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**23**

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**COMMISSION ON PRESENT-DAY GEOMORPHOLOGICAL PROCESSES  
Montreal-Vancouver 1972**

**Edited by  
Alfred Jahn and Leszek Starkel**

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## PREFACE

*The Commission on Present-Day Geomorphological Processes was set up in New Delhi, at the 21st Congress of the International Geographical Union. The Commission's objective is to examine, by means of quantitative methods, geomorphological processes according to morphological-climatic zones of the Globe.*

*In the early stages of its activities the Commission has been aiming at collecting the existing data on geomorphological processes acting in various areas, and different countries.*

*In the next stage Commission's task will consist in simultaneous investigation of some processes at selected stations located in areas varying in relief, geological structure and climate. The same methods will be applied in this research.*

*The present volume contains early papers reporting on geomorphological processes of the cold, temperate, and tropical zones.*

*It is to be hoped that in the following years more abundant systematically selected materials will be forthcoming to realise the Commission research programme.*

Alfred Jahn, President  
Commission on Present-Day  
Geomorphological Processes

## PRESENT-DAY CRYOGENIC PROCESSES IN THE MOUNTAINS OF EASTERN SIBERIA

TADEÁŠ CZUDEK AND JAROMÍR DEMEK

### INTRODUCTION

The study of present-day cryogenic processes is of considerable significance for the knowledge of the relief development of the Pleistocene periglacial zone. In 1966 and 1969 we studied the cryogenic processes in the Stanovoy Khrebet and the Aldanskoye Nagorye in the southern part of Yakutia, in the Verkhoyanskiy Khrebet, the Khrebet Kular, the Selennyakhskiy Khrebet and the Khrebet Cherskogo in eastern and north-eastern Yakutia and in the Kolymskiy Khrebet (Fig. 1) in the Magadan Region. The territory



Fig. 1. Relief of the Kolymskiy Khrebet near the Pass Dedushkina lysina. Photo  
J. Demek, June 20, 1969

investigated is heterogenous from the point of view of structure, relief, climate, soils and vegetation. Several features are typical of these areas, such as:

(a) considerable and relatively rapid temperature oscillations due to the clear atmosphere in the region of the Siberian anticyclone. For instance in the town of Verkhoyansk, the air temperature ranges between  $-54$  and  $+4^{\circ}\text{C}$  in May and between  $-45$  and  $+13^{\circ}\text{C}$  in October. Even in summer, in the month of August, the temperatures fluctuate between  $-8$  and  $+28^{\circ}\text{C}$ ,

(b) mean annual temperatures below  $0^{\circ}\text{C}$  and extremely low winter temperatures. The absolute minimum temperature is  $-67.8^{\circ}\text{C}$  at Verkhoyansk on the Yana River valley and  $-77.8^{\circ}\text{C}$  (1938) at Oymyakon on the Indigirka River valley,

(c) short transition seasons (spring and autumn) with temperatures ranging largely about  $0^{\circ}\text{C}$ , and the occurrence of numerous freeze-thaw cycles,

(d) considerable dryness and exceedingly small quantities of snow (0.3 up to 0.5 m),

(e) temperature inversions in the intermontane basins shown as examples in Table 1. In connection with the temperature inversion it was established that the active layer on the slopes of southern exposure reaches its maximum thickness at the level of the upper timber line and not in the basin bottoms,

TABLE 1. Temperature inversion in intermontane basin

Mean Temperature	Suntar Khayata (2063 m)	Baze (1350 m)	Agayakan (777 m)
January	$-29.5^{\circ}\text{C}$	$-34.7^{\circ}\text{C}$	$-48.0^{\circ}\text{C}$
July	$6.6^{\circ}\text{C}$	$11.7^{\circ}\text{C}$	$14.6^{\circ}\text{C}$
year	$-14.1^{\circ}\text{C}$	$-14.1^{\circ}\text{C}$	$-16.1^{\circ}\text{C}$
absolute maximum	$18.6^{\circ}\text{C}$ (July)	$24.8^{\circ}\text{C}$ (August)	$29.7^{\circ}\text{C}$ (August)
absolute minimum	$-43.0^{\circ}\text{C}$ (December)	$-50.8^{\circ}\text{C}$ (December)	$-58.2^{\circ}\text{C}$ (February)

(f) presence of permafrost; below large river beds (Kolyma, Indigirka, Lena, Yana) open taliks occur. In the permafrost distribution vertical zoning can be observed: with the increase of altitude above sea level the permafrost thickness increases. The high mountain regions of the Verkhoyanskiy Khrebet and the Khrebet Cherskogo are thus regions with a very low permafrost temperature. P.N. Lugovoy (1970, p. 107) states that the permafrost temperature in altitudes from 1000 up to 1600 m in depths from 15 up to 20 m ranges between  $-4.5$  to  $-9^{\circ}\text{C}$ , and in altitudes from 3000 up to 3200 m a. s. l. between  $-15$  to  $-16^{\circ}\text{C}$ .

## PRINCIPAL CRYOGENIC PROCESSES IN THE TERRITORY INVESTIGATED

### FROST CRACKING

Due to considerable temperature changes, volume reduction in rocks takes place and thermal contraction cracks develop. Fresh evidence of frost cracking was observed on granite tors and blocks at altitudes from 1800 up to 2000 m on the Nel'kanskiy Khrebet near the town of Ust'-Nera. Frost cracking also took place along bent planes (microexfoliation).



Thermal contraction cracks develop every year in soils overlying ice wedges. We located them over the syngenetic ice wedges in the Irgichan River flood plain near the town of Deputatskiy. During observations on June 30, 1969, the cracks were open and reached as far as the ice wedges.

In the bottoms of basins and in flood plains the thermal contraction cracks form macropolygonal patterns. The macropolygons can be recognized mainly from an airplane. They are distinctly developed for instance on the bottom of the Omyyakon Basin where they reach diameters of about 40 m.

FROST WEDGING

This process takes place due to the expansion of water on cooling to form ice. In cracks, ice veins and ice wedges develop splitting the rock into fragments. The size of the fragments depends on the jointing of the rock and the intensity of frost weathering. In the granites of the Nel'kanskiy Khrebet blocks of several cubic metres volume develop. In the sandstones of the Verkhoyskiy Khrebet fragments of several tens of centimetres in their longer axis prevail. In the andesites of the Kolymskiy Khrebet on the Pass Dedushkina lysina in a large part of the frost desert we found fragments 3 to 4 cm large. During weathering in the Kolymskiy Khrebet the effect of increased humidity is obvious probably due to the proximity of the Pacific Ocean.

On slopes in massive rocks with an inclination of more than 15° a special type of slope development due to frost-wedging was observed. Usually up to 3 or 4 cracks parallel with the direction of slope can be distinguished. They occur in distances from 1 to 10 m. In vertical fissures ice wedges develop

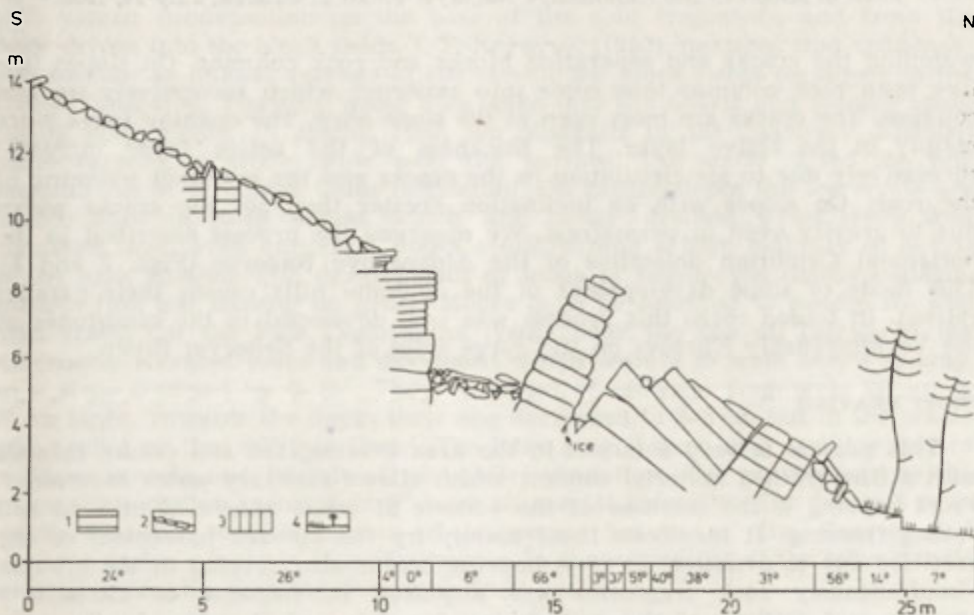


Fig. 2. Profile through the slope of the dolomite hill at the village of Yatsik near the town of Aldan in the Aldanskoye Nagorye  
 1 — Cambrian dolomites, 2 — dolomitic debris, 3 — ice wedge, 4 — vegetation (grass and *Larix dahurica*)



Fig. 3. Slope development on Cambrian dolomites at the village of Yatsik near the town of Aldan in the Aldanskoye Nagorye. Photo T. Czudek, July 24, 1969

widening the cracks and separating blocks and rock columns. On slopes fan-like open rock columns thus come into existence which successively tip and collapse. The cracks are most open at the slope edge. The opening takes place mainly in the active layer. The thickness of the active layer increases successively due to air circulation in the cracks and the resultant warming of the rock. On slopes with an inclination greater than  $50^\circ$  the cracks widen due to gravity even in permafrost. We observed the process described in the horizontal Cambrian dolomites of the Aldanskoye Nagorye (Figs. 2 and 3). This mode of slope development of the dolomite hills causes their parallel retreat. In folded rocks this process was well developed in the sandstones of the Verkhoyanskiy Khrebet on the gorge sides of the Segorym River.

#### FROST HEAVING

This process is very common in the area investigated and occurs in soils with a fine-grained material content which allows capillary water movement. Frost heaving is the increase of the volume of water saturated mineral soil during freezing. It manifests itself mainly by the upward movement of the particles due to pressure. In frost heaving cracks also develop in the ground. Simultaneously rock fragments are displaced in response to alternating heaving and settling of the soil during freezing and thawing. In the area investigated this process called cryogenic desorption is effective and displaces a relatively high quantity of material. There are no quantitative data available from the area investigated concerning the displaced material. But V. A. Voyloshnikov (1968, p. 115) mentions that in the neighbourhood of Irkutsk on

north-facing slopes a soil layer 10 to 15 cm thick is displaced by this movement at a rate of 0.35 up to 0.40 m per year. The rate of the movement decreases in a downward direction and ceases at a depth of 0.5 to 0.8 m below the ground surface. The cryogenic desorption is most effective on slopes of an inclination between 6 and 8°.

#### NEEDLE ICE

Needle ice is a kind of segregation ice developing due to the freezing of water saturated soils in places deprived of vegetation. In the area investigated it is of considerable geomorphological significance. L. A. Zhigarev's observations (1960) have shown that in the period of 36 summer days rock fragments were transferred on the slope 14–18 cm due to needle ice.

#### FROST CREEP

Frost creep is a slow downslope displacement of debris by gravity resulting from sliding on ice crusts (the debris-ice) which develop at the base of the fragments in debris mantles. The ice crystals grow upwards and the individual fragments are raised. The ice forms sliding planes on which the debris shifts in the direction of slope inclination. The debris mantles and block fields (Felsenmeere) with cavities filled up with air protect the underlying rocks from insolation making the penetration of cold air possible. Whereas the annual difference between the absolute minimum and the absolute maximum of air temperatures amounts to more than 100°C in the territory investigated it is usually only 30–35°C in the block fields (M. D. Budz, O. L. Rybak, 1968, p. 218). This is why permafrost occurs near the ground surface below block fields the whole year. The debris-ice develops through both vapour condensation on the base of the cold fragments, and from the snow driven into the block fields. I. T. Reynyuk (1959) mentions that condensation reaches an annual average of 100 mm in the block fields on slopes in the area investigated. Thawing water also leaks into the debris and freezes again during cold nights. In spring 1969, we observed in the Stanovoy Khrebet water running in block fields near the surface. The water lifted the fine material out of the block fields (suffosion). On gentle slopes the ice crusts are thicker than on steep slopes which are better drained. Frost creep thus manifests itself even on gentle slopes conditioning a high mobility of debris in the area studied. The development of lobes of non-sorted debris material without fine soil moving on gentle slopes is to be ascribed to needle ice and frost creep. On the Pass Dedushkina lysina near the spa of Talaya in the Kolymskiy Khrebet lobes had developed which were 2 m wide and 3 m long, on a slope inclined by 8–10°. They consisted of andesite fragments 20 up to 30 cm large. Towards the depth their size decreased to 2–3 cm but in the whole lobe profile no fine soil was found. The lobes moved even over isolated groups of *Pinus pumila* and islands of xerophile vegetation. The movements in the lower slope section inclined at 6° were of special intensity. The lobes, 2 m in length, consisted on their surface of fragments 3–4 cm in size. Coarse sand occurred under a layer of these fragments. In the movement of the lobes needle ice took evidently part together with frost creep.

#### SLOW SOLIFLUCTION

Some authors (e.g. S.S. Korzhuyev in: I.P. Gerasimov (ed.), 1965, p. 88) state that solifluction is fairly active in the mountain ranges of Eastern and Southern Yakutia. But our observations point to the relatively small im-

portance of solifluction. This is probably connected with the considerable dryness of this territory (low precipitation and high evaporation in the mountains). A number of forms, formerly ascribed to solifluction, are to be explained by other processes, especially by frost heaving. More intensive solifluction was established only on the cryopediments in the Stanovoy Khrebet and the Khrebet Cherskogo, and mainly in the northern mountain ranges of Khrebet Kular and Selennyakhskiy Khrebet (Fig. 4). The solifluction



Fig. 4. Slow solifluction on the slope near the town of Deputatskiy in the Selennyakhskiy Khrebet. Photo T. Czudek, June 29, 1969

movements are slow and can be observed on the bends of lower parts of the tree trunks. In the surroundings of the town of Deputatskiy solifluction lobes and terraces develop on the slopes of asymmetric valleys.

At the foot, taluses with a high ground ice content develop largely built of solifluction deposits. In the observation period (June 1969) solifluction was operating below snow patches.

We observed in the territory studied that solifluction takes always place in combination with other processes, mainly frost heaving, frost creep and sheet wash. The different processes mutually interpenetrate on the slopes mainly depending on the fine soil content and presence of water.

#### RAPID SOLIFLUCTION

Rapid solifluction develops in the case of oversaturation of the fine-grained deposits with water. We observed it in places disturbed by vegetation where the thawing of ground-ice took place. The softened soils formed lobes moving quickly downslope. In places, the rapid solifluction already changes into mud flows. For instance on the Khrebet Kular the width of the mud flows reached about 2 m, their thickness as much as 1 m and their length 10 up to 15 m.

## SHEET-WASH

We observed the effects of sheet-wash during the spring snow melt. Sheet-wash manifested itself on the bare surface mainly in non-sorted circles. In water saturated soils the effects of rapid solifluction could only hardly be distinguished from those of sheet-wash and mud flows. Both processes substituted each other in place and time.

## FROST SORTING

Frost heaving, freezing out of fragments and needle ice together lead to frost sorting. First, in debris fields islets of fine soil develop due to frost heaving. We observed numerous fine soils islets in the frost desert in the surroundings of the town of Deputatskiy. In the area investigated the most common are the non-sorted circles. They come into being in fine-grained material on watershed ridges and cryoplanation terraces. They are active mainly on calcareous soils. They were paid special attention on the flat top (1300 m a.s.l.) built of Cambrian dolomites in the Evota Group in the Aldan-skoye Nagorye. The circles are 0.3-2.0 m in diameter and mostly with circular groundplans. They consist of brown thixotropic loam with small dolomite fragments. Their surface is further fissured with small non-sorted polygons. Mainly frost heaving, and later on the bare surface even needle ice are supposed to share in their development.

The largest non-sorted circles were observed in the Gerba Pass south of the village of Orotukan at an altitude of about 900 m a.s.l. (Fig. 5). The circles were as large as 4 m in diameter and were built of greyish-brown sandy loam

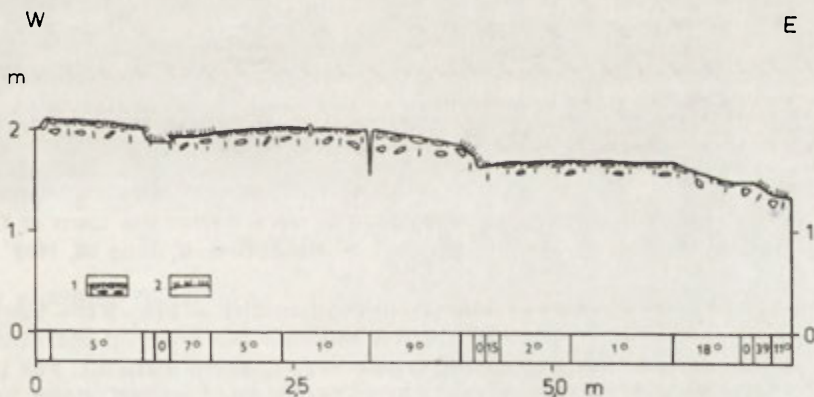


Fig. 5. Profile through non-sorted circles on the gentle slope in the Gerba Pass, south of the village of Orotukan

1—sandy loam with fragments of shales, 2—xerophile vegetation

with fragments of shales mostly of a diameter of 1-5 cm. The total thickness of the regolith in the pass was 1.1-1.5 m. Lower on the slope at an inclination of 10° the circles changed into non-sorted steps bordered with xerophile vegetation, their front reaching a height of 0.44 m. Over this vegetation rim, water saturated soil flowed and fragments fell.

In the further process sorted polygonal grounds develop from non-sorted circles. But sorted polygons are rather rare in the territory investigated. Well developed polygons of this kind were found in the surroundings of the town

of Deputatskiy (Fig. 6). Their diameters reached about 4 m. They occurred below the nivation depression in which there was snow on the day of observation (June 28, 1969). There is a terrace inclined at  $3^\circ$  with traces of solifluction below the nivation depression. On its outer margin stone polygons occur. The cores of the stone polygons consist of black-grey loam. Their rims are built of sandstone fragments, up to 0.3 m in size, in a vertical position. In the cores of the polygons, frost heaving and needle ice operate. At the town of Deputatskiy needle ice reaches heights of 4 to 10 cm and its columns are 0.25–0.5 mm



Fig. 6. Sorted stone polygon on the cryoplanation terrace near the town of Deputatskiy in the Selennyakhskiy Khrebet. Photo J. Demek, June 28, 1969

in diameter. According to our observations they lift a 1 to 2 cm thick layer of loam and small fragments and shift it towards the polygon margins.

An explanation of the sorting of large blocks is more difficult. For instance in the Aldanskoye Nagorye at the Aldan-Yatsik road sorted stone polygons 5 to 6 m in diameter overgrown with vegetation were studied on a slope inclined at  $7^\circ$ . The rim of the polygons consisted of rounded granite boulders of dimensions of about  $4 \times 2 \times 1$  m in a vertical position. The cores of the polygons consisted of brown sandy loam with granite fragments not larger than 0.8 m (to a depth of 0.8 m) Among the boulders of the rim water occurred. Frost heaving probably took part in the origin of these polygons.

On steeper slopes sorted stripes are developed. In 1966 they were studied in detail in the ridge of Zapadnye Yangi. Two types of frost sorted stripes can be distinguished; one with a concave cross profile and one with a convex cross profile. In stripes with a concave cross profile water was running. Sorted stripes are common even in the surroundings of the town of Deputatskiy on slopes with an inclination of more than  $10^\circ$ . The stripes of coarse debris here are more than 1 m wide.

## NIVATION

In mountain regions strong winds blow in winter reaching a velocity as great as 40 m/s. The wind blows the snow, and it accumulates on lee slopes and breaks of slope. Simultaneously, the strong wind consolidates the snow. In valleys snow melts in mid May but one month later in the mountains. During melting the exposure of the slopes distinctly comes into play. Snow patches (Fig. 7) last till summer. They are very important water resources because of



Fig. 7. Snow patch at the foot of the frost-riven scarp on the slope of the hill 1300 m a.s.l. south of the Evota Pass in the Aldanskoye Nagorye. Photo T. Czudek, May 28, 1969

the dry climate. Whereas the intensity of the cryogenic phenomena considerably decreases due to lack of water within a month after the snow has melted, below the snow patches cryogenic processes are very intensive until the complete melting of the snow patches. Nivation is one of the causes of the parallel slope retreat mainly in the case of frost-riven cliffs and frost-riven scarps of the cryoplanation terraces (Fig. 8).

## THERMOKARST

The development of thermokarst processes is due to the disruption of the thermal equilibrium of the permafrost and the increase of the thickness of the active layer owing to local or general (climatic) causes. In the mountains of the area investigated these processes are limited especially to thermoerosion and the melting of ground ice mainly in taluses and glacial deposits. Due to the thawing of the polygonal system of ice wedges, conical baydjarakhs developed in the moraines of the K'ub'ume River valley in the Verkhoyanskiy Khrebet.

In the territory investigated we observed both linear and the lateral thermoerosion. On the left Deputatka River valley side large polygons of ice wedges have developed. Thawing of ice wedges takes place on slope with an inclination of  $3^\circ$ . Due to linear thermoerosion a furrow, 0.6 m in depth, developed after the ice wedges. The furrow is continuously buried by sliding turf and soils. Running water removes the saturated soils. In places, the furrow is completely buried by turf. Water disappears below the turf appearing again at a distance of several metres. The length of the longest buried section was 6 m. Owing to the downslope displacement of the soil towards the thermoerosion furrow a flat prolonged depression — the dell — develops successively. The dell is about 6 m wide and 1 to 1.5 m deep.

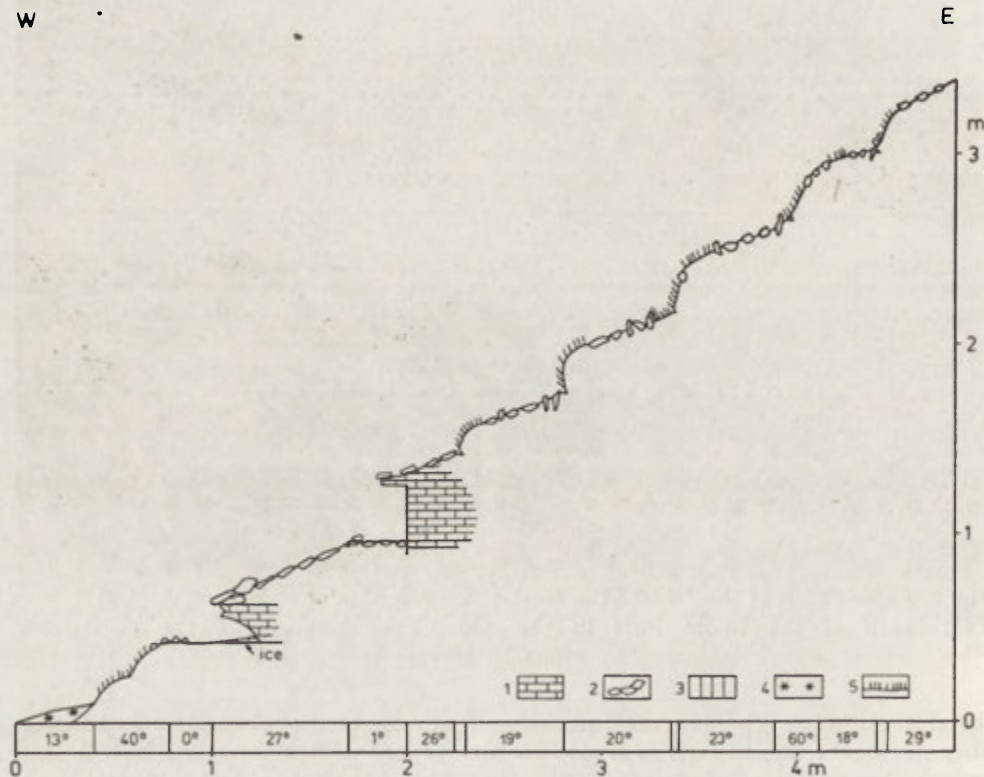


Fig. 8. Profile through the frost-riven scarp of the cryoplanation terrace on the slope of a hill 1300 m a.s.l. south of the Evota Pass in the Aldanskoye Nagorye with processes of frost creep and nivation

1 — Cambrian dolomites, 2 — dolomitic debris, 3 — ice, 4 — snow, 5 — grass

Linear thermoerosion on ice wedges was also observed on the K'ub'ume River flood plain near the village of Kamenistaya in the Verkhoyanskiy Khrebet (Fig. 9 and 10). During investigations on June 10, 1969, water was running first on the turf, and then it disappeared in a thermal contraction crack. After some metres the frost crack began to widen into a rill due to thermoerosion. The rill successively developed in a gully about 1 m deep. Its sides kept sliding and turf and blocks of frozen fine grey sand filled the gully.

Lateral thermoerosion could be observed on the banks of the Irgichan River near the town of Deputatskiy. The outer bank was 3 m high and built



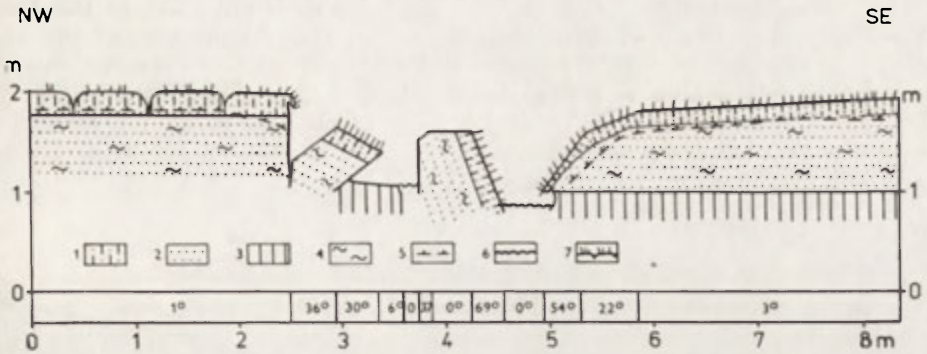


Fig. 9. Linear thermoerosion on the K'ub'ume River flood plain near the village of Kamenistaya in the Verkhoyanskiy Khrebet  
 1 — black-brown humus sandy loam, 2 — fine grey horizontally bedded sand, 3 — ice wedges, 4 — segregation ice, 5 — permafrost limit, 6 — water level, 7 — grass



Fig. 10. Linear thermoerosion near the village of Kamenistaya in the Verkhoyanskiy Khrebet. Photo J. Demek, June 10, 1969

of 2 m of greyish-brown sandy loams and of underlying gravels rising about 1 m above the water level. In loams syngenetic ice wedges have developed their ends reaching as far as in the gravels. The thickness of the active layer was only 0.4 m on the day of observation (June, 30, 1969). A thermoerosional niche developed in the bank due to lateral thermoerosion. The niche was 2 to 5 m deep and 1.5 m high. Turf bent from above and partly overlapped the niche.

Intensive lateral thermoerosion took place on the right bank of the Deputatka River near the town of Deputatskiy (Fig. 11)). At the foot of the right valley side a talus has developed consisting of black humus loams with numerous organic relics. Due to lateral thermoerosion of the Deputatka River the talus was undermined and a vertical wall developed on which ice wedges (as much as 5 m wide and 8 m high) crop out. Due to sapping blocks of ice and

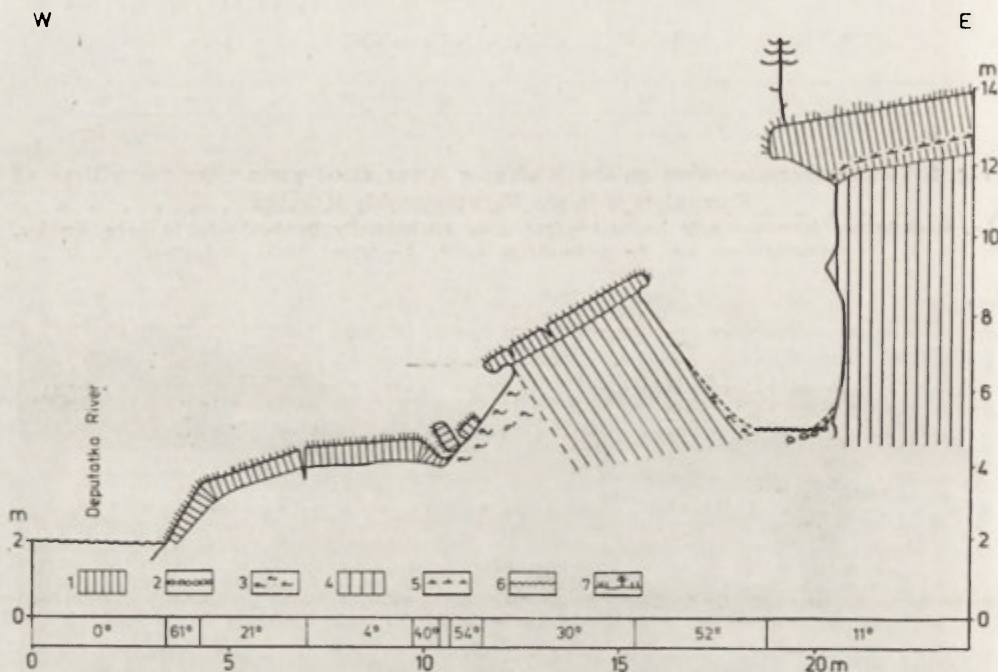


Fig. 11. Lateral thermoerosion on the right bank of Deputatka River near the town of Deputatskiy in the Selennyakhskiy Khrebet

1 — black humus loam with numerous organic relics, 2 — fluvial gravels, 3 — water saturated deposits, 4 — ice wedges, 5 — permafrost limit, 6 — water, 7 — vegetation (*Larix dahurica*, grass)

frozen soil of considerable dimensions separate and collapse into the river bed where they melt and their material is removed. The collapse of blocks and the damming of the river are typical features of the development of water courses due to thermoerosion.

#### SEASONAL INJECTION ICE

By injection of water into frozen soils during the freezing of the active layer, ice lenses of a thickness of 2 to 3 m and diameters of several metres up to several tens of metres develop near the ground surface. The ice contains numerous air bubbles and displays a prismatic texture. On the surface the ice forms flat dome-like elevations called ice-cored mounds. The ice melts in summer. We studied these processes in the Khrebet Cherskogo. The largest seasonal ice-cored mound was developed in the Nera River valley, near the town of Ust'-Nera. It had a diameter of 45 m and its centre was already sunken on the day of observation (June 12, 1969). The mound was covered with peat,

of the thickness of 0.3 to 0.8 m. The ice had a distinct stalk-like texture. There were smaller ice-cored mounds in its surroundings (Figs. 12 and 13).

Seasonal ice-cored mounds develop during the freezing of the active layer. The thickness of the active layer must be sufficiently great so that there is a sufficient quantity of water for injection. These processes can therefore take place mainly at the southern permafrost limit and in extremely continental climate as for example in the area studied.

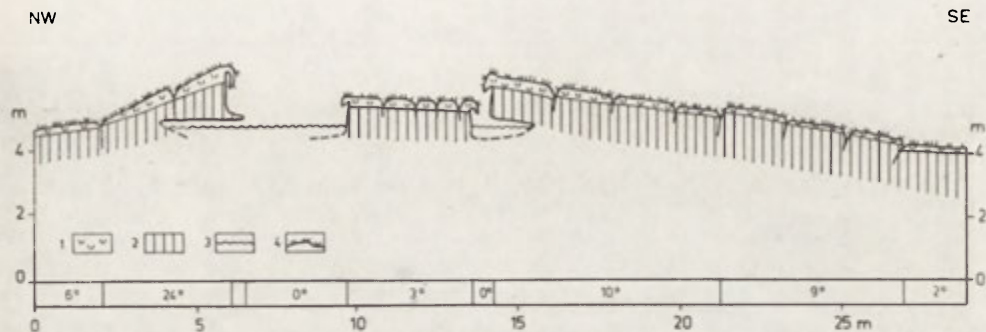


Fig. 12. Seasonal ice-cored mound in the Nera River valley near the town of Ust'-Nera

1 — peat, 2 — ice, 3 — water level, 4 — vegetation (grass and moss)



Fig. 13. Seasonal ice-cored mound in the Nera River valley near the town of Ust'-Nera. Photo J. Demek, June 12, 1969

## ICINGS

The icing is an ice mass developed due to the freezing of the surface or ground water penetrating into the ground or ice surface or into large cavities in rocks (for instance caves). The icings are up to several tens of kilometres long and about 4 m thick in the territory investigated. Ice mounds (as large as 5 m high) can often be found on icings. We observed them mainly on the icings in the Khrebet Yankan and in the Arangas River valley (Fig. 14) near the



Fig. 14. Icing in the Arangas River valley in the Khrebet Cherskogo. Photo J. Demek, June 14, 1969

placer Alaskitoviy in the Khrebet Cherskogo. In the Khrebet Yankan an ice mound was developed on the surface of the flood plain icing filling the Yeltulak River bed. The ice mound was 30 m long, 15 m wide and 4.3 m high. During observation (May 24, 1969) four water filled crevasses developed on its surface. An explanation of the origin of ice mounds could be the freezing of an accumulated pocket of water within the icing.

The icings in deep valleys are of greatest geomorphological significance. In the Vostochnaya Khandyga River valley in the Verkhoyanskiy Khrebet the widening of the river bed was observed in the places where a flood plain icing develops every year. The river is pushed aside by the ice and lateral erosion takes place. During spring thawing, water streams develop on the icing wandering over the ice surface. The position of the channels changes so that the water streams erode other bank sections every year. In places where the icing reaches as far as the valley sides it pushes and undermines them (Fig. 15). This process was also observed on the slopes of the gorge of the Arangas River near the placer Alaskitoviy. The steep rock walls were undermined on the

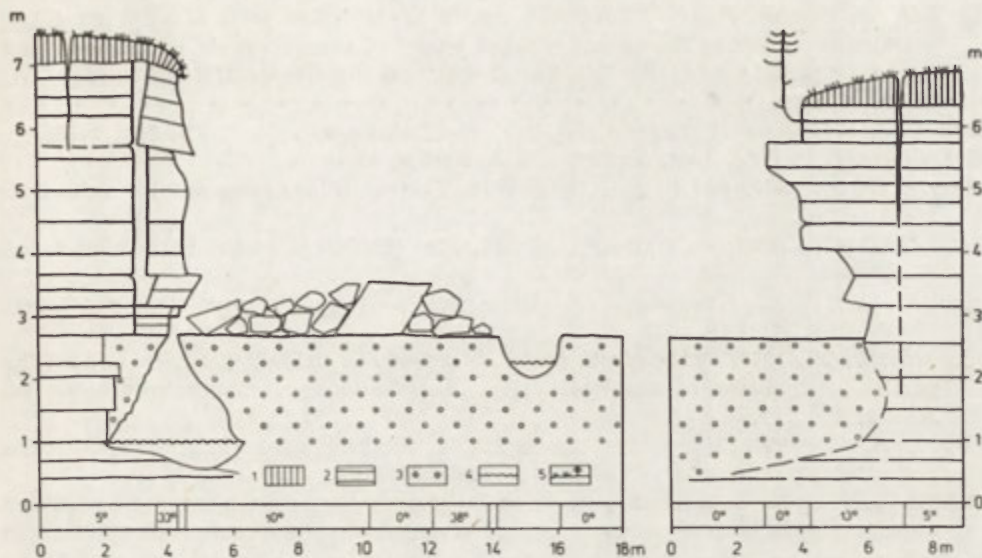


Fig. 15. Example of the effects of an icing on the gorge sides of the Segorym River (the right tributary of the Vostochnaya Khandyga River) in the Verkhoyanskiy Khrebet

1 — humus loam, 2 — sandstones, 3 — icing, 4 — water level, 5 — vegetation (grass, *Larix dahurica*)

one hand by water streams pushed to the valley side by the icing and on the other hand by the direct pressure of the icing. Considerable pressures develop within the icing often throwing out ice blocks to a distance of several tens of metres.

## CONCLUSIONS

The territory investigated represents a special type of mountain continental subnival region with an extensive occurrence of permafrost. The extremely low winter temperatures, the relatively high summer temperatures and the considerable dryness are mainly typical. Due to the shortage of water — the main agent of cryogenic processes — the intensity of the cryogenic processes is relatively small in comparison with the maritime subnival region. After the short spring period of intensive activity of cryogenic processes the desiccation of soil and a rapid drop in the intensity of the cryogenic processes take place.

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## BASES FOR THE STUDY OF PRESENT-DAY GEOMORPHOLOGICAL PROCESSES IN THE AUSTRIAN ALPS

THERESE PIPPAN

The study of recent geomorphological processes in the Austrian Alps can be based on several rather different types of data.

By comparison of air-photos of one and the same area but taken at different times morphological changes could eventually be established.

Data on fluvial erosion and load transport, filling in of reservoirs, mudflows, slides, slope movements, rock fall and landslides, recent tectonic movements and the extent of denudation are available.

### FLUVIAL EROSION AND LOAD TRANSPORT

Important details on the proportion of fluvial erosion and load transport can be found in the water supply register of the Ministries of Commerce and Agriculture in the reports of the water control offices of the governments of the federal countries, and of the Federal Experimental Station for Waterworks in Vienna and in scientific papers on this subject.

According to a lecture in 1968 by W. Kerschbaumer at the Waterworks Conference in Salzburg, the intensive downward erosion of the River Salzach between Hallein and Oberndorf is due mostly to control measures in the last century by which the river course has been shortened and confined and the transport capacity increased. Up to 1959 the average annual degree of downcutting of the Salzach at the Staatsbrücke in Salzburg was 4.4 cm, in Hallein 7 cm. Over 60 years it reached a total value of about 5 m. Devastating floods (1920, 1959) caused downcutting of about 1 m each time. The load deficit associated with downward erosion caused the dissection of the gravel bottom down to fine sand at its base by which process progressive, very quick downcutting resulted. After K. Bistritschan (1951) slumping, directed towards the river occurred. If the fine sandy-argillaceous beds at the bottom are under artesian pressure, they eventually flow out spontaneously.

Also between Taxenbach and Schwarzach the Salzach is exerting intensive downward erosion. Between 1886-1925 there was an annual downward erosion of 5.3 cm, and since 1911 a continuous downcutting of 2.1 cm was stated at the Salzach gauge near Lend. (Report of Water Control Zell am See). At the uplift-center of Embach the annual value of downcutting is attaining up to 4 cm. The Salzach has been able to dissect within 3 years the gravel dam deposited by the Embacher Plaike in 1794 which is up to 25 m high. Within the narrow

tract of the Fritz Valley at the railway-km 6.6 the fluvial downward erosion attained up to 1.27 m since the construction of the railway in 1875.

In 1934 the stream of the Pindlgraben in the Goriach Valley, Lungau cut during a flood, within half an hour, a 5 m deep file shaped gully into the slope detritus.

To intercept the vertical cutting of the Salzach below Hallein and Salzburg weirs have been constructed. The Federal Experimental Station for Waterworks in Vienna has determined the graded profile, the downcutting by means of a survey of the cross profile and the grain and quantity of the transported load. Measurements of the gravel have been taken from 1963–1965, 7 km below Oberndorf and carried out once a week. Up to 8 perpendicular measurement-poles have been fixed within a distance of 8–16 m. To obtain reliable data, measurements through 10 years would be required.

The annual gravel load attained at the assumed load density of  $1.72 \text{ t/m}^3$  for 1964 is  $87,000 \text{ m}^3$ , for 1965 is  $190,000 \text{ m}^3$ . The average annual load between 1901–1960 was  $115,000 \text{ m}^3$ . This value will decrease in the future because since the construction of the weirs the bottom erosion has stopped.

Data on the infilling of reservoirs are available from the administration of the power station companies. According to the data of the Tauern Power Station Company Ltd., one of the measurements in 1966 at the 2000 spot height of the Margaritze Reservoir reached an alluvial deposit of  $269,768 \text{ m}^3$  with a reservoir-volume of  $1,465,678 \text{ m}^3$ . The data have been obtained by echo-sounding sections. According to measurements in 1968 at the gorge reservoir of Kaprun at the spot height 897 m with a reservoir-volume of  $219,900 \text{ m}^3$  an alluvial deposit of  $15,993 \text{ m}^3$  resulted.

According to an oral communication of Dr. H. Slupetzky the Tauernmoss reservoir showed the following quantities of alluvial deposits transported by the Oedbach into the lake at the 2003 m spot height, measured at the alluvial fan:

- 1928–1938 average  $4250 \text{ m}^3$  per annum;
- 1938–1949 average  $3650 \text{ m}^3$  per annum;
- 1949–1959 average  $2000 \text{ m}^3$  per annum.

(Data from the Power Station Enzingerboden of the Austrian Federal Railways).

After G. Troppers (1950)  $300,000 \text{ m}^3$  of silt and  $50,000 \text{ m}^3$  of gravel are deposited annually in the transverse valley-section of the Mur, upstream of Graz in the Pernegg-reservoir. During flushing in the spring of 1936 the silt-load attained a maximum of 3%.

Valuable observations on recent morphological processes are presented in the papers of S. Morawetz from Graz. In 1957 the quickly developing and disappearing ravines in the Tertiary hills east of Graz have been investigated which show an available relief of up to 120 m and slope angles up to  $10^\circ$ . After a cloud burst in 1956 a small horse-shoe shaped opening developed, this process being associated with an outflow of water. With overgrown stabile lateral slopes, ravines eventually develop by headward erosion at which a cloud burst may erode some meters deep gullies.

In 1957 S. Morawetz made observations on erosion and accumulation upon a sand bank at the River Drau near Villach. A streamlet of  $4\text{--}5 \text{ m}^3$  discharge per second eroded into the sand bank within 2–3 hours a box-shaped valley which was 25 cm wide, 150 cm long and 8 cm deep, the gradient being 12 cm. Microterraces developed along the streamlet and at its mouth a deltaic fan of 120 cm length at its lower curve also developed.



## SLOPE DENUDATION

The recent slope denudation in the mountains is generally controlled by tectonic, lithological and climatic factors. With solid rock it is performed by mechanic weathering due to freeze-thaw action. Rockfall occurs which builds up the talus cones. With soft impermeable slate and phyllite the slope denudation is affected by landslips and mudflows. Landslides also contribute to denudation. In 1963 the present author collected a large number of data on these morphological processes from the avalanche and river control of the local government of Salzburg, from papers and newspaper reports on this subject.

The mudflows occur especially with slate and phyllite of the Central Alps and the Greywacke-zone and with the Werfen slates at the southern margin of the Calcareous High Alps where unvegetated steep slopes are located on the dry sunside (adret) or on the ubac. The mudflow activity is favoured by large talus cones located upon impermeable slate or phyllite. If soaked by heavy precipitation this substratum provides a slip-plane for the talus. Examples are provided at the southern slope of the Calcareous High Alps at the border of the Greywacke-zone. Mud flows are caused by summer-thunderstorms with heavy rains after a long hot dry period. Man's economic activity may eventually encourage mudflows by wood-clearing.

Landslips occur especially with rocks disintegrating into thin slabs or marked by thin foliation which produce moist talus. This material, especially when deposited on smooth argillaceous rockplanes or boulderclay starts gliding even with a slope angle of only 2-3°. W. Polt (1950) observed that in the Slate Alps landslips are increasing in the regions of alpine pastures above the level of 1800 m. The water-sides and slopes facing NE or SW are especially attacked by landslips. These movements are concentrated in valleys with steep slopes for example in the Taxenbach gorge where the largest landslip area of Salzburg is located. This is the Embacher Plaike which came down on the 8 June, 1794. It is 360 m high and up to 14 m thick. On steeply dipping, water soaked slate and phyllite moraine is deposited, above it follow clay, interglacial sand and gravel and Würm moraine on top. With this structure the gravel masses move continuously downslope some cm or even dm per year. By this process the bed of the Salzach is confined to half of the original width. Stilted tree trunks upon the slopes and fissures as well as subsidences at the upper edge of the sediments point to the continuance of the landslip.

In Vorarlberg where the largest number of landslips occur the Schesatobel must be pointed out as a large scale landslip area. Originally there was an erosional cut in Partnach and Raibler beds into which morainic material and gravel consisting of calcareous and crystalline components have been deposited with their beds dipping downvalley. Probably glacial meltwaters have washed morainic material into this dead ground.

Through wood clearing the Schesatobel has been turned into a mountain torrent. Since 1899 it has been continuously controlled. Its back wall is 200 m deep and the area covers 60 ha. From this area up to 20 m high mudflows have been discharged with a speed of 38 m/h. Up to 50 million m<sup>3</sup> of material has been transported. Gypsum funnels point to a co-operation of slumping dolinas at the development of the ravines. Due to the prevalence of loamy-silty components continuous slumping occurs with the sediments, rents, wrinkles and landslips develop. Due to the activity of water, impermeable miclamellae turn into a soft viscous slip-plane. Over 50 years the upper margin of the ravine has retreated upwards by 150 m. The loss of area attains 1600 m<sup>2</sup> per

annum, and the transport of material is 30,000 m<sup>3</sup>. By a system of concrete-barriers the slopes are being flattened so that they can be reforested.

A. Alker, H. Haas and O. Homan investigated slope movements in several regions of Styria in 1969. They occur especially in sandy-silty-loamy sediments of the Styrian Tertiary Hills and the inner Alpine Tertiary basins.

In the granite-quarry, near Weiz conspicuous slope-movements have occurred since 1959 during which a mylonite-zone between granite and slate provides the slip plane for the disintegrating granite-lense which is gliding downvalley. The intensive erosion of the Feistritz River and the quarrying activity favour gliding movements. Laminarily disintegrating finely foliated Paleozoic phyllite with a high proportion of clay favours creeping and slumping slope movements if the gradient of the valley sides and the content of slope water is higher, for example in the Enns- and Trieben Valley.

In the Tertiary Hills of the Grazer Basin with interbedding of impermeable clayey-loamy silt with water containing gravelly-sandy intercalations which are inclined to flow, favoured by fluvial undercutting at the toe of the slope large-scale creeping slope-movements occur at which under the pressure of the upper water containing beds clayey material may be extracted out of the slope. At the border of impermeable crystalline rock and of Tertiary rocks, landslips may occur during heavy rain. The use of agricultural machinery eventually causes slope movements.

By the construction of the Autobahn intensive slope movements have been caused.

If sandy-clayey silt is dissected near the surface conchoidal openings result penetrating up to 5 m into the slope. Elastic clay slumps downwards in stair-like manner. The partial movements of the slip-planes have been investigated as to the mineralogical and soilmechanic qualities. At the movement horizon, a grain distribution of 78% silt, 28% sand and 4% clay resulted. There is a clear relation between the grain structure and the mineralogic-petrographic components of the sediment. The contents of illite favour landslips.

During the construction of the Autobahn at Nestelbach a 10 m deep cut into the slope was produced. After the removal of a 2 m thick loam cap 60° towards the slope dipping slip-planes originated in the silt and clay due to relief of load associated with rents and joints.

Rockfall also is aiding the denudation of rock-faces. They occur especially in spring by frequent freeze-thaw action with bare or scarcely overgrown faces. With the recent climate talus production by rock fall is slight. Landslides occur when clayey-marly beds intercalating or underlying solid limestone are soaked with heavy, long lasting rain thus producing a slip-plane for the top beds. An example is the Sandling in the Salzkammergut where in 1920 7-8 millions m<sup>3</sup> of rock have been moved. A 90 m broad and up to 20 m thick rock stream advanced as far as 4.5 km into the Zlambach Valley during one month. The Sandling-crag was isolated and destroyed.

At the Dobratsch in Carinthia a second landslide occurred in 1348, after a postglacial huge landslide originating from three separate scars. Here 30 millions m<sup>3</sup> of boulders covered an area of 7 km<sup>2</sup> up to 5 m thick. Smaller slides occur even today.

Quantitative data on the proportion of recent denudation was presented by W. Wundt (1952), who used the measurements of the Bavarian Station for Hydrology at Burghausen. According to him, the annual denudation by suspended matter attained in the catchment area of the Salzach (6643 km<sup>2</sup>) was 269 t/km<sup>2</sup>. The denudation of 1 m by suspended material takes 9300 years. As to the 949 km<sup>2</sup> large catchment area of the Saalach, investigations at Jettenberg

showed an annual denudation by suspended material of 244 t/km<sup>2</sup>. The denudation of 1 m by suspended material takes 10,000 years. From the silt contents of the Raab River J. Stini calculated an annual denudation of the catchment area of about 0.1 mm (1920).

During investigations in the Lower Austrian Weinviertel Z. Mieczkowski (1961) pointed out the importance of heavy rains for soil denudation during which cloud-bursts especially are effective after a long hot and dry period. Soil denudation starts at a slope angle of 2-3°, on wooded slopes only at 20-30°. The average annual soil denudation in Lower Austrian vineyards at a slope angle of 15-20° is assumed to attain 0.5-0.8 cm.

In the summer of 1970 Dr. E. Stocker started investigations on slope movements in the SE-section of the Carinthian Kreuzeck Mts. This research is sponsored by the Director of the 2nd Chair of the Department of Geography at Salzburg University, Prof. H. Riedl. E. Stocker reports on these investigations: At the start of special slope investigations in a small area of the SE-section of the Kreuzeck Group it is proposed to carry on measurements of movements. For this purpose steel wires fixed on nails have been drilled into the soil. At the surface a plumbline was fastened through which wires are inserted. With a solifluidal movement the end of the wire is supposed to recede little by little; the proportion of movement can be checked with an accuracy of up to a millimeter. Difficulties exist with the placing of the nails if the horizon of decomposition is limited too strictly towards the bottom by very hard, unjointed rock. The phenomena of outfreezing of the soil can also be ascertained with this method if enough measurements have been made. This summer a total of 20 plumb-lines has been fixed at different levels and expositions in the research area. Besides this, markings on rocks of boulder streams are to state their movement. As check-points sections of solid rock located higher up have been selected. Other data are also to be obtained later; for example, grain-size spectra, soil-humidity, soil-temperature and exact mappings; up to now only easily gained data could be ascertained. The results of the movement-measurements will appear later on.

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## SOME RATES OF GEOMORPHOLOGICAL PROCESSES

LUNA B. LEOPOLD AND WILLIAM W. EMMETT

This brief report summarizes three sets of measurement data on certain processes.

The first concerns the rate of movement of soil on hillslopes, especially by mass movement or slow gravitational creep. The results are abstracted from an unpublished manuscript by the junior author who reports on the measurements which Leopold began 10 or more years ago and to which in more recent years Emmett has added new sites and has carried on the annual remeasurement. The results are those from "mass-movement lines", which consist of a series of pins or iron rods, 10 inches (25 cm) long driven vertically into the ground along a straight line-of-sight, secured at each end with stiff iron posts. The Survey consists of setting a theodolite over one of the end bench marks and orienting on the other. The distance of each individual pin from the line of sight is recorded. Resurveys are usually made annually.

These lines of pins have been established at sites mostly in the western United States at different altitudes, precipitation, and vegetation. The characteristics of the sites and the results of measurements are summarized in Table 1.

An interesting aspect of these data is the progression of downhill motion with time as can be seen in the successive resurvey data, especially in the sites for which 6 to 9 years of data are available. The cumulative downslope motion at the ground surface is shown for several sites in Figures 1 and 2. The cumulative curves show that the amount of motion in successive years was about equal.

Because various individual pins on a given mass-movement line are on somewhat different local hillslope gradients, the downhill motion and the surface erosion at different pins can be plotted against the local gradient as in Figure 3. This graph shows that downhill motion increases slightly with gradient but is less sensitive to gradient than is surface erosion. Emmett concluded that of the sites studied where local gradient is less than  $15^\circ$  the hillslopes studied do not experience either surface erosion or downslope mass movement.

The next set of data concern downslope creep measured in modified Young Pits. The measurement procedure is that described in the review of field methods for hillslope studies, published in *Revue de Geomorphologie Dynamique*, p. 157, 1967. In brief, a pit is dug in the soil and into one face of the exposed undisturbed profile a narrow slot is dug into which is placed a vertical plate consisting of small rectangles of aluminum. This column of discrete plates is placed in a plane parallel to the contours. The location

TABLE 1. Downslope movement of iron pins on mass movement lines (from Emmett)

	Average local gradient (degrees)	Rock type	Soil depth	Vegetation	Elevation (ft)	Annual precipitation (in)	No. of years of observation	Average <sup>1</sup> downslope movement (in/y.)	Surface erosion (in/y.)
Coyote Arroyo, nr. Santa Fe, N.M.	25	Alluvium	3 ft	Pinon woodland	7,000	14	7	0.12	0.46
Slopewash Tributary, Santa Fe, N.M.	35	"	"	"	"	"	9	.20	.40
Ski Basin Site, nr. Santa Fe N.M.	23	Granite	1-3 ft	Spruce	9,000	25	4	.30	0
Big View Site, Dickerson Park, Wyo.	21	Limestone	3-6 ft	Alpine grass	9,500	20	3	.14	.02
Twin Cabins Site, nr. Pine-dale, Wyo.	18	Bouldery till	—	Sage, aspen	8,000	15±	2	.12	0
Forsaken Gully Site, nr. Moneta, Wyo.	35	Shale	3-5 ft	Sage	4,000	14	6	.34	.30
Last Day Gully Site, nr. Hudson, Wyo.	16	Shale	3-6 ft	Sage	4,000	12	5	0	.05
Aching Shoulder Site, Midden Rock, nr. Four Corners, Ariz.	31	Igneous rubble	—	Semidesert grass	4,000	8	5	0	.10

<sup>1</sup> Corrected to ground surface, and data are best graphical fit.

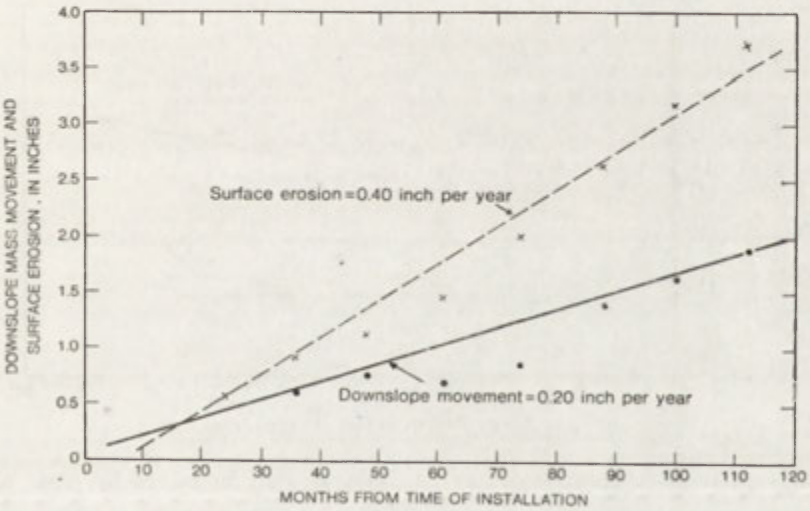
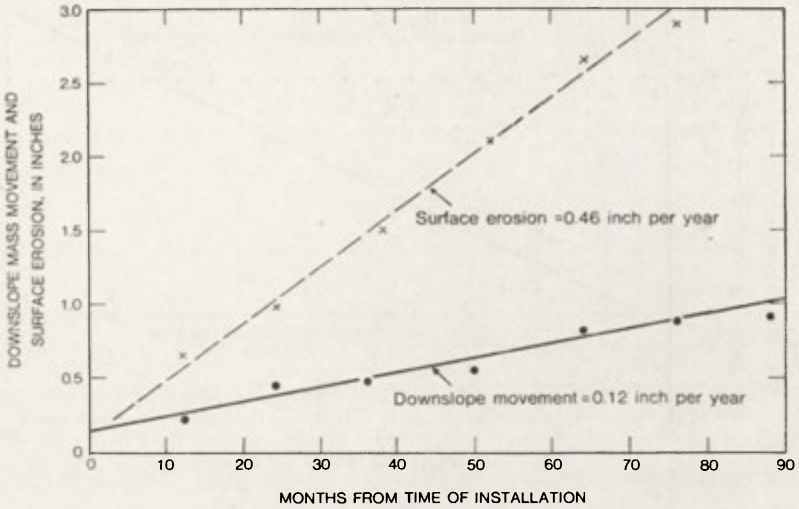


Fig. 1. Cumulative downhill creep and surface erosion measured by mass movement pins  
upper — Coyote Arroyo; lower — Slopewash Tributary, both near Santa Fe, New Mexico, U.S.A.  
(from Emmett, 1971)

of each plate in the vertical column is determined by theodolite sighting between iron rods driven deeply in the ground to serve as bench marks. After installation and recording, the pit is filled.

Preferably two years later, the pit is re-excavated and the location of the plates resurveyed.

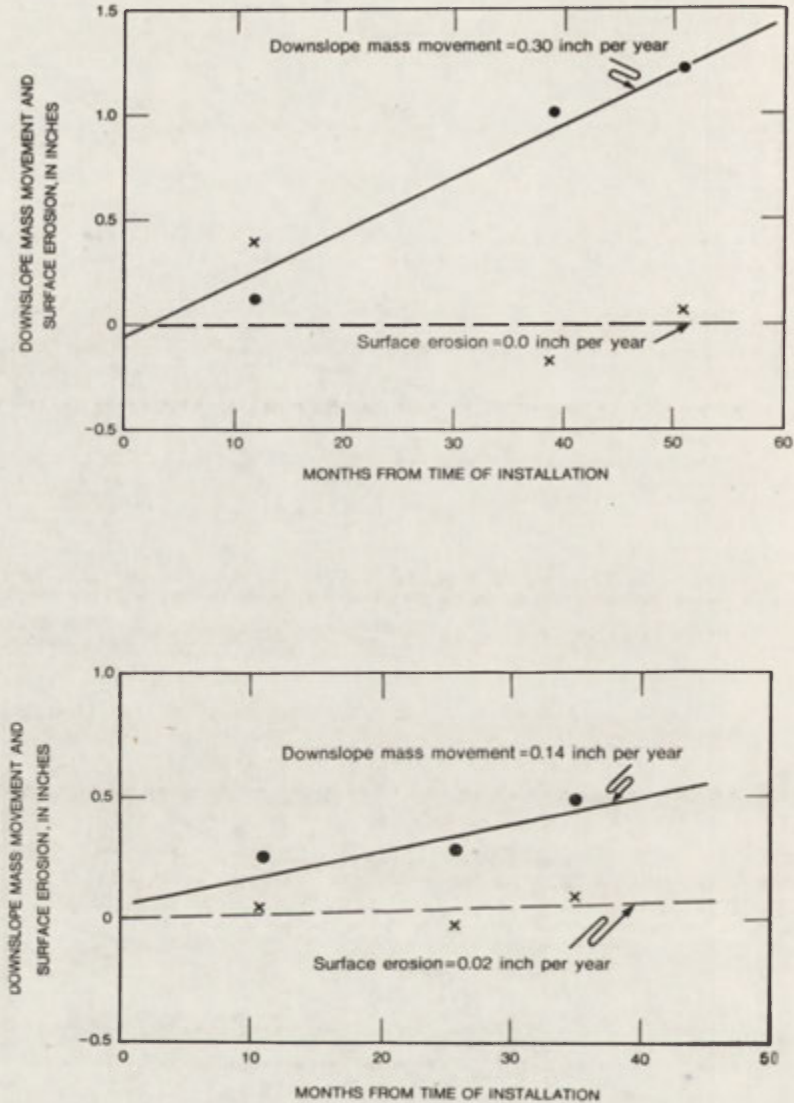


Fig. 2. Cumulative downhill creep and surface erosion measured by mass movement pins  
 upper — Ski Basin Slope near Santa Fe, New Mexico; lower — Big View Slope, Dickerson Park, Wyoming U.S.A. (from Emmett, 1971)

The data presented are for 6 such pits installed in 1964 or earlier and for which the latest resurvey was in 1971. The movement, therefore, represents that taking place over 7 years. The location is on wooded hillslopes in the headwaters of Cabin John Creek, a tributary to the Potomac River near Washington, D. C. The annual precipitation is 45 inches, most of which occurs in non-summer periods but there are heavy thunderstorms in summer which wet the soil for short periods. The soil is frozen for at least two winter months



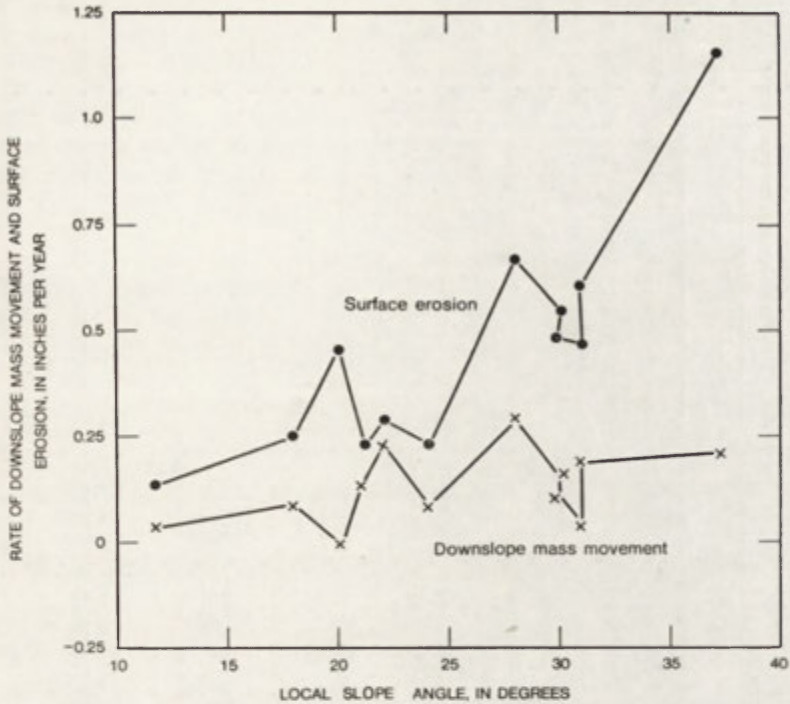


Fig. 3. Erosion and mass-movement rates as a function of local slope angle for mass movement pins, Coyote Arroyo near Santa Fe, New Mexico, U.S.A. (from Emmett, 1971)

and may remain moist for as long as four months during winter. It is presumed that downhill creep is concentrated in late winter and spring.

The soil material is deep, representing the Piedmont, which experienced long periods of weathering in pre-Pleistocene as well as Pleistocene time. Solid bedrock crops out in places within the study area but generally a weathered regolith at least 10 feet deep covers most of the hillslopes. The bedrock is gneiss containing veins of quartz which provide hard cobbles to the local streambeds. The forest is second growth, tulip-oak-hickory association.

In all except the earliest installations we drove an iron rod deep in the soil with the upper end at the level of the bottom of the excavated pit. This was to be a check on whether the bench marks (iron rods) exposed at the ground surface had been moved downhill as a result of soil creep. This procedure turned out to be most fortunate. When the 1971 surveys were compared with data at the time of installation, the buried bench marks appeared to have moved uphill in every pit where a buried rod was used. All pits for which buried bench marks were not used have been eliminated from the tabulation. These amounts of apparent uphill movement are shown at the bottom of Table 2 under the heading "Measured relative movement" and opposite "Buried Bench Mark".

A careful study was made of sources of error-instrumental, reading, and computing errors. We conclude that none of these sources could account for the apparent uphill movement of the buried rods. The best explanation is

TABLE 2. Measured and adjusted measurements of downhill creep in modified Young Pits Sisters area near Bethesda, Maryland, U.S.A., 1964-71 (7 years)

Depth (cm)	Measured relative movement (cm)						Movement corrected by reference at depth (cm)						Probable actual movement (cm)					
	Pit 1 plates	Pit 2 pins	Pit 4 plates	Pit 5 plates	Pit 7 plates	Pit 8 plates	Pit 1 plates	Pit 2 pins	Pit 4 plates	Pit 5 plates	Pit 7 plates	Pit 8 plates	Pit 1 plates	Pit 2 pins	Pit 4 plates	Pit 5 plates	Pit 7 plates	Pit 8 plates
0	+0.08	.....	+0.28	+(2.26)	+3.52	(+1.33)	+0.18	.....	+0.56	(+1.96)	+0.652	(+1.51)	0±.2	.....	+0.6±.2	(+2.0±.2)	+0.7±.2	(+1.5±.2)
3	+13	.....	+04	+43	+3.91	+07	+23	.....	+13	+32	-.091	+25	.....	+3±.2	0±.2	0±.2	+3±.2	
6	+12	+0.15	-.56	-.80	-.16	-.07	+22	+0.88	-.28	-.50	+14	+11	+9±.2	0±.2	.....	0±.2	0±.2	
9	+17	.....	+36	-.36	-.11	-.34	+27	.....	-.08	-.06	+19	-.16	.....	.....	.....	.....	0±.2	
12	-.08	-.30	-.30	-.44	-.08	-.23	+02	+37	-.02	-.14	+22	-.05	+4±.2	.....	.....	.....	.....	
15	+15	-.47	-.28	-.42	+.22	-.66	+25	+26	0	-.12	+07	-.48	0±.2	.....	0±.2	.....	.....	
18	-.01	.....	-.03	-.50	-.23	-.24	+09	.....	+25	-.20	+07	-.06	.....	.....	.....	.....	.....	
21	-.06	-.44	-.33	-.54	-.23	-.32	+04	+29	-.05	-.24	+07	-.14	.....	.....	.....	.....	.....	
24	+08	.....	-.59	-.48	-.16	-.14	+18	.....	-.31	-.18	+14	+04	.....	.....	.....	.....	.....	
27	+03	-.40	-.28	-.48	-.23	-.12	+13	+33	0	-.18	+07	+06	.....	.....	.....	.....	.....	
30	-.03	.....	-.30	-.58	-.16	-.07	+07	.....	-.02	-.28	+14	+11	.....	.....	.....	.....	.....	
34	+01	-.50	-.30	-.62	-.16	.....	+11	+23	-.02	-.32	+14	.....	.....	.....	.....	.....	0±.2	
37	-.05	.....	.....	.....	.....	.....	+05	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
40	+03	.....	.....	.....	.....	.....	+13	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
43	+12	-.44	.....	.....	.....	.....	+22	+29	.....	.....	.....	.....	.....	.....	.....	.....	.....	
46	-.16	-.60	.....	.....	.....	.....	-.06	+13	.....	.....	.....	.....	.....	.....	.....	.....	.....	
49	-.12	.....	.....	.....	.....	.....	-.02	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
52	-.20	.....	.....	.....	.....	.....	-.10	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
55	-.14	.....	.....	.....	.....	.....	-.04	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
58	-.23	.....	.....	.....	.....	.....	-.13	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	
61	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	0±.2
Buried bench mark	-.10	-.73	-.28	-.30	-.30	-.18	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....

+ is downhill motion; - is uphill.

( ) indicates possible disturbance and thus doubtful record.

"Plates" refer to pits where flat aluminium plates were the markers moving downslope.

"Pins" refer to pits where markers are horizontal rods rather than flat plates.

Ground surface gradient at each pit: Pit 1, 34%; Pit 2, 35%; Pit 4, 54%; Pit 5, 40%; Pit 6, 25%; Pit 8, 30%.

that the iron rods extending from the surface downward and marking the line of sight from which the movement of installed plates is measured have also moved downhill. Therefore, the observed movement of plates at various depths below the surface have been corrected by adding to observed relative movement that of the buried bench mark. These corrected values appear in the center section of Table 2.

In the right hand section of the table the final results are presented as probable amounts of downhill motion, with a confidence limit for each. The

TABLE 3. Erosion and deposition in channel of Watts Branch near Rockville, Maryland, U.S.A., determined by annual remeasurement of monumented cross-sections (Drainage area, 3.7 sq. mi.)

Cross-section name	Channel area		Number of years	Ratio:	Percent change channel area per year %	Bed elevation change (ft)
	original (sq.ft)	1970 <sup>1</sup> (sq.ft)		1970 original area		
1-2	73.1	56.5	17	0.77	-1.3	+0.4
1-4	106.8	53.5	17	.50	-2.9	+.8
1-5	74.8	52.7	17	.70	-1.7	-.1
44A-45	49.6	28.0	9	.56	-4.8	-.4
48-10	29.5	38.6	8	1.31	+3.9	-.5
40-41	31.7	31.8	8	1.00	0	-.2
15-17	133.0	74.3	17	.56	-2.6	+1.3
42-43	100.0	64.5	8	.64	-4.5	+.6
15-18A	85.0	59.0	17	.69	-1.8	+.3
35-34A	37.8	38.8	12	1.02	0	+.3
32-33	52.1	58.9	14	1.13	+.9	+.6
46-47	50.1	30.9	11	.62	-3.4	0
22A-47	60.2	41.7	17	.69	-1.8	-.5
25A-23A	72.8	59.0	17	.81	-1.1	+.7
22-24	58.2	41.6	17	.71	-2.4	+.4
Average all data	67.2	48.6		.78		
Average for 17-year data only	83.0	54.9	17	.66	1.7	+.4

<sup>1</sup> Includes some deposition on flood plain which is not within the channel as usually defined.

confidence, usually 2 mm, represented the average of the uphill or downhill corrected movement values at a depth where it may be assumed no movement is occurring.

The results indicate that despite the thick layer of regolith, downhill creep was measurable by our methods only in the upper 3 to 6 cm. The amount of this motion — even in the upper layers — was less than the measurement error in two out of six pits. The amount of movement varied from 0 to 9 mm in 7 years.

Our conclusion is that in this area of 45 inches (114 cm) of annual precipitation, a locality of deep soil material, and in an open hardwood forest, soil creep occurs only in the topmost 6 cm and at a rate of about  $\frac{1}{2}$  mm/year.

For such small rates of movement, measurement methods deserve careful consideration and care in their use.

The third set of data summarized here concerns changes in the channel of a small perennial stream, Watts Branch, a tributary to the Potomac River, which drains westward about 15 miles north of Washington, D. C. The measurements were made in an alluvial valley through which this stream meanders at a place where its drainage area is 3.7 square miles (9.6 km<sup>2</sup>). Land use is principally for agricultural crops. The average annual discharge is 3.22 cfs (.091 m<sup>3</sup>/s). In the study area, along a river distance of 700 feet (213 m), the width of water surface at low flow is 13 feet (4 m) and at bankfull, 19 feet (5.8 m). The river slope is .00397.

In 1953, Leopold established a series of channel cross sections, the ends of which were secured with iron posts driven below the ground surface so that they could be relocated by measurement but would not be disturbed by surface use. These cross sections have been resurveyed every year. Some had to be abandoned due to lateral movement of the channel and other new ones were installed. As of 1970, there were annual resurvey data available for 17 years for 8 cross sections and a minimum of 8 years of record for the others.

The changes in the cross sectional area of channel are summarized in Table 3. The land use in the surveyed zone has been pasture. During the 17 years urbanization has encroached on portions of the headwaters. The progressive decrease of channel area can at least in part be attributed to the increased loads of suspended sediment attributed to land opened for construction. The construction of houses has decreased the amount of drainage area devoted to farming, but even in 1970 the majority of the area is still farmland.

In the 17 years the channel has decreased in cross-sectional area an average of 1.7% per year. For those cross sections measured during the full 17 years, the 1970 channel is 0.66 that existing in 1953, a loss of 28 square feet out of an original 83 square feet. Most of this is due to narrowing by the plastering of silt on channel banks. The width-depth ratio has decreased. The streambed elevation has risen 0.4 foot but the channel slope has not changed significantly.

There has been deposition overbank, near the channel but outside the usual high water channel. This, however, is not uniform, but only in some locations. The importance of this overbank or out-of-channel deposition has not yet been analyzed.

The channel has moved laterally in some places a large amount, but in some sections no lateral change of position has occurred. The largest lateral movement on any cross section was 20 feet or about one channel width.

More details worthy of note are contained in the data but omitted here for brevity. Photographs of the channel, a description of the bank erosion process, and a map of portions of the study area have been published (Leopold, et al., p. 87, 88, 325, 468).

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## NEW METHODS OF MAPPING THE WATER BUDGET OF SLOPING AREAS

LASZLO GOCZAN

One of the objectives of the author's research programme in soil-geography has been first to determine — at an accelerated rate and in a way suitable for comparisons — the infiltration and runoff fractions of rainwater precipitating onto a sloping soil surface, the dependence on the rate of precipitation and the angle of slope, and then to develop a relevant mapping method.

The achievement of these objectives is important for science, planning and agricultural practice alike.

### EXPERIMENTAL AND MATHEMATICAL METHODS FOR THE DETERMINATION OF PERMEABILITY AND RUNOFF

#### HISTORY

Improving the techniques of J. Mattyasovszky (1953, 1957, in Di Gléria, Klimes-Szmik, Dvoracsek), B. Kazó designed a rainfall simulator (1962) which, after also being improved (1967), was suitable to permit an artificial rain precipitate rate of 20 to 40 mm/h onto an original soil monolith of 0.25 m<sup>2</sup> surface area and 20 cm thickness. The apparatus could be operated on slopes of 0, 8, 15, 21, 30 and 40% and during this artificial rain both permeability and runoff could be measured after the soil had been saturated up to its free-air water capacity.

It is this apparatus that the author made use of in order to develop a method for the determination of permeability and runoff as a function of the rate of precipitation and the slope percentage for given soil types and varieties.

Using the results of experiments with this apparatus F. Szász and the author have developed various hydrological functions (1970a, 1970b). Thus, it has been possible to determine permeability and runoff as continuous functions of both the rate of precipitation and the slope percentage.

#### DETERMINATION OF PERMEABILITY AND RUNOFF AS A FUNCTION OF THE RATE OF PRECIPITATION

Permeability as depending on the rate of precipitation (for the 0 to 40 mm depth interval) is determined as follows:

A monolithic soil sample of different slope, corresponding to slope categories used in agriculture (0, 8, 15, 21, 30, 40%), is saturated up to its minimum water capacity. Starting from this saturation state, at first artificial

rain of 20 mm/h, then at a 40 mm/h rate, is applied to it until a ready infiltration is established. In this way, a permeability value will be obtained for each of the two precipitation rates, for each slope category.

With the aid of the permeability functions of Góczán-Szász (1970a) all the permeability values corresponding to the precipitation rates of the 0 to 40 mm/h range are determined separately for each of the 6 slope categories measured (0, 8, 15, 21, 30 and 40°/0).

In the field a soil selection is exposed. It is defined in terms of its characteristics of importance for the solution of genetic problems and for crop production. In order to obtain the necessary information, samples are collected for laboratory analyses. For the determination of the volumetric weight, porosity and minimum water capacity values of the soil, 3 structurally intact samples are taken at 10 cm intervals with the aid of a cylinder of standard size. Thus the successive genetic levels of the soil are sampled continually from the surface downwards. Near the exposed section an intact soil monolith is inserted into the frame of the rainfall simulator.

(a) If the soil exposed to the artificial rain allows a precipitation of 20 mm/h intensity to be completely infiltrated in the case of any slope category, formula (6) Góczán-Szász (1970a) is used for the determination of permeability corresponding to each selected rate of precipitation.

The formula is given by:

$$y = \frac{4y_1 - y_2 - 2x_1}{3} \cdot x + \frac{y_2 - y_1 - x_1}{3x_1^2} \cdot x^3$$

where:

$y$  = permeability, or infiltration rate mm/h, corresponding to the selected  $x$  rate of precipitation,

$x$  = one of the selected precipitation rates from the 0 to 40 mm/h interval, the values between 20 and 40 mm/h being excluded as a matter of course,

$x_1$  = rate of precipitation equalling 20 mm/h,

$y_1$  = permeability corresponding to 20 mm of precipitation per hour for the slope category under consideration,

$y_2$  = permeability corresponding to 40 mm of precipitation per hour in the slope category under consideration.

(b) If the soil exposed to the artificial rain does not allow 20 mm of precipitation per hour to infiltrate completely into the soils of any slope category, formula (15) Góczán-Szász (1970a) is used for the determination of the permeability corresponding to each particular precipitation value.

We then have the formula:

$$y = x + \frac{8y_1 - y_2 + 6x_1}{4x_1^2} \cdot x^2 + \frac{y_2 - 4y_1 + 2x_1}{4x_1^3} \cdot x^3$$

For an explanation of the symbols, see paragraph (a).

The runoff as a function of the rate of precipitation is determined (in the 0 to 40 mm/h interval) as follows:

After the execution of the field surveys described above in connection with the determination of permeability, the researcher exposes a soil monolith, saturated up to its minimum water capacity, to an artificial rain until a steady runoff is attained. (The rates of runoff and precipitation should be determined during one and the same run of the simulator). In this way, six runoff values will be obtained for the monolith exposed to artificial rain. Each of these values corresponds to one of the six slope categories under consideration.



After that, by means of the Góczán-Szász' runoff functions (1970a), all the runoff values corresponding to precipitation rates of 0 to 40 mm/h are determined for each of the six slope categories.

(a) If no runoff takes place on the monolith exposed to artificial rain at a rate of 20 mm/h in any of the slope categories, i.e., if the first part of our function curve is linear, than the formula (21) is used for the determination of the runoff corresponding to each particular precipitation value.

We then have the formula:

$$y = \frac{y_2 - 2y_1}{x_1} \cdot x + \frac{y_2}{x_1} \cdot x^2$$

The symbols used here have the same significance as in the case of permeability, the only difference being that the value of  $y$  refers to the runoff instead of to the infiltration rate.

(b) If any surface runoff is observed for 20 mm of precipitation per hour in any slope category, then the formula (29) Góczán-Szász (1970b) is used for the determination of the runoff corresponding to each particular precipitation rate.

We have the formula:

$$y = \frac{8y_1 - y_2}{4x_1^2} \cdot x^2 + \frac{y_2 - 4y_1}{4x_1^3} \cdot x^3$$

In Table 1 the author has quoted as an example the precipitation and runoff rates corresponding to each particular precipitation value within the 0 to 40 mm/h range for a brown forest soil sloping at 0, 8, 15, 21, 30 and 40<sup>0</sup>/o, respectively.

#### DETERMINATION OF PERMEABILITY AND RUNOFF AS A FUNCTION OF THE SLOPE

To be able to map soils of equal permeability or those which show the same rate of surface runoff, and to find out the quantities of water infiltrating into the soil and running off its surface respectively, one has to determine the permeability and runoff values corresponding to any precipitation rate. In collaboration with F. Szász, the author approached the problem of establishing this function by developing the formulae given below (Góczán-Szász, 1970b).

The permeability and runoff values, corresponding to arbitrary precipitation rates within the 0 to 40 mm/h range for slope categories of 0 to 15<sup>0</sup>/o, are expressed by the formula:

$$y = f(x) = y_0 \cdot \frac{(x-x_1)(x-x_2)}{(x_0-x_1)(x_0-x_2)} + y_1 \frac{(x-x_0)(x-x_2)}{(x_1-x_0)(x_1-x_2)} + y_2 \frac{(x-x_2)(x-x_1)}{(x_2-x_0)(x_2-x_1)}$$

For slopes between 15 and 30<sup>0</sup>/o, the formula is given by:

$$y = g(x) = y_2 \cdot \frac{(x-x_3)(x-x_4)}{(x_2-x_3)(x_2-x_4)} + y_3 \frac{(x-x_2)(x-x_4)}{(x_3-x_2)(x_3-x_4)} + y_4 \frac{(x-x_2)(x-x_3)}{(x_4-x_2)(x_4-x_3)}$$

For slopes between 30 and 40<sup>0</sup>/o, it is given by:

$$y = h(x) = y_4 \cdot \frac{x-x_5}{x_4-x_5} + y_5 \frac{x-x_4}{x_5-x_4}$$

TABLE. 1. Permeability and runoff values of brown forest soils (*Braunerde*) as a continuous function of the rate of precipitation; (a) permeability values mm/h; (b) runoff values mm/h

(a)

Rate of precipitation mm/h	Slope %					
	0	8	15	21	30	40
1	1	1	1	1	0.9	0.9
2	2	2	2	2	1.9	1.9
3	3	3	3	3	2.9	2.9
4	4	4	4	4	3.9	3.8
5	5	5	5	5	4.9	4.7
6	6	6	6	6	5.8	5.6
7	7	7	7	7	6.7	6.4
8	8	8	8	8	7.6	7.3
9	9	9	9	9	8.5	8.1
10	10	10	10	9.9	9.4	8.9
11	11	11	11	10.9	10.3	9.7
12	12	12	12	11.8	11.1	10.4
13	13	13	13	12.7	11.9	10.4
14	14	14	14	13.6	12.7	11.8
15	15	15	15	14.5	13.5	12.5
16	16	16	16	15.4	14.3	13.2
17	17	17	17	16.2	15.1	13.8
18	18	18	18	17.0	16.4	14.4
19	19	19	19	17.8	17.1	15.0
20	20	20	20	18.44	17.7	15.6
21	20.8	20.5	20.5	19.2	18.3	16.1
22	21.6	21.0	21.0	19.8	18.9	16.6
23	22.4	21.5	21.4	20.5	19.4	17.2
24	23.2	22.0	21.8	21.0	19.9	17.6
25	24.0	22.5	22.2	21.6	20.4	18.1
26	24.8	22.9	22.5	22.1	20.8	18.5
27	25.6	23.3	22.8	22.5	21.2	18.9
28	26.4	23.7	23.1	22.9	21.6	19.3
29	27.1	24.0	23.4	23.2	21.9	19.7
30	27.9	24.4	23.6	23.5	22.2	20.0
31	28.6	24.7	23.8	23.7	22.5	20.3
32	29.3	25.0	24.0	23.9	22.7	20.6
33	30.1	25.2	24.1	24.0	22.9	20.9
34	30.8	25.5	24.3	24.1	23.0	21.2
35	31.5	25.7	24.4	24.1	23.0	21.4
36	32.6	26.0	24.4	24.0	23.1	21.6
37	32.9	26.1	24.5	23.8	23.1	21.8
38	33.6	26.3	24.5	23.6	23.1	21.9
39	34.2	26.4	24.5	23.4	23.1	22.1
40	34.9	26.5	24.5	23.0	23.0	22.2

(b) cont'd

Rate of precipitation mm/h	Slope %					
	0	8	15	21	30	40
1	0	0	0	0	0	0
2	0	0	0	0	0	0
3	0	0	0	0	0	0.1
4	0	0	0	0	0.1	0.1
5	0	0	0	0	0.1	0.2
6	0	0	0	0	0.2	0.3
7	0	0	0	0	0.2	0.4
8	0	0	0	0	0.3	0.6
9	0	0	0	0	0.4	0.8
10	0	0	0	0	0.5	0.9
11	0	0	0	0	0.6	1.1
12	0	0	0	0.1	0.8	1.4
13	0	0	0	0.1	0.9	1.6
14	0	0	0	0.2	1.1	1.9
15	0	0	0	0.3	1.4	2.2
16	0	0	0	0.4	1.6	2.5
17	0	0	0	0.6	1.8	2.8
18	0	0	0	0.8	2.1	3.2
19	0	0	0	1.0	2.5	3.6
20	0	0	0	1.3	2.8	4.0
21	0	0	0	1.6	3.2	4.4
22	0.1	0.1	0.2	1.9	3.5	4.9
23	0.1	0.3	0.4	2.3	4.0	5.4
24	0.2	0.5	0.6	2.7	4.4	5.9
25	0.3	0.8	1.0	3.1	4.9	6.4
26	0.5	1.2	1.4	3.6	5.4	6.9
27	0.6	1.6	1.9	4.2	6.0	7.6
28	0.8	2.1	2.5	4.8	6.6	8.2
29	1.0	2.7	3.2	5.4	7.2	8.8
30	1.3	3.6	3.9	6.1	7.9	9.5
31	1.5	4.1	4.8	6.8	8.6	10.2
32	1.8	4.8	5.7	7.7	9.4	10.9
33	2.1	5.6	6.7	8.6	10.1	11.6
34	2.6	6.6	7.8	9.5	11.0	12.4
35	2.8	7.5	8.9	10.5	11.9	13.2
36	3.2	8.6	10.1	11.6	12.8	14.1
37	3.6	9.7	11.5	12.7	13.8	14.9
38	4.1	10.9	12.8	13.9	14.8	15.8
39	4.5	12.1	14.3	15.2	15.8	16.7
40	5.0	13.4	15.9	16.6	17.0	17.7

TABLE 2. Permeability and runoff values of the rustbrown forest soil variation as a continuous function of the angle of slope

Slope %	Precipitation intensity (mm)			
	20		40	
	Permeability mm/h	Runoff mm/h	Permeability mm/h	Runoff mm/h
0*	20	0	34.9	5.0
1	20	0	33.6	6.5
2	20	0	32.3	7.7
3	20	0	31.0	8.9
4	20	0	29.9	10.0
5	20	0	28.9	11.0
6	20	0	28.1	11.9
7	20	0	27.4	12.8
8*	20	9	26.5	13.4
9	20	0	25.8	13.9
10	20	0	25.4	14.6
11	20	0	24.9	15.0
12	20	0	24.8	15.4
13	20	0	24.6	15.7
14	20	0	24.5	15.8
15*	20	0	24.5	15.9
16	19.7	0.2	24.2	16.0
17	19.3	0.5	23.9	16.2
18	19.1	0.7	23.6	16.3
19	18.8	0.9	23.4	16.4
20	18.6	1.1	23.2	16.5
21*	18.4	1.3	23.0	16.6
22	18.2	1.5	22.9	16.7
23	18.1	1.7	22.8	16.7
24	17.9	1.9	22.7	16.8
25	17.8	2.0	22.6	16.9
26	17.7	2.2	22.7	16.9
27	17.7	2.4	22.7	16.9
28	17.7	2.5	22.8	17.0
29	17.7	2.7	22.9	17.0
30*	17.7	2.8	23.0	17.0
31	17.5	2.9	22.9	17.1
32	17.3	3.0	22.8	17.1
33	17.1	3.2	22.8	17.2
34	16.9	3.3	22.7	17.3
35	16.6	3.4	22.6	17.3
36	16.4	3.5	22.5	17.4
37	16.2	3.6	22.4	17.5
38	16.0	3.8	22.4	17.6
39	15.8	3.9	22.3	17.6
40*	15.6	4.0	22.2	17.7

\* Measured values

List of symbols

- $x$  = the selected slope category
- $x_0$  = a slope of  $0^0/0$
- $x_1$  = " of  $8^0/0$
- $x_2$  = " of  $15^0/0$
- $x_3$  = " of  $21^0/0$
- $x_4$  = " of  $30^0/0$
- $x_5$  = " of  $40^0/0$
- $y$  = the rate of runoff or permeability being looked for and corresponding to the selected slope category
- $y_0$  = the rate of permeability or runoff observed for a slope of  $0^0/0$
- $y_1$  = runoff observed for a slope of  $8^0/0$
- $y_2$  = " " " "  $15^0/0$
- $y_3$  = " " " "  $21^0/0$
- $y_4$  = " " " "  $30^0/0$
- $y_5$  = " " " "  $40^0/0$

In Table 2 the author has quoted as an example the water quantities which filtered through the soil on the one hand, and those which ran off this surface in the order as observed for different slope angles up to a maximum of  $40^0/0$ .

THE MAPPING OF RUNOFF AND PERMEABILITY

(1) On the basis of field observations and of laboratory analyses carried out for a specially selected area unit (drainage basin, large farm, etc.), a genetic soil map of the territory is drawn up. On this map are recorded the various soil types down to soil varieties and their characteristics which influence the hydrological regime.

(2) An isoclinal map (illustrating the various slope categories present) is prepared for the same territory.

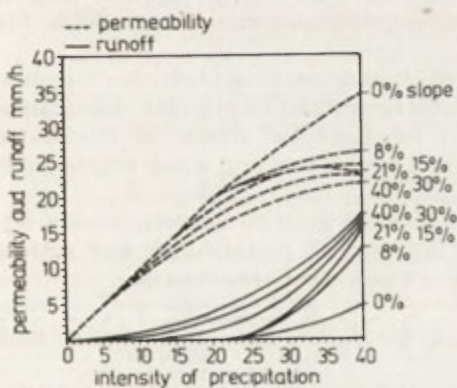


Fig. 1

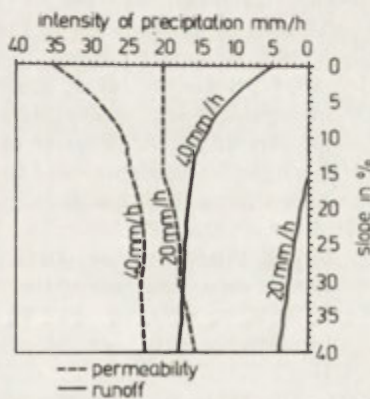


Fig. 2

(3) The genetic map is superimposed onto the isoclinal one, the former being represented by colours, the latter by ruling (hachure).

(4) By extrapolating the runoff values, measured for monolithic samples of different soil varieties or calculated with the aid of the above functions, for areas delimited according to slope categories and soil types, one can determine — by means of planimetry or calculations — the amount of water

drained off from each particular surface as a result of an arbitrary rate of precipitation, or the intensity of precipitation most frequent in the area under consideration.

Addition of all the runoff values will give the figure for the total water amount drained off from the territory under consideration, depending on the type of soil, the angle of slope and the rate of precipitation.

In the same way, the map of soil permeability of the territory can also be drafted, a map that can be supplemented with a cartogram of the minimum water capacities of the soils.

Due to lack of space, it is impossible to tackle in this short paper the problem of improvement of the above discussed method of water regime mapping. All that the author would like to point out here is that he has managed to improve Kazo's rainfall simulator to the point that it can now meet even higher requirements. In addition, the author is working upon finding out the effective influence of slope, a phenomenon resulting in the fact that in case of high-rate precipitations the soil cannot be saturated up to its minimum water capacity before the onset of runoff on its surface.

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## A COMMENT ON PRESENT-DAY CHEMICAL DENUDATION IN POLAND

MARIAN PULINA

The author reports the results obtained from his investigations of chemical denudation in Poland which he has studied by a new method, called by the author the hydrometric method. From 1960 to 1970 he carried out measurements in over a dozen thoroughly investigated experimental drainage basins built out of carbonate rocks, and in some fifty or so non-karst areas (Fig. 1). He then correlated the results of this field work with all the sources which from a theoretical point of view might affect the intensity of the processes of

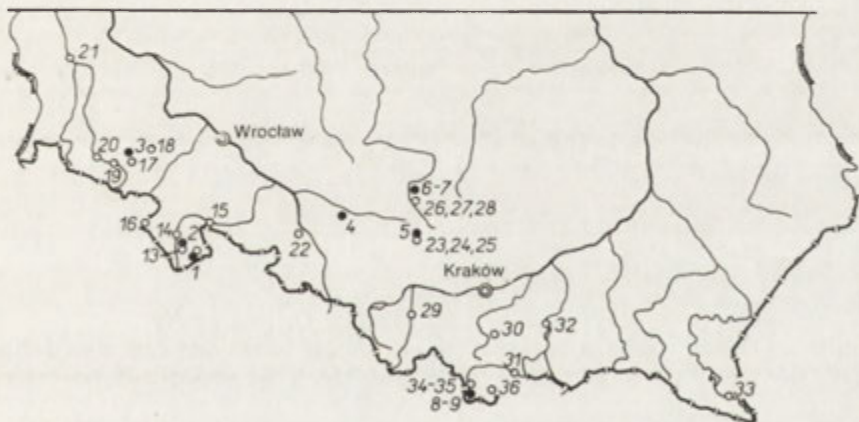


Fig. 1. Distribution of areas for which chemical denudation was determined by the hydrometric method  
1-9 karst regions, 10-36 non-karst areas

denudation involved. Next he calculated regressive formulae in connection with those agencies which bore upon the intensity of denudation. These formulae became the basis from which he determined the potential chemical denudation in force in Poland.

Chemical denudation takes place by chemical weathering and by removal of the dissolved salts. The part of the dissolved salts to be carried off and where it will be redeposited depends on the physico-chemical geographical environment. Depending on the distance over which the salts are displaced, denudation may be called local or regional. The former takes place within

small morphological units, for the most part on scarps. The latter embraces regional units and larger hydrographic basins. Investigations of local denudation involve collecting detailed record of the locally developing microforms, analyses of quantities produced within selected experimental regions, etc. On the other hand, studies of regional denudation concentrate upon the determination of hydrochemical parameters in surface streams and in springs, at the points where they leave the region under investigation. The present paper deals with the results obtained from research into regional denudation.

Everything reported here has been compiled under the guidance of Prof. A. Jahn of the Wrocław Institute of Geography, to whom the author wishes to express his sincere gratitude.

#### THE METHOD APPLIED

The area for which denudation has been calculated using the hydrometric method, is a hydrogeological drainage basin. The conditions which determine satisfactory results are that the basic parameters for the basin in question must be known, comprising:  $P$  = actual area of basin, in  $\text{km}^2$ ,  $Q$  = mean annual runoff, in  $\text{m}^3/\text{s}$ ,  $q$  = mean annual unit flow, in  $1/\text{s}/\text{km}^2$ ,  $T_a$  = mean annual mineralization of allochthonic waters penetrating the basin, in  $\text{mg}/\text{l}$ , for example: water arriving from non-karst areas, precipitation waters, anthropogenic mineralization;  $T$  — mean annual mineralization of waters escaping from drainage basin, in  $\text{mg}/\text{l}$ . The denudation taking place within the basin is expressed by the formulae:

$$D_m = 12.6 \frac{\Delta T \cdot Q}{P} \quad \text{where} \quad \Delta T = T - T_a \quad (1)$$

$$D_t = 31.5 \frac{\Delta T \cdot Q}{P} \quad (2)$$

$$D_m = 0.0126 \cdot \Delta T \cdot q \quad \text{where} \quad q = 1000 \frac{Q}{P} \quad (3)$$

$$D_t = 0.0315 \cdot \Delta T \cdot q \quad (4)$$

$D_m$  is expressed in  $\text{m}^3/\text{km}^2/\text{y}$ . or in  $\text{mm}/1000 \text{ y}$ .,  $D_t$  in  $\text{t}/\text{km}^2/\text{y}$ .

Knowing the above parameters of the drainage basin one can also define the quantity of dissolved salts which is carried off. This value can be calculated as follows:

$$A_m = 12.6 \cdot \Delta T \cdot q \quad (5)$$

$$A_t = 31.5 \cdot \Delta T \cdot Q \quad (6)$$

with  $A_m$  expressed in  $\text{m}^3/\text{y}$ . and  $A_t$  in  $\text{t}/\text{y}$ .

By the use of the hydrometric method, the extent of regional mechanical denudation can also be determined. This value is obtained by substituting for  $\Delta T$  the weight, in  $\text{g}/\text{m}^3$ , of the mineral substance carried off in suspension or dragged over the floors of streams.

Alongside the direct methods for determining denudation to which the hydrometric method belongs indirect methods are also frequently applied, one of which is, for instance, the climatic method suggested by J. Corbel (1959):

$$X = \frac{4 \cdot E \cdot T}{100} \quad (7)$$



where:

$X$  = denudation in  $m^3/km^2/y.$  or  $mm/1000 y.$ ,

$E$  = precipitation less evaporation, in  $cm$ ,

$T$  =  $CaCO_3$  content, in  $mg/l.$

In view of the difficulties in accurately defining, from the equation of water balance, the quantity of water passing the karst area, the author has refrained from reporting here the results of calculations obtained by application of this method.

Further climatic methods based on a modification of Corbel's formula are also known. Worth mentioning here are the following authors: I. Gams (1966), G. Groom and V. Williams, M. Sweeting et al. (1965), P. Habic (1968), etc. J. Corbel is also the author of a paper on the intensity of denudation observed in different climatic zones (Corbel 1959). However, his theories clash with the results obtained by research work done by F. Fournier (1960) and N. Strachow (1967).

#### CHEMICAL DENUDATION OF KARST AREAS IN POLAND

The quantitative determination of karst denudation in Poland has been the topic of papers published, among other authors, by: J. Corbel (1965), M. Markowicz (1968), H. Maruszczak (1966), K. Oleksynowa (1970), M. Pulina (1962, 1964, 1970, 1970a), J. Rudnicki (1967). All these studies refer to a variety of regions in Poland and were prepared by applying different methods. So far there was no uniform research method embracing all karst regions of Poland.

The author distinguished two types of chemical denudation occurring in karst regions. Type one, deals with vertical infiltration of atmospheric waters which principally attack the surfaces of limestones and their subsurface zones. The author asserts that, of the total corrosion caused by precipitation waters, from 80% (the Western Tatras) to 100% (the Snieznik massif in the Sudetes) represent the destruction of surfaces and subsurface zones. Hence it is this type of chemical denudation which accounts for the magnitude of denudation of limestone surfaces; these damages are expressed in  $mm/1000 y.$  of the hypothetically removed surface layer. The second type of denudation (interior ablation) takes place at the expense of fluvial waters which penetrate karst regions (this is the so-called horizontal infiltration). This type of denudation takes place particularly in those drainage basins, where limestones occur in close proximity to insoluble rocks, for the most part in the Sudetes and the Tatras. Thus, in the Klesnica valley situated in the Snieznik massif (Fig. 1, No. 1), chemical denudation due to horizontal infiltration amounts to  $106 m^3/km^2/y.$ , while the destruction of the surface caused by vertical infiltration is only  $33 m^3/km^2/y.$  (Table 1). The second type of denudation has no direct influence upon the magnitude of degradation of limestone surfaces, but it definitely affects the underground ablation of blocks of carbonate rocks. For this latter type of denudation the rate:  $m^3/km^2/y.$  will be the unit of measurement.

In his research work the author covered 11 karst drainage basins situated in the Sudetes, the Silesian Plateau, the northern part of the Cracow-Częstochowa Plateau and the Western Tatras (Fig. 1). The results of these investigations dealing with type one of karst denudation are listed in Table 1, showing values ranging from 17.2 to 48.9  $mm/1000 y.$ , and a mean value of 32.0  $mm/1000 y.$  The lowest results are the figures for the northern part of the Cracow-Częstochowa Plateau and the Silesian Plateau, the highest results are for the Western Tatras, with intermediate values being determined for the Sudetes.

In his own detailed investigations the author failed to cover all karst regions of Poland. However, the results he obtained enabled him to use these to calculate the potential karst denudation for all of Poland. In order to do this, he correlated the magnitude of denudation ( $D$ ) with structural features (density of joints and chemical composition of the limestones), with morpholo-

TABLE 1. Chemical denudation of some karst areas in Poland  
(vertical infiltration) determined by the hydrometric method  
[3]  $D_m = 0.0126 \cdot \Delta T \cdot q$  [5]  $D_m = 12.6 \cdot \Delta T \cdot q$

No. *	Area examined	CaCO <sub>3</sub> $D_{\Sigma M}^{**}$ mm/ /1000 y	CaCO <sub>3</sub> $A_{\Sigma M}$ m <sup>3</sup> /y	$T$ mg/l	$Ta$ mg/l	$\Delta T$ mg/l	$Q$ m <sup>3</sup> /s	$P$ km <sup>2</sup>	$q$ l/s km <sup>2</sup>
1.	Sudetes-Kleśnica Valley	33.0 33.9	7.4 7.6	174 187	5 15	169 172	0.0035	0.224	16.0
2.	Sudetes-Krowiarki (Wapniarka-Slupiec)	26.4 27.0	37.0 37.9	305 316	5 15	300 301	0.0098	1.4	7.0
3.	Sudetes-Kaczawa Mts. (Wapniki mountain)	20.5	57.4	311	5	306	0.0149	2.81	5.5
4.	Silesian Plateau (Strzelce Opolskie Region)	26.2 33.3		270 350	10 20	260 330			8.0
5.	Silesian Plateau Czarna Przemsza Valley	26.3		230	10	220			9.5
6.	Cracow-Częstochowa Plateau. Julianka. Wiercica and Sygatka Valleys	17.2	3160	185	10	175	1.44	183.7	7.8
7.	Cracow-Częstochowa Plateau. Wiercica, Janów	24.4	610	195	10	185	0.263	25.2	10.4
8.	Tatras-Lodowe Spring system	46.6 48.9	378 397	110 120	10 15	100 105	0.3		37.0
9.	Tatras-Kościeliska, Kiry	48.9 53.6	1429 1565	115 130	10 15	105 115	1.08	29.2	37.0

\* Order of numbering as in Fig. 1

\*\*  $\Sigma_M$  = sum of mineralization = dry residue of parts soluble in water.

gical features (mean absolute and relative altitudes), with hydrogeological properties (thickness of the *vadose* zone) and with climatic conditions. The most favourable coefficients of correlation ( $r$ ) were obtained with regard to annual sums of atmospheric precipitation ( $Op$ ) and to mean annual air temperatures ( $t^\circ C$ ).

$$r_{DOp} = 0.948$$

$$r_{D t^\circ C} = 0.884$$

The true value of the coefficient of correlation on the 0.1% level is  $r \geq 0.75$ . These high coefficients of correlation indicate that karst denudation depends, first of all, on climatic conditions and that a full understanding of climatic features may be useful for defining denudation. From the magnitude of the coefficient of determination ( $100 r^2$ ) one might assert that karst denudation

depends in 90% on the annual sum of precipitation and in 78% on the mean annual air temperature.

Further studies, aimed at establishing regression formulae, were based on the assumption that, alternatively:

$$D = f(Op)$$

$$D = f(t^{\circ}\text{C})$$

$$D = f(Op, t^{\circ}\text{C})$$

For the above cases statistical analyses were compiled in order to determine the types of the curves of regression. The chart presented in Fig. 2. was useful in examining the interrelation between karst denudation and the precipitation proved. It comprises data collected from 40 karst areas in Asia and Europe

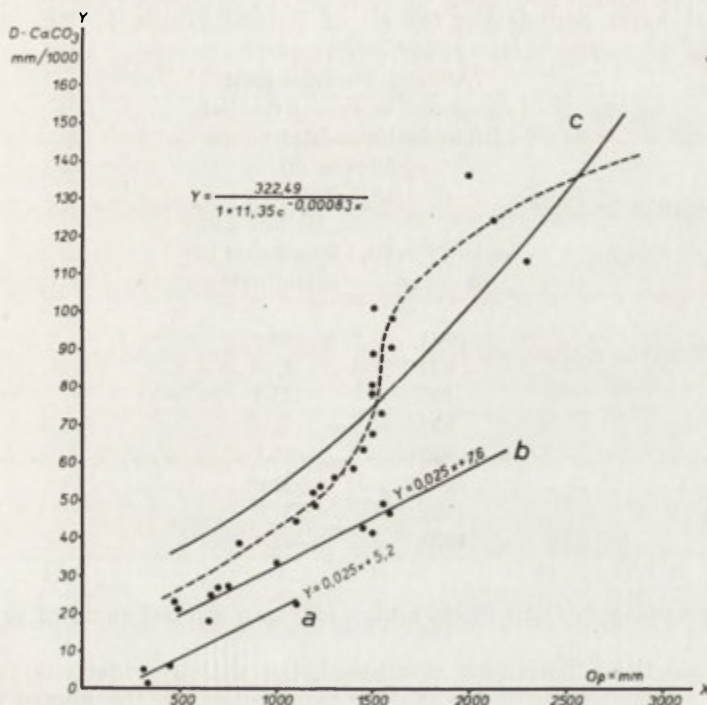


Fig. 2. Dependence of chemical denudation on the annual sum of atmospheric precipitation in 40 karst areas of Europe and Asia

a — karst in Eastern Siberia, b — karst in Central Europe, c — karst in areas of Southern Europe

(Pulina 1970). The indices of karst denudation determined for Poland lie on this chart amidst the mass of points (Fig. 2, curve b) referring to the European zone of a transitional temperate climate. The approximate position of this interdependence is best shown by the linear function:

$$D = 0.025Op + 7.6 \quad (8)$$

where:

$$r = 0.948$$

A comparison of the results obtained from applying the above equation with the direct measurements listed in Table 1 shows a high degree of conformity.

The interrelation between karst denudation and mean annual air temperature finds its expression in the formula:

$$D = -4.05t^{\circ}\text{C} + 53.7 \quad (9)$$

for  $r = -0.884$

However, this formula is less useful than (8); its conformity is worse and deviations are fairly wide.

After calculating the one-feature regressions, the author passed on to consider multi-feature regressions:

$$D = -43.44 + 0.0485O_p + 5.8737t^{\circ}\text{C} \pm \varepsilon \quad (10)$$

where  $(\varepsilon) \leq 12.4$ .

Among all the regression formulae given above, the approximations obtained from the formula (8) are the best and the author used this formula to determine the potential karst denudation for all of Poland (Table 2, Fig. 3). Figure 3

TABLE 2. Potential karst denudation in Poland (vertical infiltration) calculated by use of formula (8)  
 $D = 0.025 O_p + 7.6$

Precipitation ( $O_p$ ) in mm/y	Denudation ( $D$ ) in mm/1000 y
500	20.1
600	22.6
700	25.1
800	27.6
900	30.1
1000	32.6
1200	37.6
1500	45.1

was prepared using J. Ostromecki's map of mean annual sums of precipitation as its basis.

Apart from the calculations mentioned, the author undertook further calculations of correlation in order to determine indirectly the degree of mineralization of the karst waters ( $\Delta T$ ). Knowing this parameter would enable the calculation of denudation by means of formulae (3) and (4), because unit flow has been established for Poland by the State Hydro-Meteorological Institute. The result of these calculations gave

$$\Delta T = f(t^{\circ}\text{C})$$

for  $r = 0.902$

The true value of the coefficient of correlation on the 0.1% level is

$$r \geq 0.75$$

This interrelation is best approximated by the logarithmic function:

$$\log_{10} \Delta T = 0.0621t^{\circ}\text{C} + 1.9366 \quad (11)$$

The graphic image of this function is shown in Fig. 4. Using (11), karst denudation was calculated for particular regions  $q$  occurring in Poland, after the dependence:

$$D = f(t^{\circ}\text{C}, q)$$

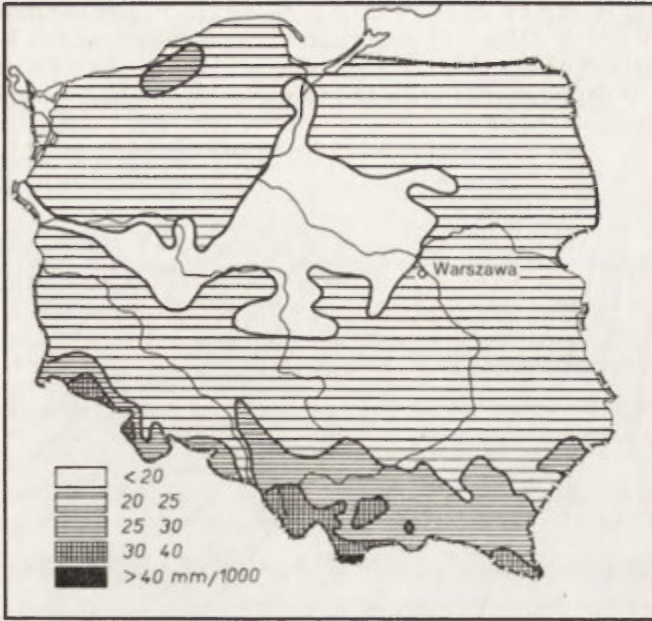


Fig. 3. Potential karst denudation in Poland calculated by the use of formula [8]  
 $D = 0.025 O_p + 7.6$

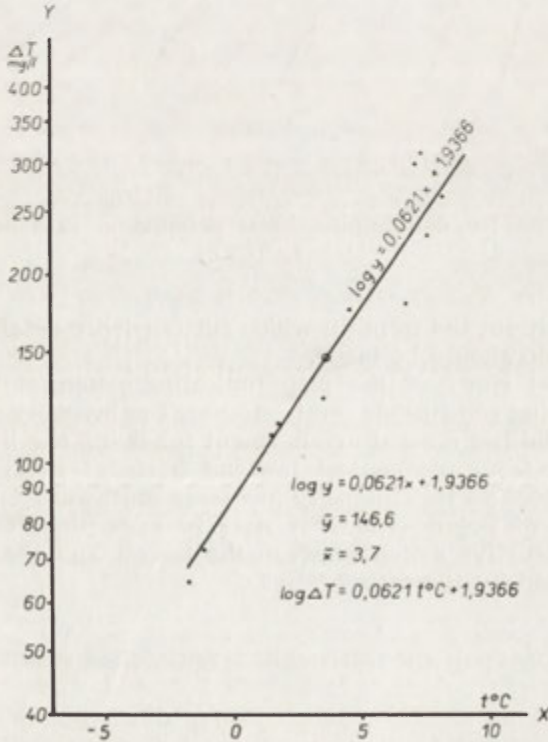


Fig. 4. Dependence of chemical composition of karst waters ( $\Delta T = Ca^{++} + Mg^{++}$ ) on mean annual air temperature ( $t^\circ C$ ) for twenty karst areas in Poland;  
 $\Delta T = f(t^\circ C)$

The graphic image of this function is shown in Fig. 5 — which is a nomogramme from which values of potential karst denudation can be read.

The comparison of the two methods (8) and (11) of indirect determination of potential karst denudation in Poland showed considerable conformity with the direct measurements made by the author. Hence it seems, that the results

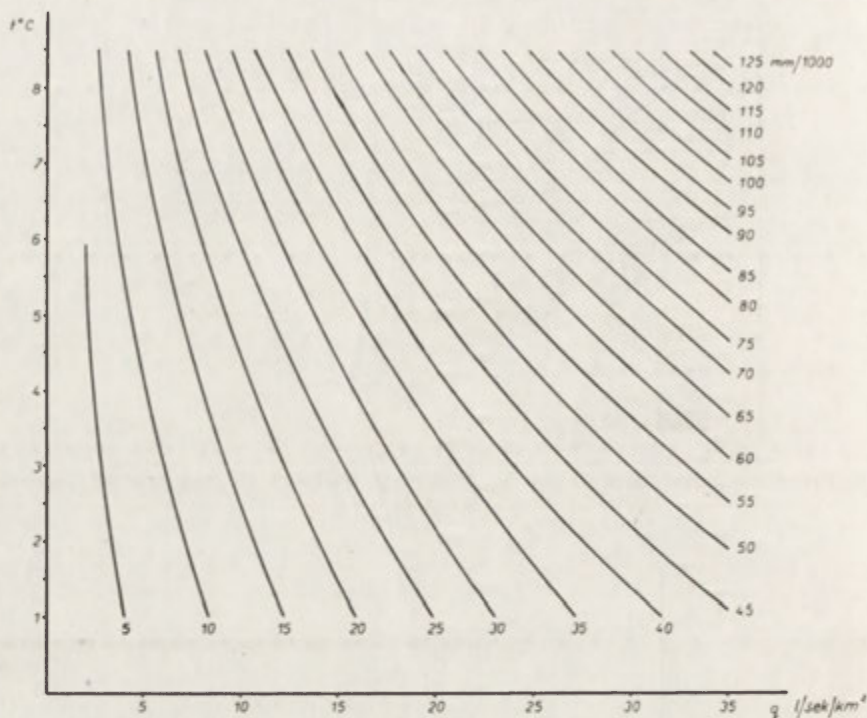


Fig. 5. Nomogramme for determining karst denudation in Poland after relation  
 $D = f(t^{\circ} C, q)$

thus obtained point out the trend in which further more detailed investigations of karst denudation should be made.

From a critical study of the map indicating potential karst denudation (Fig. 3) the following conclusions can be drawn. The intensity of karst processes occurring in Poland has a zonal arrangement following the line of parallels of latitude. Starting from the central lowland parts of Poland (where it has values of 20 mm/1000 y.) this intensity increases southward towards the higher and highest part of Poland where it rises to more than 40 mm/1000 y. The difference in destructive action between the lowest and the highest areas in Poland is expressed by the ratio of 1 : 3.

#### CHEMICAL DENUDATION OF NON-KARST REGIONS OF POLAND

Up to now nothing has been done to solve this problem. The only papers published on this subject deal with the values of what is called the ion runoff for a few Polish rivers (M. Jaworska 1968, K. Oleksynowa 1970, A. Tłałka 1970).

The author has made an attempt to determine chemical denudation by two

different methods. The first method is based on direct measurements made by the application of the hydrometric method, the second on distinguishing classes of intensity of these processes.

Chemical denudation has been determined by the hydrometric method for 22 drainage basins and for other areas situated in Southern Poland (Fig. 1), and the result of this work is presented in Table 3, indicating a range from 4.3 mm to 36.3 mm/1000 y., with a mean value of 18 mm/1000 y.<sup>1</sup> It appears that

TABLE 3. Chemical denudation of some non-karst areas in Poland, determined by use of the hydrometric method (3)

$$D_m = 0.0126 \cdot \Delta T \cdot q.$$

No.*	Area examined	$D^{**}$ mm/1000 y	$\Delta T$ mg/l	$Q$ m <sup>3</sup> /s	$P$ km <sup>2</sup>	$q$ l/s/km <sup>2</sup>
10.	Śnieżnik Massif-Kleśnica	4.3	22	0.040	2.6	15.4
11.	Śnieżnik Massif-Kleśnica I	11.9	65	0.080	5.5	14.5
12.	Śnieżnik Massif-Kleśnica II	9.6	75	0.150	14.8	10.1
13.	Krowiarki-Romanowo	10.0	186	0.041	9.6	4.3
14.	Biała Łądecka-Żelazno	14.3	77	4.43	301	14.8
15.	Nysa Kłodzka-Byczeń	21.2	203	17.65	2124	8.3
16.	Stołowe Mts.	17.2	71-140			13.0
17.	Kaczawa Mts.-Jastrowiec	10.8***	214	0.0133	3.33	4.0
18.	Nysa Szalona Mała-Bolkowice	11.3	155	0.593	102.7	5.8
19.	Bóbr-Wojanów	18.9	135	5.82	526	11.1
20.	Bóbr-Pilchowice	23.7	135	16.75	1200	13.9
21.	Bóbr-Żagań	17.4	135	43.65	4242	10.2
22.	Osobłoga-Komorniki		165			
23.	Czarna Przemsza-Przeczycze	11.6	220	1.3	306	4.2
24.	Czarna Przemsza-Przeczycze	10.8	220	1.2	306	3.9
25.	Czarna Przemsza-Przeczycze	18.0	220	2.0	306	6.5
26.	Cracow-Częstochowa Plateau, northern part	22.7	190-210	6.9	763	9.0
27.	Warta: Kręciwilk-Poraj	19.9	190	1.68	203	8.3
28.	Wiercica-Przyrów	24.9	210	1.95	207	9.4
29.	Soła-Porąbka	25.8	100	17.7	1082	16.4
30.	Beskid Wyspowy-Śnieżnica	18.6	80-125	0.132	8.5	15.6
31.	Dunajec-Krościenko	22.2	125	22.5	1598	14.1
32.	Dunajec-Rożnów	17.9	125	55.7	4883	11.4
33.	San-Dwernik	33.7	125	8.85	414	21.4
34.	Western Tatras-Kościeliska. Kiry	36.3	90		46.3	32.0
35.	Western Tatras-Kościeliska. Kiry	22.2	50		17.1	32.0
36.	High Tatras-Białka	13.7	34	2.1	65	32.0

\* Order of numbering like in Fig. 1.

\*\* Chemical denudation calculated for all salts dissolved in water ( $\Sigma M$ ).

\*\*\* Denudation calculated for CaCO<sub>3</sub>.

<sup>1</sup> Chemical denudation has been calculated for the total sum of salts (the dry residue), not as in karst denudation for the carbonates and sulphates only.

crystalline rocks are the least denuded (4 to 15 mm) and the strongest areas built of heterogeneous rocks among which carbonates occur (10 to 36 mm). Intermediate values were obtained for flysch rocks and non-karst sedimentary rocks.

The statistically ordered series of values of denudation shows a normal pattern ( $n = 1.28 d$  with a mean = 18.0 and dispersion = 7.31) indicating that the sample had been well chosen. This encouraged the author to carry out further calculations of denudation, correlating them with the following corresponding climatic agencies: the sums of annual precipitation and the mean annual air temperatures. From these calculations he obtained very low coefficients of correlation on the 0.1% level which were below the threshold of actuality,  $r = 0.6$

$$\text{For } r_{DOP} = 0.191 - 0.306$$

This result was to be expected in view of the fact that the author had omitted from consideration the differences in the properties of particular rock types, and that denudation depends to a high degree on the type of rocks involved.

Under these circumstances the author had to decide upon a different way of determining chemical denudation in Poland. For this purpose he selected

TABLE 4. Indices of chemical denudation in Poland for areas composed of eight types of rocks  
 $D = 0.0126 \cdot \Delta T \cdot q \cdot \epsilon$

Type of rock substratum	$\Delta T_{\Sigma M}$	$\epsilon$ %	$D = 0.0126 \cdot \Delta T \cdot q$						in mm/1000 y	
			$q=2$	$q=5$	$q=10$	$q=15$	$q=20$	$q=25$	$q=30$	
Crystalline rocks	12-80	0.26-1.74 74	2.9	5.8	8.7	11.6	14.5	17.4		
Carpathian flysch	60-170	0.52-1.48 48	7.2	14.5	21.7	29.0	36.2			
Loose Quaternary rocks (excluding loesses) in Central and Northern Poland	100-160	0.77-1.23 23	3.3	8.2	16.4	24.6				
Paleo- and mesozoic sedimentary rocks (excluding carbonates)	100-160	0.77-1.23 23	8.2	16.4	24.6	32.8	41.0	49.0		
Carbonate rocks: total	100-330	0.47-1.53	13.5	27.0	40.3	54.0	67.5	81.0		
old-palaeozoic	170-310	0.71-1.29 29	15.1	30.2	45.3	60.4				
mesozoic (Uplands)	200-330	0.76-1.25 25	16.7	33.4	50.0					
mesozoic (Tatras)	100-280	0.53-1.47 47	12.0	24.0	36.0	48.0	60.0	72.0		
Loesses and marls	290-340	0.92-1.08 8	19.8	39.6						
Loesses with high CaCO <sub>3</sub> content	370-400	0.96-1.04 4	9.7	24.2	48.5					
Sulphate rocks	250-700	0.53-1.47 47	10.6	26.5	53.0					

Classes of chemical denudation: I < 10 mm/1000 y, II = 10-20, III = 20-30, IV = 30-40, V > 40 mm/1000y



the eight groups of rocks which occur most frequently in Poland, and determined for them the chemical composition of the waters arriving from these areas. In this work he made use of the abundant analytical material available at the Institute of Water Economy, at the laboratories of the Sanitary Service, among numerous publications (M. Bombówna 1960, 1962, 1968, M. Bombówna and S. Wróbel 1966, K. Oleksynowa and T. Komornicki 1965, K. Pasternak 1964, 1967, 1968, M. Stangenberg 1958, S. Wróbel 1964), and among some material of his own. For his calculations he chose intervals of the most frequently occurring values of mineralization (the dry residue) from which he deduced the value of mineralization of the atmospheric waters ( $\Delta T = T - T_a$ ). This mineralization of the precipitated water he defined by using formulae (3) and (4), and the values for  $q$  he assumed in seven intervals ranging from 2 to 30 l/s/km<sup>2</sup>, so as to include all areas of Poland. In Table 4 the author presents, in mean figures, the results of these calculations. The interval of variance of denudation for each rock groups is signified by the coefficient  $\Sigma$ . The sequence of the particular rock groups in Table 4 has been arranged in accordance with the degree of susceptibility of the rocks to processes of chemical denudation.

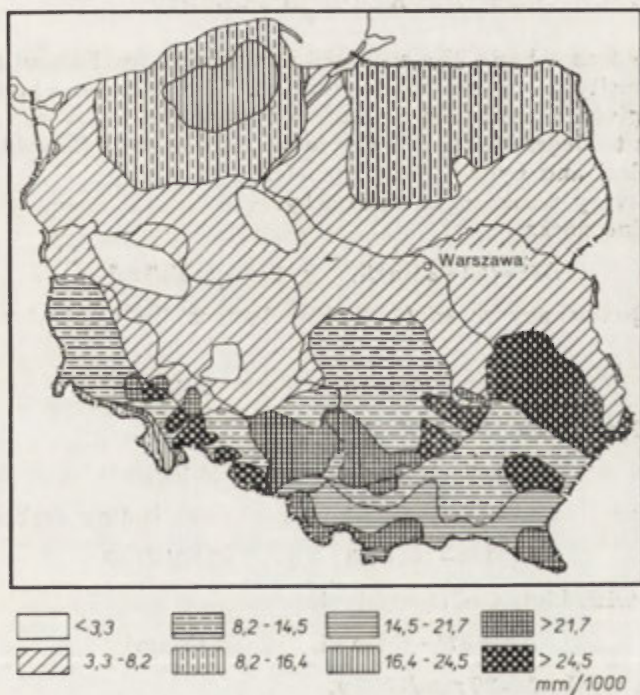


Fig. 6. Chemical denudation in Poland, in mm/1000 y.

Most resistant are crystalline rocks and the Carpathian flysch; least resistant are sulphate rocks, loesses and carbonates.

From his analyses, the author obtained a distribution of chemical denudation in Poland (Fig. 6), based on a map of unit flow (in l/s/km<sup>2</sup>) compiled by Z. Mikulski (1963) and on drift maps and geomorphological maps. Due to the absence of a more detailed map of unit flow for Poland, the pictures presented by the author must be considered as mere approximations.

Despite these manifest shortcomings, the following conclusions may be drawn from the map shown in Fig. 6 with regard to actual chemical denudation in Poland. Least degraded is a lowland belt separated in the north and south by zones of more intensive degradation. In Southern Poland the uplands are the most intensively attacked by chemical degradation, especially those situated in areas of loesses and gypsum or carbonate rocks. A zone of low values separates this area from the next strongly denuded area composed of flysch rocks and of mesozoic sedimentary rocks. An exception are the High Tatras which consist of crystalline rocks and suffer little denudation. The relatively low values of denudation observed for the Sudetes are worthy of attention, where maximum values were determined for a belt of sedimentary rocks, like the Nysa zone where denudation reaches up to 25 mm/1000 y. Obviously it is the small marble outcrops in these mountains which are denuded most strongly, as was mentioned in the preceding chapter. However, in view of the small space these outcrops occupy, they fail to markedly affect the general balance of denudation.

#### CLASSES OF CHEMICAL DENUDATION IN POLAND

Using Table 4 as a basis, the author distinguished for Poland five different classes of intensity of chemical denudation. The factors which decided to which class a given area should be assigned were the following: (1°) the type of rock subject to degradation, (2°) the magnitude of unit flow in l/s/km<sup>2</sup>, and (3°) morphological and hydrological agencies.

I. Denudation very weak: <10 mm/1000 y.

(1) Crystalline rocks:

$$\Delta T^2 = 12-80 \text{ mg/l}, \quad q < 15-20 \text{ l/s/km}^2$$

(2) Loose Quaternary rocks (excluding loesses) in Central and Southern Poland:

$$\Delta T = 100-160 \text{ mg/l}, \quad q < 5-10 \text{ l/s/km}^2$$

(3) Carpathian flysch

$$\Delta T = 60-170 \text{ mg/l}, \quad q < 5-10 \text{ l/s/km}^2$$

(4) Palaeo- and mesozoic sedimentary rocks (excluding carbonates):

$$\Delta T = 100-160 \text{ mg/l}, \quad q < 5-10 \text{ l/s/km}^2$$

(5) Loesses with high CaCO<sub>3</sub> content:

$$\Delta T = 370-400 \text{ mg/l}, \quad q < 2 \text{ l/s/km}^2$$

II. Denudation weak: 10-20 mm/1000 y.

(1) Crystalline rocks:

$$q = 20-35 \text{ l/s/km}^2$$

(2) Carbonate rocks:

$$\Delta T = 100-330 \text{ mg/l}, \quad 10 > q > 2 \text{ l/s/km}^2$$

<sup>2</sup> The intervals of values of  $\Delta T$  for each group are identical, no matter to what class they belong. This is why they are given only once. An exception is made for carbonate rocks.

## (3) Carpathian flysch:

$$15 > q > 5 \text{ l/s/km}^2$$

## (4) Loose Quaternary rocks (excluding loesses):

$$15 > q > 5 \text{ l/s/km}^2$$

## (5) Palaeo- and mesozoic sedimentary rocks (excluding carbonates):

$$15 > q > 5 \text{ l/s/km}^2$$

## III. Denudation moderate: 20–30 mm/1000 y.

## (1) Carpathian flysch:

$$15 > q > 5 \text{ l/s/km}^2$$

## (2) Carbonate rocks:

$$\Delta T = 100-330 \text{ mg/l}, \quad 15 > q > 5 \text{ l/s/km}^2$$

(3) Loesses with high  $\text{CaCO}_3$  content:

$$10 > q > 2 \text{ l/s/km}^2$$

## (4) Loesses and marls:

$$\Delta T = 290-340 \text{ mg/l}, \quad q = 5-10 \text{ l/s/km}^2$$

## (5) Loose Quaternary rocks (excluding loesses):

$$q > 10-15 \text{ l/s/km}^2$$

## (6) Palaeo- and mesozoic sedimentary rocks (excluding carbonates):

$$20 > q > 10 \text{ l/s/km}^2$$

## (7) Sulphate rocks:

$$\Delta T - 250-700 \text{ mg/l}, \quad 10 > q > 2 \text{ l/s/km}^2$$

## IV. Denudation strong: 30–40 mm/1000 y.

## (1) Old palaeo- and mesozoic carbonate rocks (uplands):

$$\Delta T = 170-330 \text{ mg/l}, \quad 15 > q > 5 \text{ l/s/km}^2$$

## (2) Palaeo- and mesozoic sedimentary rocks (excluding carbonates):

$$20 > q > 15 \text{ l/s/km}^2$$

## (3) Carpathian flysch:

$$q > 20-25 \text{ l/s/km}^2$$

## (4) Loesses and marls:

$$10 \geq q > 5 \text{ l/s/km}^2$$

## (5) Sulphate rocks:

$$10 > q > 5 \text{ l/s/km}^2$$

V. Denudation very strong:  $>40$  mm/1000 y.

## (1) Carbonate rocks:

$$\Delta T = 100-330 \text{ mg/l}, \quad q \geq 15 \text{ l/s/km}^2$$

## (2) Palaeo- and mesozoic sedimentary rocks (excluding carbonates):

$$q \geq 25 \text{ l/s/km}^2$$

(3) Loesses with high  $\text{CaCO}_3$  content:

$$q > 1-10 \text{ l/s/km}^2$$

(4) Sulphate rocks:

$$q > 5-10 \text{ l/s/km}^2$$

#### COMPARISON OF CHEMICAL AND MECHANICAL DENUDATION

Attempts at a qualitative definition of mechanical denudation in Poland, by regions, have been undertaken by hydrologists (J. Brański 1968, K. Dębski 1959, W. Jarecki 1957, Z. Mikulski 1965, B. Wiśniewski 1969), pedologists (K. Figuła 1966, A. Reniger 1950) and geomorphologists (T. Gerlach 1966, A. Jahn 1968, 1968a, L. Starkel 1960, and others).

For about twenty years the State Hydro-Meteorological Institute has been making measurements of the amount of mineral material transported by surface streams (1947-1966 Annals). A detailed classification of the intensity of soil erosion in Poland compiled by A. Reniger (1950), and the above mentioned material collected by the State Hydro-Meteorological Institute have served as the basis for the preparation of tables and maps indicating mechanical denudation in Poland. This work was done by K. Dębski (1959), Z. Mikulski (1963) and B. Wiśniewski (1969) who chose the value  $t/\text{km}^2/\text{y.}$  as the unit of denudation. For purposes of comparison the present author transformed these values into  $\text{mm}/1000 \text{ y.}$ , considering  $1.56 \text{ g/cm}^3$  the average specific gravity. This weight comes near to that of the loose material occurring in Central and Northern Poland, while it is too low for the compact rock formations of Southern Poland. This is why the indices of mechanical denudation for Southern Poland are some 30 to 40% too high. A map illustrating this is given in Fig. 7.

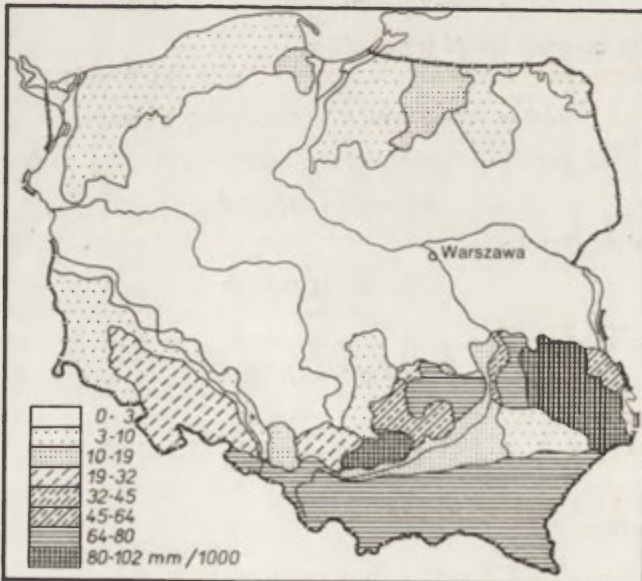


Fig. 7. Mechanical denudation in Poland, in  $\text{mm}/1000 \text{ y.}$  (after Dębski-Mikulski, partly modified)

The basic material for an analysis correlating chemical with mechanical denudation are the three maps shown in Figs. 3, 6 and 7. From this comparison the author draws the following conclusions:

(1) In the lowland of Central and Northern Poland the figures for chemical denudation are nearly similar, or higher than those for mechanical denudation. The predominance of chemical denudation in the area of the Pomeranian Lake District is worth noticing here.

(2) In Poland's upland and mountain areas mechanical denudation is higher than chemical denudation, except for karst regions where degradation mainly occurs by chemical processes.

(3) The intensity of mechanical and chemical denudation, considered qualitatively, is very similar, but the gradients of denudation differ. The differences between particular regions are greatest with regard to mechanical denudation.

(4) In the first place, both chemical and mechanical denudation depend on how far the rocks are susceptible to processes of denudation; processes caused by climatic conditions and morphological features take second place.

(5) In Poland the areas suffering greatest denudation are: the upland belt of Central Poland where degradation exceeds 100 mm/1000 y. (0.1 mm/y.) and the Carpathians where it is up to 90 mm/1000 y. Areas with the least intensive denudation are the central lowland area with 15–20 mm/1000 y. The differences between the areas in Poland where denudation is least and most intensive are expressed by the relation 1 : 7.

#### FINAL COMMENT

The quantitative studies of chemical denudation presented by the author in this paper are an attempt to applying methods which are based on direct measurements, and on a statistical analysis of the available data treated as a representative sample. The results thus obtained define the order of values of chemical denudation and illustrate the way this denudation is distributed in Poland. At the rate at which new data, especially chemical analyses of water, are being made available, it will be possible to formulate more accurately the destruction caused in areas differing in geological structure. For the most part this refers to the need for a detailed division into classes of chemical denudation. The author is well aware of the fact, that in the classification he is proposing the intervals are unduly wide, and that in his table indicating the resistance of rocks to chemical denudation (Table 4) the deviations of the coefficient  $\varepsilon$  are too great.

In spite of these shortcomings one can draw conclusions from the picture illustrating the intensity of denudation in Poland, and forecast the way the land relief may in time develop in regions of different types of rocks. The fact that in areas with carbonate rocks the sum of mechanical and chemical denudation is lower than in what are called the "more resistant" compact non-carbonate rocks in which mechanical denudation is much higher is of importance. This might appear to be contradicted by the high indices of chemical denudation of carbonate rocks which are uncomparably higher than in other sedimentary or crystalline rocks. However, it must be kept in mind that in karst regions mechanical denudation is practically negligible due to the lack of surface streams. In these facts can be seen the explanation of the "relatively high resistance" of carbonate rocks to denudation processes by which these rocks are being gradually denuded. The position occupied by the carbonate rocks in the table ranging the rocks by their degree of chemical denudation

is also interesting. Among eight types of rocks (Table 4) the carbonates occupy third place, after the sulphates and loesses. Also worthy of attention is the fact that for some of the carbonate rocks the values of denudation closely resemble those of the loose Quaternary rock material spread over Central and Northern Poland.

The comparison between chemical and mechanical denudation emphasizes the part played by chemical processes in the denudation balance of the morphological land surface. It was found, that chemical denudation is just as much, perhaps even more, responsible for ultimate degradation not only of soluble rocks but of non-karst rocks too.

In this paper the author attempted to select the analytical material used for calculating chemical denudation, so as to eliminate as much as possible any type of anthropogenic agencies such as man-made water pollution. Thus, the results of his research illustrate processes of denudation as they take place under natural conditions.

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## INVESTIGATIONS INTO PRESENT-DAY GEOMORPHOLOGICAL PROCESSES IN THE LOESS AREAS OF THE LUBLIN PLATEAU

STEFAN ZIEMNICKI AND JANINA REPELEWSKA-PEKALOWA

A considerable part of the Lublin Plateau is covered by a layer of loess deposits which have an average thickness of some 5 m. This loess is horizontally stratified and shows vertical cleavage. A characteristic feature of its mechanical composition is a high content of silty particles. On account of its physical properties this loess is readily subject to aqueous erosion because, when excessively humidified, it loses its permeability and changes into a plastic mass, easily eroded and transported.

A characteristic trait of the relief of loess-mantled areas is waviness and, in some zones, a dense network of erosive surface incisions. Particularly noticeable is the diversity of the land relief near the deeply incised river valleys. Here the slope angle oscillates between 10 and 20% and sometimes it is even steeper than 30%. On the whole, the fluvial system is scanty, but numerous dry valleys have only seasonal water flow.

The climate dominating the Lublin Plateau does not increase erosive tendencies. The mean annual precipitation totals about 600 mm, and a considerable part of it falls in the summer months (June, July, August), i.e., in the period of full vegetation. Heavy rains which occur only about 8 times a year surpass 20 mm daily. The mean annual air temperature is +8°C; in January it averages -3°C, in July +17°C.

The climate is favourable both to the growth of perennial vegetation such as trees and grasses, and to that of annual agricultural production. The soil ranks among the best in Poland, and this is why nearly all the Lublin Plateau is utilized for agriculture. Forests occupy only 11% of the land, while perennial grass lands occur only on the floors of fluvial valleys and cover about 8% of the area of the Plateau. For more than 500 years, agriculture has flourished here, but large-scale deforestation and a vast increase in the cultivated area took place only in relatively recent times, in the latter half of the 19th century. Among the farmsteads, holdings of less than 5 ha each predominate, and the size of individual fields is rarely more than 0.5 ha. Access to fields is mostly by dirt roads which often run parallel with the slope; due to water runoff and deflation of loess particles, this kind of road is easily subject to erosion, and heavy rains convert such roads into canyon-like gaps.

Each year brings two periods in which erosion threatens: snowmelt time and the period of heavy summer showers. The former occur mostly in February and less frequently in March; the latter in June and July, rarely in August. At the time snow is melting, the soil lacks a vegetation cover and surface erosion and soil incisions are a common feature; sometimes road gaps are

heavily eroded. Cases of heavy rainfall, on the other hand, coincide with the time when crops are in full growth. This is why slopewash rarely occurs -- only if a field happens to be bare of plants. As a rule, soil erosion takes place where water runs off, mainly in road cuts and over surface roads.

Detailed examinations of the actual geomorphological processes are being made near Lublin at two stations: Sławin and Elizówka. Apart from observations of the erosion process, measures obstructing slopewash were undertaken on selected slopes. In these tests, particular attention was paid to modern relief-forming processes on cultivated land, and to the effect obtained from preventive measures, considering their effectiveness to be some sort of a standard index for the intensity of erosion.

Experimental work was started at Sławin in 1948 and at Elizówka in 1957. The tests consisted of measurements of the quantities of water and soil removed from small drainage basins of the order of 5 km<sup>2</sup> each, recording the forms developed by this runoff, and observing the changes occurring in the soil on field surfaces of some 20 ha each.

## SŁAWIN

Sławin lies in the drainage basin of the river Czechowka, a left-bank tributary of the Bystrzyca River. The regions set aside for observations and experiments are periodically traversed by water escaping from a drainage area of 4.75 km<sup>2</sup> size. The surface of this basin has an undulating relief, with altitude differences of up to 30 m. The slope inclination differs: the steepest (some 17° or 30°/0) are the slopes exposed to south and west, while the slopes open to north and east are more gentle (6° or 10°/0).

For their detailed experiments the authors singled out an area of 13 ha of cultivated land. Here the principal elements of the land relief are slopes differing as to exposure and inclination, and a valley floor (Fig. 1). In 1948, before antierosive measures were applied, one of the slopes with a western exposure had already been virtually stripped of its humus cover; on another slope it was found that the thickness of this cover was less than it averaged on level upland areas. All slopes were furrowed by gullies incised by water from melting snow and from rainfall. In the floor of the valley an eroded section was observed, some 2 m wide and up to 1 m deep, filled with a limestone debris. This erosion indicates how powerful the water runoff must have been. On the slopes, proof of erosion is the thinness of the humus cover in the upper slope part and its much greater thickness at the slope base and on the valley floor (Fig. 2); here at some points the humus layer deposited had thicknesses up to 70 cm.

Anti-erosive measures were undertaken for the first time in 1948; they consisted of safeguarding the valley floor by turf, terracing the steep slopes by benches, and changing the arrangement of fields to such an extent that all principal acts of cultivation, i.e. ploughing and sowing, would be done in a direction perpendicular to the slope (Fig. 3). This alteration was accompanied by certain changes in the crop composition (more species sheltering the soil). All the new measures proved to be a success. The new field arrangement and the introduction of the revolving plough share, depositing the furrows downhill, automatically led to the gradual formation. During the first years after the changes, the width of the slope benches used to grow up to 30 cm per year, though this annual increase grew less in later years. Twenty years after the new measures were put into effect, the benches ran at some 200 cm

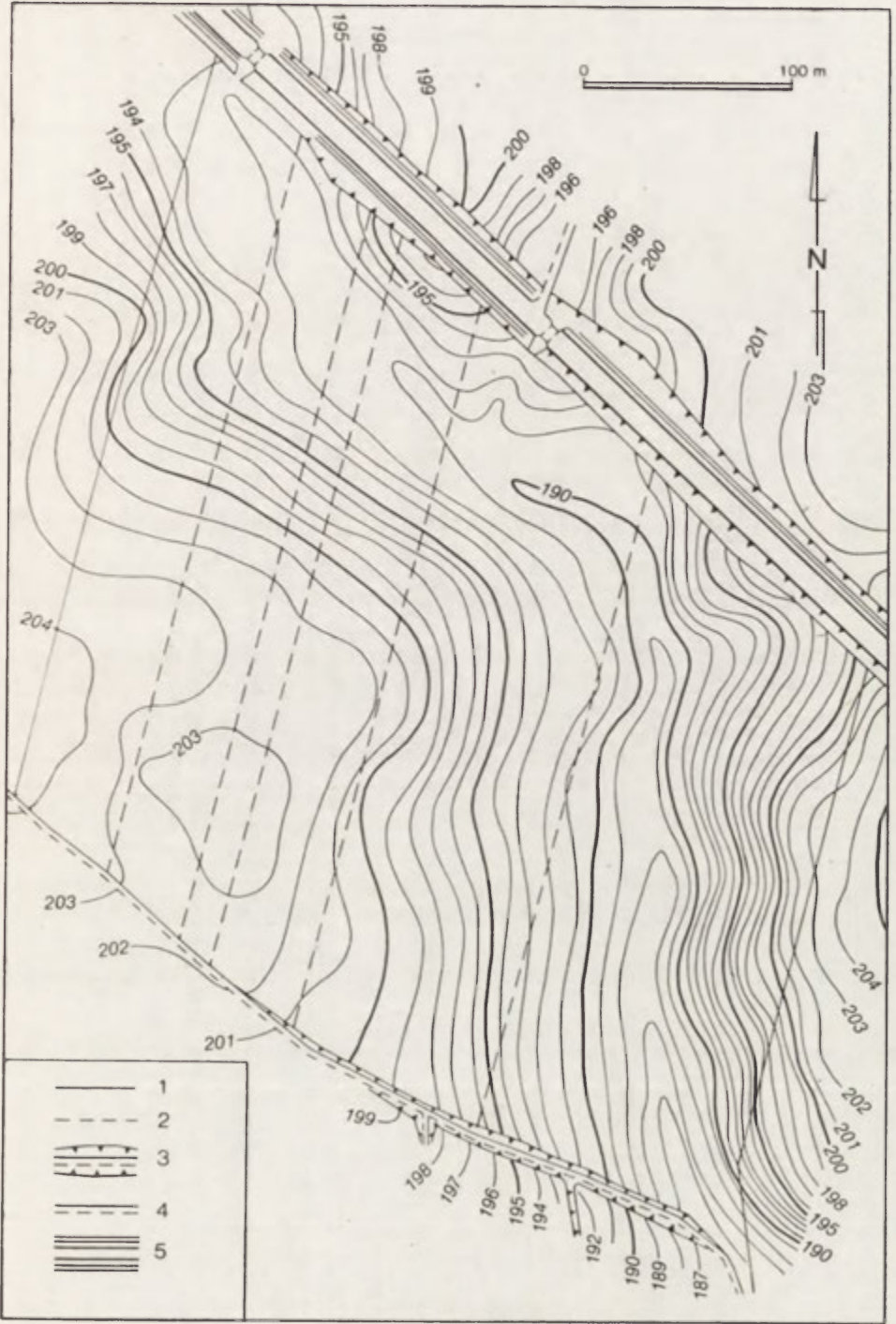


Fig. 1. Slawin Station, as it was in 1948

1 — Station boundary, 2 — boundary of crop-rotation fields, 3 — dirt road gap, 4 — dirt road, 5 — highway

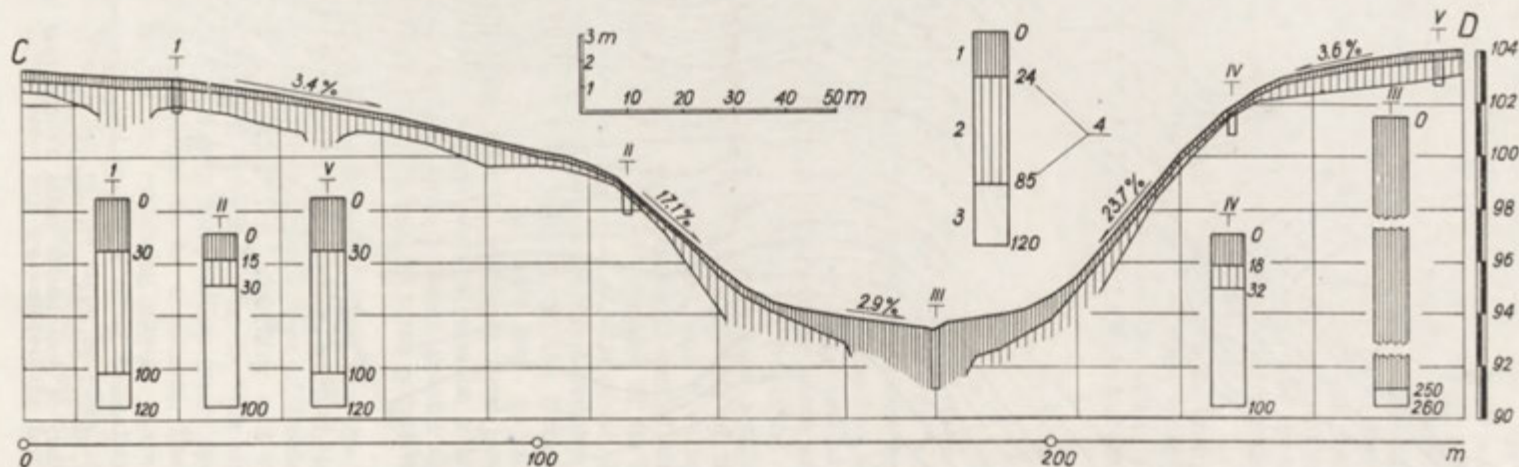


Fig. 2. Cross-section C-D, typical of loess slopes on the Lublin Plateau; its situation is shown in Fig. 5

1 — top soil layer, 2 — transition layer, 3 — substratum rich in calcium carbonate, 4 — depth of particular layers, in cm

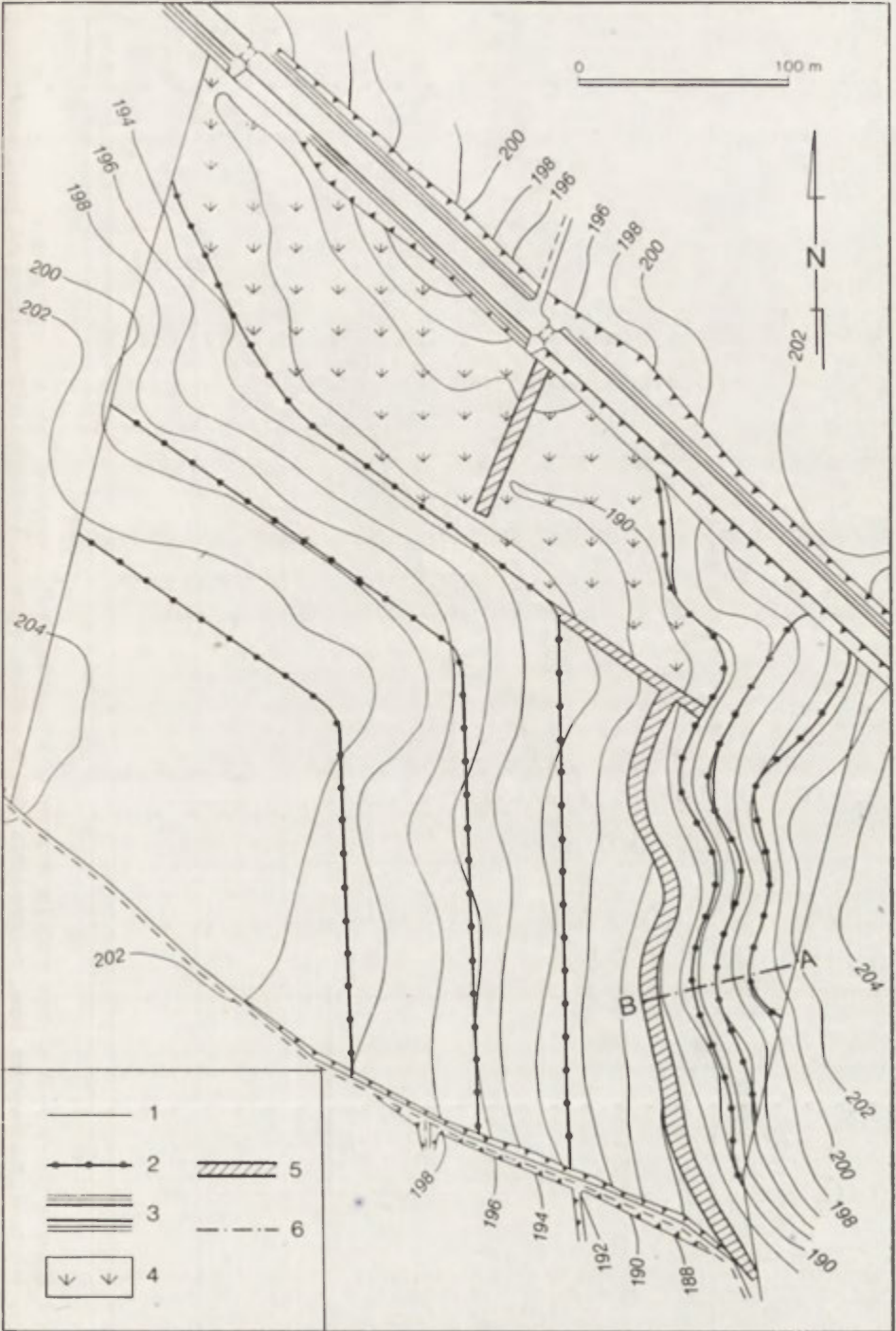


Fig. 3. Sławin Station, after the introduction of a field arrangement adjusted to land relief (after Ziernicki)

1 — Station boundary, 2 — field boundaries, 3 — highway, 4 — field sown with lucerne, 5 — turf-covered belts, 6 — situation of cross-section A — B is shown in Fig. 4

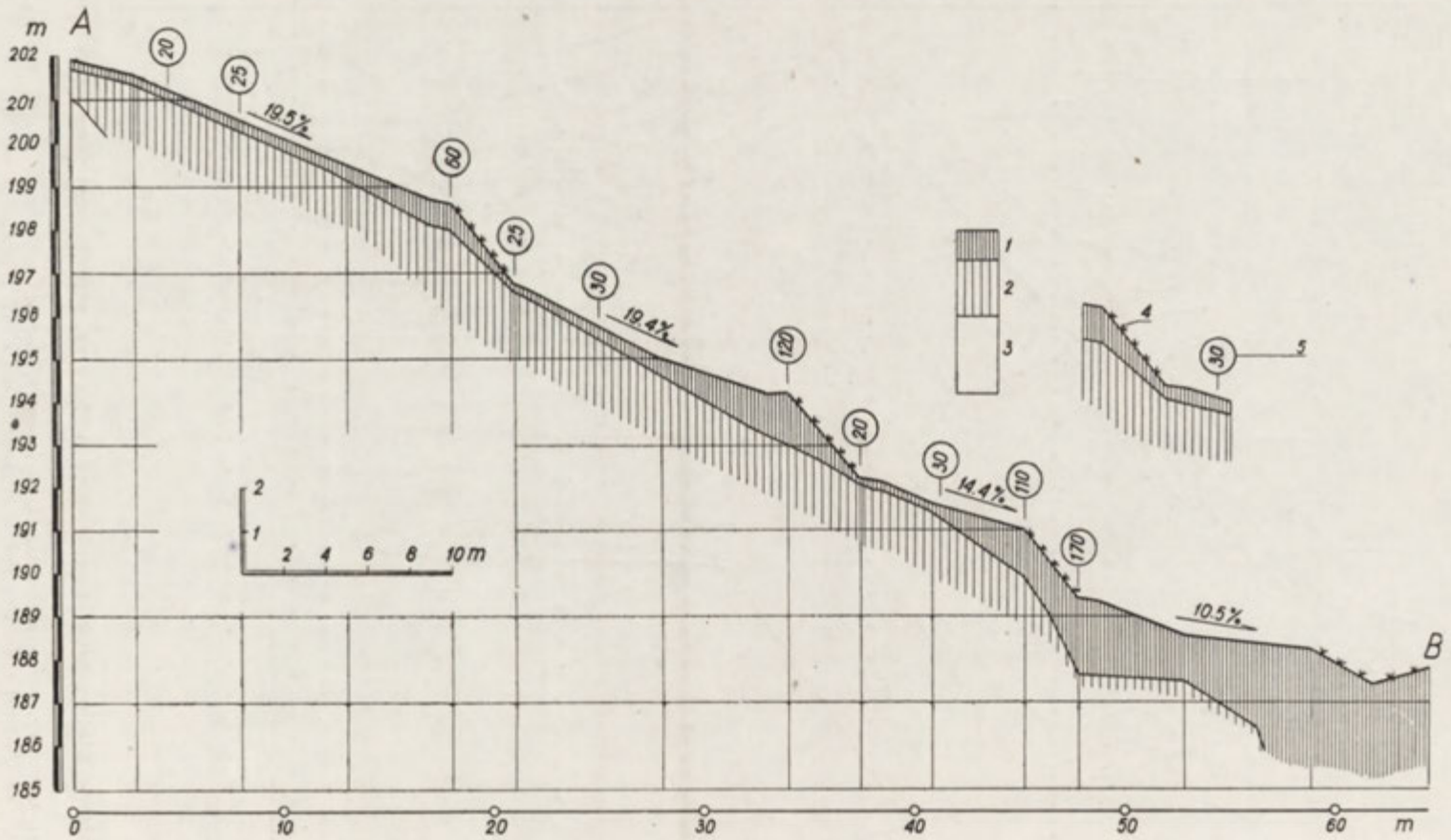


Fig. 4. Cross-section A — B of slope with its terraces which automatically developed over a 20-year period  
 1 — top soil layer, 2 — transition layer, 3 — substratum rich in calcium carbonate, 4 — turf, 5 — thickness of top soil layer, in cm

height intervals, and the steepness of the slopes was reduced from 30 to 20%. The thickness of the humus layer had grown, increasing upward from each bench but remaining unaltered downward from the benches (Fig. 4).

Ploughing with the revolving share was done for only three years after the authors started their measures; afterwards a normal plough was again used. The way in which the height of the benches was increasing was also evidence of an actual downward soil displacement. In fact, during the 20-year period a mean soil volume of 3 m<sup>3</sup> was moved an average distance of 10 m; for a slope 50 m long this means a displacement of 150 m<sup>3</sup> for a slope of 1 m width. Converted to one year's effect this soil movement involves 0.75 m<sup>3</sup>. It should also be mentioned that it was not only slopewash material which was retained in this way, but also much soil was shifted by the plough share and by other agricultural machinery.

Near the locality discussed, a different research method was applied on a slope exposed to the south and having a 20 to 30% inclination. This slope area was divided into three sections each having a different field pattern: in two sections the border lines between fields ran perpendicular to the slope, while in the third section they ran parallel with the slope "straight down". The third section was to serve as some sort of control for appraising the adjoining better cultivated sections. From 1949 to 1963, the authors observed

TABLE 1. Runoff of water and soil (in mm) from a drainage basin of 4.75 km<sup>2</sup> (at Slawin)\*

Year	Water and spring runoff		Summer and autumn runoff		Total	
	water	soil	water	soil	water	soil
1950	27.0	0.12	0.9	+	27.8	0.12
1951	3.3	0.02	0.1	+	3.4	0.02
1952	5.2	0.02	—	—	5.2	0.02
1953	2.0	0.01	+	+	2.0	0.01
1954	12.0	0.02	—	—	12.0	0.02
1955	0.15	+	—	—	0.15	+
1956	55.0	0.17	9.2	0.07	64.2	0.24
1957	5.5	+	5.0	+	10.5	+
1958	+	+	—	—	+	+
1959	—	—	—	—	—	—
1960	—	—	+	+	+	+
1961	1.8	+	—	—	1.8	+
1962	2.0	+	0.7	+	2.7	+
1963	7.0	0.01	—	—	7.0	0.01

\* Gater runoff of less than 0.1 mm, and soil losses of less than 0.01 mm are marked by +

the rate of bench growth, the change occurring in the soils, and the effect of these changes upon crop results. They found that the section lacking the protective mode of contour ploughing was being continuously incised by gullies and that the soils suffered degradation by water flow. They calculated that in Section 1 and 2 soil displacement averaged 66.3 m<sup>3</sup> per year per hectare, and that only 3.2 m<sup>3</sup> of soil was flushed away beyond the slope; this would correspond to a soil layer 0.3 mm thick. During the same time the soil

displacement in the Section 3 amounted to 90.2 m<sup>3</sup>; of this volume 29.2 m<sup>3</sup> were deposited at the base of the slope while as much as 61.0 m<sup>3</sup> were flushed off. Converted into soil thickness this value equals some 6.5 mm. Hence it appears, that a loess slope of some 20% inclination lacking a protective mode of cultivation would each year suffer surface denudation of 5 mm.

Measurements of how much water and soil was flushed off from the slopes were made at Sławin for a drainage basin of 4.75 km<sup>2</sup>. Except for 10% covered by forests, the area was fully cultivated. No protective measures against slopewash were taken, and the field borders ran at random as they do practically all over the Lublin Plateau. Table 1 records the quantities of water flow and slopewash for the time from 1948 to 1966. The flow is given in mm; 1 mm of water flow equals 1000 m<sup>3</sup>/km<sup>2</sup>, and 1 mm of slopewash — 140 t soil from 1 km<sup>2</sup>.

### ELIZOWKA

Here the land relief is diversified (Fig. 5), because five small valleys converge further down, forming one valley. The region under investigation is crossed by water from a drainage basin of 6.22 km<sup>2</sup> area, while the small valleys each have an area of the order of 1 to 2 km<sup>2</sup>. The slopes, from 30 to 60 m long, are inclined at 15 to 20% (Fig. 2). Water flow from rain and snowmelt has incised the slopes to a marked degree, and evidence of slopewash and denudation material accumulated on the valley floor is clearly seen.

Anti-erosion improvements (Fig. 6) were undertaken here in 1958, covering an area of 41 ha. The intention was to obstruct the escape of water and soil which is important in view of periodical rain shortage and excessive soil desiccation, mainly on the slopes exposed to S and W. For this purpose the direction of ploughing was changed into horizontal direction, and a suitable crop rotation was arranged creating band-shaped fields. At intervals of some 50 m slight depressions were being formed in the slopes, and these gradually developed into small valleys. In order to set limits to these forms small earth dykes were put up at the field boundaries (Fig. 7). The valley floor was stabilized by turf growth. For purposes of correlation, the previous direction of ploughing in a downslope direction was retained, and no other improvements were made.

Along the boundaries where the fields had been improved, longitudinal slanting benches were developing which changed the slope gradient locally and contributed to creating differences in the thickness of the top soil. The mean annual growth of these benches was in the period:

from 1958 to 1960	20 cm
from 1960 to 1962	15 cm
from 1962 to 1966	5 cm

From the benches upward the thickness of the top soil layer grew some 20 cm. This mode of improvement presented the flushing away of soil from the depressions in the slopes; on the average these depressions deepened by some 10 cm each year.

Calculations revealed that after these improvements had been made, the mean annual denudation of the slopes was of the order of 20 t soil from 1 ha — the equivalent of a soil layer 1.5 mm thick. This is four times less than has been determined for the unimproved loess slope at Sławin.

As at Sławin, after every heavier water flow the gullies and soil accumulations were measured, and the results observed for improved and unimproved



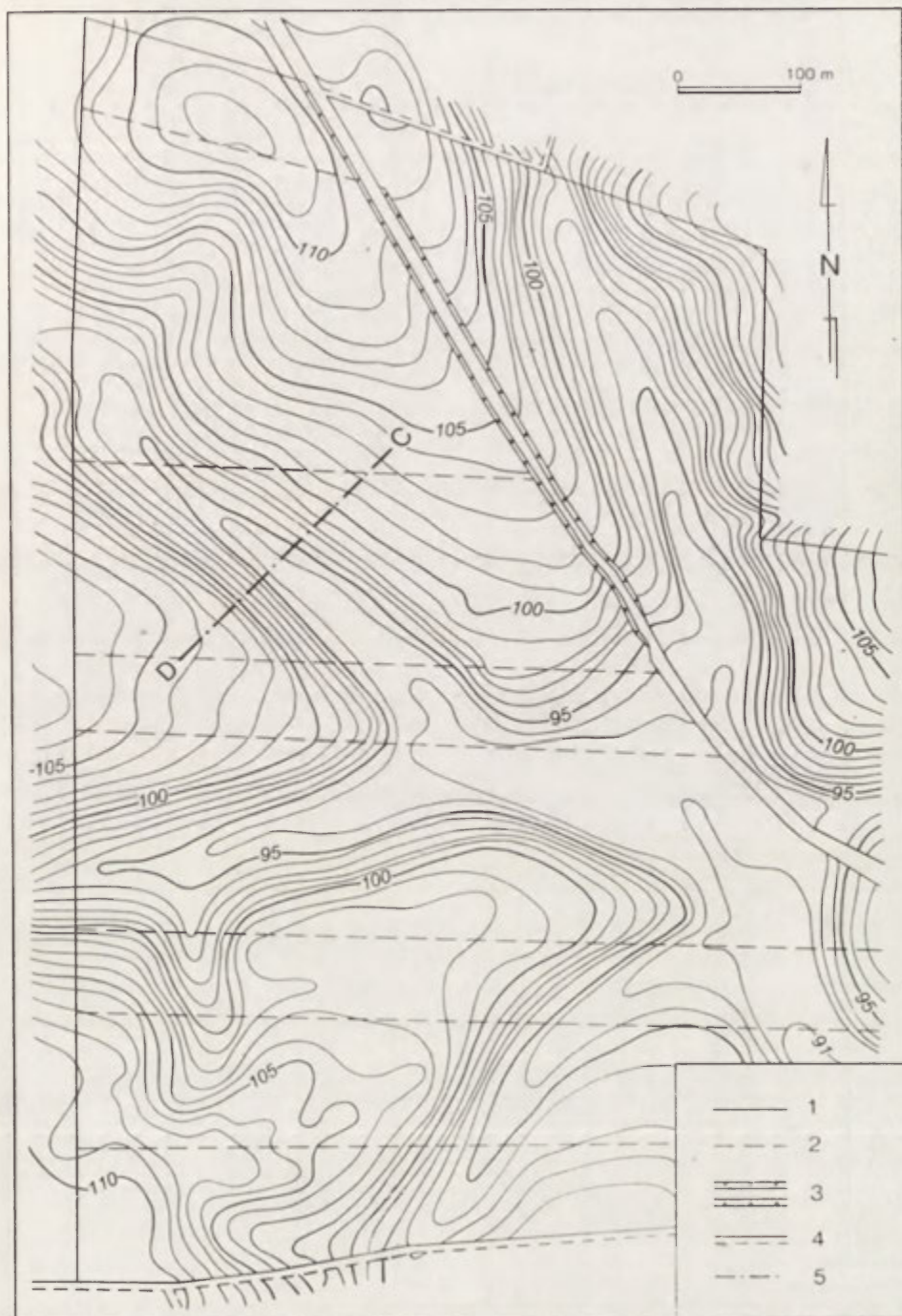


Fig. 5. Elizowka Station, as it was in 1957

1 — Station boundary, 2 — boundaries of crop-rotation fields, 3 — dirt road gap, 4 — dirt road, 5 — cross-section is shown in Fig. 2

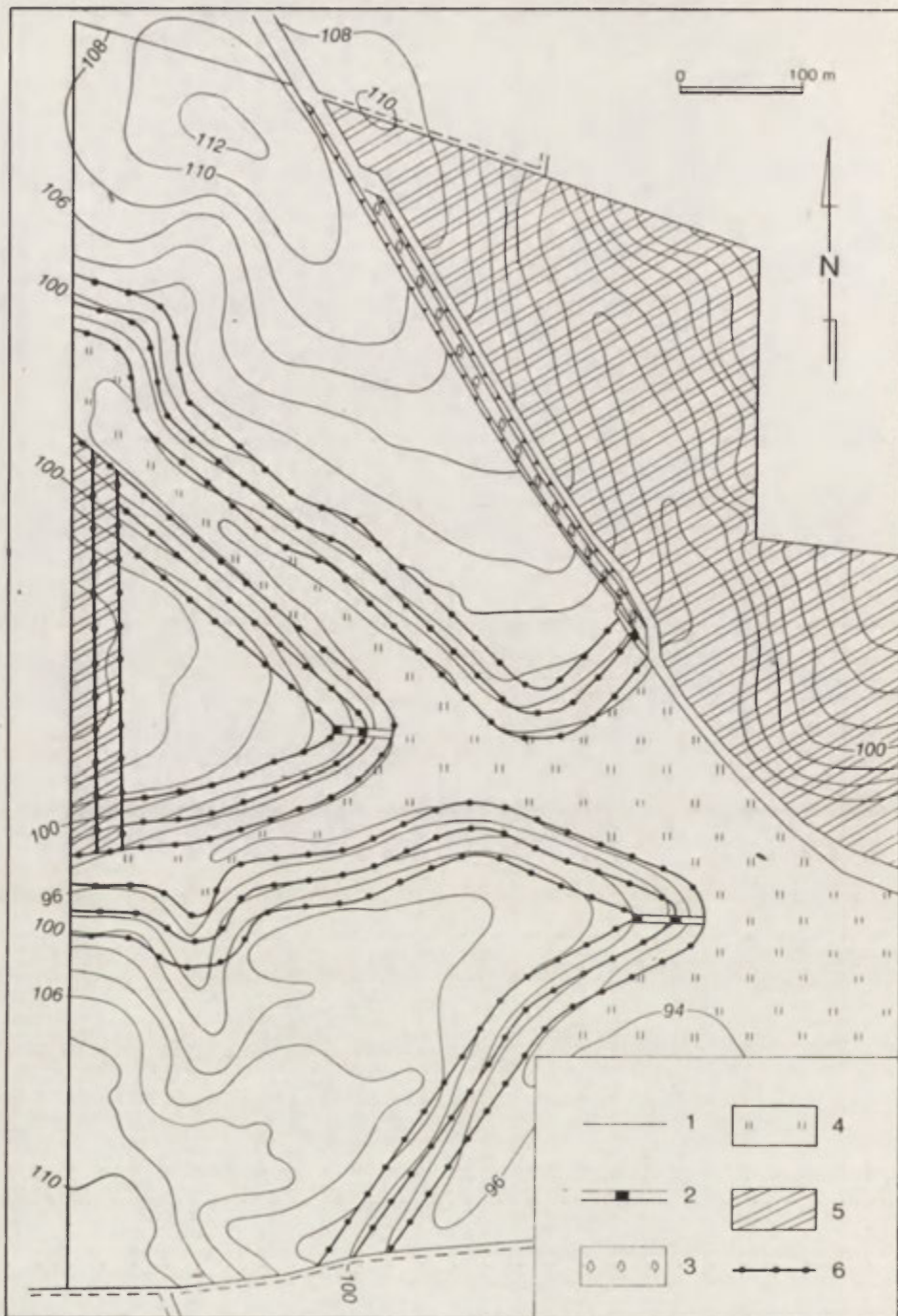


Fig. 6. Elizówka Station after the introduction of a field arrangement adjusted to land relief (after Ziemnicki)

1—Station boundary, 2—dirt roads leading to band-shaped fields, improved by pavement where they cross benches, 3—forest growth in the gap of a former incised dirt road, 4—perennial grassland (turf), 5—control fields lacking anti-erosive improvements, 6—boundaries of crop-rotation fields

TABLE 1. Runoff of water and soil (in mm) from a drainage basin of 6.22 km<sup>2</sup> (at Elizowka)

Year	Winter and spring		Summer runoff		Total	
	water	soil	water	soil	water	soil
1956	60.2	0.11	3.5	+	63.7	0.11
1957	5.6	+	0.4	+	6.0	+
1958	4.4	+	—	—	4.4	+
1959	—	—	—	—	—	—
1960	1.2	+	0.3	+	1.5	+
1961	+	+	—	—	+	+
1962	2.9	+	+	+	2.9	+
1963	9.4	+	—	—	9.4	+
1964	73.8	0.04	—	—	73.8	0.04
1965	2.4	+	—	—	2.4	+
1966	1.3	+	—	—	1.3	+

Water runoff of less than 0.1 mm, and soil losses of less than 0.01 mm are marked by +

slopes were correlated. From time to time altitude measurements by levelling instruments were made for fixed sections across slopes, and the thickness of characteristic soil layers were determined.

Furthermore, at Elizowka measurements were also made of the quantities of water and soil which were carried off from a drainage basin of 6.22 km<sup>2</sup> — this area is 100% cultivated. The runoff values are recorded in Table 2, in which the same symbols are used as in Table 1.



Fig. 7. Fragment of band-shaped fields at Elizowka. Visible are small dykes put up in the lateral valley to obstruct water runoff and retain washed-down soil. Photo by S. Ziemnicki

## GENERAL COMMENT

The long-term tests made at the experimental stations mentioned as well as random observations made at a number of further points on the Lublin Plateau suggest the following comments.

The quantities of soil removed from the smaller drainage basins of some 5 km<sup>2</sup> area varies very much for different years, from 0 to 336 t from 1 km<sup>2</sup>. During the periods of observation, none of the stations mentioned experienced catastrophic rains, as occurred in other parts of the Plateau.

During the snowmelt, as during heavy rainfall, the escaping water carries off part of the denuded soil, especially its finest particles, while a further part of the soil, the slightly larger particles, are deposited at the base of slopes due to the reduced rate of flow. The authors did not establish the ratio of soil removed by the water to the amount of soil deposited; they only found that this ratio varies depending upon how much water flow was involved. When slopewash was powerful, only a small part of the eroded soil remains inside the drainage basin; for slight slopewash, conditions may be the opposite. This is particularly easy to observe at sites where a dense turf covers the valley floor.

The effect of surface erosion and the formation of minor-size gullies are processes of slopewash, of the order of 5 mm per year. The agricultural implements used for land cultivation contribute to a high degree to this downward displacement of soil. This mainly refers to the plough which turns the earth furrows downslope, or the harrow which moves the loosened earth clods and causes them to topple downslope.

For all ground levelling done during cultivation, the deeper gullies which develop at the same place after every heavy water flow lead to the formation of characteristic depressions in the slope surfaces and, ultimately, to new small valleys which after some time develop into dry valleys. A great number of dry valleys of this type dissect the loesses of the Lublin Plateau. In the cases where a field lacks a protective plant cover, a young but rapidly developing small valley of this kind, usually slanting at some 15‰, grows in depth by some 10 cm annually; the drainage area being very small to start with, usually no more than 0.1 ha.

Catastrophic rain showers occur largely in the latter part of June and in early July; but at this time of the year flourishing vegetation makes slope denudation negligible. On the other hand, on such occasions any insufficiently plantcovered zones in which massed rainwater can rush off are strongly eroded. Dirt roads and road incisions especially are exposed to erosion (Fig. 8). In one instance after one heavy rainfall it was found that a dirt road gap increased by more than 100 m in length.

Protective measures such as tillage perpendicular to the slope, band-shaped fields, the automatic formation of terrace-like benches—all brought results almost immediately in the form of increasing yields. Obviously this was not the only effect of soil retention but mainly—as was found—the effect of the retention of water in the soil. And, provided proper fertilizer is applied, this way of increasing the water content of the soil on a slope is sufficient to make crops on slopes barely 10 to 15‰ lower than on the level expanse of the Plateau. It is worth bearing in mind that, as long as all ploughing was done “straight down the slope” and especially on fields of south facing slopes and after a rain shortage in the spring, the crops gathered on slopes used to be 30 to 50‰ of those from the level fields on the Plateau.

A method of establishing soil displacement on a slope are measures which obstruct this sort of movement. As a rule it is sufficient to lay down a horizontal turf belt at angles to the slope, or merely a horizontal fascine roll, to retain a considerable amount of soil. By using a reversible plough-share the ground surface is promptly raised on the upper side of grass-covered field borders, and in this way the slope gradually changed automatically into a set of mutually superimposed terrace benches.



Fig. 8. Dirt road gap, eroded during heavy rainfall; situated in the loess area of the Lublin Plateau. Photo by S. Ziemnicki

Damage due to water erosion includes soil diminution on slopes; reduced water resources retained in slope soils; rapid drying out of soil cover especially of those exposed to the S and W; the formation and rapid growth of new small valleys. Admittedly, the accumulation of part of the denuded soil at the bases of slope and on valley floors is a favourable feature. But, the benefit resulting from the smoothing of previously undulating surfaces is incomparably slight in the light of the soil losses suffered and the rapid evolution of a new diversified land relief.

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## PRESENT-DAY WATER AND WIND EROSION OF SOILS IN NE POLAND

MICHAŁ SKRODZKI

The surface relief of NE Poland is unusually diversified. It contains a wide variety of hill forms randomly scattered, and separated by numerous basins and by long chains of fluvial valleys and glacial channel lakes. This abundantly diversified landscape picture applies principally to the central part of the morainic hills in the Mazurian Lake District.

The parent rocks from which the soils have developed, were exclusively Quaternary deposits. They originated from rock material deposited by the last glacial ice sheet (Lencewicz, Kondracki 1963).

During deglaciation by the Baltic inland ice, a diverse land relief came into existence which afterwards, protected by perennial vegetation, retained its original character for a long time. However, the gradual development of agriculture contributed to the disappearance of the permanent meadow and forest resources which had even covered regions which were in danger of strong potential erosion, and to their conversion into cultivated land. In this way on exposed scarps processes of geomorphological changes set in (Jann 1954) whose extent and degree of intensity increased at the rate at which agriculture rose in intensity. This was especially dependent on the extent to which technique of ploughing was improved. Thus it was the impetuous of economic progress which contributed the disturbance of the equilibrium of natural agencies; and in this way led to gradual changes in the postglacial landscape of the Lake District.

In modern times geomorphological processes in the Mazurian Lake District are operating fairly intensively; for the most part they are caused by the marked variableness of climatic conditions (Chomicz 1951, Hohendorf 1956, Niewiadomski, Poradowski 1959, Niewiadomski, Krzymuski 1959, Niewiadomski, Skrodzki 1964, Rodomski 1962, Solarski 1962, Ugglą 1957), the sharply diversified land relief (Lencewicz, Kondracki 1963, Niewiadomski 1958, 1959, Niewiadomski, Grabarczyk 1962, Skrodzki 1962, Ugglą 1957, Ugglą et al. 1968), and the remarkable susceptibility of the soils to slope processes (Niewiadomski 1958, 1960, Niewiadomski, Poradowski 1959, Niewiadomski, Krzymuski 1959, Ugglą 1957, 1958, Ugglą, Nożyński 1959, Ugglą et al. 1967, 1968).

### CLIMATE

The Mazurian Lake District is situated in the transition zone between a marine and a continental climate. The clash of atmospheric properties of these two climates leads to frequent changes in the weather. During summer

the inflow of continental air causes heat waves and periods of drought; in winter the result is severe frosts. Moist air arriving from above the Atlantic brings in summer rains and cool periods, and in winter it causes the weather to become less severe.

It is worth stressing that the conflict of the two climates in the Mazurian Lake District is particularly sharp and that, in consequence, changes in weather conditions occur more frequently than in the remaining parts of the Polish Lowland (Hohendorf 1956). Also, this variability of the weather is greatly intensified by local influences brought by the specific features of this region, by the wide expanse of water basins (5%) and the considerable space occupied by forests (27.5%).

After Milata (1951) the Mazurian Lake District is the coldest region of Poland, compared with the remaining part of the Polish Lowland. The mean annual air temperature varies from 5.6°C at Gołdap to 7.0°C at Malbork. The duration of the vegetative period varies, lasting from 187 days at Olecko to 207 days at Malbork. The number of days with groundfrost is from 130 to 140, with a mean of 110 for all Poland. The first cases of groundfrost occur about October 10, the last — on May 8 to 10. However, there are years in which groundfrost occurs towards the end of August and the last as late as June (Hohendorf 1956).

Precipitation is from 550 to 600 mm (Hohendorf 1956), though in some regions it exceeds 700 mm. This precipitation is distributed fairly uniformly over the year: it is lowest in February and March (20–30 mm), slightly increasing toward spring and rapidly rising to its maximum in July and August, with 80–90 mm as mean values. The number of days with precipitation varies in the year usually being from 170 to 190 (Hohendorf 1956).

A snow cover develops as early as the middle of November and lasts from 62 days at Nowe Miasto to 82 days at Olecko. It should be added that in wintertime the snow even disappears several times, causing an abundant surface runoff of meltwater streams (Niewiadomski, Skrodzki 1964, Reniger 1952).

## SOILS

In the area under discussion brown and podsol type soils predominate: hydromorphs and black soils are much less common (Uggla 1958). The former have developed from sands, boulder clays and from clayey and silty deposits of aqueous origin, — the latter from loose and slightly loamy sands or from compacted boulder clay. The brown soils are encountered in the north-western and central part of the Lake District, mainly covering strongly undulating regions, which are very susceptible to aqueous erosion. The podsol type soils are scattered all over the Lake District, and in the northern and central part they cover undulating regions. When cultivated in a negligent manner these soils easily change into silts and lose their resistance to erosion (Uggla 1958, Uggla et al. 1968). In the south-eastern part the podsol type soils for the most part cover lowland areas and river valleys. These soils have a humus content and unless protected by vegetation they easily suffer eolian erosion (Uggla 1961). Swampy soils have developed in undrained depressions and in the valleys of some rivers like the Łyna, Pasłęka, Pisa, Biebrza, etc.; they also occur here and there in the eastern, south-eastern and some of the northern part of the Lake District. As a rule the lacustrine deposits are mantled by vegetation of some kind, because tillage would produce a silty sheet easily subject to eolian erosion. The black soils originate from heavy boulder clay and from clays, or



from lacustrine soils (Uggla 1958). They are found in the northern part of the Lake District, in Kętrzyn and Węgorzewo counties. This type of soils is fairly resistant to aqueous erosion.

In the opinion of Uggla and his co-authors (Uggla 1957, 1958, Uggla et al. 1967, 1968) soils developed from slightly loamy material and from compact and loose clayey sands are the most subject to erosion. On the other hand, soils developed from pure loose sands are much more resistant to slopewash.

## RELIEF AND EXTENT OF EROSION

As noted above, the Mazurian Lake District is morphologically a very diverse area. The absolute altitudes differ greatly by much as 312 m (Lencewicz, Kondracki 1963), and the altitude above sea level of local elevations averages 135 m. The mean relative height of the hills is from 10 to 30 m, with maxima rarely as great as, or more than, 100 m (Uggla et al. 1967, 1968). On the whole there is a predominance of short slopes of 60–80 m length, with summits resembling dome-like caps and with remarkably narrow slope bases. The average gradients vary between 20 and 15<sup>0</sup>/<sub>0</sub>, with straight or concave-convex profiles prevailing.

With regard to land relief, three zones can be distinguished in the whole Mazurian Lake District, which extend approximately along parallels of latitude: (1) the southern outwash plains, (2) the central morainic hills, and (3) the northern outwash plains.

(1) The southern outwash plains are composed of the Mazurian Plain, the Biebrza Basin, and the Augustów Plain. Level, at times slightly undulating areas dominate in the relief, with elevations varying between 120 and 160 m a.s.l. The soils are mostly sandy (outwash plains) and partly swampy. Considerable areas are covered by coniferous forest. The role of aqueous erosion is insignificant here.

(2) The central morainic hills cover the greatest part of the area and contain the highest elevations. It is here that the majority of lakes are situated, and from here issue many rivers flowing north- and southward (Lencewicz, Kondracki 1963). This zone is formed by numerous chains of end moraines, dissected by picturesque river valleys and glacial channel lakes. Firstly Garb Lubawski (the Lubawa Spur) should be mentioned with its wide spread of hills and its slopes which are partly steeper than 18<sup>0</sup>/<sub>0</sub>; its peak, Dylewska Góra, lies at 312 m a.s.l. Only 15<sup>0</sup>/<sub>0</sub> of this zone is covered by forests (Solarski, Szydłowski 1964). Next comes the Iława Lake District with its very diversified relief and an abundant system of glacial channel lakes. Here forests cover nearly 28<sup>0</sup>/<sub>0</sub> of the area. The northernmost part is formed by the Górowo Hill (216 m a.s.l.). The central part of this zone is occupied by the Olsztyn and the Mrągowo Lake Districts where the highest summit reaches 220 m a.s.l. This central chain is characterized by a very well developed network of hills which are dissected southward by long glacial channel lakes and river valleys. Here the extent of forest covering is high, being 25–30<sup>0</sup>/<sub>0</sub>. Farther east is the hilly area called the Large Lake Region. Here are found the largest group of lakes which have developed in the ground moraine. These have greatly diversified shore lines (Lakes Śniardwy, Mamry, Niegocin). Eastward is the Ełk Lake District whose northern part shows the highest elevations. The altitude differences are greatest in the northernmost part, called Szeskie Wzgórza (309 m a.s.l.). The Suwałki Lake District lies in the part of the zone extending farthest

northeastward. Its relief is very dissected and readily subject to erosion. In this part the extent of forest covering is unsatisfactory, being only 16%.

(3) The northern ice-dammed plains consist of level or slightly undulating areas of what is called the Warmian and the Sępopol Lowlands. In this part the slopes are rarely steeper than 6-8%. The soils show a high resistance to erosion, being compact and hard to erode.

The three zones discussed above roughly coincide with the natural-rural regionalization distinguishing areas subject to erosion (Fig. 2), as suggested by Niewiadomski and Krzymuski (1959).

Uggla and co-authors (1967) have divided this north-eastern region of Poland into three different zones with regard to the threat of erosion (Fig. 1). Zone 1 takes in all the areas where there is no pressing danger of erosion, because the slopes do not exceed 6%; spatially Zone 1 embraces some 52% of the total area under discussion. Zone 2, representing a medium threat of erosion, covers areas with slopes of 6-12%. The authors consider the total area belonging to Zone 2 to be about 42%. Finally, to Zone 3, the area greatly threatened by erosion, the authors assigned regions where the slopes are in excess of 12%; this zone constitutes about 6% of the total area.

Summing up it should be stressed, that in the Mazurian Lake District more than 40% of the soils occur in regions with very diverse relief, so that the danger of aqueous erosion must always be borne in mind.

#### *Intensity and causes of erosion*

Both aqueous and eolian erosion occur in the Mazurian Lake District. On the whole, eolian erosion does not cause any greater danger and its effect is barely noticeable. More often this sort of erosion can be observed in spring, more rarely in winter (Niewiadomski, Poradowski 1959, Uggla et al. 1967). It mainly attacks the lighter soils when these are thoroughly dry, picking up and carrying off the fine particles of organic and mineral substances.

Much more serious, both as to frequency of occurrence and intensity, is aqueous erosion, Niewiadomski (1964) and Uggla (1957, 1961) both believe, that the most powerful erosion is that caused by torrential rains. After Chomicz (1951), the Mazurian Lake District is second after Silesia in the highest frequency of occurrence of this type of rains; his analysis indicates that it rains most often in spring, i.e., in May and June.

Our own observations made from 1955 to 1968 confirm the high frequency of torrential rains. In 14 years these periods occurred 13 times (Table 1).

The early-spring showers falling in April and May have the most disastrous consequences. It should be remembered, that in this early season the danger of erosion is particularly great because the soil, when freshly loosened by spring cultivation, lacks a protective cover and is easily displaced. The results of the 14-year study of aqueous erosion, made by means of methodical measurements in special catchment basins (Niewiadomski, Skrodzki 1959), are presented in Table 2. Owing to the abundance of documentary material the author only reports on fragments, using the results of four years of observations (1956, 1962, 1965, 1967) which were the most characteristic with regard to the amount, the intensity and the frequency of precipitation.

The course of precipitation in 1967 as illustrated in Tables 1 and 2 is of particular interest. Torrential rains fell as often as five times, and the destruction caused was extremely heavy, especially during the first and second showers on May 23 and 24. Calculated per 1 ha, downwash was 29.5 t in areas where cultivation was done in an anti-erosive manner, perpendicular to the slope, and up to 50 tons where cultivation followed the slope of the field in the

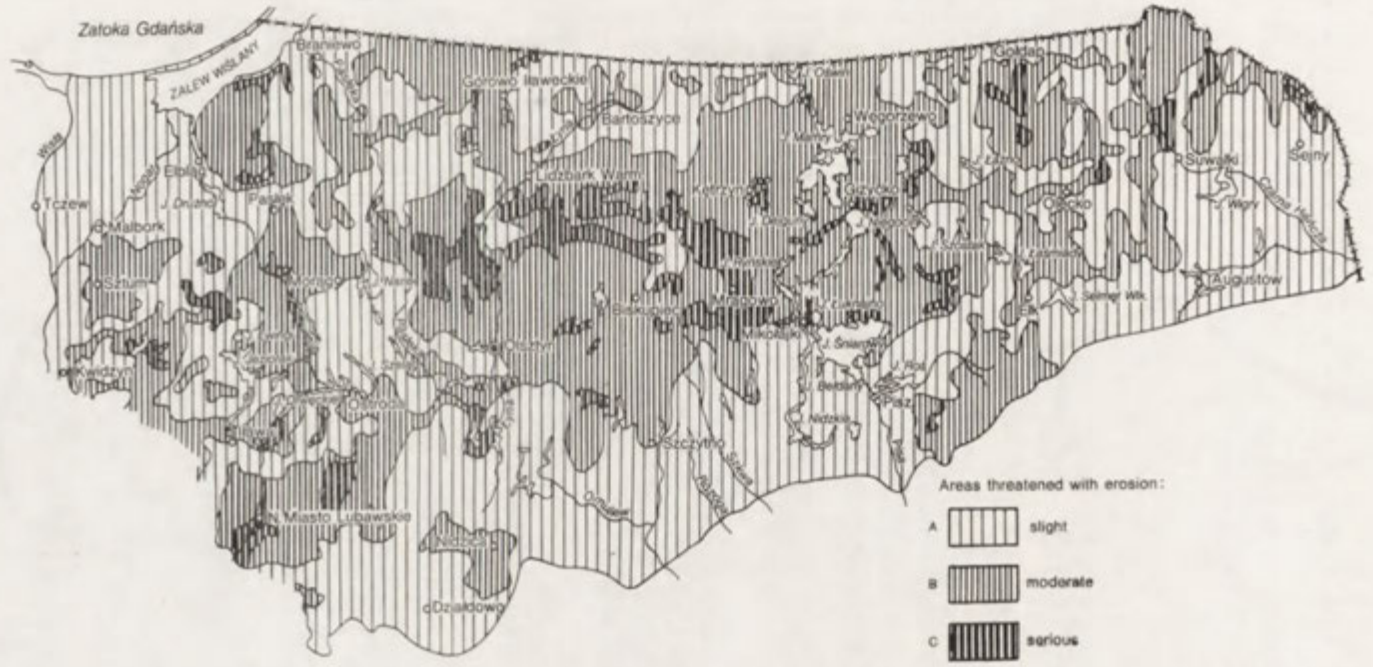


Fig 1. Zones of susceptibility of soils to water erosion in NE Poland (Compiled under the supervision of H. Uggle)

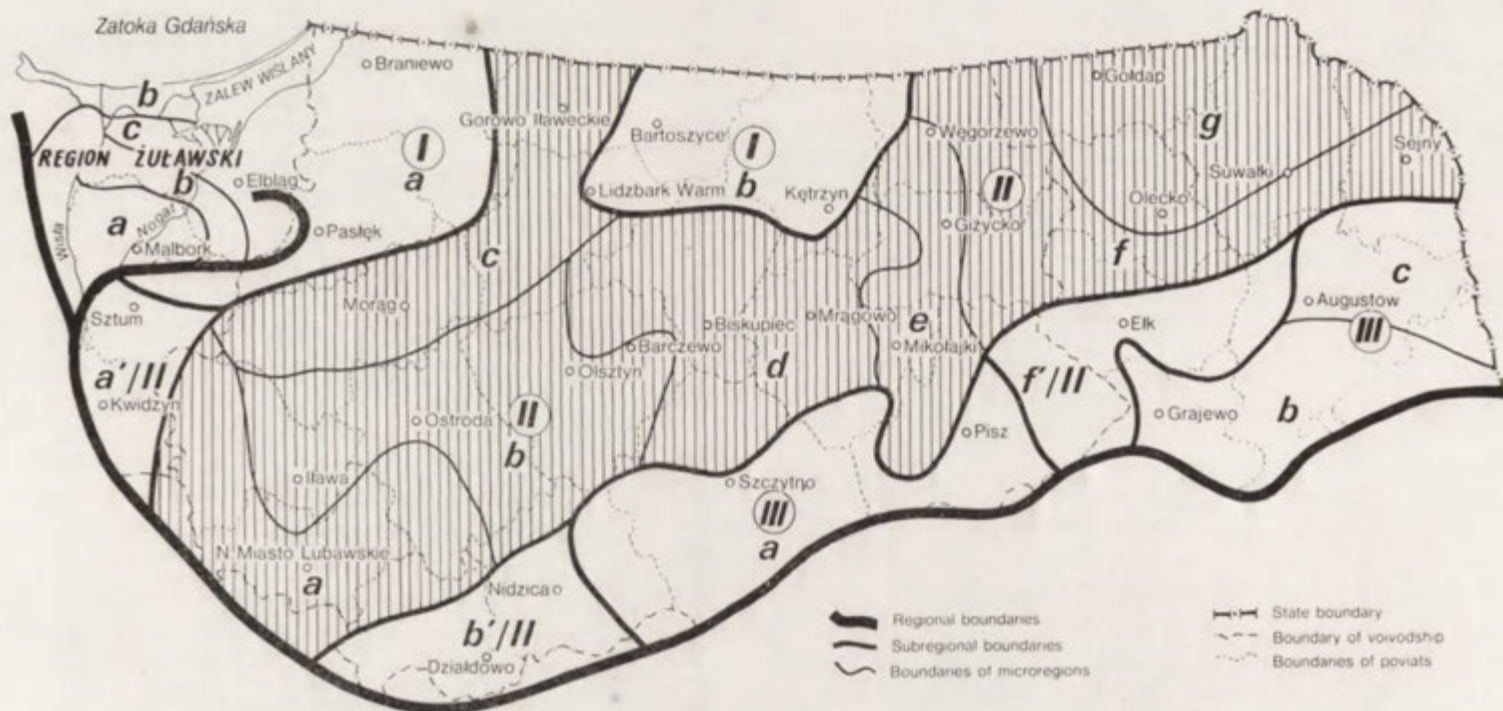


Fig. 2. Potential susceptibility of soils to water erosion in NE Poland. After W. Niewiadomski and J. Krzymuski

Żuławy Region — Microregions: a — wheat-beets; b — crops-meadows-pastures; c — meadows-pastures

North-Eastern Region — Subregions: I — Ice-dammed, pastures and wheat — Microregions: a — Braniewo-Pastłęk; b — Kętrzyn-Bartoszyce; II — Lake district, rye and fodder crops — Microregions: a — Nowe Miasto Lub.; — Ostroda-Olsztyn; c — Morąg; d — Reszel-Mrągowo; e — Large Mazurian Lakes; f — Eastern Mazuria-Sejny; g — Goldap-Suwałki; III — Outwash plain, rye-lupine meadows — Microregions: a — Szczytno-Pisz; b — Biebrza Plain; c — Augustów

Transition microregions: a'/(II) — Towards Lower Vistula Valley; b'/(II) — Towards Mazovian Lowland; f'/(II) — Towards Kolno Upland

TABLE 1. Torrential rainfall at the Experimental Station Pozorty in 1955-1968\*

Year	Date	Number per year	Quantity in mm	Duration in min	Rainfall in mm/min
1955	30.IV.	1	20.2	30	0.67
	1.VII.	2	22.0	50	0.44
1956	23.VI.	1	40.2	50	0.80
1957	10.VI.	1	72.0	90	0.80
1958			no torrential rainfall		
1959			no torrential rainfall		
1960	25.VII.	1	17.0	25	0.44
1961			no torrential rainfall		
1962			no torrential rainfall		
1963	7.VII.	1	15.5	5	3.30
1964	5.VI.	1	16.0	15	1.06
1965	22.VI.	1	20.6	20	1.03
1966			no torrential rainfall		
	23.V.	1	60.0	20	3.00
	24.V.	2	22.5	5	4.50
1967	24.VI.	3	21.7	40	0.54
	26.VI.	4	10.3	45	0.23
	27.VI.	5	23.2	60	0.38
1968			no torrential rainfall		

\* According to the Meteorological Station at the College of Agriculture, Olsztyn

TABLE 2. The impact of various modes of cultivation of the intensity of aqueous soil erosion

Year of research	Rainfall in mm	Kind of downwash	Soil erosion in t/ha		
			mode of cultivation		
			transverse to slope	parallel with slope	permanent turf
1956*	86.3	after snowmelt	0.017	0.034	0.007
	40.2	after torrential rains	1.260	5.250	—
	538.6	after moderate rains	0.091	0.152	0.019
Total	665.1		1.368	5.436	0.026
1962	82.5	after snowmelt	0.047	0.045	0.035
	—	no torrential rains	—	—	—
	464.5	after moderate rains	0.006	0.004	0.005
Total	547.0		0.053	0.049	0.040
1965	54.3	after snowmelt	0.019	0.028	0.011
	—	no torrential rains	—	—	—
	492.7	no sheetwash observed	—	—	—
Total	547.0		0.019	0.028	0.011
1967	89.1	after snowmelt	0.003	0.004	0.003
	137.7	after torrential rains	30.756	54.935	0.028
	463.0	after moderate rains	0.051	0.068	0.007
Total	689.8		30.810	55.007	0.038

\* After W. Niewiadomski and M. Skrodzki

traditional way. In crops of beetroots and winter wheat, many furrows developed of 50 or so meters long, 10 to 30 cm deep and 20 to 30 cm wide. The slopewash deposits accumulated at the base of slopes reached thicknesses of 10 to 15 cm. At the same time on turf-covered fields practically no traces of erosion were observed (Table 2).

To rather frequent features of erosion, which are not particularly dangerous in their effect, must be assigned slope caused by meltwater flow. In these cases erosion does little harm because frozen soil suffers little from slope processes; however, the loss of water is very considerable. Investigations made by Niewiadomski and Skrodzki (1964) revealed that surface runoff by melt-

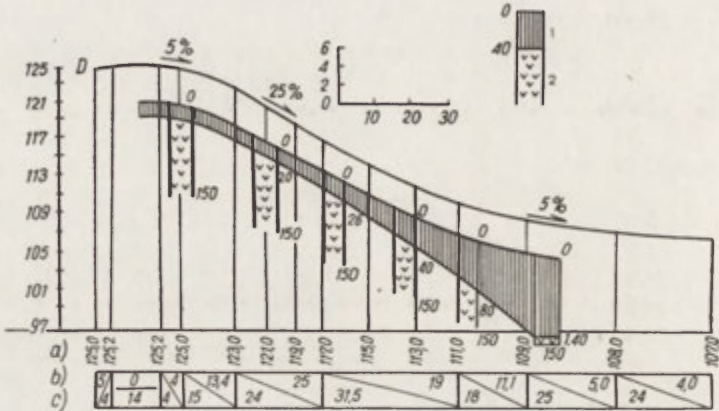


Fig. 3. Longitudinal section of slope C—D (Experimental Station Pozorty). After M. Skrodzki

1—humus horizon or deluvial layer, 2—brown horizon; a—ground datum, b—slope gradient in per cent. c—distance, in m

water may take place several times during one winter. This mostly happens when the weather abruptly turns warmer since this is usually accompanied by rains which intensify snowmelt.

Generally speaking it is rare that moderate rains cause erosion. This is only the case when precipitation lasts for a long time and when it is copious, especially when the rain falls on soil devoid of vegetation. It is also worth mentioning that surface flow over the soil surface, even if it does not actually displace much of the soil, is usually very harmful because it washes out valuable chemical components dissolved in the soil which are then unavoidably lost in streams and lakes. This fact has been confirmed by field studies made by Koter and co-authors (1963), by Niewiadomski and Skrodzki (1964) and by Solarski and Skrodzki (1966).

The effect which the long-term action of slope processes has upon changes in the surface layer of the soil is illustrated by the way a humus cover is developing on three different elements of a slope: on the summit, the middle slope and the base (see Figs. 3 and 4, and Tables 3, 4 and 5). Table 3 pictures the content of organic substance and the pH value determined for soil samples collected from 28 morainic hills situated in the central zone of the Lake District. From the figures of this table it appears that usually both the thickness of the humus layer and the percentage of the organic substance in the top 25 cm sheet of the soil are greater at the base than on the slope or the summit.

The effect of the downward displacement of the humus layer is that the content of other chemical components in the soil is also reduced (Mirowski, Lewicka 1962, Mirowski 1964, Skrodzki 1963, 1963a). Some authors hold that this difference in favour of the soil quality in the pediment zone may be 8 to 10 times that of the quality higher up. However, it seems difficult to imagine how to substitute for the losses of humus caused by aqueous erosion in undulat-

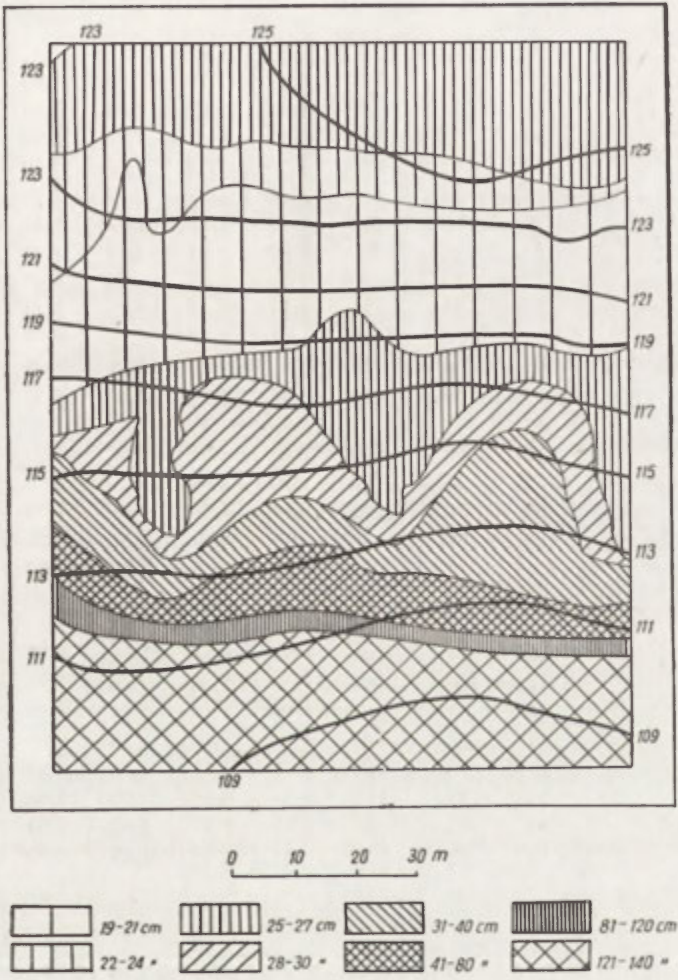


Fig. 4. Isolines of thickness of humus layer on the slope (Experimental Station Pozorty). After M. Skrodzki

ing regions. To enrich the soil requires, apart from an extensive period of time, the application of considerable quantities of organic fertilizer. It is also important to remember, that excessive amounts of humus accumulation at the slope's base and in ground depressions may not always bring favourable results, but it may also cause damage by flattening some of the produce already growing in these parts.

TABLE 3. Content of organic substance and pH value of soil, determined for 28 slopes in the Mazurian Lake District

Locality	Thickness of humus layer in cm			Organic substance in per cent (0-25 cm)			ph value of soil in 1n HCl (0-25 cm)		
				Site					
	summit surface	slope	base of slope	summit [surface	slope	base of slope	summit surface	slope	base of slope
Pozorty I	20	28	90	0.67	1.10	1.25	7.50	7.20	5.60
Pozorty II	25	38	90	1.09	1.02	1.14	7.20	7.50	5.80
Pozorty III	29	20	130	0.90	0.69	1.47	5.40	5.00	4.80
Samławki I	20	23	26	1.35	1.78	1.96	7.30	5.80	5.40
Bęsia	26	35	110	1.44	1.28	2.13	5.90	5.70	5.70
Łężany I	25	10	100	1.56	1.71	2.03	5.00	5.70	6.50
Łężany II	24	27	150	1.81	1.83	2.13	5.00	6.00	6.00
Samławki II	26	24	33	1.18	1.45	1.56	4.60	4.60	4.50
Bartązek	25	25	220	1.83	1.88	1.85	5.70	6.50	6.20
Bartąg	30	23	230	1.57	1.71	2.05	6.10	7.10	7.40
Glotowo	8	8	40	1.49	2.34	2.01	4.40	5.90	4.80
Wróble	30	26	130	1.38	1.59	2.02	4.80	4.80	4.90
Tomaryny	42	44	49	1.88	1.89	1.83	7.30	7.40	5.50
Grabin I	22	90	150	1.46	0.91	1.58	7.00	6.50	6.00
Grabin II	22	22	60	1.10	0.98	1.10	6.00	6.60	6.70
Gierzwałd	55	32	150	1.48	1.56	2.23	7.30	5.80	6.30
Bratian	15	80	125	1.36	1.97	1.36	7.20	7.70	7.70
Kurzętnik	18	55	68	0.95	1.26	1.43	7.70	6.50	5.60
Nielbark I	52	32	90	1.72	1.84	2.12	6.80	6.50	6.60
Nielbark II	26	23	140	1.22	1.53	1.88	7.40	7.80	7.70
Spręcowo	24	24	110	1.68	1.54	1.68	4.60	5.00	5.60
Radostowo I	22	20	140	1.27	1.56	1.44	7.30	7.50	5.30
Radostowo II	20	30	47	1.45	1.44	3.55	6.10	6.70	5.60
Kalis	29	28	90	1.31	1.26	1.36	6.70	6.00	5.70
Bęsia I	23	30	110	1.52	1.05	1.86	6.30	6.70	7.20
Bęsia II	26	35	110	0.36	0.63	4.09	7.60	7.60	5.50
Wola I	30	45	90	1.53	1.53	1.62	5.80	6.10	5.60
Wola II	22	25	100	1.00	1.06	1.63	6.20	7.20	6.20

TABLE 4. Dynamics of organic substance in per cent on slopes of the Experimental Station Pozorty in 1961-1965 (in 0-25 cm layer)

Site	Cultivation - by ploughing			Perennial turf cover		
	1961	1964	1965	1961	1964	1965
	Summit surface	1.22	1.15	1.13	1.15	1.17
Slope	1.04	1.04	1.03	1.02	1.06	1.11
Base of slope	1.08	1.19	1.22	1.18	1.30	1.38



The mechanics of geomorphological changes taking place in modern times is well illustrated in Fig. 4. This figure pictures the pattern of thicknesses of the humus layer observed in one of the permanent experimental plots situated on a southward scarp of the Pozorty Experimental Station (Skrodzki 1963a). From these figures it can be seen that high up, near watershed line, the humus layer is 25–27 cm thick. With the slope gradient increasing to 25% in

TABLE 5. Quantity of organic substance in t/ha, in different parts of slope (After M. Skrodzki)

Site	Thickness of humus layer, in cm	Cultivation by ploughing	Perennial turf cover
Summit surface	24–26	44.9	44.5
Slope	19–25	42.7	40.6
Base of slope	121–140	94.9	99.2

a downward direction, the humus thickness decreases to 19 cm. Afterwards, with a lessening gradient the layer deposited by sheetwash increases until it is as much as 140 cm thick at the base of the slope (Figs. 3 and 4). This illustration clearly indicates that the accumulation of considerable quantities of soil material at slope bases is the effect of aqueous erosion on slope surfaces.

#### METHODS OF PREVENTING EROSION

One of the fundamental elements in protecting soils against aqueous erosion on undulating areas is to apply the most suitable ways of cultivation, embracing a correct direction of ploughing, proper fertilizing, sowing, plant care and crop rotation (Bac 1947, 1955, Lenciewicz, Kondracki 1963, Niewiadomski 1957, 1958, 1959, 1962, 1968, Niewiadomski, Krzymuski 1962, Niewiadomski, Grabarczyk 1962, Niewiadomski et al. 1964, Niklewski 1964, Skrodzki 1962, 1963, 1963a).

For seventeen years observations were kept up at the Pozorty Experimental Station concerning the most suitable way of caring for field areas which had suffered erosion. In order to do so, one of the three known methods of cultivation were alternately applied: (1) the traditional technique, (2) the anti-erosion system of field work and (3) a system providing a perennial turf cover. These experiments produced evidence that anti-erosion cultivation is definitely superior to the traditional technique. The figures shown in Table 2 indicate that erosion-prevention methods protect the soil layer against destructive displacement much better than did the traditional method of tillage. During torrential rains in May 1967 it was found that sowing in a direction perpendicular to the slope is highly important in counteracting erosion; nearly 50% less sheetwash occurred compared with furrows running parallel with the slope. Among cultivated plants perennial mixtures proved to be the best protection against erosion, and for this type of vegetation (Table 2) slope wash was rather negligible. Further, perennial mixtures have a marked property of accumulating considerable quantities of organic substances in the soil; they enrich the physico-chemical properties of the soil and act favourably upon the water balance. This is why this sort of planting warrants the most effective

protection against erosion — an opinion fully confirmed by studies made by Kern (1957, 1964) and by Niewiadcmiski (1959). The latter author maintains that apart from the proper technique of contour tillage, trends toward erosion can also be checked by a special crop rotation comprising both winter rye and perennial plant mixtures.

It might be added that in the anti-erosion crop rotation pattern it would be advisable to use aftercrops at a higher rate because, apart from their protective properties during winter and spring, they raise the amount of crops produced (Skrodzki 1963). Furthermore, this is a problem requiring additional detailed economic investigation.

In conclusion the author emphasizes that the best means of protecting soils against erosion is a forest environment (Reniger 1950, Solarski 1962, Solarski, Szydłowski 1964, Ziemnicki, Mozoła 1966). By their dense undergrowth and high layer of litter, forests prevent violent runoff and improve water retention in the soil. They also favourably affect the climatic and hydrological conditions of the habitat. For this reason a routine of planting tree stands and forests should be applied wherever land is unsuitable for cultivation, and to all space where slope steepness prevents tillage (Niklewski 1964).

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## NIVEO-EOLIAN PROCESSES IN THE SUDETES MOUNTAINS

ALFRED JAHN

### PREFACE

In his classical essay on wind action in Sweden C. Samuelsson (1927) wrote: "The only form of erosion ... which can be put forward as a rational argument to explain soil degradation, is deflation". From autumn to spring ploughed fields are a type of desert. Eolian erosion is the principal cause of the destruction of light sand-humus soils. After C. Samuelsson the mean indices of deflation for Sweden find their expression in the lowering of the soil surface by 2 to 9 mm per year. We in Poland consider this estimate of the results of eolian erosion excessively high; still, Samuelsson's opinion deserves attention and confirmation.

For my own part, toward the end of December 1963 I happened to make my first observations on the effect of winter winds in the Sudetes. At that time, while the air was chilly and the earth covered with a thin early-winter sheet of snow, fairly strong winds were blowing at about 12 m/s for several days. This caused the snow to be swept away, bare patches of soil to appear, and snow-drifts to develop. In carrying off the snow the wind also snatched away particles of soil, so that the snow deposited in drifts and ridges was literally saturated with dirt<sup>1</sup>. And this made the snow assume a black, brown or red colouring, depending on the geological structure of the subsoil.

Starting from this observation, systematic measurements of the dirt content in the snow were made for three years (1964-1966) at 14 selected points of the Sudetes in the area between Świdnica and Jelenia Góra, thus in the Bolków Upland, the Kaczawa Mts. and the Jelenia Góra Basin (Fig. 1). Most of these control points were situated on ploughed fields, but observations were also made on meadows and in forests.

These control measurements were each made 3 to 4 times during winter and spring, and every time samples were collected:

- (1) from the surface of a snow-drift (from a 0.5-1.0 m<sup>2</sup> surface)
- (2) from inside the snow-drift (samples of set volume)
- (3) from the soil surface where deflation had taken place.

After the snow had thawed and the sample was dried, the quantity of dirt was determined by weighing.

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<sup>1</sup> In the course of this study the term "dirt" shall be used to mean all sorts of organic or mineral material like fine sand, dust and clay particles, carried by the wind and deposited inside and on top of snow-drifts.

The part of the Sudetes where these observations were made is built of granites (Jelenia Góra) or sedimentary rocks (shales and sandstones), metamorphic rocks (schists and diabases) and mesozoic age. The soils are skeletal in character and intensively coloured depending on the type of their parent rock; the altitude of the mountains in this region is 400 to 500 m a.s.l.

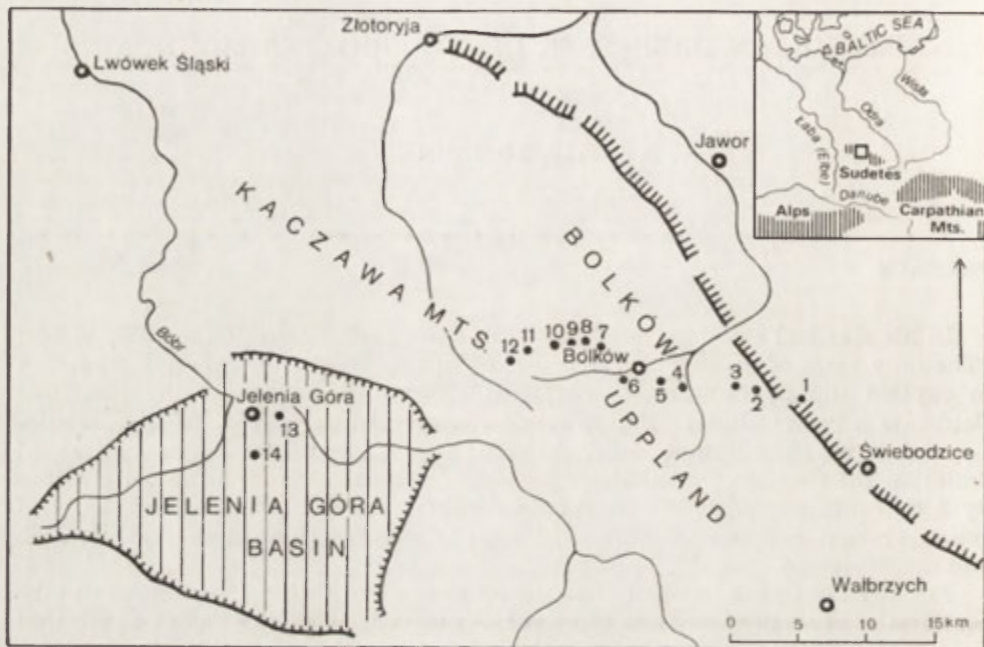


Fig. 1. Map showing distribution of observation stations in investigated part of the Sudetes

#### NIVEO-EOLIAN PROCESSES

The action of wind upon snow and, combined with snow, upon the earth surface consists of three separate processes: transport, destruction, and deposition.

Snowflakes and snow particles are swept away by the wind: (1) before they reach the earth, and (2) by being plucked off from the surface of a snow cover.

Let us now record a few important data regarding the theory of snow transport, as far as they are needed for understanding niveo-eolian processes as a whole.

While descending by gravity, snowflakes are easily swept away by the wind, and the degree of their angular deviation from a vertical drop depends on the ratio of the weight of the snow particles ( $P$ ) to the force of the wind ( $K$ ), hence on  $P/K$ . The wind-borne snow particles assume tracks approaching the horizontal, and this process of snowfall conditioned by the wind is called a snow flurry. The stronger the wind, the closer the track of snow particles follows the course of the wind. The angle between the horizontal and the track taken by the snow is called the angle of the snow flurry. This angle



results from, and can be defined as, the difference between wind velocity and the descending motion of the snow particle, thus the ratio  $P/K$ .

The sweeping away of snow particles from the surface of a snow cover does not only depend—apart from the wind force—on the weight of the snow particles but also on the force by which the particles clings to the snow base, i.e., on the cohesion or the compactness of the snow, and on friction. When the temperature is above zero and thawing sets in, an important factor is the viscosity (adhesion) of the snow particles. This shows that transport and redeposition of snow require a wind force greater than that needed for the mere transport of descending and swept-away particles.

Commonly in use is the notion of a “critical wind velocity”, that is of a lowest rate of wind force upward of which snow particles are picked up and carried off. The wind velocity close to the earth's surface, which differs from that several meters higher up must always be kept in mind. Empirically it has been determined that next to the ground, up to 20 cm high, the wind velocity is about one half (50–60%) of the velocity at 2 m height, and less than one half (40%) of the velocity at 8 m above ground surface. This indicates, that what is called the “meteorological” wind velocity measured a few meters above ground surface would have to be at least 5 m/s in order to start deflation of a snow surface. However, any sweeping away of snow particles is dependent not only on the force of the wind but also on the angle at which the wind hits the snow surface. Deflation starts most easily when the wind direction is parallel with the deflation surface; the chance of deflation decreases with the growth of the angle of incidence, and it fails altogether when the wind impinges perpendicularly upon the snow surface.

The manner in which snow particles are picked up and carried off is contingent on the degree of wind turbulence, on abrupt changes in wind force (squalls, gusts), and on changes in wind direction. There are times when the wind blows with a constant force in one definite direction; but things are different when the snow particles travel in one direction but at different rates of velocity. A third alternative involves variations in the force and the direction of the wind. Obviously, as far as snow transport is concerned, the last-mentioned mode of wind is the most effective.

Fresh-fallen snow, accumulating in the form of a powder, is swept away when the wind velocity on the ground surface is anywhere between 3 and 8 m/s; but for moving a firmly settled cover of densely compacted snow, a correspondingly higher wind velocity is required. And frozen, recrystallized snow containing an ice crust or ice layers, or trod-down snow, sometimes by the very force of the wind, will not be snatched up and carried off by a wind unless its velocity is anywhere between 20 and 30 m/s.

The quantity of snow transported during a snow flurry can be determined by using catching devices, that is, by means of receptacles the opening of which is turned into the wind<sup>2</sup>. The weight of the snow sample caught this way, considering the area of the opening (in cm<sup>2</sup>) and the time of operation (in minutes), gives us the “flow of the snow flurry” (in English “solid flux”, in Russian *tverdyj raschod*)—the counterpart to which is “flow” for flowing water (a river).

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<sup>2</sup> In the Soviet Union several types of catching devices called *metelemerye* are in use. A classical type suggested by Kuznecov consists of a cylindrical steel can which freely turns upon a vertical pole, with its opening always into the wind (P. P. Kuzmin 1960).

Three different types of snow transport are known (M. Mellor 1965): creep, saltation of the snow particles, and turbulent diffusion. The term creep is used when the snow particles move, wind-pushed, over the ground surface, combined with their being turned. Saltation, occurring at wind velocities higher than those causing creep, is a motion which consists of repeated alternating snatching up of particles from the ground and dropping of them. The leaps may be 10 cm high, but also as much as 100 cm. The third type of snow transport called turbulent diffusion presupposes turbulent air currents by which the snow, whirling like a curl of smoke, is raised high into the air, at times as much as to 100 m altitude.

Snow deflation depends on the conditions of the ground surface, i.e., on its extent and relief and on its vegetation cover. As stated by A. K. Djunić (1962), the minimum space on which a snow-drift can develop is a length of 200 to 300 m. For a distance like this, snow can be swept away forming drift ridges. The more extensive an area of deflation, the larger can be the drifts formed. However, there are limits to how far snow can be wind-borne horizontally. It cannot travel farther than 1 to 3 km, dependent on humidity conditions, because by travelling for distances larger than these the snow particles would vanish due to evaporation.

The land relief may make deflation easier or more difficult. Land forms laid open to the direct action of winds such as mountain summits, or passes, or slopes exposed to constant air currents—they all further snow deflation, and snow overhangs are apt to develop. On horizontal surfaces a variety of snow-drifts and accumulation ridges are formed, dependent not only on force and nature of the wind but on air temperature as well. The peculiar morphology of a snow cover has been the object of detailed studies reported in a classical essay by A. B. Dobrowolski (1923). Further valuable publications worthy of attention are those published by G. D. Richter (1945) and G. Seligman (1936). Among different forms of snow covers also to be mentioned are: snow waves and *sastrugi* which develop at low temperatures from powder-snow; snow barchans which are piled up on smooth surfaces like ice sheets; ridges and longitudinal drifts, and snow tongues extending parallel with the wind direction. Obstacles blocking the wind at a right angle lead to the formation of transverse snow-drifts.

Vegetation covers, especially forests, are an effective obstruction to snow deflation. Tree belts are the most appropriate protection against the effect of snowstorms (Fig. 2). The wind force slackens ahead of a forest, at a distance



Fig. 2. Forests as barrier to wind action. Two zones of calm depending on wind direction (marked by arrows) ahead of and behind the forest (the aerodynamic shade)

H — height of tree stand

of about  $3H$  from the forest fringe (where  $H$  denotes the height of the tree stand). Generally speaking, a forest is unaffected by the deflating and accumulating action of the wind, although at times here also some powder-snow, carried high in the air, may be deposited. In the lee of the forest, i.e., in its wake, the wind does not recover its force until at a distance of some  $30H$  from the trees. Assuming an average tree height of 6–7 m, the calm zone protected by a tree belt has a width of about 200 m.

Every sort of barrier or obstruction, especially artificial ones like snow-fences along highways, reduce the wind force or produce secondary inverse currents, and ultimately they cause the wind-borne substance to be dropped. At fences, snow-drifts develop on both sides but the drifts behind the fence (in its lee) are, depending on the wind force, usually two or three times larger than those ahead of the fence (Fig. 3).

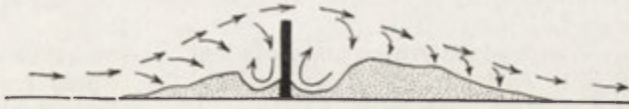


Fig. 3. Formation of snow-drifts, depending on vertical barriers set at right angle to wind direction

This concludes that was to be said about the niveo-eolian process of transport, and attention is now turned to matters of erosion and sedimentation.

Winds act destructively not only upon a snow cover but upon the top soil as well, and this takes place at points where part of the soil surface has been laid bare of snow. Vegetation such as mosses and grasses covering the soil is most easily destroyed. It often happens that the plant layer is covered by a sheet of ice, because an excess of water retained in the soil easily turns into ice. Only grass stems protrude from the ice and they become the victim of destructive winds, and of the snow and ice particles carried by these winds. Smashed and cut-up stems, leaves and plant twigs turn into a regular "chaff" which then is easily swept away for long distances. The process described here is typical of polar regions where during winter an ice crust is easily formed. On Spitsbergen in spring one encounters wide stretches covered by a sheet of snow mixed with plant chaff; here and there this layer may be 10 to 15 cm thick (A. Jahn 1961).

Wind and snow cause erosion in either of two ways: by picking up soil particles and blowing them away, or by corrasive destruction of rock or ground material. The latter action undoubtedly takes place in the Arctic and Antarctic, although it has been little investigated. Snow and ice crystals resembling fine needles, extremely hard at low temperatures, are hurled by the wind over rock surfaces as are sand clouds in deserts. But it is an open question whether in this case pure snow and ice particles are in action, or whether in this corrasion mineral soil particles are also taking part.

Blowing away (when referred to all the material held in the soil surface) or picking up (when only part of the mineral composition of the ground surface comes into play) is of much greater effect in the niveo-eolian process than is corrasive destruction which was discussed above. Generally speaking, "pure" wind-deposited snow can be seen never; admixtures of mineral or organic particles are always present, and they cause the snow to look "dirty". Sometimes we speak of "dust" although this admixture fails to answer this term in the sense of a fine fraction of some sedimentary material. The amount of dirt held in the snow varies within a wide range. At times it may be a question of solitary elements difficult to discern in the snow, but again one may observe a large-scale dirt contamination which affects the colour of the snow (red snow, black snow, etc.). Very often this sort of "mass" contamination tarnishing the snow is illusive. A small volume of dirt admixture suffices

to make the snow as a whole lose its whiteness. The cause of this optical deception is that the human eye notices not only the dirt particles held in the snow surface but also those which are embedded inside the snow and the colour of these passes through the transparent snow crystals. The sharper the contrast between the snow and the contaminating dirt, the less dirt is needed to change the snow colouring.

The niveo-eolian transport of dirt particles can proceed in two different ways: during a snowstorm together with the snow, or separately without simultaneous snowfall.

When it snows and wind is blowing, or when a snow cover is attacked by gusts of wind, soil particles are snatched by the wind, transported, and deposited some place, together with the snow. At the rate at which the snow cover vanishes, steadily growing patches of bare soil are uncovered, and in this way the share of soil particles in niveo-eolian sedimentation is increased. Normally a series of sedimentation shows dirt contamination increasing upwards. But after the snow cover has been swept off and the wind acts only upon the exposed soil, the surfaces of the snow-drifts accumulating the wind-borne material grow a "crust" consisting exclusively of soil particles. This thin sheet of soil dust is the final element of the sedimentation process. Unless new snowfall sets in, the wind piles up onto the old snow patches and drifts further soil material swept in from the bare fields, and this new deposit increases the thickness of the crust mantling the snow-drifts.

Well known from the Antarctic (A. Cailleux 1962) are sandstorms during which enormous masses of pure sand are swept away from mountain scarps and evenly deposited on snow-covered valley floors in a widespread layer. Next, new snow starts covering this sand layer and this process is repeated many times. In this way a peculiar structure develops in which pure recrystallized snow is intercalated with sheets and streaks of sand. Similar phenomena are reported from the Arctic and were observed in Greenland and also on Spitsbergen (A. Jahn 1961). This varying cycle of sedimentation of snow and rock particles leads to the rhythmically stratified geological deposits mentioned above which in French are called *grezes litées*.

In the niveo-eolian series, an ever denser arrangement of dirt contamination in an upward direction and on top of snow surfaces results not only from the way sedimentation is taking place, but also from snowmelt, thus from ablation, and from the snow becoming compacted.

The quantity of dirt, expressed in  $\text{g}/\text{cm}^2$  or in  $\text{kg}/\text{m}^3$  snow, can be defined by the degree of dirt concentration ( $s$ )

$$s = \frac{Wz}{v} \quad (1)$$

where:

$Wz$  = the weight of dirt thawed out from the snow and weighed when dry,

$v$  = the volume of snow in its natural state and in undisturbed structure, from which the dirt has been thawed out.

When the wind spreads this sort of dirt all over the snow mass its concentration ( $s$ ) is proportionally increased at the rate at which the snow density grows, due to metamorphosis and packing.

This results from simple considerations: the snow density ( $d$ ) equals

$$d = \frac{Ws}{v} \quad (3)$$

where  $W_s$  represents the weight of the snow. Substituting for  $v$  the value obtained in (1) we have

$$s = \frac{Wz}{W_s} \quad (3)$$

As a consequence of metamorphosis processes, the snow density ( $d$ ) may increase during a single winter, and the dirt concentration ( $c$ ) grows in the same proportion even if the mass of snow remains the same.

A further concentration of dirt may be caused by processes of ablation. Let us assume that  $1 \text{ m}^3$  of snow is contaminated by dirt—for simplicity's sake let the dirt be distributed uniformly in the snow—then the degree of concentration will be  $s$ . When the snow thaws and is reduced from the volume of  $1 \text{ m}^3$  to a layer of  $1 \text{ cm}$  thickness, that is, to the volume of  $0.01 \text{ m}^3$ , the dirt concentration will be  $100s$  (Fig. 4), provided the quantity of dirt remains

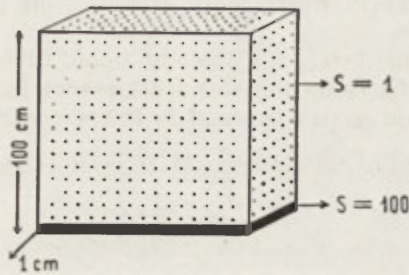


Fig. 4. Concentration of dirt admixture in the snow due to ablation

unchanged. This means that the degree of snow concentration equals the figure expressing the reduction of the thickness of the snow layer. Snow ablation which in spring reduces the snow thickness many times, in a very short period of time creates regular crusts of dirt on the surfaces of thawing snow-drifts. If the thickness of the crust and the original dirt concentration in the snow are known, it is easy to calculate the thickness of the snow cover from which the dirt has been thawed out.

This completes the description of niveo-eolian processes in their three successive stages: destructive action, transportation, and redeposition. Snow is always the active element, and therefore one must always anticipate certain results from these processes where snowy winters are the rule. This indicates that it is not only the polar climatic zone—in which the material for our observations has been collected—but also in the climatic conditions of Europe can sort of effect of snow action be expected to appear. The investigations which were made for several years in the Sudetes may serve as a convincing example; and it is the results of these studies which shall be discussed in the next chapter, supplementing the author's theory regarding niveo-eolian processes.

#### THE DEPENDENCE OF DEFLATION ON GEOLOGICAL STRUCTURE, SOIL TYPE, VEGETATION COVER, AND LAND MORPHOLOGY

How much eolian processes are conditioned by geological structure and type of soil can be seen from the following observations:

- (1) the most easily deflated are soils overlying shales, greenstones and tuffstones,

(2) the next group, less easily deflated, are soils on sandstones (Rotliegendes) and on porphyries,

(3) least subject to niveo-eolian processes are soils overlying granites.

Niveo-eolian processes are more closely linked with the vegetation cover than with geological structure and soil type. Referring to what was observed during a period of powerful deflation in December 1963, the following data were recorded:

	material deposited on the snow, in $g/m^2$	organic parts in deposit, in %
on ploughed fields	825	5.1
on meadows	49	11.4
in forests	22	11.8

Hence, deflation was 17 times more effective on ploughed fields bare of snow than on meadows.

The influence of the local morphology upon niveo-eolian processes was also distinctly noticeable. Deflation was least severe in the Jelenia Góra Basin and highest on mountain slopes exposed to the wind (Fig. 5).

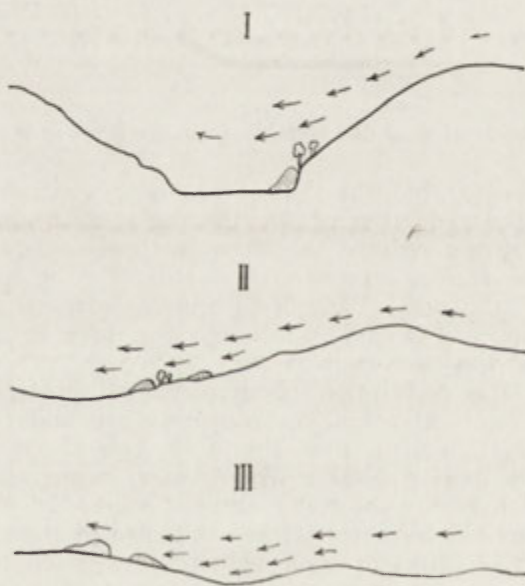


Fig. 5. Morphological conditions of the action of niveo-eolian processes in the Sudetes

I — in valleys, II — on ridge surfaces extending parallel to the wind direction, III — on ridge surfaces not parallel to the wind direction

#### THE MECHANICAL COMPOSITION OF NIVEO-EOLIAN SEDIMENTS

Analyzing the mechanical composition of the material deposited, together with the snow, by the wind and comparing this material with the composition of the soil surface of the field area from which it was swept off by deflation, one faces a difficult problem in the determination of the nature of the

mechanics of wintertime deflation. And the question arises: is the soil being carried off in its total composition, or does some process of mechanical selection take place?

In the mountains a considerable part of the soils suffering deflation are skeletal: on an average 30 to 40% of the mechanical composition are particles larger than 1 mm diameter. But it appears, that a niveo-eolian deposit is practically devoid of skeletal components or it contains very little of it, and this is the only evidence showing that deflation is connected with some selective action of the wind. On the other hand, among the "dirt" contaminating the snow (Fig. 6) all soil fractions below skeletal size are represented in percentages identical with those of the soil. The parallel run of these two curves is remarkable, and this observation involves dirt deposited both on the surface and inside snow-drifts. This tends to indicate that, after all, the selective faculty of the wind is very slight; even clay particles occur in wind-borne deposits in

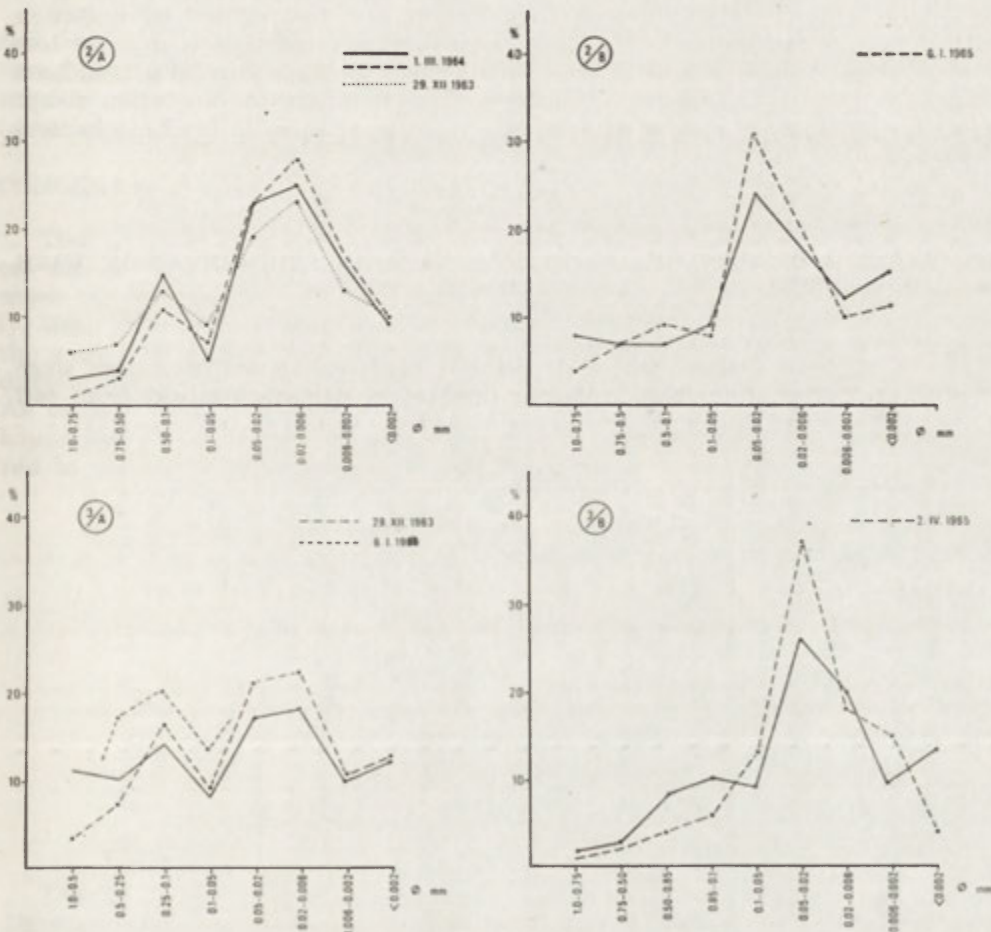


Fig. 6. Chart showing grain size of soil and niveo-eolian material at Stations 2 and 3 for different years. Concurrence of mechanical composition of dirt admixture in the snow with that of the soil

Explanation: continous line-soil, broken line-niveo-eolian material taken from snow surface, dotted line-niveo-eolian material from inside of snow-drift

the same percentage as in the soil. The explanation of this surprising fact is that the clay particles, when carried off, are not separate elements any more, but that they rather occur as aggregates held in the tiny frozen soil lumps which as whole units are swept away by the wind and behave like elements of a coarser fraction.

This similarity between the mechanical composition of the dirt particles (of sizes smaller than skeletal soil parts) and that of the soil material is typical and appears in the greater part of the samples examined. In the dirt admixture of three samples (of a total of 14), the material deposited by deflation contains slightly more coarser fractions than the soil had contained. The reason might be that the finest soil particles may at times be carried off farther — beyond the snow-drifts; but this kind of divergences occur only where bare sites were exposed to strong wind currents.

In the above summary of the results obtained from granulometric analyses attention should be called to differences between what was caused by transport and redeposition due to niveo-eolian deflation, and that caused by deflation of a purely eolian nature. Carried in snowflurries, mineral particles are less easily subject to mechanical selection than when they are carried off without snow. Once deflated from the ground surface, soil particles are often swept away for hundreds of meters without suffering any changes in their mechanical composition.

#### COMPARISON OF DEPOSITS FROM NIVEO-EOLIAN SEDIMENTATION WITH WATER DEPOSITS

In the Bolków Upland where deposits of niveo-eolian sedimentation were investigated, the author also collected samples of deposits derived from soil

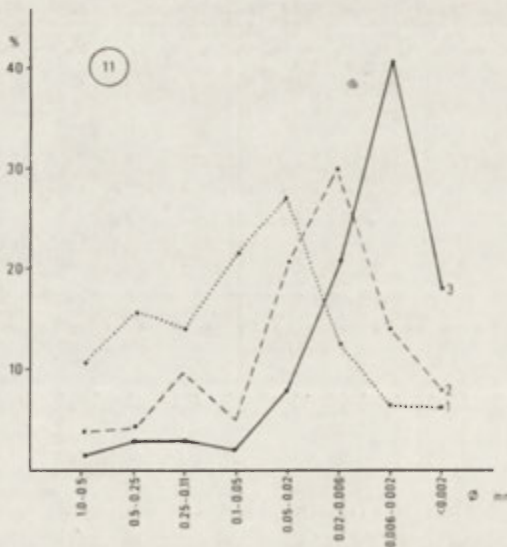


Fig. 7. Granulation curves with differing culminations

1 — sediment of field stream, 2 — niveo-eolian sediment, 3 — sediment of stagnant water



erosion. During a period of heavy rainfall in May 1965 cultivated land was eroded by water running down the slopes.

The quantity of material flushed away by the water was from 4.5 to 5.6 g/l; in part it was redeposited nearby, at the base of the slopes, but part of it was transported farther into retention basin at the outlet of the valley. These deposits of water transport and sedimentation were examined as to their mechanical composition and compared with the soil from which they had been eroded. It appeared that water can segregate the material it carries off much better than wind is capable of doing and that water separates out two groups of deposits: a coarser fraction of more than 0.05 mm diameter which is deposited on slopes, and a finer, clayey fraction with particles of less than 0.006 mm size which is removed and finally forms a deposit in a basin of stagnant water. Hence one notes, that the culmination of the granulometric curves for water deposits differs from that of curves for niveo-eolian deposits (Fig. 7), and this explains the two different curves and the two culminations mentioned.

The wind fails to segregate soil material swept away in clouds of snow, and this is why niveo-eolian deposits are some intermediate form between the two types of water deposits mentioned. The culmination of the curve for wind-borne "dirt" deposits lies between that of water slope deposits and water material deposited in storage basins.

#### CONCENTRATION OF NIVEO-EOLIAN DEPOSITS

The amount of dirt contamination held in the snow depends not only on the process of deflation but also on changes in snowfall and in the snow cover and, first of all, on the duration of the snow cover. Obviously, the concentration of dirt particles in the snow proceeds differently when the snow cover lasts throughout the winter without any marked breaks, and differently when this cover is formed and melts several times during a winter. An example of the first of the two types of snow contamination in the Sudetes happened in the winter of 1963/64, while an example of the second type occurred in the 1965/66 winter.

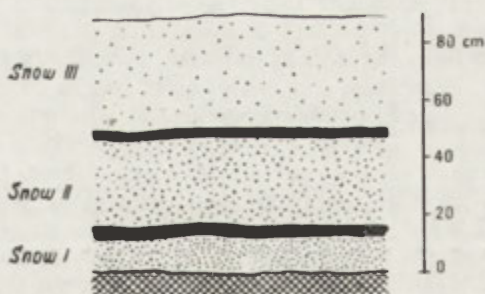


Fig. 8. Profile of snow-drift at Myslow, as determined in spring of 1966

Let us start with discussing the latter event. As early as in November and December 1965 a snow cover had been piled up and, again afterwards in January and February 1966. Interposed between periods of snowfall and low temperatures were long spells of thawing. On the Bolkow Upland the compact snow cover vanished completely several times; in the Kaczawa Mts. remnants of snow were left several times only at places where high snow-drifts had been built. The result was that in March it was easy to observe in cross-sections

and profiles of the surviving snow covers three separate snow layers, differing in their degree of metamorphosis and in colouring. These colours were caused by contaminating particles which usually grew downward in intensity, and therefore were the more concentrated the earlier the snow had been deposited. The author calls these three nival horizons according to their age, first, second, and third snow.

As an example he cites the profile of a snow-drift at locality 11, at Myslow (Fig. 8). Here the dirt concentration was highest in the December snow and lowest in the February snowfall. Between the snow layers dark crusts were distinctly seen — proof of a surface concentration of dirt caused by processes of ablation.

At many further localities these three snow layers were easy to observe. The mean concentration of dirt admixture, calculated for all snow material investigated at several localities, was:

third snow — 142 g/m<sup>3</sup>

second snow — 588 g/m<sup>3</sup>

first snow — 1600 g/m<sup>3</sup>

It is self-evident that with the course of time as a consequence of compaction, ablation and metamorphosis, the concentration of dirt is bound to correspond in direct proportion to growing density of the snow. The first snow which falls earliest and in December contained a dirt concentration of 400 g/m<sup>3</sup>, had four months later a dirt concentration four times higher, having meanwhile turned into an "old" crust-compacted snow.

An example of how dirt concentration has been changing when a snow-drift lasts for a long time, came to view during the 1963/64 winter. The quantity of dirt, in g/m<sup>2</sup> snow surface, was determined first in December and again in March and in April. The result was:

Locality	Dec. 29, 1963	March 1, 1964	April 5, 1964
4	594	812	4400
6	698	368	2012
7	1640	no record	5832
10	615	no record	5200

We note from these figures that the soil dust, brought in by the wind mostly during December, was gradually concentrated at the same rate as snow thawing or evaporating proceeded on the snow surface. The one deviation from this rule which appears at locality 6 on March 1, is most likely due to some error; another point of the snow cover, not the one considered in December, must have been taken into account. At any rate it is plainly visible that in April, after all snow had vanished, the surface of the former snow-drift was covered with a layer of dirt particles which throughout the winter had been permeating the snow pile. Part of the dirt admixture may have been flushed away during the process of gradual ablation, but it is not likely that this was very great. On flat surfaces, water from thawing snow soaks the interior of the underlying snow, like a sponge; the dust particles gradually move downward, remaining in the field of observation.

We are now approaching the problem of dirt concentration in the snow — the object of the theoretical considerations mentioned in the preface to this paper. As a snow-drift vanishes, the surface concentration of dirt grows larger in the form of a "crust" of increasing thickness. This outer dirt layer on top of the snow consists not only of thawed-out snow particles; it is rather a very thin sheet of snow overcharged with contaminating dirt. Hence one may speak

of a concentration of dirt in the crust part of the snow and calculate the degree of this concentration with regard to the snow volume. The crust varies in thickness and, while collecting samples, one may sometimes hit upon a thin section of the crust barely a few millimeters thick or, again, upon a snow crust which is perhaps as much as 0.7 cm thick.

In 1963/64 the mean proportion of dirt concentration in the crust to that inside the drift was 24. Up to March the concentration in the crust had grown to 45 times that inside the snow, and by April this figure was as much as 64 times higher.

One year later, in January 1965, measurements of the initial sedimentation of niveo-eolian material brought the same results as for the 1963/64 winter, as far as dirt concentration in the crust to that inside the snow was concerned (24).

Thus, in conformity with the gradual decay of the snow cover, the top crust accumulates a steadily growing mass of contaminating material, and this quantity is the true gauge for the niveo-eolian processes which are taking place during the whole winter. When the snow cover lasts throughout the winter or when snow keeps piling up continuously in drifts, it suffices to measure in spring the amount of accumulated dirt in order to gain a true picture of the effect which eolian soil degradation has had during the winter period. A close interrelation exists between the degree of dirt concentration in the snow and the thickness of the snow cover on the one hand, and the quantity of dirt particles on the surfaces of spring patches of snow on the other. In the 1963/64 winter which brought a high intensity of niveo-eolian processes, the quantity of dirt accumulated in the snow that had been covering ploughed fields amounted to 2–15 kg/m<sup>3</sup>, or 7.3 kg/m<sup>3</sup> on the average. Assuming a layer of winter snow 0.5 m thick (depending on local measurements), the so-called "ablation crust" on top of the spring patches of snow should amount to 3.6 kg/m<sup>2</sup>, — and this is the value obtained from theoretical figuring, based on the understanding of a process of an undisturbed geometric concentration. The actual result brought about by the concentration of a dirt crust on top of the snow, established in the field during the spring of 1964, was 4.3 kg/m<sup>2</sup>. The difference between the value determined by measurements of the dirt and by the 3.6 kg/m<sup>2</sup> index of ablation can easily be explained: when in December 1963 the winter set in, a crust accumulation developed on top of the snow which, according to the measurements mentioned above, contained an average of 0.8 kg dirt per m<sup>2</sup> snow surface. Hence, if we subtract from the spring quantity of dirt concentration which was 4.3 kg/m<sup>2</sup> and represented the total niveo-eolian sedimentation, the above 0.8 kg/m<sup>2</sup> we arrive at the true index of ablation, i.e. 3.5 kg/m<sup>2</sup>—this figure differs by barely 0.1 kg/m<sup>2</sup> from that obtained from theoretical calculations.

Here one arrives irrefutably at an important conclusion which throws light upon the mechanics of the way niveo-eolian sedimentation has been taking place in 1963/64. Much of it developed by snow thawing or evaporating, mainly during the last phase of snow decay in the spring. And for the most part the dirt sedimentation goes back to concentration by ablation, not to straight accumulation. A bare 19% of the deposit (0.8 kg/m<sup>2</sup>) is material deposited on the snow, and this deposition occurred only at the beginning of the winter; later on processes of soil degradation due to wind action practically ceased.

Taking the winters of 1963/64 and 1965/66 as an example, one can distinguish two principal types of niveo-eolian dirt sedimentation: a one-cycle and a two-cycle type. It depends on temperature, precipitation, and general weather conditions which type of sedimentation will be taking place during a winter

season. It seems that under the climatic conditions of Poland's mountains the multicycle type of sedimentation occurs more often. However, irrespective of cyclicity of sedimentation, its features are always similar. Dirt concentration by ablation prevails over straight accumulation. As put forward in our theoretical considerations, the crust-shaped concentration by ablation corresponds to the decrease in snow thickness, and therefore one might say that the value expressing this degree is the multiplier of the degree of concentration.

#### VALUATION OF THE RESULTS OF NIVEO-EOLIAN SOIL DEGRADATION

The dirt content in the snow, even when very accurately determined each year, do not permit a direct estimate of niveo-eolian soil degradation. To achieve this, one has to determine the soil surface from which deflation has been taking place, and this is by no means an easy matter. Dirt deposition occurs mainly at obstacles on the land surface at which the wind force is lessened. Starting out from the assumption — which indeed is a simplification of actual field conditions — that the snow-drifts piled up along snow fences hold all the snow (including the dirt it contains) which has been deflated from a field, we shall have to base our calculations of the effect of deflation on the size of that field. Our observations show that snow accumulating in drifts is brought in from distances of about 100 to 300 m. This same estimate can be arrived at from the following calculations: during the snowy winter of 1963/64 the mean width of the snow-drifts was 10 m and their mean height 0.5 m. The snow density in the snow-drifts is about ten times the density of fresh-fallen snow. However, considering the size of the drift and the density of the snow, we find that the area of deflation from which the drift was built — assuming the snow cover on the field surface to have been 0.5 m high — must have been 100 m long or, for a snow thickness of 0.25 m, 200 m long. Because a snow-drift piled up next to a snow fence never contains all the snow that had been covering the field, and part of this snow is retained in plough furrows and along boundary strips, or is carried off beyond the fence, it is rather the lower figure of the calculated area of deflation (100 m) which should be considered the index figure for the area from which deflation has formed a snow-drift 10 m wide. This would indicate that under average conditions the area of accumulation corresponds to one tenth the area of deflation.

This 1:10 proportion applies to summit areas and to ploughed fields exposed to the full action of air currents, hence rather flat surfaces lacking larger obstacles. This proportion holds good for niveo-eolian processes in their widely conceived operation. However, this proportion may vary, may even turn into the very opposite so that the area of accumulation may even be greater than the area of deflation when niveo-eolian processes active within micro-forms of the land relief are taken into consideration. As an example may be cited the deflation of dust from the tops of hillocks, from boundary strips of fields, etc., as has been illustrated in Figs. 6 and 7. This sort of short-distance transport and nearby deposition evokes complications and deformations in the general picture of niveo-eolian degradation of ploughed fields which most often constitute our zones of deflation. Sedimentation of the deflated material usually takes place at the margins of fields where the soil area exposed to the wind ends, or at greater obstacles where drifts of considerable size are apt to be formed.

Hence, if in the 1963/64 winter on ploughed fields an average of 4.3 kg dirt was contained in 1 m<sup>2</sup> area of accumulation, this dirt was derived from a 10 m

surface of deflation. In consequence we note that, on the average, soil deflation from fields has amounted to 0.43 kg or 430 g/m<sup>2</sup>, which would cause a lowering of the soil surface by 0.024 mm (assuming the specific gravity of soil to be 1.8). This figure applies to winters where niveo-eolian processes were exceptionally powerful. The mean quantity of niveo-eolian sedimentation taking place during an average winter in the Sudetes is, at the most, some 0.5 kg/m<sup>2</sup> as has been shown above. This corresponds to a soil degradation of 0.05 kg or 50 g/m<sup>2</sup> and to a lowering of the soil surface by barely 0.0027 mm.

The true meaning of these figures for niveo-eolian deflation can be perceived by comparing them with indices referring to aqueous denudation. Taking into consideration the deposition of material in the storage basins of the Sudetes, one can calculate that this sedimentation is of the order of 0.05 to 0.10 mm per year (A. Jahn 1968). This figure comprises the whole mountain area, no matter what its cover. In comparison with these indices, winter deflation of soil is surprisingly small — on the average 200 times less than aqueous denudation. Moreover it must be kept in mind, that the indices obtained from calculating niveo-eolian denudation referred merely to ploughed fields bare in winter, which constitute scarcely one half of the morphological ground surface, so that winter denudation is in reality much smaller yet. Finally it seems worth pointing out, that niveo-eolian deflation does not cause any loss of soil material as is the case with the action of water, but only the displacement of soil from one place to another, from open fields to points shielded against wind action. We see therefore that in sum total niveo-eolian processes, considered as destructive agencies, are very much less harmful than the effect of torrential rains.

All that has been discussed above refers only to mountainous regions, and is exclusively an estimate of eolian soil degradation during wintertime. The part which eolian degradation caused by winter processes has during the full year is unknown; this is the reason why, due to the lack of definite data, opinions vary on this subject. Generally speaking, farmers believe the effect of eolian erosion upon fields to be highest in spring and autumn, at the periods in which the soil surface is deprived of its protective cover of vegetation or snow. On the other hand, everybody knows that in the Sudetes the winds are strongest in winter and that during the cold season eolian degradation is the most effective. Far from attempting to solve these problems I wish to express the opinion that during the warm season eolian soil degradation is at least the same, or perhaps as much as twice that of wintertime deflation. Of course, compared with aqueous degradation this claim does not affect the conclusion formulated above that was demonstrated aqueous processes are very strong, perhaps hundreds of times more powerful than eolian erosion. This assertion, while undoubtedly correct, refers only to mountain regions where natural condition rather favour aqueous erosion. The steep scarps, the poorly permeable soil, the abundant precipitation — all these factors are favourable to aqueous soil erosion.

Often repeated nowadays are warnings which call attention to damages inflicted on agriculture by eolian processes; but these alarms refer exclusively to lowland areas with light soils such as predominate in Pomerania and Lower Silesia. These misgivings are justified and, although not yet fully supported by suitable data of observations, demand prompt and detailed investigation. The proportion of eolian erosion to aqueous denudation is different on flat surfaces from that on mountain areas — mainly due to the fact that the absence of steeper slopes in the morphological land surface prevents the evolution of destructive effects of flowing water. Against this background problematic terms such as "rural steppes" or "desert conditions of agricultural regions" assume

a specific meaning — referring to problems which C. Samuelsson (1927) put forward 45 years ago in his caustic and exaggerated forecast.

## RESULTS OBTAINED, AND FINAL CONCLUSIONS

The joint action of wind and snow or, as it is called, the niveo-eolian process can be examined from the viewpoint of either dynamic geomorphology, or sedimentology including Quaternary geology. In the present paper, the author gives his critical examination of the niveo-eolian phenomena occurring under conditions of a forced evolution, as observed in the mountain and submountain area of the Sudetes with its well developed agriculture.

The results of nival deflation depend, first of all, on the vegetation which happens to cover the soil surface. These results are many times more impressive on ploughed fields than on meadows or in forests. The vegetation cover supplies full protection while, at the same time, is itself exposed to destruction; here agencies of geology and pedology come into play. Under conditions as they are in the Sudetes, the soil overlying palaeozoic shales is easily deflated. In this respect, local morphological features are of marked importance: widespread summit surfaces and slopes exposed to the force of the wind are particularly subject to niveo-eolian degradation. The intensity of snow contamination is consistent with the system of existing air currents — a trait imposed not only by general tendencies of atmospheric conditions but by the local features encountered in a mountain area as well.

In their harmful action upon the soil surface the niveo-eolian processes cause to some degree a mechanical segregation of the soil material. While carrying off the snow, the wind deflates from the soil only that part of the material which is of less than skeletal size; the wind transports these particles as they are and deposits them somewhere, usually without any sort of segregation. For the majority of samples examined, the size distribution curve of this deposit shows a shape identical to that of the soil composition. However, where in the early part of winter the field surface is freely exposed to strong winds, a slight niveo-eolian segregation of the sediments does take place, and one observes in the size composition an increase in coarser fractions at the expense of the group of fine dusts.

Definite evidence is available showing that wind is less capable than flowing water of segregating the material it carries. Measurements of the effect of both these agencies, made in the same region for identical soils supplied proof that in contrast to wind, water can segregate in a subtle way the material flushed off from fields into a coarser fraction (sand and silt) which it deposits at the base of slopes, and into a fine fraction (floatable particles) which it transports farther off into larger storage basins. This double culmination in the grain composition curve for aqueous sediments does not tally with the usually single culmination curve of eolian deposits.

The wind picks up particles of snow and soil, carries them off and redeposits them. A sedimentation series shows in its profile that the dirt contamination held in the snow grows upward in intensity. During a single winter the sedimentation of this material may proceed in one or in a number of cycles. Segregation and redeposition of the dirt admixture takes place by gradual concentration within the snow. The degree of this concentration grows in direct proportion to the density of the snow. The process of snow ablation (thawing and evaporating) leads by ablation to a concentration of the dirt particles and, finally, to the formation of an ablation crust of dirt upon the snow surface.

The growth in thickness of this crust corresponds to the gradual decrease in thickness of the given snow pile.

On the surface of snow-drifts the quantity of dirt contamination, accumulated during one winter in the Sudetes averages about  $0.5 \text{ kg/m}^2$ ; the highest known quantity of this kind of sedimentation<sup>3</sup> is as high as  $4\text{--}5 \text{ kg/m}^2$ . The index of soil degradation corresponding to this figure is equal to a lowering of the soil surface during one winter by  $0.0027$  to  $0.024 \text{ mm}$ . This figure is many times smaller than the index of aqueous degradation observed in the Sudetes which varied from  $0.05$  to  $0.10 \text{ mm}$  per year.

As a rule, some results of niveo-eolian processes can easily be observed everywhere. They are conspicuous by their characteristic expressiveness, commonly appearing in spring as dull-coloured dirt-contaminated snow piles scattered over field areas. Usually their impression is more striking than the actual damage caused by these processes in the mountains, because steep slopes and high precipitation are propitious to aqueous soil erosion which greatly surpasses any damage done by wind action, not only in quantity but also in quality (segregation of material, flushing away of fine parts). Notwithstanding all this, the harmful character of niveo-eolian processes should be taken into account and counteracted as much as possible. In the Sudetes the element responsible for this kind of damage is the powerful foehn winds arriving from the south. They are dangerous to the extensive summit areas, especially where the slope surfaces are exposed to these southern winds. Under the morphological conditions mentioned, the destructive effect of winter deflation is greatest on ploughed fields devoid of obstacles to the wind, natural or artificial. This implies the necessity of midfield planting of tree belts, for which use could be made of existing boundary strips which now for the most part are devoid of bushes and trees. This particularly refers to regions where the soil of the fields rests on palaeozoic shales and sandstones, on conglomerates or on tuffstones of the Rotliegenden.

Much greater than in the mountains are the damages to be expected from niveo-eolian processes in flat lowland areas where light soils predominate, mostly sands and silts. It seems, that in examining problems of erosion we are too strongly impressed by dangers that might be caused by water from precipitation, and that we are devoting an excess of efforts, studies, meetings and consultation to this problem, whereas evidence is available that where steeper slopes are lacking and the layer of good soil is thin ("problems of steppe-formation") the wind may turn into a much more effective cause of erosive soil degradation than flowing water. In considering this wind action, particular attention seems to be due to the niveo-eolian processes discussed above which so far have been little investigated, mostly in a rather desultory fashion. The present paper which has been based on methodical observations and measurements, represents an attempt at examining certain details, and of presenting certain quantitative data, regarding the niveo-eolian processes operating in Poland's mountain regions.

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<sup>3</sup> In February 1956 a heavy erosion of loess and sand soils occurred on the Lublin Plateau, in the eastern part of Poland. At that time the amount of sand and loess material on the snow surface locally reached the value of  $10,000 \text{ t/km}^2$  (M. Strzemiński 1957). This was a phenomenon which revealed a removal of 4000 tons of dirt per  $1 \text{ km}^2$  in the Sudetes.

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## COMMENTS ON KARST DENUDATION IN BULGARIA

MARIA MARKOWICZ, VLADIMIR POPOV AND MARIAN PULINA

### OUTLINE OF CONTENTS

The authors use as background a brief comment on the geology, morphology and hydrochemical features of the three largest karst regions of Bulgaria, to present the results of their investigations made with regard to the intensity of contemporaneous degradation observed in these regions. In their paper the authors also describe the new methods they have used to examine the processes by which dissolution of carbonate rocks is taking place.

The rock areas subject to karst processes in Bulgaria cover about 25,170 km<sup>2</sup>, equalling 22.7% of the country's surface area (Fig. 1). The concentration of carbonate rocks is greatest:

(1) in the Danube plain and in Dobruja which are both built of Tertiary (Sarmatian) and Lower Cretaceous limestones (this area takes in about 66.6% of all Bulgaria);

(2) in Stara Planina built of limestones of the Lower Cretaceous, the Jurassic and Triassic (19.8%);

(3) in the Rhodope and the Rila Mts. where proterozoic and palaeogenic limestones prevail (7.8%).

In the summer seasons of 1967 and 1969 the authors made their geomorphological observations and their hydrochemical examinations in several typical karst regions of Bulgaria<sup>1</sup>. The purpose of these studies was to determine the contemporaneous intensity of karst denudation. These sorts of studies were made in the Pirin range (the Vihren Massif), in Stara Planina (the Vratsa Planina and the drainage basin of the River Panega), and at the Black Sea coast between Varna and Cap Kaliakra.

### THE PIRIN MASSIF

Proterozoic marble rocks are exposed on the flanks of the Central-Pirin Arch, the core of which contains rocks of a granite intrusion (Boyadzhiev 1957). The upper boundary of these marbles, at their contact point with the crystalline rocks, extends along the western slopes of the highest Pirin ridge. The lower, eastern boundary on the other hand is marked by the East-Pirin fault which

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<sup>1</sup> In 1969, in the investigations of the Pirin Massif and of Vratsa Planina, Mr. J. Jasiński from Kletno took part.



Fig. 1. Karst areas in Bulgaria

1 — boundary of geomorphological areas: I — Danube Plain and Dobruja Platform, II — Stara Planina, III — transitory zone, IV — Rila and Rhodope Mts. Exposed karst zones occur in the following carbonate rocks: 2 — proterozoic marbles, 3 — Triassic limestones, 4 — Jurassic limestones, 5 — Cretaceous limestones, 6 — Aptian limestones, 7 — Tertiary limestones, 8 — Sarmatian limestones, 9 — corved (buried) karst, 10 — caverns.

Karst springs discharging: 11 — up to 100 l/s, 12 — 100 to 1000 l/s, 13 — more than 1000 l/s  
 Situation of karst areas examined: A — Pirin Mts., B — Vratsa Plaina, C — drainage basin of Panega River, D — Dobruja

runs between the Pirin base and the depression of Bansko villaga. The contact zone between the marbles and the Pliocene sediments which fill the depression is shrouded by Pleistocene cones built from a marble breccia. Marble rocks also form the highest Pirin peaks: Vikhren (2915 m) and Kutelo (2908 m), and others. This marble reaches thicknesses up to 250 m; the marble bed subsides in ESE to NE direction at an angle from 45° to 80°. In its structure this marble is unequigranular, with a high degree of purity (93.3%  $\text{CaCO}_3$ ); it is greyish-white in colour and distinctly stratified.

The area under discussion is part of the Vikhren-Suchodol geological unit, covering the eastern and north-eastern slopes of the highest Pirin ridge with its altitudes ranging from 920 to 2915 m. From the south this area borders upon the Bistra River and from the north upon the Byala River (Fig. 2). The area covered by marble outcrops measures about 73 km<sup>2</sup>.

The present-day Pirin relief is the result of protracted processes which were degrading the Younger Tertiary and which suffered breaks due to intensive

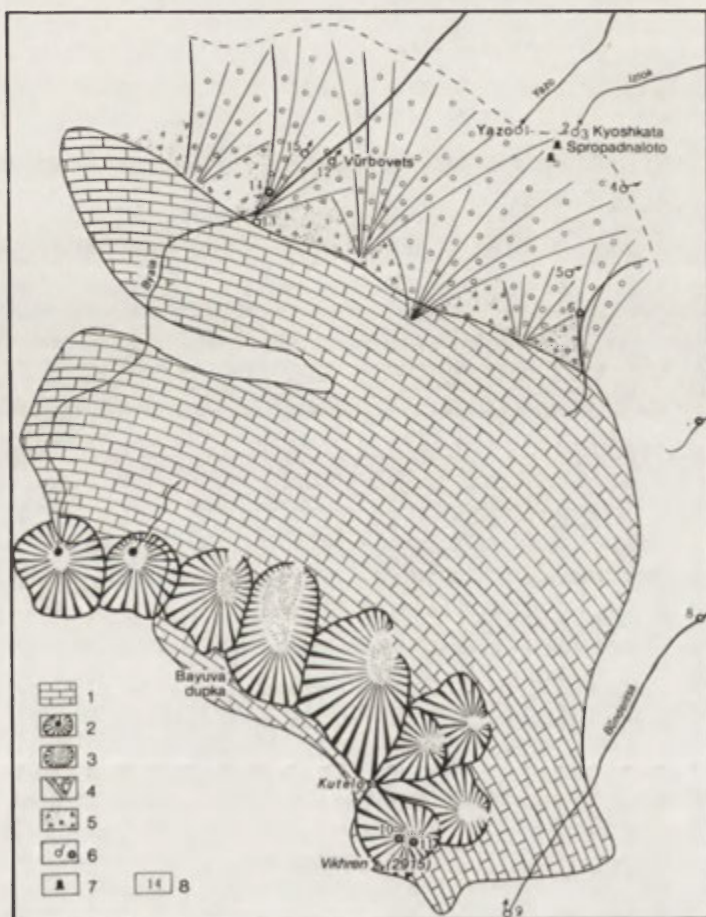


Fig. 2. Morphological map of Vikhren-Suchodol part of Pirin Mts.

1 — proterozoic marble, 2 — glacier cirque with lake, 3 — glacier cirque with perennial snow cover, 4 — cones in Pirin forefield built of old-Pleistocene breccia, 5 — deluvial deposits, 6 — karst springs and localities where water samples were taken, 7 — caverns, 8 — water samples, numbered as in Table 1

neotectonics and to the Ice Age. Scientists who investigated the Pirin range (Jankowicz 1903, Radev 1920, Louis 1930, Popov 1962) distinguished here three planation surfaces which tally with the Western European system. In the investigated part of Pirin, traces of a Miocene surface extending at 2500–2700 m altitude can be clearly observed. These traces can be seen in the eastern slope of the Vikhren and Kutelo peaks and in the upper series of the Kamenitsa, Kotechkiya, and other mountain chains. A Pliocene planation surface has survived at altitudes from 1200 to 1600 m.

During the Ice Age, certain fragments of the upper planation surface were represented by firn fields of glaciers of which some tongues descended to 1700 m altitude. The most imposing glacier cirques here are the Large and the Little Kazan (2200 m high) near Vikhren peak (Fig. 3).



Fig. 3. Pirin-area of marble rocks. On the left Vikhren (2915 m), with Large Kazan glacier cirque

Karst phenomena are in evidence in the Pirin range principally through the absence of a hydrographic network in the land surface. The glacier cirques in the marble rocks are dry. Water from precipitation and snowmelt disappears down swallow holes and sinks to find its way into the well-jointed marbles; in its further course the water follows the slant of the rock beds and re-appears in karst springs at the base of Pirin, at altitudes from 920 to 1100 m. The zone from which these springs are fed extends from 2200 to 2900 m altitude; this indicates that the vertical water circulation covers an altitude difference of the order of 2000 m.

The result of the denuding action of this water, visible on the surface, are numerous karst microforms which here failed to attain their full development. This is probably due to intensive frost weathering caused by the local specific microclimatic conditions. Of mesoforms worth mentioning are enclosed rock basins each some 50 or so metres in diameter, sinks in the bottoms of the

cirques, shallow wells, etc. So far, no larger caverns have been discovered in this region.

Among the 15 springs which the authors examined, two proved to yield the greatest discharge: the Yazo spring (No. 1 on Figs. 2 and 4), with a mean annual flow of  $2.3 \text{ m}^3/\text{s}$ , and the Kyoshtata spring (Nos. 2 and 3) yielding  $0.5 \text{ m}^3/\text{s}$  (Table 1). In spite of the similarity of the discharge curves for these springs (Fig. 5) they differ in type. The Yazo spring where the annual amplitude differences are smaller (only up to 2.5 times) is fed directly from a main karst tunnel; the Kyoshtata spring, on the other hand, with a very high amplitude (up to 17 times), drains a later karst channel which predominantly carries water from heavy rains. For the entire watershed the mean annual unit flow is estimated to be  $31.6 \text{ l/s/km}^2$ .

The Pirin artesian wells are of the high-mountain type. Their most characteristic features are as follows:

(1) spring and summer are periods of peak discharge (Fig. 5), and jointly they represent a period of maximum flow, lasting some four months; the flow is lowest in winter;

(2) in spring the quantity of flow rises abruptly and after passing its peak it drops gradually until it reaches its minimum shortly before winter;

(3) on an annual scale, the chart indicating spring discharge does not tally with the chart of precipitation; the maximum spring discharge rather coincides with the period of lowest precipitation — evidence that this type of spring is mainly fed by meltwater flow;

(4) the water from these springs has a relatively low mineral content and does not leave any carbonate deposits.



Fig. 4. Yazo karst spring in Pirin pediment

TABLE 1. Chemical denudation of Bulgaria's karst areas

Karst area Springs	m <sup>3</sup> /s	l/s/km <sup>2</sup>	P km <sup>2</sup>	E dcm/y.	Precipi- tation mm/y.	Evapo- ration mm/y.	Mineralization mg/l ppm			Denudation m <sup>3</sup> /km <sup>2</sup> /y.		A <sup>6</sup>	
							ΔT	Ta	T	D	X	m <sup>3</sup> /y.	%
Piryn	2.309	31.6	73	10.33	1523	490 <sup>2</sup>	125 <sup>3</sup>	7	118	47.0 <sup>4</sup>	51.7 <sup>5</sup>	3,433	100
mean annual							140	10	130	51.8		3,782	100
Spring No. 1 <sup>1</sup>	1.428						120	7	113			2,033	59
Yazo							141	10	131			2,357	62
Spring No. 2 and 3	0.474						123	7	116			693	20
Kyoshkata							130	10	120			717	19
Other springs	0.407											707	21
Nos. 4-15												708	19
Piryn		39.5	73				125	7	118	58.8		4,292	100
Aug. 1969	2.887						140	10	130	64.8		4,729	100
Spring No. 1	1.780						120	7	113			2,534	59
Yazo							141	10	131			2,938	62
Spring No. 2 and 3	0.990						123	7	116			1,447	34
Kyoshkata							130	10	120			1,497	32
Other springs	0.117											311	7
Nos. 4-15												294	6
Vratsa Planina	0.550	10.5	52.5	2.02	792	590	300	10	290	38.4	24.4	2,010	100
mean annual							360	20	340	45.0		2,356	100
Spring No. 27	0.339						300	10	290			1,239	62
Bistrets							390	20	370			1,580	67
Other springs	0.211											771	38
Nos. 20-26												776	33
Vratsa Planina	0.103	1.96	52.5				300	10	290	7.2		376	100
Aug. 1969							360	20	340	8.4		441	100
Spring No. 27	0.042						300	10	290			153	41

cont'd													
Bistrets							390	20	370			196	44
Other springs	0.061											223	59
Nos. 20-26												245	56
Dobruja							350	18	332	23.0		1,360	100
Kavarna	0.325	5.5	59	0.3	440	410	470	30	440	30.5	4.2	1,802	100
mean annual													
Spring No. 28	0.058						358	18	340			248	18
Kavarn. dol							516	30	486			355	20
Other springs												1,112	82
Nos. 28-44												1,447	80
Spring No. 45							185	10	175			5,479	100
Glava Panega	2.485						223	20	203			6,356	100
Aug. 1969													
River Vit (No. 47)	1.308	(53%)					95	10	85			1,401	26
Aug. 1969	1.376	(55%)					111	20	91			1,577	25
Karst waters	1.177	(47%)					285	10	275			4,078	74
(No. 46)	1.109	(45%)					362	20	342			4,779	75
Aug. 1969													

<sup>1</sup> Numbering of springs as in Table 1

<sup>2</sup> Open air evaporation was calculated by Dr. S. Lingova (Hydro-Meteorological Institute, Sofia) by method of water and thermal balance calculations

<sup>3</sup> In column showing water mineralization the carbonate contents in  $\text{CaCO}_3$  ( $\text{Ca}^{++} + \text{Mg}^{++}$ ) is given above, and full mineralization (i.e., dry residue) below

<sup>4</sup> Denudation calculated by Pulina's hydrometric methods  $D = 0.126 \cdot \Delta T \cdot q$

<sup>5</sup> Denudation calculated by Corbel's climatic method  $X = 4ET/100$

<sup>6</sup> A — runoff of dissolved salts calculated by Pulina's hydrometric methods  $A = 12.6 \cdot \Delta T \cdot Q$

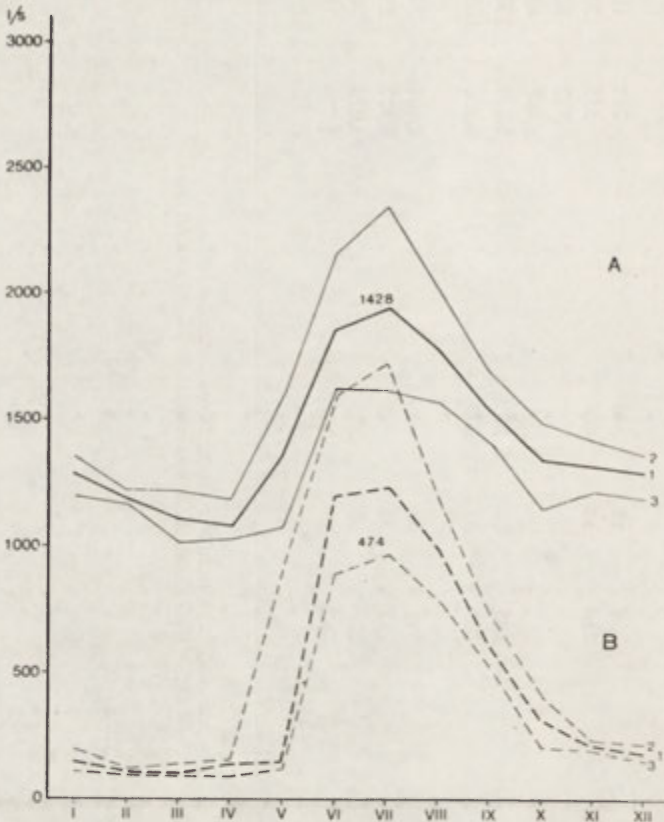


Fig. 5. Mean monthly outflow from Pirin karst springs, in 1965-1967

A — Yazo spring, B — Kyoshkata spring, 1 — mean flow, 2 — maximum flow, 3 — minimum flow

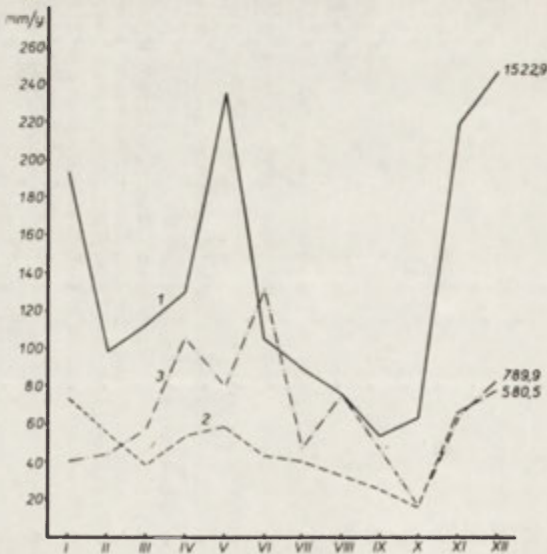


Fig. 6. Mean monthly atmospheric precipitation, in 1965-1967

1 — Virkhren tourist shelter (2060 m), 2 — Bansko (936 m), 3 — Bratsa (309 m)



The Pirin range has a mountain and high-mountain climate; it belongs to the Mediterranean continental climatic zone (Sybyeve, Stanev 1959). Characteristic data on some meteorological elements are given in Figs. 6, 7 and 8, in which the authors present records from two meteorological stations: Bansko

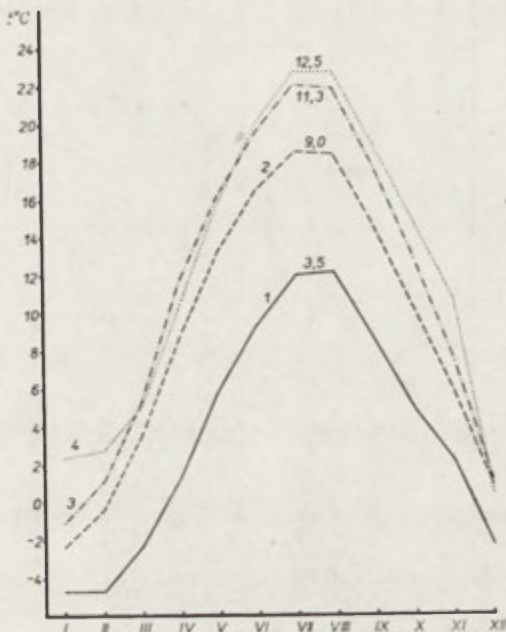


Fig. 7. Mean monthly air temperature in 1958-1968

1 — Vihren tourist shelter (2060 m), 2 — Bansko (936 m), 3 — Vratsa (309 m), 4 — Tuzlata (20 m)

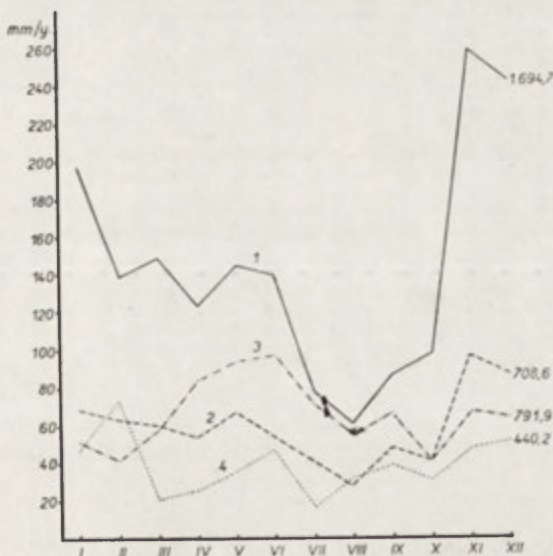


Fig. 8. Mean monthly atmospheric precipitation in 1958-1968

Explanations as for Fig. 7

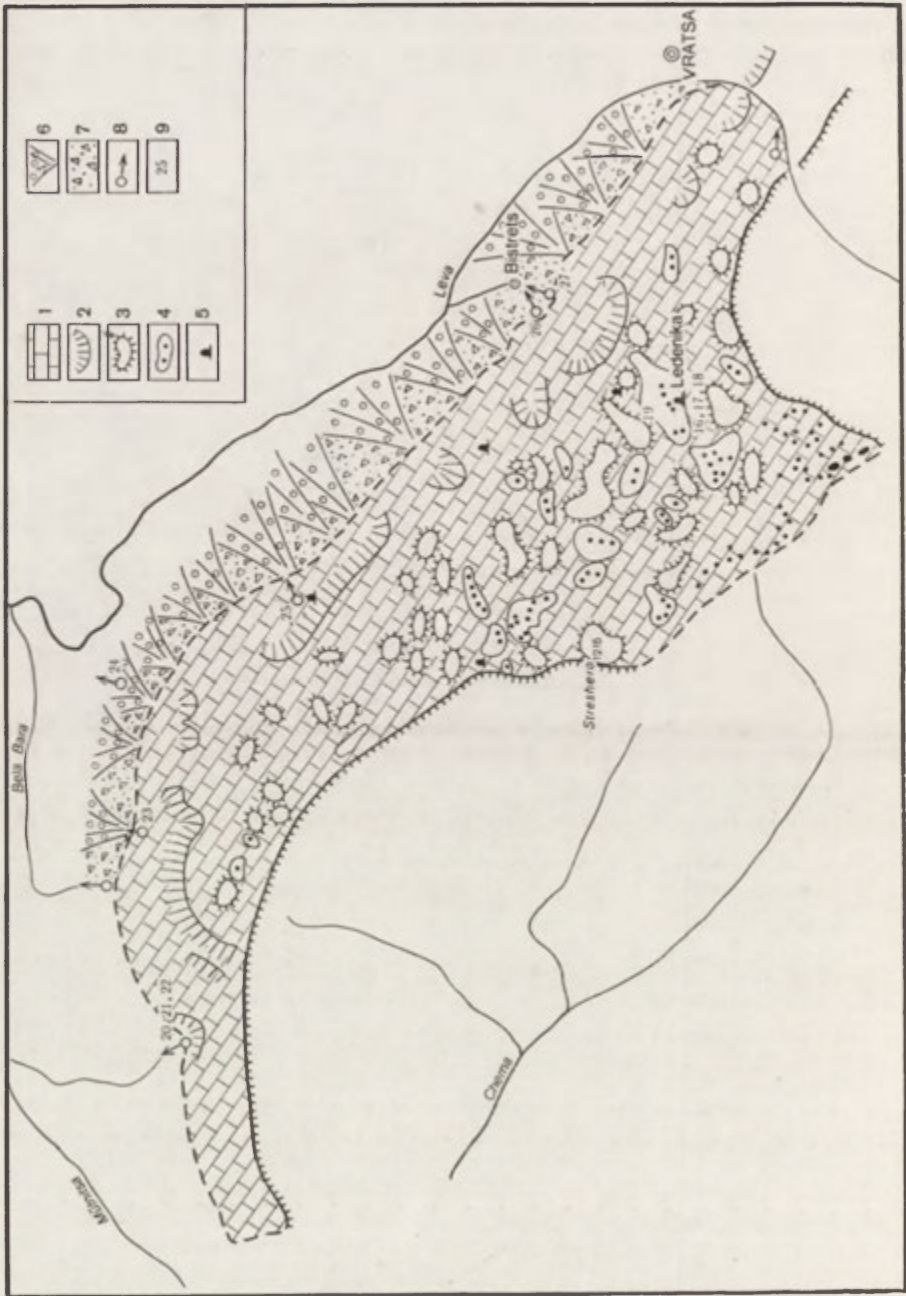


Fig. 9. Karst forms in a part of Vratsa Planina

1 — Titonian and Aptian limestones, 2 — rock walls, 3 — remnants of heaped-up karst, 4 — valoses and uvalas with sinks and potholes, 5 — caverns, 6 — cones, 7 — deluvial deposits, 8 — karst springs, 9 — water samples, numbered as in Table 1

(936 m a.s.l.) and Vihren (2060 m a.s.l.). The mean annual sum of precipitation is highest in winter — 35% (November alone has 260 mm) and in spring; it is lowest in summer, with only 28–61 mm in August. In glacier cirques the snow cover lasts 180 days (Popov 1962).

### VRATSA PLANINA

In their investigations the authors covered the north-western part of this area (Fig. 9) which northward borders on the Vratsa basin, southward on the Botuna valley and the valley of Cherna, its right bank tributary, and eastward on the deep Leva valley with its picturesque Vratcata gap.

The limestone outcrops are connected with the Starigrad Arch whose core consists of impermeable rocks of Carboniferous and Permian age. These older rocks are capped by mesozoic sediments among which are mainly represented limestones of the Upper Jurassic (Titonian) and the Cretaceous (Urgonian). In the Strescha part of this planina, covering the northern flank of the Arch the limestones dip in a northward direction. Near Ledenika Cave, situated in the

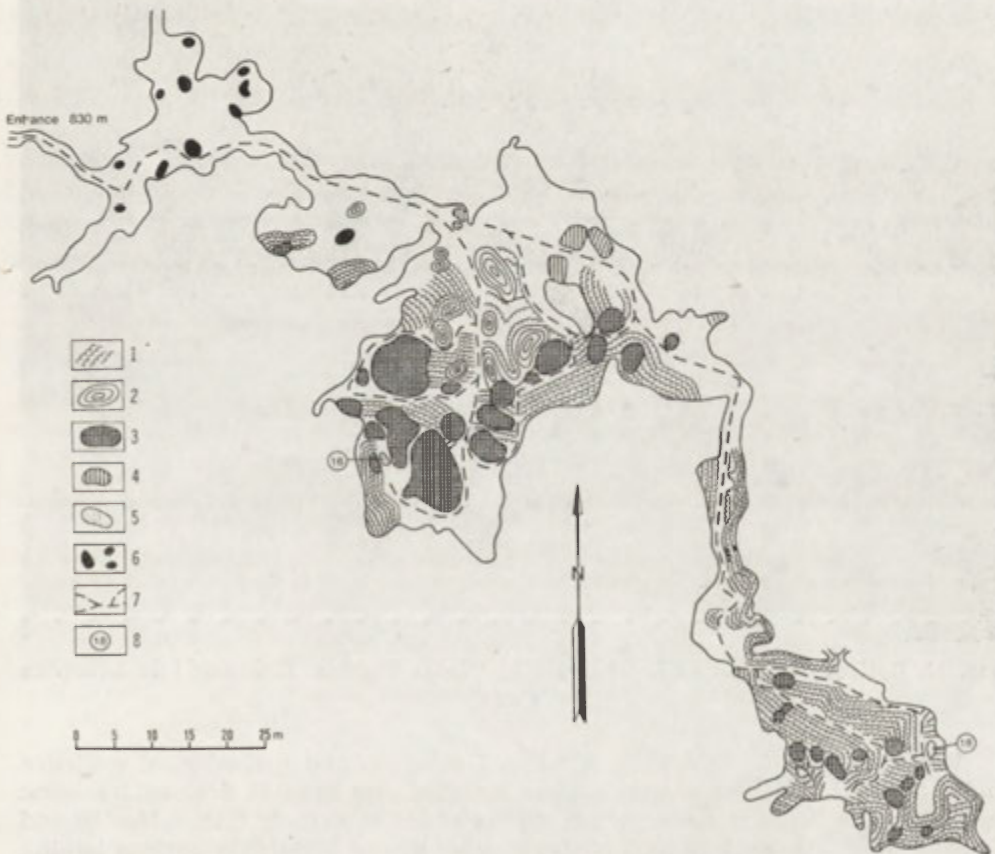


Fig. 10. Map of Ledenika Cave

1 — calcite incrustations, 2 — stalagmite, 3 — calcite incrustations forming a column, 4 — limestone block, 5 — lakelet, 6 — ice incrustation, 7 — path through the cave, 8 — numbering of water samples as in Table 1

central part of Vratsa Planina, this dip is  $26^\circ$  while along the northern rim of the planina the anticline wing dips at an angle of  $86^\circ$ . Here the greatest areas are covered by Urgonian limestones, the high purity of which reaches values of some 89.3 to 91.3%  $\text{CaCO}_3$ .

The Vratsa karst was mentioned for the first time by F. Toul and K. and H. Skorpil (1895). Later examinations made by Radev (1915) and by Mishev and Popov (1958) revealed well developed karst macro- and mesoforms occurring on an old-Pleistocene planation surface extending at 900 to 1000 m altitude. Protruding from this surface are fragments of dome-like limestone elevations which locally reach as high as 200 m. They may be considered evidence of the Tertiary karst relief which may have included towering mogote-type forms. Among land forms of surface karst, the most conspicuous forms are uvalas, valoses,<sup>2</sup> sinks, and areas of rock ribs some 2 m deep. The largest of these forms is an uvala called Ledenika Cave extending from 2260 to 1200 m. Apart from the forms of surface karst discussed, caverns are known of which the most renowned are: Ledenika<sup>3</sup> 263 m long (Figs. 10 and 11), Belyar 236 m deep and Haydushkata 108 m deep.

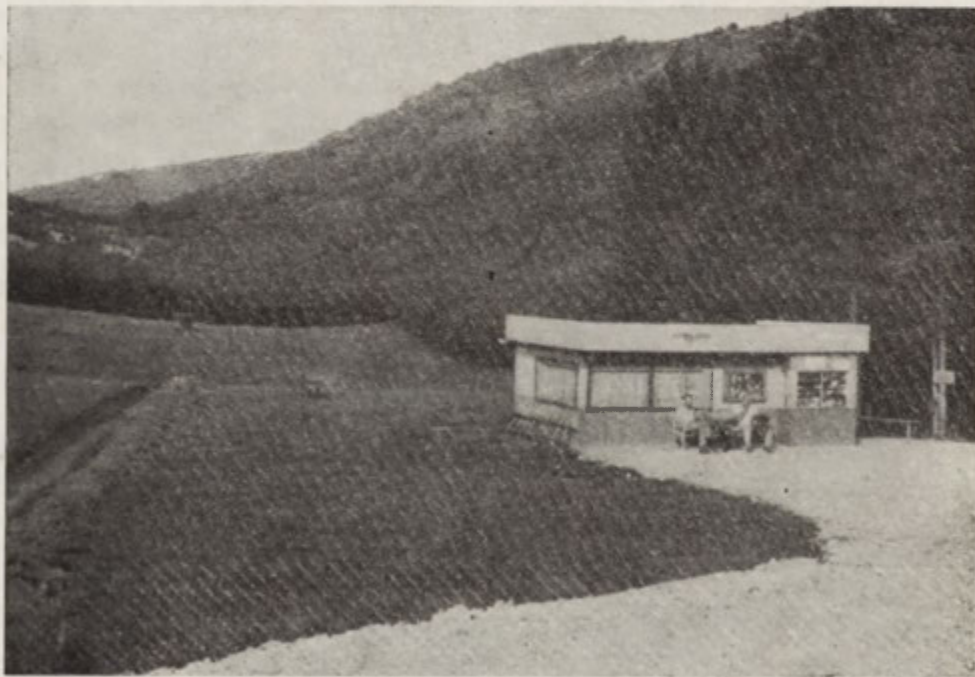


Fig. 11. Bottom of Ledenika karst uvala in Vratsa Planina. Entrance into Ledenika Cave

Vratsa Planina lacks surface streams. Geological and speleological evidence indicates that the central and eastern part of this area is drained by karst springs of the Bistritsa River which appear at 300 m altitude (Fig. 9, Nos. 26 and 27), and the north-western part of this planina by the Mytnishke springs issuing

<sup>2</sup> Valoses — a transitional form between uvala and karst sink.

<sup>3</sup> Ledenika Cave is of interest to tourists, with some 80,000 visitors annually.

at 237 m altitude (Nos. 20, 21, 22), by springs situated near the St. Ivan Rilski monastery. at 500 m altitude (No. 25), and by springs discharging their water near Byely Izvor village at 250–370 m altitude (Fig. 9, Nos. 23, 24). From these calculations one arrives at the conclusion that the underground water circulation involves a vertical altitude difference of some 900 m height.

Among the eight springs which the authors examined at the base of Vratsa Planina, only two show a fairly abundant discharge. They are the two Bistrets springs the larger of which has a mean annual discharge of 339 l/s. This spring represents an artesian well in mid-mountain altitude which is fed by both rainwater and melting snow. Its characteristic features are as follows:

(1) the flow reaches its maximum in spring (from February to June), and its minimum towards the end of summer and in autumn (Fig. 12);

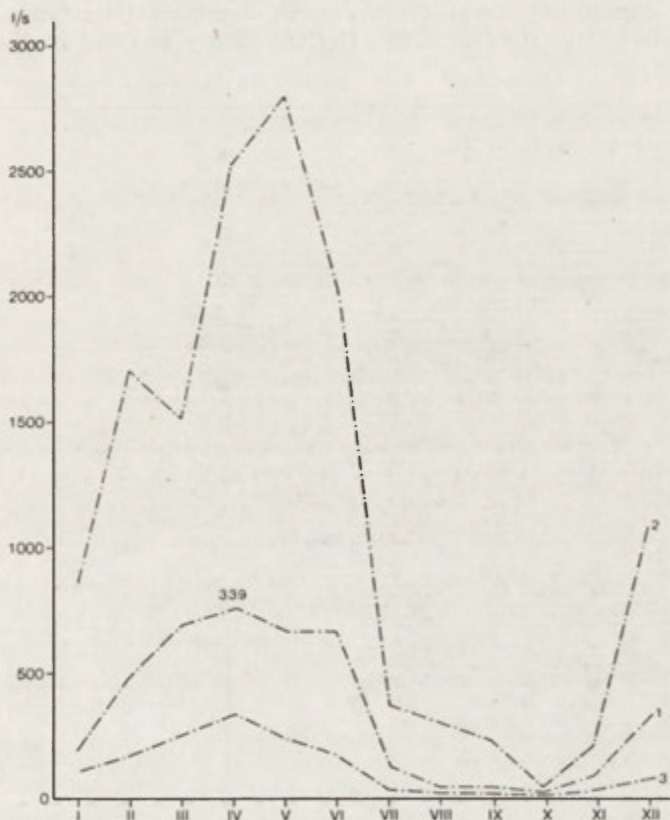


Fig. 12. Mean monthly discharge of Bistrets karst spring in Vratsa Planina, in 1965–1967

1 — mean discharge, 2 — maximum discharge, 3 — minimum discharge

(2) the increase in discharge sets in slowly, but after passing a peak it drops sharply reaching its minimum in October;

(3) the discharge curves tallies with precipitation curve (Fig. 6);

(4) the extremely high oscillations in flow quantity, from a few to 2500 l/s, indicate very large karst ducts which drain the water from the planina surface by the shortest possible routes;

(5) the water is relatively strongly mineralized and deposits calcite in the form of travertine tuffas.

Vratsa Planina is governed by a temperature continental climate (Figs. 6, 7 and 8). The long-term mean sum of precipitation is at Vratsa 792 mm; in the mountains it exceeds 1000 mm. Of precipitation 63% falls in spring (98 mm in June) and in summer. It is lowest in winter (42 mm in February).

#### GLAVA PANEGA

The Panega karst area is situated in the western part of Stara Planina, in the upper part of the large consequent rivers Iskur and Vit (Fig. 13). This area is built of mesozoic rocks constituting part of the flat Batul anticline. The deposits containing karst features are Titonian limestones which are famed for their purity (up to 98.5%  $\text{CaCO}_3$ ). They are 450 m thick and dip northwestward.

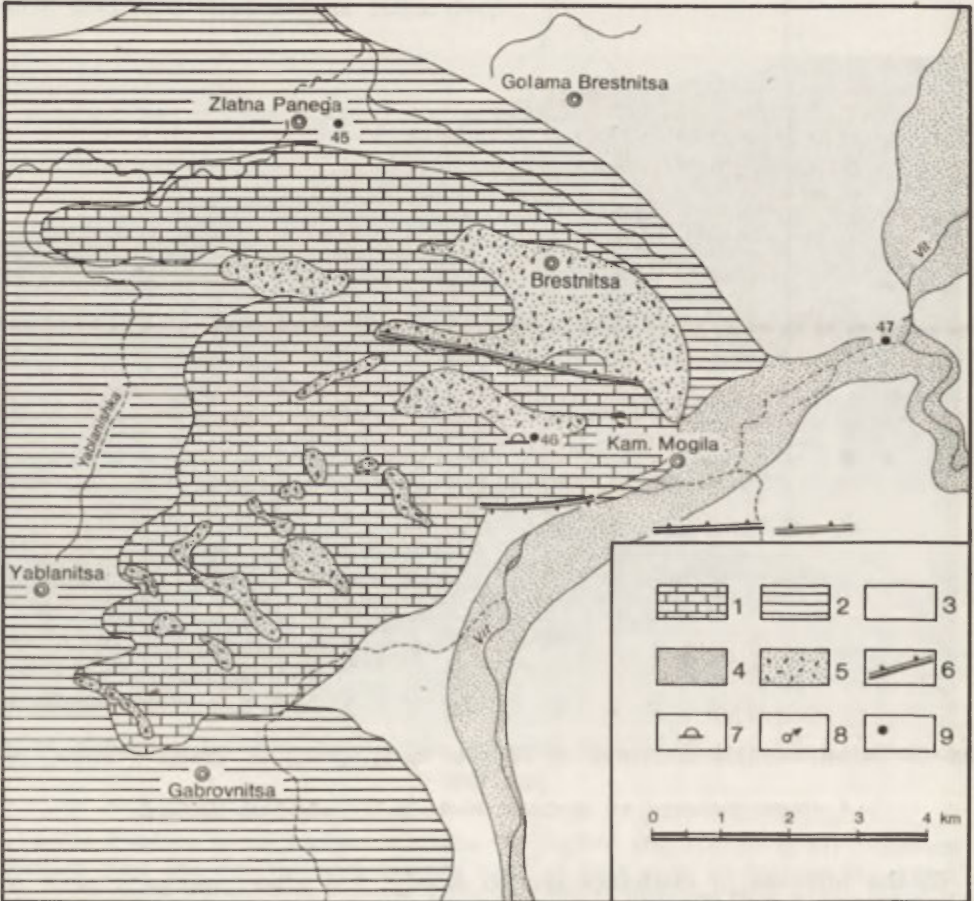


Fig. 13. Map of karst area in drainage basin of River Panega

1—Titonian limestones, 2—slightly karst limestones, 3—non-karst rocks, 4—alluvia of Vit River, 5—poljes and karst uvalas filled with terra rossa and deluvial deposits, 6—tectonic faults, visible in morphology, 7—Seva cavern, 8—Glava Panega karst springs, 9—water samples, numbered as in Table 1

The surface of these limestones is capped by a variety of karst forms such as poljes, uvalas, sinks, etc. — evidence of an extensive period of relief formation. Several stages in the evolution of this karst landscape can be distinguished (Pentshev 1965, Popov 1969). Towards the end of the Pleistocene an old-Pliocene denuding peneplanation developed which later, towards the end of the Pliocene, suffered heavy disturbances by tectonic faulting. Next, during the older Quaternary, diverse vertical movements caused part of the northern area to subside; during this period the Brestnitsa polje was formed and many caverns were filled in. Finally, in the younger Quaternary, this area underwent gradual emergence, and this resulted in the exhumation of many caverns and the revival of water flow in ancient subterranean channels.

As in other karst areas, channels of surface water are lacking in Glava Panega. The underground waters flow in a northward direction following the dip of the strata, and they end in karst springs. The largest of these springs is the Glava Panega artesian well which has a mean annual discharge of  $2.5 \text{ m}^3/\text{s}$  (Fig. 14). The slight oscillations observed in flow volume are

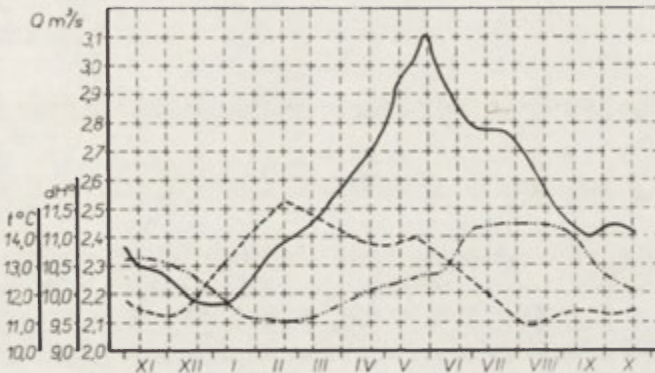


Fig. 14. Mean long-term delivery, temperature, and hardness of Glava Panega karst spring (after Pentshev)

remarkable (from  $1.6$  to  $4.1 \text{ m}^3/\text{s}$ ), in water temperature (from  $10^\circ$  to  $15.2^\circ \text{C}$ ) and in carbonate contents (from  $160$  to  $260 \text{ mg/l CaCO}_3$  — Pentshev 1965). The water flow is greatest in spring, and lowest in mid-winter — in conformity to the chart of atmospheric precipitation.

In contrast to the Bistrets springs in Vratsa Planina, Glava Panega shows rather slight amplitudes of annual discharge oscillations, and here the changes in flow volume proceed gently. These facts, besides slight water mineralization, seem to indicate that this well is partly fed by water from non-karst areas, arriving from neighbouring catchment basins. In 1955 it was found that part of the water carried by the Vit river passes by filtration into the karst region of the Panega well (Pentshev 1965).

#### THE DOBRUJA PLATFORM

The authors covered by their investigations part of the Black Sea coast of Dobruja, situated between the Batova valley and a place called Shabla. They concentrated their detailed studies upon the region of Kavarna (Fig. 15).

The area under investigation belongs to the Dobruja karst platform; its geological formation are sediments of the Upper Miocene (Beregov 1951,

Stoyanov 1960). Three horizons of the Sarmatian can be distinguished here: a lower (sandy-loamy deposits), a middle horizon some 100 m thick (arenaceous limestones), and an upper horizon 25-30 m thick (limestones). In effect these series extend horizontally, with a dip from 3 to 7° in E and NE directions. Karst phenomena have developed in the Upper Sarmatian limestones which contain up to 90.6% CaCO<sub>3</sub>.

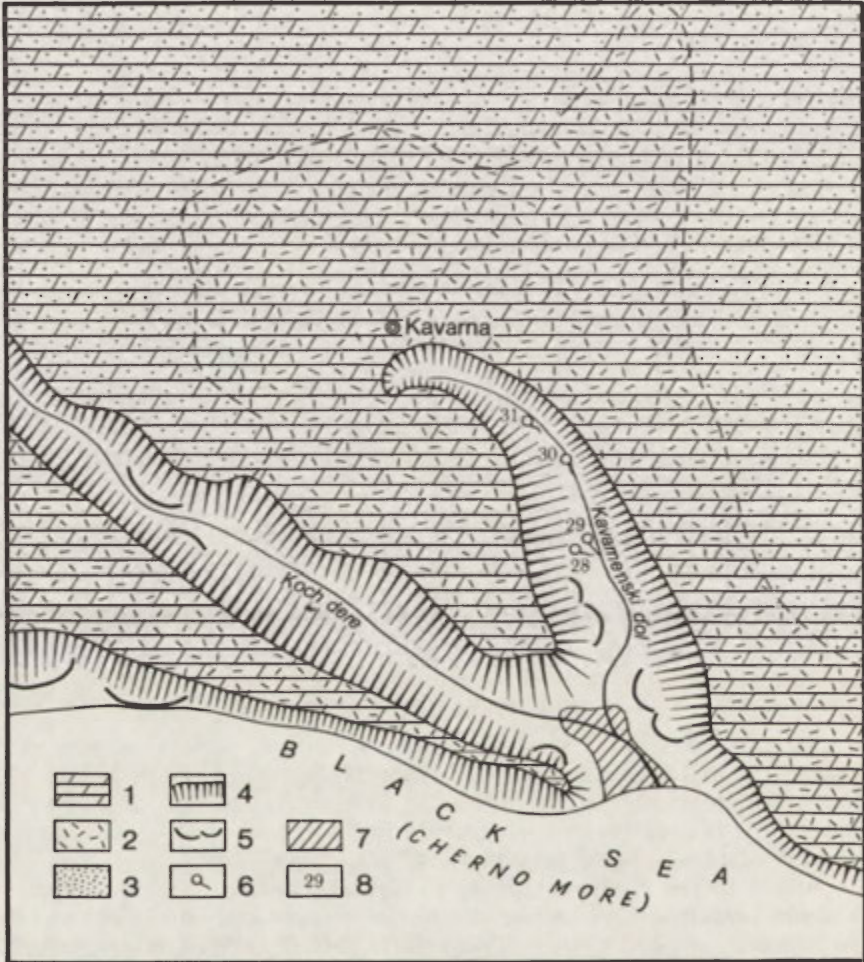


Fig. 15. Morphological map of Dobruja in Kavarna region

1—Sarmatian-Pontian erosive plain, 2—bare karst, 3—loess, 4—steep sea shore, 5—landslides, 6—karst springs, 7—marine terrace 4-5 m high, 8—water samples, numbered as in Table 1

The level surface of this platform is the result of Sarmatian-Pontian peneplanation. Along the Black Sea coast this platform is dissected by deep ravines the upper sections of which spread finger-like. These ravines are characteristic due to their flat floors which carry small streams in their lower reaches and are bordered by steep, often rocky scarps. Next to some of these narrow valleys, on top of the platform, outcrops of Upper Sarmatian limestones can be seen on which areas of karst microforms called *kayrak* are



developing. This area is some sort of rock desert extending over many square kilometres.

The platform surface lacks surface streams. All water penetrates the Sarmatian limestones, forming two water-bearing horizons overlying marls and marly clays. These horizons can be distinctly seen in a high cliff at the outlet of the Kavarna pit. The upper water-bearing horizon lies in the boundary between the Upper and the Middle Sarmatian, at about 75 m a.s.l., and the lower horizon inside the Middle Sarmatian, at 69-70 m a.s.l. Most abundant are karst springs, which issue in the bottom part of the ravines, draining the water from both the horizons mentioned. The most important of these springs delivers as much as 122 l/s; the water is strongly mineralized and deposits travertine. It was found that some of these springs discharge water which is derived from widely spread horizontal tunnels in the karst. The dip direction of these strata in E and NE directions seems to indicate that the trend of the underground water is to travel towards the coast, and that the deeply incised ravines drain wide areas of the platform. The vertical water circulation extends to a depth of 300 m.

The investigated Dobruja area is situated on the boundary of two climatic zones: a temperate continental and a tropical maritime zone. Among Bulgarian karst regions (Figs. 7 and 8) this is driest: at Tuzlata (20 m a.s.l.) precipitation is 440 m/m/y., and of this 39% falls in winter (72 m in February) and 40.9% in spring and summer. Precipitation is lowest in March and July (only about 20 mm).

#### PHYSICO-CHEMICAL PROPERTIES OF BULGARIA'S KARST WATERS

The authors carried out their physico-chemical investigations of the karst waters from Pirin, the Vratsa Planina and Dobruja in August 1967 and August 1969 (Table 2).

#### METHODS APPLIED

Temperature, pH value and amount of free carbon dioxide in the karst waters were measured directly at the localities from which samples were taken; pH was determined colorimetrically, CO<sub>2</sub> by titration with a 0.05 N sodium carbonate solution to phenolphthalein. The chemical analyses of the water samples were made in the field, in a portable laboratory. The calcium ion contents were determined by titrating, for pH = 12, with 0.02 N solution of bisodium versenate to murexide as indicator. The total hardness was found by titrating, for pH = 10, with a 0.02 N solution of bisodium versenate to eriochrome black T as indicator. Next, the contents of bicarbonate ions (the basicity) was found by titrating with a 0.05 N solution of muriatic acid to methyl orange as indicator. Finally, to determine the total mineralization, the specific conductivity of the solutions was measured by means of a portable battery-operated conductometer. The results of all these determinations are listed in Table 1 and shown graphically in Figs. 16 to 23. The numbering of the water samples given in Table 1 and in Figs. 16 to 23 tallies with the numbers shown in Figs. 2, 9, 10, 13 and 15.

Usually the aggressiveness of water to calcium carbonate is defined by the number of CaCO<sub>3</sub> milligrams which one litre of water can dissolve, in order to obtain in this way saturation in a closed system or, in other words, without increasing or decreasing the dioxide contents in the water. This is the method most commonly employed in laboratories for determining water aggressiveness, the so-called Hayer's "marble method". However, in this

TABLE 2. Physico-chemical properties of Bulgaria's karts waters

Number of the sample	Date	Hour	Temperature of air °C	Temperature of water °C	pH	Free CO <sub>2</sub> ppm	Total hardness ppm	meq/l	Δ CaCO <sub>3</sub> * ppm	Degree of saturation	Specific resistance at 18°C Ω cm	Total mineralisation ppm	ppm	Ion Ca <sup>++</sup> meq/l	contents Mg <sup>++</sup> ppm meq/l		HCO <sub>3</sub> ppm	meq/l	Mg/Ca index
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
1	12.8.69.	14 <sup>10</sup>	31.0	9.6	7.6	4	120	2.40	+5	0.96	5120	141	36	1.8	7.2	0.60	152	2.5	0.33
2	12.8.69.	14 <sup>50</sup>		7.4	7.4	8	122	2.45	+8	0.94	5500	131	37	1.8	7.0	0.60	152	2.5	0.33
3	12.8.69.	15 <sup>00</sup>			7.4						5630	128							
4	12.8.69.	15 <sup>30</sup>		12.5	7.4	22	230	4.60	-12	1.05	2760	261	61	3.0	19.0	1.6	305	5.0	0.53
5	12.8.69.	16 <sup>00</sup>		18.4	7.6	3	200	4.00	-12	1.07	3250	222	56	2.8	14.0	1.2	260	4.25	0.43
6	12.8.69.	17 <sup>50</sup>		19.1	7.6	2	122	2.45	+5	0.96	4840	149	39	2.0	6.0	0.5	168	2.75	0.25
7	12.8.69.	18 <sup>10</sup>		11.8	7.4	18	198	3.95	0	1.00	3270	220	57	2.8	13.0	1.1	244	4.0	0.39
8	12.8.69.	19 <sup>00</sup>		8.3	7.4		113	2.25	+12	0.90	6100	118	28	1.4	10.0	0.9	152	2.5	0.64
9	13.8.69.	9 <sup>00</sup>			7.0	3	125	2.50	+8	0.60	42,100	17	4	0.2	trace		18	0.3	0.00
10	13.8.69.	14 <sup>25</sup>		0.1	6.6	4	3	0.05	+20	0.05	106,000	7	1	0.05	none detected		2	0.03	0.00
11	13.8.69.	15 <sup>30</sup>		3.4	6.9	1	5	0.10	+20	0.10	110,000	6.5	2	0.1	none detected		15	0.25	0.00
12	14.8.69.	15 <sup>00</sup>	25.6	8.3	7.4	19	213	4.25	-8	1.06	3020	238	68	3.4	10.0	0.9	305	5.0	0.26
13	14.8.68.	16 <sup>00</sup>			7.4	2	43	0.85	+12	0.78	16,750	43	10	0.5	4.0	0.3	76	1.25	0.60
14	14.8.69.	16 <sup>30</sup>					50	1.00			13,650	53	14	0.7	4.0	0.3	92	1.50	0.43
15	14.8.69.	17 <sup>40</sup>	25.6	15.0	7.6	19	325	6.50	-38	1.13	2090	345	106	5.3	14.0	1.2	457	7.5	0.23
16	16.8.69.	16 <sup>00</sup>	10.2	8.6	7.4	23	245	4.90	-17	1.08	2060	350	93	4.7	3.0	0.25	350	5.75	0.05
17	16.8.69.	16 <sup>30</sup>	8.3	6.9	7.4	12	210	4.20	-10	1.05	2910	248	81	4.0	2.0	0.2	244	4.0	0.05
18	16.8.69.	16 <sup>50</sup>	10.0	9.0	7.0	41	258	1.15	+12	0.96	2520	286	102	5.1	trace		336	5.5	0.00
19	16.8.69.	17 <sup>50</sup>			7.8	10	293	5.80	-43	1.17	2290	315	96	4.8	12.0	1.0	351	5.75	0.21
20	17.8.69.	8 <sup>30</sup>	23.0	12.6	7.0	53	275	5.50	+10	0.96	2360	305	109	5.5	trace		320	5.25	0.00
21	17.8.69.	8 <sup>40</sup>	23.0	15.5	7.0	51	260	5.20	+15	0.94	2450	294	104	5.2	none detected		320	5.25	0.00
22	17.8.69.	8 <sup>55</sup>		12.0	7.3	23	238	4.75	-13	1.05	2640	273	94	4.7	trace		305	5.0	0.00
23	17.8.69.	10 <sup>50</sup>		12.5	7.0	57	320	6.40	0	1.00	2060	350	121	6.0	4.0	0.3	412	6.75	0.05
24	17.8.69.	11 <sup>30</sup>		13.5	6.8	141	382	7.65	0	1.00	1510	477	142	7.1	7.0	0.6	534	8.75	0.08
25	17.8.69.	14 <sup>15</sup>	23.4	9.8	7.0	42	285	5.70	+5	0.98	2280	316	114	5.7	none detected		381	6.25	0.00

26	17.8.69.	17 <sup>30</sup>	27.7	11.5	7.0	35	228	4.55	+20	0.92	2750	262	86	4.3	3.0	0.25	305	5.0	0.06
27	17.8.69.	17 <sup>33</sup>	27.6	12.0	7.0	48	300	6.00	+5	0.98	1850	390	105	5.3	9.0	0.8	336	5.5	0.15
28	26.8.69.	9 <sup>30</sup>	28.4	12.9	7.4	26	330	6.60	-30	1.10	1540	467	59	3.0	44.0	3.6	426	7.0	1.20
29	26.8.69.	10 <sup>00</sup>			7.5						1530	470					426	7.0	
30	26.8.69.	10 <sup>30</sup>		13.5	7.4		358	7.15	-40	1.13	1395	516	61	3.05	49.0	4.1	472	7.75	1.35
31	19.8.67.	10 <sup>20</sup>	20.5	13.0	7.6	14	380	7.60	-62	1.20			38	1.9	69	5.7	476	7.8	3.00
32	19.8.67.	10 <sup>10</sup>	23.3	13.4	7.6	18	300	6.00	-35	1.13			84	4.2	22	1.8	458	7.5	0.43
33	19.8.67.	9 <sup>45</sup>	23.3	13.4	7.6	20	450	9.00	-100	1.29			20	1.0	96	8.0	488	8.0	8.00
34***	19.8.67	12 <sup>00</sup>	27.3	15.5	7.8		340	6.80	-55	1.10			76	3.8	36	3.0	274	4.5	0.80
35	19.8.67	15 <sup>15</sup>	28.7	14.3	7.5		400	8.00	-70	1.21			92	4.6	41	3.4	396	6.5	0.74
36	18.8.67.	10 <sup>00</sup>	24.6	13.7	7.4	14	360	7.20	-38	1.12			64	3.2	48	4.0	464	7.6	1.25
37	18.8.67.	10 <sup>10</sup>	24.6	13.7	7.4	13	300	6.00	-25	1.09			32	1.6	53	4.4	488	8.0	2.75
38	18.8.67.	12 <sup>45</sup>	28.1	14.2	7.4	14	340	6.80	-28	1.09			36	1.8	60	5.0	405	6.6	2.78
39	19.8.67.	12 <sup>15</sup>	25.1	12.8	7.8	13	500	10.00	-150	1.43			40	2.0	96	8.0	427	7.0	4.00
40	19.8.67.	13 <sup>15</sup>	24.1	14.3	7.8	8	300	6.00	-50	1.20			72	3.6	29	2.4	354	5.8	0.67
41	19.8.67.	16 <sup>40</sup>	23.1	11.8	7.4	15	380	7.60	-50	1.15			92	4.6	36	3.0	398	6.5	0.65
42	19.8.67.	17 <sup>10</sup>	22.0	13.7	7.4	19	420	8.40	-70	1.20			86	4.3	49	4.1	488	8.0	0.95
43	21.8.67.	12 <sup>00</sup>	24.6	12.6	7.6	9	220	4.40	-16	1.08			44	2.2	26	2.2	324	5.3	1.00
44	21.8.67.	13 <sup>00</sup>	24.6	13.6	7.8	7	290	5.80	-50	1.20			16	0.8	60	5.0	354	5.8	6.25
45	21.8.69.	12 <sup>30</sup>	22.7	13.8	7.2	21	185	3.70	+10	0.95	3230	223	66	3.3	5	0.4	244	4.0	0.12
46	21.8.69.	14 <sup>00</sup>	11.2	11.4	7.4	20	285	5.7	-25	1.10	1990	362	101	5.0	8	0.7	388	6.35	0.14
47	21.8.69.	15 <sup>12</sup>	28.1	23.1			95	1.9			6500	111	30	1.5	5	0.4	122	2.00	0.27

\*  $\Delta\text{CaCO}_3$ : increase (+ $\Delta$ ), or decrease (- $\Delta$ ) total hardness on saturation, without gain or loss of  $\text{CO}_2$ ; - $\Delta$  signifies supersaturated water, + $\Delta$  signifies aggressive water,  $\Delta = 0$  signifies saturated water

\*\* Degree of saturation: the measured hardness proportion to the hardness on saturation (without gain or loss of carbon dioxide); degree of saturation = 1 signifies saturated water <1 — aggressive water, >1 — non-aggressive water

\*\*\* The samples not identified on the figures: 34 — Tolkliman, 35 — Bulgarevo, 36 — Bolota-dere, 37 — Bolota-dere spring, 38 — Bulgarewo, 39 — Kogpunar, spring, 40 — Kogpunar, cave, 41 — Ovcharska beach, 42 — Balchik, 43 — Vit-vit village, 44 — Vit-vit sanatorium, 45, 46, 47 — Glava Panega

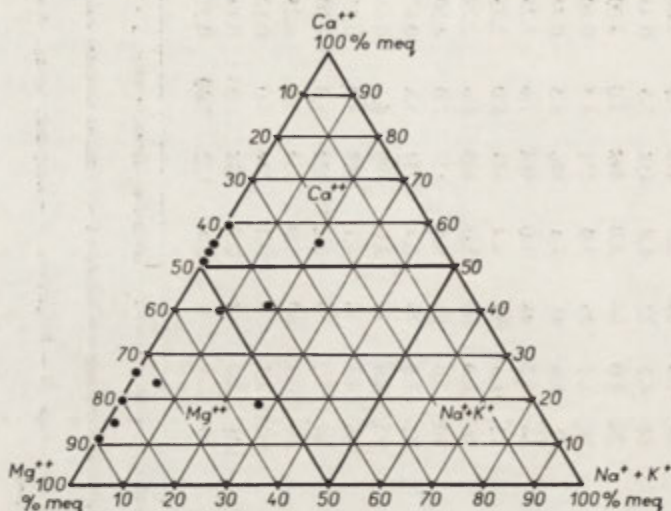


Fig. 16. Distribution of cations in Dobruja karst waters in %

method the time element is left out of consideration, and it is known that the rate of dissolving calcium carbonate depends on the physical form of  $\text{CaCO}_3$ , and on the degree of water saturation.

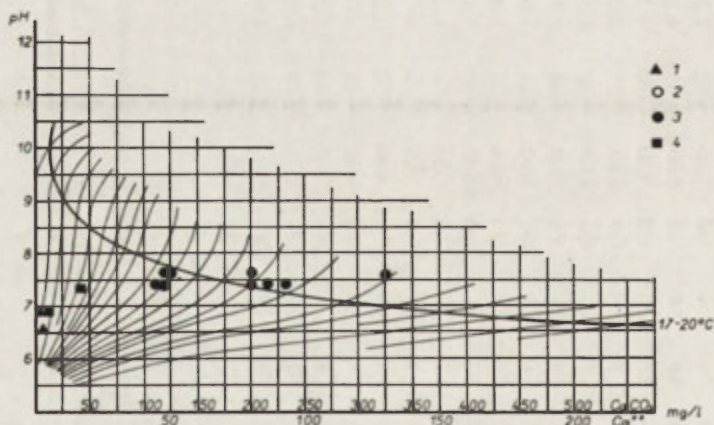


Fig. 17. Chemical aggressiveness of Pirin karst waters, shown in Tillmans-Trombe graph modified by Schmitt

1 — meltwaters, 2 — cavern waters in near-surface zone, 3 — karst springs, 4 — non-karst waters

In Bulgaria the authors determined the aggressiveness of the karst waters by an indirect, graphical method. They projected the established pH value and the total hardness upon a Trombe-Schmitt (Roques 1969) chart of carbonate equilibrium, and next they calculated the degree of relative saturation and the increase or decrease in  $\text{CaCO}_3$ , in order to obtain an equilibrium in the solution, in other words, to effect its saturation (Table 1, Columns 9 and 10). This procedure corresponds to the laboratory method of determining aggressiveness using what is called the "marble method".

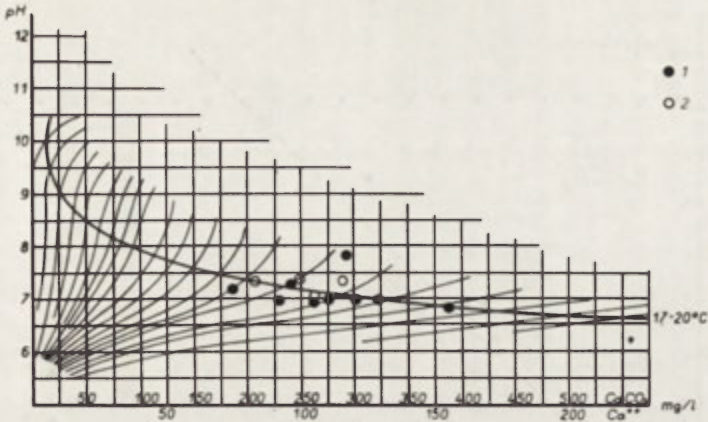


Fig. 18. Chemical aggressiveness of Stara Planina karst waters  
Water taken from: 1 — karst springs, 2 — caverns

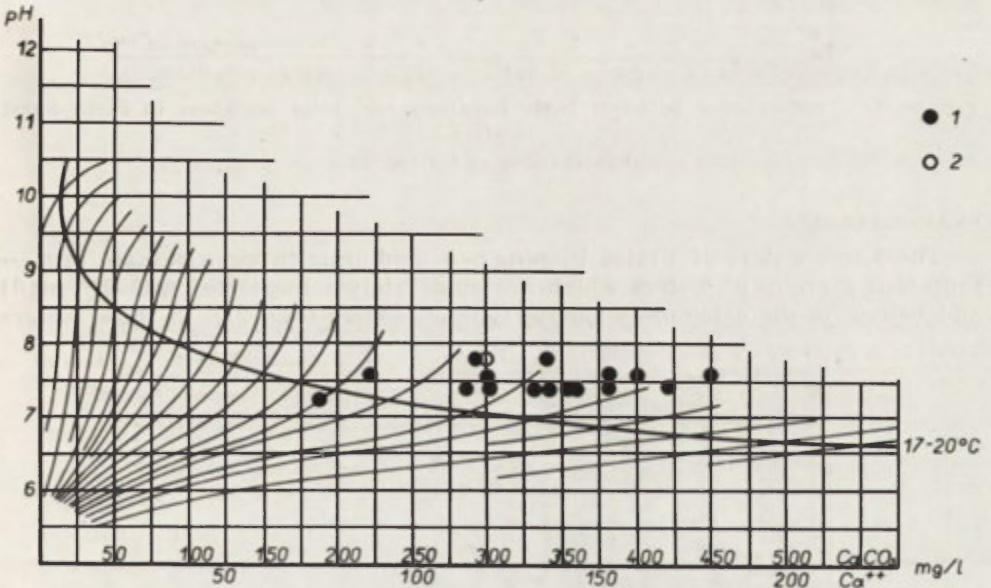


Fig. 19. Chemical aggressiveness of Dobruja karst waters  
Water taken from: 1 — karst springs, 2 — caverns

Compared with the waters of Stara Planina and Dobruja, the karst waters of the Pirin Massif—high-mountain waters—are least mineralized; they show a concentration of up to 300 mg/l and belong to the calcium-magnesium-carbonate water type (Fig. 20). The Mg/Ca ratio in the Pirin waters oscillates between 0.20 and 0.65. The contents of free carbon dioxide is never more than 22 mg/l and the pH value is 7.4 to 7.6. Regarding the quantity of dissolved carbonates two separate types of karst waters can be distinguished: samples Nos. 1, 2, 6 and 8 with 100–150 mg/l, and samples Nos. 4, 5, 7, 12 with 220–260 mg/l mineralization. The relative saturation of the water of the first group is 0.90 to 0.92 so that they are only slightly aggressive. The water of the second group has a saturation of 1.0 to 1.07; hence this water is not aggressive, being saturated or slightly oversaturated (Figs. 17 and 23).

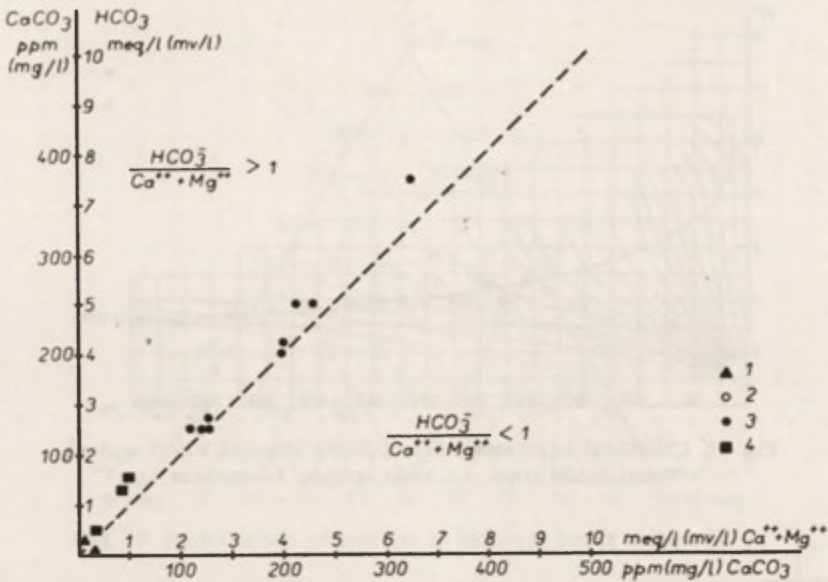


Fig. 20. Interdependence between basic hardness and total hardness in Pirin karst waters

Explanations as for Fig. 17

VRATSA PLANINA

The karst waters of Vratsa Planina — a medium-altitude mountain area — represent a group of waters which are moderately mineralized (250–350 mg/l) and belong to the calcium-carbonate type of waters (Fig. 21). In these waters

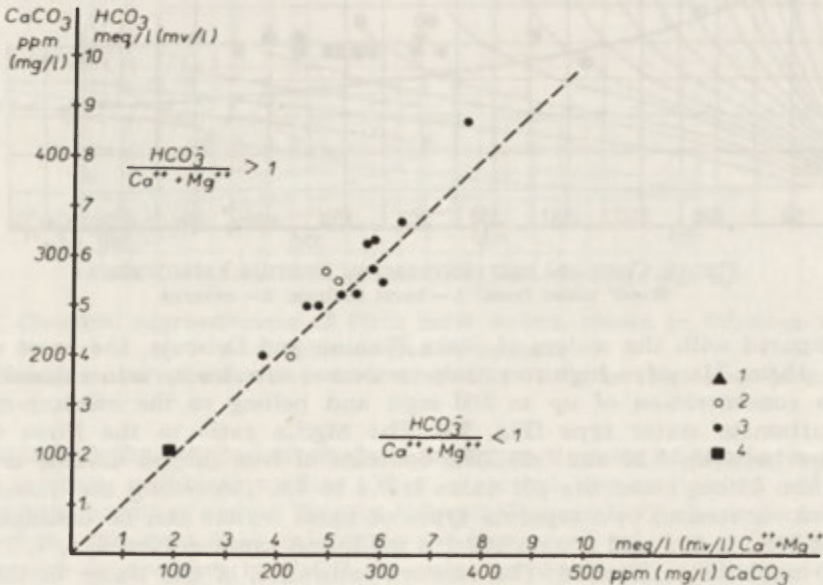


Fig. 21. Interdependence between basic hardness and total hardness in Stara Planina karst waters

Explanations as for Fig. 17

the Mg/Ca ratio is from 0 to 0.2. Their contents of free carbon dioxide is relatively high, oscillating between 10 and 57 mg/l; the pH value varies from 7.0 to 7.4. An exception is sample No. 24: this water contains as much as 141 mg/l of free carbon dioxide with a pH value of 6.8; its mineralization is 477 mg/l. Only by methodological hydrochemical examinations for at least one full hydrological year would it be possible to determine the source of this extremely high CO<sub>2</sub> contents.

The relative saturation of the water determined in samples Nos. 18, 20, 21, 24 and 25 is less than 1.0 and oscillates between 0.92 and 0.98; these waters are slightly aggressive. Regarding the remaining samples the relative saturation varies from 1.0 to 1.08 and these waters are non-aggressive; they are saturated or slightly oversaturated and apt to deposit CaCO<sub>3</sub> (Figs. 18 and 23).

#### DOBRUJA

The Dobruja waters show the highest degree of mineralization: from 220 to 500 mg/l, and are of highly diversified type (Figs. 16 and 22). Samples Nos. 34, 35, 40, 41 and 42 belong to the type of calcium-magnesium-carbonate waters where the Mg/Ca ratio fluctuates between 0.65 and 0.95; samples Nos. 28, 30, 31, 33, 36, 37, 38, 39 and 44 belong to the magnesium-calcium carbonate type where the Mg/Ca ratio oscillates between 1.20 and 8.00. The ratio of cations to anions (Fig. 22) determined for these waters indicates that the

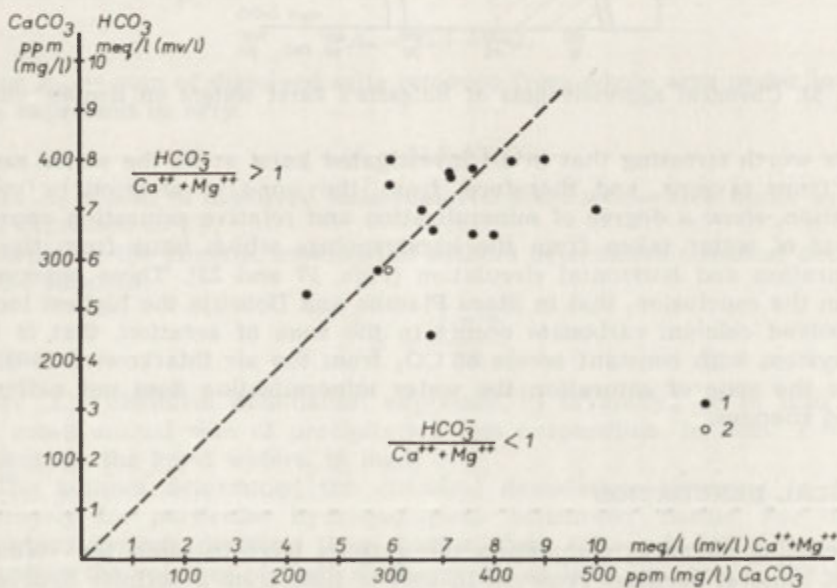


Fig. 22. Interdependence between basic hardness and total hardness in Dobruja karst waters

Water taken from: 1 — karst springs, 2 — caverns

Dobruja karst waters contain an admixture of further salts, most probably of magnesium and calcium chlorides and sulphates and of sodium and potassium carbonates. The contents of free carbon dioxide oscillates between 7 and 26 mg/l, and the pH value between 7.4 and 7.8.

The relative saturation of the Dobruja karst waters fluctuates between 1.04 and 1.43; these waters are supersaturated, non-aggressive, and liable to deposit  $\text{CaCO}_3$  (Figs. 19 and 23) to a higher degree than the karst waters from Pirin and Vratsa Planina.

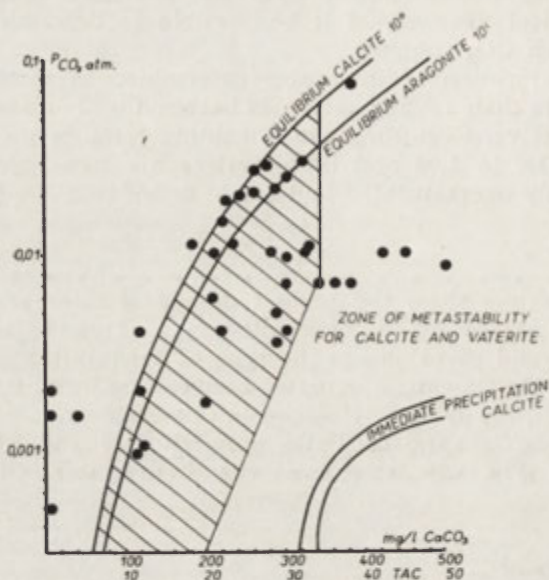


Fig. 23. Chemical aggressiveness of Bulgaria's karst waters on Roques' chart

It is worth stressing that in all investigated karst areas the water samples taken from caverns, and therefore from the zone of aeration or vertical circulation, show a degree of mineralization and relative saturation approaching that of water taken from the karst springs which issue from the zone of saturation and horizontal circulation (Figs. 17 and 23). These observations confirm the conclusion, that in Stara Planina and Dobruja the highest increase in dissolved calcium carbonate occurs in the zone of aeration, that is in an open system with constant access of  $\text{CO}_2$  from the air (Markowicz 1968), and that in the zone of saturation the water mineralization does not suffer any marked changes.

#### CHEMICAL DENUDATION

Discussing chemical denudation the authors have in mind the volume or mass of mineral particles removed in unit of time from a definite hydrogeological drainage area. Most often used as unit of measurement is the notion  $\text{m}^3/\text{km}^2/\text{y}$ . — corresponding to the hypothetical lowering of the land surface by 1 mm in 1000 years.

Mechanical denudation must be distinguished from chemical denudation. The former involves processes by which debris is dragged, or carried suspended in flowing water; the latter refers to components dissolved in the water.

Karst areas lack surface streams and the water circulates in subsurface channels. The specific subterranean conditions cause the transport of dragged or suspended debris to be limited to larger channels. Observations revealed,



that in karst springs the water is usually pure and that only exceptionally it carries a mineral suspension. These facts point to chemical denudation as the principal factor responsible for the destruction of karst areas.

#### THE METHOD APPLIED

The contemporaneous rate of intensive chemical denudation of Bulgaria's karst areas has been determined directly by applying the so-called hydrometric method (Pulina 1966). However, for purposes of comparison, calculations of the denudation were also made by what is called the climatic method (Corbel 1959).

Applying the hydrometric method, denudation was computed by means of the following formulae<sup>4</sup>:

$$D = 12.6 \frac{\Delta T \cdot Q}{P} \quad (1)$$

where  $D$  = chemical denudation expressed in  $\text{m}^3/\text{km}^2/\text{y}$ . or in  $\text{mm}/1000 \text{ y}$ .  
 $\Delta T = T - T_a$ ;  $T$  = mean annual water mineralization in karst springs, in  $\text{mg}/\text{l}$ ;  
 $T_a$  = mean annual mineralization of atmospheric water, in  $\text{mg}/\text{l}$ ;  $Q$  = mean annual discharge, in  $\text{m}^3/\text{s}$ ;  $P$  = area of outcrops of carbonate rock, in  $\text{km}^2$ .

$$D = 0.0126 \cdot \Delta T \cdot q, \quad (2)$$

where  $q = 1000 \cdot \frac{Q}{P}$ ;  $q$  = unit flow, in  $1/\text{s}/\text{km}^2$ .

$$A_m = 12.6 \cdot \Delta T \cdot Q, \quad (3)$$

where  $A_m$  = sum of dissolved salts removed from whole area under investigation, expressed in  $\text{m}^3/\text{y}$ .

$$A_t = 31.5 \cdot \Delta T \cdot Q', \quad (4)$$

where  $A_t$  = sum of dissolved salts removed from whole area under investigation, expressed in  $\text{t}/\text{y}$ .

Applying the climatic method the authors determined chemical denudation by the formula:

$$X = \frac{4 \cdot E \cdot T}{100}, \quad (5)$$

where  $X$  = chemical denudation expressed in  $\text{m}^3/\text{km}^2/\text{y}$ ., or in  $\text{mm}/1000 \text{ y}$ .;  
 $E$  = mean annual sum of precipitation less evaporation, in  $\text{dcm}$ ;  $T$  =  $\text{CaCO}_3$  contents in the karst waters, in  $\text{mg}/\text{l}$ .

The authors determined the chemical denudation occurring in Bulgaria separately for particular hydrogeological catchment basins. For the most important springs draining these basins, they collected the necessary data regarding the volumes of their discharge for at least two years. The chemical composition of their waters were determined once in August when the water flow is lowest in both Stara Planina and Dobruja and at a medium level in the Pirin area. The authors calculated approximately the mineralization of precipitation water, in conformity with Rozdztstvenski's research work (1958, 1969) and by making analyses of water obtained from melted snow.

<sup>4</sup> It is worth mentioning that these formulae can also be used for determining the mechanical denudation. In this case  $\Delta T$  equals the quantity of dragged or suspended debris, in  $\text{g}/\text{m}^3$ .

Further, the authors calculated the chemical denudation, taking into account the full mineralization of the karst waters ( $D_M$ ) as well as the carbonate contents expressed in  $\text{CaCO}_3$  ( $D_{\text{CaCO}_3}$ ). This calculation of both full chemical denudation and carbonate denudation is important from a methodological point of view when chemical denudation of non-karst areas is considered, because in the removed salts the share of carbonates is relatively small. It is worth mentioning that for karst regions the order of magnitude of  $D_M$  and  $D_{\text{CaCO}_3}$  is much alike.

In Table 1 are presented the values of denudation, based on mean annual flow and on flow observed during field examinations. The differences in the results obtained are the lowest for Pirin where the flow data for August were about the same as the mean annual data. These observations are of marked methodical significance in investigations of chemical denudation, because they show that in karst high-mountain regions measurements made in summer can be considered approaching the mean annual values.

## RESULTS

For the three karst areas which are fairly good representatives of the different types of Bulgaria's karst conditions, the index of chemical denudation varies within the following limits:  $D_M$  from 30.5 to 57.8, and  $D_{\text{CaCO}_3}$  from 23.0 to 47.0  $\text{m}^3/\text{km}^2/\text{y}$ . (see Table 2).

For Dobruja the index of denudation was found to be lowest although here water mineralization is high. On the other hand, this index is highest for Pirin in spite of lower water mineralization. It appears that the amount of water passing the given karst area is of decisive influence upon differences in intensity of denudation.

Some sort of yardstick for the rate of chemical denudation can be called the sum of soluble carbonate rock carried off (Table 2, Column A). Thus, in the Pirin range where the catchment basin covering 71  $\text{km}^2$  is largest, the amount of limestone rock removed per year is from 3400 to 3800  $\text{m}^3$ . The Yazo spring alone discharges annually 2000 to 2400  $\text{m}^3$  limestone. The volume of carbonate rock dissolved annually in the Pirin area is the equivalent to that of a tunnel of 1  $\text{m}^2$  section and a length of 3.4 to 3.8 km.

In view of all the above results the authors are inclined to consider the value of 30  $\text{m}^3/\text{km}^2/\text{y}$ . a fair approximation to the mean index of denudation for all of Bulgaria's karst areas. Converted to the surface area of soluble rock outcrops, this means the removal of some 0.76 million  $\text{m}^3$  per year. A certain part of the carbonate salts are deposited near the springs and in nearby valleys as travertine tuffas; particularly favourable in this respect are dry and warm regions where—as demonstrated by Ivanov's research (1964)—the fluvial waters show the highest degree of mineralization. However, the major part of the dissolved salts are carried off into the Black Sea.

For the indices of present-day chemical degradation in Bulgaria's karst areas, the authors found it possible to draw tentative conclusions as to former palaeogeographical conditions. Using as starting point the assumption that during the Quaternary denudation must have proceeded mostly as it does today, they believe that the thickness of the hypothetical rock layer removed in the meantime is: 30 m for Dobruja, 45 m for Stara Planina, and 52 m for Pirin.

Applying the hydrometric method for determining denudation enabled the authors to establish how much allochthonic water from non-karsted areas

penetrates the investigated catchment basins from outside. As an example they present the results of their calculations for the basin of the River Panega in Stara Planina (Table 1).

The Glava Panega spring is fed from both precipitation water permeating karst massifs and from non-karst waters of the River Vit. The per-cent share of each of these waters has been defined from the ratio of mineralization of the two types of water to the mineralization of the spring water: these calculations were verified by means of a denudation balance (Table 1, Column A). The authors made their field measurements at a time at which the water was at about its mean annual level; and they found that in August 1969 the River Vit supplied 1.3 to 1.4 m<sup>3</sup>/s as her part of the flow of the spring, the equivalent ow 54% of the total discharge of the Glava Panega spring. In spite of this ample flow, the River Vit plays a rather secondary role in the balance calculation of chemical denudation in the drainage basin of the River Panega — amounting to a bare 25 to 26%. Water derived from vertical infiltration of precipitation is the principal destructive agent, accounting for 74–75% of chemical denudation.

The above data confirm Pentshev's (1965) calculations who believed the mean annual infiltration of Vit waters into the Panega basin to be 49%.

## CONCLUSIONS

The karst areas of Pirin, Stara Planina and Dobruja may definitely be considered as representative areas of karst encountered in Bulgaria: a high-mountain karst — Pirin, a medium-altitude karst — Stara Planina and a platform karst — Dobruja. These three types have passed through different stages of evolution and their karst reliefs have been developing at different rates.

The areas of Stara Planina and Dobruja are the oldest revealing typical macro- and meso-forms of surface karst and well developed caverns. Pirin is the youngest, where the forms of surface karst are poorly developed and where the underground drainage tunnels are of Pleistocene and Postglacial origin.

Contemporaneous destructive processes act upon the three karst areas with different intensity, and for each the rate of denudation depends not only on differences in climatic conditions but on structure and present-day land relief as well.

The authors had been given the task of determining the intensity of the modern processes of denudation, assessed upon a wider background of morphological and hydrochemical evidence. On top of this, the authors went further and described a further variety of carbonate aggressiveness of karst waters and defined the manner by which they established the thare of allochthonic waters in the discharge of karst springs.

Concise conclusions resulting from all the research discussed above may be formulated as follows:

(1) The contemporaneous coefficients of chemical denudation show the values of 23.0–30.5 m<sup>3</sup>/km<sup>2</sup>/y. for Dobruja and up to 47.0–51.8 m<sup>3</sup>/km<sup>2</sup>/y. for Pirin. For Stara Planina the values are intermediate, 38.4–45.0 m<sup>3</sup>. The above coefficients of denudation determined for Bulgaria come near to the values known from other European karst areas situated in zones of subequatorial and temperate climates (Pulina 1970).

(2) The differences in the intensity of denudation are caused, for the most part, by differences in the quantities of fluvial waters crossing the particular karst areas and in the aggressiveness of these waters. Further factors such as the lithological features of the rocks are of rather secondary importance.

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## A QUANTITATIVE COMPARISON BETWEEN RAINFALL EROSION CAPACITY IN A TROPICAL AND A MIDDLE-LATITUDE REGION

JAN DE PLOEY

### INTRODUCTION

From 1964 to 1968 tracer experiments using  $Sc^{46}$ -labelled glass were carried out in the neighbourhood of Lovanium University (Kinshasa, Congo) in order to obtain a better understanding of rainfall erosion mechanisms (J. De Ploey 1966, 1967, 1969). Splash erosion modalities were also examined in the laboratory applying tracer methods (J. De Ploey and J. Savat 1968, J. De Ploey 1969).

Field experiments took place on experimental plots corresponding to convex slope segments covered by a substeppic grassy savannah which characterizes vast expanses in the southern Congo, in Angola and Zambia. This vegetation is characteristic of the fine and permeable sandy soils, the so-called arenoferrals, which are developed on mesozoic formations and on the cainozoic Kalahari System. A detailed description of the experimental crest-slope has been given in former publications. Nevertheless, some characteristics may be recalled: (1) hilltop ( $15^{\circ}18'E$  and  $4^{\circ}26'S$ ) at an elevation of 450 m (2) declivity;  $0^{\circ}$  to  $18^{\circ}$  (3) fall from hilltop; 25 m (4) slope length; 300 m (5) base level; a rectilinear portion ( $20^{\circ}$ ) passing into a concave foot-slope (6) microtopography; absence of rilling (7) ground cover; low to mid-height grasses and a few shrubs (*Landolphia*, *Hymenocardia*...) (8) crown density of vegetation cover; 25% to 50% (9) bedrock; homogeneous silty sands (mode between 150  $\mu$  and 300  $\mu$ ) (10) topsoil; an arenoferral with weakly developed humus- and B-horizon (11) infiltration capacity; minimum 500 mm/h.

Since soils and vegetation are remarkably uniform over the whole area, it is assumed that differentiation of pluvial erosion is essentially a function of pluviometric parameters and slope. In order to evaluate the role of these parameters, the field experiments were spread over three rainy seasons, on convex slope segments of  $0^{\circ}$ - $6^{\circ}$ ,  $9^{\circ}$ - $12^{\circ}$  and  $18^{\circ}$  respectively. A pluviograph was mounted near the experimental plots in order to check pluviometry, and the pluviogram of each rainfall was divided into several pluviophases, each of which corresponds to a specific value of rainfall intensity and duration (Fig. 1). As one could expect, much of the convective rainfall in Kinshasa was shown to be characterized by a relatively high degree of intensity and by a well-defined period of duration.

Some samples of tracer sands were placed below a shield to eliminate splash erosion and reveal the part played by runoff. The erosive effect of each rainfall was qualified in this manner: (1) total erosion, when splash erosion and discontinuous, intermittent runoff were active over the whole slope seg-

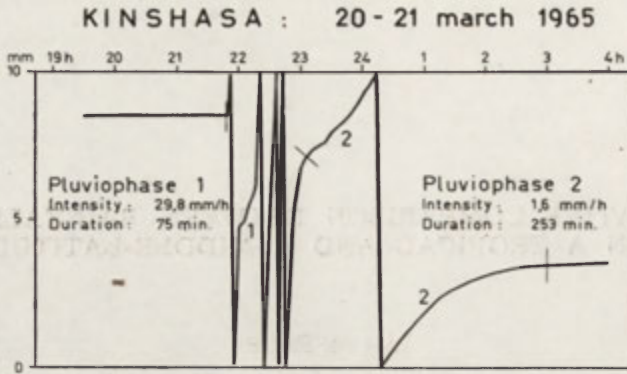


Fig. 1.

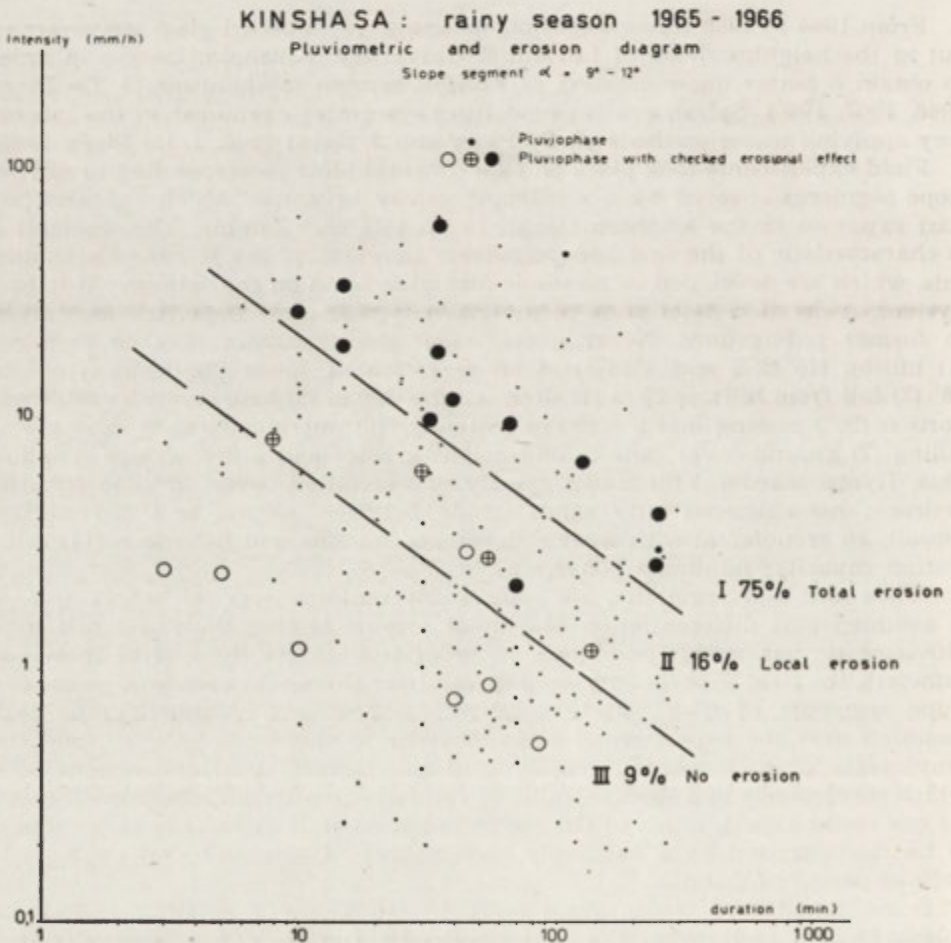


Fig. 2.



ment; (2) local erosion, erosion was mainly the result of raindrop impact on the uncovered spots; (3) no erosion, when no measurable removal of tracersands occurred. No steady continuous runoff was observed on the sandy experimental slope, not even on the 18° plot.

The results of field experiments with three convex segments have been sketched in pluviometric diagrams showing the intensity—duration values of all pluviophases in one rainy season—including the pluviophases of rainfalls whose erosive effect was not determined. An erosion diagram of each plot has been obtained by subdividing the pluviometric diagram into three erosion zones according to the above-mentioned threefold qualification of the erosive effect (Fig. 2).

The relative importance of each erosion zone is indicated by the percentage of precipitation recorded for each slope segment during the rainy season. These results have been discussed in other publications and can be compared with the erosion activity in a middle-latitude region:

	Total erosion	Local erosion	No erosion
Slope 0°–6°	57%	21%	22%
9°–12°	75%	16%	9%
18°	87%		13%

The climate of the western Congo is an Aw-4 climate according to the Köppen classification. It is marked by a dry season lasting from June to September. The area of Kinshasa is still slightly influenced by the cold Benguela current so that mean annual rainfall is only of about 1350 mm. Much of this rainfall is convective in origin and during each rainy season main pluviophases have been registered with rainfall rates of more than 50 mm/h, a few surpassing 100 mm/h. But even during the heaviest showers no steady continuous runoff affected the sandy hills. A discontinuous flow was detected transporting particles over a distance limited to several centimeters or decimeters.

The erosion diagrams of the Kinshasa region give rise to the question of evaluating the relative erosion capacity of middle-latitude rainfall—assuming the Kinshasa area was affected by precipitation of cyclonic origin. It was therefore decided to re-interpret the erosion diagrams with their fixed zone boundaries, taking into account the specific pluviometric conditions of a middle-latitude region.

#### COMPARATIVE PLUVIOMETRY KINSHASA-ANTWERP

The Antwerp area is characterized by a precipitation regime which is representative for a humid marine west-coast climate. Total annual rainfall is of about 700 mm. In 1968 the station of Deurne near Antwerp recorded 685.3 mm. The pluviophases of that year have been analysed and divided into pluviophases in the same way as was done in Kinshasa. A pluviometric diagram was thus obtained, comparable to the 1964–1965 diagram of Kinshasa where 1156 mm of rainfall was recorded during the major part of the rainy season (Fig. 3): 1) The Antwerp diagram is marked by a relative concentration of pluviophase-intensities under 2 mm/h of a duration varying between 15 and 800 minutes. These intensities occur rarely in Kinshasa 2) High intensities observed in Antwerp during summer and early autumn range from 10 to 30 mm/h while the corresponding values for tropical rainfall vary between 30 and 100 mm/h. Maximum intensity recorded in Antwerp amounted to

68 mm/h, compared to the Kinshasa rate of 115 mm/h. The pluviophases with medium or high intensities (5–100 mm/h) last for longer or shorter periods in Kinshasa than in Antwerp.

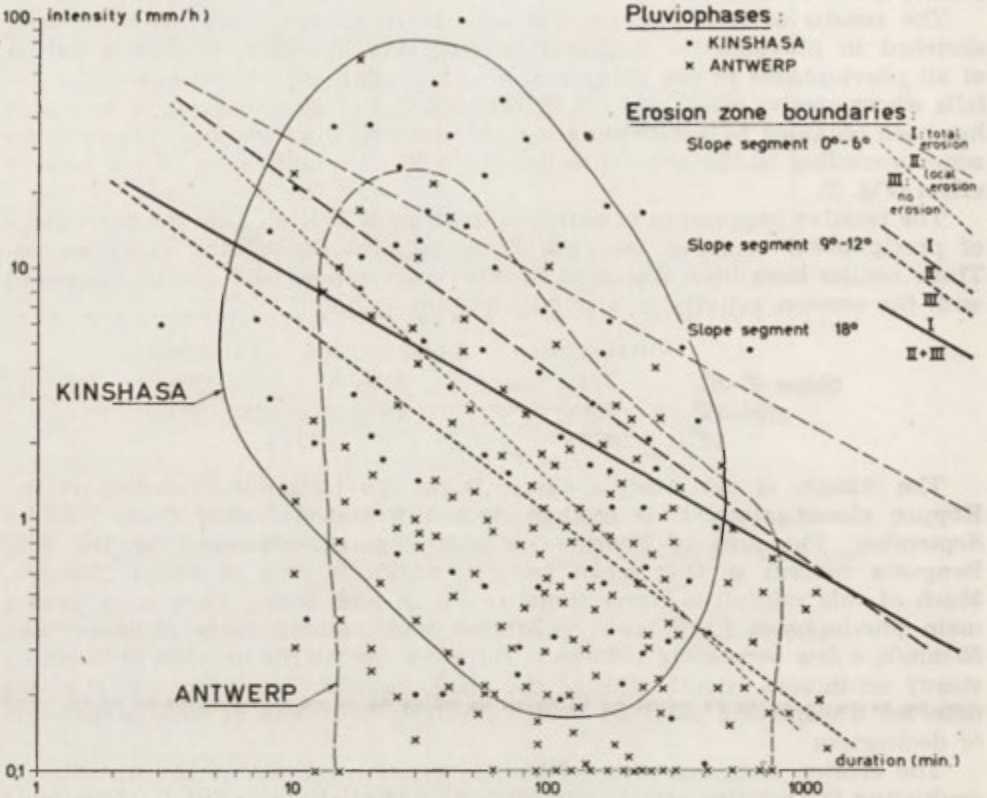


Fig. 3.

The pluviometric diagrams of both stations reveal a significant difference between the rain characteristics of the corresponding climatic belts. What then is the impact of this pluviometric discrepancy on rainfall erosion capacity?

#### COMPARISON OF EROSION DIAGRAMS KINSHASA-ANTWERP

A valid comparison between both stations implies that pluvial erosion is acting on identical slopes controlled by the same environmental conditions as related to lithology, vegetation and slope profile. This means that the Kinshasa experimental slope has to be considered as the reference slope for both stations. In order to evaluate the erosive effect of cyclonic rainfall on this slope, the pluviometric diagram of Antwerp is converted into a synthetic erosion diagram by interpolating the erosion zone boundaries for 0°-6°, 9-12° and 18° plots, as resulting from the field experiments in Kinshasa. In this way the amount of rainfall corresponding to each erosion zone and relative to each slope segment is calculated, taking into account a total annual rainfall of 685.3 mm, recorded in Antwerp in 1968. Data of Kinshasa are recalculated on the basis of mean annual rainfall (1350 mm). It is also possible to compare the erosion diagrams

of both localities where rainfall has had the three above-mentioned erosive effects (Fig. 3). Diagram A (Fig. 4) shows the contrasting effects on an absolute basis, whereas diagram B compares them on a percentage basis. Antwerp data have been corrected, as part of winter precipitation occurs in the form of snow which is assumed to have no erosive effect. According to Dr. Poppe (Climatology, Leuven) snow represents the following mean percentages of total monthly precipitation: December 15%, January 30%, February 30%, March 15%.

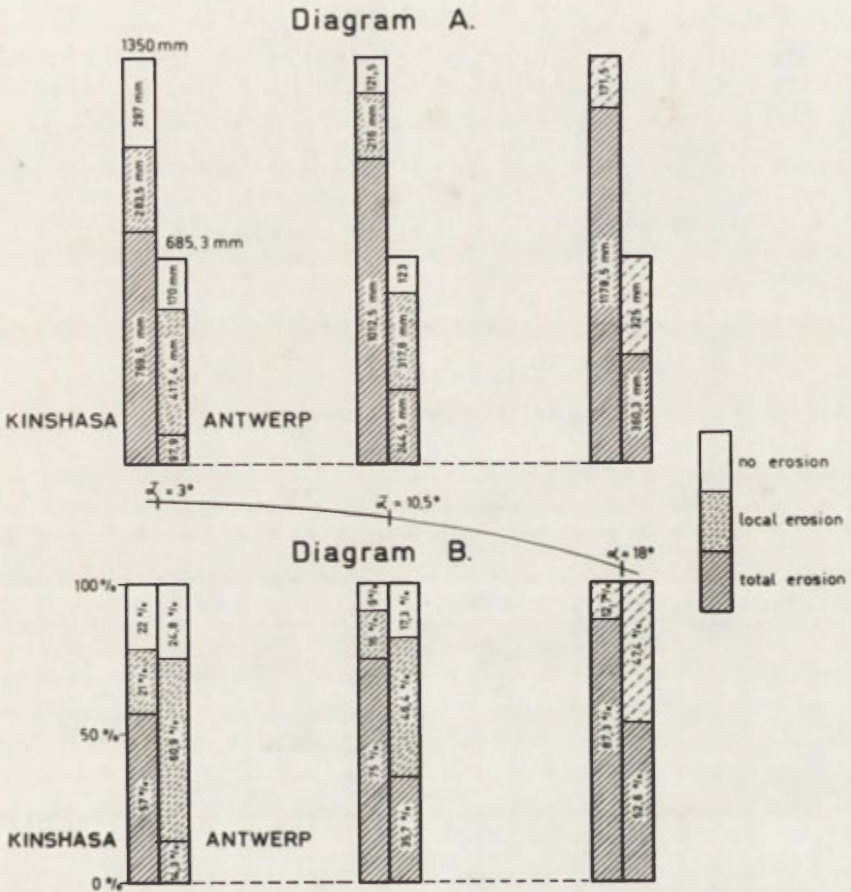


Fig. 4.

Analysis of both diagrams leads to the following conclusions:

- (1) On the experimental slope middle-latitude rainfall causing soil erosion in Antwerp should act essentially in the form of splash erosion except on medium slopes (18°), where total erosion rains become slightly dominant. Tropical storms, on the contrary, have a predominant total erosion effect on all segments; this means that intermittent runoff is active even on gentle crest slopes.
- (2) As one could expect, there is an increase in erosion where there is an increase in slope — in both climatic belts. But as the diagrams show, the erosional discrepancy between tropical and middle-latitude rainfall diminishes

downslope on the convexity. This formulation is confirmed by the ratio of total erosion rains in both localities:

$$\text{slope segment } 0^{\circ}\text{-}6^{\circ}; \frac{769.5 \text{ mm (Kinshasa)}}{97.9 \text{ mm (Antwerp)}} \approx 8 \quad (1)$$

$$\text{slope segment } 9^{\circ}\text{-}12^{\circ}; \frac{1012 \text{ mm (Kinshasa)}}{244.5 \text{ mm (Antwerp)}} \approx 4 \quad (2)$$

$$\text{slope segment } 18^{\circ}; \frac{1178.5 \text{ mm (Kinshasa)}}{360.3 \text{ mm (Antwerp)}} \approx 3 \quad (3)$$

A more representative comparison of both climates can be drawn by calculating the ratio of the percentages of total precipitation causing total erosion:

$$\text{slope segment } 0^{\circ}\text{-}6^{\circ}; \frac{57\% \text{ (Kinshasa)}}{14.3\% \text{ (Antwerp)}} \approx 4 \quad (4)$$

$$\text{slope segment } 9^{\circ}\text{-}12^{\circ}; \frac{75\% \text{ (Kinshasa)}}{35.7\% \text{ (Antwerp)}} \approx 3 \quad (5)$$

$$\text{slope segment } 18^{\circ}; \frac{87.3\% \text{ (Kinshasa)}}{52.6\% \text{ (Antwerp)}} \approx 1.5 \quad (6)$$

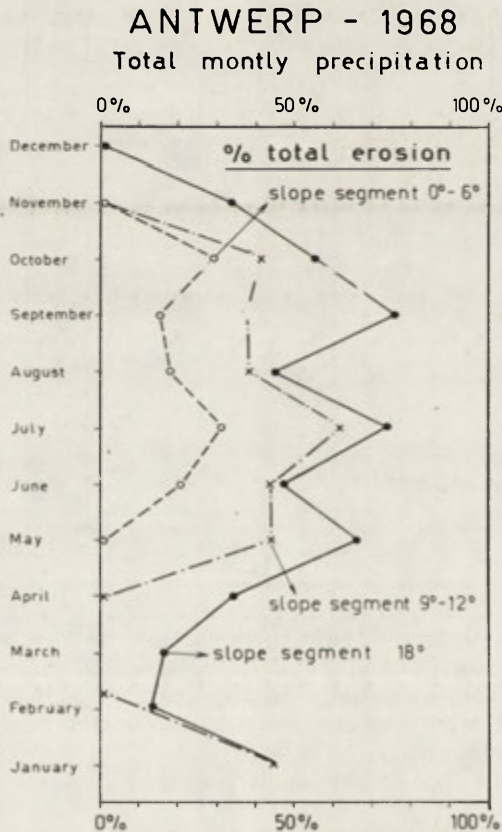


Fig. 5.

(3) Downslope erosion increases on the convexity, yet — as stated in Kinshasa (J. De Ploey 1969) — the total erosion increment diminishes downslope. Antwerp data confirm this statement, as illustrated by the following ratios:

$$\frac{\% \text{ total erosion on segment } \alpha = 10.5^\circ}{\% \text{ total erosion on segment } \alpha = 3^\circ} = \frac{35.7\%}{14.3\%} \cong 2.5 \quad (7)$$

$$\frac{\% \text{ total erosion on segment } \alpha = 18^\circ}{\% \text{ total erosion on segment } \alpha = 10.5^\circ} = \frac{52.6\%}{35.7\%} \cong 1.5 \quad (8)$$

The corresponding values for Kinshasa are approximately 1.5 and 1.15.

A diagram (Fig. 5) shows the seasonal evolution of erosion patterns as the result of middle-latitude rainfall on the experimental slope. The diagram makes it clear that total erosion, meaning erosion by combined splashing and discontinuous runoff, occurs essentially during the summer and early autumn, especially on the gentle crest slopes. In fact, convectional heavy showers are most frequent during that period, while drizzling rain is relatively more abundant during the rest of the year. In Kinshasa on the other hand, the intensity of pluvial erosion is more equally spread over the whole season lasting from September to June.

#### A RAIN ENERGY RATIO

If field factors (soil, vegetation) remain constant and correspond to those of the grassy savannah in Kinshasa, differentiation of rainfall erosion capacity or erosivity in both areas is dependent on pure rain characteristics. Both the amount of the precipitation and the intensity of rainfall are important parameters in determining total energy available for soil erosion. P. C. Ekern (1953) has discussed the impact of these parameters on raindrop erosion capacity. A. Feodoroff (1965) mentions a report by N. W. Hudson (1963) according to which soil loss resulting from splash erosion is a linear function of rainfall energy. Recently M. A. J. Williams (1969) discussing soil loss by rainsplash erosion, concluded that raindrop momentum rather than kinetic energy of rain may explain a greater proportion of the variation in soil loss. Authors seem to find difficulty in establishing a precise relationship between soil loss by runoff and rain characteristics. Nevertheless, one may reasonably assume that a positive relation exists between rain energy and runoff activity as well as splash erosion. According to W. H. Wischmeier and D. D. Smith (1958) the best variable found for prediction of soil loss from cultivated fallow is the product of the total rainfall energy of a storm and its maximum 30-minute intensity. These authors published a rainfall energy table and a regression equation of kinetic energy:

$$Y = 916 + 331 \log_{10} I \quad (9)$$

where  $Y$  is the kinetic energy in foot tons per acre inch, and  $I$  is the rainfall intensity in inches per hour. These interesting data permit us to estimate total annual rain energy  $Er$ :

$$Er = Pt(916 + 331 \log_{10} I)$$

where  $Pt$  is total annual precipitation and  $I$  is mean intensity of all rains calculated as an arithmetic mean, taking into account intensity and duration of all pluviophases fitted in the pluviometric diagram.

Comparing rain energy of tropical and middle-latitude precipitation a rain energy ratio  $Re$  may be delineated as follows:

$$Re = \frac{Pt(916 + 331 \log_{10} \bar{I})_{\text{Kinshasa}}}{Pt(916 + 331 \log_{10} \bar{I})_{\text{Antwerp}}} \quad (10)$$

For Kinshasa  $\bar{I}$  was calculated on the basis of pluviometric data of the rainy season 1964–1965 and found to be equal to 4.74 mm/h = 0.19 in/h. Antwerp data of 1968 result in a  $\bar{I}$  value of 0.96 mm/h = 0.038 in/h. Finally using Wischmeier and Smith's rainfall energy table values, a global  $Re$  value for both stations is obtained:

$$Re = \frac{54 \times 677}{27.4 \times 437} = 3.05 \approx 3$$

where 54 inches and 27.4 inches correspond respectively to Kinshasa and Antwerp mean total annual rainfall.

Values of  $Re$  can be calculated for rains with specific erosive effect according to zonation marking the erosion diagrams of both localities. The  $Re$  value of total erosion rains may be considered first, as these rains are the most energetic. Corresponding  $\bar{I}$  values were calculated taking into account the erosion diagram data:

	$\bar{\alpha} = 3^\circ$	$\bar{\alpha} = 10.5^\circ$	$\alpha = 18^\circ$
$\bar{I}$ Kinshasa	19.4 mm/h = 0.79 in/h	12.2 mm/h = 0.48 in/h	10.2 mm/h = 0.41 in/h
$\bar{I}$ Antwerp	13.7 mm/h = 0.59 in/h	6.1 mm/h = 0.25 in/h	4.9 mm/h = 0.20 in/h

Values of  $Pt$  have already been determined in the foregoing paragraph where the evolution of total erosion in relation to slope was discussed. Final computation gives the following  $Re$  values:

$$\text{slope segment } 0^\circ\text{--}6^\circ (\bar{\alpha} = 3^\circ); \quad Re = 8.27 \quad (11)$$

$$\text{slope segment } 9^\circ\text{--}12^\circ (\bar{\alpha} = 10.5^\circ); \quad Re = 4.67 \quad (12)$$

$$\text{slope segment } 18^\circ (\bar{\alpha} = 18^\circ); \quad Re = 3.76 \quad (13)$$

Since rain erosion capacity must be proportional to rain energy, the rain energy ratio in turn may be a significant parameter in a comparative study of pluvial erosivity in different climates. The global energy ratio (10) and ratios (11), (12) and (13) indicate a striking erosive discrepancy between both climates under consideration which in fact is partly due to the difference in total annual precipitation amounts. It is very interesting that this discrepancy, as far as concerning total erosion rains, decreases on increasing slopes of the experimental convexity.

The effect of the so-called local erosion rains, occasioning only splash erosion must also be taken in account. However, considering the relation between rainfall and splash erosion in Kinshasa and in Antwerp (Fig. 4, Diagram A) and the role of the intensity factor in the energy ratio, one must admit that the splash erosion effect could only slightly reduce the erosive discrepancy between both areas. This statement certainly does not mean that splash erosion is of no importance on the sandy slope (J. De Ploey und J. Savat, 1968).

## CONCLUSIONS

Tracer experiments carried out in Kinshasa on a sandy convex hillslope covered by a substeppe savannah led to interesting information on superficial

hydrology and related pluvial erosion. In order to compare the erosivity of tropical and middle-latitude rains, erosion diagrams obtained in the Congo have been re-interpreted going out from the supposition that the experimental slope has been exposed to annual rainfall registered in a weather station near Antwerp.

Confrontation of erosion diagram data of both localities brings out a striking difference between rain erosion capacity in both areas. Experiments in Kinshasa showed that combined intermittent, discontinuous runoff and splash erosion are active even on gentle crest slopes, whereas this type of so-called total erosion under middle-latitude rainfall should only occur down medium slopes of about  $10^\circ$ . This means — as illustrated by numerical data — that the erosive discrepancy between both areas decreases on increasing slopes. This statement can only be explained in function of different rain characteristics, especially concerning the distribution of high intensities in the erosion diagrams. On the  $0^\circ$ – $10^\circ$  slope of a sandy hill, runoff is conditioned by relative high values of intensity  $X$  duration which rather rarely occur in Antwerp. As proven by the balance theory (J. De Ploey 1968, 1969), ablation by splash erosion and intermittent runoff is most intense at the crest zone of convex slopes, which in this way tend to flatten. We must now conclude that the flattening of convex slopes by discontinuous pluvial erosion, occurs more quickly in the savannah climate than in the marine middle-latitude climate. This phenomenon therefore may be invoked as a factor contributing to the development of flattened multiconvex landscapes characterizing some tropical regions. In Antwerp runoff should essentially occur during summer and early autumn partly as a result of convectional thunderstorms.

Total erosion increments decrease in both localities with increase of slope. Hereby one may consider the role played by gravity parameter  $g \sin \alpha$  for  $\sin \alpha$  increments also decrease on increasing slope. Moreover intensity of rainfall theoretically decreases downslope according to  $\cos \alpha$  values and this factor may also intervene but in a minor extent.

A rain energy ratio is proposed which, from a theoretical point of view, may be considered as a significant parameter to comparison of pluvial erosivity in different climates.

As a matter of fact the mentioned data and conclusions apply to a singular slope model marked by specific and constant field parameters. The intention was primarily to evaluate the role of rain characteristics of given climates in pluvial erosion on a given type of slope. Similar investigations on other slope models should be necessary to obtain a better understanding of pluvial erosion patterns in relation to the discussed climates.

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## THE MODELLING OF MONSOON AREAS OF INDIA AS RELATED TO CATASTROPHIC RAINFALL

LESZEK STARKEL

The Indian Subcontinent together with the Himalayas is a terrain of rarely encountered contrasts in the type and intensity of morphogenetic progresses, this being determined by the differentiation of orography and geological structures, as well as by the considerable range of total annual precipitation and its intensity<sup>1</sup>. The common features of this area (with the exception of the Kashmir and the southern part of India) are the changes of atmospheric circulation between summer and winter (Ananthakrishnan, Krishnan 1962), the concentration of precipitation during the period of 3-5 summer months, and the periods of many weeks duration in the winter half-year (Figs. 1 and 2). The season of the summer monsoon, during which on average over 90% of a rain falls, constitutes the principal period for the shaping of the relief by flowing water and gravitation processes. This is well illustrated by the elongated curves of the climatographs (Fig. 1). The role of heavy rainfall is enhanced by the period of torrid heat and droughts preceding the monsoon, when the soils (especially of cultivated land) dry up and become cracked. Hence the greatest transportation of suspended matter takes place at the beginning of the monsoon (cf. Fig. 4). The participation of other processes depends on the occurrence of minus temperatures (frost weathering, cryonival processes) encountered not only in the Kashmir Himalaya as different from the monsoon thermal and precipitation regime, but also in the semi-desert Rajasthan lowlands, in the eastern Low Himalaya and in the Assam Upland (above an altitude of about 1300 m). In the dry areas devoid of plant cover wind plays an important role (Western Rajasthan).

The differentiation of rainfall, which has a bearing on the circulation of water and development of plant cover, is distinctly marked in the east-west section (Figs. 2 and 3). Simultaneously in this system, the exposure of elevated slopes has an essential influence on precipitation. Hence, while in the areas of lowlands and low uplands the mean annual total of precipitation decreases from

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<sup>1</sup> In the winter of 1968/69 the author carried out field observations, collecting comparative material in several regions of India, described in the text. His journey to India was organized within the framework of an exchange between the Polish Academy of Sciences and the Council for Scientific and Industrial Research of India. The author wishes to express his sincere gratitude to many Indian geographers, geologists, climatologists, forest officers, agronomists and other experts for their help in collecting interesting data and site observations.

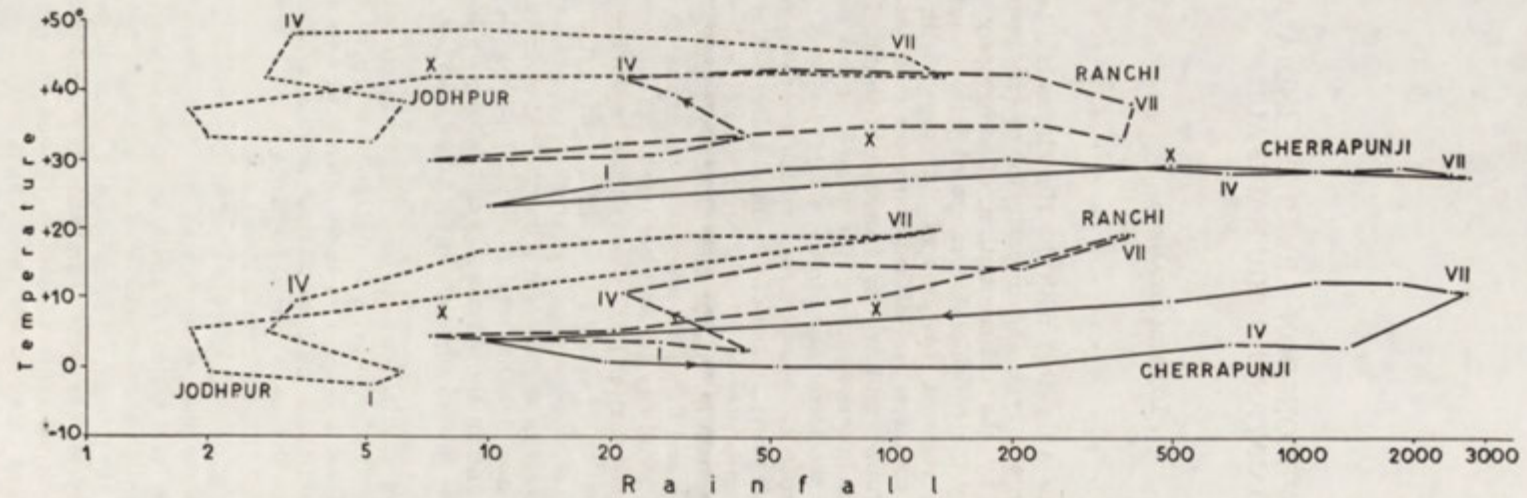


Fig. 1. Double climatographs presenting climate-process systems for Cherrapunji, Ranchi and Jodhpur. Rainfall for each month. Extremes of maximum and minimum temperatures

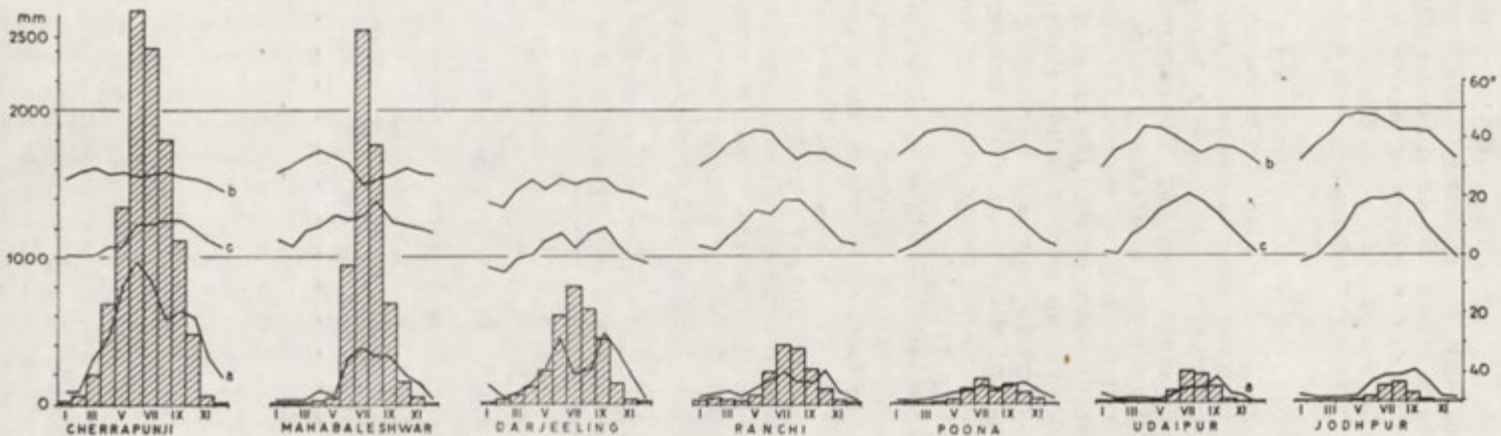


Fig. 2. Monthly rainfall, heaviest rainfall in 24 hours (a) and extremes of maximum (b) and minimum (c) temperatures based on data up to 1959 (after Indian Meteorological Department)

about 2000 mm in the eastern part of the Ganga-Brahmaputra Plain to 200 mm in the western part of Rajasthan, the contrasts in the mountain and upland areas are greater. The southern slope of the Khasi-Jaintia Hills (Assam Upland) has up to 11,000 mm of the rainfall, the scarp of the Western Ghats 6000 mm (Mahabaleshwar), whereas on the opposite slopes the precipitation is only of the order of 1500–2000 mm (Shillong, Gauhati) and even below 700 mm (Poona, cf. Fig. 2).

At the same time, however, the length of the period of precipitation and its intensity undergo changes. In the region of Cherrapunji or on the scarp of the Ghats the rainfall in some summer months is up to 2500–2800 mm, the number of days with precipitation amounting to 26–30 days, i.e., 50–100 mm of precipitation falls on the average every day. The maximum recorded daily precipitation at Cherrapunji amounted to 974 mm (in June 1956)<sup>2</sup>. The mean annual maximum daily precipitation here reaches 550 mm (Krishnan et al. 1959, Extremes 1959). As can therefore be seen the range of intensity of daily precipitation is considerable, though rain is falling almost every day. The highest daily and several days' unintermitting rains bringing about catastrophic floods occur at the beginning and at the break of the summer monsoon (Dutta-Choudhury, Das 1968); this is shown among others by the example of the Darjeeling region (Starkel, 1970, 1972).

The monthly total of precipitation in winter decreases even to 10 mm, the rains having occurred during 23 days in a 50-year period (according to monthly and annual normals 1962). In fact, every winter there are months without a drop of rain. On slopes in the rain shadow and on the Deccan Plateau (cf. Fig. 1, Ranchi) the mean monthly number of days with precipitation during the monsoon period falls below 20 at a total of precipitation of 200–400 mm and mean maximum daily precipitation of the order of 100 mm. Towards the north-west, together with the decreasing precipitation, the number of days with rain grows smaller (up to 4–6 days per month during the period of the summer monsoon) with a simultaneous increase in the suddenness of precipitation and no less high yield of daily precipitation (Fig. 2). Downpours occur when the mass of humid air breaks into their terrain, this often taking place also in winter. In the course of 2–3 days 50–80% of the annual precipitation falls on average, the daily rainfall sometimes exceeding the mean annual total of precipitation.

Apart from the amount and intensity of precipitation in the monsoon circulation a characteristic trait is the varying from year to year of the stability of precipitation periods in the east-west section, which was noted by Gowindaswamy (1953) — cf. Ramdas 1968. This author calculated the total of precipitation in a weekly series and showed its distribution in the successive years 1908–1950, comparing it with the mean of the whole series. He considered as "flood" periods (during monsoon) the years when precipitation exceeded at least twice the annual mean of the given week and as drought periods when precipitation did not reach even 50% of the mean. The diagram plotted by the present author on the basis of these calculations (Fig. 3) distinctly shows that in the period 1908–1950 the number of flood and drought weeks increases towards the west together with the decreasing total of precipitation, whereas the duration of "normal" periods with smaller deviations from the mean increases towards the east. This means that according to the situation in relation to the monsoon circulation, areas with small rainfall (e.g. Rajasthan) have from year to year the least stable precipitation regime, this being

<sup>2</sup> In the region of Bombay 983 mm were recorded (Parde 1961).

reflected in the shrivelling up of the already poor vegetation, and subsequently in the destruction of soils by wind or by a sudden catastrophic rainfall sometimes of local range. On the other hand, in spite of the dry season, in areas with a high annual total of precipitation their regime is the most stable in the multiannual course, which does not stand in opposition to the occurrence of heavy rainfalls of an intensity and duration not encountered in areas of more scanty precipitation and bringing about heavy floods.

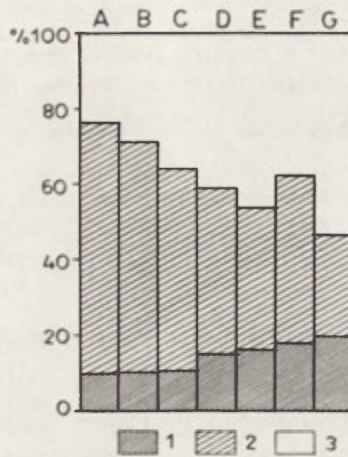


Fig. 3. Rainfall abnormalities of week by week rainfall during years 1908-1950 in percentage of duration (after Ramdas 1968)

1 — percentage of flood weeks, 2 — percentage of normal weeks, 3 — percentage of drought weeks, A — Assam, B — Bengal, C — Bihar, D — Uttar-Pradesh W, E — Punjab NE, F, G, — Rajasthan W

Catastrophic rainfalls are of varying intensity. According to this and to the degree of saturation of the substratum with water, as well as to the conditions of the whole environment, the surface runoff and infiltration determining the geomorphological effects of floods have a varying course. The discharges of Indian rivers themselves give the best information about the scale of annual changes in the circulation of water (Vij, Shenoy 1968). The largest among them, such as the Ganga and Brahmaputra, carrying water from the Himalayas distinctly show seasonal oscillations (Table 1). The Brahmaputra, rich in water, in 1962 reached discharge of 72,460 m<sup>3</sup>/s at an observed minimum of 2680 m<sup>3</sup>/s (Desai 1968). The maximum specific runoff amounted to only 0.17 m<sup>3</sup>/s, this being related to the taking over by catastrophic rainfalls of only part of the river basin. The specific flood runoff in catchment basins of the marginal part of the Himalayas of an area of the order of several to some twenty thousand km<sup>2</sup> already reaches 0.5-4 m<sup>3</sup>/s/km<sup>2</sup>, and from the areas of some twenty to several hundred km<sup>2</sup> it even exceeds 10 m<sup>3</sup>/s/km<sup>2</sup> (Parde, 1961 noted a maximum value of 21.8 m<sup>3</sup>/s/km<sup>2</sup>). In the Assam Plateau with the highest amount of precipitation the smaller streams dry up completely in the winter months, while the larger ones carry water only in 1/10 or 1/20 of the width of the channel cut out in the rock. Also rivers of the Deccan uplands show greater variations than the Himalayan ones. For example the River Mahanadi during a flood carries up to 46,000 m<sup>3</sup>/s (0.35 m<sup>3</sup>/s/km<sup>2</sup>) with a minimum observed discharge of 6 m<sup>3</sup>/s

(runoff coefficient of the order of 33%). In the River Narmada during the maximum discharge the specific runoff amounts to  $0.7 \text{ m}^3/\text{s}/\text{km}^2$ , while the River Subarmati in the Gujarat, draining the semi-arid areas, becomes seasonally dry, though during floods it carries more than  $10,000 \text{ m}^3/\text{s}$ . The runoff coefficient, high in the case of the River Narmada (40%) amounts to only 3% for the River Subarmati.

TABLE 1. Discharge and specific runoff of some Indian rivers

River (station, year)	Catchment area in $\text{km}^2$	Discharge in $\text{m}^3/\text{s}$		Specific max. runoff	References
		minimum	maximum		
Ganga (Farukka)		1568	59,500		(Vij, Shenoy 1968)
Brahmaputra (Pandu) 1962	424,309	2680	72,460	0.17	„ „
Kosi (max. in 1968)	59,540		25,840	0.434	(after official information)
Tista (in 1950)			18,680		(Krishnan 1960)
Tista (Brigde 1968)	7,200		27,500	3.8	(calculated by Starkel 1971)
Tista (Jalpuiguri 1968)			18,800		(after official information)
Little Rangit (1968)	75		5,625	75.0	(calculated by Starkel 1971)
Lish River (1952)	49		255	5.176	(Dutt 1966)
Sutlej		78	12,000		(Vij, Shenoy 1968)
Joba Khas	161		1,601	9.93	(Soil erosion 1948)
Damodar	22,000		18,410	0.84	(Vij, Shenoy 1968)
Mahanadi	132,090	6	46,000	0.35	„
Narmada	93,180		65,100	0.7	„
Subarmati	54,610	0	10,874	0.2	„

The smaller rivers of the Deccan Plateau dry up almost completely in spring, whereas those in the semi-arid Rajasthan at a mean annual precipitation total of 400–600 mm carry water only several days in a year during the period of heavy rainfall, or even once in some twenty years.

The course of morphogenetic processes is dealt with within regions of varying relief, plant cover, amount, and intensity of precipitation for

(1) Middle-high mountains with steep slopes, and monsoon rains extreme on a world scale, amounting to 5000–11,000 mm

- (a) the Assam Plateau;
- (b) the scarp of the Western Ghats.

(2) Middle-high mountains with steep slopes, and abundant monsoon rains of the order of 1500–4000 mm (the marginal part of the Himalayas).

(3) Uplands and low mountains with a relatively large amount of rainfall of the order of 500–2000 mm

- (a) the Ranchi Plateau;
- (b) the Poona Plateau;
- (c) the Aravalli Mts.;
- (d) the western part of the Siwaliks.

(4) Uplands and depressions of the semi-arid zone (Western Rajasthan) with rainfall below 500 mm.

(5) Lowlands and submontane depressions collecting flood waters from the surrounding mountains and uplands (the Ganga and Brahmaputra Basin).

## MIDDLE-HIGH MOUNTAINS WITH MONSOON RAINS AMOUNTING TO 5000-11,000 MM

The Khasi-Jaintia Hills in Assam form a vast plateau at an altitude up to 1900 m. They are built of crystalline rocks, partly of granites, with a summit area sloping eastwards to about 1200 m a.s.l. (Chatterjee 1968). They are covered here with a mantle of Cretaceous and Palaeogene sedimentary rocks. The plateau is cut from the south by a complex system of faults, sloping in a sharp scarp to the Bengal Plain. The scarp formed chiefly by Pliocene-Quaternary movements continues to be active and is dissected by deep (up to 1000 m) canyons, whose valley heads form steep-wall cirques.

The source sections have not as yet been rejuvenated. The rivers flow down in 300-400 m waterfalls. The late-mature structure of the wide valleys, long slopes, and flat erosional plains with single monadnocks (the relative heights rarely exceeding 100 m) sharply contrasts with the vertical walls of valley heads (Murthy 1968). This zone of scarps at a height of 1000-1300 m a.s.l. receives annually up to 12,000 mm of rain (Cherrapunji, Mawsynram). The summer season is a period of intense runoff and of morphogenetic processes depending, among others, on the energy of the relief; the winter season is a period of relative stagnancy of processes, even the channels of larger streams drying up. The area is covered with forests of the monsoon jungle type and above the altitude of 1500 m with forests in which conifers have a share. Unfortunately, these communities have been completely destroyed in many regions and nowadays, according to H. P. Das' (1954) estimate, 65% of the Assam Plateau is occupied by cultivated land, 8% by forests, while vast areas (15%) constitute barren land, generally devoid of soil cover.

In slope forming processes the position of the base level plays an important role. The slopes of canyons are modelled by the erosion of flowing streams, by big rockfalls and rock slumps, especially in the source cirques. The waters flowing down from the plateau penetrate into the fractures formed chiefly in sandstones and limestones during the earthquakes frequently occurring here, and by isolating big blocks, contribute to their falling or slumping. The non-rejuvenated channels on the plateau, forming local base-levels, condition the development of slopes. In the almost completely deforested granite areas there is a destruction of the primarily thick (up to 10 and more meters) lateritic covers, an exposure on the surface of core-stones, and a deposition of the excess of material carried down during heavy rainfall in the broad depressions of valleys. The downwash is facilitated by the cultivation of potatoes on slopes inclined at 25-30°. Ploughing is carried out concordantly with the direction of the flow of water, in order that the potatoes will not rot. Thus, the frequent desert-like landscapes almost completely devoid of soil cover are formed. Under these conditions with a lack of humic acids, chemical weathering cannot make very rapid progress. In winter the water percolating through a thin waste mantle brings about a fall in temperature near the ground below 0° (at night) the formation of needle ice (the author observed needles 1.5 cm high) and a loosening of the soil structure accelerates the downwash during heavy rainfall. In areas built of flat lying sandstones and shales a retreat of scarps built of resistant beds takes place from the valleys, with the aid of mass movements. The presence of thick sandstone beds determines the formation of high tablelands with vertical abrupt slopes and with rockfalls. When the sandstone beds are thin and separated by thick shales, the plasticized shales creep down the breaks during downpours, tearing up the overlying sandstones. The single sandstone blocks creep,

forming a characteristic pavement on the slopes, which is protective against slope wash. The destruction of the resistant bed often due to a flow under it (of suffosional type) brings about a rapid removal of the less resistant series with the structural surface becoming exposed. Periodical rivers also participate in this exposure. In spite of the mature profiles, a rejuvenation through the interbed wash-off takes place from the scarps, as well as a piercing of resistant beds by potholes, the formation of subsurface channels, roof subsidence, bridges formation and finally a removal of debris up to the lower resistant bed. Simultaneously, lateral erosion is progressing in the less resistant series — hence many channels have a width out of proportion to the catchment basin and to the amount of material carried. This is also favoured by the hardening of rocky channels due to the enrichment in iron compounds, which — as the author observed — proceeds from water sources at the base of slopes. In the winter season the water from these sources covers the dry river beds with a thin layer of this while evaporating during hot days brings about the formation of ferruginous crusts on the rock surfaces.

Observation carried out in the Charrapunji region showed that under conditions of exceptionally heavy monsoon rains present-day tectonic movements play a vital role in the formation of the relief. Thus a different trend of development of young canyon valleys, more intensively deepened and widened than retreating, and a different one from the mature relief of the plateau ensues. In spite of favourable conditions for chemical weathering, the lack of plant cover and torrential rains bring about a lowering of the surface of slopes and the carrying away of material. This is why, apart from the closed intramontane depressions, there is a lack of soils covering the slopes. The tendency to slope retreat is conditioned by the occurrence of the alternate rock beds of high and low resistance. Due to the seasonal excess of water, the modelling by waters flowing over the surface, as well as in the debris mantle and in the rock takes place simultaneously.

Similar observations were carried out by the author on the edge of the Southern Ghats near Mahabaleshwar (1400 m a.s.l.), where the annual total of precipitation amounts to 6000 mm (Fig. 2). The intensively dissected basalt escarpment sharply contrasts with the flat interfluves, which as a result of deforestation are completely devoid of soil cover, the soil being removed up to the foot of the scarp mountains.

#### MIDDLE-HIGH MOUNTAINS WITH MONSOON RAINS OF THE ORDER OF 1500-4000 MM

The Darjeeling Hills representing the Low Himalaya group dealt with by the author in a separate study (Starkel 1970, 1972). This region with a height of 2500-3500 m a.s.l. is dissected by young V-shaped valleys with narrow floors and convex slopes 1000-2000 m deep. It is built for the most part of gneisses and crystalline schists of medium resistance. Up to the height of about 2000 m the forest cover is being strongly destroyed, while vast areas (often more than 50% of the slopes) are occupied by tea gardens. Owing to the absorptive silty-sandy waste mantles 1-5 m thick, in spite of considerable inclinations (20-40°), the water during the summer downpours (on the average 11 days with more than 50 mm of precipitation) flows down in a subcutaneous runoff, no surface runoff or mass movements on a large scale being observed. It is only the catastrophic downpours of the order of 500-1000 mm within 2-3 days, occurring about 4-5 times a century (June 1950, October 1968), when



the hourly rainfall exceeds 50 mm, which brings about a violent liquefaction of covers as well as mudflows, and sometimes a deep infiltration into the rock, together with big landslides. As a result there occurs a single tearing down of rock mantle from about 20–30% of the deforested areas (while from only 1–2% of the forest ones), which leads to the retreat of slopes, and to the lowering and narrowing of divide areas. In the valley floors the annual floods bring about an accumulation of gravel and sand. The catastrophic floods described, with a unit discharge exceeding  $10 \text{ m}^3/\text{s}/\text{km}^2$ , lead to a deepening and widening of smaller valleys. On the other hand, the big valleys of the order of the Rangit or Tista carried masses of water (Table 1) which removed considerable quantities of material outside the mountains, in some sections deepening and in others accumulating. A flood wave of the order of 20 m in height, overflowing the valley floor, undercut the two slopes, hence the tendency to widen the valleys and carry away the material from the slopes. Catastrophic floods play a very important role here, removing the material accumulated as a result of the annual chemical weathering on slopes and of the poor fluvial transportation in the valley floors. Here also the uplifting of mountains and the lithology of the substratum are the basic factors regulating the role of the annual catastrophic precipitation. As in Assam, earthquakes are an additional factor during the period of floods and saturation of waste mantle with water (M. S. Krishnan 1960, H. J. Desai 1968). The high relief conditions the rate of processes, the rates being particularly high on unconsolidated sedimentary rocks (Dutt 1966, Jain 1966). The calculations of the transportation of suspended load in the Himalayan rivers of this zone (Manas, Kosi et al.) and the estimate of the volume of the material carried away show that while the annual denudation lowers the mountains by 0.7–2.0 mm, this value during the October downpour in 1968 was of the order of 100 mm (Starkel 1971).

#### UPLANDS AND LOW MOUNTAINS WITH MONSOON RAINS OF THE ORDER OF 500–2000 MM

Areas assigned to this type of region are the largest on India territory. On account of the varying character of the relief, the author describes 4 regions successively, which personally he had the possibility of inspecting. They are the Ranchi Plateau, the Poona Plateau, the Aravalli Mts., and the western part of the Siwaliks (Subhimalaya).

##### THE RANCHI PLATEAU

This covers the areas of the western Deccan slope falling down in steps towards the Bengal Plain (Ahmad 1958, S. P. Chatterjee et al. 1968). On the foreland at the edge of the Khamarfat hills (at an altitude of 1050 m) lies a slightly undulating plain 500–600 m high. This is built of gneissore granites and dissected by rivers flowing down towards the east and south (Barakar, Damodar, Subarnarekha, South Koel). The plain, composed of long slopes with inclinations rarely exceeding  $1-2^\circ$ , is diversified by residual and structurally controlled ridges on interfluves and disorderly scattered mounts of dome, of bornhardt type (Ahmad 1958). The old destructional surface, covered with remains of mantles of laterite type (M. S. Krishnan 1960) is nowadays dissected by rivers slowly incising, forming waterfalls shifting upwards which is characteristic of tropical areas (Machatschek 1955, Chatterjee 1963). The area receives an annual rainfall of the order of 1300–1600 mm,

concentrated in July and August (50% of precipitation). The rainfall in 24 hours rarely amounts to 200 mm. The rivers, which in winter are almost dried up, in the summer season carry large quantities of water, e.g. the Damodar up to 18,000 m<sup>3</sup>/s (Table 1) when in 1943 it formed a new channel, flowing locally at a speed of up to 26 km/h. This is an agricultural land, the monsoon forests losing their leaves for the winter season being strongly destroyed.

The present-day processes lead to the degradation of plains by wash off and to the exposure on the surface of ferruginous horizons which are also often washed out. In the parts in which the material was removed denudation already reached the massive rock, especially in the surroundings of inselbergs, whose floors inclined at 1-5°, often with single corestones, are grooved by shallow channels. In the proximity of valleys an undercutting of the foot of slopes covered with mantles takes place. The scarps formed thus are dissected by a network of gullies, whose borders retreat, becoming lower and tending at the same time to a more rectilinear free-face profile. The new undercuts give an impulse to the formation of the new generation of gullies. The author often observed 3 systems of simultaneously retreating scarps. In this way soil covers are carried away from areas lying near valleys (cf. Tejwani 1963). The magnitude of this removal is evidenced by measurements of the suspended load. For example, the River Barakar, a tributary of the Damodar, every year from 1 km<sup>2</sup> carries away 501 m<sup>3</sup> of material, which is equal to the annual degradation of this undulating land by 1.5 mm (Table 2). The measurements

TABLE 2. Annual degradation in some river catchments on basis of silt load

River, station	Period	Silt load per km <sup>2</sup> in m <sup>3</sup>	Mean annual degradation in mm	References
Brahmaputra (Pandu)	long period	770	0.77	Desai 1968
Kosi (Barahakshetra)	1948—1953	1742	1.5—2.5	Ahya 1955
Manas	long period	667	0.67	Desai 1968
Sutlej (Bhakra)	„	604	0.60	after Central Water
Barakar (Maithan)	„	1501	1.50	Power Commission of India
Darjeeling Hills (flows on slopes)	heavy rain (Oct. 1968)	200,000	300	based on measurements of slope degradation (Starkel 1971)

of the suspended load from the neighbouring catchment basins of Kangasabati and others showed that the culmination of its transportation occurs at the beginning of the monsoon, considerably outpacing the period of the greatest floods (Fig. 4, Sen 1968). It is a value approximate to the transportation of the Himalayan rivers.

Rivers with considerable variations of discharge overflow widely, flowing in numerous arms within flood plains of a meandering course and only in waterfall sections do they run into narrow canyons. The edges of waterfalls retreat slowly, since the river at the most carrying coarse sand has nothing with which to erode.

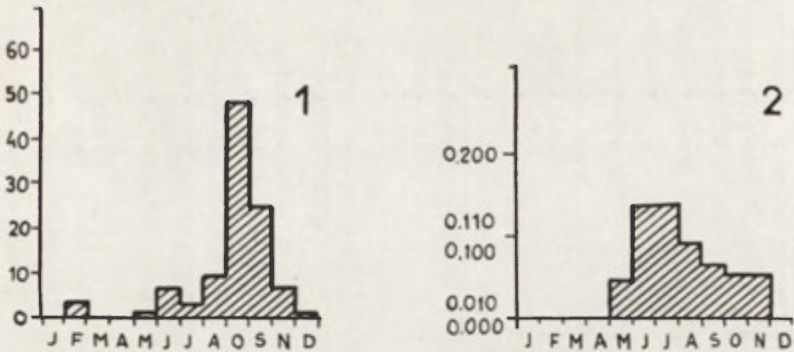


Fig. 4. Typical hydrograph (1) and silt charge diagramme for Chotanagpur area — Kangasabati River at Ambikanagar (after S. Sen 1968)  
1 — monthly discharge in 1000 m<sup>3</sup>/s, 2 — silt percentage of discharge volume

The Ranchi Plateau presents an example of a slowly rejuvenating plantation surface denuded by monsoon downpours under conditions of plant cover destruction by man.

#### THE POONA PLATEAU

The hinterland of the Western Ghats is built by horizontally lying covers of basaltic trap. Rivers of the Krishna system have formed broad valleys here which are widened by pediplanation processes into vast basins. They are surrounded by guyots of a relative height of 200–400 m protruding from their floors, with flat structural interfluves, steep scarps of slopes, inclined often at more than 50°. These slopes dissect gullies, while at their foot pediments are lying with an inclination decreasing from 10° to 2°, covered lower down with sandy-silt mantles. In the valley floors rivers often cut down to the hard bedrock carrying away large quantities of waste. The area lies in the rain shadow of the Western Ghats; it receives 500–700 mm of rain yearly (Fig. 2). In the summer months the rainfall never exceeds 200 mm. However, downpours occur which carry away large quantities of suspended load, similar to the Ranchi Plateau. Near Sholapur the Dry Farming Research Station carries out investigations on soil erosion. It was noted there that on average 20% of rain flows down, while soil denudation amounts to 4.2 mm/year. Detailed investigations carried out in the years 1934–39 on plots (cf. Table 3) showed that soil covered with natural vegetation is degraded by about 0.05 mm/y., that with freshly cut vegetation by 6 mm/y., and cultivated soil by 13 mm/y. (Soil erosion 1948). The downwash takes place during downpours when often 50% of the water runs off (Table 3); the amount of material carried away in the case of a particular downpour amounting to 1/3 of the annual denudation. The results of investigations carried out at the Hagari station in the State of Madras (where the rainfall amounts to 1000 mm) showed that colloid clay is chiefly washed out, while the sand remains (the soils contain on average 44.9% of clay, which in the material removed is present in 56.8% — Soil erosion 1948).

The present-day processes on the Poona Plateau therefore lead to the removal of waste material similar to the Ranchi Plateau. Here, however, in the semi-arid climate where the plant cover is poor, there simultaneously occur a retreat of slopes and development of pediments. This is inhibited by

TABLE 3. Runoff and soil degradation on some experimental slopes in India

Station	Land use	Area	Period	Rainfall in mm	Runoff in %	Silt loss t/km <sup>2</sup>	Silt loss m <sup>3</sup> /km <sup>2</sup>	Degradation in mm	References
Sholapur Dry Farming Research Station	natural vegetation	50 m <sup>2</sup>	1934-39	659 (mean annual)	3.44		48	0.048	(Soil erosion 1948)
	natural vegetation after cutting under cultivation		--		15.50		6070	6.07	
			--		16.61		12,920	12.92	
			28/9.IX.1938	91	51	812.6	406	0.406	
Kota, Soil Conservation Centre (slope 1%)	natural cover	40 m <sup>2</sup>	1957-60	502-778	5.56	39.2	19.6	0.020	(Singh, Dayal, Bhola, 1967)
	cultivated fallow		--	--	19.33	372.4	186.2	0.186	
	natural cover		1957	778	4.82	133.2	66.6	0.067	
	cultivated fallow		1959	622	30.24	955.4	477.7	0.478	
Jodhpur Central Arid Zone Research Institute	Prosopis spicigera community	68 m <sup>2</sup>	1963	179.6	0.4				(Krishnan, Blagoveschensky, Rakhecha 1963)
			10.V.- 6.X.1964	179.6	23.0				
			18.VIII.64	138.2	43.8				
Punjab Forest Dep. Research Station (slope 25%)	scrubs, gras (90%) grass (80%) soil without vege- tation			1092	9.16	558.5	279.2	0.279	(Soil erosion 1948)
				--	14.55	545.3	272.7	0.272	
				--	47.51	1026.1	513.0	0.513	
Dehra Dun Soil Conserva- tion Centre	forest,	1-1.5 ha	VII-VIII.1967	1038.4	0.6-1.5				(Singh et al. 1968)
	maize terra- ced slope	.	VI-IX.1966	1735	60.0	770	385	0.385	(Gupta et al. 1968)
			--	--	19.8	207	103	0.103	--
slope 8%	.	1967/8	1928	22.0	130	65	0.065	(Tejwani et al. 1968)	
		--	--	3.5	10	5	0.005	--	

the cultivation of soils leading to the dissection of these plantations by a network of gullies, probably formed on the lines of roads.

#### EASTERN RAJASTHAN

The Aravalli Mts. and plateaus lying east of them, at an altitude of 300–1000 m, form a number of basins from which isolated structurally controlled ridges protrude, built of ingenous and metamorphic rocks of Precambrian age (Machatschek 1955). The basins which are often closed and are surrounded by rocky floors cutting the steeply folded rock beds, developed intensively in a climate that was probably drier than the present one. The slope retreat continues to occur, though a general tendency to the dissection and deepening of valleys is observed. Nowadays this area received annually 500–600 mm of rainfall. Forests of the “dry deciduous forest” type are strongly attacked. The water, not being able to stop in the thin cover, flows down — hence the majority of streams carry it only seasonally. Data for the River Sabarmati which is characteristic of this region show that this seasonally dry river with a catchment basin of 54.6 thousand km<sup>2</sup> during flood carries even more than 10,000 m<sup>3</sup>/s (Vij, Shenoy 1968).

Observations carried out at the station in Kota (Soil Conservation Research, Demonstration and Training Centre) show that the plant cover plays the most important role during downpours (Table 2). On sodden slopes inclined at 1% about 5% of water flows down, bringing about downwash of the order of 0.02 mm/y., on cultivated slopes 20–30% of water and a downwash of the order of 0.2–0.5 mm/y. (Singh, Dayal, Bhola 1967). These are values lower than in the Bihar, but soils here are more sandy and the amount of rainfall is smaller by half. The water flowing down over slopes with a minimal inclination leads, when the steepness of the slope increases, to the formation of gullies occupying immense areas in the basin of the River Chambal (Gorrie 1957).

On the uplands of the Eastern Rajasthan a similar contemporaneous evolution of the relief is observed as in the other parts of the Deccan already discussed. The tendency to a deepening of valleys and retreat of slopes is accompanied by a washing down of soils and formation of gullies within the thicker covers.

#### THE WESTERN PART OF THE SUBHIMALAYA (SIWALIK)

The marginal part of the Himalayas is built of poorly cemented conglomerates, sandstones, and mudstones of Neogene and Lower Quaternary age. These sediments, uplifted to a height of 1000 and more metres, form a number of slightly undulated folds and dislodged slices flatly or steeply overthrust towards the south. These persisting uplift movements and overthrust (Krishnan 1960, Shome, Dyal 1966) favour the intensive dissection of the tectonic relief often to 500 and more metres, and to the formation of scarps, monoclinical ridges, resequent ramparts, or inversional mesas. With precipitation concentrated in the summer season of the order of 1000–2000 mm, occurrence of steep slopes, and often of convex slopes this area is dissected by a dense system of valleys. Presumably, the substratum does not generally favour the formation of richer water reservoirs, as periodical rivers prevail here. Only the big Himalayan transit rivers (Sutlej, Beas and others) carry water all the year round, though the variations of discharge in them are also considerable (e.g. the Sutlej from 78 to 12,000 m<sup>3</sup>/s — Vij, Shenoy 1968).

The greatest degradation occurs during the period of summer downpours, though measurements of suspended load in big rivers do not reveal it (e.g. the Sutlej  $600 \text{ m}^3/\text{km}^2/\text{y}$ . — Table 2). Violent mass movements, especially at the fronts of active thrusts (cf. Kalagarh landslip — Nossin 1967) take place simultaneously with the river floods, as happened near Darjeeling. Hence the formation of wide, flat valley floors entirely occupied by braided rivers displacing sandy or gravelly bars and convex slopes. Observations in the basin of the Joba-Khas stream showed that during the flood it carried about  $1600 \text{ m}^3/\text{s}$ , which equals a specific runoff of  $10 \text{ m}^3/\text{s}/\text{km}^2$ . At smaller floods it was noted that the turbidity amounted to 20%, and about  $630 \text{ t}/\text{km}^2$  were carried down within one hour, i.e., the catchment basin was lowered on the average by 0.3 mm (Soil erosion 1948). Vegetation here plays a very important role. The investigations of the Forest Department Research Station in Punjab showed that on a slope inclined at 25% when the precipitation amounted to 1092 mm, the flow on a field devoid of vegetation was five times greater (Table 3) than on a slope with shrubby vegetation, but the wash off was only twice as great (of the order of 0.5 mm).

Observations carried out in the Dehra Dun Basin at an agricultural station (Tejwani et al. 1968) showed that from forest plots only up to 5% of precipitation flowed down, on meadow 3.5%, and on maize with ploughing transverses to the contour lines as much as 60%. The magnitude of the lowering is of the order of 0.005 mm on the meadow to 0.4 mm on maize (cf. Table 3).

These observations prove irrefutably that the magnitude of denudation increased incommensurably after the deforestation of the Siwaliks, as a result of the flow of water not being regulated by the vegetation. That is why, in contradistinction to the divide areas, the relief of valleys of this active orogenic zone lost the traits of a young landscape, while the areas of intramontane depressions, often being synclinal zones as well, are littered with covers of torrential cones.

#### UPLANDS AND DEPRESSIONS OF THE SEMI-ARID ZONE WITH RAINFALL BELOW 500 MM

The western Rajasthan forms a platform lowering towards the west from a height of 400 to 150 m a.s.l. From this plain dissected by periodical rivers (Luni River) and in the part composed of undrained basins (e.g. the Sambhar Lake basin) isolated inselbergs, mountain belts, and upland outliers of plateau character are rising built of old igneous, metamorphic, and effusive rocks at a relative height of the order of 50–200 m (Pandey 1965, 1968; S. Singh et al. 1966). This plain is apparently homogenous. It is composed of pediments inclined at 1–5° surrounding the hills, passing into accumulation plains covered with river and deluvial sand often some twenty centimetres thick (P. C. Chatterji et al. 1968), and having in their floor an older fossil relief of valleys. To these plains, nowadays transformed by aqueous and eolian processes, slightly lower valley terraces relate, as well as flood plains, and terraces surrounding closed lake basins.

The amount of rainfall increases towards the east, however, not exceeding 500 mm (mean for the years 1901–1960). The precipitation is violent (its intensity amounting to more than 5 mm/min was noted) followed by long periods of drought. In some years no torrential rains occur. Catastrophic rainfalls are often of local character. For example at the station Nawa on the border of the Sambhar Lake in the years 1963–1968 226 to 935 mm of rain

was falling annually. In the peak year 1968, 18 days with precipitation were recorded there, 8 days of which with a rainfall exceeding 20 mm in 24 hours. The minimal slopes and the thick permeable sandy covers almost completely devoid of vegetation inhibit the transportation by water on long sections—hence the greater importance of eolian erosion (Ghose et al. 1968). The dry western winds of considerable strength, particularly frequent in spring and summer (on the average more than 10 days with wind of a velocity  $> 20$  km/h in a month) blow out only the soils, forming migrating systems of dunes entering even on the slopes of rocky hills, but also blow away the material from river beds or else cover the latter and contribute to the development of depressions of the taphoni type on the walls of monadnocks (they are particularly well developed on the windward side).

The action of flowing water limited to several days in a year is concentrated within the exposed bedrock. The waters flowing down from the steep ridges run superficially over plains of rocky pediments carrying coarse, sand and gravel deposited in the form of rhythmically stratified sediments. The catastrophic downpours wash out these covers, raising big fans in the lower parts of the plains. Within the pediments often more than 40% of the heavy rainfall amounting to over 100 mm flows down superficially (cf. Table 3). Station investigations showed that runoff already starts at a rainfall of 6 mm (A. Krishnan et al. 1963). In dry years almost no runoff is noted. The cover of herbaceous plants and shrubs plays a very important role in the degree of degradation, this vegetation being continuously destroyed by the grazing of big herds of sheep, cattle, goats and camels. Experiments carried out by the Central Arid Zone Research Institute showed that after suspending grazing for some years the plains grew green with grass and numerous shrubs appeared on them (*Prosopis spicigers* and others). The rivers of this area flow only occasionally, usually during several days in a year, often once in several years. Sand-bars and levees deposited in the channel are blown by wind (Pandey 1965, Singh et al. 1966). However, in spite by their short duration catastrophic downpours direct the development of slopes and valleys in this area.

In February 1969 the author observed in the vicinity of Sambhar Lake the effects of catastrophic rainfall, which had occurred on the 12th July 1968. This greatly exceeded the mean for the several previous years (Fig. 5). The Sambhar Lake forms a separate catchment basin surrounded by inselbergs, on the slopes of which are traces of lake terraces testifying to a more humid climate. This also corroborates investigations carried out by Singh (1967), who demonstrated that the nowadays salt lake about 3000 years ago was a fresh-water lake.

After three days of precipitation, during which 182 mm of rain had fallen, and the sandy soils were saturated with water in the night of the 11th–12th July rain fell uninterruptedly for several hours. An exceptional rainfall amounting to 488 mm was then noted at Nawa (greatly exceeding the probable extreme daily rainfall expected once in 250 years). The runoff was so violent that the sandy levels falling down over slopes inclined at about  $15^\circ$  towards the basin of the Sambhar Lake became cut through by gullies up to 10–15 m deep. Some of them had formed at the contact with inselbergs and sandy covers, the rocky slopes being excavated. The waters flowing down the channels of ephemeral streams widened and deepened them. For example over the bottom of the Trutmati stream a wave of water several metres high flowed for 14 hours (the first time for 40 years). As a result, the bottom was deepened by 4–6 m, while the only cultivated plots in this area localized just

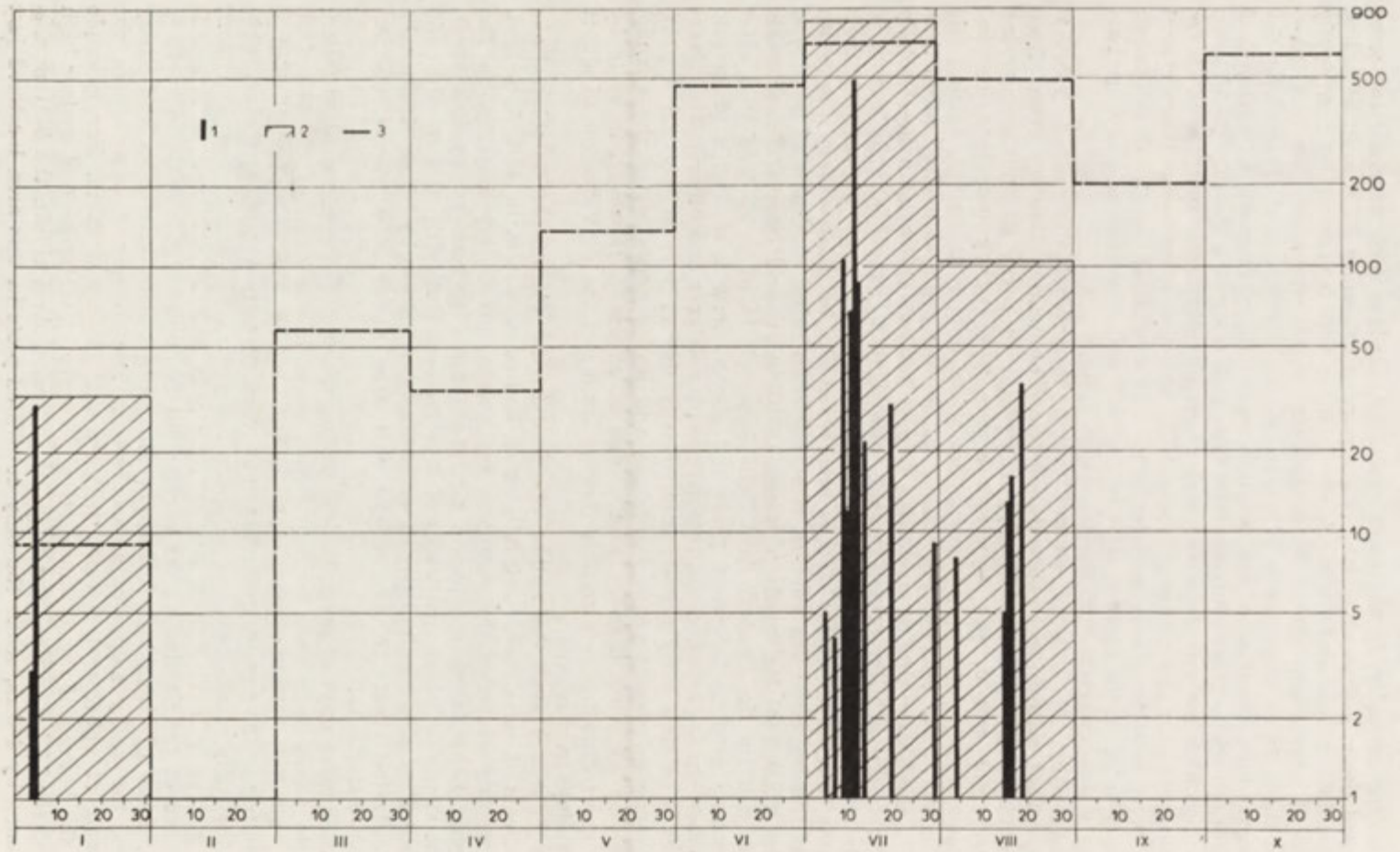


Fig. 5. Daily and monthly rainfall at Nawa near Sambhar Lake in 1968

1—daily rainfall at Nawa, 2—monthly rainfall at Nawa, 3—monthly rainfall at Darjeeling in 1968 (for comparison)



in this valley, were completely destroyed. The level of water in the Sambhar Lake (with an area of about 200 km<sup>2</sup>) rose by 1.5 m, and the plots of salt exploitation were flooded. Into one of the deepened channels, in the course of 7 months after the downpour, a cover of eolian sands 0.5 m thick was blown near its border. This testifies to the role of wind in levelling the depressions.

The part played by catastrophic downpours in the semi-desert Rajasthan though ephemeral and occurring on the scale described above once in many years, is still essential to the trend of slopes evolution (cf. the observations of the effects of the heavy rainfall in Tunisia, Mensching et al. 1970). The causes of the limited range of annual downpours are the considerable permeability of the substratum and the minimal gradients, thus the lack of uplifting tendencies observed if only in the neighbouring Aravalli mts.

#### LOWLANDS AND SUBMONTANE DEPRESSIONS

Waters flowing down from the Himalaya and from the Deccan and Assam Plateaus carry a considerable amount of material. The suspended load itself carried away by rivers amounts annually to 600–2500 m<sup>3</sup>/km<sup>2</sup> not counting the coarse material building the roots of the Sub-Himalayan cones. The Deccan rivers carry such considerable quantities of suspended load that for example Krishna River enlarges its delta into the sea by about 20 m annually on a front of 22 km. Here we shall characterize more closely the big foredeep drained by the river systems of Ganga-Brahmaputra and Indus. This foredeep lying along parallel of latitude is distinctly asymmetrical — the big cones of the Himalayan rivers drive the waters of rivers draining the graben of the Ganga and Brahmaputra under the border of the southern uplands. On the other hand in the Indus basin the parallel Himalayan rivers form a converging system of the complex cone of the Punjab.

The asymmetry of the river system in the eastern part reflects the flood violence of the Himalayan rivers and the carrying by them of large quantities of coarse material often 20–50 cm in diameter. This is the result of the considerable gradient of these rivers up to the border of mountains (effect of the lasting uplift) and of the supply of material from the deepened valleys in various climatic altitudinal belts with the glacial one inclusively. A sandy fraction is often carried in suspension, especially mica sands proceeding from metamorphic rocks (the so-called flash floods — Dutta-Choudhury et al. 1968). The monsoon downpours are not responsible for all floods in mountain rivers, especially at the beginning and at the end of the monsoon.

Apart from rain floods occurring every ten or more years within the fan of the given river, there also occur floods related to the violent break of lakes due to landslide damaging activities, as for example in the Birch Ganga valley in 1894, when the velocity of the wave flow amounted to 13.5 m/s and the discharge was of the order of some ten thousand m<sup>3</sup>, or else the big landslip and flood in the Indus valley in the years 1840–41 (M. S. Krishnan 1960). Earthquakes also contribute to the violence of floods. Thus, for example in the Assam Brahmaputra basin out of the 16 earthquakes in the years 1895–1960 as many as 12 occurred during monsoon floods (according to Dutta-Choudhury). The fans of Himalayan rivers with an average length of 100–200 km form systems related to one another. Their gradient at their base varies from 1 to 5%, decreasing near the outlet to the Ganga or Brahmaputra often to less than 0.1% (Vij, Shenoy 1968).

During catastrophic floods the river (as can be seen on the example of

the Kcsi or Tista at a discharge of over 20,000 m<sup>3</sup>—Table 1) grooves a vast zone of a fan, accumulating here the coarse material dragged along its bottom—it resembles a proglacial outwash tract, composed of a number of arms. During the annual smaller floods of the order of several thousand m<sup>3</sup>/s the river adapts its bed to smaller discharges and velocities. Within a wide zone it forms a channel with a meandering course, developing by the accretion of bars and raising of levees stabilizing the channels. The next catastrophic flood either transforms the channel again into a wide zone of braided rivers, or else the water breaks the levee immediately at the outlet from mountains, forming a new system of braided channels in another part of the alluvial fan. Sometimes 2 or 3 new channels develop and then the rivers in the following years form meanders of a correspondingly smaller curvature. However, many Himalayan rivers continuously carrying coarse material permanently maintain a rectilinear course. Through the migration of channels within the fans the latter grew, retaining their convex profile. Sometimes the migration has a unidirectional trend, as for example the River Kosi which shifted in the course of 200 years westwards by 112 km (Fig. 6, Ahya 1955, Sen 1968a, and others).

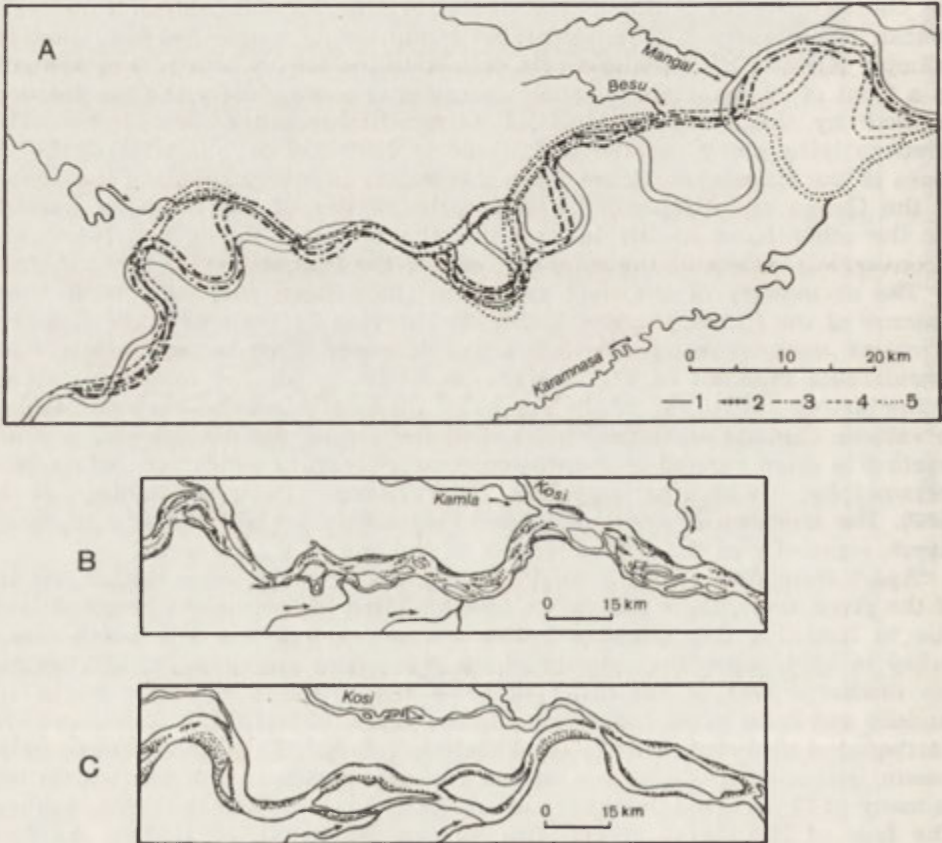


Fig. 6. Changes in the course of Ganga River (U. Singh 1968)

A — changes between Varanasi and Buxar (1-5 river course of different age), B-C — changes near the confluence with Kosi River, B — Ganga channels in 1801 (after Colebrooke), C — Ganga channels in 1964

Rivers carrying away waters from the lowland of the mountain foreland — the Ganga and Brahmaputra have smaller gradients (often below 0.1‰) — carry large quantities of suspended load in summer (Fig. 7). They are rich in water i.e. show no such variations of discharge as their tributaries (cf. Table 1). Floods never occur in their whole basins. The flood channels of these rivers are adapted to contain considerable discharge, but they still rise every year. In the Brahmaputra valley in Assam an average of 8100 km<sup>2</sup> were flooded annually and 25,100 km<sup>2</sup> during a maximum flood (Desai 1968). This

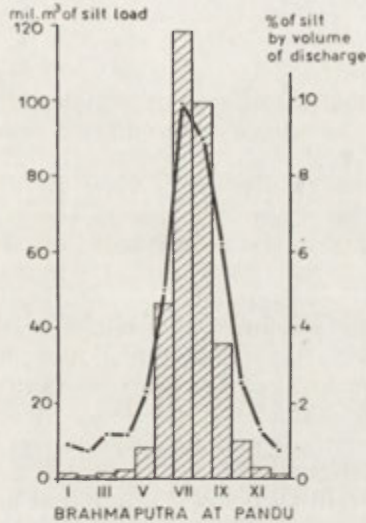


Fig. 7. Monthly silt load and percentage of silt by volume of discharge in Brahmaputra at Pandu (after records presented by Desai 1968)

is the reason why these rivers support the rarely rising tributaries with levees which for a number of years flow parallel to the main river (Fig. 6). Numerous studies concerning changes of the Ganga and of the other rivers, based on historical sources and old maps (U. Singh 1968), showed that these rivers rapidly adapt the system of their channels to the changes of the regime. In sections not reached by the larger inflows from mountains for example near Varanasi, they form gentle arches of meanders, whereas in inflows supplying large quantities of sand they form systems of braided channels — this being corroborated by the relation noted by Leopold et al. (1964) of the type of channel with a ratio of the suspended load to the bed load. For example each new flood on the river Kosi brings about considerable changes in the lower lying section of the Ganga.

In the development of mountain foreland fans and river channels a change of the regime in the last century is marked, being the result of advancing deforestation. For example, the arm of the Ganga-Brahmaputra until the sixteenth century formed regular meanders (Sen 1968b), while later the increase in the bed load led to the breaking of meanders and to the rivers becoming erratic. The areas of waste land recorded in the Punjab and related to the erosive-accumulative action of Sivaliks zone rivers, the so-called "cho", covered 194 km<sup>2</sup> in 1852, while owing to deforestation in 1897 they increased to up to 388 km<sup>2</sup>. This was the reason why in 1900 the so-called

"Chos Act" was issued in Punjab forbidding the cutting down of vegetation (Soil erosion, 1948).

The submontane depression, covered by rivers rising during the periods of monsoon downpours, has distinct traits of an active depression of the mountain foreland filled with clastic material proceeding from the dissected Himalayas. The present-day violence of processes, the amount of material carried, and the braiding of rivers were intensified by the deforestation of mountains. Water plants built at the outlets of valleys from mountains begin to counteract this detrimental phenomenon.

## CONCLUSIONS

The review of morphological effects in various kinds of monsoon climate of central and northern India with a differentiated relief permits the following conclusions to be drawn:

(1) From the geomorphological point of view the distribution and intensity of precipitation in the course of a year is a more essential trait than its amount. Catastrophic rainfall leading to the development of new forms or to the rejuvenation of old ones is a characteristic trait of a monsoon climate. During the period of such rainfall there simultaneously occurs a modelling of slopes and valley floors, this not being a feature of a temperature or periglacial climate for example (cf. Tricart 1962, Rapp 1960, Büdel 1969). The lack of precipitation in the winter months, even in Assam is evidence that one cannot speak in the dry season of processes typical of the dry zone. Strahler's (1965) or Wilson's (1968) classification does not adequately characterize the morphogenetic specificity of the monsoon climate with a large amount of precipitation. It is only in Rajasthan with an annual rainfall below 500 mm that processes typical of a dry zone occur.

(2) Catastrophic downpours of the summer rainy season are the principal factors of the transformation of the relief (Fig. 1), both in the humid and dry variety of the monsoon climate. In the humid variety the preparatory role (often levelling the forms) is performed by the annual monsoon rains of a long duration, causing a deep chemical weathering, a slow gravitational transportation on slopes and accumulation in valley floors, as well as by the periods of spring drought bringing about deep cracks in the soil, facilitating the wash off and formation of gullies. In the dry variety this levelling and preparatory role is performed by wind and physical weathering.

(3) The trend of transformations of the relief depends on the type of forms, on the geological structure, and position of the erosional base and its fluctuations related to tectonic movements.

Catastrophic downpours produce a different effect in a mountain area with considerable oscillations and being uplifted, leading to the lowering of slopes with simultaneous deepening of valleys, and another effect in the poorly rejuvenated area of planations or inselbergs where the retreat of slopes may occur with a simultaneous accumulation of alluvia and erosional deepening of river channels carrying away large quantities of suspended load.

One cannot, therefore, set apart either the contemporaneous processes conditioned by the climate (Klimatisch-dynamische Geomorphologie, Büdel 1969) from the type of relief conditioned by changes of climate (klimatisch-genetische Geomorphologie) or the present-day tectonic tendencies of the relief. The so-called typical landscapes of the tropical, savanna or desert zone were described from stable, slowly uplifting shields of Africa, Asia or South America.

(4) The devastation of plant communities by man also brought about in the monsoon climate of India an increase in denudation not only often dozens of times greater (e.g., Darjeeling), but also hundreds and thousands times greater in areas with gentle, almost flat slopes. Here the sheet runoff favours the retreat of slopes. This is due to the surface runoff and to the change of the soil structure (nowadays almost completely destroyed). Under natural conditions the soils of the Assam uplands or Darjeeling Hills were adapted to receive considerable quantities of water and to a deep infiltration or to a relatively delayed subcutaneous runoff. In the upland areas of India terrains occupied by an intensively developing net of gullies are estimated at over 22,000 km<sup>2</sup> (Tejvani 1963).

In semi-arid areas with an unstable hydrological regime the intensive animal husbandry and tillage brought about, beside a violent though local water erosion, an increase in eolian processes.

Apart from uplift movements, the present-day catastrophic course of morphogenetic processes of India is the measure of the effects of man's interference in the environment of the monsoon climate.

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## MÉTHODES D'ÉTUDE DES VERSANTS ET DE LEUR ÉVOLUTION GEOMORPHOLOGIQUE ACTUELLE EN MILIEU INTERTROPICAL

JEAN ALEXANDRE ET JULES ALONI

Les méthodes que nous présentons ici sont celles qui sont ou vont être utilisées au Centre d'Etude des Sols de l'Afrique Centrale de Lubumbashi. La plupart de ces méthodes ont déjà été décrites ailleurs: nous avons fait un choix en fonction du milieu intertropical. Ces méthodes peuvent d'autre part être regroupées en trois ensembles où le degré de détail va croissant. Le chercheur en adoptera un suivant le temps et les moyens dont il dispose.

L'étude a été centrée sur le ruissellement aréolaire (du ruissellement diffus au ruissellement en film suivant l'épaisseur et la continuité de la lame d'eau). Ce processus est le plus répandu, mais ses effets quelquefois importants sont toujours trop discrets. Des phénomènes spectaculaires comme le ravinement ou les glissements de terrain demandent une analyse spécifique en apparence plus aisée et qui sera envisagée ultérieurement. Par contre, les mouvements dus à l'impact des gouttes de pluie, à la biosaltation, au microglissement et aux coulées boueuses de très petite taille sans lesquels les effets du ruissellement en surface se réduiraient à peu de chose, ont été considérés comme formant un tout avec ceux-ci.

Enfin, les méthodes considérées n'ont été appliquées jusqu'à présent que dans des milieux subnaturels où les associations végétales ont été modifiées par l'homme soit à la suite d'une exploitation plus ou moins systématique (cueillette, bois de chauffage), soit par la pratique du feu de brousse. Les cultures, en modifiant la végétation, la partie superficielle du sol, la microtopographie et le cycle de l'eau, imposent une adaptation voire un changement plus profond de l'analyse.

Le milieu intertropical offre au ruissellement un cadre spécifique en ce qui concerne la topographie, la géologie et l'hydrologie (précipitations exclues).

(a) Un complexe pente douce-versants raides où les valeurs intermédiaires sont peu fréquentes relativement à d'autres milieux. Dans les zones occupées aujourd'hui par la forêt claire, la savane et la steppe, les surfaces d'aplanissement tant par pénélplanation que par pédiplanation, imposent leurs lignes horizontales ou presque. Les pentes fortes restent très localisées: les buttes résiduelles, les versants dominant le piémont, des vallées encaissées, peu nombreuses car les niveaux de base locaux respectés par des rivières au faible pouvoir érosif limitent la dissection des surfaces planes. Dans certains secteurs de forêt dense ou dans des régions fortement soulevées par une activité tectonique, les surfaces planes cèdent le pas aux pentes raides, domaine des glissements de terrain.

(b) Quelle que soit la pente, le manteau d'altération est rarement négligeable. Il dépasse très souvent une dizaine de mètres et constitue une proie facile pour l'érosion dans la mesure où les conditions hydrologiques, elles-mêmes contrôlées par l'épaisseur des altérites, ne constituent pas un facteur défavorable. En outre, il tend à estomper les différences de nature pétrographique du substratum.

(c) Enfin, dans la partie supérieure de ce manteau d'altération, se trouve très fréquemment un "horizon", stone-line ou cuirasse latéritique qui en modifie le drainage et la résistance à l'érosion.

Certaines de ces formes de terrain sont rarement fonctionnelles et ces dépôts ne sont pas toujours en voie de formation. La plupart du temps, ils sont dus à des systèmes morpho-climatiques antérieurs, quaternaires voire très souvent tertiaires et dont l'action a pu se répéter à plusieurs reprises et être par conséquent, assez longue. L'érosion des versants dépend, quelquefois dans une grande mesure, de ces caractères hérités. Ce fait doit rester présent à l'esprit lorsque l'on veut comparer l'activité du ruissellement sous des climats différents ou le travail accompli sous un même climat à des moments différents de l'histoire géomorphologique. Dans la plupart des cas, une courte étude du passé géomorphologique sera non seulement nécessaire à l'interprétation des observations plus précises sur le ruissellement mais devra aussi guider le choix de leur localisation.

Les autres caractères du milieu intertropical qui influencent le ruissellement et son action géomorphologique ont heureusement une réponse plus rapide aux changements de climats. Ce n'est toutefois que par leur aspect quantitatif qu'ils constituent une particularité vis-à-vis des autres milieux.

(a) Les précipitations présentent les débits instantanés les plus grands et les intensités calculées sur de longs délais (plusieurs jours) peuvent être importantes. Lorsqu'elle existe, l'alternance de la saison des pluies et d'une période de sécheresse continue est souvent mieux marquée que partout ailleurs.

(b) L'activité biologique dans les sols assure une évolution trop rapide de la matière organique morte si bien que la quantité de matières humiques est dérisoire vis-à-vis de la biomasse qui en est l'origine. Par ailleurs, des termites plus nombreux remuent un volume de terre beaucoup plus grand que celui qui est utilisé par les lombrics en Europe Occidentale.

(c) Quant à la végétation, il faut noter qu'à côté de la forêt dense ou de la savane dont la masse et la structure sont très spéciales, il existe une forêt claire ou des prairies où la protection des sols vis-à-vis du ruissellement est peu différente de celle exercée par d'autres associations végétales.

A l'intérieur du milieu tropical, les facteurs de différenciation restent, comme pour tout autre milieu pour les phénomènes qui nous occupent:

(a) La végétation avec la nature et la densité de chacun des écrans qu'elle interpose entre la pluie et le sol. Dans le Haut Katanga par exemple, se trouvent des échantillons de la forêt dense (muhulu), de la forêt claire (miombo) de la savane herbeuse à *Hyparrhenia*, de la prairie (dembo, bianco) et d'un type de végétation substeppe dans les faciès xériques de ces derniers. Le miombo lui-même présente de nombreux stades de dégradation anthropique.

(b) La nature de ce sol, plus particulièrement quant à la résistance qu'elle oppose à l'érosion par la cohésion des particules dans un horizon supérieur notamment et quant à ses qualités vis-à-vis de l'infiltration et de la percolation.

(c) La pente topographique qui sera fréquemment soit inférieure à 2° soit supérieure à 18°.

A ces trois critères, il faudrait joindre quelquefois un quatrième:

(d) La date à laquelle se produit le feu de brousse: hâtif ou tardif suivant qu'il se produit dans la première ou la seconde moitié de la saison sèche. Le feu de brousse hâtif n'a pu éliminer tous les chaumes ni les feuilles tombées pendant le repos végétatif: il peut donc être reconnu aisément. Malheureusement, la végétation herbacée ne brûle pas toujours au même moment en un lieu déterminé. Il faut donc opter pour le feu hâtif que l'on allume soi-même ou pour des parcelles à feux contrôlés, ce qui suppose une certaine vigilance et un certain degré d'organisation.

Même en éliminant ce dernier critère et en considérant qu'un grand nombre de cas relevant des trois premiers sont incompatibles entre eux, il subsiste un grand nombre de possibilités parmi lesquelles il faudra choisir avec un discernement d'autant plus grand que l'on veut détailler l'analyse en comparant une série de champs d'observation.

Quelles seront les observations à effectuer dans ce champ que nous précisons dans la suite? On peut étudier des éléments statiques, la forme topographique et la matière minérale ou autre qui se trouve sous cette forme (dépôt, résidu d'érosion mécanique, résidu d'altération chimique) et essayer d'en déduire quelle part constitue les résultats de l'action de ruissellement. Une telle démarche présente l'avantage de ne nécessiter qu'un court séjour sur le terrain; elle est à la portée des "pèlerins de la saison sèche", chercheurs en mission pendant la période favorable aux déplacements. Toutefois, très rapidement on se heurte à des problèmes de datation: les colluvions que l'on a détectés grâce à un pourcentage de sable plus élevé, se forment-elles encore actuellement? La bioturbation élimine systématiquement toute stratification et tend à homogénéiser les dépôts voire à les inclure dans d'autres formations. Les caractères subactuels se fondent dans un héritage plus ou moins ancien.

Nous avons tenté antérieurement de démontrer la formation des terrains de couverture surmontant la stone-line par le ruissellement étalant les matériaux des constructions épigées dues aux termites, en nous fondant sur la variation verticale de la composition granulométrique et les structures de migration des colloïdes observées en lame mince. Malgré la complexité des moyens mis en oeuvre, une étude de la dynamique des phénomènes s'est révélée nécessaire pour déterminer notamment le niveau d'origine des argiles qui migrent dans le sol et la proportion de limon superficiel qui est intégré au sol, soit par dépôt et bioturbation, soit par infiltration par des pertuis relativement larges. Lorsque toutes les notations à faire dans le délai le plus court après l'action géomorphologique seront réunies, on les confrontera aux données statiques. Celles-ci auront alors une importance accrue, non seulement comme moyen de vérification, mais encore comme base de quantification, les observations simultanées, comme nous le verrons, montrant dans ce domaine quelque faiblesse.

Quelles sont les observations relatives à l'action morphologique du ruissellement diffus en milieu intertropical et simultanées ou presque de celle-ci? Nous allons tenter d'en faire l'inventaire.

#### OBSERVATIONS PORTANT SUR LES FACTEURS DU RUISSELLEMENT ET DE SON ACTION MORPHOLOGIQUE

(a) Végétation. Rôle: interception des pluies, infiltration. Densité des différentes strates: branches et feuillages des arbres (sur photo prise de bas en haut par temps clair si possible) éventuellement surface occupée par les cimes

lorsque celles-ci laissent des vides entre elles (sur photo aérienne verticale). Notation quant à la conformation: forme et dimension des feuilles (uniquement lorsque celles-ci sont grandes), forme des branches, épaisseur du tapis de mousse.

(b) Aspect superficiel du sol. Rôle: infiltration, résistance à l'érosion. Pourcentage de la surface couverte par la litière, des chablis (dus à l'activité des termites plutôt qu'au vent), une roche meuble nue (recherche de croûte, crevasses, bouche-pores, glaçages), des cailloux, des blocs, la roche en place relativement fraîche, la cuirasse latéritique, les termitières (dimensions, nues ou couvertes de végétation) et autres amas, de trous (terrier ou action des termites).

(c) Profil du sol quant à ses propriétés vis-à-vis de l'eau. Rôle: percolation facilitant l'infiltration.

- litière, épaisseur et nature (destruction lors de pullulement de termites);
- humus, épaisseur de l'horizon humifère, concentration;
- structure du sol (accompagné d'une détermination des minéraux argileux);
- texture du sol (analyse granulométrique complète en laboratoire avec intérêt spécial pour l'argile et pour le silt s'il y a très peu d'argile);
- racine, densité suivant profondeur.
- galeries principalement de termites, densité suivant profondeur;
- couche(s) de porosité différente: cailloutis (stone-line), horizon induré, cuirasse latéritique, roche en place plus ou moins altérée, et plus ou moins perturbée mécaniquement;
- niveau supérieur de la nappe aquifère; celle-ci vient quelquefois en affleurement dans les creux de la topographie (dolines, dembo).

Une partie de ces facteurs varient au cours de l'année et ces variations doivent être observées ou mesurées au cours de visites de fréquence appropriée. La position dans le temps du cycle de variation sera éventuellement définie vis-à-vis du régime des précipitations. D'autre part, chaque fois que la chose est possible, des évaluations quantitatives seront faites. Pour des éléments de faible densité (chablis, bloc, grosse termitière), il faut des mesures dépassant les limites du champ que l'on s'était assigné au départ. Le mieux est de procéder par triangulation successive et d'estimer la distance moyenne entre deux éléments de même nature.

D'autre part, une attention toute spéciale doit être dédiée à la remontée des matières minérales réalisées par les termites jusqu'à niveaux situés parfois à plusieurs mètres au-dessus du sol. Les constructions revêtent les formes les plus diverses: galeries épigées le long de troncs d'arbre, de chaumes, à l'intérieur de souches, de troncs abattus. Les données quantitatives relatives aux masses transportées en dehors des termitières proprement dites sont surprenantes. L'étude de la vitesse avec laquelle s'édifient les termitières doit être faite avec beaucoup de prudence car la destruction faite dans ce but stimule les ouvriers et les estimations risquent d'être exagérées.

Parallèlement à ces observations quantitatives, des facteurs du ruissellement et de son action, une série de mesures relatives à la partie du cycle de l'eau concernée:

- l'intensité des précipitations en fonction du temps (d'après pluviogrammes).
- l'interception des pluies suivant la nature et la densité du couvert végétal, l'intensité des pluies) à l'aide d'un grand nombre de petits pluviomètres convenablement étalonnés.
- le spectre du diamètre des gouttes de pluie (C. Barat, 1957) (mesure délicate qui pourra peut-être être évitée lorsque l'on connaîtra mieux ses relations

- avec l'intensité des précipitations et la nature de la surface d'interception).
- l'écoulement le long des troncs, des tiges ou des feuilles jusqu'au collet pour les plantes basses, à l'aide d'anneaux de caoutchouc ou de papier d'aluminium raccordés à des réservoirs.
- la vitesse d'infiltration à l'aide de deux tubes concentriques, l'infiltration à partie de la zone externe obligeant l'eau du tube interne de pénétrer verticalement. Les débits obtenus sont à comparer aux débits non interceptés.
- le comportement du sol superficiel (là où il est mis à nu) lors de l'humectation (par exemple limites de plasticité et de liquidité d'Atterberg).

#### OBSERVATIONS RELATIVES AU RUISSELLEMENT ET A SON ACTION

Se promener sous l'orage et voir de ses yeux le ruissellement le départ de matière en certains endroits, le cheminement des eaux troubles, le déplacement de différents objets, l'engouffrement des eaux dans une crevasse ou dans le trou laissé par l'élimination totale d'une souche d'arbre par les termites, est une chose indispensable. Toutefois, une foule de phénomènes nous échappent soit que leur échelle soit trop petite, soit que le chercheur ne puisse les observer sans perturber le cours des choses. En outre, l'aspect quantitatif lui échappe presque totalement.

Après les notations faites immédiatement, il faudra d'une part revenir ou attendre pour terminer les constats et d'autre part installer des repères et des collecteurs pour tenter une évaluation.

##### (a) Observations sur le champ

G. Rougerie (1960) a donné une bonne description de la plupart des effets directement observables: disposition particulière des feuilles et des brindilles de la litière avec barrages quelquefois, rebondissement de particules minérales sur la base des végétaux (trace de splash sans ruissellement important) déchaussement des racines avec ou sans escalier accompagné quelquefois d'un plonge pool en miniature (attention, l'existence de radicelles au-dessus du sol peut-être provoqué par le seul départ d'une ou deux feuilles de la litière!), traces de rills (amorces de rigoles essentiellement labiles) et éventuellement d'un cône de déjection qui lui est associé (matière organique, quelques débris grossiers) dépôts dans des flaques ou d'autres pièges à sédiment (une galerie par exemple), dégâts produits aux dépens de constructions de termites.

##### (b) Observations dans le sol

Surveillance de différents types de galerie à l'aide de trous de différentes profondeurs.

##### (c) Observations à l'aide de repères et de collecteurs

Faute de traceurs radioactifs (J. de Ploey 1966), il faut se rabattre sur des moyens plus traditionnels. Les repères que nous avons utilisés sont les suivants:

- (1) des tiges graduées en métal.
- (2) des plaquettes en métal déposées à même le sol.

Il est nécessaire d'exécuter un bon repérage car dans certains cas, elles peuvent disparaître assez rapidement.

Il faut faire le départ entre l'apport direct des termites et le matériel qui a été remanié par le ruissellement.

Par contre, d'autres plaquettes seront portées en relief.

(3) des socles en béton, profondément ancrés dans le sol afin de suivre au théodolite les mouvements du sommet des tiges.

Le but final est de mesurer le tassement consécutif à l'abandon des galeries.

Des essais de placement de collecteurs nous ont montré combien l'utilisation de ceux-ci est délicat. Un certain temps est nécessaire avant que disparaisse la perturbation introduite dans le milieu. Le calcul du volume des réservoirs qui reçoivent les eaux du collecteur doit également faire l'objet d'une enquête préalable.

Ces méthodes ne doivent nécessairement pas être utilisées toutes d'une manière concomitante. Un choix s'impose suivant que l'on aborde ou non l'aspect quantitatif et dans le premier cas suivant le degré de précision auquel on désire arriver, précision d'ailleurs toute relative dans les conditions actuelles d'observation.

Un programme réduit s'appliquera à un champ linéaire perpendiculaire à une zonation établie suivant des critères énumérés plus haut. Les caractères de chaque zone (végétation, pente, nature du sol) doivent être définis avec précision. Lorsque la pente est faible, cette zonation ne suit pas nécessairement la ligne de plus grande pente. Les observations directes à l'aide de repères doivent se poursuivre pendant une année entière au moins.

Un programme moyen inclura des observations complètes sur les facteurs, les repérages au théodolite.

Un programme plus développé, doit comprendre les mesures à l'aide de collecteurs, les paramètres du cycle de l'eau, les mesures relatives aux réactions du sol vis-à-vis de l'eau, une étude détaillée (granulométrie et micro-structure) des modifications introduites dans le sol à la suite de migrations de particules fines et de la bioturbation.

Un programme moins élaboré peut servir d'étude préalable à un programme plus complexe. Ils constituent une approche graduelle d'un problème de géomorphologie quantitative.

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