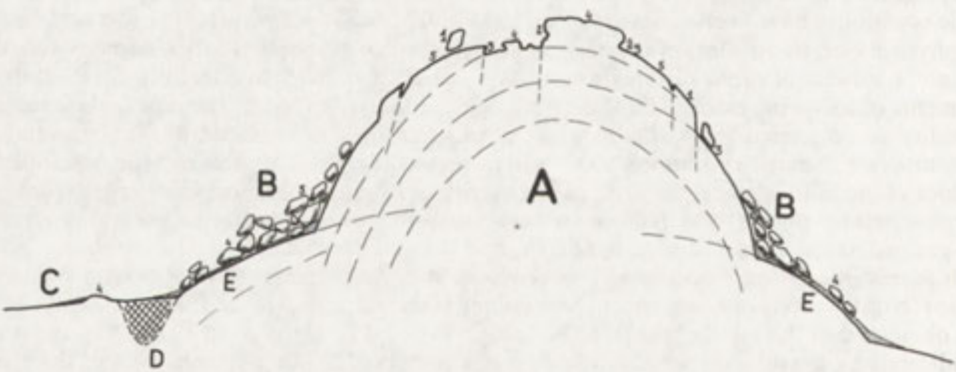

Fig. 2. Schematic profile of a granite dome showing relief elements of various age
A — bornhardt form developed on domical structure — at least of Middle Miocene age, B — block fields due to disintegration of dome, C — rock wash pediments, D — deep weathering profiles of Pliocene age, E — fine-grained periglacial slope covers. Occurrence of microforms: 1 — karren, 2 — flared slopes, 3 — basis-tafoni, 4 — weathering pits, 5 — pans and rinnen

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CUESTA LANDSCAPE IN THE MIDDLE PART OF THE SUDETES MTS.

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ABSTRACT. The cuesta landscape has been analysed in the border area of the Upper Bóbr River drainage basin and the Upper Ścinawka Kłodzka River drainage basin. Thanks to differentiations within the monoclinal basement of the area, it was possible to observe the structural control of relief forms and influence of the structure on the evolution of those forms. On the other hand, the localization of the study area in the border zone of two different drainage basins makes it possible to analyse the influence of the intensity of river erosion on the shapes and morphodynamic of escarpment forms in both drainage basins of the same rocks. Cuestas, a valley step and resistance scarps have been found within the study area. Evidence of the lively morphodynamic of those forms in the Pleistocene is dealt with.

I. INTRODUCTION

Due to the variety of the geological basement, the landforms developed on sedimentary rocks in the study area provide good examples for an analysis of the dependence of relief on different geological factors. Since the area is localized out of the range of the Pleistocene icesheets, an evolution of its morphology should occur under the strong influence of the periglacial conditions favouring the development of cuestas (Schunke, Spönemann 1972). Adjacent parts of the Sudetes were twice covered by icesheets (Szczepankiewicz 1953; Jońca 1975). The author has compared conclusions drawn in the articles dealing with the origin and dynamic of cuesta relief in Europe (Barth 1975; Blume 1972; Büdel 1957; Schmitthenner 1954; Mortensen 1947; Schunke, Spönemann 1972; Tricart 1951) with his own field researches in the study area.

Studies concentrate on the question whether evidence of significant changes in the morphology of that region in the Quaternary can be found. Pulinowa (1989) has already described such changes in the escarpment relief of the Stołowe Mts., south-east of the study area.

II. ESCARPMENT FORMS IN THE STUDY AREA

The study area comprises ca 40 km^2 in the central part of the Sudetes. It is situated on the north-western limb of the intra-Sudetic synclinorium. Its southern and middle parts belong to the Ścinawka Kłodzka River basin while the northern and north-western ones are drained by the tributaries of the Bóbr River (Fig. 1). A Permian volcanitoe ridge borders on the study area in the north-east. The following sedimentary rocks outcrop parallel to this ridge (Fig. 2): shales and conglomerates of the Lower Perm, dolomitical arkoses of the Upper Perm, Lower Triassic

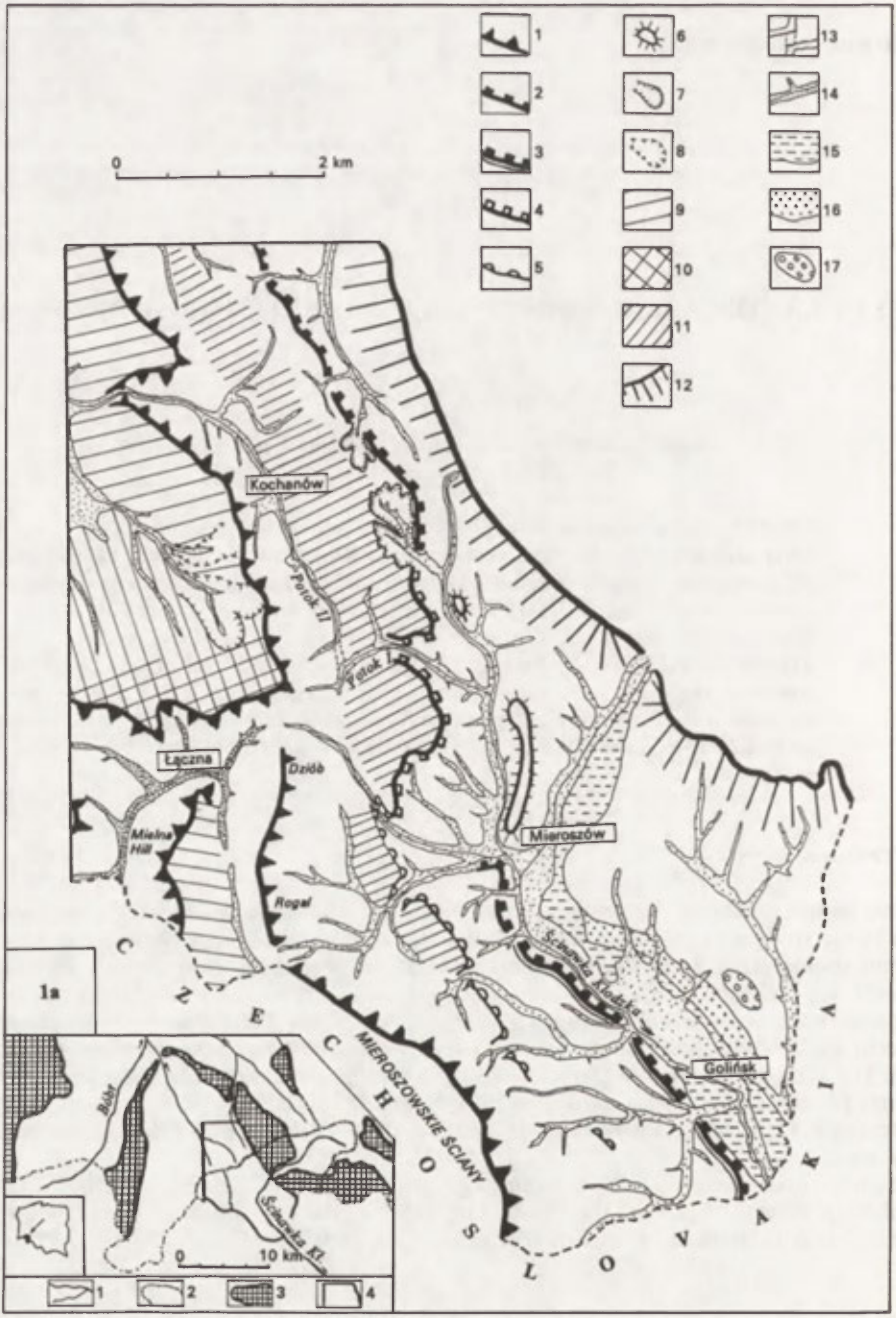

Fig. 1. Morphological sketch of the study area

1 — Cretaceous sandstone scarps, 2 — Lower Permian conglomerate scarps, 3 — valley step, 4 — composite scarp, 5 — resistance scarps, 6 — outliers, 7 — amphitheatral valley heads, 8 — remains of valley heads, 9 — intersecting blackslopes, 10 — structural top surfaces, 11 — exhumed infra-Triassic surface, 12 — Permian volcanitoe ridge, 13 — capture, 14 — Holocenian alluvium, 15 — Weichselian terrace, 16 — Saalian terrace, 17 — Elsterian terrace

Fig. 1a. Localization of the study area in the Middle Sudetes

1 — main rivers, 2 — watersheds of the Bóbr River and Ścinawka Kłodzka River drainage basins, 3 — mountain ridges, 4 — borders of the study area

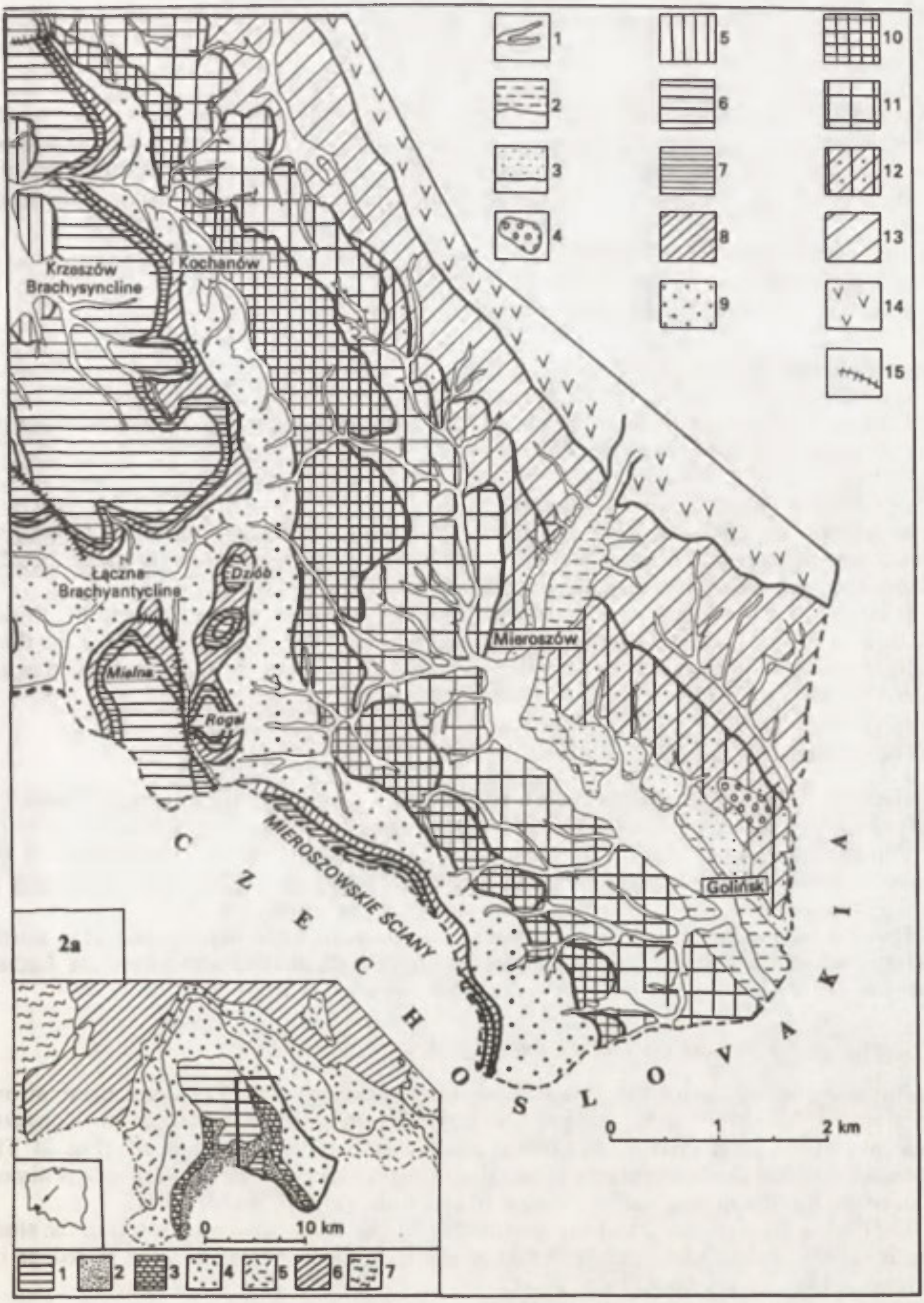

Fig. 2. Geological sketch of the study area

1 – Holocenian alluvium, 2 – Weichselian terrace, 3 – Saalian terrace, 4 – Elsterian terrace, 5 – Turonian sandstones, 6 – Cenomanian sandstones, 7 – Cenomanian mudstones, 8 – Cenomanian tabular sandstones, 9 – Lower Triassic clayal sandstones, 10 – Upper Permian dolomitical arkoses, 11 – Lower Permian conglomerates, 12 – Lower Permian shales and sandstones, 13 – Lower Permian shales, 14 – Lower Permian volcanitoes, 15 – faults

Fig. 2a. Localization of the study area in northern part of the intra-Sudetic synclinorium

1 – Cretaceous rocks, 2 – Triassic sandstones, 3 – Upper Permian dolomitical arkoses, 4 – Lower Permian sedimentary rocks, 5 – Lower Permian volcanitoes, 6 – carboniferous rocks, 7 – metamorphic massive

clayal sandstone and a set of Cretaceous sandstones and mudstones. The courses of the rock layers range between 320 and 350 degrees and layer inclinations between 3 and 30 degrees with 5–15 degrees the most frequent (Grocholski 1973; Jerzykiewicz 1971; Krechowicz 1968).

Three main sets of scarps have developed on the outcrops of resistant rocks in the study area:

– Cretaceous sandstone scarps,
– Upper Permian arkose scarps,
– Lower Permian conglomerate scarps.

A. THE CRETACEOUS SANDSTONE SCARPS

The following characteristics of the sandstones have facilitated the shaping of these scarps:

– regular and close jointing,
– high level of compaction,
– high content of a siliceous binder.

Those features make the rock far more resistant to destructive processes than the underlying Triassic sandstones do, despite the fact that both rocks have a similar mineralogical composition and grain size distribution (Grocholski 1973).

Because of the local differences in their morphology, the sandstone scarps can be divided into three sets. This is determined by the Cretaceous beds in the region belonging to the three small tectonical units described by Jerzykiewicz (1969). Changes in the shape of the scarps are clearly related to a variation of the tectonical structure.

1. The Mieroszowskie Ściany scarp

The form is the highest scarp in the study area with a relative altitude ranging between 100 and 140 m. A slope face dips at an angle of 20–30 degrees. In the northern section of the scarp, the altitude decreases gradually. This is caused by a declining of the participation of the Triassic sandstones in the structure of the slope face. Consequently, there the rock starts to form a very broad footslope dipping gently towards the north-east.

The Mieroszowskie Ściany scarp is built of Cretaceous beds outcropping at a north-eastern limb of the Policka brachysyncline. The form ends at the north where the Łączna brachyantycline (Fig. 2), the next tectonical unit begins.

2. Scarps of the Łączna brachyantycline

The complicated tectonical structure in the brachyantycline has determined strong variations in the shape of the scarps. The Rogal-Dziób ridge is formed at an elongated fragment of the Cenomanian rocks uplifted along a fault running south-north (Fig. 2). The fault has facilitated the development of two deep stream valleys. Their dipward side is almost as steep as the updip directed slope face of the ridge (Fig. 3, profile G-H).

The backslope of the neighbouring Mielna Hill, unlike the other similar forms in the study area, is exposed to the south-east. This fact results from the localization of the backslope on a dropped limb of the fault.

The Rogal-Dziób ridge and the Mielna Hill are parts of the southern head of the Łączna brachyantycline. The head is isolated from its northern analogue by a broad valley in which the Cretaceous beds are so deeply intersected that Triassic sandstones occur in the valley floor (Fig. 2).

A 100–120 m high scarp has evolved on the northern head of the brachyantycline. Since the bed inclinations range here from 0 to 3 degrees, the scarp has a flat top surface at the back of a slope face (Fig. 1). A typical for cuestas, backslope begins further towards the north where a basement is formed in a flexure limb. According to Jerzykiewicz (1969), the backslope does not coincide entirely with the strata, but cuts off a significant part of the raised beds.

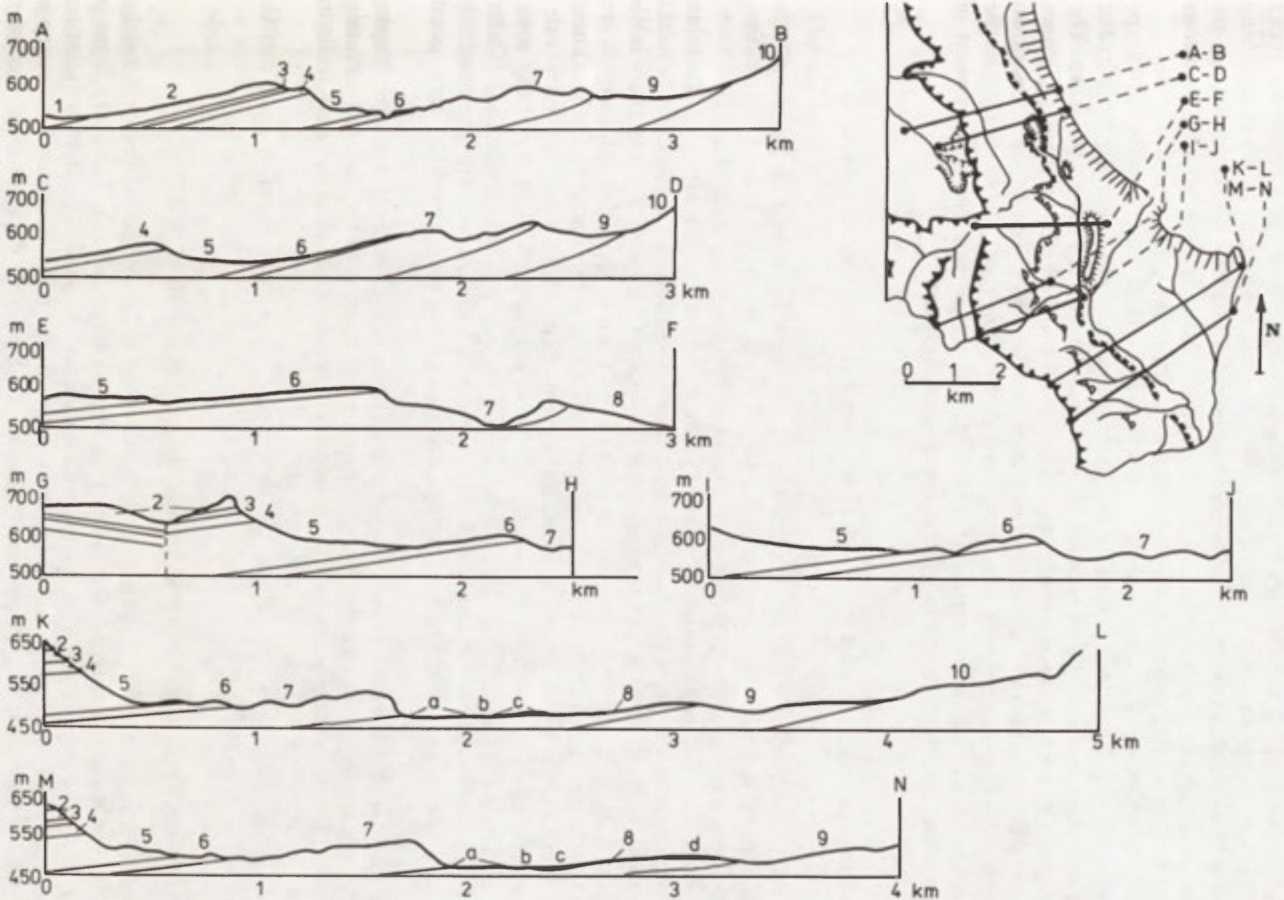

Fig. 3. Morphological profiles

1 — Turonian sandstones, 2 — Cenomanian sandstones, 3 — Cenomanian mudstones, 4 — Cenomanian tabular sandstones, 5 — Lower Triassic clayal sandstones, 6 — Upper Permian dolomitical arkoses, 7 — Lower Permian conglomerates, 8 — Lower Permian shales and sandstones, 9 — Lower Permian shales, 10 — Lower Permian volcanitoes

3. Cuestas of the south-eastern side of the Krzeszów brachysyncline

The sandstone cuestas of this region are much lower than those described above. Their relative altitude also decreases distinctly from 80 m at the southern edge to 30 m at the northern one. While the first fact results from a reduced thickness of the Cretaceous rocks within the Krzeszów brachysyncline (Jerzykiewicz 1969), the latter is related to the diminishing participation of the Triassic sandstones in the creation of the cuesta front. The sandstones are gradually wedging out along this scarp and they vanish finally under a cover of Cretaceous layers just near the northern border of the study area (Fig. 2).

The Cretaceous sandstone cuesta continues to be a well-marked feature in the landscape further to the north too. Despite the fact that the Cretaceous rocks are underlain there by the Lower Permian conglomerates which are also resistant enough to build their own scarp. The backslope of the Cretaceous sandstone cuestas in the region are typical intersecting planes running across different strata of the Cenomanian and Turonian age. The valley network on the backslopes consists of moderately dense localized dells. Their direction is parallel to the strata inclination. Both facts are determined by the existence of a well-developed system of joints allowing the good permeability of the Cretaceous basement.

B. THE LOWER PERMIAN CONGLOMERATE CUESTAS AND THE UPPER PERMIAN ARKOSE SCARPS

The forms are described commonly, because there is no stratum separating the conglomerates and dolomitical arkoses. This determines the strict correlation in the spatial distribution of the scarps or even the joint creation of a cuesta with composite face (Fig. 1).

The resistance of the conglomerates results from the high participation of Permian volcanitoe fragments in the petrographic-mineralogical composition of the rock (Grocholski 1973). Since the lithofacial structure of the conglomerates varies very strongly and there is no regular system of joints in this rock, the relief forms built on the outcrops of conglomerates are very different in shape. The existence of the conglomerate cuestas is facilitated by the weakness of the underlying shales which made possible the development of an extensive inner vale.

The Upper Permian dolomitical arkoses are the most resistant rocks in the study area. They have a high coherence, mainly due to the strong dolomitical and calcareous binder forming up to 70% of the mass (Grocholski 1973). A system of 2–40 cm wide crevices resulting from jointing widened by underground water dissolving the rock provides very good permeability.

This extremely high resistance of the arkoses allows them to create distinct scarps, despite the lack of an underlying weak stratum. These forms are not cuestas in the classical meaning (Schmitthenner 1954), but merely resistance scarps which normally develop in valley walls in the presence of intensive erosion (*Resistanzstufen* – Büdel 1957).

Like the Cretaceous sandstone scarps, the Lower and the Upper Permian ones can be divided into three groups too.

1. The scarps between the Golińsk village and Mieroszów town

The conglomerate scarp here runs along the monoclinal valley of the Ścinawka Kłodzka River (Fig. 1). The scarp is deprived of a footslope along its whole length. The scarp front is either undercut by the river channel or it ends with a distinct knickpoint in an accumulation plain of the valley. The entire slope face is formed entirely in the conglomerates. This scarp is an example of a valley step. The origin and dynamic of such forms depends on the erosive activity of a river flowing at the forefront (*Talstufe* – Brosche, Schultz 1972).

The valley step has a relative altitude of 40–60 m. The slope face is very steep (30–35 degrees) in those places where the river runs directly at the foot of the slope. There, two slope segments can be distinguished. The upper one, slightly dipping and undulated by dells, is cut off by the lower one, which is much steeper and plain. The backslope of the valley step is intensively intersected by deep stream valleys.

Arkose resistance scarps are to be observed only in the northern part of this region. In the south, outcrops of the arkoses are included in the long footslope of the Mieroszowskie Ściany scarp.

2. The composite cuesta

The form is built of the Lower Permian conglomerates and the Upper Permian dolomitical arkoses (Fig. 3, profile E–F). The latter outcrop at the broad backslope and in the most upper part of the slope face which has a lower segment made of conglomerates. The lower segment is carved into many promontories by long stream valleys. The deep valley of he main stream in the area isolates a huge fragment of the conglomerates which forms an elongated outlier. A smaller outlier is to be observed somewhat farther towards the north too (Fig. 2).

3. The conglomerate cuesta in the northern part of the study area

The cuesta here is 20–40 m high and its slope face has a distinct convex-concave shape. The slope face is devoid of the dells which are so common in analogous forms in the above described areas. On the contrary, the backslope is intensively undulated by dells and shallow valleys. Two sets of the amphitheatral valley heads have developed in an area where the watershed between the Bóbr River and the Ścinawka Kłodzka River crosses the cuesta.

III. THE MORPHOGENESIS AND THE MORPHODYNAMIC OF THE ESCARPMENT IN THE STUDY AREA

The first problem demanding consideration is whether evidence of a retreat of the scarps can be found. River terraces at a cuesta forefront have been used in some cases as such evidence (Schmitthenner 1954; Barth 1975). This is possible to analyse in the study area only for a monoclinal stretch of the Ścinawka Kłodzka River valley, where 4 terraces of a different age exist (Fig. 2).

The lowest one, Holocenian, lies 1–1.5 m over a channel bottom on a flood plain of the river. The next one is situated at height of 2–3 m and consists of periglacial gravels dated on the Weichsel glaciation (the North-Polish glaciation) (Krechowicz 1968). Both are underlain by a rocky basement and cover a flat valley floor.

Two other terraces placed on the gently dipping left valley side are made of gravels, too. One of them extends along the whole monoclinal stretch with the height decreasing from 20 m in the south to 13 m in the north. The terrace originates from the Saale glaciation (the Middle-Polish glaciation) (Grocholski 1973; Krechowicz 1968).

The highest terrace is preserved only as a small patch situated on top of a low hill, 25–36 m over the channel bottom. There are many views on its age. The most probable estimation is that of Jahn (1960) dating the form on the Elster glaciation (the South-Polish glaciation). The results of the analysis of gravel samples seem to confirm this opinion. Two samples of 100 gravels each have been collected, one from the 13–20 m terrace another from the 25–36 m terrace. A petrographic composition and evidence of the weathering of gravels has been compared (Table 1).

The considerably bigger number of quartz gravels in the highest terrace suggests that the material has been redeposited from preglacial river sediments. They consist almost entirely of quartz and are common in adjacent big valleys, but lacking in the Ścinawka Kłodzka River valley (Berger 1931; Szczepankiewicz 1953).

TABLE 1. Petrographic composition of gravels (20 mm) from the Saalian terrace (13–20 m) and the Elsterian terrace (25–36 m)

	Saalian terrace	Elsterian terrace
Permian volcanitoes	80%	72%
Quartz	9%	24%
Cretaceous sandstones	9%	4%
Permian conglomerates	2%	0%

Far more gravels are cracked and have sharp edges in the highest terrace than in the 13–20 m terrace. That could be proof of a longer post-sedimentary weathering of the material of the highest terrace. The facts show that the 25–36 m terrace should be younger than the preglacial deposits, but significantly older than the Saalian terrace. The conditions under which its sedimentation occurred must have been similar to those causing the sedimentation of the Saalian gravels. This was the case in the Elster glaciation. In both periods the icesheets were stagnating in their maximal extent only several kilometres to the north of the study area (Jońca 1975).

If such an estimation is accepted, the following order in the terrace system is to be seen: the older the terrace, the higher it is over the channel bottom and the farther from the valley step. This could be the result of a gradual deepening of the valley and the simultaneous retreat of the valley step in the Middle and Late Pleistocene. The fact is additionally confirmed by the lack of both higher terraces on the right side of the valley. Here, the terraces must have been destroyed by the lateral erosion of the Ścinawka Kłodzka River sliding dipward on the monoclinal lying stratum (Fig. 4).

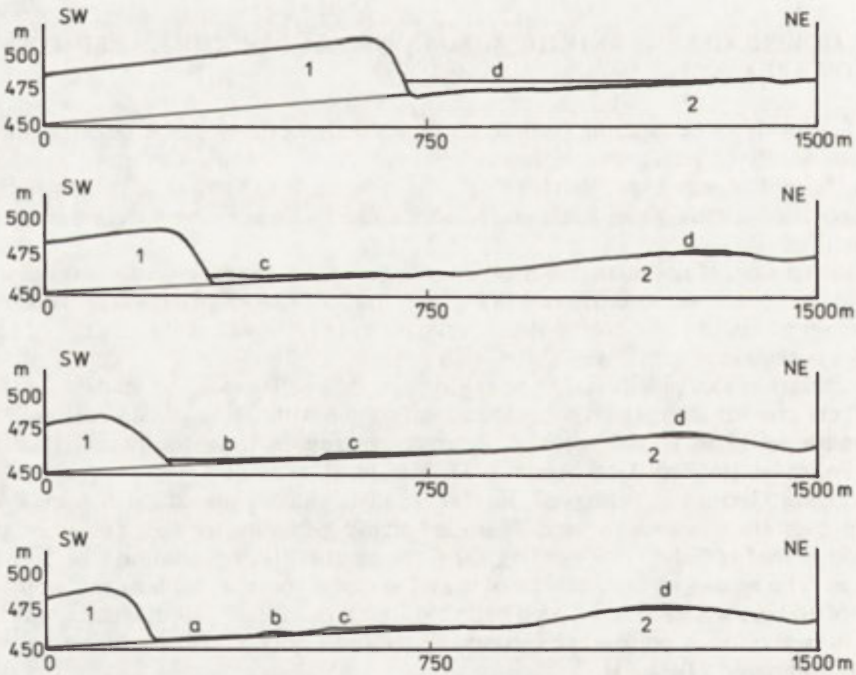

Fig. 4. Reconstruction of the development of the monoclinal valley and the valley step on the ground of river terraces

1 — Lower Permian conglomerates, 2 — Lower Permian shales and sandstones. River terraces: a — Holocenian terrace, b — Weichselian terrace, c — Saalian terrace, d — Elsterian terrace

The process also seems to have been continuing in the Holocene. The Weichselian terrace has been completely removed from the forefront of the valley step in its northern and middle parts. In the southern one, river action is destroying the two remaining patches of this terrace.

Where the river channel runs directly at the foot of the step, the slope is undercut so that a very steep segment contrasting clearly with a flatter upper one came into being (Fig. 3, profile K-L). The steeper segment could be created only after a period of accumulation of the Weichselian terrace as a result of the retreat of the lower part of the step.

The above conclusions are in accordance with the theories on the development of cuestas and valley steps in the Quaternary (Schunke, Spönemann 1972; Tricart 1951). Under periglacial conditions in cold periods, the rivers were overloaded with sediments and have accumulated thick alluvial valley fillings covering, an important argument for morphodynamic contact between resistant and weak strata (Blume 1971). Since river erosion was feeble or even missing, scarps have been becoming flat under the influence of slope processes. The rivers started to erode only when the climate changed to a warmer one. At first they removed their own deposits and then began to follow the direction of the dip in the rocky basement and finally to undercut the scarps.

Exact information on the scale of retreat can not be gained from the present spatial distribution of the terraces. The width of the past accumulation plains and the previous position of the valley step are hard to discover. Nevertheless, some estimations can be made.

Assuming that the foot of the valley step was situated at the time of the accumulation of the Saalian terrace at the level at which the rocky basement of the terrace is now lying (5–8 m over the channel bottom), and taking into consideration that the strata inclination is 10–12 degrees in the valley (Krechowicz 1968), the step front should be placed 50–150 m further north-east in the Saale glaciation (Fig. 4).

The facts, that the elevation of the top of the terrace declines from 20 m at the south to 13 m at the north and that the width of the valley decreases rapidly in the same direction, indicate that both processes, the valley deepening and the retreat of the step, are strictly correlated and that they have reached the lowest intensity in the northern stretch of the valley.

An estimation is more difficult for the highest terrace, since this form has been strongly transformed by post-sedimentary degradation. There is no information on the primary thickness and height of the terrace. Assuming that the thickness was similar to that of the Saalian terrace, the front has retreated about 200–400 m since the Elster glaciation.

Similar analysis can not be made for the Cretaceous sandstone scarps, since there are no river terraces at their forefronts. The only evidence of their retreat is here provided by the amphitheatral valley heads situated at the backslope south-west of the Kochanów village (Fig. 1).

A contemporary active valley head intersects the rocks which are part of the Łączna brachyantycline. The form is 350 m in diameter, 50–60 m in depth with 25–35 degree steep walls. While the bottom of the head is composed of tabular sandstones, mudstones and Upper Cenomanian sandstones appear in the walls. The impermeable mudstones force the ground water circulating in the overlying rock to form springs. These are causing a backward erosion.

A similar sequence of rock outcrops is to be seen slightly to the north between the dip-parallel scarps running across the backslope (Fig. 2). The scarps are the remains of two huge valley heads. Only the lack of back walls distinguishes them from the present active head. The destruction of those valley heads could not be the result of backward erosion only. Since while the forms have been approaching the slope face, their drainage areas have been diminishing and erosion holding up. The final stage of the destruction of the heads must have arisen from the retreat of the Cretaceous sandstone scarp (Fig. 5).

Because the basement here is made of rocks with a regular jointing, frost wedging caused rockfalls and rockslides can be seen as a mechanism of scarp retreat (Blume 1971). The common presence of block fields on the slope faces of the sandstone cuestas confirms this fact. If the process of creation of the blocks has been repeated several times during the cold periods of the Pleistocene, the few hundred metre long retreat required for the acceptance of the

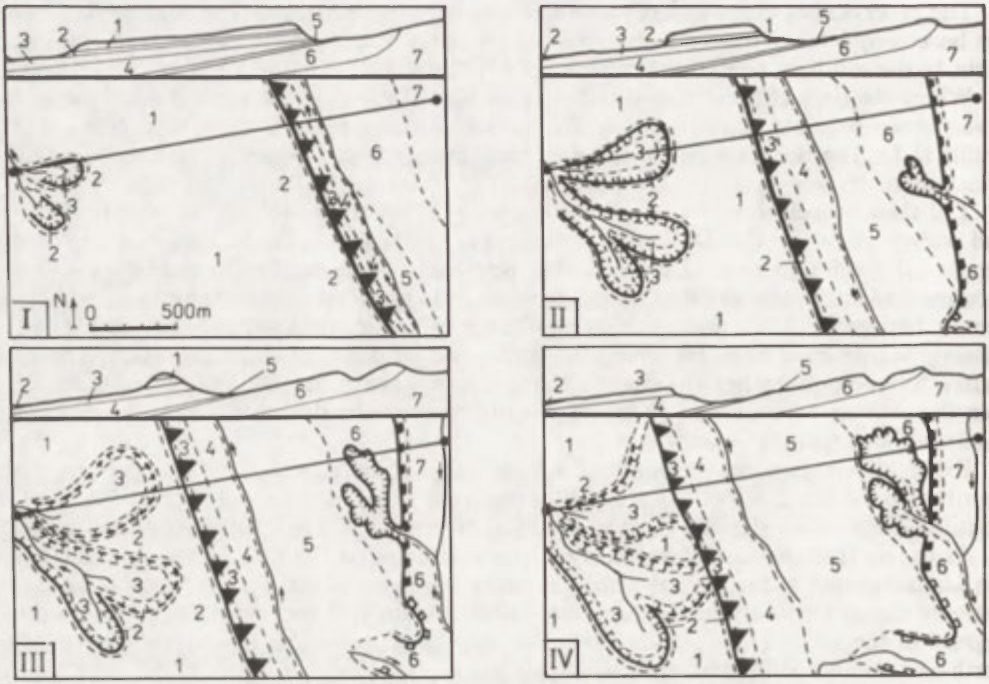

Fig. 5. Stages of development of the amphitheatral valley heads at the backslope of the Cretaceous sandstone scarp

1 — Cenomanian sandstones, 2 — Cenomanian mudstones, 3 — Cenomanian tabular sandstones, 4 — Lower Triassic clayal sandstones, 5 — Upper Permian dolomitical arkoses, 6 — Lower Permian conglomerates, 7 — Lower Permian shales. Explanations of morphological symbols see Fig. 1

development of the valley heads presented by Fig. 5, could have taken place. The straight-line course of the sandstone scarps proves that if a retreat occurred, it must have had a similar intensity in all of their parts.

The facts, that the Lower Triassic sandstones are entirely included in the cuesta faces and that their outcrops end on the axis of the valleys situated at the forefronts (Figs. 2 and 3), lead to the conclusion that during the retreat, the clayal sandstones must have been removed more or less as quickly as the Cretaceous ones. The retreating scarps have been exhuming an infra-Triassic surface (Figs. 1 and 5). Such a process can easily occur if the underlying rocks are extremely resistant (Tricart 1976) like the dolomitical arkoses in the study area.

The exhumed surface is well preserved on the backslope of the conglomerate cuesta in the vicinity of the Kochanów village. It is almost completely absent on the forefront of the Mieroszowskie Ściany cuesta. The surface here was intersected at the same time as the retreat by streams having a base level situated much lower than the surface.

If the retreat of the Cretaceous sandstone scarps was intensive and occurred throughout the Pleistocene or even earlier too, a major part of the landscape between the Permian volcanitoe ridge and the fronts of the scarps is a result of the dissection of the exhumed surface. The acceptance of this opinion allows the understanding of the two functions played by the dolomitical arkoses in the structure of the conglomerate cuesta near the Kochanów village. Outcrops of the rock are included entirely in the backslope plain together with less resistant conglomerates in the northern and middle parts of the cuesta. Just several dozen metres to the south, the arkoses start to be a major part of the steep slope face of the same form (Fig. 3, profiles A-B, C-D, E-F).

The arkoses play the latter function only within the Ścinawka Kłodzka River basin where evidence of the strong influence of river erosion on the relief exists. The 15–20 m thick cover of the arkoses protects against the further destruction of the underlying conglomerates which are eroded after the retreat of the sandstone scarp from the area. The origin and development of the conglomerate cuesta and the composite cuesta are similar to the schema described by Büdel (1957). A scarp is evolving as a result of the young erosion of the edges of a much older plain. But a Tertiary peneplain from Büdel's theory is here replaced by the exhumed infra-Triassic surface.

Both the outliers situated at the forefront of the composite cuesta must have arisen through a strong vertical erosion of a stream intersecting the edge of the surface. The fact, that the course of the stream valley does not follow the outcrops of the weaker shales on the whole stretch, suggests that the valley has kept at least in some of its parts to the course it had before the stage of dissection of the surface. The process of isolation of the outliers could have been facilitated by the unusually great inclination (35 degrees) of the conglomerate stratum in those places. A. Jahn (1946) has proved in the theoretical model that a high inclination of monoclinal strata can cause the development of a valley despite the resistance of the strata.

Similar reasoning is to be used when aiming to explain the sliding down of the Ścinawka Kłodzka River in the monoclinal stretch. The sliding occurred on the basement made of the weak shales. The river had to undercut, at the same time, the conglomerate scarp built of the rock which is more resistant than the shales. But not only 10–12 degrees of strata inclination could cause the river to continue following a contact shales-conglomerates, instead of cutting in the basement vertically. Simulations of landscape development by Ahnert (1976) showed that an accordant slope tends to be degraded several times slower than a slope intersecting strata at a big angle. This could be the next factor which created equilibrium between the rate at which the valley step was undercut and the rate of the valley deepening on the monoclinal stretch.

Evidence of the development of the monoclinal valley exists only for the period from the Elster glaciation to the present. But since there are no remains of buried Late Pliocenian canyons in the Ścinawka Kłodzka River valley, though they are common in adjacent valleys (Szczepankiewicz 1953; Walczak 1952), it is to be accepted that the energy which has been expended elsewhere for vertical dissection was used in the monoclinal valley for dipward development.

Evolution of the valley could thus have taken place throughout a long period starting in the Late Pliocene. The evolution was presumably accelerated by the tectonic movements which occurred at the mouth of the Ścinawka Kłodzka River in that period (Berger 1931; Jahn et al. 1984).

A lowering of the base levels of the tributaries of the Ścinawka Kłodzka River resulted from the development of the monoclinal valley. That has resulted in the creation of a dense network of dells and in the steepening of the scarps. The fact, that the landscape has been shaped in the Ścinawka Kłodzka River drainage basin under conditions of high relief energy (*Relief-energie* – Schunke, Spönemann 1972), is proved by the existence of many water gaps and the capture of the most upper part of the I-stream by the II-stream flowing in that drainage basin (Fig. 1).

The capture occurred because the valley situated at the forefront of the conglomerate cuesta is placed 40–50 m lower than the I-stream valley running along the sandstone scarp. The II-stream flowing to the first of the valleys has, in such a situation, a low lying base level and a steep slope. Both are securing energy for intense headward erosion. That erosion could be additionally strengthened by the feeding of the stream by groundwater being drained not only from the nearby neighbouring backslope, but from the highly placed sandstone scarp, too.

The differences in the elevation of the main valleys also caused the development of another water gap, north of the Kochanów village (Fig. 1). One of the valleys running dipward across the backslope of the sandstone cuesta has dissected the basement so deeply that this has created a connection between the valley at the front of the scarp and the 40 m lower lying bottom of the Krzeszów Basin. The mechanism of that dissection could be the same as that of the evolution of the amphitheatral valley heads.

Local differences in the shapes of the relief forms in the study area result not only from the variability of geological features, but from the different intensity of erosion in the basin of the Bóbr River and in the Ścinawka Kłodzka River basin.

Both factors also caused important differences in the evolution of the conglomerate scarps in the Pleistocene. The valley step was cyclically undercut in warmer periods and was becoming flatter in the colder ones. That means that an alternating degradation had taken place in the form known from other regions (Mortensen 1947; Tricart 1951). On the contrary, the conglomerate cuesta in the northern part of the study area shows no evidence of such a process. Degradation by periglacial mass movements has created its shape with many dells on the backslope and the concave-convex slope face. Most probably, the course of the cuesta has not substantially changed since the form was dissected from the exhumed infra-Triassic surface.

The block fields of the Weichselian age covering the sandstone scarps prove that the present shape of the scarps was formed under the periglacial conditions of the last glaciation. The brief Holocene had little influence. According to laboratory analysis, only the periglacial sediments on steep slopes have been transformed by selective erosion accelerated after deforestation.

Unlike the conglomerate valley step, the sandstone scarps retreated and oversteepened in the cold periods of the Pleistocene and were much less active in the warmer ones. River erosion could only have had an insignificant effect on the scarps in the latter periods, since there are only small streams at their fronts.

The backslopes of the sandstone cuesta in the vicinity of Kochanów village, as slopes localized closest to the centre of the syncline, should be the oldest elements of the landscape in the study area. Their surfaces intersecting a few Cretaceous strata probably originate from the last of the three stages of the creation of the Tertiary relief generations in the Sudetes. That stage occurred during the Pliocene (Jahn 1980).

IV. CONCLUSIONS

The great variety of escarpment forms evolved in the study area as a response to the complicated geological structure and to the unequal relief energy in the different drainage basins. While Cretaceous sandstone cuestas and the Lower Permian conglomerate cuesta exist in both main drainage basins of the region, the occurrence of the dolomitical arkose resistance scarps and the valley step is limited to the Ścinawka Kłodzka River basin where evidence of strong erosion exists.

Observations from the area suggest that a primary factor determining the role of rock in the escarpment relief is permeability. Both the most resistant rocks here have a very high permeability provided either by regular jointing (tabular sandstones) or by crevices (dolomitical arkoses).

The fact, that the Ścinawka Kłodzka River has been sliding down on the stratum of weak shales, proves that not only the geological characteristics of the basement determine the rates of degradation of the slopes in the monoclinal structure, but the angle at which the slopes intersect the strata too. This confirms the conclusions drawn from the theoretical models by Jahn (1946) and Ahnert (1976).

The terrace system at the forefront of the valley step provides evidence of the retreat of this step. This retreat amounts to 200–400 m since the Elster glaciation (South-Polish) and 50–150 m since the Saale glaciation (Middle-Polish). The Cretaceous sandstone scarps were retreating in the Pleistocene too. Huge amphitheatral valley heads were destroyed at the backslope of the scarps during the retreat.

The retreating cuestas left an exhumed infra-Triassic surface at their forefronts. Presumably, all of the relief forms between the sandstone cuestas and the Permian volcanitoe ridge were dissected from the infra-Triassic surface in the Quaternary and the Late Pliocene. Older relief forms are the backslopes of the sandstone cuestas only.

Three schemes of development of the scarps in the Pleistocene can be created:

– rapid retreat caused by river erosion in warmer periods and stagnation and flattening of the scarps in the cold ones (the valley step);

– retreat resulting from frost-wedging caused rockfalls and rockslides in cold periods and stagnation and degradation of the block fields on the slope faces in warmer ones (the sandstone scarps);

– stable development throughout the Pleistocene with strong influence of periglacial mass movements on slopes.

The scarps of the area have not been developing in accordance with Schmitthenner's theory (1954), since the retreating sandstone scarps leave at their forefronts an exhumed surface not a plain intersecting weak and resistant strata and because of the minor importance of the action of springs for the mechanism of the retreat. The significant backward movements of the scarps contradicts the principles of Büdel's theory (1957).

Observations made in the study area confirm the opinion that an analysis of the morphogenesis and morphodynamic of cuestas must be made with the exact consideration of the various influences of different local factors and that general uniform theories can hardly be matched in such an analysis (Schunke, Spönemann 1972).

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GENESIS AND AGE OF THE TERRACES OF THE DNIEPER RIVER BETWEEN ORSHA AND SHKLOV, BYELORUSSIA

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ABSTRACT. The Dnieper river valley, in spite of being and old one, has only young terraces preserved in the section between Orsha and Shklov: tow Vistulian-Young Pleniglacial terraces and two levels of the Late Glacial-Holocene flood plains. With a large horizontal stability of the channel, the modelling of flood plains resulted only form a vertical accretion of flood sediments. Phases of the intensified activity of the Dnieper which have been identified on the border of the Atlantic and Subboreal and during the Subatlantic about 1000 years BP manifested themselves here by changes of the type of sedimentation on flood plains. These phases are conditioned by the climate, although in the younger one a remarkable influence of human activity is evident.

INTRODUCTION

Research in the Dnieper walley is a continuation of the studies on Byelorussian rivers of the Black Sea catchment area, which, in the period of the Last Glaciation carried away waters of the melting ice sheet south-bound (Kalicki 1991b). In the paper we present the results of our own research compared with previous studies scattered in numerous sources. The results of the former research were systematized from the point of view of the evolution of the Dnieper valley between Orsha and Shklov in Vistulian and Holocene.

The research was carried out in summer 1988 during the geological expedition of the Institute of Geochemistry and Geophysics, Byelorussian Academy of Sciences in Minsk led by Dr M. A. Valchik. Similarly, like on the Berezina river, it was limited to description and study of outcrops on the banks of the channel and in gravel-pits. Malacological analyses were made by A. F. Sanko whereas ^{14}C dating in the Institute of Physics, Silesian Polytechnic in Gliwice. We made the aerometric analyses of sediments and analyses using Tokarski's method of the humus content. Parameters of grain-size distribution were estimated by Folk and Ward's method (1957).

We would like to express our gratitude to all colleagues from the expedition for their help and cooperation during the researches, to Professor S. W. Alexandrowicz for the discussion on malacological spectra and to Professor L. Starkel for the all-round discussion.

LOCATION

Dnieper river being 2201 km long (700 km of which lie on the territory of the republic) an having the drainage area of 504 000 km^2 (105 000 km^2 in the republic) belongs to the biggest rivers of Byelorussia (Matveyev et al. 1988). Its springs are situated on the Valdai Plateau and it has its estuary in the Black Sea near Kherson.

The area of research included the c. 75 km section of the river between Buroe (upstream Orsha) and Shklov (Figs. 1, 2). In this section, the drainage basin only slightly increases from 18 000 km^2 in Orsha to 20 000 km^2 in Shklov. The mean river gradient is 0.12–0.15‰ and the channel width is 80–150 m. The valley is very narrow and its width oscillates between 0.5–1.5 km. This is the reason why the river can not freely meander, like for instance below Mogilev, and the meander indicator is very small: 1.1–1.3 (Matveyev et al. 1988).

Upstream of Orsha the Dnieper valley has a parallel course and continues slightly south-bound from the Orsha Plateau which marks the boundary of the maximum extent of ice sheet in Vistulian (Valdai). The valley has a slightly sinuous course here – its shape resembling valleys of "underfit" rivers with confined meanders (Fig. 1; cf. Lewin, Brindle 1977). Downstream of Orsha the valley hass a meridional and straight course. There are two "dead valleys" in this section (Fig. 2). The Dnieper valley dissects here the fluvioglacial-moraine Orsha-Mogilov Plain dating from the Oshmyanski stadial of the Sozhski Glaciation (Middle Polish).

This plain is covered by loesses and loess-like deposits, 3–12 m thick (Motuz 1986) being dissected by a complicated network of deep (15–20 m) ravines, which are best developed on the slopes of the Dnieper valley. A very sudden and distinct change of the direction of the Dnieper in the region of Orsha is explained by the neotectonic elevation of a local structure – the Goretski's structural peninsula included in intensively uplifted East-Byelorussian zone (Gorelik 1960; Goretski 1970; Levkov, Karabanov 1987). It has led to occurrence of the system of cataracts, so called Kobelyaki Porogi, connected with outcrops of Devonian limestones in the channel, 9 km upstream Orsha (Matveyev et al. 1988).

The Dnieper valley, situated within Oshmyany-Loev zone of concentration of lineaments (Levkov, Karabanov 1987) must have been the subject, in lower Quaternary, at least locally, of subsidence movement the evidence of which are thick series of alluvia, known from borings in Buroe or Selishche region (Meshcheryakov 1961; Goretski 1970). Only in Vistulian and in Holocene we observe the tendency to uplifting the region which has led to deepening of the valley and to straightening of the river (Nechiporenko, Pavlovski 1989; Pavlovets 1989).

HYDROLOGICAL REGIME

The river basin of Dnieper is located in the area of temperate continental climate, with a domination of western circulation. The mean temperature of January oscillates between −7.5 and −8.5°C, and the mean temperature of July oscillates between 17.5 and 18.0°C. Annual precipitation is 600–650 mm (*Priroda*... 1986).

Annual mean discharges of Dnieper in Orsha is 125 m^3/s (*Resursy*... 1966). Dnieper is characterized by the snow regime with a distinct snow-melt flood (Fig. 3). During such a flood discharges increase on the average to 937 m^3/s, and water level is 1.5–3.5 m higher. Yet, catastrophic floods (maximum discharge 2000 m^3/s in 1931), can raise water level even to 8 m, which leads to submergence of 6 m terrace. The flood starts most often at the end of March (24 III), reaches its maximum in mid-April (19 IV) and ends in May (31 V), i.e., it lasts, on the average, 69 days (44–100 days). There also happen floods connected with summer rains during which discharges can increase even to 713 m^3/s (1927). The mean period of ice lasts 111 days (41–147 days), it begins on 10 December and continues till the end of March.

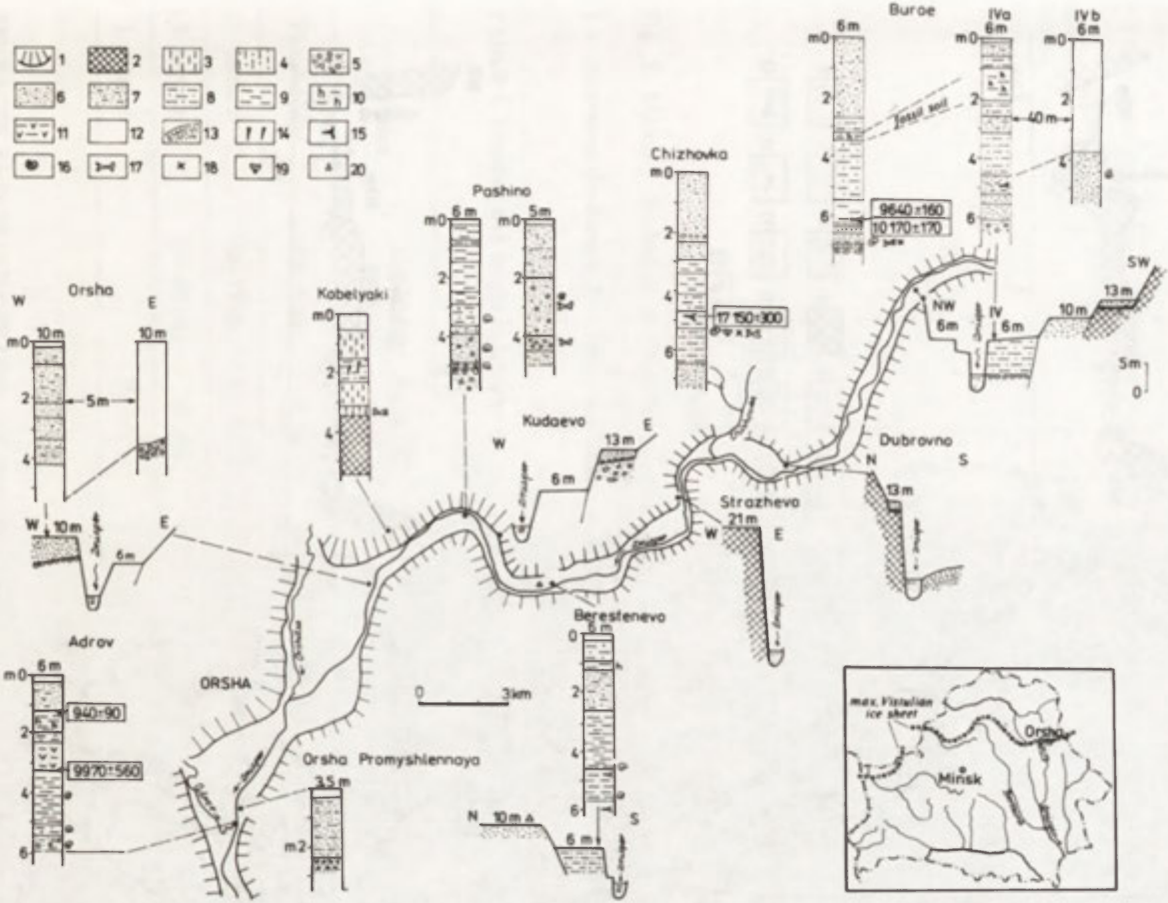

Fig. 1. Sketch of Byelorussia and map of the Dnieper valley between Buroe and Adrov

1 — valley side, 2 — till, 3 — loess, 4 — sandy loess, 5 — sand and gravel, 6 — sand, 7 — silty sand, 8 — sandy silt, 9 — silt, 10 — organic silt, 11 — peaty silt, 12 — soil, 13 — sandy alluvial fan, 14 — fossil ice wedge, 15 — organic detritus, 16 — malacofauna, 17 — bones, 18 — ostracodes, 19 — insects, 20 — archaeological site

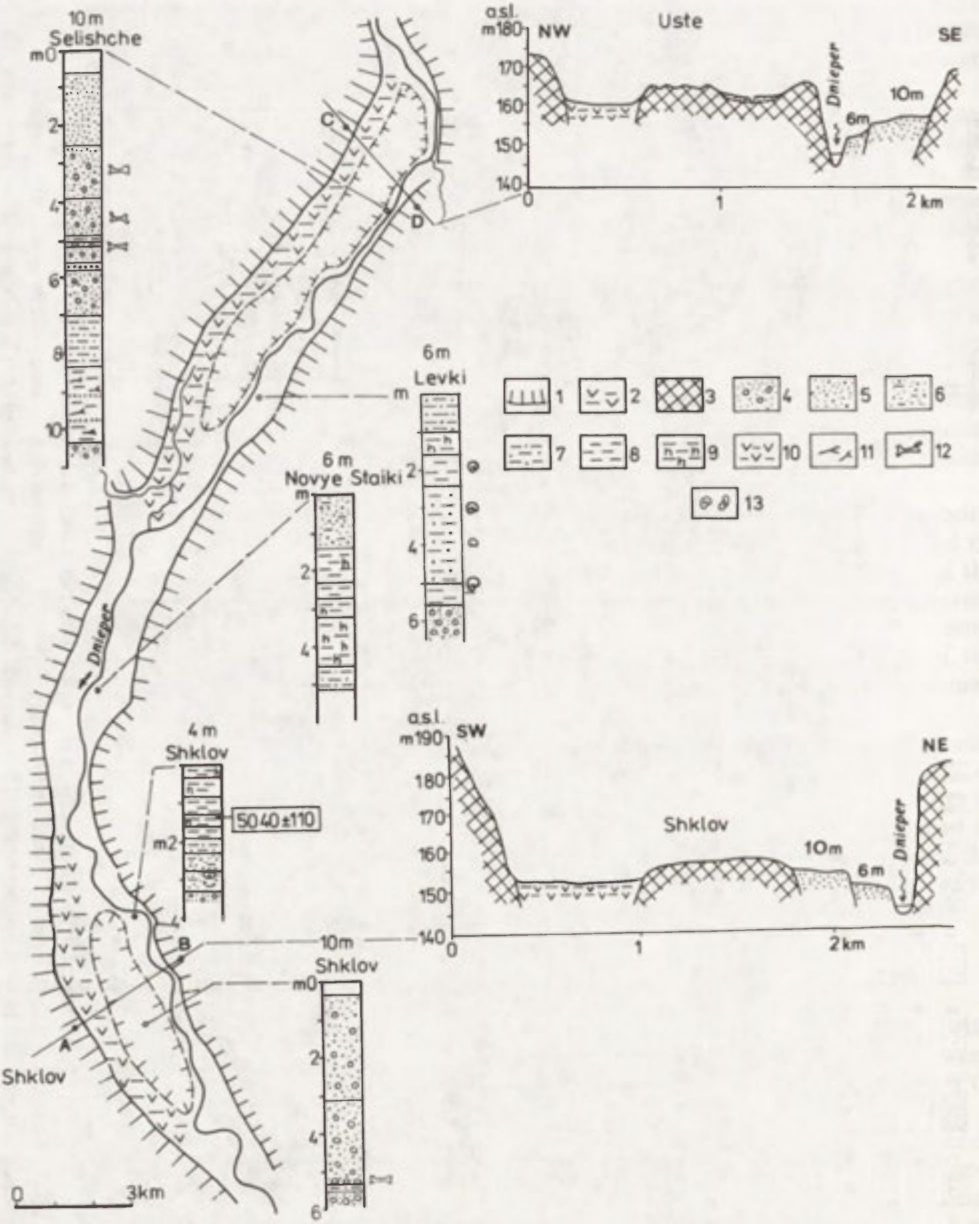

Fig. 2. Map of the Dnieper valley between Adrov and Shklov
1 — valley side, 2 — peat and clayey peat, 3 — till, 4 — sand and gravel, 5 — sand, 6 — silty sand, 7 — sandy silt, 8 — silt, 9 — organic silt, 10 — peaty silt, 11 — organic detritus, 12 — bones, 13 — malacofauna

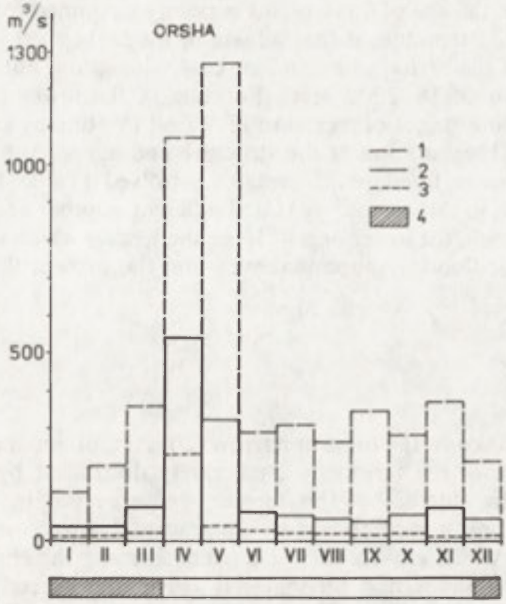

Fig. 3. Monthly discharges of Dnieper in Orsha during the periods: 1881–1922, 1924–1941 and 1945–1970 (by T. Kalicki)

1 — maximum discharges, 2 — mean discharges, 3 — minimum discharges, 4 — period with ice cover

TABLE 1. Number and height of terraces in the Dnieper valley between Orsha and Shklov according to different authors

Terraces	Lichkov 1928	Isachenko 1964	Tsapenko Mander 1968	Motuz 1972 1975	Voznyachuk 1973	Matveyev et al. 1988
Flood-plain a b	2.5–5 m	8 m	6–7 m	5–7 m	5–7 m 8–10 m	5–6 m
1. terrace	10–15 m	12–13 m	15–16 m	13–15 m	12–15 m	13–15 m
2. terrace	36–40 m	25 m	40 m	30–40 m	18–20 m	23–25 m
3. terrace		50–60 m				

TERRACES AND FLOOD PLAINS OF DNIEPER

In spite of the fact that studies in the Dnieper valley were carried out still in 19th century (Dokuchayev 1878), the opinions on the quantity and height of terraces in the region of research are very much different and contradictory (Table 1). It probably results from the fact, that some researchers took flat moraine plains for terraces of Dnieper. However, there is a complete lack of fluvial sediments on them, that is why they can not be treated as terraces. The outcrop in Strazhevo is instructive here (Fig. 1).

Also, the question of the age of flood plains is poorly examined. Although G. F. Mirchink (1926, 1927, 1928) already maintained that alluvia of the highest 30–40 m terrace interfinger with outwash aprons of the period of maximum Last Glaciation, but A. F. Sanko (1987) sees this connection on a lower, 18–20 m level. Forming of the lower terraces G. F. Mirchink connected with succeeding stages of recession of Würm (Vistulian) ice sheet and epeirogenic uplifting of the region. The question of the structure and age of 5–6 m terrace, regarded by many authors as Holocene flood plain, remains unsolved (Table 1).

During field research in this region, we stated different number and height of terraces. The highest terrace is 13 m high, the lower one — 10 m, the terrace which is 6 m high is submerged only during catastrophic floods (compare above), and the present flood plain forms narrow strips 4 m high.

13 M TERRACE

This terrace is preserved in the form of narrow (20–200 m), separate strips on both banks of Dnieper. The surface of the terrace is even, rarely diversified by dunes or thermokarst depressions (Buroe). The stucture of this terrace was observed in three sites (Fig. 1).

In Buroe, in a gravel-pit it is composed of fine-grained, brown-rusty sands of the thickness of 1.25 m in the bed of which gravels to 30 cm occur forming the channel pavement. Alluvia are lying here either on sands and fluvioglacial gravels, or directly on till.

In the gravel-pit of Kudaevo a profile of this terrace similar to Buroe is exposed. Alluvia with small thickness of 1.2–2.5 m are composed of yellow, fine-grained sands covering channel pavement. The thickness of alluvia increases in the direction of the river bed. The alluvia cover sands and fluvioglacial gravels within which single boulders occur.

In Dubrovno, opposite the mouth of Zadubrovianka a narrow (20 m) strip of the terrace has been preserved. In the outcrop along the distance of 250 m silt deluvia (0.2 m) covering fine-grained silt sands (1.1–1.3 m) with lag deposits in the bed, are exposed. The alluvia rest here on till erosional socle.

From the presented materials it is evident that 13 m terrace is of an erosional character, and the thickness of alluvia is very small and oscillates from 1.5 to 2.5 m.

10 M TERRACE

10 m terrace is the best preserved overflood terrace in the Dnieper valley. Its width oscillates between 50 and 100 m, disappearing in places (for instant near Kobelyaki Porogi). The terrace is composed of gravel-sandy channel sediments described by the authors on several sites. In the gravel-pit of Buroe it is built of yellow, fine-grained sands with an apparent thickness of 2.5 m. It has a similar structure in the region of Berestenevo (sands to 2.25 m), where there is a Mesolithic site, archaeologically dated at about 10 000 years (Fig. 1).

This terrace covers a large area in the region of Orsha (Fig. 1). In the outcrop, the upper 0.8 m is constituted of white-rusty silt sands. Below, to 2.1 m, yellow silt sands are deposited with horizontal bedding of rusty silts and ferrous sands forming rhythmic sequence. Below, to 2.6 m, there occur yellow sands with normal stratification, coarsening towards the bed, bedded diagonally, bed of which is sloping towards the river. Below these, again yellow, fine-grained sands occur (to 3.35 m), and in the bottom part (to 4.2 m) they are interbedded by thin (to 2 cm) inserts of silty sands. In another place a channel gravel pavement appears inclined towards the slopes of the valley.

On this terrace the sites earlier described are located. In Dubrovno, in alluvia of this terrace cryogenic involutions were stated (Sanko 1987). Under 2.5 m stratum of diagonally and horizontally stratificated sands and a thin insert of channel pavement, we find fine sands with a system of pseudomorphoses after ice-wedges (to the depth of 2.5 m) of repeated freezing, which is the evidence of the occurrence of large polygons (to 7 m). In the top part of these fine sands stretches almost a continuous stratum with involution-injectional cryoturbation of the thickness of 0.5 m. Clusters of large boulders cemented by clayey sediment occur in places. These boulders are often pushed out to 1 m upwards into covering them upper sands, and there they constitute a system of relic rubble rings. This site proves that alluvia were accreted in two phases, and the accumulation took place in a cold climate, the evidence of which are pseudomorphoses and cryogenic involutions.

TABLE 2. Molluscan assemblages from the profile Mitkovshchina (by W. M. Motuz)

E	Species	Number
10	*Lymnaea truncatula* (Müller)	1
11	*Gyraulus* cf. *acronicus* (Ferussac)	2
11	*Valvata piscinalis* (Müller)	3
11	*Pisidium* cf. *ponderosus* (Stelfox)	1
11	*P. henslowanum* (Sheppard)	106
11	*P. subtruncatum* (Malm)	20
11	*Sphaerium corneum* (Linne)	21
11	*S. scaldianum* (Norinand)	8
12	*S. rivicula* (Lamarck)	18
12	*Pisidium amnicum* (Müller)	1

E – Ecological groups of molluscs (Lozek 1964, Alexandrowicz 1987): 1 – forest species, 2 – snails inhabiting mainly forests, 5 – species of open environments, 7 – mesophile snails of moderately moist habitats, 8 – mesophile snails of moist habitats, 9 – higrophile species, 10 – species occurring in shallow, periodic water bodies, 11 – molluscs living in permanent bodies with stagnant water, 12 – molluscs occurrring in streams and rivers.

Similar duality of sediment of this terrace is observed in nearby Mitkowshchina. Top 3 m are composed of yellow, fine-grained sands, passing below into coarse-grained sands with gravels – grey, stratificated, in the bed of which a layer of channel pavement occurs. There malacofauna in sands with gravels which was marked by V. M. Motuz (Table 2) was found. This fauna is characteristic of a cold climate, indicating a basin of stagnant water, which seems to be odd taking the character of sediment into consideration. It is possible that we deal with a paleochannel of braided river. Below the channel pavement we find vari-grained sands with single gravels, bedded diagonally, with frost wedges. The thickness of sediments with pseudomorphoses reaches 2 m, and they cover 2 m layer of sands with gravels. It is not unlikely that in both of these sites we deal with a fossil terrace (old-Vistulian?) covered by younger sediments.

The site in Selishche gravel-pit is also interesting (Kalinovski 1983; Fig. 2). In the outcrop, upper 7 m are composed of sandy-gravel sediments in which on the depth of 2.5–5.0 m single bones of large tundra mammals were found: those of mammoth, horse, reindeer, hairy rhinoceros and musk-ox. Below the thin insert of sandy silts (5.0–5.2 m) numerous fragments (42) of bones of rodents were also found, mainly of lemmings and field-voles (57%) which is the evidence of periglacial climate, though there is also one specimen of *Microtus arvalis*, which now lives in Byelorussia (Kalinovski 1983). Below the sandy-gravel sediments, on the depth of 7–10 m, sandy silts occur. The palynological profile of these sediments reveals a small content of pollen, a high fraction of grasses (64%) and distinct blend of redeposited pollens of thermofil plants with arctic and steppe species (Sanko 1987). Also, studies of macrofossils from the inserts of detritus indicate the occurrence of two complexes of vegetable residua: periglacial one, with a remarkable diversification of species and redeposited one, containing remains of interglacial, mainly Eemian vegetation, coming from erosion of older sediments (Velichkevich 1982).

Also in the region of Shklov in sediments of this terrace bones of mammoths and rhinoceroses were found long ago (Polikarpovich 1932; Shcheglova 1963). Recently, on the depth of 5.3 m, on the contact of sandy-gravel sediments with underlying sandy silts, teeth and skull of mammoth were found (Kalinovski 1983; Fig. 2).

Sandy silts and fine-grained sands described as lacustrine-alluvial sediments were examined in detail on Chizhovka site (Fig. 1). It is not situated in the main Dnieper valley but in the valley of a small tributary – Chizhovka, 3 km south from terminal morains of maximum stadial (Sanko 1987). The profile is situated on the terrace elevated indeed 15–17 m above Dnieper level (Sanko 1987) but the true height of this terrace in relation to the bed of Chizhovka is 8–10 m and that is why it can be regarded as an equivalent of the 10 m terrace of Dnieper. Like in Selishche, silts are covered here by sandy-gravel alluvia. Paleobotanical studies of these silts (N. A. Makhnach, T. V. Yakubovska) showed, similarly to Selishche, a small content of pollen and the occurrence of two complexes of vegetable residua: cold, periglacial and warm, redeposited (Sanko 1987). Residua of lemmings and field-voles found in deposits (A. N. Motuzko) indicate tundra conditions. Also, several thousands of insects residua (V. I. Nazarov) among which tundra beetles prevail, indicate a very cold (mean temperature of July below 10°C, of January −26° to −30°C) and dry (200–500 mm per annum) climate. The main set of examined ostracodes (S. F. Zubovich) create characteristic for a cold climate *Ilyocypris* and *Candona*. Dating of the organic residua (17 150 ± 300 BP, Tln–329) confirms that they were assembled during exceptionally severe, periglacial climatic conditions, i.e. the maximum of the Last Glaciation (Sanko 1987).

10 m terrace is, therefore, an accumulative terrace created in a cold, glacial climate, after the maximum of glaciation. In some of the profiles (Dubrovno) two phases of alluvia accumulation occur which are divided by a stratum with cryoturbation and a channel pavement. It is a very interesting thing that on this level in the Dnieper valley there are two characteristic erosive rectilinear "dead valleys", to 1 km wide (Fig. 2). The first one stretches from Uste to Kopys along the distance of more than 14 km, the second one, in the region of Shklov, is 9 km long. They both are parallel to the present-day Dnieper valley. They are hanging valleys and are separated from the valley by moraine elevations (Fig. 2). At present they are covered by vast peatbogs. "Dead valleys" were probably formed simultaneously with the 10 m terrace, when Dnieper carried away waters of melting ice sheet south-bound, and having multiplied discharges, flowed in two valleys. After ceasing to be supplied with proglacial waters, the river cut in and abandoned one of the branches of the valley. Further evolution took place only in the active valley, where forming of the lower terrace occurred, whereas "dead valleys" were covered by peatbogs.

6 M TERRACE (HIGH FLOOD PLAIN)

6 m terrace is 100–200 m wide and stretches along the whole Dnieper valley in the entire examined section. It is regarded as a flood plain by many researchers (e.g. Motuz 1972, 1975; Matveyev et al. 1988). However, as hydrological data show, at present it is drowned only in periods of extreme floods. Characteristic of this terrace are common in its top silt deposits of large – several metres – thickness. We examined them in a few sites.

In Buroe, this terrace constitutes a plain a few scores of metres wide, distinctly sloping towards Dnieper (Fig. 1). In the outcrop Buroe 1 the top metre is composed of clayey muds and brown clayey sands with ferruginous concretions. Below, to 2.05 m, characteristic of this terrace layer of black-brown humus silts occurs, being probably a fossil soil. From 2.05–4.6 m

TABLE 3. Molluscan assemblages from the profile Buroe 1 and Buroe 2 (A. F. Sanko)

E	Species	Buroe 1		Buroe 2					
		sample							
		1	2	1	2	3	4	5	6
2	*Vitrea crystallina* (Müller)			1	1	2	1		1
5	*Vallonia pulchella* (Müller)	2	1	3	7	2		3	4
7	*Cochlicopa lubrica* (Müller)					1	2		
7	*Nesovitrea hammonis* (Ström)			1					
8	N. cf. *petronella* (Pfeiffer)	1							
9	*Succinea* cf. *putris* (L.)	1		1	1			1	
	Succinea sp.		3	2	10				
10	*Limnaea peregra* (Müller)	1							
	Limnaea sp.	1							
10	*Planorbis planorbis* (L.)		8						
10	*Valvata pulchella* (Studer)	34	24	16	49	72	50	20	18
11	*Acroloxus lacustris* (L.)	4	4		3	1			
11	*Armiger crista* (L.)	17	19	1	17	26	2	11	12
11	*Bathyomphalus contortus* (L.)	1	21						
11	*Bithynia tentaculata* (L.)	54	50	39	70	43	123	15	13
11	*Gyraulus albus* (Müller)	2							
11	*G. laevis* (Alder)	6	16	4	55	59	9	10	7
11	*Pisidium* cf. *casertanum* (Poli)							10	
11	*P. henslovanum* (Sheppard)	5	5				20		100
11	*P. moitessierianum* (Paladilke)					25			
11	*Pisidium sp.*	15	50	100	100	100	100	10	100
11	*Sphaerium corneum* (L.)						2		2
11	*S. lacustre* (Müller)	1							
11	*S. subsolidum* (Clessin)					2			
	Sphaerium sp.	5			5	2	7	1	
11	*Valvata cristata* (Müller)	30	106	3	32	20	3	7	8
11	*V. piscinalis* (Müller)	15	20	7	28	21	41	11	5
11	*V. piscinalis antiqua* (Sowerby)	5	2	1					
12	*Pisidium amnicum* (Müller)	7		22	84	65	85	50	15
12	*P. pulchellum* (Jerzyna)	1							
12	*P. supinum* (Schmidt)	75	10	200	500	500	300	500	400
12	*Sphaerium rivicola* (Lamarck)					1			

E – Ecological groups see Table 2.

we find sandy silts passing below into clayey sands, white with rusty spots. In the upper parts of these sediments black traces of roots can be observed. The bed of these sediments (4.6–6.1 m) is composed of fine-grained sands, growing more coarse towards the bed, white in the upper part, interbedded with silts, in which wood and shells of malacofauna are met. The very last sediments are sands with gravels showing already on the level of the river. However, 40 m downstream of Dnieper channel sediments (sands diagonally stratified) are elevated to 1.5 m above the river level (Buroe 2).

Malacological samples from the profiles Buroe 1 and Buroe 2 were taken from bed parts, from silts and sands (Table 3, Fig. 4). In both profiles water fauna dominates constituting about 90% of tanatocenoses. These tanatocenoses differ from one another in the share of the

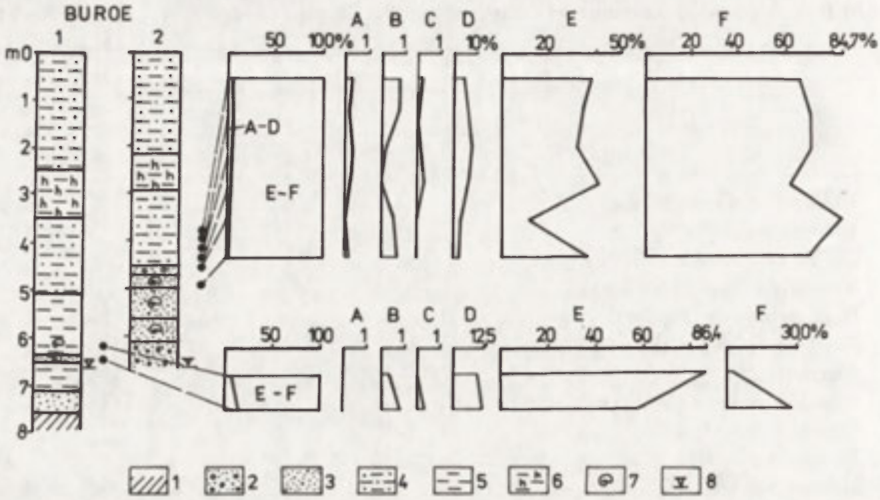

Fig. 4. Malacological diagram at Buroe site (by A. F. Sanko)
1 — till, 2 — sand and gravel, 3 — sand, 4 — sandy silt, 5 — silt, 6 — organic silt, 7 — molluscs, 8 — level of river. Type of malacofauna according to Ložek (1964) and Alexandrowicz (1987): A–D – land molluscs: A — forest snails, B — open-site snails, C — mesophile snails, D — hygrophile snais, E–F — fresh-water molluscs: E — stagnofile molluscs, F — reofile molluscs

stagnofile and reofile species, from which the first ones dominate in the silts of Buroe 1 and the second ones in the sands of Buroe 2. Water basin in the profile Buroe 1 was gradually losing contact with active river channel and it overgrew. The basin in Buroe 2 preserved constant contact with the channel and the evidence of periodic water flow is a sudden increase of reofile forms (to 85%) and simultaneous decrease of stagnofile forms. Contrary to Buroe 1, this basin was filled not with silts but with fine-grained sands and overgrew less thickly.

In both tanatocenoses in Buroe thermofil species are missing, and species living in a cold climate prevail – Paleoarctic, Holarctic, and Eurosiberian. Species characteristic for Late Vistulian also occur here in great number: *Gyraulus laevis, Valvata piscinalis, Pisidium sp.*, on

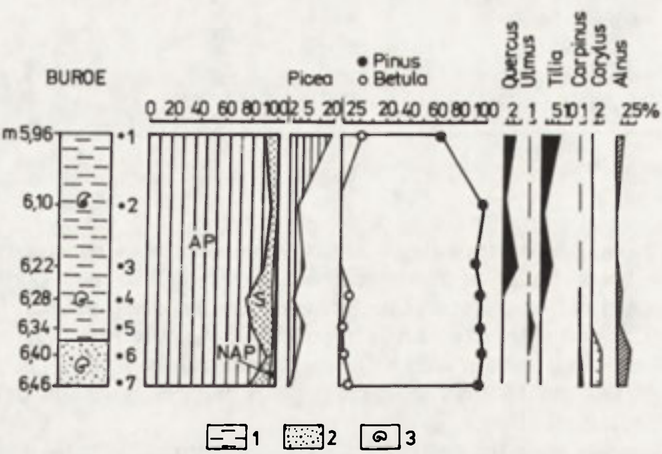

Fig. 5. Pollen diagram at Buroe site (by A. P. Shostak)
1 — silt, 2 — sand, 3 — malacofauna, AP — trees, NAP — herbs, S — spores

the other hand *Bithynia tentaculata, Valvata cristata* are also numerous, which is the evidence of gradual warming of the climate (cf. Alexandrowicz 1987). Characteristic is almost a complete lack of forest species and the occurrence of loess species, which were carried into the valley from the surrounding morain plateaus.

In Buroe the profile, having similar sequence of deposits, was already earlier described in 1975 (Kalinovski 1983; Sanko 1987; Fig. 1). From this profile, from the silts lying on the depth of 5.9–6.5 m, the palinological analysis was made by A. P. Shostak (Fig. 5). The pollen diagram is typical of a forested area with a very distinct domination of the *Pinus* (to 98%). Other trees met are *Betula, Picea, Quercus, Tilia, Alnus* and *Corylus*. Yet, these pollens do not constitute ordered accumulations and occur together with representatives of Paleogene vegetation (to 1.8%) and also with gymnospermous of Mesosoic and spores of Carbon. The last ones are met in big number on the depth of 5.96 m. The occurrence of old pollens and spores in forest diagram allows to assume that, together with them a part of Quaternary pollens might have been deposited on the secondary deposit, first of all foreign for the pollens spectrum of thermofil deciduous trees. If we exclude them we obtain pure pine diagram with a small admixture of birch and maybe fine-leaved species. Comparison of this spectrum with full Holocene diagrams proves its attachment to Preboreal period. Datings of wood found in grey silt in the bed of this terrace confirmed these conclusions and gave the results 9640 $\pm$ 160 BP (GIN–2308), and the second date of the same sample 10 170 $\pm$ 170 BP (GIN–2309) (Sanko 1987). In sands lying below the bones of rodents were found. The bones of field-voles decidedly prevail (*Microtus arvalis, M. agrestis*) being the representatives of early Holocene fauna, but the share of *Microtus gregalis* – a representative of tundra biocenoses is also remarkable, though residua of lemmings are missing (Kalinovski 1983). The studies of ostracodes gave similar results (Zubovich 1981), among which species characteristic for a cold climate prevail and the fauna of Buroe is similar to Lateglacial – Early Holocene complex from Volosovo (Sanko 1987). We can distinguish a few stages in the development of the water basin. In the initial phase it was relatively cold and deep. Later, the waters became moderately warm and rich what caused intensive development of organic life. The process of warming progressed further until the disappearance of the lake itself.

Similar profile to Buroe was described by us in Pashino (Pashino 2) (Fig. 1). Top sandy muds have here slightly smaller thickness (0.85 m). Organic, brown-black clayey silts constituting a fossil soil are deposited below these sandy muds on the depth to 2.5 m. Under them, to 3.15 m flood plain deposits occur, brown-rusty silts with yellow spots, which below pass into sandy silts and clayey sands, white with rusty spots and ferruginous laminas (to 3.9 m). The outcrop ends with fine-grained sands with gravels, white with rusty laminas in the bed of which, on the depth of 5.1 m, a channel pavement occurs.

Malacofauna occurs in the alluvia on the depth of 3.4–4.7 m (Table 4, Fig. 6). A few stages can be distinguished in its development. In the first phase it was a shallow water basin (water species above 90%), initially having quite remarkable contact with the river channel (reofile species to 42%). In the second phase a distinct reduction of contact with the river channel follows (decrease of reofile species to 20% and increase of stagnofile to 75%). In this period the basin begins to overgrow intensively and drys up temporarily. On the depth of 4 m a very distinct and sudden change and an almost complete vanishing of the basin (water fauna below 20%) is noticed. An intensive development of snails of open and mesofile environments follows, which indicates, that the flood plain of Dnieper was covered by moist and mid-moist meadows. This sudden change might be a result of drainage by an incised river with simultaneous gradual accretion of the flood plain. Another cause could also be climatic changes leading to a decrease of floods and to the lowering of the level of ground waters. This caused "drying" and change of sites on the flood plain of Dnieper. Pashino tanatocenose gives the evidence of a cold climate. There are no thermofil species, and the only forest species is *Vitrea crystallina* though its share is slightly bigger than in Buroe. The age of silts with malacofauna can be referred to the decline of Vistulian and the beginning of Holocene.

TABLE 4. Molluscan assemblages from the profile Pashino (A. F. Sanko)

E	Species	Sample						
		1	2	3	4	5	6	7
2	*Vitrea crystallina* (Müller)			2		2	15	5
5	*Vallonia pulchella* (Müller)			10		7	50	87
7	*Cochlicopa lubrica* (Müller)			3			19	20
7	*Limacidae sp.*			1			1	4
7	*Nesovitrea hammonis* (Ström)							2
7	*Punctum pygmaeum* (Drap.)					2	17	23
8	*Succinea oblonga* (Drap.)			1				
9	*Carichium minimum* (Müller)						2	
9	*Succinea putris* (L.)			2	4		4	6
	Succinea sp.	4		7				
9	*Zonitoides nitidus* (Müller)			1				
10	*Lymnaea truncatula* (Müller)			2			5	4
	Lymnaea sp.			2		13		
10	*Planorbis planorbis* (L.)			11	1	2		
10	*Valvata pulchella* (Studer)	31	7	107	22	3	1	
11	*Acroloxus lacustris* (L.)	1		2				
11	*Armiger crista* (L.)	2	1	83	58	4		2
11	*Bathyomphalus contortus* (L.)				2			
11	*Bithynia tentaculata* (L.)	50	97	340	92	17	2	
11	*Gyraulus laevis* (Alder)	2	1	103	72	4		
11	*Lymnaea stagnalis* (L.)			8	3			
11	*Pisidium casertanum* (Poli)						3	
11	*P. henslovanum* (Sheppard)				20			
11	*P. moitessierianum* (Paladilhe)				10			
11	*Pisidium sp.*	50	50	250	50	15		7
11	*Sphaerium sp.*			2				
11	*Valvata cristata* (Müller)			175	101	16	2	2
11	*V. piscinalis* (Müller)	16	4	37	9	1		1
12	*Pisidium amnicum* (Müller)	20	15	37	8	2		
12	*P. supinum* (Schmidt)	50	70	800	100	25	9	15
12	*Unio sp.*	1						

E – Ecological groups see Table 2.

Earlier examined profile of sandy alluvia situated about 250 m downstream indicates the similar age (Pashino 1) (Fig. 1). Numerous residua of rodents were found in sands with gravels on the depth of 2–4 m. Among them lemmings and a field-vole *Microtus gregalis* (35%) prevail representing the fauna of tundra biocenose. However, also Holocene species *Clethrionomys glaredus, Microtus sp., Arvicola sp.* occur in great numbers (Kalinovski 1983).

Almost analogous to Pashino and Buroe is Berestenevo profile (Fig. 1). Under sandy muds (to 0.9 m) a fossil soil, black-brown organic silts but of a small (30 cm) thickness, and below, to 2.6 m clayey sands in bed, very strongly ferrous occur. They cover a thick, over 3 m, series of flood plain deposits of white silts with single wood and malacofauna.

Malacological studies showed that water molluscs dominate almost in the whole profile, especially stagnofile species (to 73%) (Table 5, Fig. 7). The water basin in which silts sedimented was shallow and was filled with flood waters, initially with quite a strong relation with river channel (reofile species to 36%). This basin began to overgrow very strongly in the

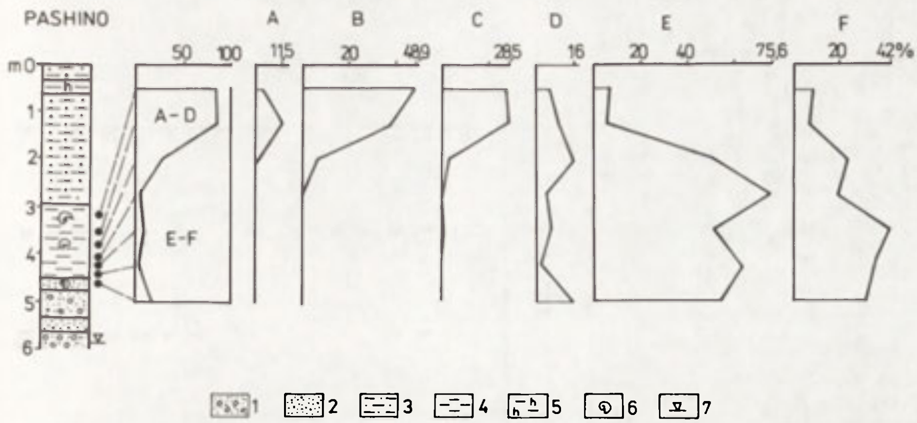

Fig. 6. Malacological diagram at Pashino (by A. F. Sanko)
1 — sand and gravel, 2 — sand, 3 — sandy silt, 4 — silt, 5 — organic silt, 6 — molluscs, 7 — level of river, A–F — type of malacofauna see Fig. 4

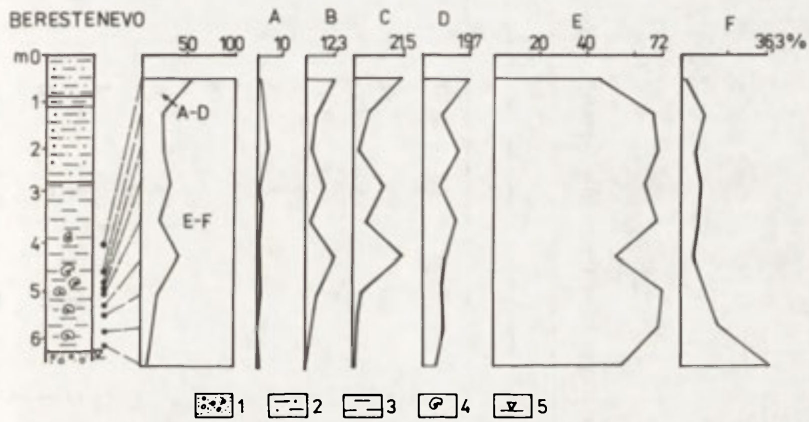

Fig. 7. Malacological diagram at Berestenevo (by A. F. Sanko)
1 – sand and gravel, 2 – sandy silt, 3 – silt, 4 – molluscs, 5 – level of river, A–F – type of malacofauna see Fig. 4

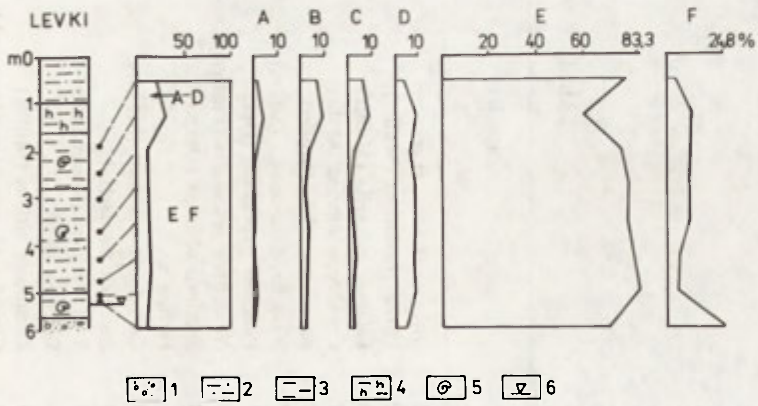

Fig. 8. Malacological diagram at Levki (by A. F. Sanko)
1 — sand and gravel, 2 — sandy silt, 3 — silt, 4 — organic silt, 5 — molluscs, 6 — level of river, A–F — type of malacofauna see Fig. 4

TABLE 5. Molluscan assemblages from the profile Berestenevo (by A. F. Sanko)

E	Species	Sample								
		1	2	3	4	5	6	7	8	9
1	*Discus ruderatus* (Ferussac)				1					
2	*Vitrea crystallina* (Müller)	6		5	2	3	2	7	7	2
5	*Vallonia pulchella* (Müller)	6	20	21	51	8	25	4	15	49
7	*Cochlicopa lubrica* (Müller)		11	9	32	4	34	5	8	27
7	*Euconulus fulvus* (Müller)		1							
7	*Nesovitrea hammonis* (Ström)	1			1	5	8		1	
7	*Punctum pygmaeum* (Drap.)		2	5	28	12			5	46
8	*Nesovitrea petronella* (Pfeiffer)				13					
8	*Succinea oblonga* (Drap.)	41			18				6	
	Vertigo sp.									1
9	*Carychium minimum* (Müller)			3	6	16		1		12
9	*Succinea putris* (L.)	13	2	10		7		4		5
9	*S.* cf. *pfeifferi* (Rossmässler)	22			4		3	2	15	
	Succinea sp.	2	10	4		14	11	8		25
9	*Zonitoides nitidus* (Müller)	1	18	2	10	2		4	3	27
10	*Gyraulus rossmaessleri* (Auerswald)	1							1	
10	*Lymnaea truncatula* (Müller)	6	23	5	19	12		10	4	5
10	*L. peregra peregra* (Müller)	1								
	Lymnaea sp.		4							

10	*Planorbis planorbis* (L.)		3	1						
10	*Valvata pulchella* (Studer)	161	16	8	1	3	10	1	3	5
11	*Acroloxus lacustris* (L.)	8	17	8	2	4	5	5	2	1
11	*Anisus spirorbis* (L.)	1	11			2				1
11	*Armiger crista* (L.)	98	190	62	51	60	67	38	33	37
11	*Bathyomphalus contortus* (L.)	1	1							
11	*Bithynia tentaculata* (L.)	1004	72	26	6	15	5	5	17	5
11	*Gyraulus laevis* (Alder)	179	170	63	43	15	27	18	25	7
11	*Lymnaea peregra ovata* (Drap.)		2							
11	*Pisidium henslovanum* (Scheppard)				10				10	
11	*P. moitessierrianum* (Paladilke)	20		20	5	65		10	15	25
11	*P.* cf. *subtruncatum* (Malm)						4			
11	*Pisidium sp.*	250	85	100	100	100	100	50	100	100
11	*Sphaerium solidum* (Normand)	11		10				2		
11	*Sphaerium sp.*	1								
11	*Valvata cristata* (Müller)	133	21	13	1	9	6	7	15	3
11	*V. piscinalis* (Müller)	243	28	17	1	1	1	1	1	2
12	*Pisidium amnicum* (Müller)	827	30	24	2		2	2	2	
12	*P. supinum* (Schmidt)	357	107	25	20	30	25	10	30	10
12	*Sphaerium rivicola* (Lamarck)	52								
12	*Unio sp.*	55	6						1	

E – Ecological groups see Table 2.

TABLE 6. Molluscan assemblages from the profile Levki (by A. F. Sanko)

E	Species	Sample							
		1	2	3	4	5	6	7	8
2	*Vitrea crystallina* (Müller)	1	2	1		7	3	15	16
5	*Vallonia pulchella* (Müller)	6	10	12	13	17	12	41	62
7	*Cochlicopa lubrica* (Müller)	6	6	8	3	5	10	30	43
7	*Euconulus fulvus* (Müller)							3	
7	*Nesovitrea hammonis* (Ström)				1			3	8
7	*Punctum pygmaeum* (Drap.)	1	2	4	3			3	9
9	*Succinea* cf. *pfeifferi* (Rssm.)					2	2		
9	*S. putris* (L.)		4		12	30		10	3
	Succinea sp.	7	8	10	15	15	5	15	13
9	*Zonitoides nitidus* (Müller)	1		3	4			2	
10	*Gyraulus rossmaesslezi* (Auerswald)	1							
10	*Lymnaea truncatula* (Müller)		1		2	3		1	1
10	*Planorbis planorbis* (L.)					1		1	
10	*Valvata pulchella* (Studer)	9	27	20	6	29	17	10	15
11	*Acroloxus lacustris* (L.)	2	2	2	6	1		3	1
11	*Armiger crista* (L.)	49	81	65	41	56	78	49	74
11	*Bithynia tentaculata* (L.)	8	17	6	5	10	2	6	6
11	*Gyraulus laevis* (Alder)	15	71	29	28	150	75	40	22
11	*Pisidium casertanum* (poli)				4			2	
11	*P.* cf. *henslowanum* (Scheppard)	20	40	40	50	100	100	50	500
11	*Pisidium sp.*	150	200	200	250	500	100	100	150
11	*Sphaerium* cf. *nitidum* (Clessin)					13			
11	*Valvata cristata* (Müller)	29	15	16	6	4	7	19	17
11	*V. piscinallis* (Müller)	2	2	1		7	2		1
12	*Pisidium amnicum* (Müller)	1	4	4		2	2	1	
12	*P. supinum* (Schmidt)	100	20	20	50	100	50	50	50

E – Ecological groups see Table 2.

final phase the result of which was a remarkable decrease of stangofile species (to 40%) and the development of hygro- and mesofil species. In the whole profile Paleoarctic, Holarctic and Eurosiberian species characteristic of a cold climate occur what indicates that silts sedimented in the decline of Vistulian and in the beginning of Holocene.

Also in Levki village, 6 m terrace has the similar structure (Fig. 2). Top metre is constituted of grey, slightly sandy muds covering black-grey humus silt (fossil soil) of the thickness of 0.6 m. There are below, down to almost 6 m, flood plain deposits: grey-rusty silts with the shells of malacofauna covering sands with gravel. Tanatocenose and its changes resemble these in Berestenevo profile (Table 6, Fig. 8). Sorts of stagnant waters dominate here decidedly (60–80%). Only in the initial stage, the basin preserved contact with the river channel (reofile species to 25%), whereas in the final stage it overgrew what indicates an increase of palustrine species. Similarly to previous profiles, fauna is composed of species living in a cold climate, indicating the decline of Vistulian and the beginning of Holocene.

The following profile originates from the site Adrov, described in detail (Fig. 1). It is situated at the mouth of the tributary Adrov to Dnieper. At the top the outcrop (0.0–1.2 m) light-brown silty sands occur (M_z = 4.1–5.0 ø, δ_I = 1.7–1.9, Sk_I = 0.1–0.2, K_G = 1.5–1.8), mildly passing at the bed part into flood plain deposits of black-brown organic silts (1.2–3.25 m). We can distinguish two complexes of them. The upper one (1.2–2.25 m) is composed of organic silty muds (the content of the organic substance drops from 7% at the top to 3% in the bed), slightly sanded at the bottom (M_z drops from 6.2 at the top to 5.9 ø in the bed), very badly sorted (δ_I = 2.1–2.5, Sk_I = 0.4, K_G = 0.9–1.0) the top of which was dated 940 + 90 BP (Gd-4570). The lower complex (2.25–3.25 m) is composed of peaty silts (the content of organic substance equals 5%), silts (M_z = 6.4 ø) very badly sorted (δ_I = 2.4, Sk_I = 0.3, K_G = 0.9), the bed of which was dated 9970 ± 560 BP (Gd-4402). Below the organic series clastic sediments occur, coarsening towards the bed, containing malacofauna. From 3.25–5.05 m these are silty muds (M_z = 5.9 ø), very badly sorted (δ_I = 2.3, Sk_I = 0.3, K_G = 1.1), with a small content of organic substance (to 2.5%) and with single inserts (to 1 cm of thickness) of yellow sands inclined towards the river and downstream in the bed. In the bed (5.05–5.45 m) of silts, at the contact with the underlying sandy silts (M_z = 5.0 ø, δ_I = 2.3, Sk_I = 0.2, K_G = 1.5) there occur strongly ferrous sediments and deformational structures connected with reversed density gradient. In the bottom parts of the profile, to a depth

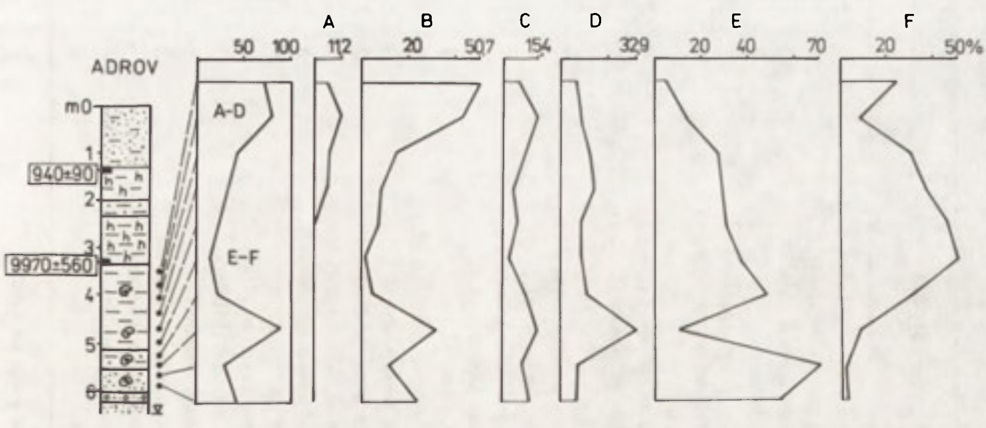

Fig. 9. Malacological diagram at Adrov (by A. F. Sanko)
1 — sand and gravel, 2 — sand, 3 — silty sand, 4 — sandy silt, 5 — silt, 6 — organic silt, 7 — peaty silt, 8 — molluscs, 9 — level of river, A–F — type of malacofauna see Fig. 4

TABLE 7. Molluscan assemblages from the profile Adrov (by A. F. Sanko)

E	Species	Sample									
		1	2	3	4	5	6	7	8	9	10
2	*Vitrea crystallina* (Müller)	4	3	1	1			20	36	100	40
5	*Vallonia pulchella* (Müller)	58	24	57	41	28	31	30	85	378	401
7	*Cochlicopa lubrica* (Müller)	15	2	8	3	13	5	5	20	59	37
7	*Cochlicopa sp.*	3						7	7	38	15
7	*Euconulus fulvus* (Müller)	1									
7	*Nesovitrea hammonis* (Ström)	4	2	7	6	14	5	5	28	40	10
	Vertigo sp.		1	1	1		1				
8	*Succinea* cf. *oblonga* (Drap.)	6	11	10	31	39	16				
9	*Carychium minimum* (Müller)										1
9	*Succinea* cf. *putris* (L.)							28	32	68	44
10	*Limnaea truncatulla* (Müller)	17	7	5	3			19	18	5	3
10	*Planorbis planorbis* (L.)				1	2			1		
10	*Valvata pulchella* (Studer)		8	53	89	189	38	3	20	7	12
11	*Acroloxus lacustris* (L.)							3	1		
11	*Armiger crista* (L.)	9	42	7	142	343	65	47	63	50	11
11	*Bathyomphalus contortus* (L.)	1									
11	*Bithynia tentaculata* (L.)	3	2		23	46	13	5	2	1	
11	*Gyraulus laevis* (Adler)	2	2	2	148	212	31	32		33	20
11	*Pisidium* cf. *lillieborgi* (Clessin)	5									
11	*P.* cf. *subtruncatum* (Malm)	5									
11	*Pisidium sp.*	100	100	10	13				80	25	
11	*Sphaerium corneum* (L.)					1					
11	*Valvata cristata* (Müller)	3	2		11	15	8	14	6	11	5
11	*V. piscinalis* (Müller)	1			72	202	12			2	3
12	*Pisidium amnicum* (Müller)	2	2			26	5	1	40	1	2
12	*P.* cf. *pulchellum* (Senyns)				1						
12	*P. supinum* (Schmidt)		2	15	257	1100	191	125	130	75	187
12	*Unio sp.*	5									

E – Ecological groups see Table 2.

of 5.95 m grey, clayey sands are lying, passing in the bed into sands with gravels with the diameter to 4 cm.

Studies of malacofauna show, that it is possible to distinguish several stages in its development (Table 7, Fig. 9). In the first stage (to the depth of 5.5 m), when clayey sands sedimented, a shallow water basin almost completely deprived of contact with river channel occurred on the flood plain (reofile species below 10%). Because of a very strong overgrowing almost a complete disappearance of the basin followed and palustrine species and those of open environments dominated in this period. In the next stage repeated formation of a shallow water basin followed, however maintaining a remarkable contact with the river channel the evidence of which is the presence of reofile species up to 50% and sandy inserts in silty sediments. On the depth of 4 m, similarly to Pashino, a sudden change of the conditions on the flood plain occurs and almost a total disappearance of water basin though caused not by overgrowing but by drying of the environment. It could result of climatic changes but also of drainage of the terrace by incised river. As the dating of the bed of organic silts (9970 $\pm$ 560 BP) shows, this change occurred on the border of Holocene.

In the tanatocenose thermofil species are missing. The share of species indicating warming of the climate is less than in previous profiles (*Bithynia tentaculata, Valvata cristata*), and the share of typical cold species is remarkable (for instance *Gyraulus laevis*). Forest snail *Vitrea crystallina* appears in larger number only in top samples. All this is evidence of sedimentation of clastic sediments below organic silts and should be referred to the decline of Vistulian. The dating of the bed of organic series also confirms this conclusion.

6 m terrace is an accmulative plain. As datings and studies of residua of rodents, ostracodes and malacofauna indicate, the terrace began to be formed at the end of Vistulian. Initially, numerous depressions with water (channels of braided river?, depressions after permafrost?) were filled with clayey sands and sandy silts (Buroe, Adrov). In the later period silts sedimented on the waterlogged flood plain with numerous small water basins. This accumulation must have proceeded relatively quickly, as silts reach the thickness of 2–3 m. On the border of Holocene we see a distinct change of the character of flood plain, the evidence of which are the profiles in Adrov and Pashino. It probably resulted from quick accretion of the plain with simultaneous incision of the river and change of climate for a dry one of the early Holocene. This caused strong drainage and drying of the sites on the flood plain, though sedimentation of peaty silts continued in place (Adrov). Almost during the whole Holocene 6 m terrace functioned as a flood plain. Silt muds deposited on it while in the period of smaller frequency of floods the fossil soil (found on almost all the sites on this terrace) developed. About 1000 years BP fairly sudden change of the type of accumulation occurred and silt muds were substituted by sandy-silt muds (Adrov). We also observe a similar change on other sites.

LOW FLOOD PLAIN

The present-day flood plain is a narrow (a few to several metres) strip of a level elevated 3.5–4.0 m over Dnieper. Its structure is shown by two profiles.

In Orsha Promyshlennaya the top of the plain (0.0–2.4 m) constitutes characteristic flood sediments. These are alternating strata of yellow-brown sands 10–25 cm thick, and brown silty muds 2–5 cm thick. Below the muds (2.4–2.65 m) sands with gravels of diameter to 2 cm deposit in the bed of which channel pavement inclined towards the river and elevated 80 cm above the level of Dnieper occurs.

In the second profile, in the region of Shklov, the flood plain elevated 4 m above Dnieper is a very narrow, a few-metre strip attached to a vast 6 m terrace. In the outcrop we have here silty muds (0.0–1.85 m) with two fossil soils at a depth of 0.3–0.8 m and of 1.25–1.85 m, divided by about 0.5 m stratum of brown silts. The top of the bottom soil was dated 5040 $\pm$ 110 BP (Gd-4403). Below (1.85–2.25 m) sandy silt, grey with rusty spots occurs, and silt and clayey sands with interbeddings of fine-grained and in the bottom parts coarse-grained sand (2.25–3.25 m). In these sediments a single boulder of the diameter of 15 cm was found on the depth of 2.75 m. The bed of alluvia, below 3.25 m, is composed of sands with gravels.

The flood plain of Dnieper is accumulative and it was formed in Holocene after incision of the 6 m terrace. In mud sediments an increase of intensity and height of floods, about 5000 years ago, is marked.

CONCLUSION

The Dnieper valley is an old valley in the section Buroe-Shklov (Goretski 1970). A big influence on its development in Later Pleistocene was exerted by a very near neighbourhood of the front of the ice sheet of the Last Glaciation during the maximum of which (Orsha stage) the Surazh lob almost rested as far as the Dnieper in the region of Novy Tukhin. Here, a large ice-dammed lake Orsha was formed (Kvasov 1976; Sanko 1987), and the valley carried proglacial waters south-bound. All this caused that older terraces were destroyed in the examined river section. They are preserved only downstream of Shklov where in the region of Krasnaya Gorka near Rogachov 2. overflood terrace (18–20 m) is of mid-Valdaian age, and its alluvia were dated 30 000–46 000 years BP (Arslanov et al. 1971; Zimenkov, Kuznetsov 1985).

Also, in the later stage by the recession of ice sheet during the Orekhovsk oscillation and Gornaya Luchosa phase Orekhovsk and Luchosa dam lakes were formed in front of the ice sheet and the waters of melting ice sheet flowing in pradolina (ice marginal valley) on the northern slopes of Orsha Plateau were carried away into Dnieper by the Luchosa, Chizhovka, Orshitsa and Adrov tributaries (Sanko 1987). In the region of Orsha both these latter troughs unite with the Dnieper valley and not accidentally, Dnieper changes the direction from parallel to meridional here. Lake Luchosa dropped in three stages and the highest level (170 m a.s.l.) A. F. Sanko (1987) connects with the highest terrace in the valley of Chizhovka elevated about 20 m over the level of Dnieper but about 11–13 m above the level of the river. It seems probable then, that the formation of 13 m terrace of the Dnieper should be connected with this period; apart from their similar height the terraces in the valleys of the rivers Chizhovka and Dnieper also show similar features. These are erosional levels with a very small thickness of sandy-gravel alluvia (1–3 m), without a loess cover. Loesses were deposited until Braslav (Gorodok) stage, i.e. until about 15 000 years BP (Voznyachuk 1973). The lack of loesses on the terrace is the evidence, that it must have been in that period still an active accumulative plain. Similar view concerning the age of two overflood terraces of Byelorussian rivers is expressed by Zimenkov and Kuznetsov (1985) and by Matveyev (1990) who connected them with Orsha stadial.

As the profile in Chizhovka shows, 10 m terrace of Dnieper is younger than 17 000 BP. It was formed in a cold climate what is indicated by the type of alluvia deposited by braided river (cf. e.g. Shantser 1951, 1982), cryogenic structures in the alluvia (Dubrovno, Mitkovshchina) and findings of fauna and vegetation in this terrace (Selishche, Shklov). It was formed simultaneously with the increased discharges caused by carrying away waters of melting ice sheet of which the evidence are "dead valleys" preserved in this level in Uste and Shklov. Cutting of 10 m terrace must have happened in Late Vistulian what is shown by the oldest deposits of 6 m terrace coming already from this period (Buroe, Adrov, Pashino). In general this is consistent with previous views that incision of the first overflood terraces on Byelorussian rivers occurred in Allerød (Zimenkov, Kuznetsov 1985) though Matveyev (1990) assumes that it happened even earlier, in Braslav stage. About 10 000 years ago 10 m terrace must have been an overflood terrace or flooded very rarely since Mesolithic population settled on it (Berestenevo). We meet a similar situation in other river valleys of Easteuropean Lowland, where Mesolithic sites are found on first overflood terraces – for instance in the valley of Kama (Baler 1957; Ivanova 1982).

6 m terrace was formed by a river with concentrated channel what is distinctly indicated by facial diversity of its alluvia, from Late Vistulian through almost the whole Holocene. With a stable – in horizontal plane – channel of Dnieper and climatically and tectonically distinct tendency of the river to incise this terrace functioned as flood plain and was vertically accreted by fine-grained overbank sediments often reaching big thickness (Buroe, Berestenevo, Adrov, Novyye Staiki). They sedimented from the decline of Vistulian through the whole Holocene (Adrov – 9970 BP). On the border of Holocene a distinct change of the character of flood plain is marked, i.e. its drainage and dryness (Adrov, Pashino). Most probably it was the result of overlapping of a few factors: concentration of the channel which caused its incision, accretion of flood plain with muds and drier climate at the beginning of Holocene. The change of conditions on flood plain, less frequent floods and slower sedimentation of muds enabled the forming of a fossil soil on this terrace. Its formation most probably began in Eoholocene (Levki) and continued in the Atlantic period when small variability of discharges and wooded drainage basin were favourable conditions for this process.

A constant tendency to Dnieper incision caused that most probably in Eo- and Mesoholocene the forming of the lower step of flood plain began what is proved by Atlantic flood sediments deposited on this terrace (Shklov). At the same time, a fossil soil dated 5040 ± 110 BP indicates an increase of floods frequency in this period. However, their range was relatively small because they didn't affect 6 m terrace.

Remarkable climatic changes on the border of Atlantic and Subboreal are noted on the whole Easteuropean Lowland (Khotinski, Starkel 1982). They led to stimulation of activity of rivers, what resulted in changes of channels and sedimentation type on flood plains (for instance Chebotareva et al. 1965). A distinct change of sedimentation from silty mud into sandy-silty mud on the upper flood plain of Dnieper occurs about 900–1000 years BP (Adrov). These changes indicate an increase of frequency of large floods submerging this terrace what, in consequence, led to fossilization of soil. A very distinct similarity of sandy-silty muds in Adrov and muds covering fossil soil in Buroe and Berestenevo leads to such a conclusion. Their relatively small thickness would also support this conclusion. The period of sedimentation change coincides with taking these terrain by Kiev Russia (from mid-9th c.), location of Kopys (1059 AD) and Orsha (1067 AD), and what follows, with increasing changes of environment caused by man cultivating fertile loess soils of Orsha Plateau and Orsha-Mogilov Plain. Increasing transformations led to almost total deforestation and changes of this region are one of the greatest throughout the whole Byelorussia (Matveyev 1989). Since the 10th and 11th centuries a distinct influence of man on environment of the whole Easteuropean Lowland is marked (Khotinski, Starkel 1982). Yet, simultaneously in this period a distinct phase of cooling and wetting of the climate occurs which marked itself in the valleys of rivers in Byelorussia and Easteuropean Lowland, among others by changes of the sedimentation type on flood plains and by fossil soils (Zolotokrylin et al. 1986; Klimanov, Serebryannaya 1986; Aleksandrovski et al. 1987; Zernitskaya, Kozharinov 1988; Kalicki 1991b). These changes occurred both in basins changed by man – like Dnieper – and in basins having almost natural environment – like Berezina (Kalicki 1991b). That is why man's activity fulfilled a preparatory role, which manifested itself in periods climatically conditioned. At present, on a low flood plain of Dnieper alternating lamines of sands and silty muds are sedimented. These muds are very similar to flood deposits described in the valley of Berezina (Kalicki 1991b).

Finally, it must be noted that the examined, seemingly monotonous section of the Dnieper valley, with poorly developed terraces and missing preserved paleochannels, is interesting and at the same time very specific on the whole course of the river. There is a lack of older terraces here, only Young Plenivistulian terraces and two levels of Holocene flood plains occur. Narrow valley and incision of river caused big stability of the channel in horizontal plain and meanders practically have not developed. That is why modelling of flood plains was mainly the result of only a vertical accretion of flood deposits. We deal here with a completely different type of development than in the valley of Berezina near Borisov (Kalicki 1991b) or

Vistula near Cracow (Kalicki 1991a) where within the bottom of the valley a row of inserted, multiaged series of alluvia occurs next to one another.

These researches lead also to a conclusion that, in spite of such a distinct difference between the two mentioned valleys, in Dnieper valley, changes created by short-lasting periods of cooling and wetting of climate in Holocene occur, too (cf. Starkel 1983; Kalicki 1991a). In the Dnieper valley phases from the border of Atlantic and Subboreal and from about 1000 years BP manifested themselves by changes of sedimentation type on flood plains. Probably, it was induced by an increase of frequency and height of floods in these periods and what follows by an increase of the rate of mud sedimentation. This led to the fossilization of soils developing on flood plains in still periods. It is very interesting that stages identified in Dnieper valley occur in the same periods in Berezina valley and in other valleys of Easteuropean Lowland (Kalicki 1991b) and also in the valleys of Polish rivers (Kalicki 1991).

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A MODEL OF DEPOSITION OF LOESS IN CENTRAL PART OF THE POLISH FLYSCH CARPATHIANS DURING THE LAST COLD STAGE

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ABSTRACT. The paper deals with a proposal of constructing a model of formation of the specific Carpathian variety of loesses during the Vistulian. It is based on the observation and measurements of the actual activity of winds, the occurrence of blowouts containing lacustrine-boggy deposits of Allerød-Holocene age, as well as on the occurrence of ventifacts and covers of radiocarbon dated Plenivistulian anemoclastic silty-clayey sediments, a dozen or so metres thick.

Silty and silty-clayey sediments, defined as loams or loess-like deposits, are covering considerable part of the Polish Carpathians, mainly the areas of Foothills and adjacent Foreland regions.

There are divergent opinions of the origin and age of these deposits in the Carpathians. In earlier papers: Łoziński (1909), Kuźniar (1912), Tokarski (1936), Tokarski et al. (1961), Klimaszewski (1948) they were considered to represent aeolian distant-transport deposits. In more recent papers (Malicki 1950; Jahn 1956; Maruszczak 1963, 1985; Jersak 1976; Różycki 1968, 1976; Chlebowski and Lindner 1989) this aeolian deposit is treated as the product of short-distance transport. On the other hand, Łoziński (1934), Cegła (1963), Uziak (1964) and Zasoński (1983) consider them to represent the deposits of water basins or even weathering crust of rocks of the flysch basement. As far as the age is concerned, these sediments are supposed to be related with South-Polish glaciation (Malicki 1950), pleniglacial of the last cold stage (Klimaszewski 1948) or with the stadials of the last glaciation (Malicki 1961, 1972; Maruszczak 1972, 1985, 1986; Jersak 1973, 1976).

There are three kinds of premises which were the basis of constructing the proposed model of formation of anemoclastic sediments in the Carpathians:

1) The results of observations and measurements of the phenomena of recent activity of winds in the Low Beskid Mts. and its foreland and the evaluation of its effectiveness in the period of human activity;

2) The occurrence of fossil blowouts in the Jasło-Sanok Depression, containing lacustrine-boggy deposits of Allerød-Holocene age with the ventifacts on the divides between fossil lakes, as well as spindle-shaped forms of some humps in central part of the Jasło-Sanok Depression, showing S — N orientation;

3) A dozen or so metres thick cover of anemoclastic silty-clayey deposits in southern part of the Dynów Foothills of Vistulian age (after radiocarbon dating).

These studies were carried out in the Low Beskid Mts. and its northern foreland. From S to N there occur here four main geomorphological units showing E — W strike (Gerlach et al. 1985). In the south we observe low mountains (500–900 m a.s.l.) and further to the north there occur successively:

1) large intramontane Jasło-Sanok Depression showing the morphology of low foothills and wide Pleistocene plain terrace (250–360 m a.s.l.);

2) Strzyżów Foothills, also called the Brzozów Beskid Mts., representing low mountains (450–600 m a.s.l.);

3) Dynów Foothills (350–400 m a.s.l. – Fig. 1).

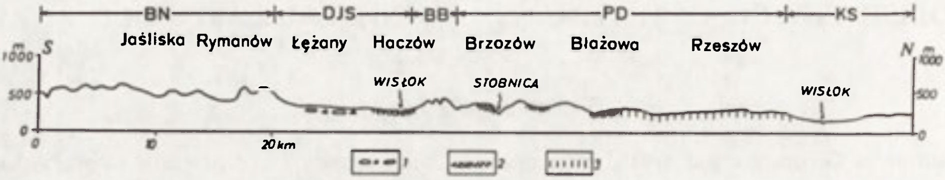

Fig. 1. Section across the central part of the Polish flysch Carpathians showing zones of deflation and aeolian deposition of the Carpathian variety of loesses during the Vistulian

1 – terrace plains bearing ventifacts, and blowouts containing lacustrine-boggy deposits of Allerød-Holocene age, 2 – sheets of the Carpathian variety of loesses, 3 – loess sheets in the Carpathian Foothills. BN – Low Beskid Mts., DJS – Jasło-Sanok Depression, BB – Brzozów Beskid Mts., PD – Dynów Foothills, KS – Sandomierz Depression

This part of the Polish Carpathians consists predominantly of folded Oligocene and Lower Miocene Krosno Beds (Koszarski 1985). They are composed of very thick complex of gray, strongly calcareous, very fine-grained micaceous sandstones, intercalated with similarly gray and calcareous shales. When compared with the Magura Beds or other flysch complexes, they are very easily subjected to weathering processes.

The sandstones of the Krosno Beds consist predominantly of grains less than 0.1 mm in size (Obuchowicz 1957) whilst in the shales they are essentially below 0.06 mm (Ratajczak 1990). Another important feature of Krosno Beds is their mineral composition. According to Obuchowicz (1957) and Pinińska (1977), the major rock-forming minerals of sandstones are: quartz (30–45%), carbonates (20–30%), micas (20–40%) and heavy minerals (about 3%). Among the latter minerals garnet is the most abundant. The shales consist of smectite-chlorite and smectite-illite groups (Ratajczak 1990).

It should be emphasized that the Low Beskid Mts., about 100 km long and 500–900 m a.s.l. high, represent the narrowest and the lowest part of the whole Carpathian mountain range. Morphologically it is something like a broad gate. Eastwards (Eastern Beskidy Mts. – Howerla 2061 m a.s.l.) and westwards (Tatra Mts. – Gerlach 2654 m a.s.l.) from this gate, the Carpathian range is broading and becomes much more elevated. In view of such orographic situation, the area of the Low Beskid Mts. is characterized by the occurrence of strong south winds, called here the Dukla and Rymanów ones. Usually, these winds are blowing during late autumn, in winter and in spring, when the majority of fields is plought and there is no vegetation. These winds are strong (more than 10 m/s) or very strong (more than 15 m/s) and are lasting from one to several days. They are destroying the soil and winter corn on the windward slopes. On the other side, the deposition of snow and soil material takes place on leeward slopes, as well as in front and behind various natural and artificial obstacles. The measurements of mineral sediments and those mixed with snow or deposited on it have shown that on leeward slopes and behind the obstacles the average thickness of these deposits increased by more than 5 cm during one winter (Gerlach and Koszarski 1968a, 1968b). These sediments consist predominantly of soil microaggregates and fragments of shales, up to 5 mm in size. After the melting of snow, these sediments cause increases the thickness of loose deposits. Therefore, actually on leeward slopes, as well as before and behind the obstacles, there are growing anemoclastic clayey covers. Their thickness, corresponding to about 600 years of human activity, amounts to ca 1 m. One the contrary, during the same period on the windaward slopes considerable part of the soil layer was removed, often down to the flysch basement (Gerlach 1976, 1977).

The second premise for constructing of this model of formation of anemoclastic sediments in this area during the Vistulian period are the results of studies on the evolution of Late Quaternary fossil lacustrine-boggy basins. They have documented the occurrence in this area of a dozen of basins filled with Late Glacial and Holocene deposits. The shape of depressions, showing predominantly S–N orientation, spindle-shaped hills exhibiting the same orientation, as well as the age of sediments filling these depressions and the occurrence of ventifacts at the inter-lake divides – all these data allowed to explain the origin of these depressions as blowouts (Gerlach 1990a, Gerlach et al. 1991).

The third premise for the construction of this model resulted from detailed studies of a dozen or so metres thick silty-clayey cover in Humniska near Brzozów. This stand is localized in southern part of the Dynów Foothills just behind the northern margin of the Jaslo-Sanok Depression, i.e in the Brzozów Beskid Mts. (Gerlach et al. 1990). The silty-clayey sediments in Humniska, 25 m thick, can be subdivided into 3 horizons: 1) lower, 15 m thick, yellow, 2) middle, about 4 m thick, dark gray, and 3) upper, about 7 m thick, brown-yellow.

The characteristic feature of the sediments in question is their silty-clayey mechanical composition, high muscovite content, considerable proportion of clay minerals of the smectite-illite-chlorite group and relative abundance of the most resistant heavy minerals. All these features indicate their strict relation with grain size and mineral composition of rocks of the flysch basement in this area. Consequently, it is reasonable to conclude that the alimentation source of material for the sediments in Humniska were weathering Krosno Beds, consisting of strongly calcareous, very fine-grained sandstones and siltstones, intercalated with clayey shales.

Morphological-geological position of loams in Humniska, as well as the results of analyses and radiocarbon dating have shown that these sediments should be related with aeolian deposition during the Vistulian. Deluvial origin of these deposits is difficult to accept because of lack of connection with the slopes of more elevated hills.

The occurrence of ventifacts in the region situated south of Humniska, as well as spindle-shaped hills and blowouts filled with lacustrine-boggy deposits of Late Vistulian and Holocene, are sufficient arguments to conclude that the main deflation area was the Jaslo-Sanok Depression.

Aeolian deposits in Humniska differ from loesses of the Lublin Upland in grain size distribution and mineral composition (Dolecki 1986; Harasimiuk 1986). The former are distinguished by finer grain size, high content of clay minerals and muscovite, as well as by different heavy mineral assemblage. These differences, when compared with loesses of the Lublin Upland, and distinct similarity of loesses in Humniska with grain size and mineral composition of the Krosno Beds, are evidencing their autochtonous position in the Carpathians. This is the basis for distinguishing a separate, so called Carpathian, variety of loesses (Gerlach et al. 1991).

Two periods of more intensive deposition of the Carpathian variety of loesses, separated by that of slower sedimentation, can be distinguished. They correspond to lower and upper Plenivistulian. The period of slower increase of their thickness, marked by distinct traces of pedogenesis and continuous blow of mineral and organic materials, corresponds to Hengelo and Denekamp interstadials of Dutch authors (Kozarski 1981). Documented origin of such thick series of aeolian near-transport deposits in the Dynów Foothills during the last glaciation clearly evidences that in the close surroundings during their growth there had to be larger areas devoid of vegetation (Starkel 1988).

During the last glaciation the routes of cyclones took place more to the south than actually. The gradients of temperature and air pressure were higher. Consequently, the frequence and strength of winds were more intense. Strong winds from southern sector were causing intense deflation on the windward slopes, formation of blowouts in depressions and remodelling of pebbles into ventifacts. On the other side, abundant aeolian deposition of silty-clayey mineral sediments took place on lee slopes and behind natural obstacles of structural escarpment type. Consequently, in the Dynów Foothills, thick beds of anemoclastic loams were formed, representing the Carpathian variety of loesses (Gerlach 1990b). The northern margin of the Carpathians was the transition zone from the Carpathian loess (whose material was transported from the south) to central European loess originated from the material transported along E — W (parallels of latitude) axes.

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DESERTIFICATION IN THE LIGHT OF SEDIMENTOLOGICAL FEATURES OF DUNE DEPOSITS

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ABSTRACT. The purpose of the present study was the answer to the question whether sedimentological features of dune deposits make a record of the duration of desertification processes. According to the author, the features of dune sediments showing the duration of aeolian processes, indirectly indicate the period of lasting of desert conditions. Particularly informative of the duration of aeolian processes is the degree of transformation of dune sediments in relation to source sediments as for their grain size distribution, mineral composition, as well as the features of abrasion and frosting of quartz grains (with the use of electron microscope analysis as one of the research methods).

INTRODUCTION

Within of continental areas of the Earth hot and cold deserts are distringuished. According to the definition which is to be applied for deserts and semideserts, they cover 1/5 to 1/3 of Earth surface (Glennie 1987). The major part of desert areas are rocky surfaces, some of them covered with fluvial sediments deposited in local depressions. Only 4% of these areas are covered with dune sand.

The purpose of investigations presented below is to answer whether in sedimentological features of dune sediments the length of desertification processes has been recorded. Among hot deserts the dune sands of six deserts of south-western Asia (Dymowska et al. 1984, Fig. 1), sands from a dune of central part of Syria near Palmyra (Mycielska-Dowgiallo 1980) and dune sands from the central part of Sahara (vinicity of Djanet oasis) were examined.

For the sake of comparison the results of investigations dunes from cold desert of the decline of last glaciation from the area of Poland, Belgium and Sweden (Mycielska-Dowgiałło in press) will be also presented.

An increasing interest in features of aeolian grains of desert areas date from the end of previous century (Sorby 1877; Phillips 1881, 1882; Cayeux 1926; Cailleux 1942; Kuenen and Perdok 1962; Pye 1987; Pye and Tsoar 1987). Main interest is recently focused on quartz grain abrasion (Kuenen 1960; Kuenen and Perdok 1962; Kaldi et al. 1978; Krinsley et al. 1979; Wallendorf and Krinsley 1980; Lindé and Mycielska-Dowgiallo 1980; Whalley et al. 1982, 1987), the origin of dusty fraction (Goudie et al. 1979; Nahon and Trompette 1982; Pye 1983, 1987; Smith et al. 1987; Whalley et al. 1987) and the genesis of silica coats on grains (Folk 1978; Pye 1983; Pye et al. 1987; Whalley et al. 1987).

BRIEF CHARACTERISTIC OF SAMPLING SITES

The examined dunes of deserts of south-western Asia were situated at the Salty Desert, Lotha, Kharan, Thal, Thar and the Baktria part of Kara-Kum, Fig. 1 (Dymowska et al. 1984)

1. The Salty Desert extends in the mountain basin, its bottom being covered up with prolluvial and sandy loam alluvial salty deposits. In the basin dominates continental-dry climate (Table 1). Sand samples were collected in the Shurab region, where dunes are formed as short transverse ridges converted into sickle-shaped formes. The length of individual ridge is 140–150 m, the height 3 m. The source of dune sand are fresh sediments of alluvial fans of periodical streams flowing down from the mountains and the sand od the local rocks.

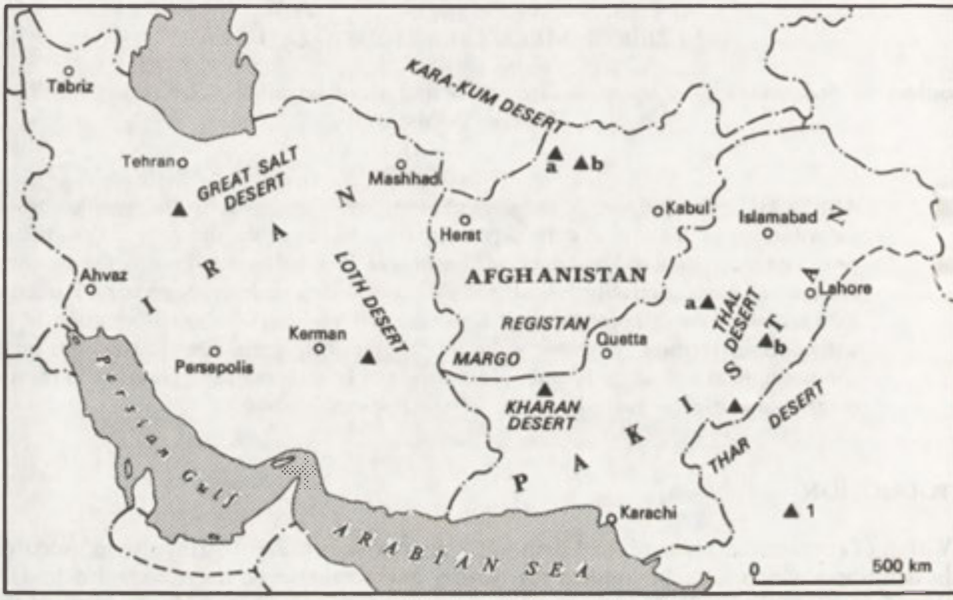

Fig. 1. Distribution of investigated areas in South-Western Asia
1 — sampling site

TABLE I. Climatic data after K. S. Ahmad (1951) and M. H. Nour (1965)

Desert	Measuring Station	Altitude (m a.s.l.)	Climatic zone	Average air temperature		Average precipitation (mm)
				January	July	
Kara-Kum	Mazar-i Scharif	487.0	temperate, subtropical	6°	31.6°	21[illegible]
Salt	Yazd	1200.0	subtropical	6°	33.0°	12[illegible]
Lotha	Zahedan	1370.0	subtropical	6°	31.0°	7[illegible]
Kharan	Dalbandin	849.0	subtropical	10°	33.0°	12[illegible]
Thal	Multan	128.0	tropical	13.6°	34.6°	18[illegible]
Thar	Khnapur	99.0	tropical	13.8°	33.8°	16[illegible]

2. Lotha Desert between the Kuh-Ruda and East Iranian mountains extends in many basins, filled with sediments of contemporary periodical rivers. The climate is continental, extremely dry (Table 1). The samples of sand were taken from the surface of single barchan dune in Noalabad region. The height of this dune was 21.5 m, length in profile 238 m. Between the arms of the dune a shallow depression with sandy and mady material of small periodical streams was situated. The main bulk of the dune sand originated from winnowed sand and gravel plains on the barchan dune foreland.

3. Kara-Kum Desert in the Baktrian part covers wide plains of the old flood-free terrace of Amu-daria river and rivers flowing down from Band-i and Turkestan Mts., which losing water on the terrace level of Amu-daria river accumulate wide sand fans. The climate is continental with transitional features between temperate and subtropical (Table 1). The winds are variable with northern ones prevailing. Graminous plants develop during the rain-season.

The sand samples represent two sites: one situated on the plain of high terrace of Andhui river and the second on alluvial fan of Balkharr river. The plain of flood-free terrace rises about 40 m above contermporary valley bottom and is covered almost totally with drift sands forming barchan type dunes, transversal ridges and rake forms. The height of barchans does not exceed 5 m.

4. Thal Desert covers the region between Indus, Chenab and Jhelum rivers built of Pleistocene alluvia winnowed on the surface, with dominating dry continental climate. Nevertheless a fairly rich plant cover occurs here with shrubs and single trees prevailing.

Two sites were situated on the desert, one in the central part to the East of Bhakkar, the second in southern part near Rangpur. Low dunes of irregular shapes covering one another in result of variable monsoon winds, prevail in the first site. The samples originate from a barchan dune of 3.5 m height. The direct source of a dune sand are local sandy deposits of the region between rivers. More distinct are the dunes in southern part, in the vicinity of Rangpur, where their conglomerations form huge ridges elevated up to 56 m above the plain. The height of individual barchan dunes reaches 28 m.

5. Thar Desert, similarly to the Thal Desert, spreads out from Indus to the East, on the plain of flood-free terrace. Its climate conditions are similar but with higher air temperatures. Shrubby plants occur here in clumps between dunes. Drift sands form significantly larger surfaces. They are diversified with barchan dunes of small height of 2–2.5 m. The sand samples were taken from an asymmetric barchan dune in the vicinity of Khanpur.

6. Kharan Desert in northern Baluchistan comprises dunes and gravel plains. The samples originate from the surface of symmetrical barchan dune on the Sultan plain which is bordered by Chagai Mountains from the North and the Kharan Basin, with saline lake Maszkel and dunes in its eastern part, from the South. The surface of the plain is built of a series of conglomerates lying under the sandy deposit. With dry continental climate the plain is almost entirely deprived of plants. The dunes wandering on the plain represent the shape of classical barchan dunes of free movement. Their height reaches 5.5. m, maximum width 102 m and length 108 m.

The second area of exploration is the desert of Central Syria, district of Palmyra (Mycielska-Dowgiallo 1980). It is situated in a basin, bordering on the North on the Dżaba-Abu Rudżmajn mountain ranges. The mean annual temperature for this area is 18.8°C, mean temperature of January 6.9°C. Significant temperature differences occur between day and night. Annual precepitations average 93 mm. From aeolian forms nebkhi are the most frequent, occurring both in the mountains as well as on the border of sebkhi in Palmyra vicinity. The typical dune ridge was found only in the border area of mountain range, about 2 km to North-West from Palmyra at the level of about 100 m above the bottom of sebkhi. It extends from West to the East along low mountain pass. Its maximum height is 7.5 m. From southern (proximal) side, the inclination of the surface reaches 30° and from the northern (distal) one it is bent at the angle of 40°.

On the southern slope of the dune, a series of layers is exposed steeply inclined (from 11° to 20°) to NE. It is built of sand laminae with strongly comminuted plant remnants. The

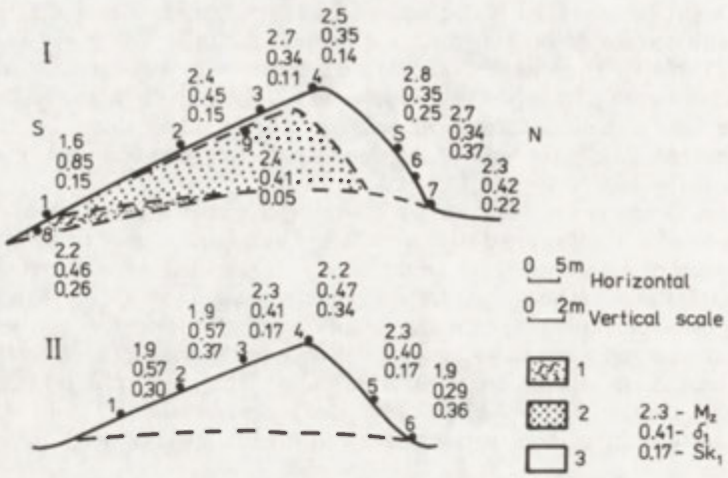

Fig. 2. Transversal sections of a dune ridge nearby Palmyra with marked sampling sites and values of granulometric indices

1 — waste series underlying a dune form, 2 — deposits of an older dune generation, 3 — recent dune sands. Values of inices: M_z — mean size, δ_1 — standard deviation, Sk_1 — skewness

character of stratification of this series indicates, that it makes a record of the distal side of the earlier dune form, which is simultaneously the core of a contemporary dune ridge. The latter is displaced to the North in relation to the former one (Fig. 2). Below both generations of dune forms sandy series are preserved, the stratification character and petrographic composition of which show, that they are slope deluvia.

Djanet oasis between Tassili-N-Ajjer and Erg Admer plateau in Central Sahara, extending as a continuous stretch of 450 km of huge dunes to Hoggar Mts. in the West was the third studied area. It is an area of hot, extremely dry climate. Temperatures above 40°C are recorded during more than 50 days of the year. The diurnal amplitudes of temperature are 15–20°. Mean yearly precipitation do not exceed a few mm. In the vicinity of Djanet oasis, when samples were taken (1986) the last rain was noted 16 years ago. The relative moisture below 30% in Tamanrassat is noted on the average during 310 days of the year, but it fluctuates from higher in the morning to lower in the afternoon.

Timraz is the first place from which samples were taken. It is an area of high, steep monadnocks surrounded by conoid taluses. A few meters high dunes are formed between them from blown material of the taluses. Two samples for analysis in electron microscope (SEM) were taken from the surface of these forms. The sample was taken from the surface of one of the dunes on its proximal side.

GRANULOMETRIC COMPOSITION, QUARTZ GRAIN ABRASION OF SANDY FRACTION AND CALCIUM CARBONATE CONTENT

From the dunes of south-western Asia and Syrian desert, samples of sand were taken along transversal profiles of the dunes. In the first area of exploration (SW Asia) they were taken from depth of 20 cm in case of dune sand and 20 and 40 cm in case of sediments of matrix. In the second investigated area (Syria) samples were taken from surface layers. In the third area (Central Sahara) only three samples were taken from characteristic sites also situated at the surface, for electron microscope analysis (SEM).

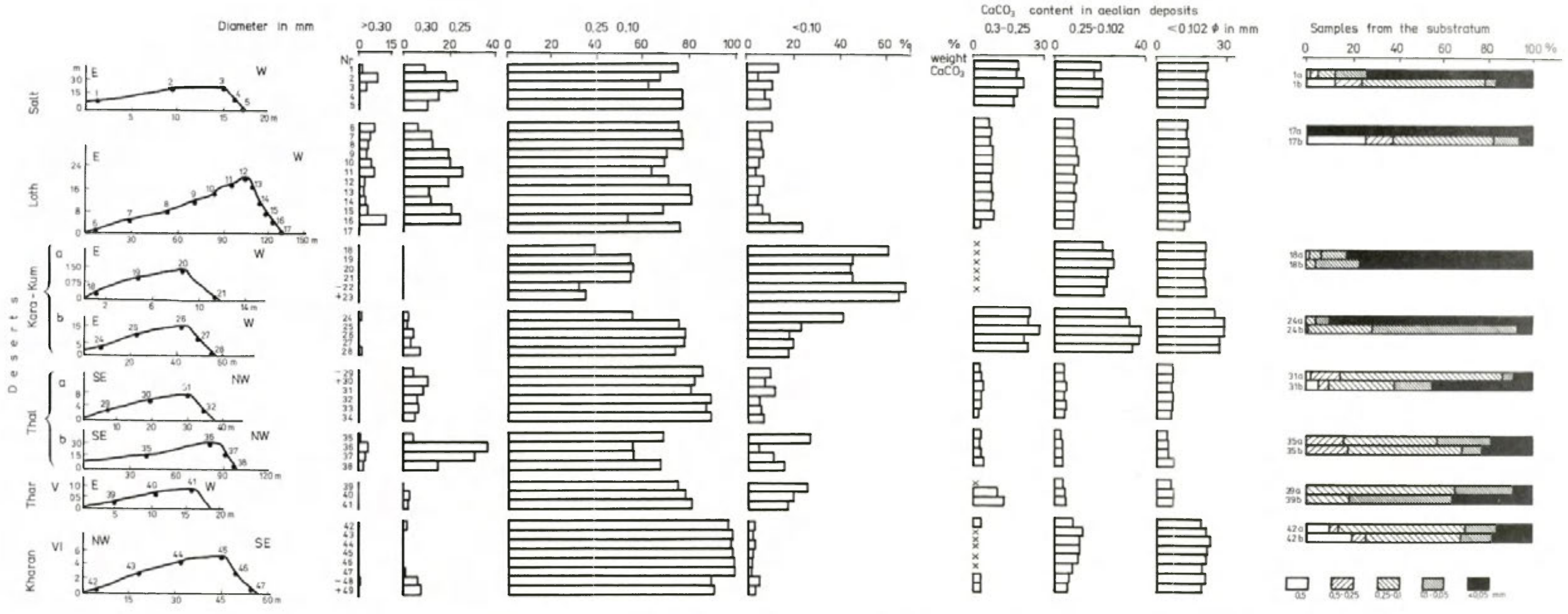

Fig. 3. Granulometry of dune sands and their substratum, as well as $CaCO_3$ content in aeolian deposits (−) samples taken from left-hand side of the dune, (+) samples taken from right-hand side of the dune.
Sample of the substratum: a — from the depth of 20 cm, b — from the depth of 40 or 60 cm

THE SECOND AREA OF EXPLORATION (THE DESERT OF CENTRAL SYRIA)

The analysis of graining of dune sediments from Palmyra region was conducted basing, both, on values of graining indices of Folk and Ward (1957) as well as on plotted frecuency curves (Mycielska-Dowgiallo 1980). Within the both profiles the best sorting is noted in the crest parts and the weakest one at the basis of proximal side (Fig. 2). Simultaneously ore may see in the crest part, similarly to the dunes of deserts of south-western Asia, a clear enrichment in 0.25–0.1 mm fraction. The thickest grains are generally noted at the basis of proximal side. It is possible to suppose, that they are accumulated both in the process of blowing out of the fine material, as well as the result of rolling by wind, from the foreland of dune area, the thicker grains and placing them at the base in front of an obstacle which a dune forms for them. The abrasion of quartz grains was studied using a graniformameter (Krygowski 1964) for fraction of 0.5–0.8 mm. Unfortunately, only single samples had sufficient amount of quartz grains of this fraction. We managed to select three samples: from the basis of proximal side, from the central part of the proximal side and from the base of the distal side. Generally, the grain abrasion is very low, the index of abrasion (Wo) ranges from 494 to 745 with participation (2–3% only of very well rounded grains) γ – type of rounding. A relatively highest index of abrasion is observed at the basis of the proximal side, the lowest one – at the basis of the distal side.

Many authors think (Trembaczowski 1968; Nowaczyk 1976, 1986; Jaśkowski and Kowalski 1977) that the process of selective transport has a decisive influence on the distribution of grains in transversal profile through the dune, according to their abrasion degree. The angular grain is more susceptible to air transport in the saltation process and in suspension. It is more readily carried up on the proximal slope, the crest of the dune and out of it. We shall return to this discussion later.

THE COMPOSITION OF HEAVY MINERAL FRACTION

In order to specify the degree of sand transformation in aeolian process as well as the types of weathering, the analysis of heavy minerals was carried out in 48 samples from the first area of exploration (SW Asia) and in 6 samples from the secound area (Central Syria) in the 0.1–0.2 mm fraction. In the case of the second area this analysis was supplemented with five petrographie analyses from selected sites of the dune profile. This was done for the 0.43–0.5 mm fraction.

THE FIRST AREA OF EXPLORATION (SW ASIA)

Within five out of six investigated desert regions: Lotha, Kharan, Salty, Kara-Kum and Thal, the epidotes exhibit clear predominance in the fraction of heavy minerals. At the same time it was possible to notice a certain increase in their share in the crest zone of the dunes (within three first mentioned deserts). On the contrary the content of amphiboles is lowered in the same zone.

The different share of epidotes and amphiboles in samples from transversal profiles through the dunes may be explained as the effect of aeolian corrasion, selectively destroying less resistant minerals, in this case – the amphiboles (Mycielska-Dowgiallo 1980b). At the same time both these minerals have a similar specific gravity, thus we can not attribute this phenomenon to selective removing by wind of a lighter one. Moreover a relatively large participation of the mentioned minerals in relation to small content of garnet, a mineral very resistant to corrasion (Morawski 1988; Mycielska-Dowgiałło 1980a) suggest that the intensity and duration of this process could not be significant. This is also shown by generally a low degree of rounding of the grains of heavy minerals as well as the lack of clear tendency

TABLE 2. Selected data of heavy minerals analysis

Desert	Average percent of nontransparent minerals	Average ratio of transformed epidotes to nontransformed ones
Salt	38.0	1.8
Lotha	31.4	1.5
Kharan	29.0	0.5
Kara-Kum	22.5	0.4
Thar	10.7	0.9
Thal	10.3	0.8

towards enrichment in garnets of dune sediments in realtion to the sediments of foreland of the dune. The composition of heavy minerals found in the dune sediments of the discussed deserts points to its considerable dependence on the source of these sediments, much more than on corrasion processes.

Another problem is the mutual relationship between processes of chemical weathering occurring in the source sediments and within the formed dunes. For this purpose, the average share of non-transparent minerals, as well as the relation of nontransformed to transformed epidot grains were calculated for dunes of particular desert areas (Table 2). It appeared that the discussed six desert areas may be classified into three groups characterized by similar features. These features are determined by processes occurring contemporarily and in previous geological periods as well. It is then necessary to specify the conditions in which the contemporary sediments are formed and to reconstruct approximately the conditions of formation of sediments being the source of the contemporary ones.

The dune sediments of Salty and Lotha deserts originate from alluvial sediments of periodical streams flowing down the surrounding hills. These are sediments continuously renewed, derived from the rock weathering material from the basin. The highest proportion of nontransparent epidotes in dune sediments is probably due to short duration of weathering processes in the alimentation material. In dunes already formed, in hot and dry climate, the intensive processes of deposition of iron compounds probably occur. This may probably explain the highest proportion of non-transparent grains.

The dune sediments from the second of the three distinguished groups of deserts (Kara-Kum, Kharan) seem to originate from sediments which have a long earlier history and underwent prolonged process of chemical weathering also under conditions of hot and wet climate. The epidote grains might undergo significant transformation in such conditions. This is highly probable in the case of Kara-Kum desert, because there the dunes being formed contemporarily originate from blown-through alluvial sediments of terrace surface which had been formed in much more wet periods. The dunes of Kharan desert also derive sediments from the old alluvia filling the Kharan basin. Contemporary conditions of dry climate prevail in this area which is marked by accumulation of iron compounds on the grains.

The dune sediments of the third group of deserts (Thar, Thal) are characterized by the lowest content of non-transparent minerals with simultaneous slight predominance of transformed epidotes over those non-transformed. This situation seems to depend on three factors: a low content of $CaCO_3$ in sediments, a slightly higher value of mean yearly precipitation (comparable to Kara-Kum precipitation) and the presence of old river alluvia as a source of dune sediments. The first two factors may lead to the depletion of iron compounds that results in lowering of the content of non-transparent minerals in sediments, the third factor may be responsible for the predominance of transformed epidotes over the non-transformed ones.

THE SECOND AREA OF EXPLORATION (THE DESERT AREA OF CENTRAL SYRIA)

For analysis of petrographic and mineral content, five samples were taken from characteristic sites of transverse profile of the examined dune. Biomicritic fines of dolomitic limestones predominate in all samples. They are often silificated and contain fragments of cherts. All this fines are sharp-edged without traces of abrasion. The content of this group of rock fines is differentiated in the transverse profile of the dune. Their lowermost content (47%) is found in the lower part of the proximal side together with the highest share of the quartz grains (52%). In the crest part their content increases (67%) and the quartz content decreases (31%), whereas at the basis of the distal side it reaches its maximum (80%) with only 17% of quartz.

In may be presumed that the grain shape has an essential influence on the distribution of mentioned rock-mineral groups in the transverse profile through the dune. The fines of carbonate rocks are much more sharp-edged than quartz. Because of the difference in the character of grain rounding in both groups, most probably takes place in the process of aeolian transport.

In the group of heavy minerals, terrigenous ones occur in trace amounts only (single grains of tourmaline and zircon). Iron hydroxides and fines of carbonate rocks covered with them form the remaining part. The content of heavy minerals does not show any substantial differentiation in the transverse profile of the dune.

Rock-mineral analysis carried out for deluvial sediments underlying the examined dune shows their considerable similarity to the overlying aeolian sediments. It seems to indicate the short duration of aeolization process.

ANALYSIS OF THE SURFACE TEXTURE OF QUARTZ GRAINS IN ELECTRON MICROSCOPE (SEM)

The analysis of the surface texture of quartz grains for the first two investigated areas was performed for 0.5–0.8 mm fraction, but in the case of lack of this fraction for 0.25–0.30 mm fraction. From the third area, the analysis was performed for grains of 0.7–1.0 mm fraction. A number of 100–200 grains selected for analysis, were subsequently divided under light microscope into characteristic group with respect to rounding and frosting of the surface. Then, 12 grains were selected altogether, for analysis in electron microscope, proportionally to the amount in particular groups. The grains were washed with 10% HCL before analysis.

FIRST AREA OF EXPLORATION (SW ASIA)

The analysis of quartz grains carried out by the SEM technique did not show any clear differences between the quartz grains from different desert areas and from different parts of the investigated dune forms (proximal or distal sides, crest). Irrespective of magnification the same types of texture were recognized on all grains. At low magnifications all grains showed weak rounding (Photo 1, 2). However at the magnification 400–1000x on the majority of grains, the presence of homocentric crusts, wrapping earlier forms of surface texture, have been revealed (Photo 3, 4, 5). Within the local depressions these crusts were intensively cracked (Photo 6, 7, 8) showing their laminar structure (Photo 9, 10, 11).

Many authors were interested in the process of formation of crust on aeolian grains. In its formation an essential role plays the process of etching and precipitating of silica, in which dew (as well as rainfalls if they occur) and high temperature participate actively. The water trapped in the dune sands in the form of films covering the surface of grains, may have during drying up a high alkalinity and salinity. It causes dissolution and precipitation of free silica

(Pye and Tsoar 1987; Pye 1987). This complexity of the process (dissolution and precipitation) is visible on the examined grains in places where the crust came off. The surface of clearly directed texture, indicating an active process of chemical etching simultaneous with precipitation, are visible under the crust (Photo 12).

Mechanical abrasion in the process of aeolian transport of the grains may participate in the formation of silica crust. In this process, the quartz grains may become electrically charged and a part of the crushed fines may attach to the grains, particularly filling depressions on their surface (Lindé and Mycielska-Dowgiallo 1980). These fines can be secondarily cemented with free silica (Photo 13, 14, 15).

Traces of typical aeolian abrasion are visible on small parts of the grain surface and only within the already existing crusts (Photo 16, 17, 18). This seems to indicate that aeolian processes were short episodes in the area of the investigated deserts.

THE SECOND AREA OF EXPLORATION (THE DESERT AREA OF CENTRAL SYRIA)

The majority of quartz grains building the examined dune of Palmyra region are angular or semi-angular (Photo 19, 20). The analysis of the quartz grains surface texture in SEM proved, that the main role in formation of this surface was played by etching processes with simultaneous precipitation of silica crust. Within surface depressions this crust is cracked or porous, (Photo 21) due to the process of drying up of hydrated silica. The forms evidencing aeolian abrasion are met merely on small fragments of grains. Sometimes marks of aeolian abrasion and silica precipitation, occurring alternatively, are visible. When the silica crust was removed with the use of ultrasounds, the surface indicating intensive processes of chemical corrosion has become visible (Photo 22). This surface has clearly directed forms revealing the crystallographic net of the quartz (Photo 23, 24). This type of quartz grain texture is typical of environments with considerable activity of chemical etching.

THE THIRD AREA OF EXPLORATION (CENTRAL SAHARA)

Three samples for SEM analysis were taken from this area. First two were taken from the neighbourhood of a high steep rock monadnock (vicinity of Timraz). It is surrounded by a conoid talus. The weathering material from this talus is the source of sediment for an actively winnowed dune (7–8 m of hight). At a distance of 10 m from the base of the talus, already on the flat surface, the first sample was taken and the second, 30 m further at the base of the dune. The third sample was from any other area, that is from an 80 m high dune, one of the huge Erg Admer dunes.

The first sample consistes of angular and semirounded grains (Photo 25, 26). In the surface texture of these grains the presence of silica crust, of different degree of its development, is common (Photo 27, 28, 29). Some grains are crushed but the surfaces of fractures already show initial forms of silica precipitations (Photo 31). Aeolian processes are marked only within the silica crust. Small magnification shows incisions crumbling the silica crust (Photo 32, 33). At higher magnifications incisions are visible; they initiate the silica crust destruction (Photo 34, 35, 36).

The second sample consists of round, frosted grains and semiround, partly frosted (Photo 37, 38). Within the grain surface the aeolian texture predominates, occurring on the whole surface of the grains (Photo 39) or only on its convex fragments (Photo 40, 41). Larger magnifications show fresh incisions and chipped, ground surface of the grain characteristic for aeolian processes shaping the noncrusted quartz grains. (Photo 42). Silica crust was not observed on the grains of this sample. It may be assumed that it was removed in the active, long-lasting process of aeolian abrasion.

The third sample from Erg Admer dune consists of round and semiround frosted grains (Photo 43). At small magnification the major part of the surface is covered with aeolian pitting (Photo 44) formed within the silica crust (Photo 45). Aeolian abrasion crumbling the emerging crust (Photo 46, 47) exposes sometimes an underlying surface of directed texture (Photo 48). At the same time, aeolian abrasion may parallel further silica precipitation that causes gradual smoothing of the outline of arising abrasion depressions.

The comparison of grain pictures of the three examined samples seems to prove that the sediment, covering the flat reg area between the talus and dune base, reflects strongly both the processes of chemical weathering and superimposing them aeolian processes. The chemical processes from the grain surface within the talus, whereas aeolian processes contribute in secondary grain modelling. A marked difference of the grain character is visible at the further 30 m distance towards dune base. The remnants of intensive processes of chemical weathering disappeare totally, replaced by abrasion aeolian texture. Such an important change indicates the long duration and intensity of the aeolian process surpassing the influence of the chemical weathering process.

The record of processes on quartz grains from huge dunes of Erg Admer is different. Here, the weathering and aeolian processes are simultaneous and of similar intensity. Predominance of round grains with distinctly formed silica crusts and forms of aeolian texture indicate the persistence of the proecesses.

AEOLIZATION OF DUNE SEDIMENTS OF COLD DESERTS OF LATE GLACIAL OF SELECTED AREAS IN EUROPE

To determine the duration of aeolian processes the dunes from Central Poland, Belgium and Southern Sweden were chosen. The substratum and dune sediments were examined each time. It was found, that, in the case of prolonged aeolian processes, due to selection, the share of quartz and of total heavy minerals resistant to mechanical abrasion (garnet, zircon, rutile, tourmaline, staurolite, disthene) increases, whereas the share of minerals of the mica group (biotite, chlorite, glaukonite, muscovite) decreases. The amount of round frosted grains increases exceeding 70%. In the case of short aeolian processes typical of Northern Europe, the sole difference observed was the increase of grain content in dune sediment with aeolian abrasion. In the case of Swedish dunes, the quartz grains of the substratum and of dunes do not differ when examined macroscopically and under light microscope. It is only the analysis using SEM that reveals, that the grains from dunes show slight aeolian abrasion of most edges and angles which is not observed on the grains from the substratum.

CONCLUSIONS

The results presented above are based on investigations conducted during a long time and not everywhere the full set of analyses of sediments was done, which nowadays the author would consider necessary. It creates some limitations in drawing conclusions. Nevertheless, the investigations seem to enable to determine the duration of aeolian processes.

In case of the first two explored areas (the deserts of South-Western Asia and the desert of Central Syria) aeolian processes have not been able to change the shape of quartz grains as well as have not blurred the features of the sediment indicating its close relation with the substratum.

Aeolian processes within the examined dunes were recorded only in an increase of the content of fraction 0.25–0.1 mm (optimum for aeolian transport), a low mineral selection (the increase of epidotes and the decrease of amphiboles content in the crest parts of the dunes) and in the shape selection (accumulation in the dune crests of grains of weaker rounding in relation to the base of the proximal slope). The latter feature in dunes of long duration of aeolian activity is blurred by long lasting abrasion in the crest parts and is generally invisible. The differentiation of grain shape observed in transverse profiles of the examined dunes seems then to indicate a short duration of aeolian processes.

The analysis of grains in the electron microscope (SEM) showed also, that only a small part of the surface of the examined quartz grains exhibits the aeolian texture, marked only within already existing silica crusts. It was never observed the aeolian abrasion to remove this crust. This seems to indicate a relatively short duration of aeolian activity in the examined area.

The duration of aeolian activity seems to be indirectly an index of the length of desertification processes. On the other hand, the presence of silica crusts without distinct aeolian texture is not such an index. They may occur also on grains originating from temperate climate with well-marked dry season.

In the case of Central Syria a short duration of aeolian processes and indirectly of desertification of this area is historically confirmed. The existence of the desert in Palmyra region is estimated for the last 200–300 years.

Quite different was the third explored area (Central Sahara, foreground of Tassili-N-Ajjer plateau) where the impact of aeolian transport on the quartz grains of the examined dunes was common. It was visible as dominating over the process of chemical weathering (Timraz) as well as equivalent to the influence of this process (Erg Admer). This indicates a long duration of aeolian activity and indirectly of desertification processes. There are many archaeological proofs for several thousand years of desert lasting in this area.

The present results show many similarities with those of the sedimentological investigations carried out on dunes of the cold desert of Europe of the Late Last Glacial. In this region where cold deserts lasted several thousand years (e.g. Central Poland) the dune deposits differ distinctly from those of the areas where the duration of cold deserts was limited to several hundred years (Southern Sweden).

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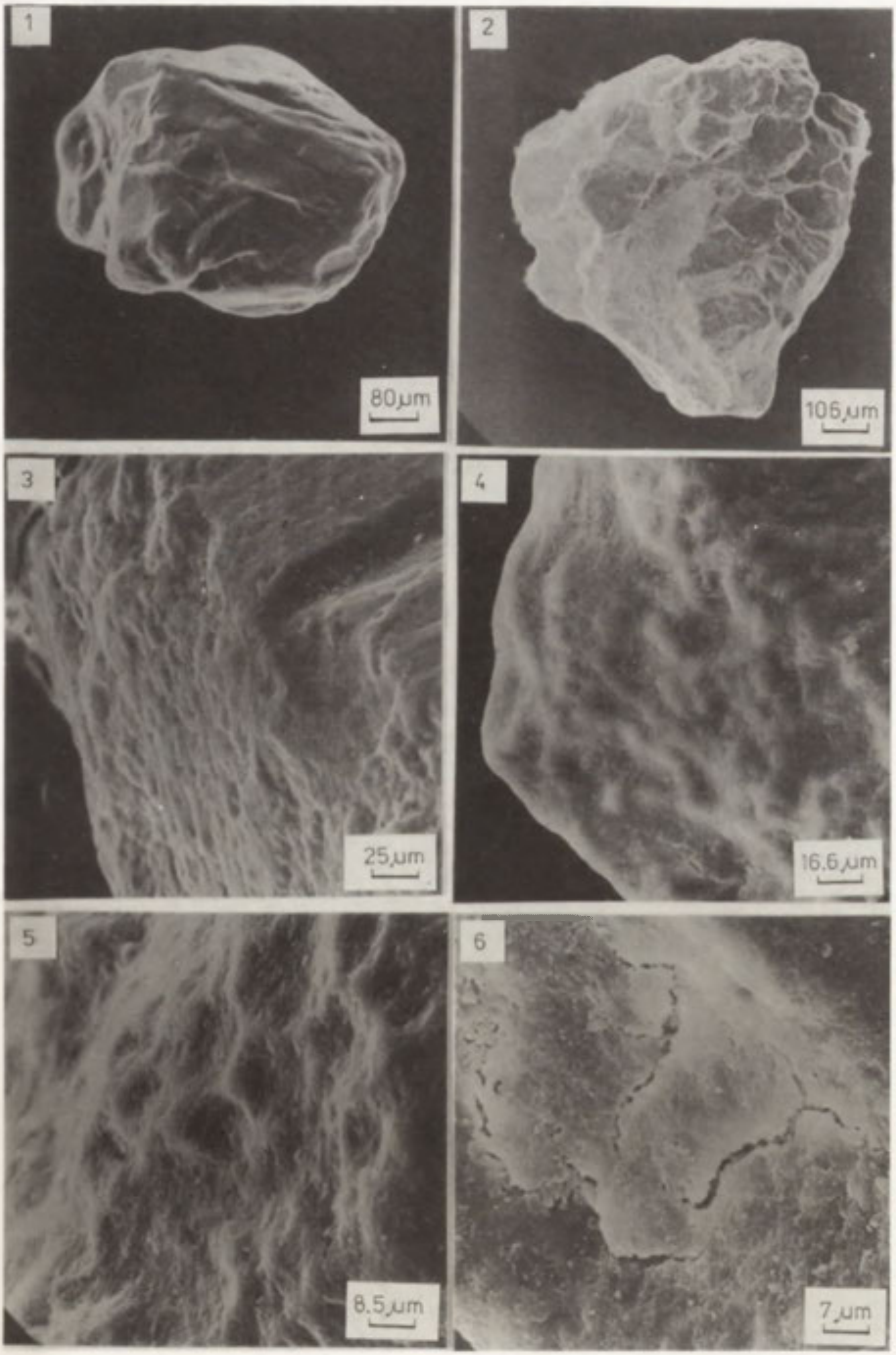
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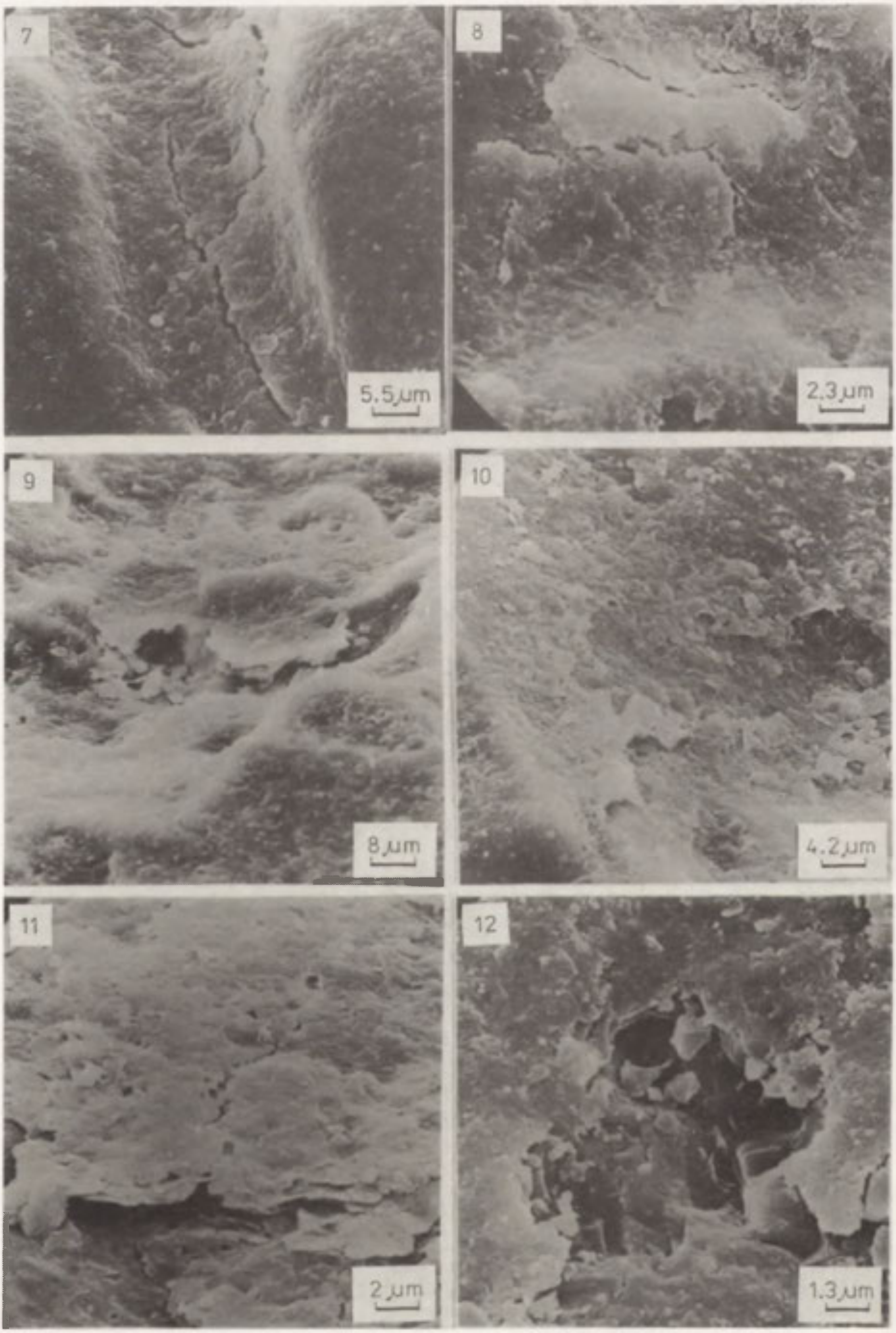
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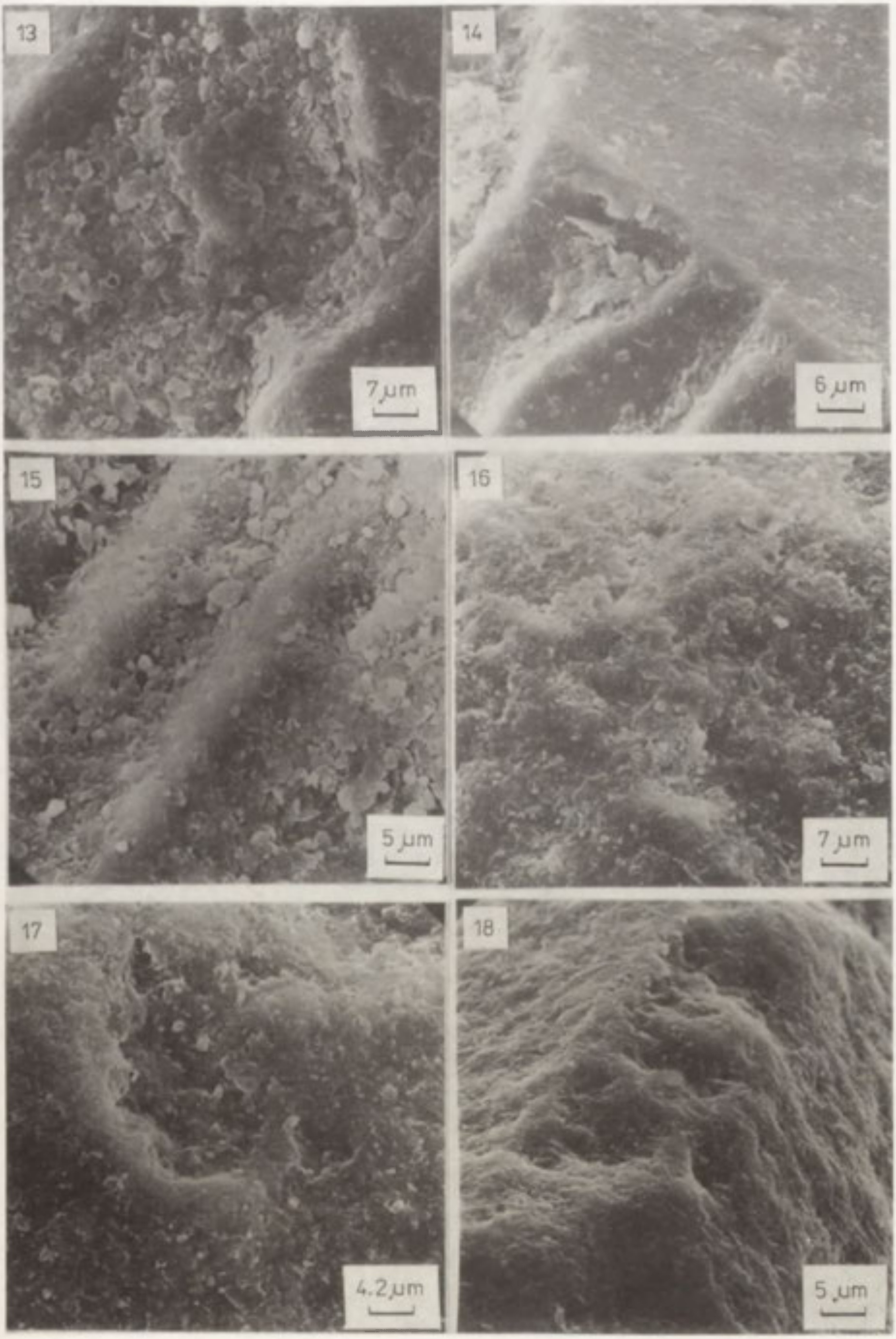
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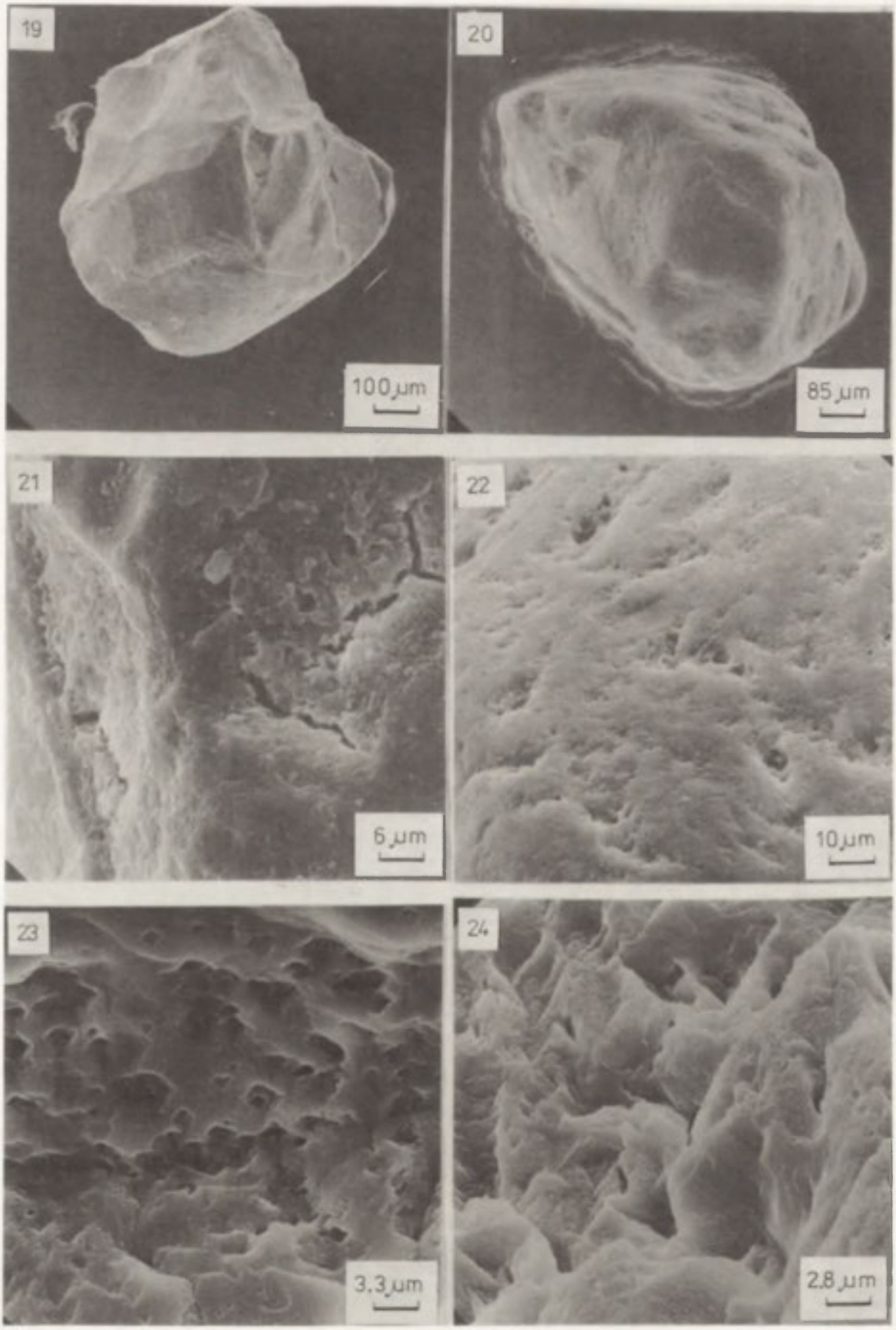
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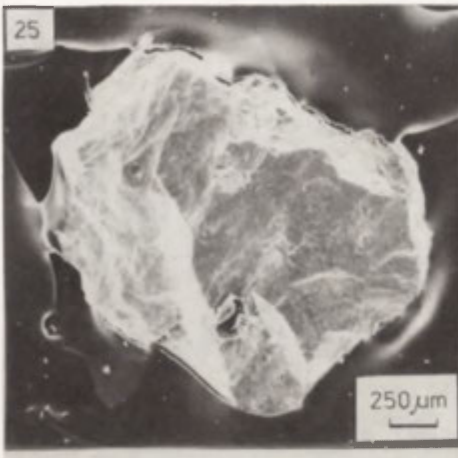
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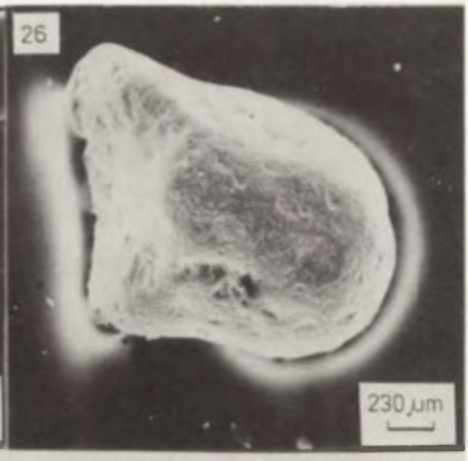
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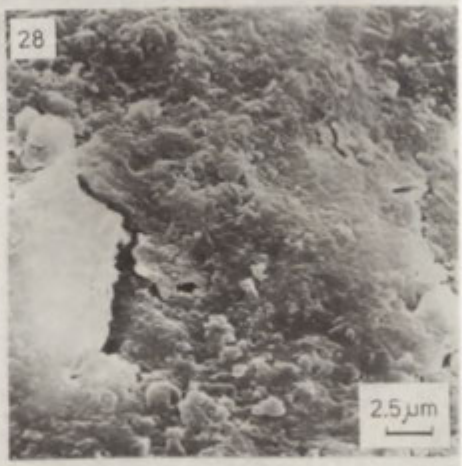
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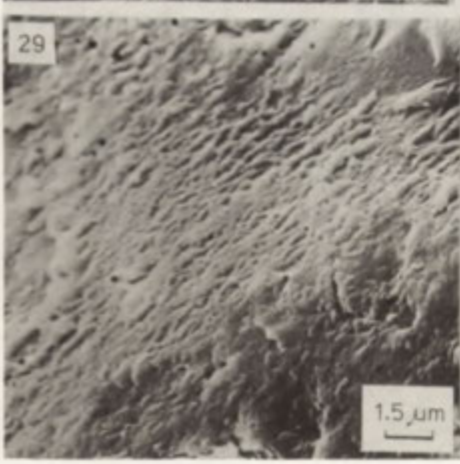
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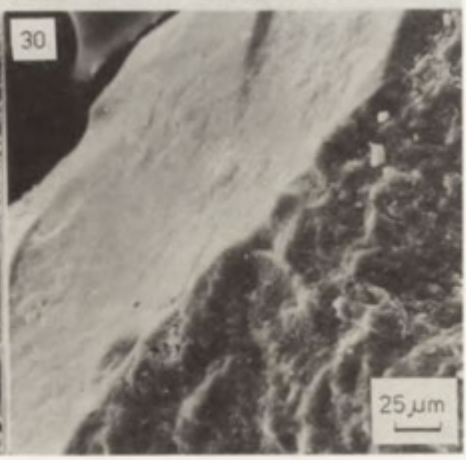
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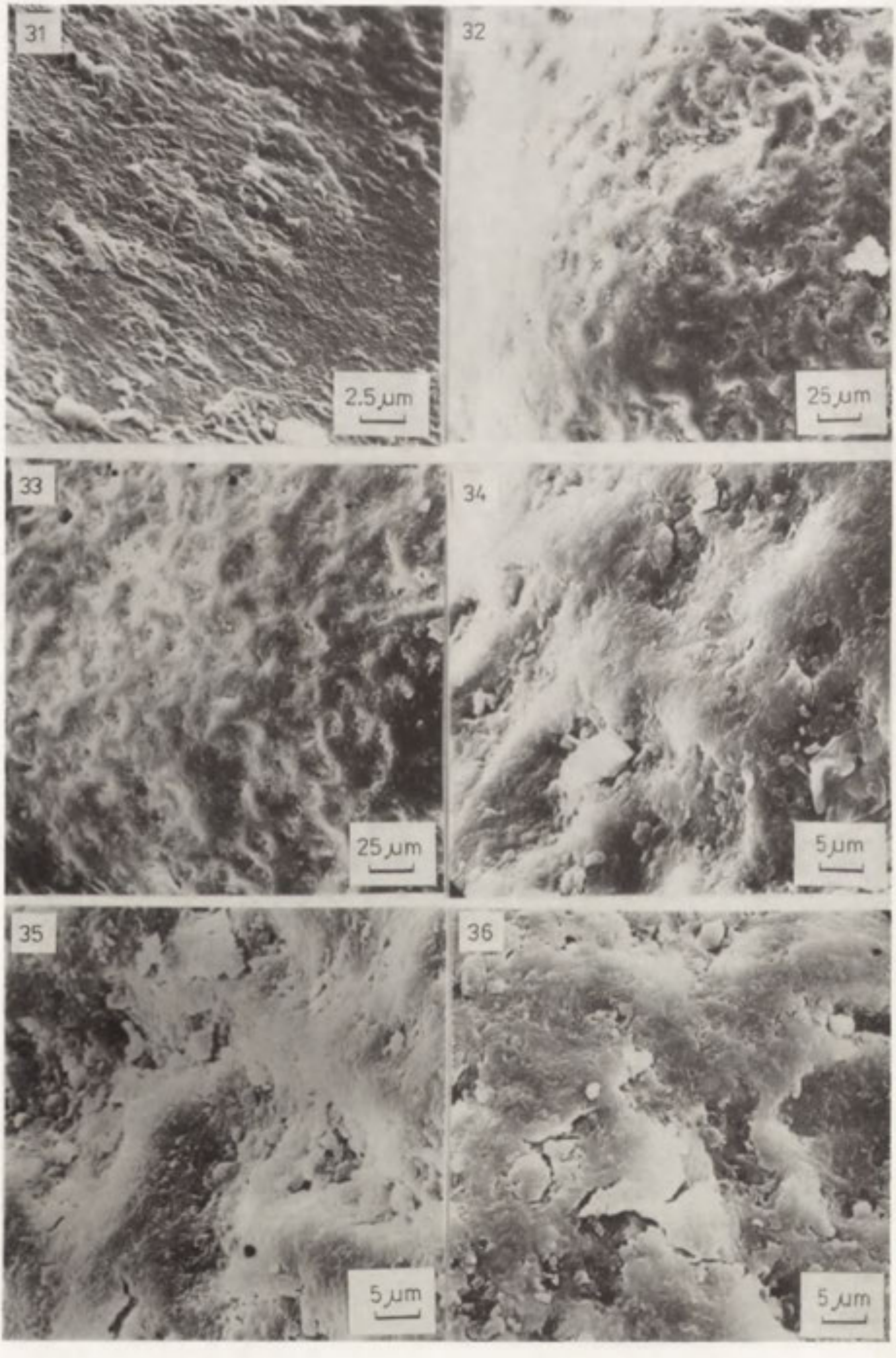
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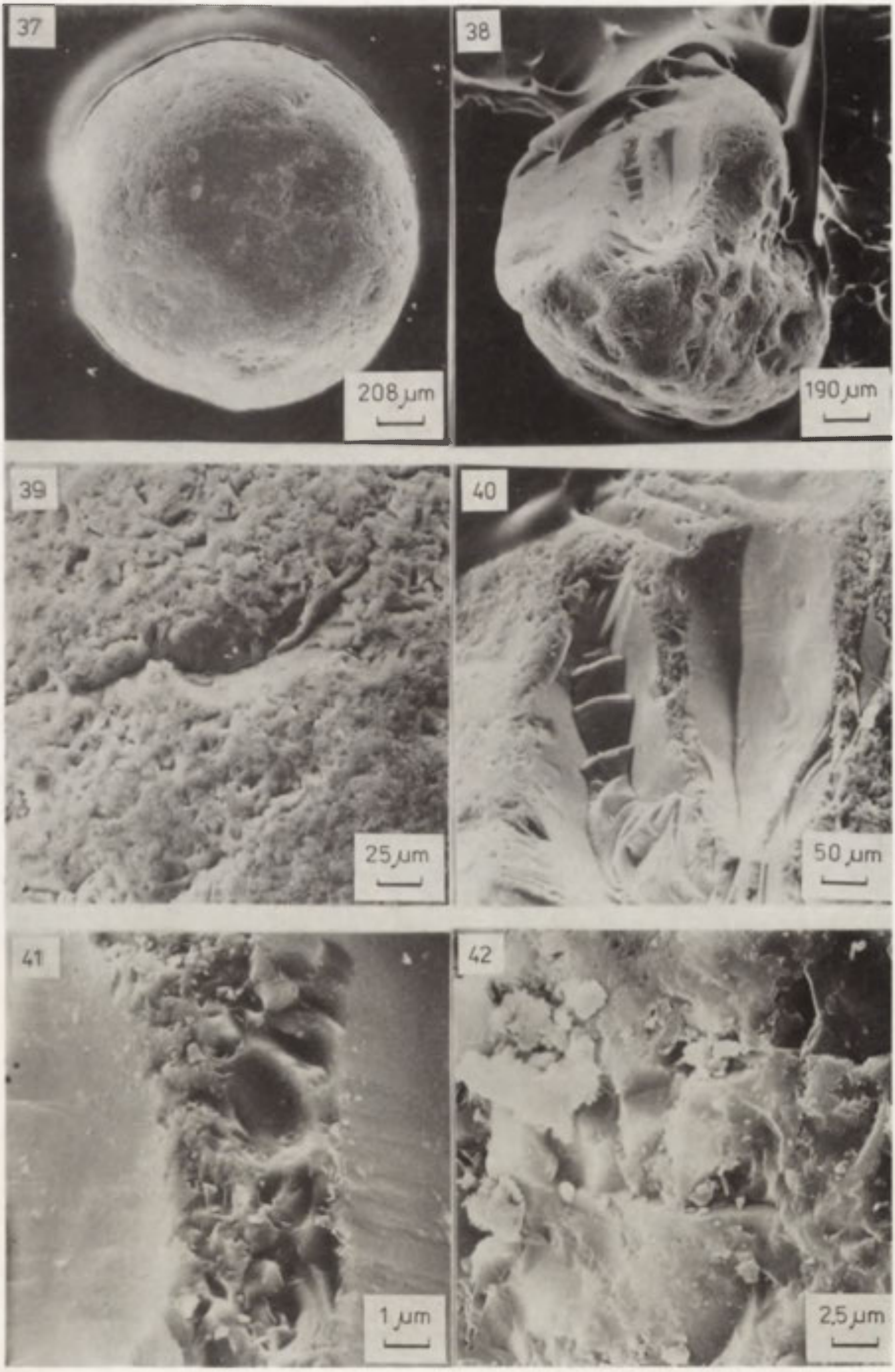
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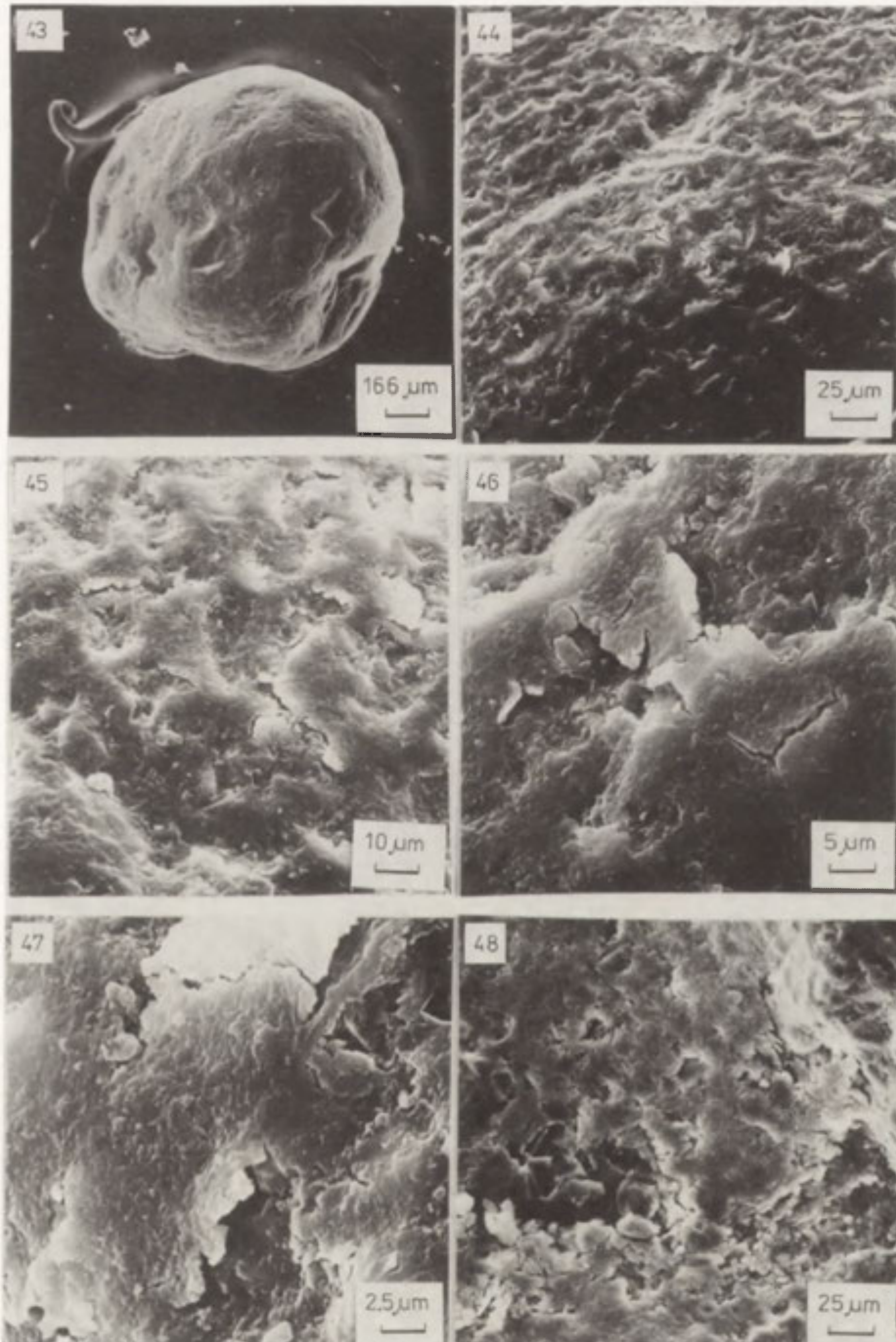
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LIST OF PHOTOGRAPHS

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TROPICAL KARST AND CHEMICAL DENUDATION OF WESTERN CUBA

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ABSTRACT. The results of geomorphological and hydrological investigations, carried out in the Western Cuba, have been presented. The investigations were carried out in the catchment basin Cuyaguateje located in the Sierra de los Organos and two small subterranean catchment basins in the massif Pan de Guajaibon in the Sierra del Rosario. The differences between ways of water circulation in a well developed upland karst region (Cuyaguateje) and in a well isolated limestone massif (Pan de Guajaibon) have been shown. The former is dominated by an underground transit water flow through mogot type hills, the latter is dominated by a local circulation of waters in the subterranean basin discharged by a central conduit. According to several years of investigations, elements of the hydrological balance, changes in chemical composition have been determined and values of chemical denudation have been calculated. Large amounts of underground waters have been emphasized. They result from the intensity of the rainfall, and from fast retention of rain waters in a karst massif. It causes an intensive process of chemical denudation, which reaches in mean perennial values, 90 $m^3/km^2/Y$. Such results are representative for tropical karst and they have been compared with chemical denudation values in karst regions in other climatic zones.

INTRODUCTION

Cuba is one of the best known karst areas in the Caribbean. Low mountain ranges, belonging to the Alpine system, spread along the axis of the island. They are mainly composed of Jurassic limestone and chalk. Large surfaces of marine terraces are located on the coast of the island (including shallow Pleistocene submarine terraces), they are built of Tertiary limestones. Genuine tropical karst has developed in these two graeat morphological units (Fig. 1). Mogote type karst and tower type karst have developed in the western part of Cuba. These two types of karst are the most thoroughly investigated in the Sierra de los Organos.

Tropical karst of the Caribbean is mainly known thanks to H. Lehmann's works, particularly a few geomorphological maps prepared for a karst atlas (Lehmann 1960). Recent knowledge of the karst areas of Cuba is significantly developed by Cuban investigations, carried out in cooperation with a large international group which included, among others, Czechoslovak and Polish scientists. It resulted in preparation of karst plates included in the

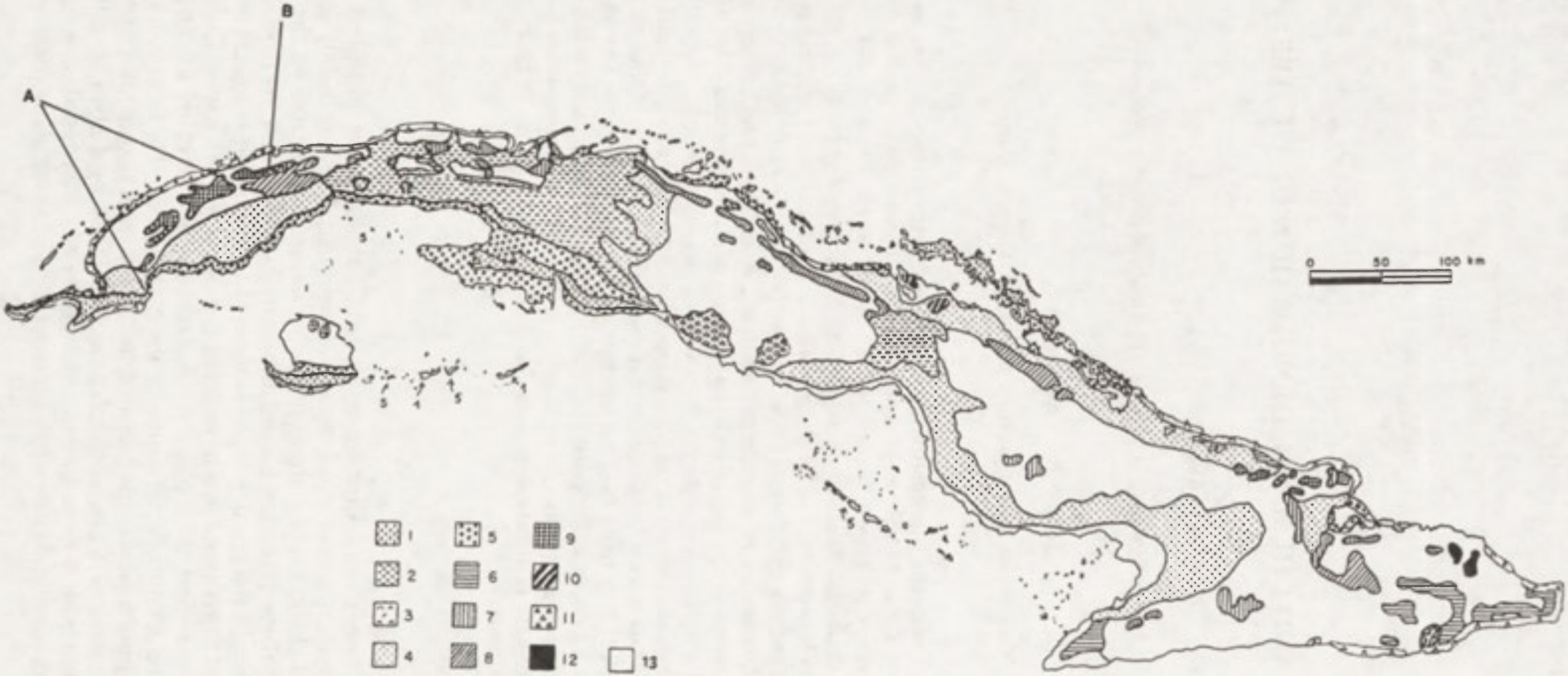


Fig. 1. Karst areas of Cuba after Nuñez Jimenez (1984)

Plains: karst in exposed limestones: 1 — naked karst, 2 — karst covered by soil, 3 — lithoral karst in marine terraces; buried karst: 4 — karst covered by thick fluvial and laterital sediments, 5 — coastal swamp karst. *Uplands* and *Mountains*: karst in limestones and marbles: 6 — platform karst, 7 — "stack" karst, 8 — table karst, 9 — cone and tower karst. Karst in limestone and gypsum: 10 — flattened "stack" karst. Karst in interbeded limestones and marbles: 11 — table karst and stone karst. Karst in serpentinites: 12 — high platform karst. Nonkarstic areas — 13. A — Sierra de los Organos (the river Cuyaguateje basin), B — Sierra Rosario (Pan de Guajaibon)

New National Atlas of Cuba (1988) and numerous joint reports[1], as well as geological maps of karst regions[2]. Apart from geomorphological research, hydrological and hydrochemical[3] investigations are carried on. Previously, they were carried out in the western part of Cuba, at present they are carried on in the central part of Cuba (Matansas province and Zapata peninsula) and in the Havana region.

Polish and Czechoslovak scientists are greatly interested in this tropical karst area of Cuba[4], as there in an urgent need to reconstruct paleokarst relief of Cracow-Silesia Upland, Northern Moravia and the Sudetes Mts. Usefulness of such comparative research has been proved by M. Klimaszewski (1958), S. Kozarski (1963) and J. Głazek (1966) in their previous investigations concerning tropical karst of south-east Asia, especially, due to the fact that Cuban tropical karst seems to be more similar to the tropical paleokarst of central Europe, than the other ones.

The article presents results of geomorphological and hydrological investigations carried on in selected catchment basins of western Cuba. They represent two types of the tropical karst: classical upland karst of the Sierra de los Organos (catchment basin of the Cuyaguateje river) and an isolated limestone massif of the Pan de Guajaibon. The values of contemporary chemical denudation being the coefficient of geomorphological activity, have been calculated in several catchment basins.

CLASSICAL TROPICAL KARST – SIERRA DE LOS ORGANOS

Classical tropical karst covers over 500 km^2 in the western part of Cuba. It is located in the Sierra de los Organos (Fig. 1). Mountain ridges and isolated limestone tables rising 100–300 m above a flat level of large polje like depressions are characteristic for the area. Those hills, of cupola shapes, called mogotes, are cut through by horizontal caves situated on at least four levels. After M. Acevedo (1971) the levels date back to the Paleogene (the highest one – 350–400 m a.s.l.), the Miocene (350 m a.s.l.) the Plio-Pleistocene (300–195 m a.s.l.) and the Holocene (80–100 m a.s.l.). The lowest one is located inside the flooded terrace of the Cuaguateje river and its tributaries.

The mogotes are perforated also by vertical shafts called *hojos* of the diameters from several to less than a hundred metres. Old levels of the caves appear on their walls and the bottoms may reach the subterranean rive. Old *hojos* forms are similar to deep sinkholes.

Large karst depressions occur both inside the karst area (typical polje) and on its bordeers (contact or border polje). All mesoforms from classical dolines and uvalas, karst spring valleys and sinkhole valleys up to typical for Caribbean karst – large shafts called *senotes* or *cassimba* (a kind of karst depression) and *caleton* (a kind of karst spring velley) are quite numerous there. Surfaces of the mogotes are covered by various types of karst microforms, beginning from oval forms with characteristic pans called *tinejitas* up to all possible linear forms – from classic karren to sharp microforms called here *diente de perro*.

Geomorphology of this classical tropical karst has been presented in several papers (i.e. Acevedo 1971; Nuñez Jimenez, Panoš, Štelcl 1969; Nuñez Jimenez 1984), however the first

[1] Bibliography of previous papers is collected by Nuñez Jimenez (1984).

[2] Information on geological reports and their bibliography are collected by A. Pszczółkowski et al. (1982). Karst investigations in that geological programme were carried out by J. Rudnicki from Warsaw and J. Dzulyński from Cracow.

[3] They are carried on in the international project PIGEK, affiliated by UNESCO. Direct cooperation is established between scientists from the Silesian University and Centro Nacional de Investigaciones Cientificas in Havana.

[4] Results of karst investigations in Cuba were presented many times during Speleological Schools organized by the Silesian University and the Wroclaw University.

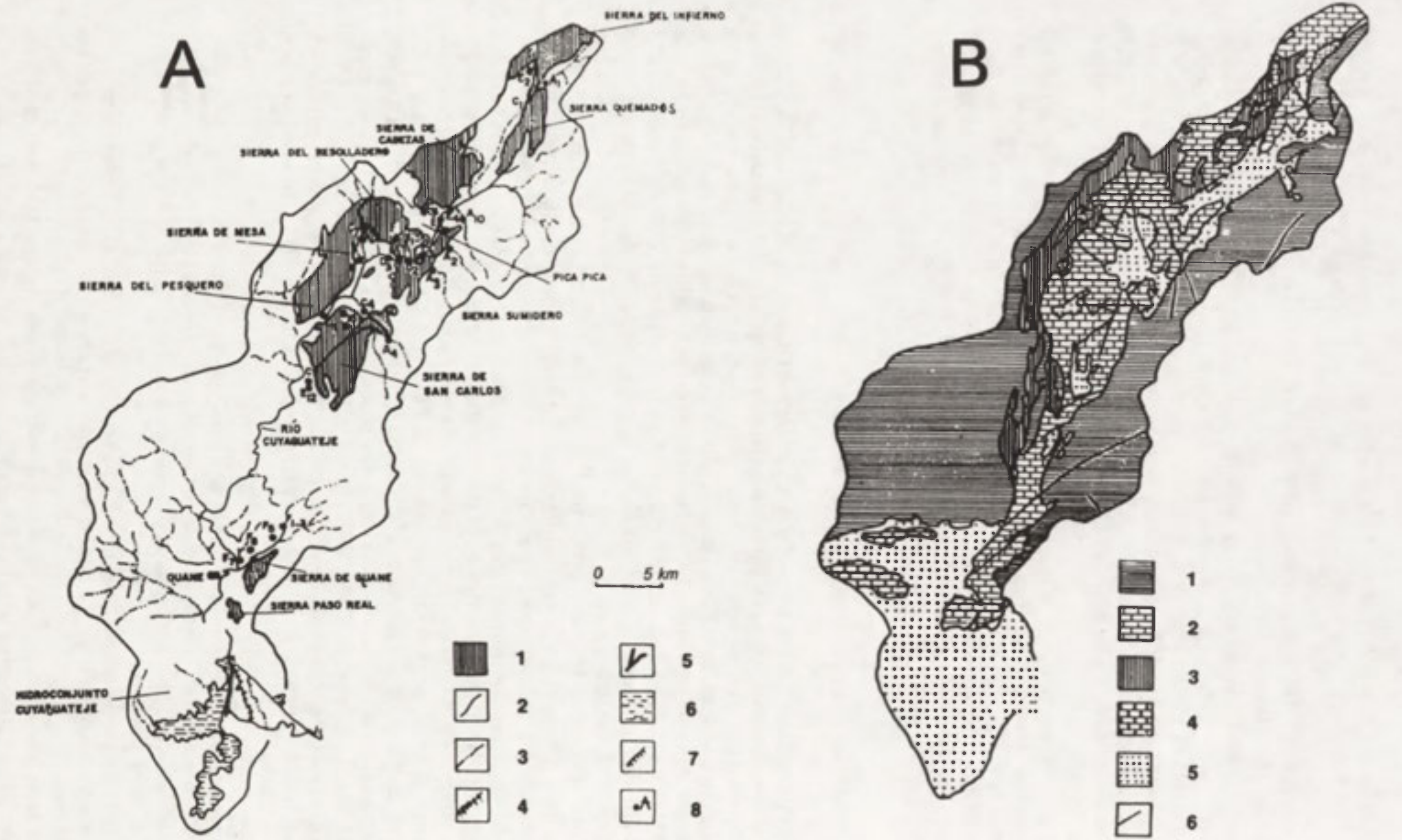


Fig. 2. The basion of the Cuyaguateje river in the Sierra de los Organos (W Cuba)

A — topographic sketch: 1 — karst massifs, 2 — permanent streams, 3 — intermittent streams, 4 — dry river bed, 5 — water conduits in subterranean caves, 6 — lagoons and lakes, 7 — bank of a dam, 8 — points of hydrological and hydrochemical measurements. B — geological sketch (geology after Piotrowska and Pszczółkowski 1978): 1 — terrigenic fromations (J_{1-3}), 2 — limestone fromations (J_3-P_g), 3 — terrigenic formations (P_g-N), 4 — Peso-Real limestone formation (N_1), 5 — loose fluvial sediments, 6 — faults

TABLE 1. Chemical denudation in karst regions in the Sierra de los Organos and the Sierra del Rosario in W Cuba

Basin	Year	S km^2	Q m^3/s	q $l/s \cdot km^2$	ΔT[1] mg/l	A $10^3 m^3/y$	D $m^3/km^2 \cdot y$	D $t/km^2 \cdot y$	$D\varepsilon$[2]
Sierra de los Organos (Cuyaguateje basin)									
V Aniversario	1979–1981	145	4.0	28	150	7.6	52.9	132.3	0.84–1.15
La Guira	1979–1981	279	7.7	28	195	18.9	68.8	172.0	0.88–1.12
Portales II	1979		22.7	45		68.6	136.1	340.2	
	1980		12.5	25		37.8	75.6	189.0	
	1981		8.9	18		26.9	54.4	136.1	
	1979–1981	502	14.7	29	240	44.4	87.7	219.2	0.97–1.04
Sierra del Rosario (Masizo Pan de Guajaibon)									
Ancon, Canilla	1984–1985	9.45	0.307	32.5	144	0.537	59	147	
Ancon – autochtonous basin	1984–1985	4.35	0.207	47.6	154	0.402	(92)	231	

[1] ΔT – $CaCO_3(Ca^{2+} + Mg^{2+})$ [2] the range of denudation values = $D\varepsilon$

geomorphological investigations on tropical karst carried out by H. Lehmann seem to be the most significant. The map of the Siera de los Organos prepared for the atlas of karst regions (Lehmann 1960) was the result of the investigations.

There are large horizontal cave systems known in the Sierra de los Organos, among them Majaguas-Cantera and Santo Tomas are longer than 20 km. The caves have been discovered and explored by Cubans themselves as well as by international expeditions, among them one of the first was the Cuban-Polish expedition in 1961. The paper by A. Gradziński and A. Radomski (1963), concerning the origin of caves in the Sierra de los Organos, dates back to that period. The book by A. Nuñez Jimenez (1984) presents a complete speleological bibliography concerning Cuban karst.

The hydrographic system of the Sierra de los Organos is thoroughly investigated. There is a net of long-period hydrometric stations of the National Institute of Hydroeconomy, which has a department in Pinal del Rio. The interest in this region has grown significantly for the last few years mainly due to the plans to construct water dams and water intakes for household and farm needs. Systematic hydrochemical analyses of water, also for scientific purposes, have been carried on since 1979 (Fagundo et al. 1981, 1987). Hydrological investigations were carried out in three sub-basins of Cuyaguateje (Fig. 2A, B), closed by water level indicators, in 1979–1981 (Table 1). The examinations have been repeated later.

HYDROLOGICAL AND GEOLOGICAL SKETCH OF BASIN CUYAGUATEJE

The south-wastern part of the Sierra de los Organos is discharged by the river Cuyaguateje (Fig. 2A). It begins in the Sierra del Infierno and after about 100 km ends in the Caribbean Sea at the Bahia de Cortes. The basin covers over 840 km^2. Table 1 presents chosen hydrological data form the basin. The mean perennial flow in the river is 14.7 m^3/s, it gives elementary run off $q = 29$ l/s/km^2 for the whole basin. Such relatively high run off coefficient is not only caused by the amount of precipitation — more than 1700 mm per year, with 20–22% of them in dry season (between November and April). High mean annual air temperature (22–24°C) causes intensive evaporation. However, most of the precipitation waters immediately infiltrates into the massif, where surface evaporation does not exist. On the contrary, water vapour condensation is very intensive there, especially in caves.

Internal part of the basin is built of carbonate rocks formations (Fig. 2B), mostly Upper Jurassic (Jagua formation — J_3) and Paleogene (P_g). These rocks are covered, in depressions, by loose fluvial or lateritic deposits, including thick covers of terra rossa. The basin is surrounded by nonkarstifiable rocks both sedimentary, mainly sand-clay types, and volcanic, from Lower and Middle Jura (J_{1-3}) and Cretaceous. Among them, there is a typical for western Cuba formation San Cayetano. The tectonic structure of the Sierra de los Organos is typical for young folded mountains of Alpine type. The influence of tectonics can be observed on the shape od large karst massifs such as e.g. Sierra Sumidero and in separation of single mogotes.

There are three hydrogeological zone in the karst massifs of the region. The upper-vadose zone reaches in large massifs and mogotes the depth of 300–350 m. However in most cases it does not exceed the depth of 100 m. The largest horizontal caves and *hojos* have developed there. The largest caves of western Cuba, mentioned above, are also there. The vadose zone is, still unexplored, an immense underground world. The surfaces of the massifs are intensively perforated and inflow of percipitation waters into this zone is instantaneous. The flow of waters is gravitational. Precipitation waters reach a transitional hydrogeological zone, situated between a vadose zone and a phreatic one, very quickly. The zone is built of a system of water table caves filled up periodically with water. Large caves, with the subterranean river Cuaguateje and its tributaries, are located in the upper part of this zone. During cyclonic rains the water level rises from several to almost twenty metres. The poljes are partly flooded, as the caves cannot discharge the whole amount of waters flowing through them. The caves are typical subterranean conduits formed mainly by mechanic activity of water.

Numerous exploratory drillings for the dams and hydrochemical investigations have proved occurrence of a phreatic zone. Its deep part is discharged by a system of springs Portales II. The phreatic zone in the mouth part of the basin contacts directly with the sea waters. The shallow part of the phreatic zone consists probably of large caves which have been formed in the transitional zone. Such concept is supported by recent results of under-water explorations in the caves. They were probably formed in glacial periods when the denudation level was lower than the present one. It was connected with the oscillation of the world ocean in the Pleistocene.

KARST PHENOMENA IN THE MASSIF PAN DE GUAJAIBON IN THE SIERRA DEL ROSARIO

Pan de Guajaibon is an isolated limestone massif, located in the northern part of the Sierra del Rosario. The massif, covering less than 20 km^2, consits of two parallel ridges: Pan de Guajaibon and Sierra Chiquita (400–692 m a.s.l.) divided by a morphological depression (about 300 m a.s.l.). The northern and the western slopes of Pan de Guajaibon from a rock cliff, falling down to the base of the massif at 100 m a.s.l. The southern slopes are also steep, but the walls are lower. The south-eastern part forms wide surfaces above 300 m a.s.l., connected with the main ridge of the Sierra del Rosario (Fig. 3).

The ridge of the massif Pan de Guajaibon is formed by large karst cupolas like mogotes, separated by deep passes. The highest of these mogotes reaches 692 m a.s.l. and the relative altitude 0.5 km. There are numerous shelves, with cave entrances, on slopes of the cupolas. The ridge of the Sierra Chiquita is lower and forms a regular mountain ridge. The surface between the ridges is covered by many karst depressions passing into vertical caves. The most common hare are large megadolines, with steep walls and shaft openings. There are also depressions with centripetal drainage systems with sink-holes at the contact between impermeable deposits and limestone substrata. A great blind valley has been formed between two ridges, in the western part of the massif. It is connected with headward erosion of the Ancon cave karst spring (Fig. 4).

Pan de Guajaibon, as an isolated massif steeply rising above a hilly area, reaching the Atlantic coast, has its specific microclimate. It is characterized by the abundant precipitation with a large vertical gradient. According to perennial data (Trusov et al. 1983), the annual sum of precipitation in the area reaches 1800 mm at the foreland of the massif and 2500–2600 mm at the ridges (Fig. 5). The precipitation is distributed into two seasons, a dry one betwen November and April when 25–30% of annual precipitation occurs and a wet one between May and September. Most of the rains, typical of Caribbean, are cyclonic rains (a stormy, extremely intensive rain, lasting from a few to twenty hours). The mean perennial air temperature is equal 23°C at the foreland and 20°C in the mountains (Fig. 6). The lowest temperatures occur in January (only a few degrees lower than the mean annual). In that time, 24 hours run of air temperature is characterized by high amplitude, often exceeding 20°C. At night, air temperature in the mountains may reach only few centigrades. Large differences in the air humidity, in 24 hours cycle, can be also observed. Despite the relatively high humidity in this climatic zone, the highest humidity can be observed at night and in the morning (supersaturation and condensation of water vapour).

Hydrometeorological investigations carried out in the Pan de Guajaibon in the years 1984–1985 brought precise data. Some of them, for the hydrological year 1984/1985 are presented in Figs. 5 and 6. That hydrological year, compared to the mean perennial values, had the annual sum of precipitation 75–80% lower and the air temperature slightly higher. The sum of the precipitation in this year varied from 1700 mm at the foreland to 2200 mm on the main ridge, including 500 mm and 700 mm in dry season respectively.

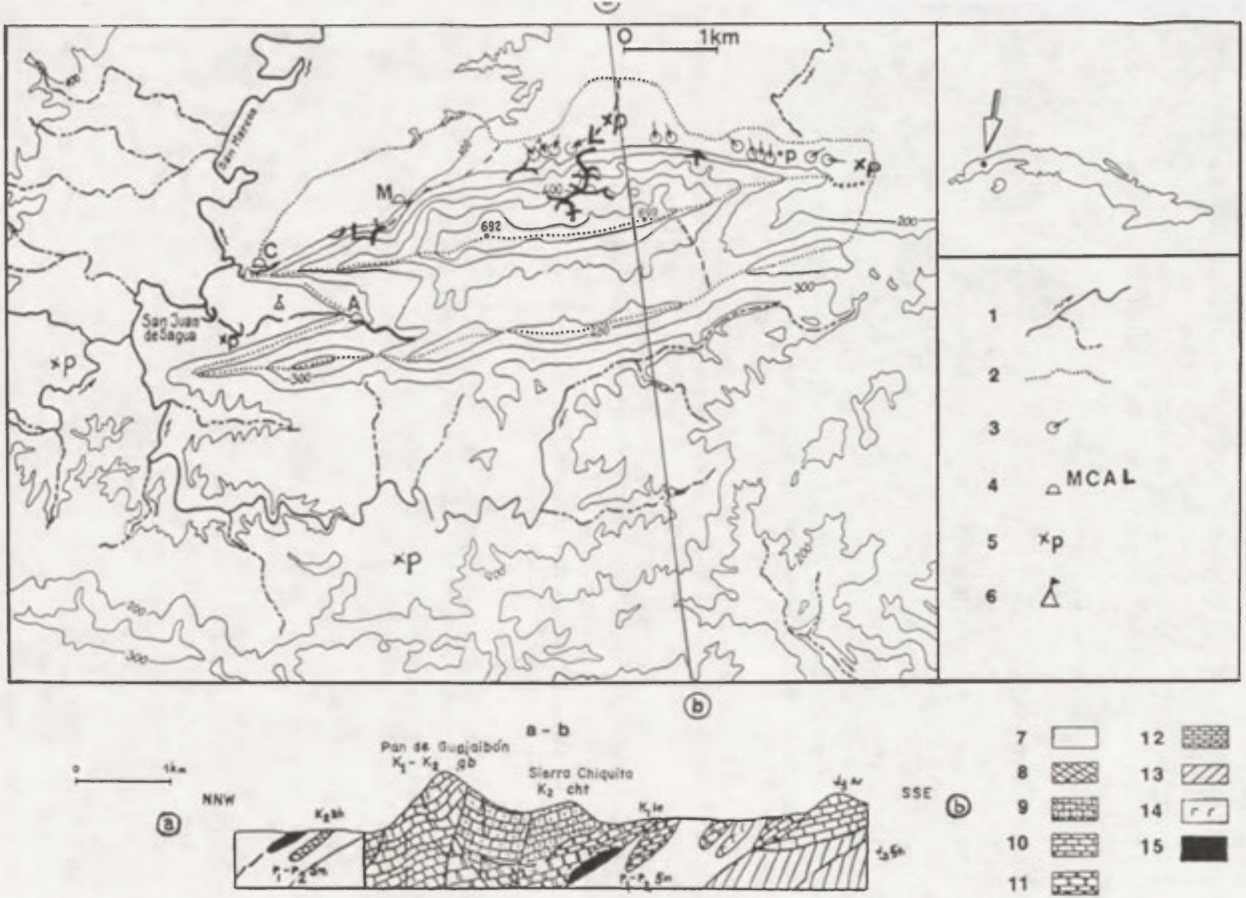


Fig. 3. Topographic sketch and geological profile of the Pan de Guajaibon massif (Sierra Rosario – W Cuba)
1 – surface water courses, 2 – limits of the subterranean basins Ancon and Canilla, 3 – sinkholes, 4 – caves (M – Mamey, C – Canilla, A – Ancon, L – Lechuza), 5 – location of hyetographs and rain gauges, 6 – field scientific station at San Juan de Sagua. *Nonkarstic rocks*: 7 – San Marcos formation ($P_1 - P_2$). 8 – Bahia Honda formation (K_2). *Karstic rocks*: 9 – Chiquita formation (K_2), 10 – Guajaibon formation (K_1, K_2), 11 – 12 – Lucas and Artemisa formations ($K_1 - J_3$). *Nonkarstic* formations: 13 – San Cayetano (J_3), 14 – gabbro intrusions, 15 – serpentinites (geology after Maksimov et al. 1978)

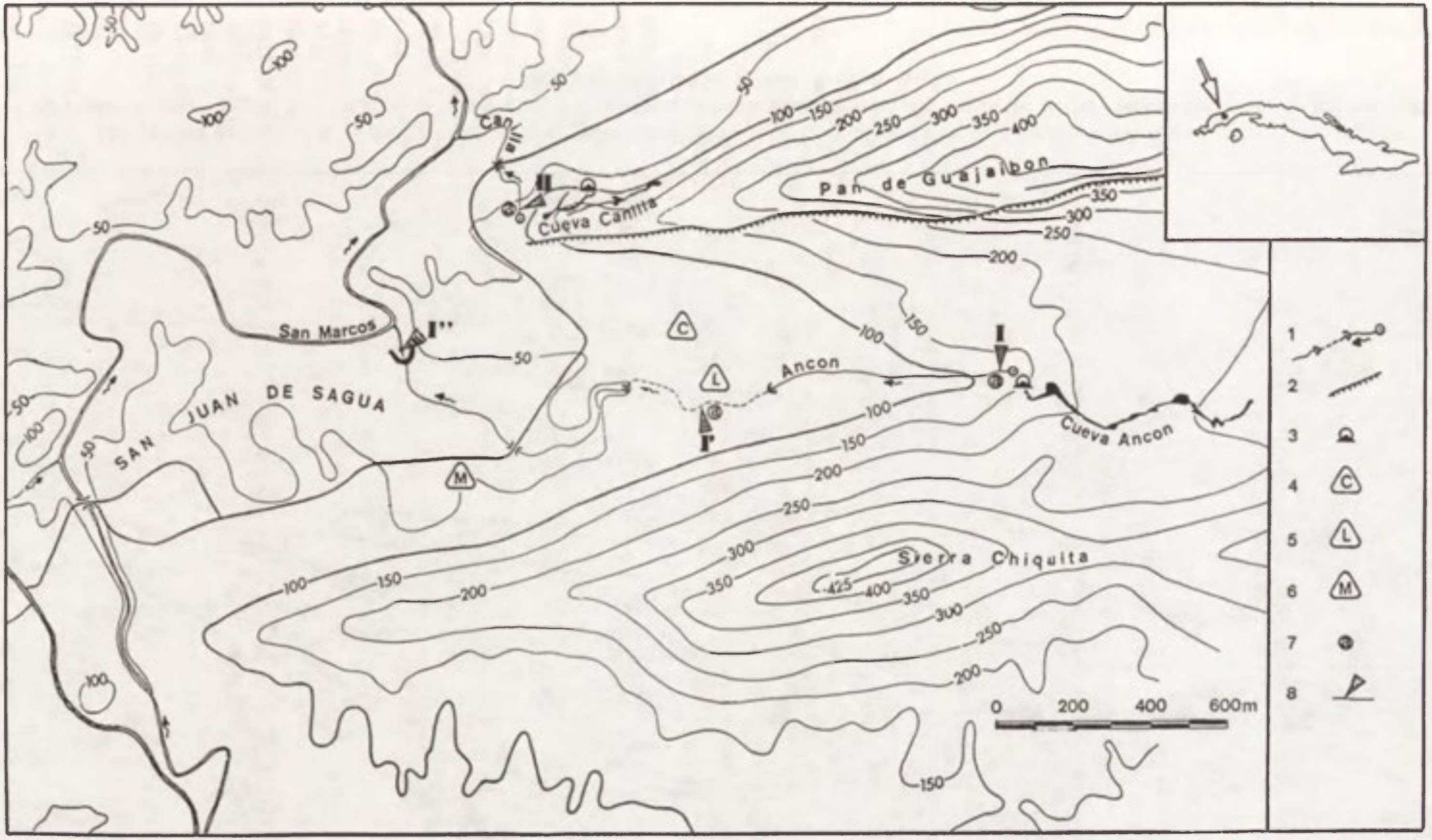


Fig. 4. Topographic sketch of the western part of mogotes Pan de Guajaibon and Sierra Chiquita
1 – surface stream with ponors and a karst spring, 2 – rock cliff, 3 – cave entrances, 4, 5 and 6 – scientific station, laboratory and meteorological station at San Juan de Sagua, working from 1984 till 1988, 7 – location of an automatic hydrochemical station (working from 1984 till 1988), 8 – water gauges; I – karst spring Ancon, I′ – river Ancon below first sinkholes in the vicinity of the scientific station, I″ – lower karst spring Ancon, II Canilla cave karst spring

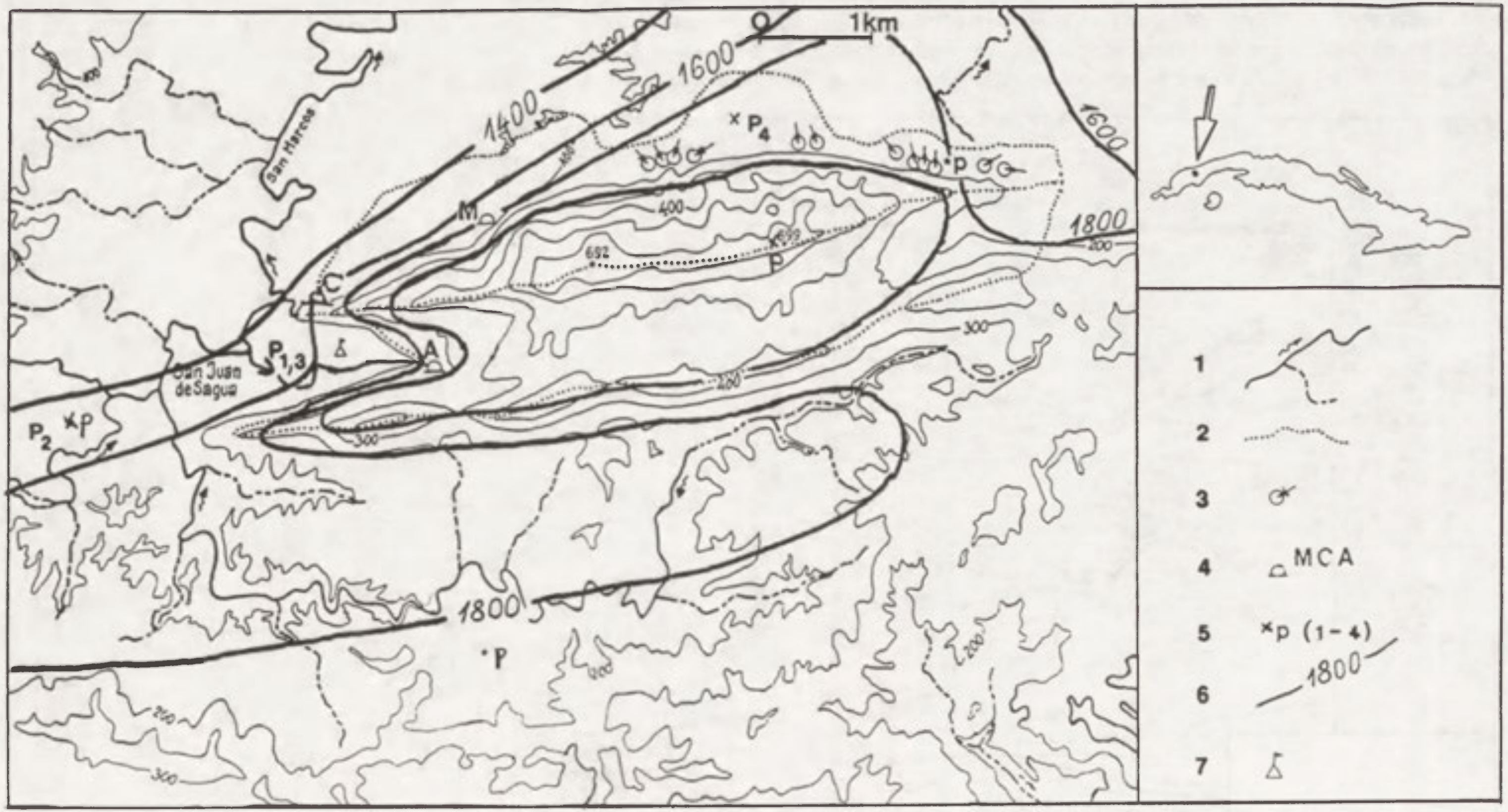


Fig. 5. Izohyets of annual sum of precipitation in the hydrological year 1984/1985 in the Pan de Guajaibon (W Cuba) after J. Rodriguez
1–4 see explanations to Fig. 3; 5 – location of rain gauges, 6 – isohyets of annual sum of precipitation in the period: November 1984 – October 1985, 7 – field scientific station at San Juan de Sagua

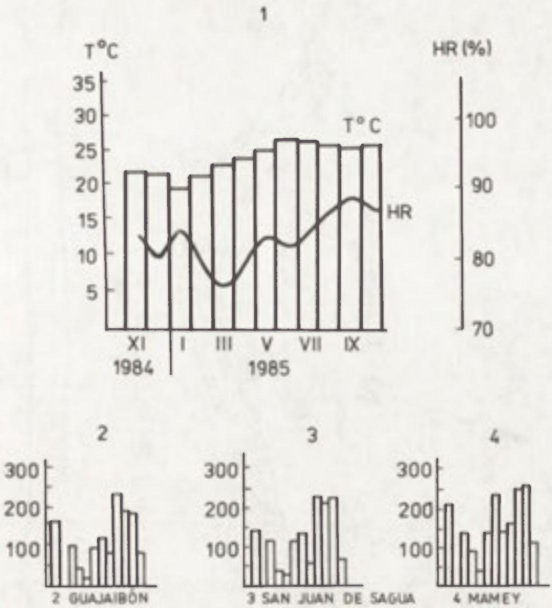


Fig. 6. Monthly changes of air temperature (T°), humidity (HR in %) and precipitation in the hydrological year 1984/1985 in the Pan de Guajaibon massif (W Cuba), prepared by J. Rodriguez
1 and 3 — meteorological station San Juan de Sagua, 2 — rain gauge Guajaibon, 4 — rain gauge Mamey. Location of the meteorological station and hyetographs in Fig. 9

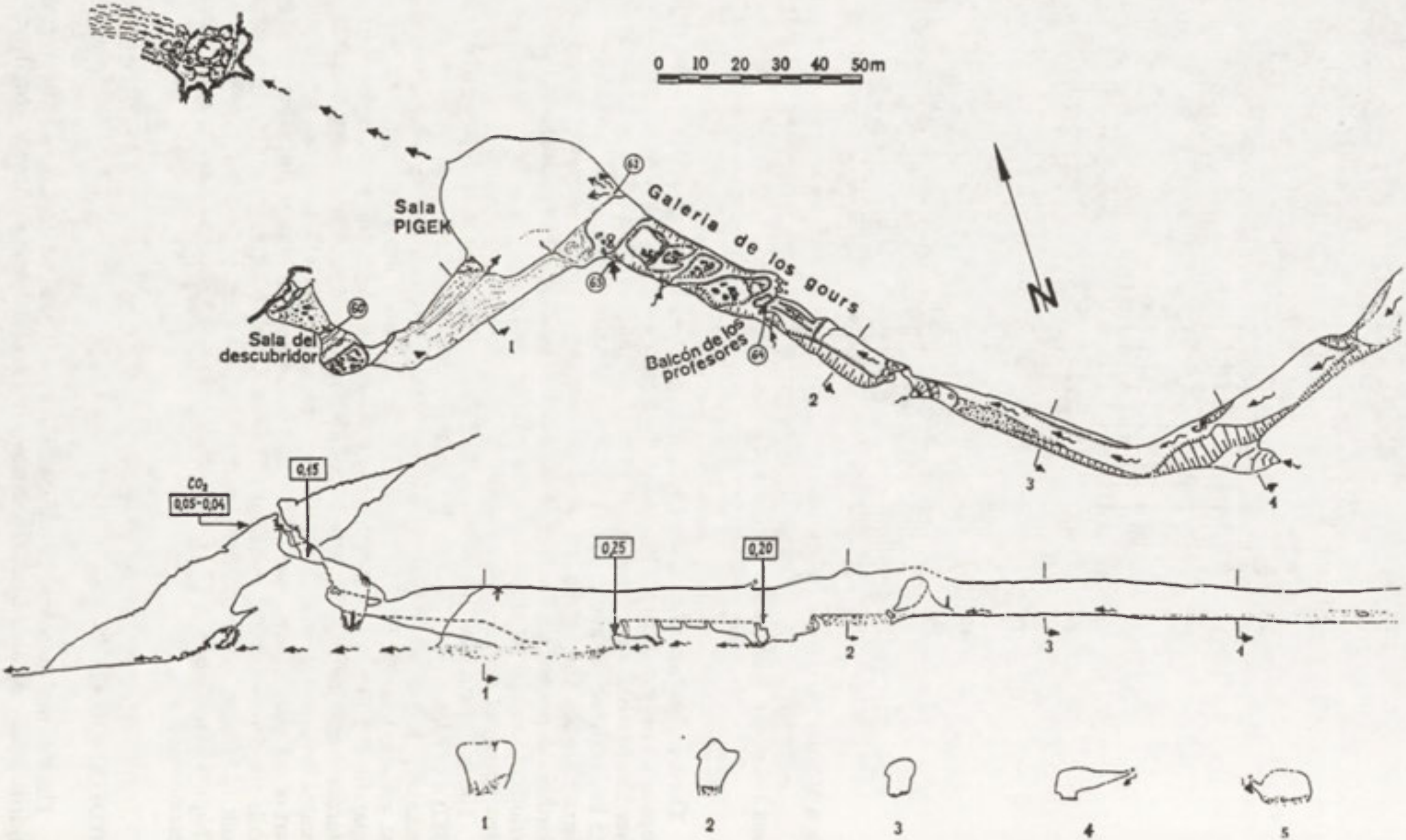
The discussed massif is covered by a dense, hardly accessible vegetation cover typical of the tropical Caribbean Region. The massif Pan de Guajaibon is discharged by a system of the river San Marcos, one of the largest rivers of the northern coast of western Cuba (basin area 155 km^2). The upper part of the valley of San Marcos discharges southern slopes of the ridge Sierra Chiquita and catches karst waters emerging from the springs of Ancon and Canilla. Northern slopes and the foreland of the massif are discharged by surface water courses, mostly periodical, drained by sinkholes of an underground system of conduits connected with the karst spring of Canilla.

The geological structure of the massif (Fig. 3, Maksimov et al. 1978; Pszczólkowski et al. 1982) is of Alpine type. The Pan de Guajaibon and the Sierra Chiquita are built of Cretaceous rocks (K_1-K_2) mostly massive limestones with bauxite intercalations. Conglomerates, marls, etc. are also found there. The massif forms a large horst limited by faults. The northern fault separates the limestone block from the Paleogene nonkarstic rocks belonging to the San Marcos formation ($P_{g1}-P_{g2}$), with gabbro and serpentinites intrusions. Central depression, located between ridges, is also limited by large faults. Tectonic faults caused that the northern part of the massif, mainly the ridge Pan de Guajaibon, is built of thick limestofne formations while the depression and the ridge of Sierrra Chiquita, apart from limestone formations, is built of breccia, marls and nonkarstic rocks. The syncline between these ridges is filled by Chiquita formation (K_2). The faults are accompanied by breccia zones, ore mineralization and thermo-mineral springs.

HYDROGEOLOGICAL KARST ZONES

The Pan de Guajaibon, in contrast to the karst basin of the Cuyaguateje river, is a massif hydrologically isolated from the surrounding areas. Therefore an autochtonous hydrologic

A


0 10 20 30 40 50m
Sala PIGEK
Galeria de los gours
Sala del descubridor
Balcón de los profesores
N
CO_2
0,05-0,04
0,15
0,25
0,20
1
2
3
4
5

B

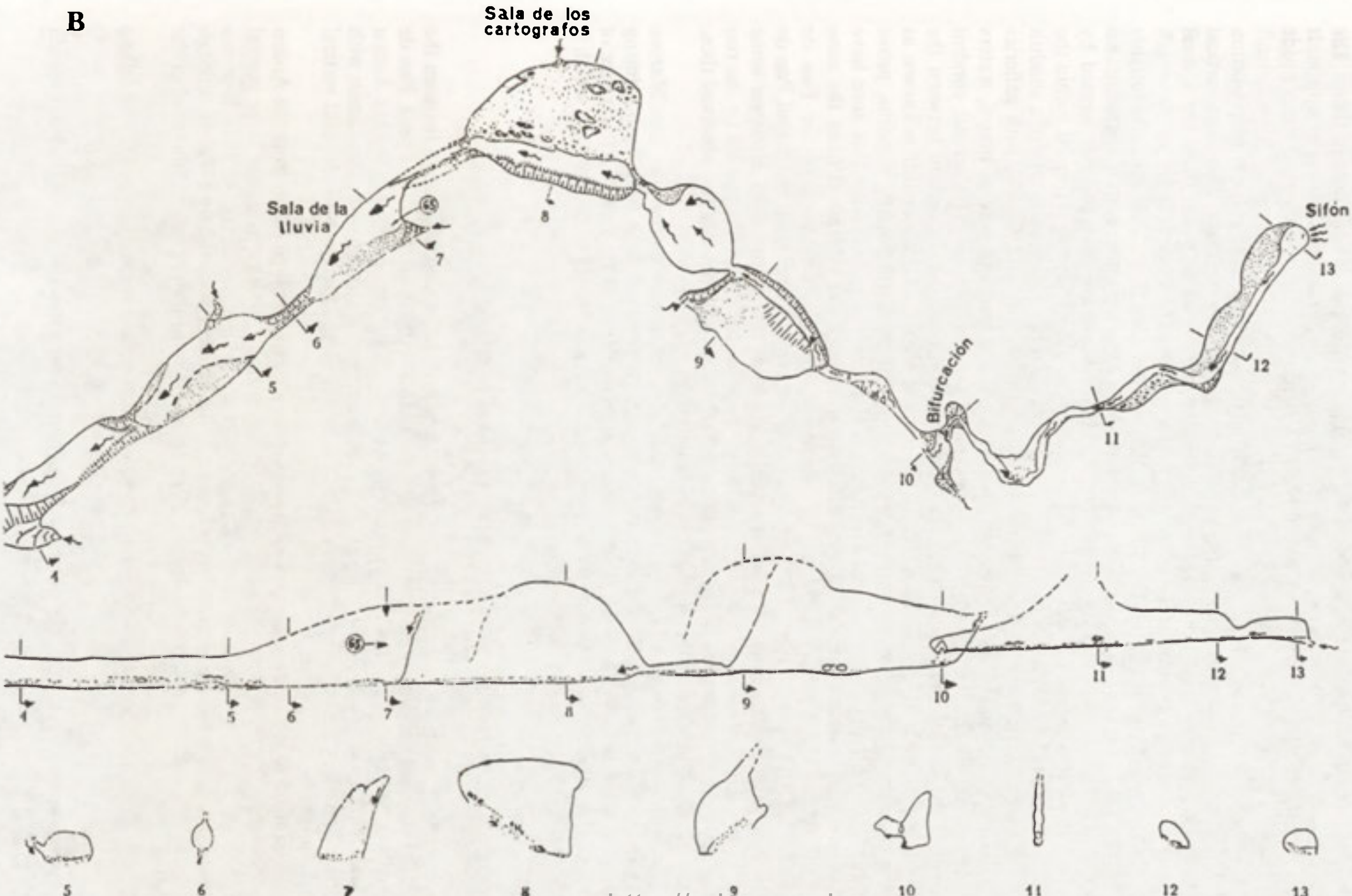


Fig. 7. Topographic plan and longitudinal cross-section of the Ancon cave in the Pan de Guajaibon massif — W Cuba (carto ra hy after A. Kozik). Location of the cave in Fi . 3

network, formed by precipitation waters falling directly on the massif, occurs there. The northern part of the massif charged by water from nonkarstic foreland is the only exception. It forms original subterranean basin of mixed type. There are three hydrogeological zones inside the massif.

A vadose zone is vast and reaches the depth of 0.5 km. The zone collects precipitation waters from the surface and quickly passes them to the transitional zone. Therefore vertical caves and fragments of old horizontal ones dominate in the vadose zone. There are several caves, currently known, with entrances located in the depression between the ridges that reach the transitional zone. However the caves in the vadose zone have not been thoroughly investigated. The caves of the transitional zone have been explored much better. There are two types of conduits in the zone mentioned above. The first one are cave galleries formed by transit waters from exterior of the massif. The second one, autochtonous, is formed within the isolated basin as a central conduit. There are several caves representing such type of a conduit in the Pan de Guajaibon. Such are the caves Lechuza, Mamey and Canilla with galleries several kilometres long. They are located at the foot of the northern wall of massif, waters from them are discharged through the karst spring of Canilla cave. The central conduit discharges the autochtonous subterranean basin formed under the depression between the Pan de Guajaibon and the Sierra Chiquita. The outflow part of this conduit is known as Ancon cave (Fig. 7). Geological drillings, made here to investigate bauxite deposits, prove existence of deep karst aquifer of the phreatic zone. However the springs from that zone have not been found in the vicinity of the massif. Probably the water is drained from the zone through surrounding areas hydrologically connected with the formation of the Pan de Guajaibon. There are systems of karst springs in the depression separating the massif Pan de Guajaibon from the Sierra Azul which may originate from the vadose zone. However some waters from the phreatic zone may be drained by Ancon spring. It may be proved by the run of outflow from this spring in the periods prior to high water level periods and observed then, typical changes in chemical composition of waters (Fig. 8).

There are thermo-mineral springs in the vicinity of the massif (within the San Marcos valley, near Mil Cumbres). Similar waters, under a high pressure, have been found during drillings in the hilly area along the road from San Juan de Sagua to La Palma. Appearance of that type of waters proves the existence of a deep water zone, probably of karst origin, which is connected with immense tectonic lines.

ANCON CAVE, AUTOCHTONOUS SUBTERRANEAN BASIN

Topographic surface of the Ancon cave basin, which includes vast depression between the ridges of the Pan de Guajaibon and the Sierra Chiquita, is equal 4.4. km^2. The peak Pan de Guajaibon (692 m a.s.l.) is the highest point in the basin, the karst spring of the cave Ancon (165 m a.s.l.) and the spring (130 m a.s.l.) at the lower part of the stream at the confluence with the river San Marcos are the lowest ones. Hence the difference in relative altitudes and vertical hydraulic gradient is above 0.5 km.

Water from the subterranean basin flows out through a siphon type outlet from the Ancon cave. The cave[5], explored at the distance of 0.8 km, forms an outlet part of a large central conduit (Fig. 7). The cave consists of a water conduit which, at the outlet part becomes a multi-level gallery ending with an oval chamber. The chamber is closed by a siphon through which the water flows towards the spring. High levels of flood waters can be observed in the chamber.

The water outlet from the Ancon spring is located in a gully cutting the crock wall falling

[5] The cave was discovered by the members of the Polish-Cuban expedition in 1984 (Pulina, Fagundo et al. 1984).

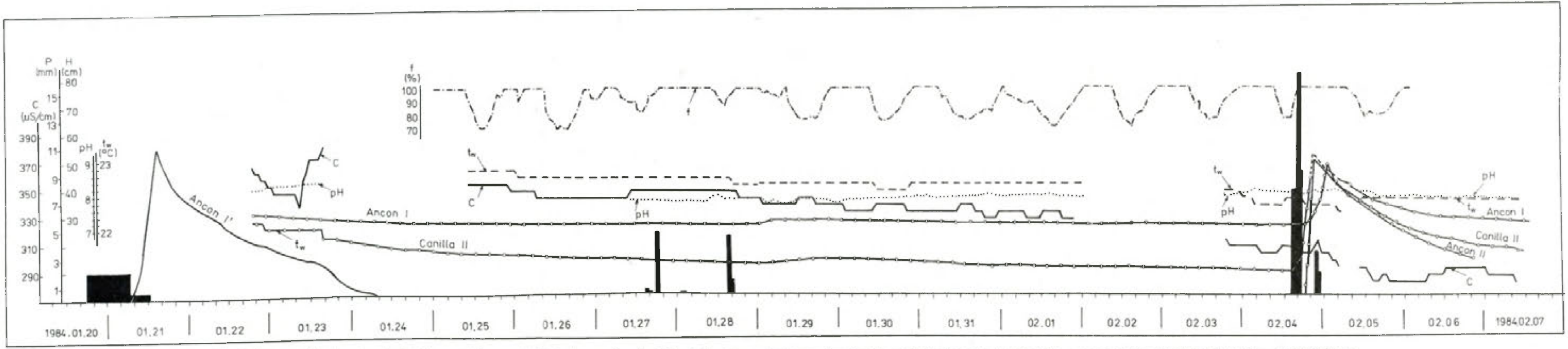


Fig. 8. Changes of hydrometeorological elements (P – precipitation in mm, f – humidity in %, t_w – water temperature in °C) and hydrochemical elements (pH – concentration of hydrogen ions, C – conductivity at 25°C – S/cm) in karst streams and karst springs of the caves Ancon and Canilla in January and February 1984 (Pan de Guajaibon massif – W. Cuba).
Location of water gauges: I, I', II in Fig. 5. Records of the hydrochemical station: 22–23 01 1984 at the water gauge Ancon I'; 25 01 – 1 02 1984 at the water gauge Ancon I; 3–7 02 1984 at the water gauge Canilla II

down from the depression between the ridges. A stream with travertine cascades appears among large gravitational blocks. It flows towards a flat surface of an alluvial cone covering the final part of a large karst spring valley. The stream loses water in certain places where fine-grained deposits disappear. All the water emerges in a large karst spring before the confluence with the river San Marcos.

A permanent hydrometric point was installed at the spring Ancon (limnigraph working from January 1984 till the end of 1988). Periodical points with limnigraphs were installed at the stream Ancon, near the scientific station (Ancon I′) and at the lower spring (Ancon I″). Automatic hydrochemical station worked several times at the spring Ancon. In winter 1988 and 1989, systematic hydrochemical investigations were carried out in the main cave, in vertical caves above the central conduit of Ancon, as well as in the stream Ancon and in the lower spring. The analyses were repeated in particular places each month during four hydrological years. A meteorological station, included in the Cuban meteorological net (PRI 222), was installed at the mouth of the valley. The part of the waters below the spring has been directed to the water supply system of San Juan de Sagua. The water intake was developed in 1986 and a small water power plant was built.

SUBTERRANEAN BASIN OF THE CANILLA CAVE

It is a subterranean karts basin charged mainly by waters from the northern, nonkarstic foreland of the Pan de Guajaibon massif. Part of the waters comes also from the northern slopes of the massif. The topographic surface of the basin is equal 6.4 km^2, half of it covers a nonkarstic area. The peak Pan de Guajaibon is the highest point of the basin — 692 m a.s.l. and the resurgence Canilla is the lowest point — 90 m a.s.l. A large part of the underground system has been explored. It is a system of galleries with transit streams flowing in the conduits located along the foot of the northern slope of the massif. A few kilometres of the system have been explored. It is developed on several levels with an acitve low level, flooded during high water level periods. The waters flow out through a siphon of the cave Canilla and they transport a great amount of suspension derived from fluvial deposits from the nonkarstic foreland of the massif. A hydrometric point was installed at the karst Canilla spring (working from January 1984 till December 1988). Samples for chemical composition of waters are taken at about twenty points. Systematic examinations of chemical contents of the waters were carried out in the spring and in the Canilla cave as well as in other caves and nonkarstic water courses in January and February 1984 and 1985. Systematic analyses were carried out each month for four hydrological years at several water points.

CHEMICAL DENUDATION IN TROPICAL KARST OF WESTERN CUBA

The amount of carbonate rocks dissolved and transported outside the karst massif is determined by a real chemical denudation. This quantity is presented both in volumetric values $m^3/km^2/y$ or in the equivalent linear value as a hypothetic layer scraped from the surface in mm/1000 years.

The quantity of the mass removed in a given time unit: t, depends on the following physical values:

$$D = f(Q, T)$$

where: D = chemical denudation in $m^3/km^2/y$ or in mm/1000 years,
Q = quantity of water discharged from a given area in m^3/s,
T = quantity of salt dissolved in waters discharged from the basin as a mean value in a given time period.

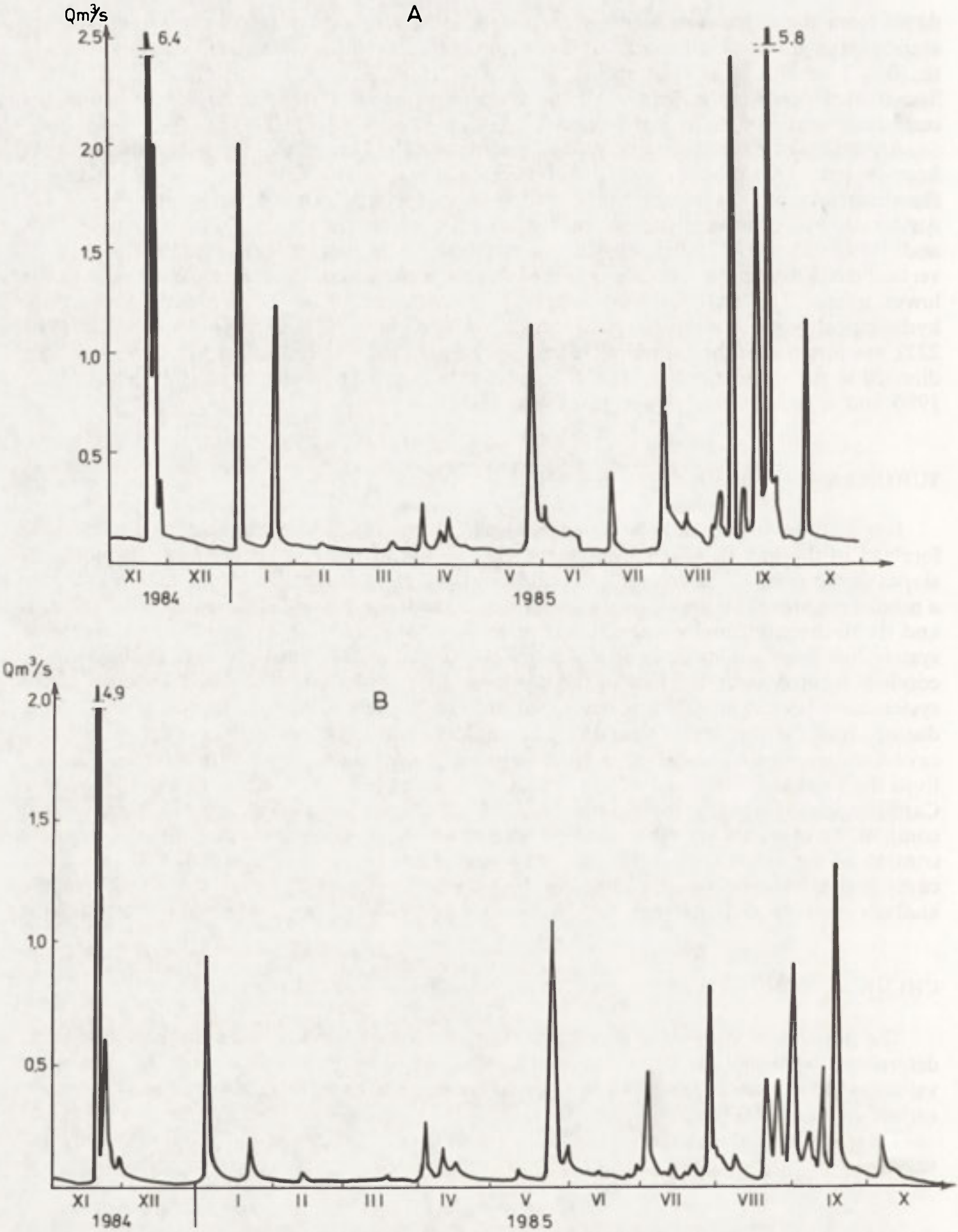


Fig. 9. Changes of water outflow in the hydrological year 1984/1985 in the subterranean basin of the Pan de Guajaibon massif (W Cuba) prepared by J. Rodriguez

A — karst spring of the Ancon cave (water gauge I), B — karst spring of the Canilla cave (water gauge II)

Location of water gauges in Fig. 5

In our case calculations have been based on the contents of $CaCO_3$ determined by calculation of the contents of cations Ca^{2+} and Mg^{2+} presented in mg/l. This dependence for an annual period may be presented as the following formula (Pulina 1974):

$$D = 12.6 \frac{\Delta T \times Q}{S} \quad \text{or } D = 0.0126 \times \Delta T \times q \quad \text{as } q = \frac{1000}{S}$$

where: D = $m^3/km^2/Y$ or mm/1000 years,
$\Delta T = T - T_a$ (relationship between the real contents of dissolved salts – T and allochtonous salts coming from outside of the basin – T_a, e.g. salts from precipitation, anthropogenic ones etc.),
S = surface of the basin in km^2.

Quantity of denudation can be also presented in weight unit t/Y, in this case the obtained values should be multiplied by mean specific gravity of carbonate rocks – 2.5 g/cm^3. It is necessary to carry on special investigations in experimental basins to obtain real quantities of Q and T in a given time period and calculation of ΔT Such investigations were carried out both in the Sierra de los Organos and in the Sierra del Rosario. Thoroughly, hydrologically and geologically investigated subterranean basins were chosen in both areas. Three ones in the basin Cuaguateje and two in the Pan de Guajaibon (Figs. 1, 2a, 3, and 8). The Sierra de los Organos represents a large, well developed region of tropical karst with predominance of transit waters. The Pan de Guajaibon is a small, isolated karst massif denudated mainly by autochtonous waters. The paper contains only a chosen part of the hydrological and hydrochemical material from both areas, to present several characteristic processes taking place in tropical karst.

Figures 8, 9A and B present hydrological diagrams based on records from karst springs in the basins Ancon and Canilla in the Pan de Guajaibon obtained in 1984 and 1985. The first diagram presents run of water levels and precipitation and air humidity in relation to the chosen physico-chemical elements (pH and C – conductivity). The diagram covers the end of January and the beginning of Febrauary 1984, when two great floods occurred, divided by a period of low water level. Special attention should be paid to the springs response to high precipitation level. Increased outflow appears almost immediately (in 2–4 hours), but changes of chemical contents of waters occur much later after several or even twenty hours. Hence, it may be assumed that in the initial phase of high pressure, caused by the inflow of large amount of water into the phreatic zone, old waters from the upper part of the zone are forced out and then the waters from contemporary precipitation flow out. That phemonena is multiplied during heavy rains occurring during wet seasons. Figures 9A and 9B show a typical diagram of water discharge from a subterranean basin in an isolated karst massif. Minor rains occurred only in the initial five months of 1984 (dry season). In other months, especially in the second half of August, and in September but mainly in November, cyclonic rains occurred. The outflow from the Ancon cave was equal 6.4 m^3/s (!) in November, when at lower levels it is usually equal 53 l/s. In that time 4.9 m^3/s was flowing out from the cave Canilla compared with 15 l/s at lower levels.

Mineralization quantity and chemical contents of waters vary greatly in western Cuba. Precipitation waters and those originating in nonkarstic areas are much different. Waters in karst massifs vary only in mineralization quantity and chemical aggressiveness. It is quite easy to determine, by means of Stiff's diagrams, particular types of waters in Fig. 10; from nonkarstic waters with significantly different chemical composition (diagram 1 and 3), through typical waters form the interior of a karst massif (diagram 4 and 6), waters with relatively high mineralization from a phreatic zone (diagram 16) up to karst waters influenced by marine intrusions (diagram 17 and 18). Precipitation waters infiltrating into a karst massif become fully saturated already several metres below the morphological surface. They deposit large amount of calcite. Changes of their aggressiveness may occur when reaching conduits and fissures with high concentration of CO_2. Contents of CO_2 varies greatly in tropical karst as

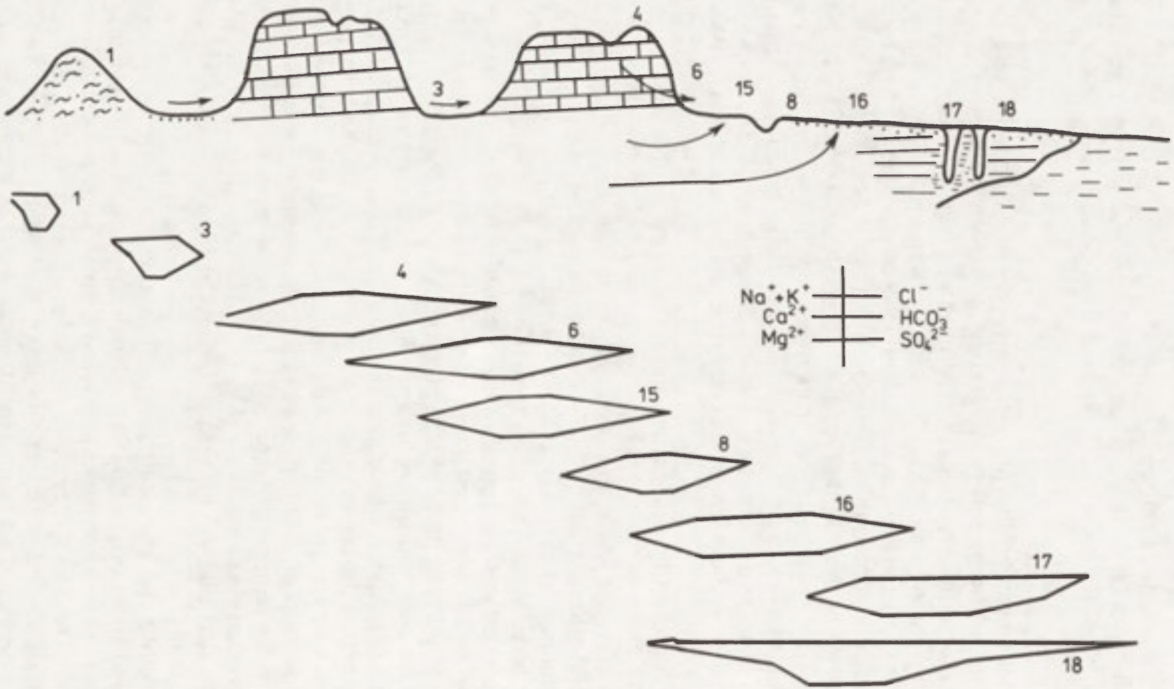


Fig. 10. Changes of chemical composition of surface and subterranean waters (Stiff's diagrams), along a schematic longitudinal profile in the Cuyaguateje 1–2 waters from non-karstic areas, 3 — waters from the river Cuyaguateje and in Pica Pica polje, 4 and 6 — waters in the mogotes Sierra de San Carlos, 8 — waters from the river Cuyaguateje at the lower part of the basin, 15, 16, 17 — waters from karst springs and wells reaching the phreatic zone, 18 — subterranean karst waters influenced by sea water intrusions

TABLE 2. Quantities of chemical denudation in karst regions located in various climatic zones (Pulina 1992) $m^3/km^2/y$ or mm/1000 years

Climatic zones	Lowlands and uplands	Mountains	High mountains
Tropical	Sierra de los Organos, Cuba 53–88	Pan de Guajaibon, Cuba 59 (92)	
Mediterranean	Upland Karst, Slovenia 80–136	Vercors French pre-Alps 44–98 Venetian pre-Alps Italy 23–70	Central Apenines Italy 53–130 Julian Alps Slovenia 51–67 SW Caucassus Georgia 114–139
Temperate warm	Dobrogea, Bulgaria 23	Vracanska Planina 38	Pirin, Bulgaria 47
Temperate, transitional	Cracow-Silesian Upland, Poland 24–26	Sudetes Mts. Poland 20–33	Tatra Mts. Poland 49
Temperate cold	Irkutsk Upland E Siberia 1–6	Khamar Daban E Siberia 22	Tunkinsk Alps E Siberia 13
Subpolar	Hornsund region Unglaciated area Spitsbergen 11	Hornsund region Glaciated area Spitsbergen 32	

there are large amounts of carbon dioxide of organic origin there. Therefore, saturated and oversaturated waters with relatively high mineralization predominate in karst massifs. Waters with low mineralization, chemically active and with various ionic composition occur in karst massifs with transit water flows. Those waters, especially during heavy rains, form mechanically large galleries.

Table 1 presents data necessary to calculate chemical denudation as well as calculated quantities of denudation. Particular digital data come from thousands of data collected systematically in few consecutive hydrological years. Outflow quantity has been calculated from everyday measurements of water levels (water level indicators in the hydrographic net) or registration (limnigraph). The chemical composition was determined in selected points each hydrological season and in particular areas once a month. Special hydrochemical investigations were carried out in the Pan de Guajaibon. The determination of the subterranean basin area was the most difficult problem. It was easier to solve it in the Sierra de los Organos, while mean values for two basins had to be used in the Pan de Guajaibon. The result for the autochtonous basin Ancon, given in Table 1, may be a bit too high (92).

In the period 1979/1991, chemical denudation in the basin Cuaguateje at the final profile Portales II was equal 87.7 $m^3/km^2/y$ (from 54.4 in 1981 – the lowest outflow 8.9 m^3/s to 136.1 in 1979 with the highest outflow 22.7 m^3/s). The hydrometric profile Portales II, which closes the basin covering 502 km^2 collects transit waters as well as karst waters from the upper part of the phreatic zone where the retention period is smaller than one hydrological year. Differences in chemical denudation between particular partial basins are not caused by various amounts of water but by different degrees of their mineralization. Transit waters from nonkarstic areas predominate in higher located basins.

The chemical denudation in the isolated Pan de Guajaibon massif in the hydrological year 1984/1985 was equal 59 $m^3/km^2/y$. This value is probably underrated due to the relatively dry year, when compared with the longer period, and the fact that it was charged by low mineralized allochtonous waters from outside of the karst area (Canilla basin).

After comparing the values of chemical denudation in two different types of tropical regions the following conclusions can be formed:

1. Large subterranean outflow of waters, when compared with precipitation, e.g. autochtonous basin Canilla about 48 $l/s/km^2$ (1500 mm), precipitation 2000 mm (evaporation equivalent about 500 mm) is significant.
2. Autochtonous basins, which are quickly discharged by a subterranean system of a well developed central conduit, contain waters of relatively low mineralization. Mean value of $CaCO_3$, for the Ancon basin, is a little more than 160 mg/l and full mineralization Σ_M = 206 mg/l. While in a large basin in a well developed karst charged by waters from a phreatic zone mineralization reaches 250 mg/l $CaCO_3$ and Σ_M is even equal 300 mg/l.
3. Relatively low chemical denudation is observed in mountain autochtonous basins and relatively high in basins developed in lowland and upland tropical karst.

Table 2 presents values of karst chemical denudation in about twenty basins located in different climatic zones. Chemical denudation in western Cuba is rather high, comparable with areas of mountain karst in the Mediterranean Region. However the values are lower than in the Mediterranean high mountains regions.

FINAL REMARKS

The article presents geomorphological, hydrological and hydrochemical characteristics of two different regions of tropical karst. The basin Cuyaguateje in the Sierra de los Organos represents a well developed upland karst and Pan de Guajaibon an isolated mountainous karst massif. Differences between these two areas are seen not only in the relief but mainly in circulation of subterranean waters. When comparing these two regions with morphologically similar areas located in the zones of Mediterranean and moderate climate in Europe, e.g.

Sierra de los Organos versus Karst Upland, Dobrogea, Cracow-Silesia Upland and Pan de Guajaibon versus Vercors massif, Vracanska Planina or Venetian pre-Alps, we can find many similar karst forms, but also many other original forms, especially in macromorphology of karst massifs (cone and tower karst relief). Special attention should be drawn to well developed interiors of mogotes, cut by numerous vertical shafts and *hojos* type of dolines, with old galleries with many large chambers. The morphological surface is filled with various karst microforms. Hydrogeological zones are very well developed in karst massifs.

The chemical denudation is active mainly upon the morphological surface and the shallow zone under the surface. Inside the massif, within the vadose zone, waters are saturated. They are corrosive only locally. Hence, the morphological surface is strongly perforated and numerous dripstones occur in the vadose zone. The mechanical activity of water is a characteristic process for the transitional zone (transit caves, central conduits). A large role in development of subterranean karst conduits is also played by the phreatic zone, especially along fault zones.

The quantity of chemical denudation was investigated in selected basins, thoroughly known geologically and hydrogeologically. They represent two basic types of subtropical karst. It was observed that chemical denudation varies greatly, e.g. the values for Cuyaguateje range from 27 to 130 $m^3/km^2/y$ in the period 1979–1981. They result from the differences in the quantity of annual precipitation between dry and humid years (1 : 3) and in the degree of mineralization.

In spite of the high evaporation, a relatively high amount of water is remarkable in the karst massifs within that climatic zone. It is caused by a rapid drainage of precipitation waters form the surface inside the massif as well as large share of waters from condensation, mainly in the caves of the vadose zone. Therefore mean perennial chemical denudation reaches the value of 90 $m^3/km^2/y$. This value is much higher for the morphological surface and the zone under the surface, as part of the denudated rock mass is deposited in the vadose zone as calcite dripstones. This phenomenon is less intensive in climatic zones at higher latitudes. It does not occur in high mountains and in a subpolar climate, where the whole amount of the dissolved mass is transported outside the karst region. Hence, the values presented here show the quantity of rock mass transported outside the massif, i.e. regional denudation.

The results of measurements of the chemical denudation in western Cuba solve the problem initiated by J. Corbel (1959) concerning real chemical denudation in karst regions at different latitudes. They also support the importance of statistic models, prepared for potential denudation by one of the authors (Pulina 1974), which could predict similar quantities of chemical denudation in the tropical climate zone.

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