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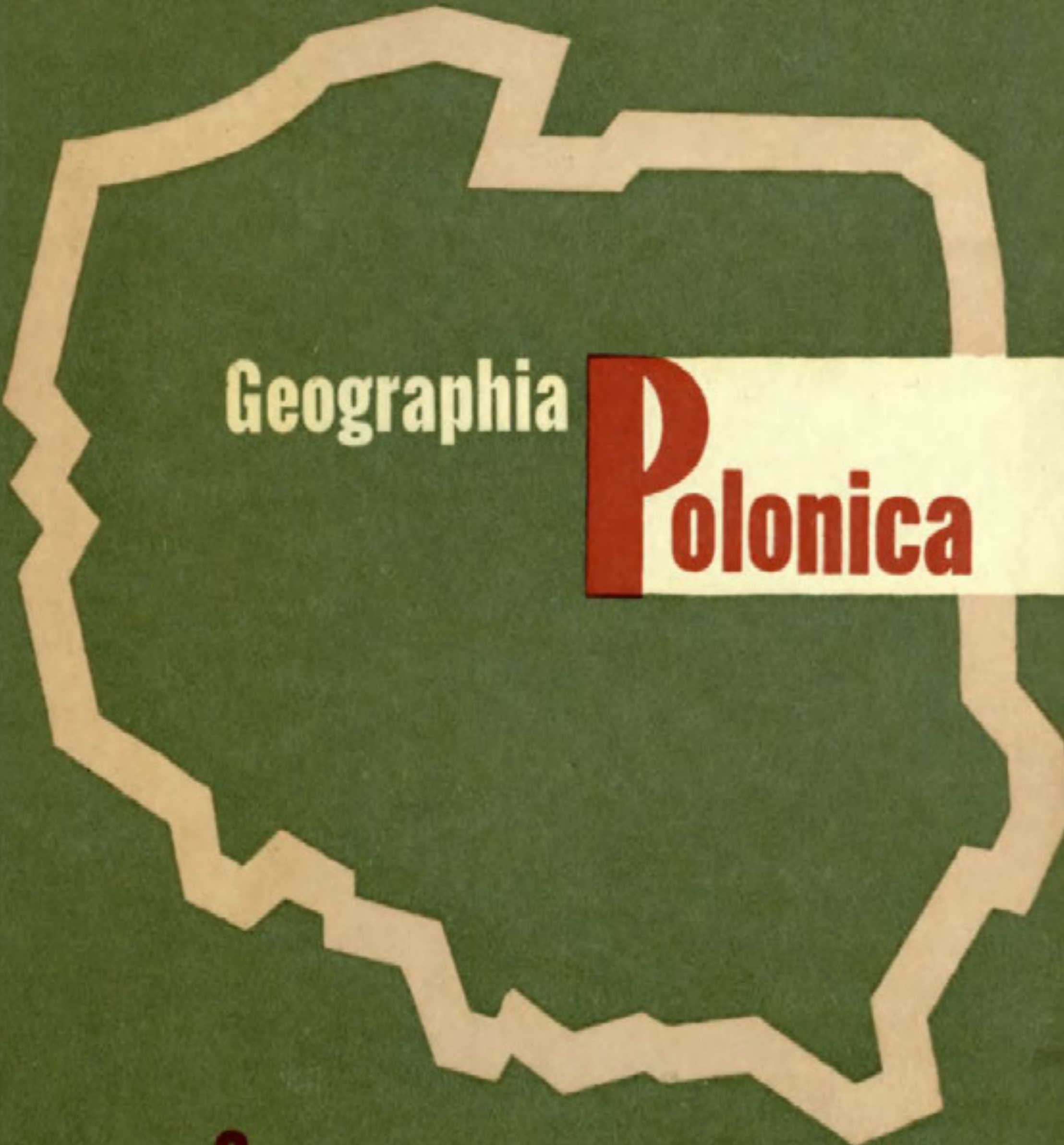

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INTRODUCTION

The present issue of GEOGRAPHIA POLONICA contains some recent studies of Polish geographers on the geomorphology and geology of the Quaternary. It is published in connection with VII Congress of the International Association for Quaternary Research (INQUA) which will be held in August and September 1965 in Denver (Colorado, U.S.A.). The preceding congress of this association, which took place in Poland, 1961, offered an opportunity of presenting the results of Polish researches, among others those of Geographers, in the field of the Quaternary, either in form of reports delivered at the congress, or in form of extensive guide-books of excursions, or in special publications. This collection of papers, in which are represented almost all geographical centres in Poland, gives evidence to the permanent interest of Polish geographers in the geomorphology and geology of the Quaternary. The purpose of this selection is to inform about the widening scope of the problems investigated and the development of research methods. Quaternary researches continue to be the favourite subject of Polish physical geographers and especially geomorphologists. Obviously the present volume of GEOGRAPHIA POLONICA is not the only Polish publication prepared in connection with the INQUA Congress 1965.

GEOGRAPHIA POLONICA 6,1965

CONDITIONS OF OCCURRENCE AND DISTRIBUTION OF KAME LANDSCAPES IN THE PERIBALTICUM WITHIN THE AREA OF THE LAST GLACIATION

WŁADYSŁAW NIEWIAROWSKI

INTRODUCTION

While the term "kame" is known for a long time in glaciological literature, little attention continues to be paid to kames and their role in the glacial relief. Evidence is the fact, that this term is differently interpreted in different countries. In agreement with many other writers, I consider kames to be landforms produced in stagnant and dead ice by the accumulating action of meltwater streams, which deposited the material transported in crevasses, depressions, and in any kind of hollows existing in this ice and between blocks of dead ice, or in the intervals between dead or stagnant ice and slopes of elevations existing at that time. This is the reason, why kames are built of deposits more or less distinctly stratified, such as gravels, sands, silts and, sometimes, even of clays laid down by meltwater flow (fluvioglacial kames); other kame deposits were laid down in lakes existing in the dead ice (limnoglacial kames). Sometimes, the structure of kames appears to be more complex, because of they have been produced partly by accumulation from fluvioglacial waters and, partly, as the result of sedimentation of deposits in ice lakes encountered in stagnant or dead ice.

By the term "stagnant ice" I have in mind large masses of dying ice, usually represented by wide slabs or zones spread out mostly in the foreland of active ice which, however, have not yet lost contact with the active ice; thus, in stagnant ice there may occur insignificant local movements. On the other hand, I consider "dead ice" to be usually minor slabs and blocks of fully dead ice which, separated from the main ice body, have in this manner completely lost contact with it.

Kames produced in ice depressions, hollows, caves and voids, as well as between blocks of dead ice, are mostly developed in the shape of hillocks, hills or — more rarely — plateaus distributed at random. Those kames which took their origin from the accumulation of deposits in ice-lakes, appear usually in the shape of plateaus with a flat top surface and relatively steep slopes. As a rule, they also are distributed fairly erratically.

On the other hand, kames produced in ice crevasses gradually widened by melting ice and flow of meltwaters, show usually the shape of ridges and ramparts; these forms are elongated and oriented parallel with the direction of cracks in the ice. Kame deposits laid down between the slopes of elevations existing at the time that the kame material was deposited, and stagnant or dead ice masses, appear in the form of kame terraces. The morphology and the internal structure of kames depends, among other factors, on whether these deposits were laid down supraglacially, that is, in depressions, crevasses and open hollows not reaching the ice sole, or englacially, meaning, in any opening reaching as far as the base of the ice.

Where kames occur in larger agglomeration and in a given region clearly prevail over other glacial forms, their area of occurrence may be called a kame landscape or kame region. The knowledge of kame landscapes and of their distribution is a very important matter; among other reasons, because the occurrence of wide kame landscapes is one of the foremost indicator of the past existence of extensive expanses of stagnant and dead ice, at the time deglaciation was taking place in that region. Again, the problem of determining the occurrence of large dead ice stretches of glacier ice is closely connected with a problem that continues to be very actual and is very much disputed recently: the character of deglaciation proceeding in areas which during the Pleistocene had been covered by the Scandinavian inland ice. Detailed examination undertaken in a number of countries indicate that, apart from frontal deglaciation, that is, the retreat of the margin of mobile ice due to ice melt prevailing over the supply of new ice masses, there existed many areas that were covered by large masses, at times even by extensive zones, of stagnant and dead ice. These zones were melting from the top downwards all over the ice area, producing a characteristic glacial relief. It seems that in some countries the part, played by stagnant and dead ice in shaping the glacial relief, is not fully appreciated. In virtue of the continued lack of more detailed knowledge, which land-forms and deposits definitely imply the last-named type of deglaciation, we are facing great difficulties in delimiting regions, in which such large masses of stagnant and dead ice have been melted away; therefore also it is a difficult problem to apprehend correctly the role of this type of deglaciation in areas of past glaciations. While kame landscapes are by no means the only symptom of the occurrence of large masses of dead ice in a given region, still, their occurrence signifies to a certain degree the intensity of this phenomenon.

The purpose of the present paper is to analyze, on the basis of descriptions of kame landscapes within the Peribalticum, the relation of kames to other glacial forms and, in this way, to scrutinize which among them did originate in dead ice; further, to present examples, what were the topographical conditions of areas in which kame landscapes developed — a study which also would indicate conditions leading to the formation of dead ice — and, finally, to comment on the distribution of kame landscapes within the Peribalticum, from Denmark to Karelia.

Fig. 1. Subglacial channels and end moraines of the last glaciation in the Polish Lowland (after Maria Klecha)

1a — limits of stages of deglaciation; 1b — minor marginal forms of the inland ice; 2 — subglacial channels; 3 — subglacial channels transformed by rivers;
L: Leszno (Brandenburg) stage; Pz: Poznań (Frankfurt) stage; Pm: Pomeranian Stage

This analysis is intended to embrace merely that part of the Peribalticum, which had been ice-covered during the last Scandinavian glaciation. The distribution of kames in this area has been shown diagrammatically on the attached map (Fig. 1), and this is the first attempt of illustrating the occurrence of kames in an area as extensive as this. Drawing this map was not possible until now: it was only for the VIth INQUA Congress held in Poland in 1961, that for the first time there were shown, by separate symbols, kame landscapes occurring in the whole area of the Valdai glaciation. These data were entered in a map prepared jointly by 16 Soviet authors, with K. K. Markov as editor in chief [34]. For Polish territory, it was the author of the present paper who mapped the distribution of kame landcapes [43]; in this map (Fig. 1) he also took into consideration the most recent research work done in Poland. As to Danish territory, information was presented on the basis of Danish literature and of personal observations made by author who, in 1964, was in position to inspect the glacial relief of great part of Denmark. As to the area of Germany, I mainly took into account P. Woldstedt's map [57]. I am fully aware of the fact that the map attached to my paper is incomplete and that, with the successive appearance of new detailed research data, this map will be getting more complete and more detailed.

In my map I also entered, diagrammatically, the principal ice border lines indicating halts of the inland ice during the main stages and phases of the last glaciation. Even so, parallelization of these data within the Peribalticum is still controversial. It proved impossible to trace the principal ice border lines for a large part of the Russian Plain, because Soviet authors [34, 18] insist that, while a number of ice border lines have been distinguished, up to now their parallelization still remains very controversial. In marking the principal ice border lines I intended to present a general orientation as to those stages and phases of the last glaciation, in which kame regions came into existence.

Relations between kames and other glacial forms

Most frequently, closed depressions (kettles) occur within kame landscapes; these depressions developed after ultimate melting of blocks of dead ice, thus locally confirming the past existence of dead ice. However, kettles also occur in regions where active ice has been receding, thus kettles in themselves are no proof of the past existence of wide zones of stagnat and dead ice; they only indicate, that relatively minor blocks of dead ice have occurred in a given region.

For the most part, kames appear surrounded by a ground moraine. In some instances, they are superimposed on the surface of a ground moraine, that is, they developed after its formation; this shows that the ground moraine originated in the stage of active ice, the kames during the period of formation of dead ice. In some instances, however, only part of a ground moraine has developed in the stage of active ice, while its upper part, called ablation moraine, was melted out during the stagnant period of the ice, simultaneously with the formation of the kames. On and off, small lobes of an ablation moraine rest on the slopes of kames, indicating that areas of ground moraines covered by a thick mantle of an ablation moraine may serve as proof for local melting of major slabs of stagnant and dead ice.

Within some kame regions, hummocky moraines and dead ice moraines occur together with the kames. It is my opinion that, in such instances, land-forms of this type developed at the same time as the kames, in the same stagnant and dead ice, as the mutual interlinking of these forms seems to indicate. Amidst the hummocks, hills and ridges (hummocky moraine) developed in stagnant or dead ice, some at least of these forms can be clearly distinguished: (a) land-forms produced by morainic material melted out, slid down and accumulated in depressions in the surfaces of dead ice [1, 26, 28, 41]; (b) land-forms produced by accumulation of morainic material at bases of dead ice slabs [28]; (c) land-forms produced by squeezing morainic material into glacier crevasses by the weight of the glacier ice [17, 19, 24, 26]; the sediments thus squeezed out may additionally be covered by fluvioglacial material or by an ablation moraine [12, 17, 26, 41]; (d) land-forms containing stratified fluvioglacial deposits, laid down subglacially or englacially in widened crevasses and dead ice hollows, mantled on their surfaces by an ablation moraine [1, 31, 41, 54]. Some authors [7, 11, 26] assign to kames the forms containing stratified fluvioglacial deposits and covered by a mantle of morainic material, but initially developed in stagnant or dead ice. It is a matter of dispute, whether these forms should not rather be assigned to dead ice moraines. As to myself, I am willing to look upon them as transition forms between typical kames and dead ice moraines. Obviously, dead ice moraines, whether they appear together with kames or independently, indicate that they melted out in situ from stagnant or dead ice. It also is commonly assumed, that hummocky moraine with their numerous kettles were produced within the dead ice [40, 48, 49 and others]. Even so, it seems conceivable — as shown by studies made by Hoppe [25], that some areas of hummocky moraine might also have developed with the participation of active ice.

Within drumlin fields, no kames are encountered as a rule. It was only E. Markus [35], who described from Estonia an instance of occurrence of kames on top of drumlins. Thus, here the drumlins must have been produced in the active ice stage, with the kames developing afterwards, that is, in the period of stagnant or dead ice.

No kame landscapes are found in zones of terminal moraines, formed along the margin of active ice. Most frequently, they are met within the backland of terminal moraines, or in the foreland of terminal moraines, of the successive stage or phase of glaciation. Thus, there may occur zones of land-forms produced by active ice and, between them zones of land-forms of stagnant or dead ice. As palpable example of a phenomenon of this kind may serve the Dobrzyń and the Chełmno moraine plateaus [41]. At times, kame regions occur in the outer zone of the maximum range of a given glaciation stage or phase; in such cases, it is the kames and other land-forms of stagnant ice, not the end moraines that delimit the maximum range of the inland ice. Numerous examples of this type have been cited from Poland [4, 5, 6, 52, 54], and from Soviet territory [13]. It also is known from many examples, that kame landscapes occur in near distances from end moraines, or in intervals between such moraines, thus in identical marginal zones. This would mean that, given favourable topographical conditions, active ice existed in some places, stagnant ice in others. Such examples can be cited, among various other areas, from Latvia and from the Leningrad region. In uplands and elevations of these areas, extensive masses of stagnant and dead ice were spread out, in which a kame relief developed, while the lowland still contained active ice tongues or lobes in front of which end moraines developed [26, 53], or large ice-lakes [33]. In some places, great masses of stagnant and dead ice from a preceding ice stream still persisted, while new ice streams, incapable of covering the older ice, surrounded it. In the dead ice, kames and other glacial forms developed and, at the same time, in front of active younger ice tongues new land-forms of active ice were produced. As classical example may serve Denmark where, in the central part of Funen, great masses of dead ice of the NE glacier were spread out, surrounded by the Great Belt and Little Belt glaciers of a little younger age [36, 37, 39, 40, 51, 55], and the Zemaitija Uplands in Lithuania, on which dead ice was still left behind, surrounded by ice lobes of Middle-Lithuanian end moraines then forming [7, 8, 20].

Within the range of larger kame landscapes, eskers frequently occur, undoubtedly developed in the same stagnant and dead ice bodies [3, 4, 41, 42, 50, 56]. This indeed by no means precludes, that eskers, being polygenetic forms [15, 16], may also have developed independently of kames, and in active ice also. Some authors consider eskers to be mutually related with kames. As regard this interrelation, it should be kept in mind that kame hillocks, hills and plateaus, usually distributed at random, as well as kame terraces, neither in morphology and interior structure nor in their origin disclose any resemblance to eskers. On the other hand, kame ridges show a close relationship only to eskers produced in dead ice. In my opinion, kame ridges are some sort of transition form between kames and eskers; at times, it is very difficult indeed to distinguish them, and therefore some writers assign them to eskers, others to kames. Even so, whereas eskers were principally produced in ice tunnels, kame ridges originated exclusively in ice crevasses open upwards. It seems admissible to assign to kame ridges forms called "crevasse fillings" in American literature. Most precisely was defined the difference between crevasse fillings and eskers by R. F. Flint [15] and A. K. Lobeck [30].

At times, kames occur together with glacial channels. On record are descriptions implying, that these channels originated at the same time as the kames [33]; there also exist many data [2, 3, 41, 56] suggesting, that glacial channels developed subglacially in the phase of active ice and that later, in the stagnant and dead ice phase, they were filled in with ice derived from the ice vaults breaking down or from freezing of meltwater (winter ice), while in crevasses or depressions in the stagnant ice kames were formed alongside of, or on top of, preserved channels. After melting of the stagnant ice and of the ice filling the channel, an exhumation of the channels took place. In some of the glacial channels kame terraces occur, developed between channel slope and dead ice.

A further complex problem is the interrelation between kames, and some parts of outwashes. Within the range of outwash plains and valley-like outwashes, no kames are found; in the proximal part of an outwash, however, where on and off numerous gravelly-sandy hillocks with frequent kettles occur, that is, where fluvio-glacial deposits have at times accumulated on the surface of dead ice or between blocks of dead ice, it is sometimes difficult to decide, whether we are facing a kame relief or an outwash relief. Difficulties are also encountered with plateau-like fluvio-glacial plains, developed within slabs of stagnant and dead ice. Some authors are assigning them to an outwash, others to kames. In my opinion, forms of this type should be considered forms of transition between kames and outwashes. I do not believe, however, that it is admissible to assign to kames what is called "transition cones" described by W. E. Boerman [10]; these must have originated outside of the ice margin, without participation of dead ice or of the so-called "pitted outwash plain". The above survey shows that, apart from classical forms, there also exist in nature transition forms representing a transition from certain forms into others.

Interrelation between kame landscapes and topographical conditions

From descriptions of kame landscapes it results, that they must have been produced in divergent topographical conditions. They occur both in ranges of high elevations, where diversified relief and impeded inflow of new ice masses created conditions favourable to the formation of large masses of stagnant and dead ice strongly fissured (like in Zemaitija, Kurzeme, Vidzeme and other uplands), and in older large depressions. In such depressions, kame regions occur most frequently in the zone of maximum ice range, especially in instances, when these depressions are situated at a certain angle to the direction of the ice movement, or when they are inclined in the distal direction, thus obstructing ice inflow and causing the formation of widespread slabs of dead ice. At times, kame landscapes occur on a relatively flat moraine plateau, but in its substratum an older, diversified relief exists which had obstructed the supply of ice [41]. As indicated by Markov [33], kames found in the Leningrad area developed principally on distal sides of older elevations. This shows that, especially in the period of deglaciation, the local relief of the ice base bears upon the dynamics of ice movements and, thus, on its turning into dead ice, and on the probability of kame landscapes developing.

Distribution of kame landscapes in the peribalticum within the area of the last glaciation

From Denmark, P. Smed [51] reported the largest agglomeration of kames to exist on Funen. In view of numerous kames occurring among land-forms produced in the dead ice of the NE glacier situated in central part of Funen, one may speak of "the Vissenbjerg kame region" [45]. Most probably, sandy-gravelly hillocks which K. Milthers [37] describes and believes to have been formed in crevasses of dead parts of the Little Belt glacier at Horne Land, as well as on Björnö and Avernakö islands, are kames also. Minor kame fields were described as early as in 1922 by V. Milthers [38] from Hornsherred, North Zealand. In the region of Farum, Birkeröd, Hilleröd and Lilleröd, among what is called plateau-clay hills described by V. Milthers [38] and S. Hansen [23] there occur what I consider some of them also to be limnoglacial kames; therefore, here also may be spoken of a minor kame landscapes. It is worth pointing out, that in Denmark kames as well as eskers occur mainly on the islands of Funen and Zealand, that is, they were produced during the younger stadials of the last glaciation. Evidently, during these stadials, particularly favourable conditions existed here for the development of extensive dead masses of inland ice.

For a long time the German literature contains numerous notes on the existence of kames, but the absence descriptions of kame landscapes in present-day Germany is surprising. The only description of kame regions, known to me, is a paper by P. Woldstedt [56] discussing kames in the Potsdam region*. This is the more astonishing, because in the neighbouring countries a much larger number of kames have been distinguished. The reason may be, that German scientists rather assigned kames to what they called *Satzendmoränen*.

From the part of Poland embraced by the last glaciation, numerous descriptions of kame landscapes exist. Particularly plentiful are data on kames contained in papers by T. Bartkowski [2—6] and W. Niewiarowski [41—45], mentioned are kame hillocks, hills, ridges, and kame terraces. In the structure of these forms, principally fluvioglacial sands and gravels participate, but also known are many kames built of fine sands and silts, sometimes even containing an admixture of ice-lake clays. Among kame landscapes described so far, most numerous are those found within the area of the Leszno (Brandenburg) and Poznań (Frankfurt) stages. Within the Pomeranian stage and younger phases of the last glaciation, relatively few kame regions were distinguished so far. The reason may be, that here fewer detailed field examinations were made in recent time. Therefore it seems premature to disscuss, whether the area of the Pomeranian stage and of younger phases of glaciation differ distinctly as to the number of kame landscapes from earlier stadials.

* Recently H. J. Franz [59, 60] denied the existence of a kame landscape in this region and he is convinced that these forms described by Woldstedt as kames are in reality mostly push end moraines.

Numerous and extensive are the kame landscapes described from the Lithuanian and Latvian Soviet Socialist Republics [7, 8, 11, 12, 26, 27, 53]. They occur principally in upland regions, such as the Baltic Uplands and the Zemaitija Uplands in Lithuania, and the Kurzeme Uplands, the Vidzeme Uplands and the Selia kame region in Latvia. Particularly important is the part played by kames in the glacial relief of Latvia. Several Latvian authors [14, 26, 27, 53] believe that in Latvia deglaciation took place mainly by melting of stagnant and dead ice, in which characteristic complexes of land-forms developed; among them, an important role to have been played by kames. The Latvian authors mentioned assume, that participation of active ice in the formation of the youngest glacial relief of Latvia was rather insignificant.

In the territory of Estonia, kame landscapes are less numerous and extensive that in Latvia; however, many minor kame fields are also known here [22, 35, 46]. They occur principally within the range of Estonian elevations [46] and, among other regions, in the Haanja and the Otepää Uplands [22].

Probably the largest kame landscapes within the area of the last Scandinavian glaciation are found on the Karelian Isthmus. The southern part of this kame region has been described by S. A. Yakovlev [58] and K. K. Markov [33]. In Markov's opinion, the kames of the Leningrad area developed in large ice-lakes existing within the dead ice body. Apart from the kame landscapes of the Karelian Isthmus, there also should be mentioned the Koltushi kame region and the Shapki and Kirsino kame region [47]. Further, in many other regions of the NW part of the Russian Plain, within the range of the Valdai glaciation, numerous kame landscapes occur [9, 13, 21, 32]. It can even be shown, that here participation of kames and other land-forms, developed in stagnant and dead ice, is probably larger than it is in Poland, Germany or Denmark.

An analysis of the map (Fig. 1) discloses, that kame landscapes occur within diverse stages and phases of the last glaciation. In Poland and Germany they are principally found within the area of the Brandenburg and Frankfurt stages. It must be added, however, that land-forms produced during these stages are here developed best (occupying the largest areas), whereas in other areas marginal forms of the Pomeranian stage are found relatively near the limit of the maximum extent of the last glaciation. This explains, why in other regions kame landscapes occur principally in the glaciation zone of the Pomeranian stage and of younger phases of glaciation (this refers to Denmark, Lithuania, Latvia, Estonia, and to the NW part of the Russian Plain). Because we still lack parallelization of individual stages and phases of the last glaciation taking in the entire glaciation area, it is impossible now to state, whether there appears a differentiation in the occurrence of kame landscapes in areas affected by the successive stages of shrinkage of the last Scandinavian inland ice. In other words, whether during the retreat of the inland ice certain periods occurred, during which — under favourable climatic conditions — stagnation of large zones of inland ice took place, leading to the formation of kame landscapes simultaneously all over the Peribalticum area. Data on hand seem rather to indicate, that decay of the various parts or zones of inland ice took place during deglaciation of the regions in question; ice decay was mostly of local character, depending on local conditions of ice dynamics which in turn — apart from general climatic conditions — were influenced by the relief of the ice base.

In the Peribalticum, regional differences in the occurrence of kame landscapes may be observed. In the glacial relief of Denmark, Northern Germany and Poland push end moraines and other land-forms produced by active ice have played a greater part than forms produced in stagnant and dead ice. In the Russian Plain, push end moraines are rare [1, 29], while land-forms produced in stagnant and dead ice, among them kames, are more frequent and occupy wider spaces. This must have been caused by differences in the dynamics of the last Scandinavian inland ice, that is, by its greater activeness in the west than in the east and, therefore, the more frequent occurrence of stagnation of large masses of inland ice in the east. The western part of the Peribalticum lies in a line with the Baltic Basin, a basin which facilitated the down flow of ice and through which the most active ice streams moved. A. Asejew [1] ascribes this differentation in inland ice dynamics, particularly during its period of shrinkage, to climatic differences. In his opinion, the western part of the Peribalticum was more subject to oceanic and cyclonic influences, while farer east the effect of a continental and anticyclonic climate predominated. These climatic differences must have had their effect on alimentation and ice balance of the Scandinavian inland ice, revealed in differentiation in the dynamics of the ice cover. The problems discussed require further research.

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SOME NEW PROBLEMS CONCERNING SUBGLACIAL CHANNELS

RAJMUND GALON

As it is known, in the time of deglaciation every longer halt of inland ice, marked by marginal forms produced an intense erosional activity of subglacial waters along its edge. This activity resulted in the formation of a corresponding system of subglacial channels which constitute one of the most prominent features of the young glacial landscape. There are many works on this subject written by Danish, German [20], Polish and other authors. A synthetic description of the subglacial channels was presented by S. Majdanowski [11] who took into consideration the whole Peribaltic area from Copenhagen to Leningrad and established the boundary of the last glaciation on the basis of the distribution of these forms.

In accordance with their genesis subglacial channels are generally perpendicular to the ranges of end moraines. It was stated, among others, by S. Majdanowski [10]. This principle may serve in the reconstruction of local marginal zones of inland ice as well as of the direction of its movement, where end moraines or marginal forms are not clearly developed. It was applied [12] to the reconstruction of marginal zones of the retreating inland ice in Chełmno Moraine Plateau which is in the foreland of the Pomeranian stage of the last glaciation.

Subglacial channels are one of the objects of researches carried out by the Department of Physical Geography, Toruń University, and the Department of Geomorphology and Hydrography of the Polish Lowland, Institute of Geography of the Polish Academy of Sciences.

The point of departure of the present considerations is a map worked out by Maria Klecha in her diploma thesis [7], illustrating the question of the relation of subglacial channels to the course of marginal forms in the area of the last glaciation in Poland (Fig. 1). The course of marginal zones was presented according to the available studies and syntheses [5], but the picture of channels is based on the existing geomorphological maps, special monographs and an analysis of detailed topographic maps. The picture of the distribution of channels contains, no doubt, several errors, the more so that it was not always possible to separate subglacial channels from subaerial valleys of meltwater the course of which is rather independent of the direction of the advancing inland ice. On the map a distinction was made between subglacial channels and those transformed by rivers and partly changed into river valleys. This transformation does not affect the problem posed.

As it follows from figure 1, subglacial channels in Poland run SSE, S, and SSW, so their range is rather limited. This was stated already by S. Pawłowski [15]. The channels are connected with only some halts of inland ice and they dissect at different angles younger, even more important ranges of end moraines. Thus it appears that the course of advancing ice was rather permanent. But the oscillations of small ice lobes, adapted to local conditions, differed in many places from the main direction of the transgression of inland ice. The map in question shows (Fig. 1) that certain channel ranges, independent in their course of differently situated numerous halt stages of inland ice, stretch from Gdańsk Bay to the Noteć *pradolina* (ice marginal streamway) along a distance of about 200 km. This testifies to the persistence of certain directions of subglacial channels connected perhaps with non-glacial processes (tectonic or predisposition of the quaternary bedrock). The parallel course of the channels range in question to the lower Vistula valley suggests that the origin of the gap in this valley may be found in the existence of a similar channel range in place of the present lower Vistula valley.

A crossing of channels may also be observed, e.g. within the Pomeranian stage. We can observe here two successive phases of the formation of subglacial channels overlaying one another. The older phase is represented by channels stretching independently of the direction of younger end moraines, especially of marginal forms of the Pomeranian stage, i.e. of the last appearance of glacial processes in this area. Those channels follow the meridian direction while the younger ones run NW-SE, that is perpendicularly to the main moraines of the Pomeranian stage. In this way the above two channels systems represent the growing (transgressive) and shrinking (regressive) glacial phase. We know but little about the course of the growing ice sheet so far. Perhaps there is a possibility of learning more about this process.

On the basis of his investigations of Jutland N.V. Ussing [18] was the first to find out that channels often formed radially in the vicinity of the inland ice edge, the meltwater flowing through so-called ice-gates. This question was later developed by P. Woldstedt [19]. More often, however, as it follows from the map (Fig. 1), the channels run parallel to one another, the distances between them being different. Sometimes parallel channels occur in agglomerations, e.g., according to the investigations of Z. Churski [2], within the region of the Wdzydze Lake in Pomerania (Fig. 2). Some of these channels are deeper, others are shallow and of initial character (cf. also the distribution of channels SE of Poznań, Fig. 3). In many places it was stated that channels are accompanied by a kind of terraces, e.g. along Ruziec valley (Dobrzyń Moraine Plateau) or at Byszewo, NW of Bydgoszcz (Fig. 4). These terraces, which have no permanent inclination and very diversified, change in places into lateral channels of initial character, as in case of Byszewo (Fig. 5). It may be assumed that central channels had formed first and when they were filled with ice

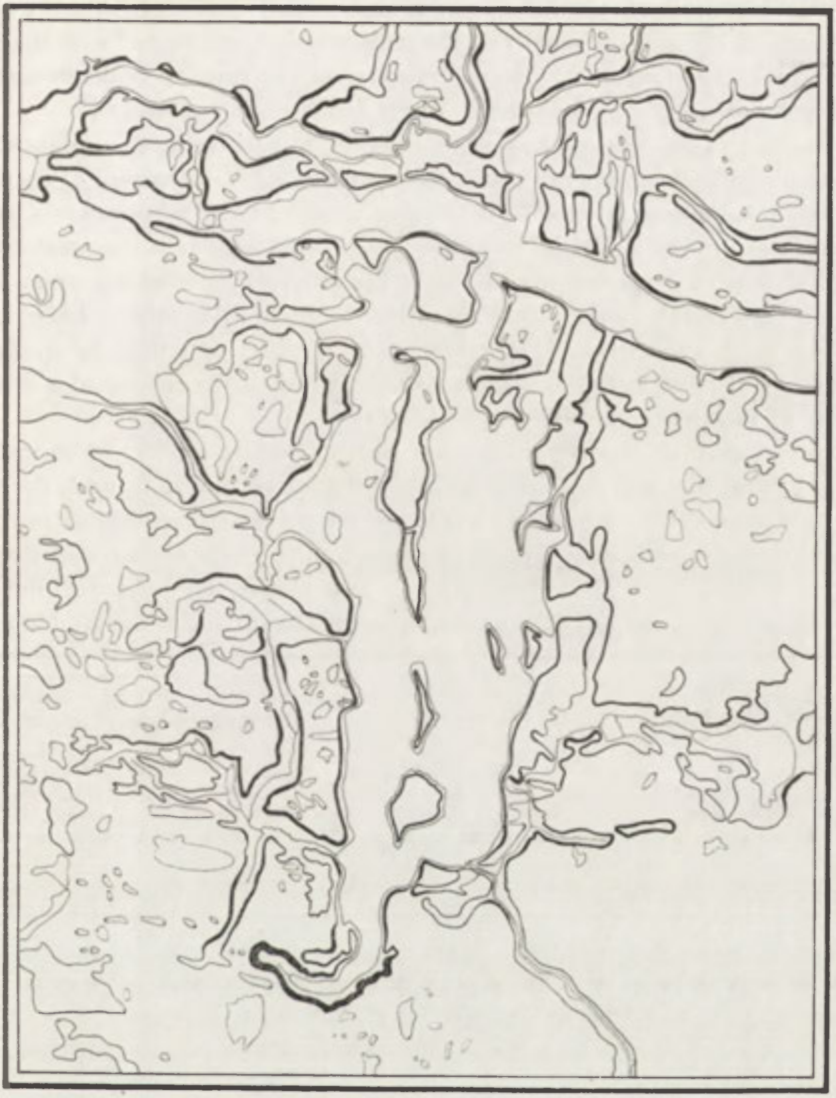


Fig. 2. Subglacial channels of the Wdzydze Lake (after Z. Churski)

the erosional activity of subglacial water moved towards the sides. Thus the question of the varying quantity of subglacial channels is strictly connected with the process of continual formation of new ways under ice as the existing channels were filled with blocks of dead ice or so-called winter ice (frozen meltwater).

In case when two subglacial channels are close to each other the fragment of the moraine plateau which separates them is lowered until it is deprived of its moraine cover and takes apparently the form of esker deposited at the bottom of a spacious channel (Fig. 6). The similarity of cutted off ridge to a esker is the greater that its geological structure reminds of that of esker (Fig. 6). Several examples of such pseudo-eskers can be mentioned, e.g. near Miszewo, Kashubian Lake District, described by J. Sylwestrzak [17], esker near Oslonino, described by B. Zaborski [21], or Turtul esker in Suwalki Lake District, which is the subject of a discussion between S. Pietkiewicz and C. Pachucki [14].

In special deglaciation conditions, viz. the stagnation of ice, when instead of marginal forms like end moraines, deposits of kame type were formed, the activity of meltwater occurred on two levels (Fig. 7). Above the channels filled with winter ice developed the landscape of dead ice, conditioned by the existence of stagnating ice. The disappearance of ice took place in two phases. First, i.e. still during the deglaciation period, slabs of dead ice which lay on the surface of the ground moraine

Fig. 3. Subglacial channels and esker in the vicinity of Poznań (fragment of a morphological map elaborated by E. Tomaszewski)

Fig. 4. Morphological sketch of a section of the Byszewo subglacial channel (elaborated by R. Rybacki)

melted away uncovering kames, eskers and other land forms originated among ice blocks. Then subglacial channels emerged from under the winter ice filling. Both forms occur in varying arrangements. Thus W. Niewiarowski [12] writes about the melting out of a subglacial channel partly situated under kame which resulted in a subsidence of the kame deposit (Fig. 8). Z. Churska [1] and W. Okołowicz [13] describe the occurrence of a channel along the erosional edge dividing the former outwash level from a lower lying outwash surface. The ridge cut off from the higher lying outwash looks like esker.

The problem of the period when dead ice filling the subglacial channels melted away is still open. We can add a few facts and remarks. From the point of view of the morphological age of forms it is interesting to note the occurrence of channels in the valleys, sometimes on very low terraces. Their melting away certainly took place after the formation of a given terrace. Cases are known when channels dissected in a ground moraine cross with outwash and river valleys, e.g. Brda river valley [3] or Drwęca river valley [9] where the channel also dissects low river terraces. It may be considered as a kind of epigenesis [4].

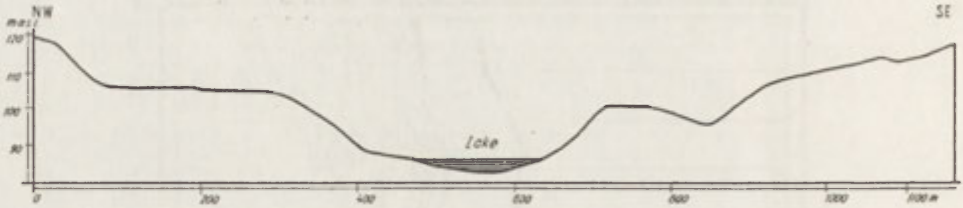


Fig. 5. Cross-section of the Byszewo subglacial channel (after R. Rybacki)

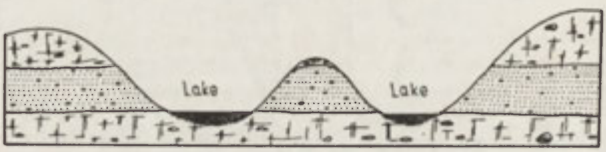
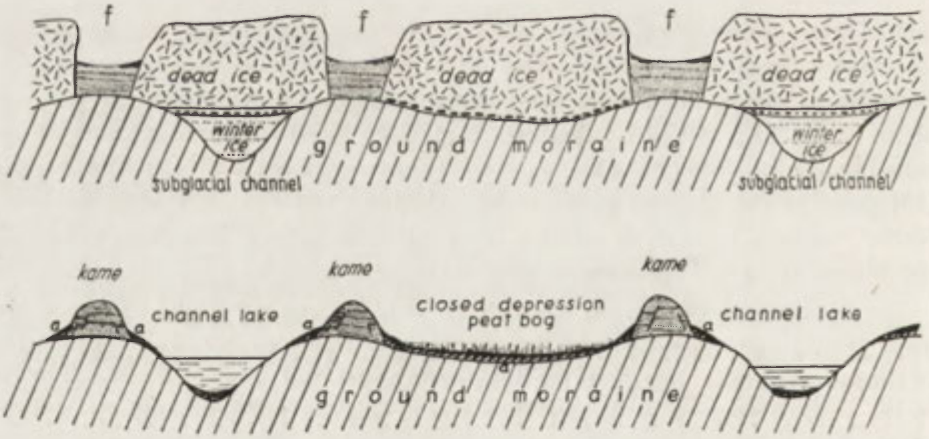


Fig. 6. Erosional ridge between two subglacial channels forming a pseudo-esker

A



B

Fig. 7. Preservation of subglacial channels by ice and formation of kames

A — dead ice slabs covering ground moraine with subglacial channels filled with winter ice; f — filling of ice crevasses with fluvioglacial sands and ablation deposits, B — appearing of subglacial channels (lakes) and dead ice landscape; a — ablation deposits

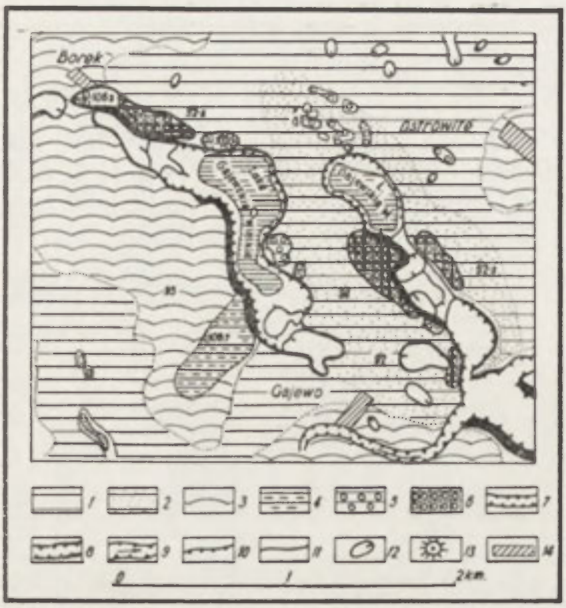


Fig. 8. Subglacial channels and kames in the vicinity of Ostrowite (after W. Niewiarowski)

1 — boulder-clay flat ground moraine; 2 — flat ground moraine covered by sands and gravels; 3 — undulant ground moraine; 4 — dead ice moraine; 5 — kame hummocks; 6 — kame ridges; 7 — subglacial channels; 8 — steps and hollows in subglacial channels; 9 — meltwater valleys; 10 — Pleistocene erosive edges; 11 — Holocene erosive edges; 12 — kettles; 13 — Holocene gorges and other small erosional valleys; 14 — settlements

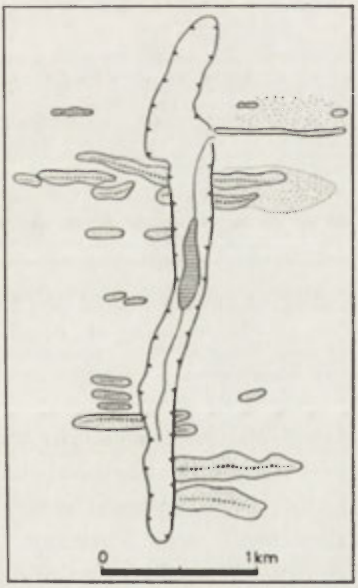


Fig. 9. Subglacial channel near Fletnewo (Grudziądz valley basin) with dissected dune ridges (after L. Roszkówna)

Fig. 10. Fragment of a morphological map of the area of Riss Glaciation (S of Masurian Lake District). Elaborated by M. Bogacki

The appearance of subglacial channels began in warmer late-glacial phases, especially in Allerod. A general melting of subglacial channels took place beginning from Preboreal time and lasted, in exceptional cases, until post-glacial climatic optimum. In Grudziądz valley basin near Fletnowo, a melting channel which appeared on the lower terrace caused the dissection of dunes (Fig. 9). Lake deposits which fill out this channel, investigated by B. Noryśkiewicz (pollen analysis), show that the deposition of gyttia began in Preboreal time, while higher lying peat was deposited during the Atlantic period. Thus the dunes must have been formed in Dryas time.

Finally the role of melting subglacial channels must be stressed in the rejuvenation of the glacial landscape partly changed by later periglacial processes. The young glacial landscape was formed in two phases: the first phase, during which developed marginal forms, mainly end moraines, outwash, kames, eskers, marginal streamways; the second phase during which emerged subglacial channels and similar forms. These phases were separated by a period of activity of periglacial processes which caused a transformation and above all a denudative softening of forms of the first phase. In this way subglacial channels and related forms laid bare already after the above periglacial period constitute the least transformed part of the glacial landscape and channel lakes owing to their considerable depth are of longest duration [6]. Many channels assumed the function of river valleys (on Fig. 1 interrupted lines) without essentially changing their original course. Actually the river pattern shows divergences with the general sloping of the region and of the course of main rivers which is characteristic of the young glacial landscape.

But in the foreland of the last glaciation, in the area of the old glacial landscape subglacial channels which underwent complete fluvial transformation are already very indistinct and not easy to reconstruct (Fig. 10). The lack of typical channels still remains the best criterion of establishing the morphological boundary of the last glaciation.

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DIFFERENTIAL BALTIC ICE-STREAM ACTIVITY ON THE EXAMPLE OF THE ODRA LOBE

Stefan Kozarski

During the Pomeranian stage of the Baltic (Würm) glaciation on the Lower Odra and Lower Vistula there were two great lobes as side branches of the chief Baltic ice-stream. For the first time both lobes — based on the distribution of end moraines of that stage — have been determined by K. Keilhack [16, 17]. The extent of the Odra lobe, according to K. Keilhack, reached the surroundings of Bytów, where the junction with the Vistula lobe existed. The research work done lately and recent views [3, 4, 7, 21, 27, 28, 34] have brought a more detailed determination of the extent of both lobes. The Odra lobe, for instance, in the north-east reached only the vicinity of Ińsko, i.e. where the ending of the great end moraine arch occurs coming from Feldberg (GDR) and passing the Odra valley near Stara Rudnica. On the elevation of central Pomerania there existed separate ice lobes well marked by smaller than in the Odra lobe, but nevertheless distinct end moraine arches. To such lobes belong the Bytów lobe [7] and the better developed Parsęta lobe.

The end moraines of the eastern part of the Odra lobe developed into hills and hillocks of a relative altitude 5-40 m, and are chiefly built of boulder and sand material, deposited rather chaotically (Fig. 1). In one place only, at Objezierze near Choszczno, glacitectonic disturbances were encountered. Similar features of internal structure are encountered in the end moraines of the western part of the Odra lobe, though, according to H. Liedtke [24], push moraines seem to be more frequent there.

After the phase of the maximum extent of the Pomeranian stage the margin of the Odra lobe found a new line, marked out by the end moraine belt running in the area from Fürstenwerder through Gerswalde and Angermünde to the Odra valley south-west of Schwedt. On the Polish side of the Odra, German geologists [29, 31, 32, 33] marked this belt up to the surroundings of Moryń, where, according to them it vanished in the hinterland of the chief end moraine of the Pomeranian stage. P. Woldstedt called this belt *Angermünder Staffel* or *Raduhner Endmoräne*. The investigations of the present writer have shown that this belt of end moraines, named by him Chojna phase, cannot have its end near Moryń. It approaches to within 2 km of the chief end moraine, and then it runs further in the form of distinct hills and hillocks (Fig. 2) — overlooked by former investigators — arranged in arches up to Barlinek, and there it is connected with the chief end moraine (Fig. 3). The Tetyń end moraine (*Beyersdorfer Endmoräne*) also belongs to this phase. Comparing the end moraines of the Chojna phase on both sides of the Odra according to their internal structure we see the same phenomenon as in the chief end moraine belt of the Pomeranian stage. In the west the glacitectonic disturbances occur much more often than in the east, where the thrust features are only encountered sporadically.

The facts mentioned above concerning the internal structure of end moraines allow us to conclude, that during the maximum extent of the Pomeranian stage and during the Chojna phase there were differences in the activity of the Odra lobe margin; it was more active in the west than in the east.

The presence of the long end moraine belt in the hinterland of the chief end moraine, with which it is joined near Barlinek, shows that, while the west and south parts of the Odra lobe retreated about 2—10 km, the east part did not change its previous position. There is some reason, therefore, to speak of the bipartition of the Pomeranian stage on the Lower Odra, that appears not only in end moraine belts, but in two outwash plain levels and also in two terraces in the Noteć-Warta *pradolina*. We can assume that this phenomenon was not confined to the Lower Odra region only, because in Mecklenburg, going northwest of Fürstenwerder, the end moraine belts in the hinterland of end moraines of the maximum extent

Fig. 1. Moryń. Internal structure of end moraine of the Pomeranian stage

Fig. 2. Brwice. End moraine hillocks of the Chojna phase

of Pomeranian stage are rather common, i.e. south of Neubrandenburg [13, 15], south of Teterow [26], as well as in the region of Krakow am See [30], which probably form a continuation of end moraines of Chojna phase. The present writer's investigations strengthen the views of H. G. Ost [25], K. Beurlen [1], H. Liedtke [24] and R. Galon [5] of an earlier recession of the inland ice in the west than in the east. It is only proved to be true in relation to the initial period of the Odra lobe recession. However, contrary to H. Liedtke, a some what larger range of recession in the east should be agreed to, i.e. it must be supported by the new part of the line of extent of the Chojna phase, found by the present writer. This evidence shows that it is difficult to agree with K. Beurlen [1] who accepted too a great recession for the initial period of disappearance of the Odra lobe and of its simultaneous shrinkage from the east, south and west which has not been confirmed by facts. The same concerns the views of K. v. Bülow [2] based on K. Richter's conception.

Taking as the basis of comparison data from papers of R. Galon [6], R. Galon and L. Roszkówna [7], W. Schulz [30], it must be stated that end moraines of the phase of maximum extent of the Pomeranian stage to the east and west of the Odra lobe show the presence of glacitectonic disturbances much more often. In the area of the Odra lobe considered the type of *Satzendmorane* prevails. The reason, probably, is that the margin of the Odra lobe was less active than the inland ice of the Pomeranian stage in other regions. The smaller activity of the Odra lobe

Fig. 3. End moraines of the Odra lobe. After Galon and Roszkowna, Hurtig, Janke, Karczewski, Kliewe, Kozarski.

1 — Pomeranian stage, 2 — Chojna phase, 3 — late glacial

was caused by the fact stressed by some authors [1, 9, 12, 13, 23] that it formed only a side, and thus worse alimented, branch of the Baltic ice-stream, whose chief part was directed to the Danish straits, entering with its margin west Mecklenburg. The great passivity characterizes the Odra lobe, especially in its eastern part, and also in the later period, after the phase of maximum extent, which is proved by the rare occurrence [4, 8] of end moraine belts in the interior area, which it formerly occupied. Its margin became more active only when it retreated nearer the main ice-stream on the Baltic shore, which is shown by the numerous belts of push moraines in north-east Mecklenburg and on Uznam (Usedom) and Rügen islands [13, 14, 18, 19, 20]. The high push moraines near Szczecin are not discussed in the present paper, because of their uncertain chronological position [22, 34] on the other hand in the interior of the Vistula lobe many of push moraine belts were found [6, 8, 28]. This phenomenon shows the greater activity of the Vistula lobe margin, which can be explained by the fact that it lay in the extension of the axis of the chief Baltic ice-stream, which flowed from the centre of the last inland ice situated in the Gulf of Bothnia [10, 11] and thus, during the Pomeranian stage and its decline, it could be better nourished than the Odra lobe.

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PATTERN AND DEVELOPMENT OF ICE MARGINAL STREAMWAYS OF THE KASHUBIAN COAST

Bolesław Augustowski

The *pradolinas* (ice marginal streamways) are a very dominant relief feature of the configuration of Middle European Lowlands. Numerous theories, characterized by generalized hypotheses were expounded to explain the genesis of those forms. All the large systems of the *pradolinas* spread over the nearer or the more remote foreland of the zone arrangement of the glacial sculpture forms — which is correlated with the definite substages of the Pleistocene glaciation. The development of one of the systems, namely: the Warta-Noteć *pradolina* was presented in a convincing manner by R. Galon [6].

There is a *pradolina* system in the north-eastern part of Pomerania, that is to say: on the Kashubian Coast, which differs from the schemes accepted for the remaining large systems; it differs from them in arrangement and dimensions, in conditions of its development, and also in many morphological characteristics. The *pradolinas* there, form a very complicated pattern. They run in two directions: meridionally and latitudinally parallel.

The meridional direction of the *pradolina* does not make an uniform or even a complete entity of form, but consists of several segments, mostly separated from each other which are the segments that had remained from the vaster system which was destroyed by sea abrasion and by successive transgressions of the Baltic Sea. Only the meanders curving westward remained; the easterly ones found themselves under the waters of the Gulf of Gdańsk and the open sea. The north-south direction of the *pradolinas* consists of: the Oliwa-Sopot Seaside Terrace, the Redlowo Depression, the Kashubian Meander, the *pradolina* of Plutnica and its extension, but running in the westerly direction, the Plain of the Seaside Marshes. The Kashubian *pradolina* is the name given to the entire system.

The latitudinal system of the *pradolina* is uniform and continuous in form. It is represented by Reda-Łeba *pradolina*. In the east it has connections with the middle segment of the Kashubian *pradolina*, or so called Kashubian Meander, in the west it runs into the Baltic Sea. Fig. 1. illustrates the described above system of the *pradolinas*. Those streamways cut deeply (50-100 m) into the Pleistocene plateau in many planes, called here *kępa* (isolated moraine plateau). The *pradolinas*

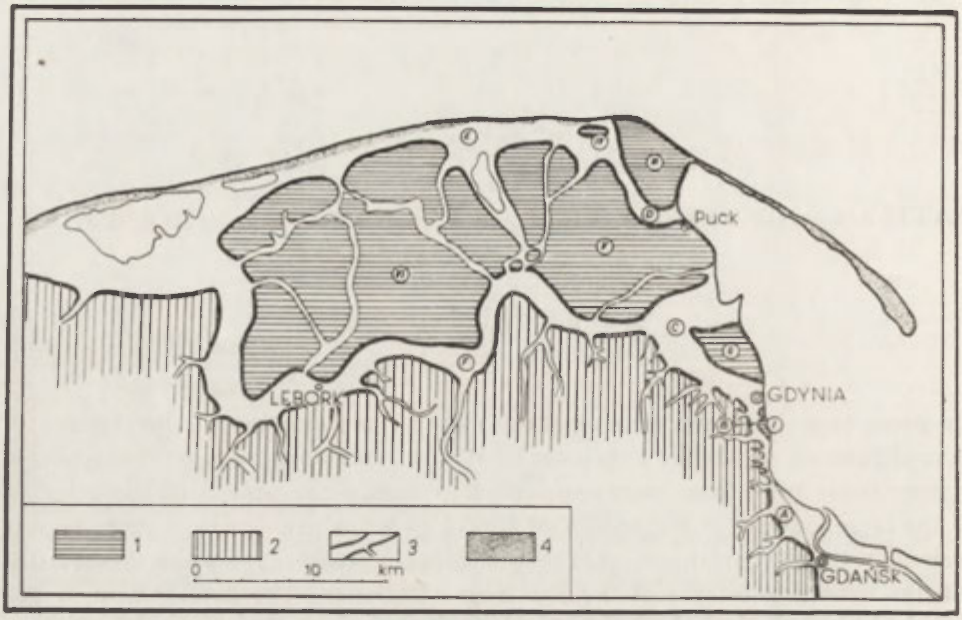

Fig. 1. The *pradolinas* and the isolated moraine plateaux of the Kashubian Coast

1 — the isolated moraine plateau of the Kashubian Coast, 2 — plateau of the Kashubian Lake District, 3 — streamways, valleys, gullies, 4 — dunes and sand spits
I — the Orłowo Isolated Moraine Plateau, II — the Oksywie Isolated Moraine Plateau, III — the Swarzewo Isolated Moraine Plateau, IV — the Ostrowo Isolated Moraine Plateau, V — the Puck Pleistocene plateau, VI — the Lębork Pleistocene plateau, A — the Oliwa-Sopot "coastal terrace" B — the Redlowo Depression, C — Kashubian Meander, D — the Plutnica *pradolina*, E — the litoral peats plain, F — the Reda-Łeba *pradolina*

and the mounds make two basic complexes of contrasting forms on the Kashubian Coast. They stamp the whole area in an individual way, and form a distinguishing characteristic, so that they form the basis for the separation of a geomorphological region.

The changeability of the height of the valley floor of each individual member of the *pradolina* is very characteristic and also important for the reconstruction of the development of the system. The Oliwa-Sopot Seaside Terrace is located at a level of 10-20 m above the sea. It consists of an erosion surface marked by moraine pavement and covered with alluvial cones. The width of the formation reaches 3 km. The floor of the Redlowo Depression is located at 40 m and is suspended from the south as well as from the north. The erosion floor is carved out in fluvio-glacial materials and reaches 800 m in width. The Kashubian Meander forms a beautifuly developped arc cut by the seashore. The inlet of that form from the Gulf of Gdańsk is located in the vicinity of Gdynia, and the outlet into the Puck Bay — between the villages Machelinki and Oslonino. The erosion floor of that *pradolina*, tested in numerous borings, is located 2 m below the sea level. At the outlets of the lateral valleys it is covered with widespread alluvial cones, sediments and peats. They form a cover well over ten meters deep. Thus the Pleistocene erosion floor is suitable for mining. In relation to the scanty demensions of the Redłowo Depression, the Kashubian Meander reaches 22 km in length and from 2·2 to 5·5 km in width.

Even the comparison of those three *pradolina* segments allows us to establish two basic conclusions. The first one concerns the establishment of three unquestionable levels, at which the runoff of the *pradolina* waters had been accumulating. They are: the 40-m level of the Redłowo Depression, the about 20-m on the Oliwa-Sopot Seashore Terrace, and the 2-m erosion floor level of the Kashubian Meander. The second conclusion, deduced from the varying widths of the discussed forms, refers to the amount of the water masses carving the high as well as the low *pradolina* levels. Here at the initial periods of the formation of the system of *pradolina* and marked by high terraces and valley floors found up to the 40-m level inclusively, the amounts of waters had been considerably smaller than in the periods of the development of the *pradolinas* at low levels.

The *pradolina* of Plutnica, located on the extension of the Kashubian Meander across the Puck Bay, forms further segment of the north-south *pradolina* alignement. Then it passes in the westward direction into the Plain of the Seaside Marshes. The actual floor of those forms is correlated with the floor of the Kashubian Meander, not only in its height but also by the peat layer coverage of the same age, dating its beginnings from the end of the Boreal period and the early Atlantic period. The Pleistocene floor descends to 4 and 6-m below the sea level in the Plain of the Seaside Marshes and so it may be correlated with the Kashubian Meander erosion floor at 2-m below the sea level and the same period of formation and development may be accepted for both of them.

The morphological characteristics of the Reda-Łeba *pradolina* are contrary to the system described above. It has a continuous form, about 90 km long. Its width increases westwards: at the inlet, near Reda, it is 1·2 km, 2·5 km in Łężyce, 3·5 km near Cecenowo and 5·5 km at the outlet. The increase in width is the result of the increase in volume of the masses of waters draining into the *pradolina* through numerous lateral streamways. Very numerous terrace levels may be found in the Reda-Łeba *pradolina*. The high levels, above 50 and 60 m, appearing mostly at the outlets of the lateral streamways, were preserved least. The 40-m level is represented much more distinctly. In the neighborhood of Boże Pole and Strzebielino it forms a floor of the *pradolina* covered by an extensive alluvial cone of the Łeba river, which ends its Pleistocene plateau course here. This fragment of the valley floor forms the *pradolina* watershed because Łeba river flows down the ancient streamway westward while the Reda flowing out of the cone (hidden bifurcation) heads eastward. The *pradolina's* floor, carved by waters flowing down in two opposite directions begins at the watershed at the 40-m level, and descends both westwards and eastwards. Two lower terrace levels were formed as a result of adjustments to a changing erosion basis. They both show a close correlation with the lower levels of the Kashubian *pradolina* that is with the average 10-20 m level and also with the low 2-m level, covered with alluvia to the height of the actual floor. The alluvia appear only in the lower segments of the Łeba and Reda river valleys. Fig. 2 illustrates the arrangement of the terrace system in the Łeba-Reda *pradolina*. It demonstrates

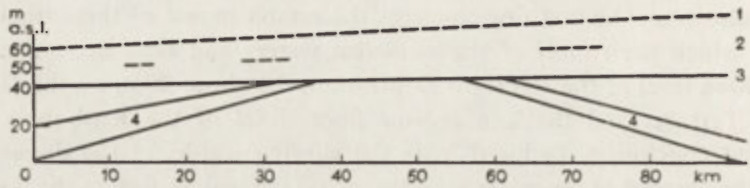

Fig. 2. The long profile of terrace systems of Reda-Łeba *pradolina*

1 and 2 — high terraces levels, 3 — height of the watershed, 4 — directions of the downflow of waters

that the flow of the meltwaters down that streamway had been taking place at levels up to 40 m in height (the height of the watershed divide), and the further development of that form is the work of Łeba and Reda rivers, which carved two lower terrace levels as a result of the changes of the erosion basis.

The 40 m height of the Reda-Łeba *pradolina* and the relatively well preserved terrace system at this level, obliges us to search for an older correlation of this form with the Redlowo Depression, the floor of which is also located at the 40-m level. This relation is also confirmed by the narrowness of the Reda-Łeba *pradolina* at its inlet (about 1200 m) and the Redlowo Depression (800 m).

Many different conditions contributed to the peculiar arrangement and development of the *pradolina* of the Kashubian Coast, but a very important one is the configuration of the surface, strongly dependent on the configuration of the substratum and the course of the front of the inland ice. The tilt of the initial surface left by the Scandinavian ice sheet from the Pomeranian stage as it retreated northward was of decisive importance in the formation of the system of the water runoff. The tilt of the land surface, as it was proven by R. Galon [7] and Z. Pazdro [11], was closely correlated with the geological structure of the older substratum conforming in general to the configuration of the Cretaceous and Tertiary surfaces. The depression of the lower Vistula river, lower Łeba or the Żarnowiec Lake and the Wejherowo Depression are classical examples of that. Two directions are marked in the general tilt of the surface. One of them is the tilt of the scarp of the Pleistocene plateau of the Lake District and the Kashubian Coast eastward, that is in the direction of the Gulf of Gdańsk and towards the contemporary valley of the lower Vistula river. The other direction is the general northward tilt. Thus the meltwater streams were flowing from the dilluvial plateau in the easterly and northerly directions after the retreat of the Vistula lobe from the Pomeranian stage into the Kashubian Coast, the Gulf of Gdańsk and the alluvial plains of the Vistula. In

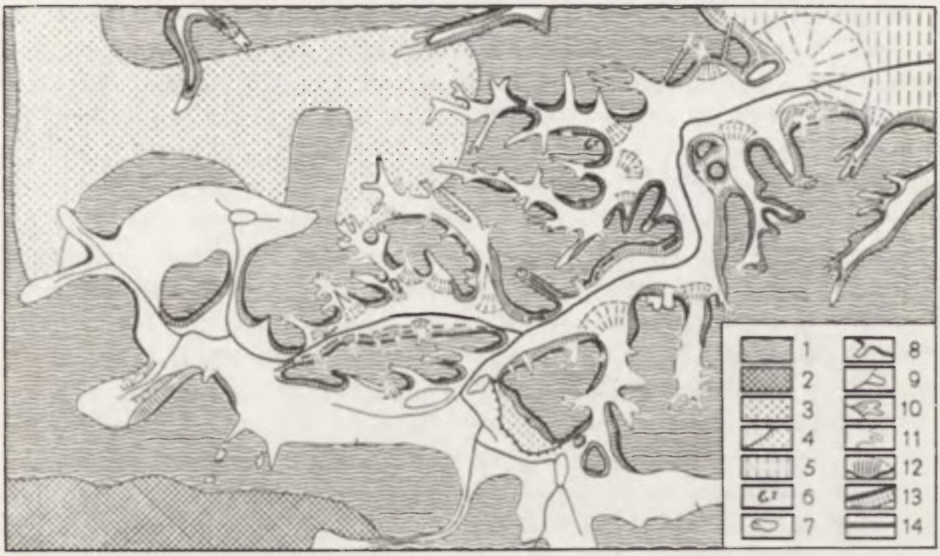

Fig. 3. The morfological map of Brentowo Valley

1 — undulant ground moraine, 2 — frontal moraine, 3 — outwash plain, 4 — kames terrace, 5 — fragment of Oliwa-Sopot "coastal terrace", 6 — Gravel pit, 7 kettle holes, 8 — lateral valleys, 9 — hanging valleys, 10 — rejuvenated hanging valleys, 11 — denudation valleys, 12 — alluvial cones, 13 — slopes, 14 — edges above 20 m of relative hight

such a situation, the meltwaters did not flow away from the front of the glacier, in the direction of the more remote foreland, as was the case in a typical proglacial phase, but according to the surface tilt were flowing in the direction of the glacier. Therefore they were eroding their own drainage pattern at first northward across the area adjacent to the Gulf of Gdańsk and Żuławy, and then, near the contemporary settlement of Reda, westward directly alongside the ice barrier. In such conditions the *pradolina* had been shaped as a typically marginal form. Very numerous consequent streamways cutting the marginal zone (eastern and northern) of the Kashubian Lake District and the subsequent, arrangement in relation to the tilt of the *pradolina*, are the forms which had been left by the drainage pattern that had been developped in this manner. Therefore the consequent character of the streamways is caused by the tilt of the initial surface, and the subsequent course of the *pradolina* was forced by the course of the front of the ice barrier.

The valleys, cutting the marginal zone of the moraine plateau, originate from vast kettle holes or from the outwash plains. In the case of the streamways originating from kettle holes, their outlet — in relation to the present kettle hole is suspended. It shows that the development of those streamways had originated in a period when the meltwaters were flowing over dead ice filling the kettle hole depressions. The age of those streamways is Pleistocene without any doubt. The group of Brentowo streamways, presented in Fig. 3, is an illustration of that.

On the basis of studies of many lateral valleys connected with the *pradolina* arrangement, it is possible to state that there exist, in those streamways, terrace systems which can be correlated with the terrace levels in the *pradolinas*. At the outlet segments of the streamways they descend to a height of about 65 m, above 50 m, 40 m, below 20 m and to the heights of the alluvial floors. Of course, the quantity and the quality of the terrace levels is different along different valleys depending on the height of the main streamway segments from which the lateral valley emerges. There are also very numerous, smaller lateral valleys without any water flow, with floors that are suspended above the *pradolina*. The height of the suspension is correlated as a rule with one of the terrace levels. That correlation of the terrace levels of the lateral valleys and the *pradolinas* gives proof of an uniform and simultaneous development of the drainage patterns.

This very characteristic arrangement of the Kashubian Coast *pradolinas* is the result of a complex development of the pattern of the postglacial water runoff. This development may be groupped in several stages, defined by the terrace levels. The first stage began with the activity of the meltwater streams in the period following the regression of the inland ice from the Pomeranian substage and the temporary stabilization of the ice barrier, directly north of the Łeba-Reda *pradolina*. The meltwater streams flowing down from the Pleistocene plateau were developing their own drainage pattern along the ice barrier and were modelling the ice streamway of the marginal type. This stage lasted through out the development of the terrace systems at the levels above 60, 50 and 40 m. The Redlowo Depression bad fulfilled its function as an *pradolina* during this developmental stage and it did not undergo any considerable changes in the later stage. The absence of further water flow caused it to became a dead form and a document of a definite geomorphological event. Although the flow of the meltwaters at the 40 m level (Fig. 2) was finished in the Reda-Łeba *pradolina*, none the less this streamway has been and still is being modelled by the rivers which have used it as a part of their valleys.

The second stage of the development of the water runoff pattern was marked by a change in its arrangement and the forming of the lower terrace levels. The change of the arrangement of the water runoff was caused by the slight regression of the front of the inland ice to the north of the contemporary shore line, the Swarzewo Isolated Moraine Plateau and the west shores of the Gulf of Gdańsk. The flow of the waters in that arrangement was taking place along the Oliwa-Sopot Seaside Terrace, and after that down a segment, now flooded by the waters of the Gulf of Gdańsk, on the east side of the Orłowo Mound, down the Kashubian Meander, along the Puck Bay, down the Płutnica *pradolina* and over the Plains of the Seaside Marshes. The water runoff in that segment was shaped at the 10-20 m level, and then at 2 to 6 m level. The considerably greater width of that arrangement (2 and 3 times) in comparison with the first arrangement indicates the participation of quantitatively different masses of water in modeling both systems. In the first developmental stage of the drainage pattern only small masses of waters, coming exclusively from the flow of the meltwaters from the adjacent diluvial plateau were active, but in the second stage of development, much greater water masses participated. They were most likely not only the local waters, as was the case in the first stage of the *pradolina* development but also some extra waters. They may have come through the Fordon gap (R. Galon's bifurcation phase [6]) and been a part of the Vistula waters from the Noteć-Warta *pradolina* or the waters of the Pregoła system, or even both of those rivers together. The evidence for this is the cutting off of the Elbląg Moraine Plateau from the Iława Lake District. The evidence of such tributary flow of "outside waters" is the 20-m erosion terrace system on the west and east sides of the Vistula delta (Żuławy).

The first stage of development of the water runoff pattern on the Kashubian Coast was a local one in extent, but the second one with the participation of the "outside waters" took advantage of the elder one, remodeled it and created a new pattern.

The presented above development of this peculiar *pradolina* pattern allows itself to be placed in certain chronological order on the basis of the existing materials, helpful in determining the age of the forms and the processes. These materials consist of: the proved participation of the outwash waters and meltwaters in iniciating and developing the lateral valleys, the appearance of the periglacial structures on the slopes of those valleys, the statement of Z. Pazdro of the existence of the *pradolina* filled with fresh water sediments of Yoldia period and suitable for mining, the determination by R. Galon [6] of the age of the Vistula Fordon gap and its turning to the seacoast during the Alleröd, the determination of the age of the floor parts of the peat layers by the pollen method as being of the Atlantic period, and the C_{14} method (in the Płutnica *pradolina*) for the end of the Boreal period [12]. In reference to the peat layers, it is necessary to point out that they do not lie directly on the erosion floor of the *pradolina*, but on Pleistocene cones and other water sediments which cover the floor. This fact shifts considerably the terminal period of the *pradolina* water runoff to an earlier one than the Boreal and the Atlantic periods [1, 8]. The age of the modelling of the *pradolina* system on the Kashubian Coast may be referred to the period of the late Pleistocene or more exactly to the regression phases of the halting of the Scandinavian ice sheet on the coastal areas to Alleröd period, according to the climatic chronology or to the late Yoldia period in reference to the history of the Baltic Sea.

The later period of the development of the *pradolina* system could be recognized as the third stage of this development in which the fragments of the *pradolina* preserved untill today are dead forms because of the lack of the larger flows of water, with the exception of Reda and Łeba rivers. The sea abrasion moving the shore line rapidly inland and destroying the *pradolina* as well as the isolated moraine plateau, is performing a much greater transformation.

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PERIGLACIAL ELEMENTS IN THE SLOPE RELIEF OF OUTWASH VALLEYS AND LATE-GLACIAL RIVER VALLEYS OF NORTHERN POLAND

ZOFIA CHURSKA

The zone of Northern Poland, extending from the Noteć-Warta *pradolina* (*Urstromtal*, ice marginal streamway) and the southern boundary of Masurian Lake District northwards as far as Poland's frontier line, has been distinguished as an area of glacial relief, with traces of a periglacial environment [3]. Up to the fiftieth of this century, commonly endorsed was the opinion, that in the area of the youngest glaciation periglacial alterations are lacking. However, in the light of more detailed examinations, this opinion proved fallacious. Numerous exposures were discovered in Northern Germany [14, 24], in Holland and in Poland [3, 11, 15, 16, 17, 19,] in which both the structures and the deposits had developed in a periglacial environment.

The major part of the publications mentioned refers to the zone comprising the older stages of the last glaciation. For this zone in Poland J. Dylik [3] coined the term "periglacially retouched region".

The periglacial alterations of the relief of the zone of the last glaciation has been discussed by H. Lembke [13], with dry valleys of the Odra gap as example; these valleys had developed upon fossil ice, and Lembke considered them of late glacial age. Similar conclusions were reached by M. Hannemann [7] as well as by G. Richter [22] as to similar forms occurring in the hinterland of the principal chain of Pomeranian moraines.

In Poland, the present author [2] reported examples of the occurrence of small late-glacial valleys, some of them asymmetrical in shape, from the slope of the Notec-Warta *pradolina*. Asymmetrical small dry valleys from the Leba-Reda *pradolina* were also described by A. Marsz [15] who confirmed that they must have developed during the Bölling and the Alleröd periods.

Further mention of small dry valleys is found in papers published by T. Bartkowski, Z. Churski, J. Szukalski, etc.

In earlier literature dealing with the occurrence of small dry vall eys these were looked upon as forms of erosion by atmospheric waters — processes of higher intensity compared with today. In Poland, the first author to point out their glacial age, was in 1947 W. Okołowicz who linked their origin with frontal precipitation [18]. R. Galon [5] distinguished two periods of development of these land-forms: he connects them with outwashes of the Pomeranian stage, and with older river terraces. B. Zaborski [26] considered them subglacial channels erosively altered.

Up to now, research dealt mainly with small dry valleys disregarding other concave forms encountered on valley slopes. J. Dylik [3] implied an absence of small denudative troughs in Northern Poland; in dry valleys — of sporadical occurrence in his opinion he failed to note the asymmetry of slopes, that feature so characteristic for a periglacial environment.

However, J. Dylik [3] and Ł. Pierzchalko [19] observed ice wedges and structures of involution and free congelifluction in the area examined; this led them to the conclusion that, in Northern Poland; the periglacial environment is conspicuous due to certain modifications in the initial structure of sediments of glacial accumulation, but that in the relief it did not leave any characteristic traces such as long, concave valley slopes.

This conclusion is correct only to a certain extent. As a matter of fact, no long concave slopes are found in Northern Poland; still, there do occur asymmetrical small valleys and above all, denudation small troughs, features so characteristic for classical periglacial regions.

The slopes of pradolinas and ancient river valleys in Northern Poland are most intensively dissected by concave forms of various types and sizes. An example of this type of slope sculpturing may be considered the Drwęca valley (Fig. 1).

On the basis of her detailed field studies carried out for several years, the author distinguishes among the forms mentioned three principal types: denudative small troughs, small denudative valleys, and — hitherto not distinguished at all coombes or niches produced by permafrost waters, as are known solely from contemporaneous periglacial areas [12].

1. Denudative small troughs are forms characteristic for slopes of outwash and late-glacial high river terraces. They also occur in slopes of small denudative valleys, failing, however to reach the valley base. Within these land-forms, three subgroups may be distinguished: (a) nivation throughs, most often found on outwash slopes; these are short forms, more distinctly enclosed than other depressions and, sometimes, resembling miniature glacier cirques, (b) slope troughs encountered most often in the area discussed, and usually occurring exclusively on slopes; (c) the third type are elongated troughs, sharper in contours, representing transition forms between troughs and small denudative valleys.

A feature common to all denudative small troughs are their gentle, slightly concave slopes, the absence of floors, and the marked inclination of their longitudinal axis, often differing but little from the slant of the slope in which these land-forms are scooped out.

In view of the opinion prevailling, that Pomerania is devoid, of periglacial denudative small troughs, the geological structure of these forms has been investigat-

Fig. 1. An example of the intensely dissected slopes of the Drwęca valley in the neighbourhood of Nowe Miasto

ed on a particularly large number of natural exposures and in test pits and borings made on purpose in these land-formes.

Most frequently, denudative small troughs disclose an asymmetrical development of slopes and what is called a "warm" asymmetry. The "cooler" slopes are more gentle. The initial outline of this form, apparent from under the mantle of later fill-deposits, is much sharper defined, showing more distinctly than today the feature of asymmetry, also a "warm" asymmetry (Fig. 2).

The deposits now filling the troughs can be distinctly divided into series of periglacial and of Holocene origin. The series of periglacial sediments are made up of: deposits disturbed by congelifluction (Fig. 3), of structureless sands with pebbles and boulders usually found in the bottom part of the sands, and of differentiated deposits depending on slope exposure. On the warm sunlit slope, light-coloured clays are spread out with a high $CaCO_3$ content, disclosing a variable ragged base; on the cool slope, on the other hand, at the same time clayey sands were deposited, streaked with sands containing much iron oxides. These deposits show distinctly a rhytm of sedimentation (Fig. 4). Further, they are proof of a major participation of water.

The Holocene series consist of brown clayey deluvia with much iron compounds; their thickness increasses uniformly towards the axis of the form. On top of these clayey deluvia, there is usually either a fairly well developed soil profile with a layer of humus sometimes containing charcoal or there is a thick layer of humus

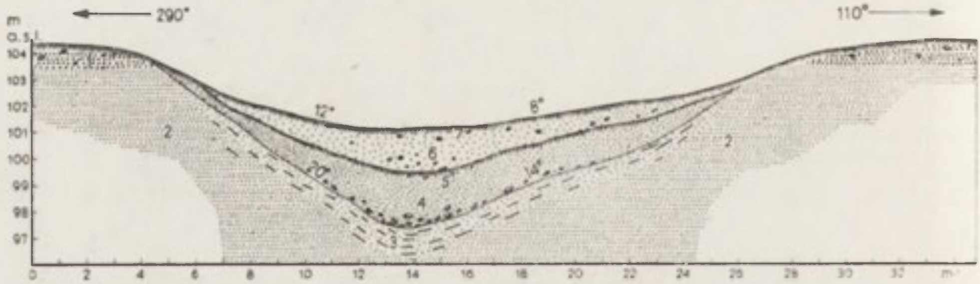

Fig. 2. Profile of the asymmetrical small denudative trough in the neighbourhood of Nowe Miasto

1 — fluvioglacial sands and gravels, 2 — horizontally stratified sands, 3 — unstratified sands with the iron concretions, 4 — unstratified sands with stones at the bottom, 5 — fossil soil with charcoal, 6 — unstratified sands with single pebbles and gravels, 7 — recent soil

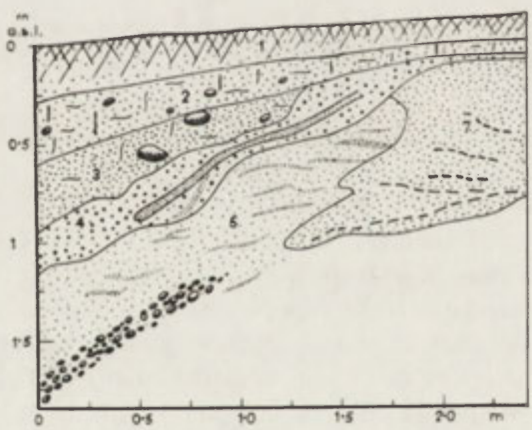

Fig. 3. Deposits occurring in the denudative trough, near Nakło

1 — recent soil, 2 — unstratified, brown iron-stained clayey sands, 3 — iron-stained clayey sands with strongly weathered stones, 4 — light sands with silt lenses, 5 — fine sands with silt streaks, 6 — coarse-grained sands and gravels, 7 — unstratified fine sand with iron streaks

deluvial. The layers terminating this series are streaked or disjunctively stratified sands, and grey unstratified sands with solitary pebbles and boulders, occurring at random in the sand mass. On today's surface of these land-forms usually a soil profile of fairly uniform thickness is found, not increasing in thickness

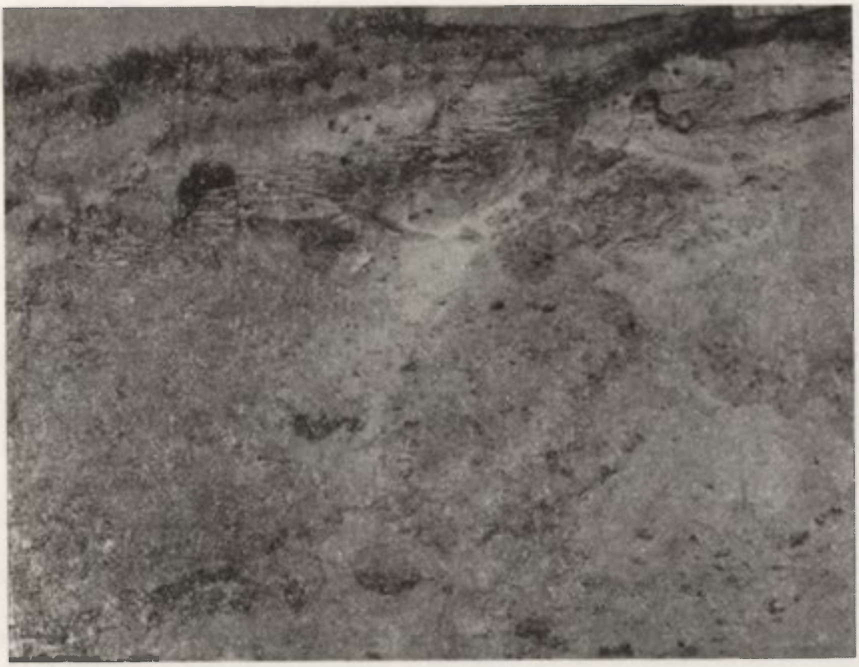

Fig. 4. Rhytmically stratified clayey sands and sands with iron streaks. Photo Z. Churska

towards the axis of the form. This is distinct evidence of today's stabilization of the slope processes.

Only in a few of the troughs inwestigated, the full series of filling deposits enumerated was discovered; even so, the succession of the layers proved always much alike.

The denudative small troughs are often developed in coarse grained sands and gravels. The marked permability of these deposits implies, that they must have been frozen. This presumption seems to be confirmed by the presence of large quantities of iron compounds, accumulated in the shape of ferruginous concretions and of coatings on quartz grains, found in the boundary lines between floors of troughs and substratum gravels or sands. This indicates the presence of a filtration layer for atmospheric waters, laid down on an impervious frozen base.

The age of the denudative small troughs can be determined from their relation to outwashes and late-glacial river terraces, onto whose surfaces the troughs issue. Thus it can be seen, that the oldest denudative small troughs developed in near vicinity to the ice margin during the Pomeranian stage. In the troughs dating from this age, no disturbances of bounded congelifluction were observed, most frequently, they are filled with unstratified sands with boulders. Younger denudative small troughs issuing upon high valley terraces and occurring in slopes of small denudation valleys also have developed under periglacial conditions, as may be concluded from geological evidence reported. We still lack sufficient data to reconstruct climatic conditions as they were during the Bolling. The Allerod has been a period of definite stoppage in the development of denudation forms. Many arguments seem to indicate, that the younger denudative troughs should be dated from the Oldest Dryas.

It is a difficult task to reconstruct today the character of the initial processes taking place in the development of troughs. This may have been fore instance during late spring or early summer, congelifluction flow choosing its paths over irregularities in the slope surface and accetuating them. The denudative troughs disclose a one phase development, and later phases of glacial decline and of Holocene processes may have only contributed to flattening these forms by filling them with slopewash material. Active in the development of troughs were mainly processes of denudation. The character of the forms and the type of deposits filling them make it clear, that the troughs had no linear flow passing through them, and rather were produced by mass movements of free or bound material.

The asymmetrical development of these forms is a further proof of their periglacial origin. The type of this asymmetry seems to indicate periglacial conditions more moderate. The differentiation of deposits on cooler and warmer slopes is the geological confirmation of the asymmetry existing, and enables us to draw conclusions as to the character of the processes leading to an asymmetrical slope pattern of the troughs. On a warm slope thawing proceeded rapidly. The clay covering this slope, saturated with water in its surface layers, slid down in the form of slumps over the frozen slope substratum. On the other hand, on the cooler slope thawing went forth at a slower rate; with participation of larger water amount, clayey sands were deposited, poorly sorted and showing a specific type of streaking with sand layers containing iron compounds.

The morphological and, in particular, the geological features discussed above indicate a formation of the troughs under conditions of a periglacial tundra.

2. Small denudative valleys are a very characteristic element of the slope relief in *pradolinas* and older river valleys. They are principally linked with late-glacial river terraces. Their characteristic features are: a wide (0·5 km in an average) and relatively flat floor, and steep slopes (up to 20° inclination) of concave-convex shape (Fig. 2), frequently asymmetrical, of the warm asymmetry type.

These small valleys disclose traces of having developed in at least two phases. In their older phase of development, there took part troughs, which had originated on the slopes and reached to the very bottom of the valleys; traces of these older forms may be recognized in terrace shelves sporadically surviving, which even today appear to be linked with the denudative small troughs mentioned. Where no shelves remain, the troughs have flattend out somewhere in the middle of the slope and issue "into open air".

In their initial phase of development, the small denudation valleys had contours more gentle, resembling the shape of the troughs. They were linked up with the upper step of the higher river terrace. Their formation probably dates back to the

Fig. 5. The Śleska valley outlet to the Notec-Warta *pradolina*. Photo Z. Churska

Oldest Dryas. After the river had been incised to the present day lower level of the high terrace — which may have taken place during the Bolling [10] — there occurred a temporary stoppage in the course of slope processes and intensive sculpturing of forms. It has been suggested that, at that time, a soil development may have been in progress; some trace of it may be observed in a slope of the small Sleska denudation valley, exposed in the slope of the Noteć-Warta *pradolina*[1]. Here a two colour clay (dark-brown and light-brown) is visible, shoving specific irregular cracks; along these cracks, manganese efflorescences can be seen.

In the Older Dryas, intensive denuding processes set in again. Climatic conditions had become more harsh: congelifluction processes took place on the valley slopes, an importent part was played by downwash and surface water runoff; this flow became concentrated in the axes of the small denudative valleys, leading to an incision into the floors down to the levels of river flow as they were in those times. A trace of the denuding — even congelifluctional in its wider sense [3] — character of this surface flow is the cover of unstratified clayey sands, found on the slopes of these small valleys or in the floors of the troughs and, also, on the slopes of *pradolinas* and older river valleys. This cover is a product of intensive thermal weathering and thermal segregation. Remnants of these structures may be traced in the subsurface layer of Pomeranian outwashes and higher river terraces (Fig. 6), as well as on the moraine plateau.

[1] On the basis of suppositions voiced by members of the INQUA Congress from the German Democratic Republic partaking in an excursion and inspecting this exposure.

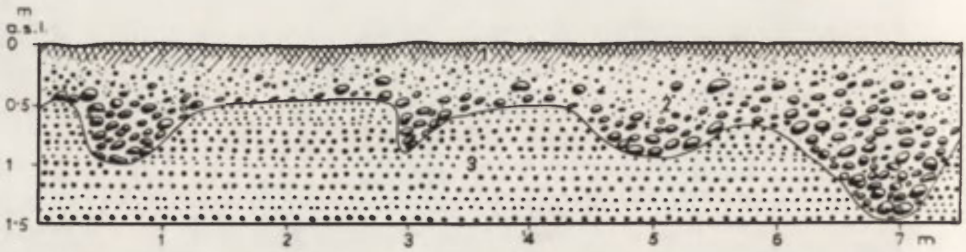

Fig. 6. Upper part of the outwash connected with the Pomeranian stage in the neighbourhood of Brodnica

1 — soil, 2 — unstratified sands with strongly weatherd stones, 3 — stratified fluvioglacial sands and gravels

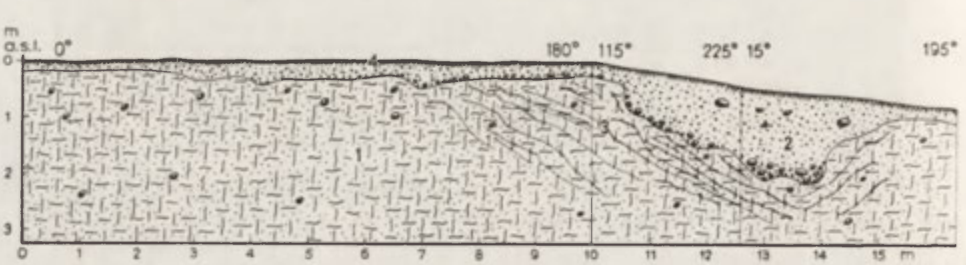

Fig. 7. Geological structure of the slope of the Notec-Warta ice marginal streamway (*pradolina*)

1 — grayisn-green boulder clay, 2 — unstratified sands with weathered stones, 3 — calcareous streaks, 4 — soil

The products of periglacial weathering were put in motion by congelifluction on all inclined surfaces (Fig. 7), or they were deposited by random water flow on valley slopes or floors. The mutual relation of two cover layers of different age may be pursued on the slope of one of the small denudative valleys occurring on the slope of the Notec-Warta *pradolina*. The upper cover consists of congelifluction deposits, mainly of the bound type, while the lower cover layer contains free congelifluction deposits laid down by random water flow. An intensive action of water is also revealed by changes in slope inclination and in the convex-concave pattern of the slopes.

The profile of a section across the small denudation valley of Śleska creek, determined by surveying (Fig. 8), enables us to study in detail the asymmetrical development of the slopes, and the differentiation in intensity of denudation processes on both the warm and the cool slope. The cross-section shows the cool slopes to have suffered much heavier denudation. Displacement of a large amount of deluvial material on the cool slope resulted in shifting the channel of seasonal water flow towards the base of the warmer slope. This process undoubtedly accentuated the valley's asymmetry, brought about by the difference in denudation on warm and on cool slopes.

Thus, the small denudation valleys took shape due to congelifluction processes and the action of surface waters, which were taking place in a periglacial environ-

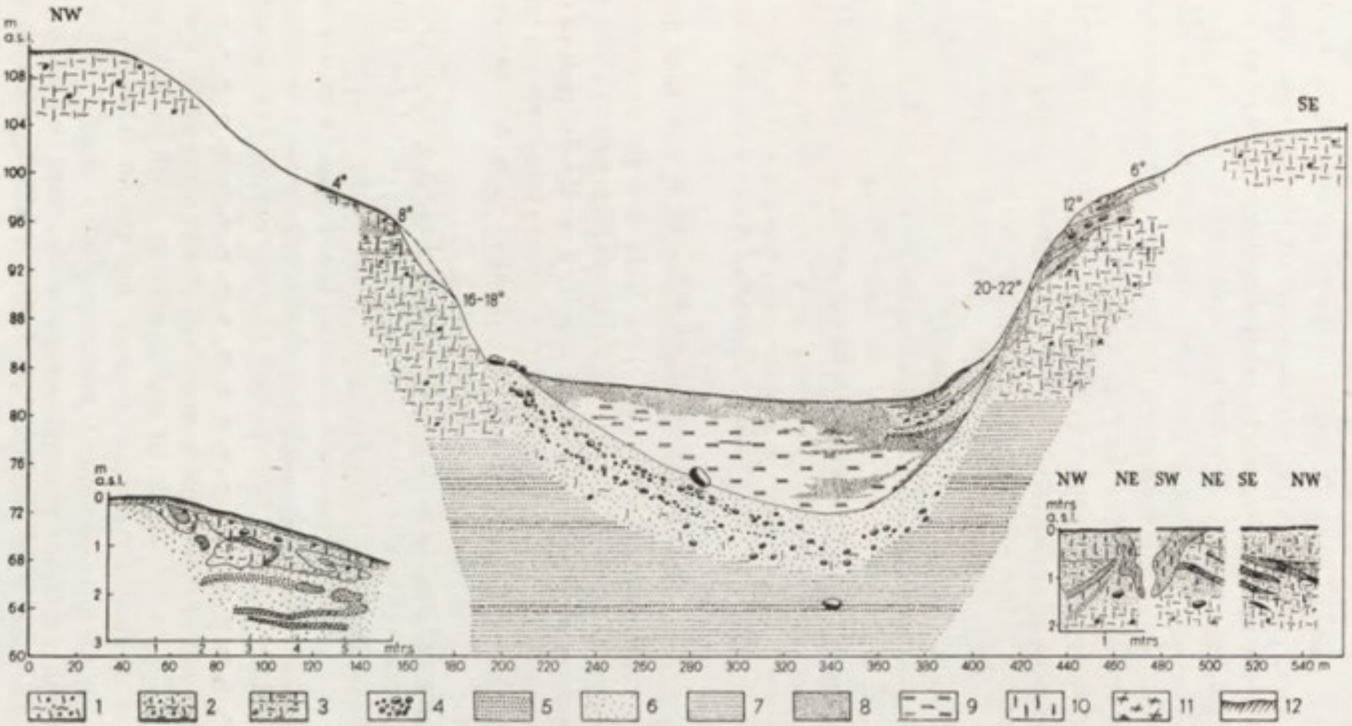

Fig. 8. Cross-profile of the Śleska valley and outcrops with the periglacial structures

1 — brown boulder clay, 2 — grayish-green boulder clay, 3 — dark-brown and light-brown clay, 4 — stones, 5 — gravels, 6 — medium-grained sands, 7 — fine sands, 8 — silts, 9 — peat, 10 — iron streaks, 11 — wood pieces, 12 — soil

ment. The age of these forms, determined morphologically by linking them with dated river terraces [6, 21], has since been corroborated palynologically. T. Przybylski dated the beginning of peat formation in the Śleska valley from the Allerod. Therefore, this small valley itself, as well as others of similar type, must have developed very much earlier. Frequently, small denudation valleys were formed often upon on fossil ice embedded below the surface (traces of these ice blocks are left in the shape of kettle-like depressions in valley floors); they are older than the dunes that sporadically invaded them. In Northern Poland, these dunes are dated from the Younger Dryas or the Preboreal.

In the Allerod period a deep incision by waters of the ancient Vistula took place, reaching down to the level of the present-day upper middle terrace (that is, the floor of the Noteć-Warta *pradolina*). As revealed by longitudinal profiles, of the small denudation valleys only their lowest sectors adapted themselves to the change in base level of denudation. Some of these small valleys are terminated by alluvial cones. In peaty parts of the *pradolina*, these cones are covered by peat; up to now, therefore, no close definition of the age of the cones was possible.

3. The third type of land-forms developed by denudation are coombes or niches produced by permafrost waters. They are found under specific topographical conditions in river cut-offs. Their greatest number was observed in the slopes of the Noteć-Warta *pradolina*. They tie in with the *pradolina* floor; often they are peat-bogged.

The coombes show specific shapes: they are short, deep, with wide flat floors (the ratio of length to width being often 1:1 or 1:2); they are steep-sided, both at their rear face and their sides. Often, these niches are asymmetrical, with a warm asymmetry. Most frequently, the niches start directly from the flat plateau surface. Compared with their depth, the thickness of the deposits filling them is relatively small; the deposits consist mainly of sands with pebbles, clayay deluvia, peats and, laid down on top of the peats, Holocene deposits.

The topography and the shape of the coombes implies an origin similar to the one ascribed to forms of similar features in a modern periglacial zone [12]. This would mean, forms produced by melting ground ice and by congelifluction flow over the slope that produced these characteristic land-forms. Later on, covering of the deep-seated ground ice by congelifluction deposits caused a stoppage of the denudation processes that had been initiated by ice melting. Presumably, this explains the relatively small dimensions of the niches discussed.

On the background of investigations made and results obtained we may assert, that traces of a short-lived periglacial environment in Northern Poland can be seen not only in changes of the initial structure, but also in producing a relief characteristic solely for this environment: minor denudation forms on valley slopes. Simultaneously with the decline of periglacial conditions, there also set in a change of sculpturing processes which from then on started developing different types of land-forms.

Fig. 9. Periglacial small denudative troughs on the slope of the Śleska valley. Photo Z. Churska

Fig. 10. Holocene erosive furrows on the slope of subglacial channel. Photo Z. Churski

Under Holocene conditions, the most important part in slope sculpturing was played by underground waters which, at times, dissected the small periglacial valleys, forming deep V-shaped incisions consistent with the change of base level. The difference between periglacial and Holocene relief can also be seen in the comparison of slopes of old valleys with those of subglacial channels. In the latter, no elements of periglacial sculpturing are found, rather only Holocene elements.

As classical example may serve the periglacial denudative troughs developed in the Sleska slopes (Fig. 9), and erosive forms of identical size found on slopes of channel lakes (Fig. 10). Indeed, comparison of late-glacial land-forms and Holocene land-forms is the most distinct indication of the difference in their morphogenesis.

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GEOGRAPHIA POLONICA 6,1965

FORMATION OF OUTWASH VALLEYS AS SHOWN IN THE VALLEY OF THE PISA

Mirosław Bogacki

Large sandy outwash areas are encountered in Northern Poland. They are connected with the maximum development of the successive phases of the last glaciation. Those most important are the outwash area of the Pisa, those of the Brda, Wda, Gwda and Drawa streams, the Kurpie Forest, and the Augustów Forest. These outwash areas slope S or SE, in conformity with the meltwater outflow at that period. The areas here considered are now drained by the so-called outwash streams. Within the Kurpie region they are the Pisa, the Szkwa, the Rozoga and the Omulew; within the Pomeranian lake district the Drawa, the Gwda, the Brda and the Wda (Czarna Woda).

The writer's geomorphological investigations have been carried out in NE Poland within the sandy outwash areas of the Pisz and that of the Kurpie region (Fig. 1). They deal mainly with the sandy outwash horizons and the fluvial terraces.

The glaciofluvial sediments in the two areas mentioned above are of great thickness, gradually decreasing from the north to the south. In the vicinity of Pisz no such glaciofluvial sediments have been reached by borings to a depth of 30 m (Fig. 2).

In the eastern part of the Pisz sandy outwash area both an upper outwash horizon and a lower one are distinctly discernible (Fig. 2, 3).

During the Poznań stage the front of the glacier had reached south of the Śniardwy lake. Glaciofluvial sands that build the upper outwash horizon accumulated south of the end moraines of this stage.

The upper outwash horizon that had been dissected in the vicinity of Pisz and the accumulated glaciofluvial sediments of the lower outwash horizon filled in the depressions. Northwards across the Śniardwy lake it joins the outwash area on the north-eastern shore of that lake, then farther north the kamelike terrace on the Jagodne lake and that on the southern shore of the Niegocin lake. Southwards the lower outwash horizon is superimposed on the sediments of the upper one as a widely spread outwash fan. The genesis of the lower sandy outwash horizon are connected with those of the frontal moraines of the Pomeranian stage and indicate a small climatic oscillation within this stage [9, 10].

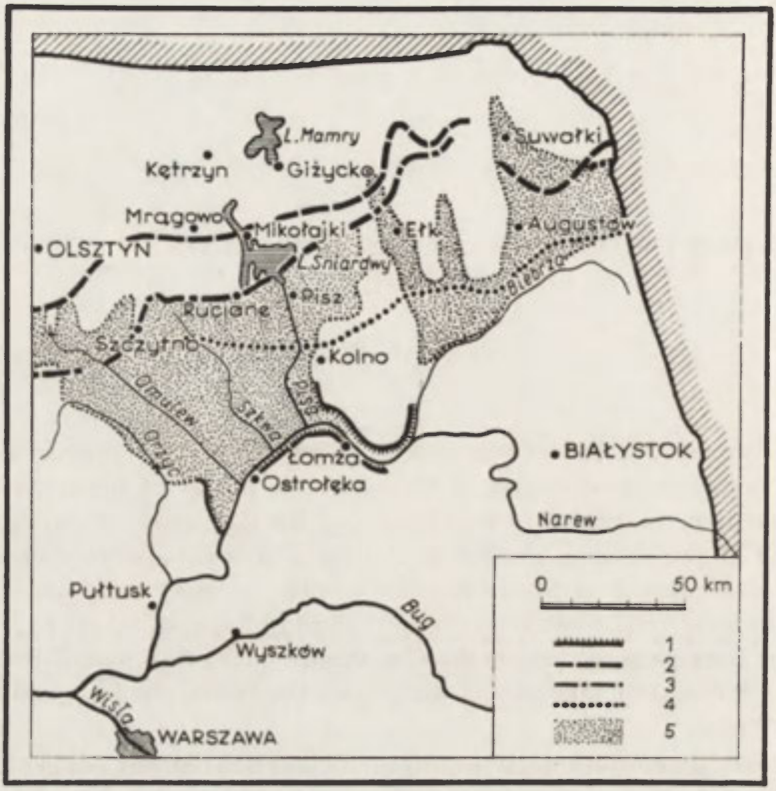

Fig. 1. Major sandy outwash areas in NE Poland

1 — margins of valleys, 2 — extent of the Pomeranian stage, 3 — extent of the Poznań stage, 4 — extent of the Leszno stage, 5 — sandy outwash plains

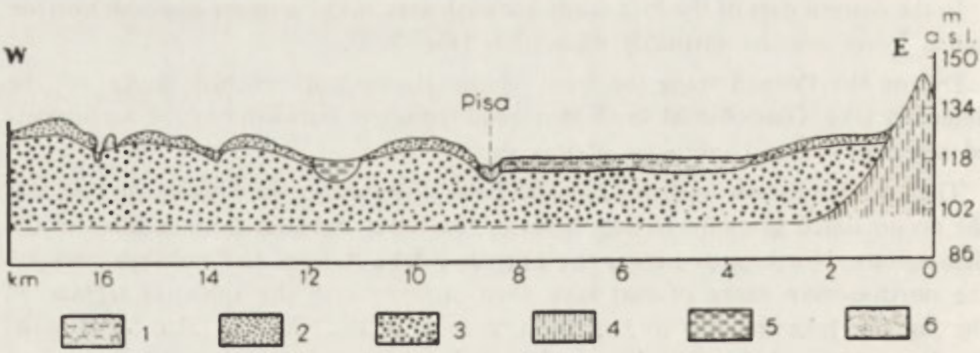

Fig. 2. Geological section S of Pisz

1 — fine-grained sands and silts of the outwash horizon, 2 — coarse-grained sands, upper outwash horizon: 3 — very coarse sands and gravels, 4 — clay, 5 — peats, 6 — fine-grained sands in the Pisa valley

The accumulation of the glaciofluvial sediments of the lower sandy outwash horizon is followed by the development of recent valleys.

In the Pisa valley, similarly as in the valleys of all streams of the Kurpie region, the terraces are incompletely developed, thus suggesting a low intensity of the vertical component of erosion within this area. Accumulation and lateral erosion were the predominant processes here. During the late glacial stage and in the Holocene the above area was stabile or perhaps subject to slow subsidence; slight uplifting movements on the other hand would have led to stronger subsurface erosion. The valley of the Pisa was incised in glaciofluvial sands. The depth of the incision of the sandy outwash horizon ranges from 9 to 11 m. Owing to the great accumulation

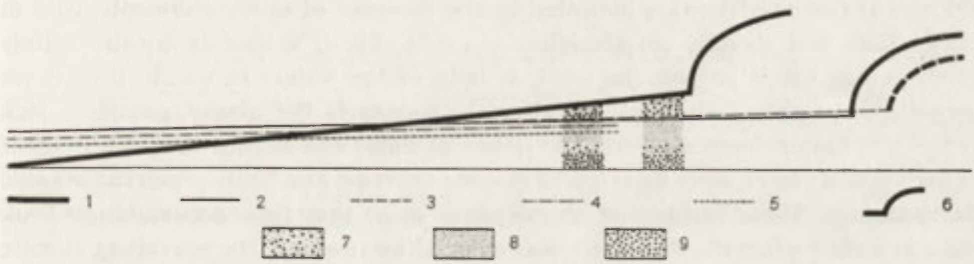

Fig. 3. Diagrammatic longitudinal section of the outwash horizons and of the Pisa terraces

1 — upper outwash horizon, 2 — lower outwash horizon, 3 — accumulation on the lower horizon, 4 — Pisa terrace II, 5 — flood terrace of the Pisa, 6 — front of the glacier, 7 — unequigranular sands, 8 — fine-grained and extremely fine-grained sands, 9 — coarse-grained sands and gravels

of the alluvial sediments in the valley, the present depth is markedly smaller, i.e. 4-7 m, as compared with the lower outwash horizon. In the middle and lower course the Pisa has completely cut through the glaciofluvial sediments and has slightly eroded the deposits underlying the sandy outwash plain, the silts and the boulder clay horizon (Fig. 4).

Two horizons may be differentiated in the valley of the Pisa: the flood terrace and the high terrace. The flood terrace occurs as a continuous and unbroken surface on either side of the river bed while the high terrace consists of ledges on the marginal parts of the valley or as islands on the flood terrace.

Altogether six phases have been observed in the development of the Pisa valley.

Phase I (erosion). Owing to the lack of reliable morphological and palynological data it is hardly possible to determine when the incision of the streams into the surface of the outwash plain began. In view, however, of our fairly good knowledge of the later processes of accumulation and of the morphological development of other valleys in contiguous areas it may reasonably be concluded that, within the sandy outwash plain of the Kurpie region, the incision of the outwash horizon by the local streams began during the warm Allerőd interstage. Recent investigations in Poland and in the neighbouring countries [4, 8, 16] show Allerőd as a comparatively warm period and its final phase as one associated with rapid dead-ice thawing. During the Allerod the waters of the Vistula retreated into the ice-free Bay of Gdańsk in the lower horizon of the middle terrace [6]. At that time the river Niemen similarly changed its course to the Baltic glacial lake [1]. Erosion in the valley of the Narew also occurred during the Allerod.

The general lowering of the erosional base during the period considered above and the increasing amount of water supplied by the thawing of the dead ice undoubtedly intensified the erosive action of the surface run-off water from the sandy outwash plain. In the Pisa valley the work of erosion completed during that period is up to 4 m.

Phase II (accumulation) is indicated by the presence of sandy sediments, ca 3 m thick. They rest directly on glaciofluvial sands. These sediments consist mainly of fine sands which suggest the weak activity of the waters in which they were deposited. Another feature of the alluvial deposits is the almost complete lack in them of humus intercalations. The pollen of *Pinus* and *Betula*, also of a number of herbaceous plants, have been found in some sporadic and badly preserved organic interbeddings. These features of the deposits show that their accumulation took place at a time when the Pisa basin was covered by tundras. The prevailing climate

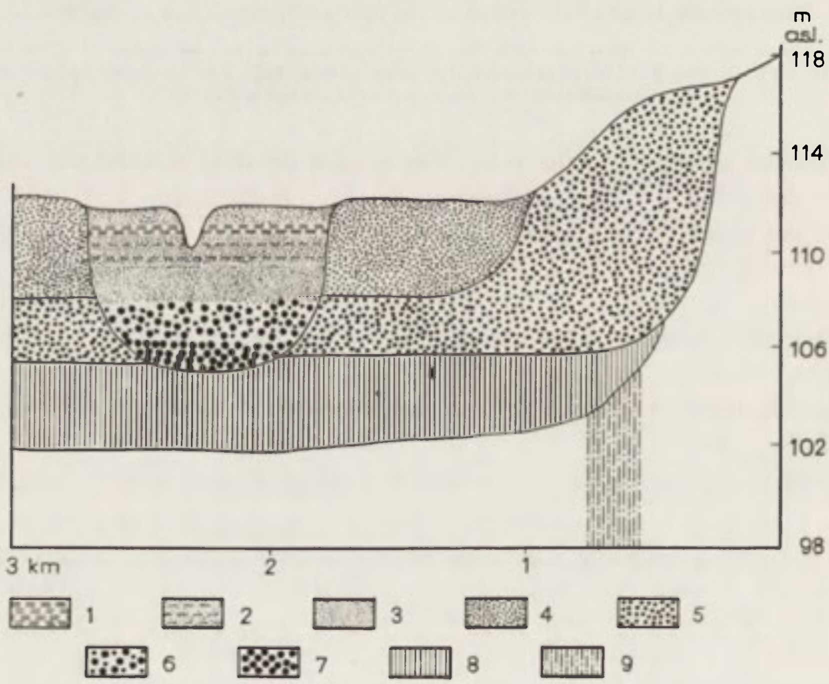

Fig. 4. Section through the Pisa valley S of Ptaki

1 — peats, 2 — silts and sands, 3 — very fine-grained sands of the flood terrace, 4 — fine-grained sands of terrace II, 5 — outwash sands, 6 — coarse-grained sands and gravels, 7 — gravels and pebbles, 8 — loams and silts, 9 — clay

was cool and relatively dry. Hence this phase of accumulation may reasonably be associated with the Younger Dryas. The accumulation of alluvial deposits may have begun at the boundary of the Allerod and the Younger Dryas when the climatic conditions of northern Poland deteriorated rapidly in connection with the movement of the glacier margin to the Salpausselka moraines.

According to radiocarbon dating [2, 11] the Younger Dryas lasted a relatively short period of 1000 years. This is also confirmed by the small thickness of alluvial deposits laid down at that time in the outwash valleys.

Phase III (erosion) is that most readily distinguishable in the Pisa valley. Vertically downward erosion here reached well below the present valley floor and was much more active in the lower course of the river than in its upper part. In the latter case the incision of the Pisa is 4 m below the present valley floor while in the former it ranges from 6 to 7•5 m. Differences in the extent of erosion are connected with the erosional base, the river Narew being the erosional base for the Pisa. Closer to the erosional base the work of river incision was stronger.

The intensification of erosion in the Pisa valley occurred at the beginning of the Preboreal period. This is clearly shown by the pollen analysis of organogenic deposits from the floor of the valley [3]. The bottom of organogenic sediments at a depth of 5 m is dated as Preboreal. The incision of the river occurred somewhat earlier, i.e. probably at the beginning of the preboreal period.

During that period erosion is chiefly associated with a general lowering of base level, due to the fall of the level in the Baltic during the Yoldia stage.

Phase IV (accumulation). Strong erosion was followed by a period of accumulation in the Pisa valley. The valley was filled by alluvial sediments which were formed in three facies: the river-channel facies, the flood facies and the ox-bows facies. The thickness of sediments deposited at that time ranges from 4 to 7•5 m (Fig. 4).

The accumulation of alluvial deposits in the valley began towards the close of the Preboreal period and continued throughout the Boreal and the Atlantic periods. Current opinion, based on the supposed uplift of the base level during the Littorina, believes that the process of accumulation in the valleys did not begin before the Atlantic period [1, 11, 14]. The more recent palynological investigations show that the Atlantic stage, or, at least, its start, did not differ very much from the Boreal stage. During the initial phase of the Atlantic stage the water level in the Masurian lakes did not change as compared with the Boreal stage. It fell gradually in the course of the later Atlantic phase. A similar phenomenon was observed in the pollen diagram of sediments from the Pisa valley.

More important changes in the erosional base and in climatic conditions had started in the Boreal stage and continued throughout the Atlantic stage. As a result the alluvial deposits were accumulated in considerable amounts in many of the river valleys.

Phase V, which coincides with the sub-Boreal stage, is characterised by a pause in both the erosional and accumulative processes in the Pisa valley. According to many authors [4, 15] this was a relatively dry period. Low water level predominated in the lakes and streams. The exposed organogenic deposits in the Pisa valley were partly destroyed in this period and only an intermittent accumulation of muds and fluvial silt have occurred.

Phase VI. The sub-Boreal stage was followed by a gradual increase in the humidity of the climate and the beginning of a new — the sub-Atlantic — stage [12, 15]. It is hardly possible to trace the successive developmental stages of the Pisa valley at that time with accuracy, in view of the lack of direct evidence. Archaeological finds could supply the most reliable data but these are also lacking in this area. Settlers did not enter the Kurpie Forest until rather late, since the first scattered settlements in the Kurpie region date back to the 15th and 16th century.

The growing humidity of the climate during the sub-Atlantic stage raised the water level in the Pisa valley. The ancient ox-bows, partly filled with peat and covered by alder woods were overflooded and transformed into closed water reservoirs. Organogenic sediments, such as clay, gyttja, detritus gyttja and peat [3] were normally accumulating in these basins. The Pisa valley was at that time completely filled with peat. Besides the accumulation of peat, fluvial silts, mainly arenaceous, were also being deposited. These occurred at high water levels when the river overflowed its banks, thus inundating the whole valley floor. The present relief of the valley is the produc of erosion during that phase when the river bed was incised 3-4 m into the alluvial deposits that filled the valley floor and a constant outlet was provided for the stream that drained the Great Masurian Lake District.

The formation of sandy outwash plains and valleys differed somewhat within the particular regions. Their development pattern in the Pisz and Kurpie areas, as presented above, comes closest to the evolutionary pattern of the sandy outwash plains at the boundary of Mecklemburg with Brandenburg. There the glaciofluvial sediments of the Pomeranian stage are also superimposed on deposits of the Poznań stage [7]. The evolution of the surface relief in sandy outwash plains and valleys in the Pomeranian Lake District differed from the one described above.

Four sharp steps occur the outwash area of the river Brda: two in the open outwash plain, the two others in the outwash valley. The origin of the two former steps are, according to R. Galon [5, 6], referable to the ice-sheet melting phases. During the first of these phases the area nearer to the end moraines was covered up with glaciofluvial sediments. During the second phase the meltwaters flowed on the previously formed surface, and became concentrated in several routes and incised into the older outwash surface. During the next melting phase the glaciofluvial waters were divided: one branch followed the Brda valley another that of the Wda, the third one that of the Gwda. R. Galon [5] connects the origin of the two steps within the outwash valley with oscillations of the erosional base in the *pradolina* of the Noteć-Warta streams.

The formation of river valleys that drained the Pisz and Kurpie sandy outwash areas also differed slightly from the evolutionary pattern of the valley of the Brda and of other valleys draining the Pomeranian Lake District. The valley of the Brda is very distinct, well developed, with numerous terraces. In the Brda valley R. Galon [5, 6] has differentiated a flight of 9 terraces. Terrace 9 was formed during a period of intermittent bifurcation in the *pradolina* of the Noteć-Warta streams. At that time the waters partly followed the westward *pradolina*, and partly the northward valley of the lower Vistula. The formation of the lower terraces had followed closely the changes of the erosional base, e.g. the Vistula valley in the case of the Brda. According to R. Galon [5, 6] the lowering of the erosional base of the Brda, or of the Vistula, was a continuous process, as is suggested by the transitional terraces in the Brda valley. At times the lowering process was more rapid and this is expressed by the unbroken distinct terraces in the Brda valley. At other times the lowering of the level of the Vistula, and causally of the Baltic, ceased, and was followed by a period of accumulation on the erosional terraces.

Differences between the formation of valleys that drained the sandy outwash areas of the Pisz and Kurpie regions and the development of the Brda valley may reasonably be associated with the distance from the erosional base, i.e. the Baltic and the Vistula valley.

The Brda valley, being situated in close proximity to the Vistula valley, was probably more susceptible to every change in the level of the Vistula connected with changes in the level of the Baltic. The formation of river valleys at the close of the last glaciation and during the Holocene was doubtless strongly influenced by crustal movements. Upheaval in the Pomeranian Lake District was probably stronger than that in the Pisz and Kurpie sandy outwash areas. According to B. Rosa [13] the isostatic positive movement of the southern coast of the Baltic occurred during the Yoldia stage. Also the vertical land movements — either isostatic and confined to small areas, or epeirogenic and involving larger areas — probably occurred throughout the Holocene and at the close of the last glaciation.

The extent and strength of the positive crustal movements during the period considered here are not adequately known. The available morphological evidence suggest a greater intensity in the Pomeranian Lake District as compared with the southern part of the Masurian Lake District and with the basin of the Narew, as expressed by the differentiated development of the river valley.

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AREAL DEGLACIATION IN THE WIELKOPOLSKA LOWLAND

TADEUSZ BARTKOWSKI

Generally speaking during the last glaciation the forms of the marginal zone of the Wielkopolska Lowland can be divided, according to the well-known division by K. Gripp [11], as follows: inland ice push forms, i.e. push end moraines — *Stauchendmoränen* and, forms of its recession, the so-called accumulation end moraines — *Satzendmoränen*. This division, as far as can be seen, is based on the criterion of ice-sheet balance, push forms being forms of positive balance while the recession forms are those of negative balance.

The subject of this paper is to consider the negative balance forms of the ice-sheet, i.e. recession forms, in the Wielkopolska Lowland as well as their origin during the last glaciation. All the recession forms of the Baltic inland-ice may be grouped along two lines of the inland-ice margin of the Leszno (Brandenburg) and Poznań (Frankfurt) stages.

Forms of the marginal zone of the Leszno stage (like those of the Poznań stage) are found in two subzones: intraglacial and extraglacial ones.

Forms of the extraglacial subzone are the following:

1. Morainic hillocks, built of fluvioglacial, stratified material (usually in the flat or dune sedimentation phase) with steep slopes, especially when declining in the proximal direction. Inside these forms, in the proximal parts the stratification of fluvioglacial deposits shows some deformations caused by the contact with the ice-wall supporting the hillock slope. The strata dip mostly in the distal direction, though dips of stratification are also met in all directions (Fig. 2). These forms have been regarded as end moraines [10].

2. Sometimes, instead of hillocks, plateau-rampart forms are met. They are built of fluvioglacial deposits mostly horizontal, or dipping slightly towards distal direction. They represent forms of outwash accumulation, being short outwash cones, grown together and forming a border outwash rampart [4] (Fig. 3). It is worth noting that these forms occur in places where the Baltic (Würmian) inland ice margin met the substratum depressions, e.g. Brody-Drewitz Rampart — in the Lubsko-Gubin-Cottbus depression (in the lower basin of the Nysa Łużycka). A similar rampart blocks another substratum depression of the Baltic inland ice, i.e. between Sława Śląska and Nowa Sól (between the Odra and Obra rivers, west of Leszno).

Fig. 1. Morainic hillock built of stratified drift, south of Ostrowieczno Lake (149 m a.s.l.). The strata are dipping towards south, towards the interior of the hillock. Note slight deformations of stratification due to ice contact in the proximal part of the hillock

3. In places where the inland-ice border ran along the foot of high, large surface forms, as, for instance, at the foot of Zielona Góra Rampart (push end moraine) or of the Old Pleistocene restling, Kolo Restling [3,5], kame terraces may be found while in places, where meltwaters of the inland-ice could flow freely into the region of the morainic plateau of the foreland of the inland ice, typical flat, outwash fans can be encountered (incised into the substratum), as, for instance, near Leszno (Fig. 4). Along the interior part of the marginal zone, in the intraglacial subzone of the marginal zone there occur:

4. Marginal dissection forms, i.e. marginal tunnel valleys and morainic plateau erosional restlings, dividing the valleys. These take the shape of ramparts, as we can see between Leszno and Gostyń-Dolsk (Dolsk Restlings), near Lubsko (Lubsko-Bobrowice Restlings), near Bucz (west of Kościan Bucz Restlings), or between Konin and Lubotyń [4,5].

5. Numerous kames and eskers, occurring in lowerings as, for instance N of the Brody-Drewitz Rampart, and N of Zielona Góra near Sława Śląska, may also be found in other places, not only in depressions, as in marginal tunnels near Gostyń, in the Gosławice Lake tunnel valley, north of Konin, etc.

6. At the most interior intraglacial part of the marginal zone numerous, rather low, morainic hillocks, built only of till, frequently with a cover of till resting on a nucleus of stratified fluvioglacial drift, or lacking this cover. These forms are

rather insignificant, but in the marginal zone of the Poznań stage they become magnificent.

The inventory of the marginal zone forms of the Poznań stage is largely the same as in the marginal zone of the Leszno stage, though, because of different morphogenetic conditions of its formation these forms developed in a slightly different way. The factor that was quite unimportant in the marginal extraglacial zone of the Leszno stage, viz. dead ice, became quite prominent here. This ice covered large portions of the extraglacial zone of the Poznań stage, above all in lowerings (e.g.

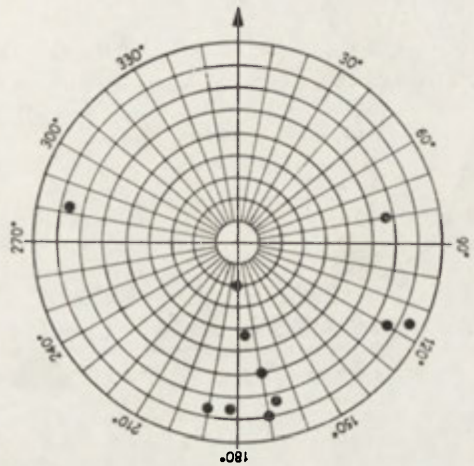

Fig. 2. Dipping of strata in the hillocks, surrounding the hillock 149 m a.s.l. south of Ostrowieczno Lake and in the latter itself. Note the dipping of strata in all directions with preponderance of southward (distal) dipping

Obra lowering near Międzyrzecz and Pszczew [2]), though it also occurred on the morainic plateau, as can be concluded, for instance, from the occurrence of a large closed depression in the hinterland of the Lewice outwash zone, SSE of Międzychód (Fig. 5). The edge, 8-10 m high, which is seen here is a structural edge (developed by the melting of a portion of dead ice abutting upon the cone) and is not erosional in origine. That there was dead ice on the morainic plateau is witnessed by numerous kame hillocks west of Poznań, near Buk or Pniewy, regarded hitherto as recession end moraine hillocks.

The presence of dead ice can be seen in the "beginnings" (at the heads) of outwash cones. A very strong flow of meltwaters first melted this dead ice and then much fluvioglacial material was accumulated. After the dead ice melted, small hillocks were found. They have a nucleus of stratified, fluvioglacial material because of the excess of the fluvioglacial mass in this place in relation to the amount of fluvioglacial outwash plains. Because of this the „beginning" of outwash plains are hillocks, hitherto taken as end moraines (e.g. at the outlet of Kiekrz Lake, south of Lusowo Lake etc.)

Fig. 3. Geomorphological sketch-map of a section of the marginal zone of Leszno (Brandenburg) stage, west of Lubsko, near the frontier with East Germany, at the Nysa Łużycka river
1 — fluvioglacial deposits building a fluvioglaciai plateau-border outwash (extraglacial), 2 — fluvioglacial deposits building kame and esker hillocks (intraglacial), 3 — contourlines (5 m) and height in m a.s.l., 4 — meadows

Another new factor shaping the marginal zone of the Poznań stage is the formation of the inland ice substratum. At the border of the high niveau level of the morainic plateau, at the south edge of the depression of the Gorzów Basin there is a zone of morainic hillocks, regarded till now as the main end moraine belt of the Poznań stage [10 and others]. However these hillocks are built of fluvioglacial material for the most part; this material is stratified with a till cover on top. Sometimes the morainic till cover is very rich and then the hillocks have either a very thick layer of morainic till or are entirely built of it. There are, however, hillocks without

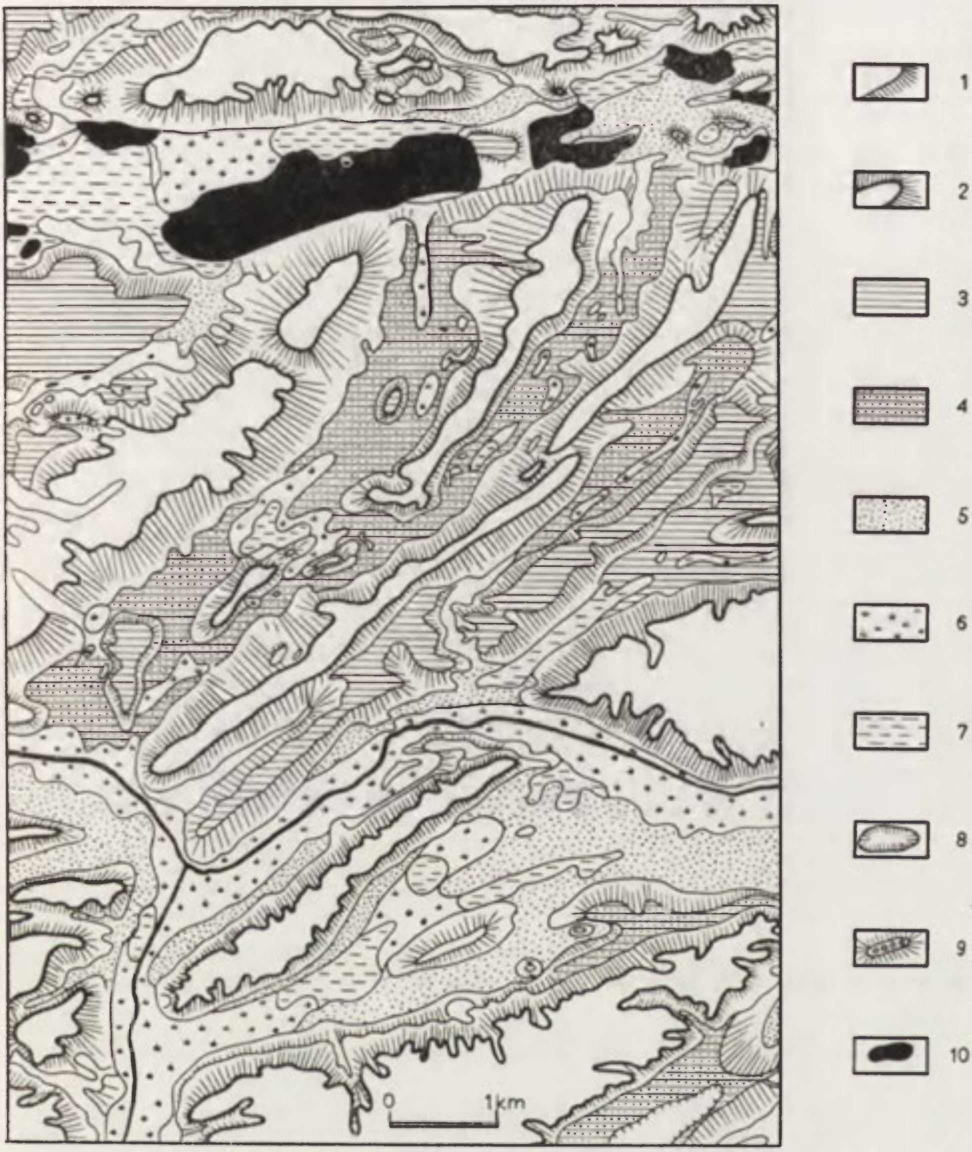

Fig. 4. Geomorphological map of a section of the marginal zone of Leszno (Brandenburg) stage south of Dolsk, east of Leszno

1 — edges, 2 — highest morainic plateau level, 3 — "middle" morainic plateau level, 4 — "low" morainic plateau level, 5 — sandy terrace level in marginal tunnel-valleys, 6 — flat, meadow bottom in marginal tunnel-valleys, 7 — flat, peaty bottom in marginal tunnel-valley — effect of an organogenic accumulation in former closed depressions, 8 — "dry" closed depression, 9 — esker, 10 — lakes

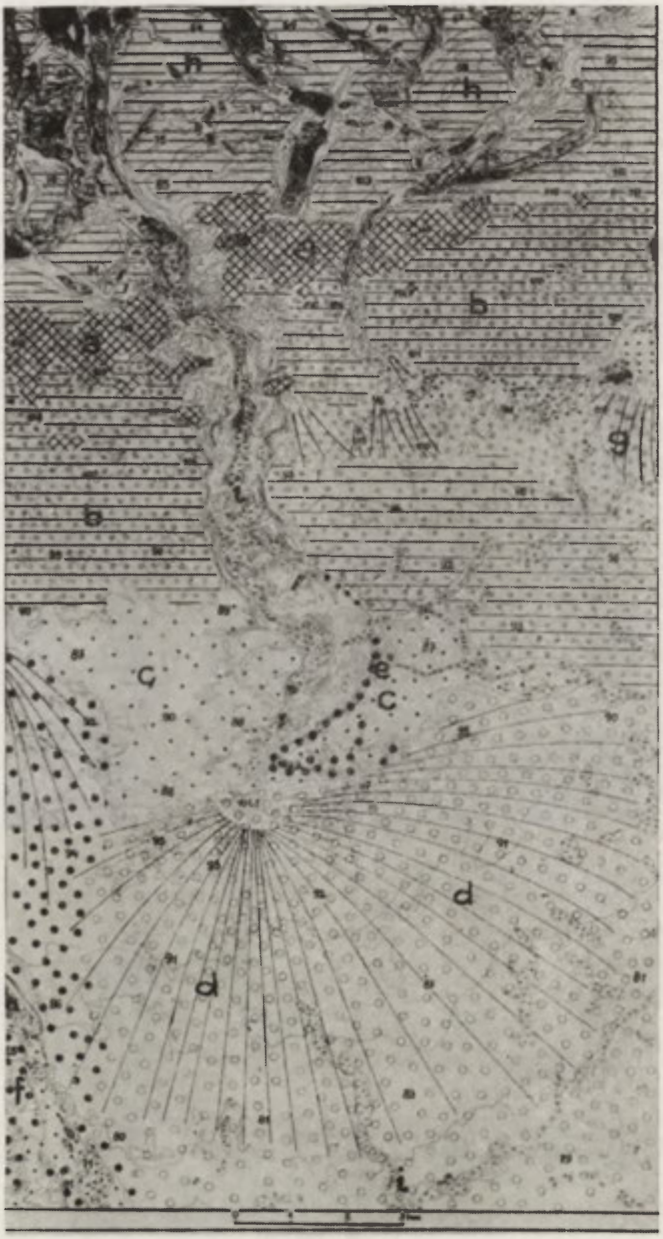

Fig. 5. Geomorphological sketch map of the marginal zone of Poznań (Frankfurt) stage, SSE of Międzychód with Lewice Outwash Fan

a — morainic hillocks with stratified drift core, b — morainic plateau with fluvioglacial cover, c — large depression without superficial outlet in proximal side of Lewice Outwash Fan, d — Lewice Outwash Fan, e — esker, f — Dormowo Outwash Fan and kame plateaus, g — secondary, minor outwash fans, h — flat and undulant morainic plateau in proximal part of marginal zone, dissected by "marginal" and "perpendicular" tunnel-valleys, i — meadowy, mostly peaty, flat bottoms of depressions any till cover at all. The stratified fluvioglacial deposits are usually horizontally stratifled with dips of strata near the slopes according to their inclination as seen in the hillock 118·8 m a.s.l. SSE of Międzychód (Fig. 6).

As the structure of these hillocks shows, they are forms developed in caves and crevasses in the inland ice. The till cover shows that the fluvioglacial material was deposited in the lower parts of the inland ice, containing still unwashed morainic material (boulders, morainic till or unsorted sand, gravel, stone material).

These are forms described by W. Niewiarowski as dead ice moraines in the region of the Chełmno Moraine Plateau [13]. A. A. Asejev [1] called them *Moränenhügel mit sortiertem Kern*. These hillocks occur here—as in the intraglacial zone of the

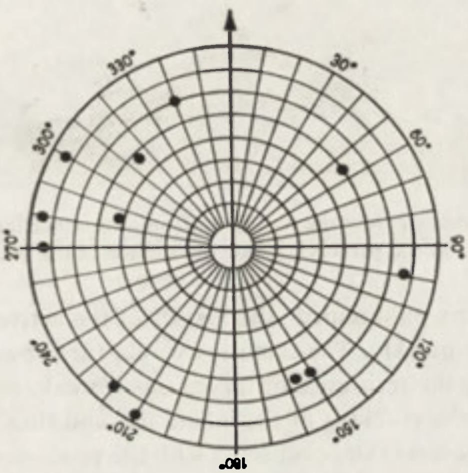

Fig. 6. Dipping of strata in sand pit of the hillock 118·8 m a.s.l. Note the dipping in all directions, typical for kame hillocks

Leszno stage—as wide hillock chains superimposed upon the edge zone of the highest morainic plateau level dipping towards the north, to the depression centre of the Gorzów Basin, S of Międzychód.

Considering the fact that these forms were created in caves and crevasses in the inland ice, as well as the situation of the hillock zone on the edge of the substratum of the inland ice, the author assumes that the formation of crevasses in the inland ice was brought about by the existence of a plateau step. Its existence, together with the action of ablation, caused the thin (owing to ablation) inland ice to die *en bloc* in the distal part, being, however, active in the proximal part for some time afterwards. The hillock zone was formed where the active ice came into contact with the passive or dead ice. This process is well illustrated in the included diagram (Fig. 8).

It should be added that in order to cause an intensive formation of crevasses in the inland ice, the existence of a substratum step in the substratum of the inland ice is not necessary because, as a result of the ablation activity, the marginal parts of the

Fig. 7. Structure of the sand pit. Note the mostly horizontal stratification of fluvioglacial deposits and absence of till layer in the parts lying near the surface of the pit, in top of the hillock

inland ice being thinner would impede the masses of the active ice in the hinterland from moving on, more quickly. These masses would turn towards the upper strata of the inland ice, where the resistance of the masses is weak, and this upward movement of the ice caused the cracking of the inland ice, and thus we get the formation of crevasses. This is a general rule, connected with the process of inland ice recession which in the area of the Wielkopolska Lowland receded areally, and not frontally, as was assumed. An evidence of this is found in the general occurrence described above of the threefold constitution of the marginal zone in the Wielkopolska Lowland:

a) contact zone of active with passive ice (hillock zone),

b) zone of occurrence of passive or dead ice (kames, eskers, vast closed depressions, etc.),

c) zone of extraglacial outflow of inland ice melt waters in the place where the "inglacial" water level „crossed" the ice surface (outwash cones, hillocks at the head of outwash cones, structural edges.)

Such is the development of the marginal zone of the Poznań stage between Międzychód and Ostoróg (southern margin of the Gorzów Basin *sensu largo*, west of Poznań, near Gniezno, in the marginal zone of the Leszno stage, in the region of Leszno, Gostyń, Dolsk, as has been examined).

The conception of areal deglaciation, although quite new to the region of the Wielkopolska Lowland, is not new to other lowland regions. In 1927 it was accepted by K. Bülow for the North German Lowland [6[, and by A.A. Asejev [1] for the Russian Lowland. In the opinion of R. F. Flint ice margins without end moraines

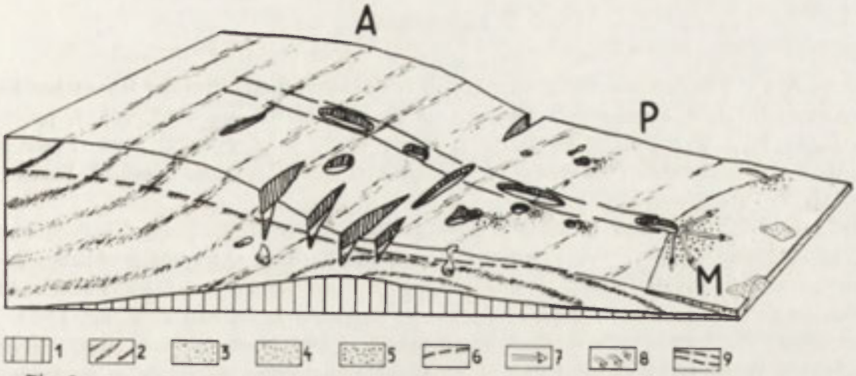

Fig. 8. Formation of marginal zone (based on circumstances encountered in the Lewice Fan marginal zone section

1 — inland ice substratum, 2 — shear plane surfaces with admixture of morainic material, 3 — fluvioglacial material, mostly well sorted, 4 — "inglacial" pure ice, formed of frozen meltwater, 5 — outwash, "extraglacial" deposits, 6 — "inglacial" water level, 7 — direction of extraglacial outflow of meltwater 8 — direction of intra- and supra-glacial outflow of meltwater, 9 — contours of inglacial crevasse, forming a glacial tunnel. A — active ice P — passive ice M — dead ice

called "attenuate drift borders" are, at least, as common as ice margins with such forms [7]. R. Galon [9] ascribes to deglaciation an important role in the process of inland ice recession by wastage of large portions of dead ice, though he thinks, too, that deglaciation also took place by means of a number of oscillation of the edge of inland ice that left numerous recessional end moraine chains with glacitectonically disturbed fluvioglacial material, especially in the Pomerania [8].

In the author's opinion areal deglaciation is conditioned both by climatic causes and obstacles in the inland ice alimentation (both are strictly dependent on each other) and because of this he accepts the possibility of a greater role of the active ice in the inland ice recession in such places where the configuration of the inland ice substratum favoured a long existence of streams of very active ice, as could be the case in the lower Vistula valley, lying in the prolongation towards the south, of the chief depression of the Baltic Sea. In the region of the Wielkopolska Lowland, however, forms created in the area of the Baltic glaciation by the active ice during recession, in the shape of thrust fluvioglacial material, are not found anywhere.

Disturbances of stratified material in hillocks, regarded by S. Kozarski [12] as end moraine hillocks may also be explained as saggings of the fluvioglacial masses as a result of the thawing out of dead ice upon which those deposits were resting. We need not, therefore, accept numerous oscillations of the edge of the inland ice in order to explain those disturbances. R. F. Flint [7] explains similar disturbances in the same way, calling them ice contacts.

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FOSSIL PLEISTOCENE RELIEF IN CENTRAL POLAND

CECYLIA RADŁOWSKA

Geomorphological investigations carried out in the NE foreland area of the Holy Cross Mts have shown the occurrence of fossil Pleistocene topography. The area is dissected by the left bank tributaries of the Vistula in a zone where the river breaks through the southern uplands (Fig. 1).

The relief is wholly periglacial in character. In the more southern parts of the upland the periglacial morphogenesis included the exhumed Tertiary forms, while in the northern peripheral areas the periglacial processes developed on the mantle of glacial deposits and forms. Hence we can observe here a mingling of relief elements from the uplands with those from the plains of central Poland. In the valleys the dissimilarity of features is also repeated though apparently there are no differences between them. All the valleys readable in the present surface were formed after the Riss glaciation had invaded that area and on the whole their forms are simple with only poorly terraced slopes. Their outer appearance does not suggest differences in age at all. The character of the valleys, however, differs and so does the degree of dissection of the terraine. In the southern part, where the Pleistocene covers have been destroyed, all the rivers flow in accordance with the dissections in the bedrock, while in the north they have been cut only in Quaternary material and have no direct contact with the erosional forms in the fossil relief which is covered by sediments of the Riss glaciation.

As a rule it can be shown that wherever Tertiary forms have been exhumed the relief of the sub-Quaternary bedrock is virtually the same as that of the present surface, only the valleys are more shallow and the denivelations smaller. Farther north, however, with an increase of the thickness of the Pleistocene deposits, the present morphology of surface becomes more independent of the buried older forms. Here we have two reliefs, the recent and the fossil, each often displaying quite a different valley pattern. The relief of the Quaternary bedrock as well as the recent one are of a polygenetic character. Important changes of the pre-Quaternary surface occurred during the glacial epoch, such as the filled in erosional forms cut in Mesozoic and Tertiary rocks, and also the Pleistocene deposits and the presence of country rocks in the morainic boulder clay and in valley sediments. The only

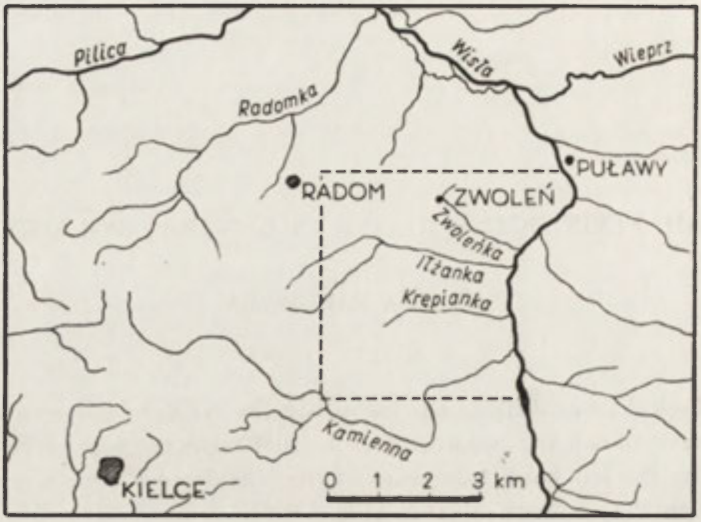

Fig. 1. Sketch map of the investigated area

valleys that may reliably be considered as Tertiary structures are those not completely filled to their bottoms with deposits of Pleistocene waters.

In the antecedent valley of the Vistula, and also in its tributary valleys, the deepest incisions are of interglacial origin [8, 9, 11, 12, 13, 15, 17].

The NE foreland area of the Holy Cross Mts was involved in the Mindel and the Riss glaciations, and was influenced by the cold climatic conditions of the Würm glaciation. In support of this there is evidence supplied by the pre-Mindel cooling of the climate which led to an extensive filling in of the valleys with preglacial material. The preglacial deposits do not, however, contain Scandinavian material, hence it may reasonably be supposed that the area here considered was not covered with an ice-sheet until the time of the Mindel glaciation.

Elements of fossil relief from the older Pleistocene

The bedrock relief is buried under a cover of Quaternary sediments. The deeply incised valleys are in contrast with the rather monotonous relief of the upland watershed area. Evidence of the Tertiary origin of valleys will be found where preglacial deposits are encountered. The extensive preglacial cover near Zwoleń is a record of the accumulating process of the old valley of the Vistula (Fig. 2A).

The petrographic character of the preglacial sediments (dominance of country rocks with an admixture of Carpathian pebbles) clearly indicates a connection between the origin of that covering and the action of a cold climate. Preglacial accumulation occurred in two cycles separated by a horizon that bears traces of

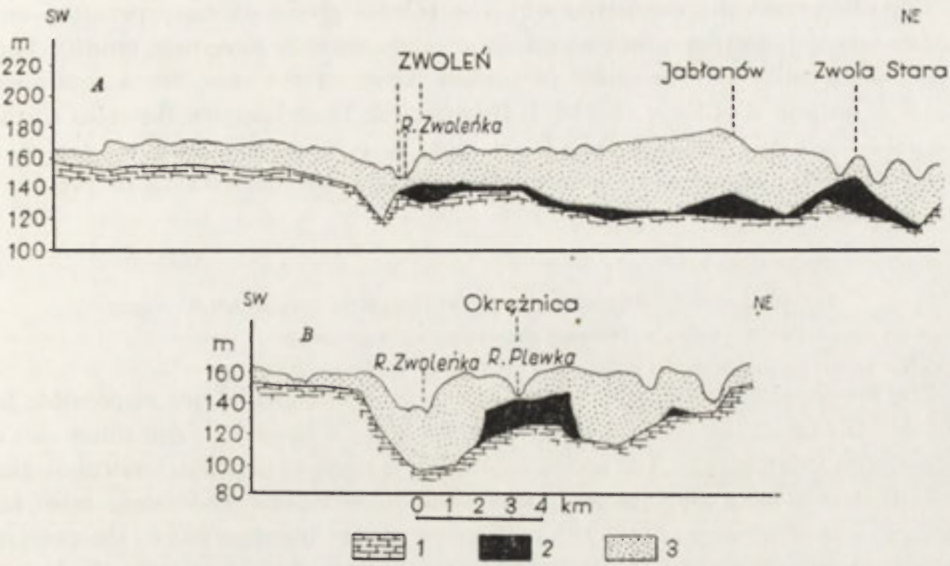

Fig. 2. A — position of preglacial deposits in the fossil valley of the Vistula, B — position of preglacial deposits in relation to present valleys

1 — Quaternary bedrock, 2 — preglacial deposits, 3 — Pleistocene deposits

plant remains [15]. In accordance with the investigations of A. Jahn in the Lublin Plateau [6] and in Lower Silesia [7], also those of S. Z. Różycki in Warsaw [14], the preglacial period of the NE peripheral area of the Holy Cross Mts, has been chronologically referred to the older Pleistocene.

Preglacial accumulation resulted in the shallowing of valleys formed during the Upper Pliocene. This filling in of the valleys corresponded with a simultaneous denudation of the upland area. These two contrasting processes lowered the denivelations of the previous relief by at least a score or so of metres.

Preglacial deposits do not lie at the bottom of present-day valleys but on their slopes, or they fill in the buried valleys completely (Fig. 2A, B).

On the slopes of some valleys, sediments preceding the transgression of the Mindel ice sheet are deposited on partly eroded preglacial gravels. This suggests a period of erosion prior to the Mindel glaciation. The extent of the dissection cannot be reconstructed since the valleys which developed along the same depressions as forms of earlier erosion were effaced by later deep erosion of greater strength.

The Mindel moraine is underlaid by fluvial and lacustrine deposits. Two clay horizons separated by fluvioglacial sediments belong to that glaciation. Outwash deposits from the Mindel moraine rest directly on bedrock in the deepest incisions of the valleys or on valley slopes. Such is the position of remnants of the Mindel moraine in all the Vistula tributaries that have been analysed.

Deposits genetically connected with the Mindel glaciation have persisted until today only in the valleys, because on the upland area they have been eroded. Their very scanty relics may be found in various points of the area, but a continuous cover is missing. For these reasons it is impossible to reconstruct the relief of that period. A gradual modification of the relief is perhaps suggested by geological evidence from the valleys, expressed by considerable destruction of the lower boulder clay horizon.

The importance of erosion of the Mindel/Riss interglacial period in the formation of the relief

The erosional processes during the Mindel/Riss interglacial are responsible for the destruction of the boulder clays from the Mindel glaciation, and often also of deposits underlying them. The period named above plays an important morphological role. It terminates a long period in the evolution of the old Pleistocene relief and lends it a modified appearance although its netway of streams follows the previous valley pattern. Processes of deep erosion then attain their maximum. It dissects and erodes not only Pleistocene forms but also reaches the bedrock. Hence, the Mindel moraine has not been preserved *in situ* at the bottom of valleys. Material eroded from it occurs in the deepest incisions.

The depth of the incision of interglacial rivers is shown in the attached sections (Fig. 3). The valleys then were deeper by some tens of meters than are the recent ones. Some outcrops (Fig. 4) are also a source of information about erosion on

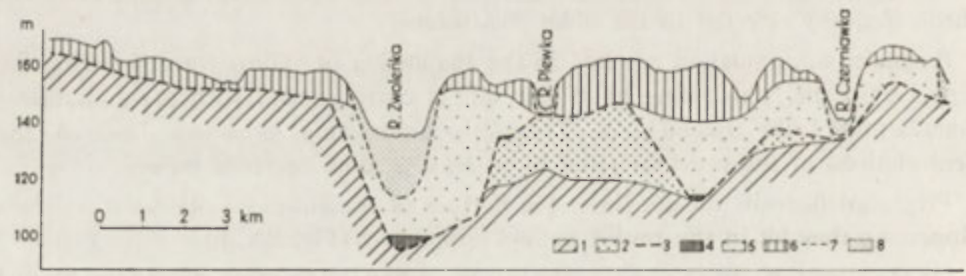

Fig. 3. Pleistocene fossil relief

1 — ceiling of pre-Quaternary rocks, 2 — preglacial deposits, 3 — incision from the Mindel/Riss interglacial, 4 — outwash sediments from the Mindel moraine, 5 — fluvial deposits prior to the Riss transgression, 6 — Riss sediments, 7 — incision from the Riss/Würm interglacial, 8 — Riss/Wurm interglacial deposits

the slopes. The lowering of the level of streams results in the formation of terraces due to the alternating processes of erosion and accumulation. These terraces are covered with old-Pleistocene material.

The increased energy of deep erosion during the Mindel/Riss interglacial has also been observed in other parts of Poland [17].

"The Mindel/Riss interglacial is characterised by the formation of deep erosional valleys. These valleys cut not only deposits of the older Pleistocene, but also those of the Pliocene and even all of the Tertiary" [3]. This citation also applies to the valleys of the Masovia and Kujawy regions.

In the Silesian Plateau interglacial events follow the same pattern as those in the NE marginal area of the Holy Cross Mts. [5].

The Mindel/Riss interglacial period is the subject of special investigations by S. Z. Różycki [13]. According to that author during the climatic optimum of that interglacial deep erosion attained its maximum. In many places it exceeded the

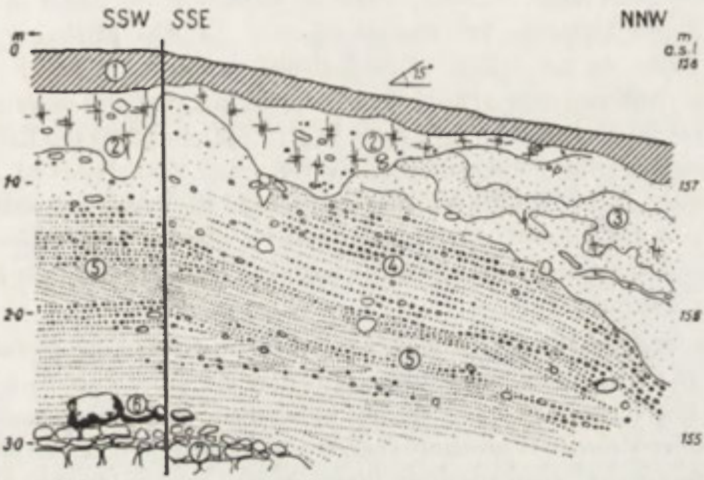

Fig. 4. Relics of the Mindel/Riss interglacial erosion on valley slopes

1 — soil, 2, 3 — slope deposits from the Würm glaciation: boulder clay with structureless sand transported by solifluction, 4, 5 — glaciofluvial deposits prior to the Riss transgression, 6 — pebbles of Scandinavian material/limestones and crystalline rocks\ of the Mindel moraine, 7 — Quaternary bedrock

level of erosion from the older Pleistocene. In the uplands of the provinces of Lublin and Małopolska the depth of the incision during the Mindel/Riss interglacial period occurs 80 m below the level reached during the Mindel glaciation.

The results of the strong erosional activity in the NE foreland area of the Holy Cross Mts. are indubitable. The causes are, however, unknown. It may possibly be due to some tectonic movements that have not as yet been investigated. This hypothesis is not in principle contradictory with the marked increase of the erosional energy.

S. Z. Różycki is in favour of a tectonic interpretation. He thinks that after the recession of the Mindel glaciation uplift occurred within the Pomerania-Kujawy-Holy Cross Mts. anticlinorium.

So far there is no currently accepted interpretation of the data obtained from the investigations concerning the erosional activity in the Mindel/Riss interglacial period. The problem as to whether this phenomenon only occurred regionally or on a much wider scale is still an open question. Its solution calls for additional systematic studies in different regions.

The covering of erosive-denudational relief from the Mindel/Riss interglacial

Thick fluvial series, occasionally of the lacustrine type obscure recently developed forms in interglacial relief. Records of this accumulation are visible in every valley within the Polish Uplands. The transgression of the Riss glaciation is preceded by the deposition in the valleys of fluvioglacial sediments in which the country rock plays a dominant role. This is reliable evidence that the morainic cover of the Mindel glaciation had been torn away in a period preceding the Riss glaciation. The uncovering of the sub-Quaternary surface may have followed not directly after the retreat of the Mindel ice sheet but during the break between glaciations owing to the erosional and denudational activity of the Mindel/Riss interglacial period. These processes first removed the morainic material and then attacked the exhumed rock forms.

In the NE foreland area of the Holy Cross Mts. more than one morainic horizon was left by the Riss glaciation. In the north they are more numerous than in the south [12]. The lower horizon is mostly outwashed. The differences observed in the number of clay horizons probably suggest a more advanced degree of denudation in areas farther north, particularly in those situated at considerably higher levels. Neither should we exclude a longer duration of the Riss glaciation or greater wealth of events within the northern peripheries as compared with the southern part, most likely including the whole complex process of deglaciation. This problem, scarcely at all investigated in areas under the Riss glaciation, calls for additional studies. The large number of the Riss morainic boulder clay horizons which have been observed at the foot of the Upland where the bedrock reaches deeper down, need not be interpreted as being solely the result of the many phases of this glaciation. A more accurate analysis of profiles shows a greater amount of boulder clays in depressions of the topographic surface. This suggests fossil evidence of the processes of deglaciation. Speculation concerning the extent of the successive phases of the Riss glaciation should examine this fact more closely.

The Riss glaciation produced a somewhat different pattern of relief evolution. The occurrence of thicker series of glacial deposits will be a decisive factor in the formation of a new network of valleys, initiated by erosion of the Riss/Würm interglacial period, and it will continually mask the older relief of Pleistocene age here.

Changes in relief during the Riss/Wurm interglacial period

The period of denudation and erosion of the Riss morainic boulder clays had already started during the kataglacial recession of this glaciation and passed into the erosion of the Riss/Würm interglacial. On elevated highland areas morainic boulder clay was then removed from the valleys. The streams returned to their former depressions while on the northern slopes of the Highland new valleys were

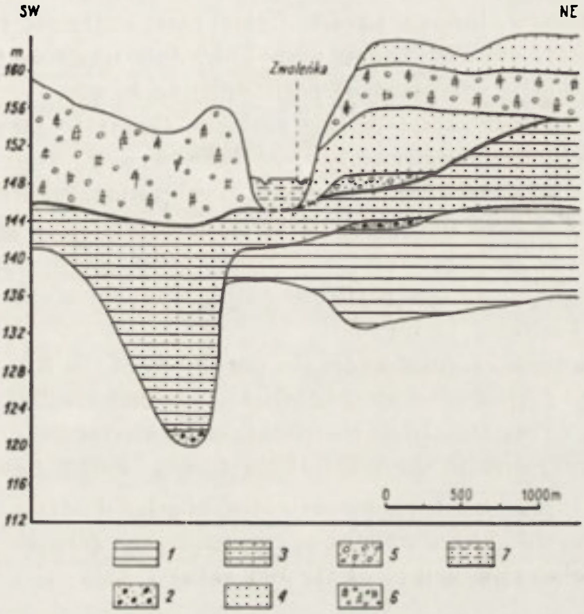

Fig. 5. An example of valley evolution during the Pleistocene

1 — preglacial deposits, 2 — outwash mateiial from the Mindel moraine, 3 — fluvial deposits prior to the Riss period, 5 — residuum of the Riss moraine, 6 — bulder clay of the Riss moraine, 7 — silty-sand deposits accumulated in the valley incised during the Riss/Würm interglacial

formed not connected with the previous valleys. This would suggest a strong planation of the terrain and the complete burial of valley forms that existed there prior to the Riss glaciation.

In valleys which developed along the same lines not only the morainic cover but also the sediments on which it rests were dissected and removed by interglacial erosion. Very typical truncation surfaces are then formed on valley slopes.

The depth reached by this erosion is indicated foremost by newly formed valleys (Fig. 3). They are deeper than the present valleys. A similar situation has been observed in the gap valley of the Vistula [9], and in the Lublin Plateau [6]. Nevertheless erosion of the Riss/Würm interglacial never reached the dimensions attained during the previous interglacial period.

The main problem connected with this interglacial lies in changes of the drainage pattern. Alterations in the valley pattern and the formation in new places of erosion-denudation bases indicate a different trend in the development of the relief.

The attached sections show several deep fossil valleys which were not exhumed either during the Riss/Würm interglacial period (Fig. 3) or in more recent times. Obviously, at the beginning of interglacial erosion the terrain was so completely covered by the Riss moraine that streams did not find their old depressions. Hence valleys were formed in new places, occasionally even on the watersheds of the fossil relief. Sometimes the new streams follow parts of the old river course so that new depressions are cut out in some parts only. As a result we have valleys whose longitudinal section is composed of parts differing in age.

There is still another aspect of this problem. Owing to the fact that material from the Mindel/Riss interglacial has not been removed from deep valleys and that valleys have been incised in new places, two individual reliefs, both Pleistocene in age, are superimposed one on the other. The Mindel/Riss and the Riss/Würm interglacial periods each produces a different relief, partly independent of one another. The evolution of one particular valley throughout the individual Pleistocene stages is illustrated in Fig. 5.

The style of the relief formed under the conditions of the Riss/Würm interglacial period gradually deepens the morphological differences within the NE foreland area of the Holy Cross Mts. In its morphological character the northern peripheries of this area come closer to the relief of the Central Polish Lowlands [12].

Important changes also took place in the Silesian Plateau. "The network of valleys formed on the surface which was covered by the deposits of the Riss glaciation does not follow the course of the old valleys" [5].

Further modifications in the pattern of relief

During the Würm glaciation, both in the area here considered and in other areas previously involved by the Riss glaciation, morphogenetic processes, connected with cold climate, are responsible for partial modifications in the relief [1, 2, 4, 5, 16]. They are expressed by the denudation of the upland area and by the accumulation of various slope deposits on the valley sides where they are interbedded with fluvial material. Today the accumulation of Würm deposits forms the flood plain terrace. The appearance of the valleys changed as they were filled in. Periglacial processes gradually obscured the older elements of the topography of the valleys [10, 11, 12].

The character of the relief in this part of Poland was not influenced to any substantial extent by Holocene morphogenesis.

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GEOGRAPHIA POLONICA 6,1956

ON THE REGULARITIES IN COMPOSITION AND MODE OF ORIGIN OF KAME TERRACES IN THE NORTH-WESTERN PART OF THE MAŁOPOLSKA UPLAND

Kazimierz Klimek

Kame terraces which have been recognized as long ago as the close of the 19th century are common features in areas where the waning ice-masses lay on a rock-floor of varied relief. Kame terraces occurring at different levels are best preserved in regions that had an inland-ice-cover during the last glaciation. Kame terraces are known from North America [5, 6], the British Isles [2, 13] and the Baltic region [12]. They also are evidenced in the foreland of existing glaciers [10, 14].

In Poland kame terraces are found in its northern part which was occupied by the Scandinavian ice-sheet during the last glaciation [1, 15, 16]. They also occur within the limits of the earlier glaciations in south Poland [3, 7, 8, 9, 11, 17]. In the mountains and uplands of south Poland that were covered by the ice-sheet during the earlier glaciations abundant kame terraces have been formed but only a few of them survived because they were subjected to heavy modification during the Upper Pleistocene. In South Poland kame terraces are found only in the Sudety Mts [4, 7, 9, 17] and throughout the Małopolska Upland [3, 7, 11]. Little attention has up to now been paid to the recognition of the composition of kame terraces and to the precise determination of their modes of origin depending on the local orographic conditions. The paper by A. Jahn [9] forms an exception.

The north-western part of the Małopolska Upland lies within the Silesian-Cracow monoclinorium consisting of Mesozoic sedimentary rocks inclined generally north-eastwards at 1-3°. There occur several structural escarpments separated by great subsequent depressions (vales). Detailed analyses of the composition and distribution of kame terraces have been made in the Upper Warta and Upper Liswarta stream basins which are sited upon the Jurassic rocks.

The area discussed was twice covered by the Scandinavian inland ice i.e. during the Cracovian (Mindel) and Middle Polish (Riss) glaciations. During the second glaciation the ice-margin lay against the northern slopes of the Małopolska Upland. Only lobes intruded into the depressions. The area studied was then lying in the ice-marginal zone.

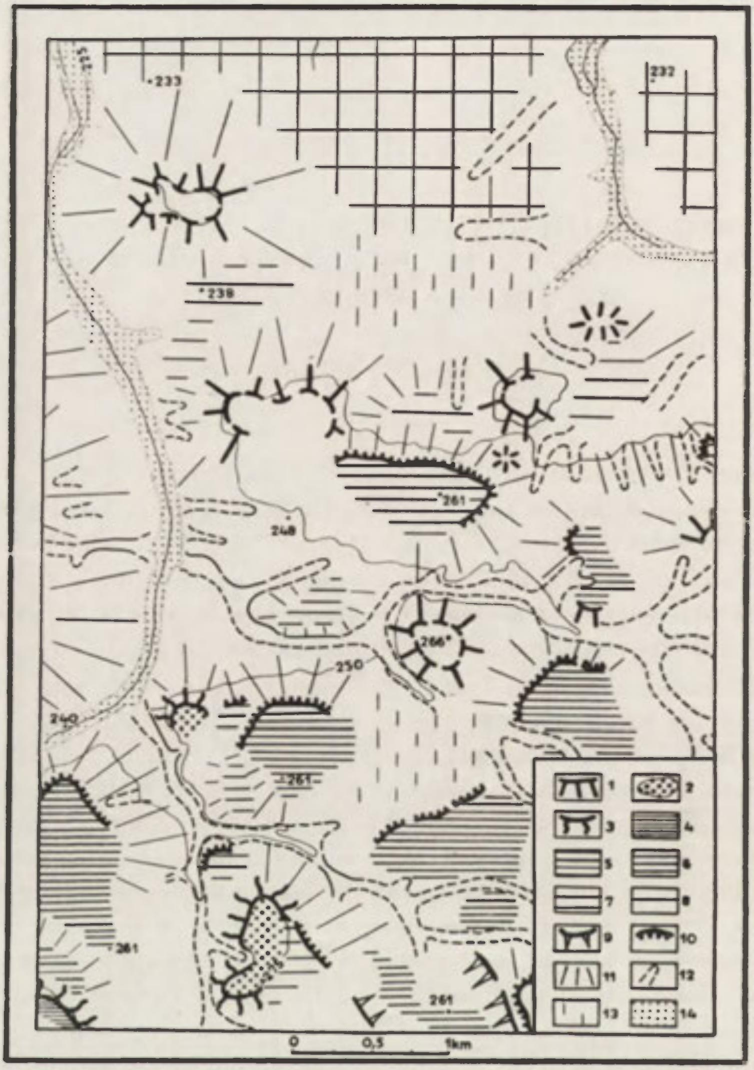

Fig. 1. Sketch to show the geomorphology of the hinterland of Upper Jurassic Escarpment

1 — hills composed of pre-Quaternary bedrock, 2 — ridges of depositional moraines, 3 — ice-contact slopes, 4-8 — kame-terrace plains, 9 — channel-sides due to proglacial water erosion, 10 — edges of kame-terraces (badly preserved), 11 — erosive-denudative sides, 12 — trough-like valleys, 13 — erosive-denudative plains, 14 — valley floors

In the area considered there prevailed the areal deglaciation. The varied rock floor has been favourable for dead ice formation associated with the production of a characteristic group of glacial land-forms including kames and kame terraces. There occurs evidence of certain regularities in the composition and the distribution of these forms, especially of kame terraces.

At an early stage of the deglaciation the escarpments first emerged above the surface of the ice whereas the subsequent vales were filled with dead ice. Between the ice-edges and ice-free hill-sides narrow sandy plains were built by the pro-glacial and extra-glacial waters. These plains were left as kame terraces after melting of the ice.

Both the plains of kame terraces and the related sandy plains resulting from the destruction of the former occur as flats a few kilometres long on the older hill-sides. They may be found on the scarp faces and in the escarpments hinterland. For the most part kame terraces and sandy plains occur one above the other at four or five levels (Fig. 1) situated between heights of about 10 and 65 m, on sides of the subsequent vales whose floors are infilled with a ground moraine. Kame terraces vary in width up to 0·2-2 km. The terrace surfaces rise above each other at 8-12 m intervals. The slope of these terraces along their length does not exceed 3 ‰. Terrace surfaces markedly diverge from each other where the escarpments are dissected by valleys.

Kame terraces that are found in the north-western part of the Małopolska Upland are composed mostly of stratified sand, less frequently of silt and gravel. These are flanked by occasional small morainic mounds up to 8-10 m high. Most of the kame terrace slopes may be recognized as ice-contact faces. Most of the

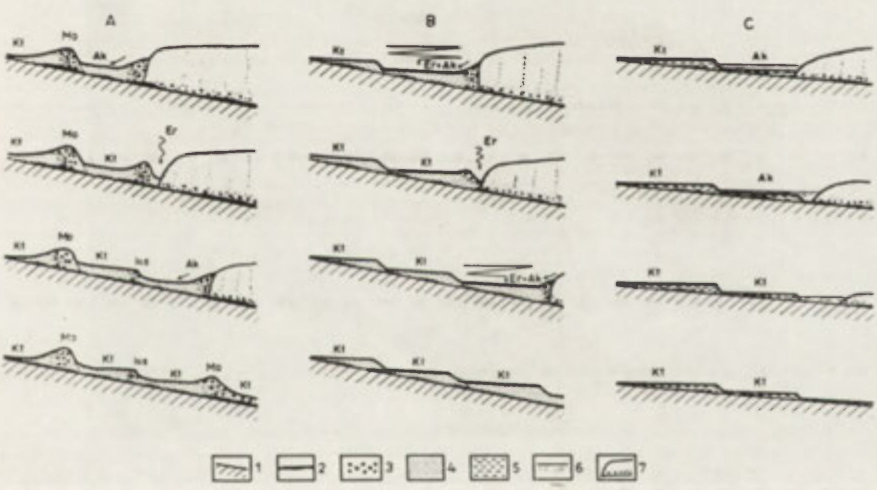

Fig. 2. Scheme of kame-terrace formation in the north-western part of the Małopolska Upland

1 — sub-drift surface, 2 — ground-moraine, 3 — gravel 4 — sands, 5 — silt, 6 — lake-level, 7 — ice, Kt — kame-terrace, Mo — moraine, Ics — ice-contact slope, Ak — accumulation, Er — Erosion

kame terrace slopes are, however, erosive-denudative faces. But this seeming changeability in composition and development of the kame terraces is strictly dependent upon the place and conditions of formation of the terrace cover. On this basis the present author recognized three distinct types of kame terraces being characteristic of the north-western part of the Małopolska Upland.

The first type (Fig. 2 A) includes terraces composed of either fluvioglacial or limniglacial deposits. Characteristic features are ice-contact slopes and mounds of depositional moraines occurring at their edges. Those kame terraces are found mostly on the sides of the small basins and of marginal resesses of the escarpments. For the most part the level of these basin floors is above that of the subsequent vales. During the deglaciation these depressions were infilled with secondary lobes of the stagnant or dead ice. As a consequence of ablation small morainic mounds were accumulated against the ice-edges. The scanty and, therefore, less powerful pro-glacial streams produced by melting of these small lobes washed out the ablation material from the small sandy plain resting between the distal slope of the morainic mound and the basin sides. Periods of warmer climate with increased ablation involved the rapid fall of the accumulation base. As a consequence the meltwaters began to flow between the ice-edge and the proximal slope of the morainic mound referable to an earlier phase. During a repeated cooling of climate the successively lower and younger sandy plain was produced by the meltwaters. The erosive power of these was, however, too small to destroy completely the earlier mound. They undercut it only. The fluvioglacial deposits were left as narrow kame terraces with

Fig. 3. Mstów — Excavation in the edge of a kame-terrace. Photo K. Klimek

ice-contact slopes or flanked by a morainic mound when the ice has completely melted. Small lakes were impounded along the ice-lobe where the glacial drainage was blocked by rising ground. These lakes were filled with silt and sand which reveal a very well assortment and a varve-like bedding (Fig. 3, 4).

The second type of kame terraces (Fig. 2B) includes terraces consisting of fluvioglacial deposits. These terraces are abundant in the area studied and may be followed

Fig. 4. Mstów — Structure of limniglacial deposits forming the kame-terrace. Photo K. Klimek

along the slightly dissected scarp facesfoe several kilometres. These terraces with erosive-denudative edges are much wrdir than those of the preceding type. During the cool periods with relatively small ablation very slow retreat of the ice-masses that occupied the subsequent vales has occurred. The channels between the ice-edges and the scarp faces carried both pro-glacial waters produced by ice-melting and extra-glacial waters draining from the already ice-free parts of the Silesian and Cracow Uplands. These abundant and powerful waters which took on the form of a marginal river had flowed along the icemargin for considerable distances. They eroded and transported the material melted from the ice and, therefore, no morainic mound has been built at the ice-margin. On the other hand, these marginal streams undercut the slopes of the higher kame terrace and destroyed the ice-contact slope. In this way an extensive sandy plain has been formed along the ice-margin. Periods of increased ablation involved the rapid fall of the accumulation base. As a consequence the marginal streams followed the retreating ice-edge and the sandy plain referable to an earlier phase of the deglaciation was left as a kame terrace without a morainic mound at its slope. The ice-contact slope was easily destroyed by the streams that were building the successively lower sandy plain. The composition of the fluvioglacial deposits which form the cover of those kame terraces is infinitely variable in the vertical profile (Fig. 5, 6).

Fig. 5. Wyczerpy — Structure of fluvioglacial deposits forming the kame-terrace.
Photo K. Klimek

Fig. 6. Grabowka — Structure of fluvioglacial deposits forming the kame-terrace.
Photo K. Klimek

The third type of kame terraces (Fig. 2C) includes terraces developed on the basin-sides and slopes of valleys that dissected the floors of subsequent depressions. These kame terraces are composed entirely of limniglacial deposits showing a well assortment. At a final stage of deglaciation of this area only small detached masses of ice occupied the lowermost depressions and thus blocked the free extra-glacial drainage. At the margins of the detached ice masses conditions were favourable for the formation of small pro-glacial lakes. These were filled with materials brought by the extra-glacial waters. Low kame terraces were formed on the sides of basins and of depressions when the dead-ice has completely melted.

Hence it follows that kame terraces occurring in the north-western part of the Małopolska Upland vary in composition (fluvioglacial and limniglacial deposits). They have been formed in different ways under different conditions. This kame terrace formation was a function of the irregularities in the sub-glacial surface and of the climatic conditions.

The varied relief of the bedrock conditioned: (a) the hydrodynamic regimen prevailing at the melting ice-margin, (b) the character of morphological processes leading to the formation of kame terrace covers, and (c) the width and the gradient for the length of kame terraces. The climatic conditions which were changing during the deglaciation determined: (a) the rate of morphological processes, together with (b) the climatic changes and their frequency. These controlled the retreat stages of the ice-edges and formation of kame terraces at different levels.

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GEOGRAPHIA POLONICA 6,1956

DEVELOPMENT CONDITIONS OF THE RELIEF OF LOESS AREAS IN EAST-MIDDLE EUROPE

Henryk Maruszczak

1. Introduction

The presented here analysis of development conditions of the loess areas relief is mainly based on the results of the author's own investigations in Poland and on comparative studies carried out in Czechoslovakia, Hungary, Rumania and Bulgaria. The loess in this part of Europe dates mainly from the last glaciation. Older loessy sediments are, as a rule, hidden under a cover of upper loess and occur in much smaller amounts. Owing to this fact the absolute age of the relief connected with loess is about the same throughout the whole studied area. This fact makes analysis easy.

The considerable extent of the mentioned areas in the meridional direction — 8° i.e. nearly 900 km — gives the opportunity of comparative estimation of the developmental conditions of relief (Fig. 1). In principle the author confined himself to the comparison of extreme areas, namely those lying in the upper Vistula basin (Poland) and in the basin of the lower Danube (Rumania and Bulgaria). The former are presently characterized by a transitional climate (from maritime to continental climate) and belong to the mixed forests belt of the moderate zone; in the period of loess accumulation ıhey are a part of the periglacial tundra zone. The latter constituted a part of the moderate steppe and forest-steppe zone at the time of loess accumulation and after its end [14]. The loess areas of Bohemia and Moravia represent, in this respect, a type related to that of the Vistula basin, while the loess of south-east Slovakia and of Hungary resemble the type characteristic for the lower Danube basin.

2. Characteristic relief forms of loess areas

In 1961 the author distinguished two groups among relief forms of Polish loess areas: 1. larger forms conditioned by the configuration of the loess covers substratum, or maybe connected with the distributional character of these covers, and 2. smaller forms, sculptured in loess covers [12].

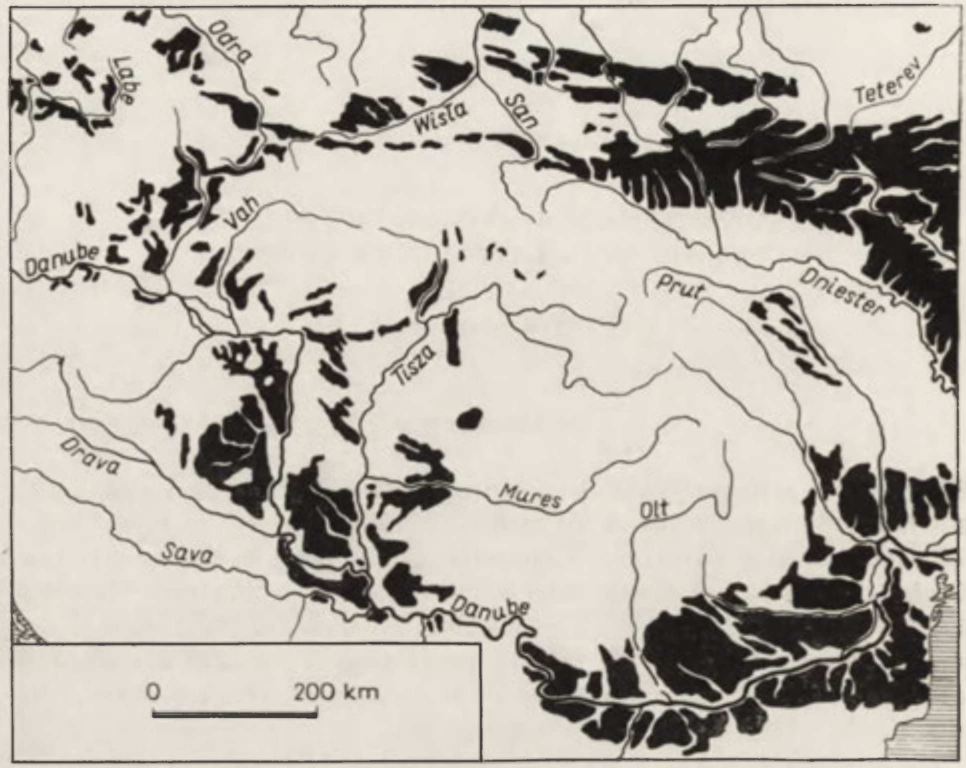

Fig. 1. Loess distribution in the eastern part of Middle Europe. Prepared by H. Maruszczak according to maps drawn by E. Rühle and M. Sokołowska, D. Jaranoff, P. K. Zamoriy and G. M. Molavko, G. Raileanu et al., J. Sekyra, M. Pécsi

Larger forms. In the upper Vistula basin loess occurs mainly in highland areas, where relative altitudes usually exceed 30-40 m, and even 100 m. The loess thickness reaches generally about 10 m and exceeds 20 m only exceptionally. That is why the cover surface of these deposits reflects all the larger forms of substratum relief and particularly valleys and escarpments. Such forms cannot be considered as determining the type of loess relief. But beside them edges of loessy lobes should be distinguished as characteristic (Fig. 2). A. Jahn has noticed that they are closely connected with the pecularities of the loess distribution [8]. Such edges are completely independent of the substratum configuration and they often cross high situated flat surfaces of watershed zones [8, 12]. Some of them have been formed almost exclusively by the particular system of forces controlling the dust accumulation, others were at least partly remodelled by later action of destructive forces.

Smaller forms might be divided into: a) forms developed during the loess accumulation period (trough-like valleys and small closed depressions), b) forms modelled after the accumulation ended (dry valleys, recent erosion cuts, kettles, pits, and

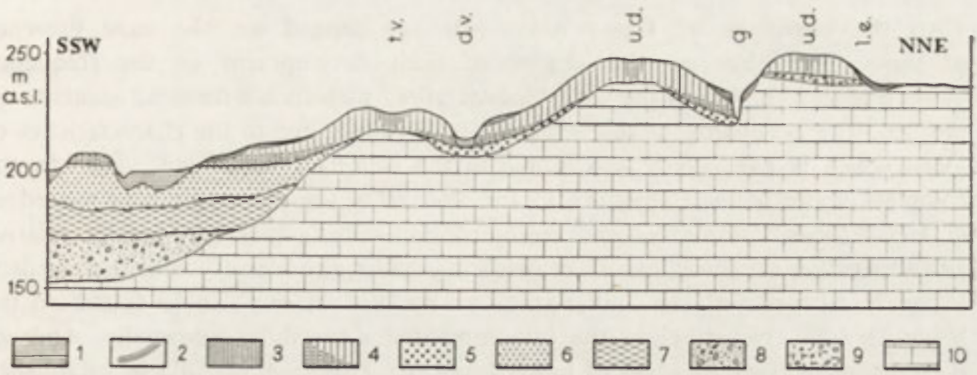

Fig. 2. Schematic cross-section of a loess lobe in the western part of the Lublin Upland (after H. Maruszczak)

1 — mud alluvium of valley bottoms (Holocene), 2 — slope sediments (Holocene and late glaciaı stage), 3 — decalcified and clayey loess (decalcification in the Holocene and late glacial stage), 4 — loess and loess with sand inter-beddings (last glaciation), 5 — sheet-wash and solifluction sediments (last glaciation), 6 — fluvial sands (last glaciation and last interglacial), 7 — sandy muds (Middle Pleistocene), 8 — gravel and sandy fluvial of fluvioglacial sediments (Middle and Lower Pleistocene), 9 — boulder clay (Middle or Lower Pleistocene), 10 — calcareous bedrock (Cretaceous), u.d. — basin-like undrained depressions, t.v. — trough-like valleys, d.v. — dry valleys, g — gullies, l.e. — loess edges

antropogenic forms) (Fig. 2). A detailed description of these forms was presented by the author in 1961 [12] who gives here only a short comparison of their most important characteristics. Trough-like valleys were formed through the operation of solifluxion, nivation and ablation; they represent a type of forms called "dellen". Closed depressions (pans) developed mainly as a result of unequal dust accumulation. However, in the course of this accumulation and particularly after it had ended they were strongly modified and deepened. This was due to the leaching of carbonates and the subsidence of porous sediment resulting from the infiltration of water gathering in depressions (suffosion processes). Such depressions known in Polish literature as *wymoki* represent the type of forms determined by Russian authors as *stiepnyie bludtsa* or, less frequently, as *pody*. Dry valleys differ from troughs by a rather distinct bottom and high inclination of slopes. In the initial stage they were formed by deep erosion and probably resembled the present-day gullies. Only after the remodelling of slopes by washing, solifluction and mass movements did they acquire a form approaching their present shape. Recent erosion cuts (gullies) belong to forms now arising, fast developing and creating in some places strongly branched systems. They are mainly formed by periodical streams. Gullies are closely connected with small kettles and pits formed as a result of the subterranean erosion development (mechanical suffosion) and of evorsion. Among antropogenic forms the greatest role is played by road ravines and larger exploitation pits.

It should be stressed that all the smaller forms mentioned above may also be found beside loess covers, on the substratum of other deposits. But there, as a rule, they are less developed and occur much more seldom or even only sporadically.

Thus the character of loessy areas does not depend on the mere presence of these forms but on the degree of their development or the frequency of occurrence. The use of the term "loessy relief" with such a meaning seems to be justified. The peculiarity of this sculpture is obviously due to the characteristics of loess, which is particularly susceptible to the action of water and wind. In the complex of the forms characteristic for that relief the main role play the edges of loessy lobes, dry valleys and recent erosion cuts (gullies). At greater relative altitudes gullies are more important, while at smaller altitudes dry valleys come first.

Nearly all forms of loess relief areas on the Vistula can also be found on the lower Danube. Nevertheless, the two compared areas differ essentially. And so, if larger forms, such as edges of loessy lobes, are discussed, it could be established that they rather do not occur on the Danube. Instead, the peculiar loess humps, sorted out by Bulgarian authors can be classified among large forms. They extend mainly along the high right bank of the Danube and rise above the principal surface of loessy plateaux adjoining this bank [9, 16, 17]. According to the unanimous interpretation of the named authors these are forms of loess accumulation, not dependent on the substratum (Fig. 3). Since their origin is connected with the dynamics of a dust accumulation they may be considered as a peculiar analogue of the loess edges of the Vistula basin. The author will try to justify this thesis in the next chapter.

Smaller forms of the Vistula loess relief are all found without exception on the lower Danube. Besides, Bulgarian authors also distinguish loess dunes and loess slides [9, 17]. It should be noted that beside sand dunes which lie on the loessy cover and are younger than the cover, there occur loessy dunes on the Danube which seem to constitute organic aggregates with loess [15]. However, their extension is rather limited [14, p. 44]. Loess slides on the other hand, represent secondary forms developed after the end of the loess accumulation. They are connected with the high banks of the lower Danube and its tributaries. But their origin is closely associated with the occurrence in the substratum of loesses of clayer Neogene sediments of the Ghetic Depression or of the Moesic Plateau (Cretaceous sediments) [9]. The characteristic features of loess and especially its great porosity and a tendency to form vertical cracks are favourable, however, to the development of such forms.

The presence of peculiar humps, dunes and loessy slides constitute one of the important moments determining the distinctive loess relief in the Danube areas as compared with those of the Vistula. This peculiarity is strongly reflected in the fact that the most characteristic of all forms of loess relief on the lower Danube banks are the closed depressions and not dry valleys or gullies. Those depressions are known in Rumanian literature as *crovuri* (sing. *crov*), and in Bulgarian works as *padini*, *panichishta* or simply — following Russian terminology — as *stepnyie bludtsa* [9, 17, 18, 19]. Bulgarian and Rumanian authors value so highly the role of these forms that they distinguish their occurrence areas even on geomorphological

Fig. 3. Loess humps along the high Bulgarian bank of the lower Danube (after Tz. Michailov)

A — Extension and thickness of loesses on the Danubian plain between the Isker and the Ogosta: 1 — 0-10 m (or 5-10 m), 2 — 10-20 m, 3 — 20-30 m, 4 — over 30 m, 5 — maximal recognized thickness of loess within individual humps, B — Cross-section along the line running on the east of Orekhovo, in a nearly meridional direction: 1 — loess cover, 2 — substratum of loess cover

maps, on scales smaller than 1:1,000,000 [7, 18]. As for dry valleys and gullies, these are nowhere so strongly developed as in Vistula region though in some other parts considerable relative altitudes and the loess thickness constituted very favourable conditions for the development of such forms [5].

The differences existing between the compared areas result in the first place from the different geographical conditions of relief development during the loess accumulation and after its end.

3. Conditions of relief development during loess accumulation in the Upper Pleistocene

At the time of loess accumulation, the upper part of the Vistula basin was within the compass of the tundra and forest-tundra periglacial zone. The character of processes, controlling dust accumulation was of particular importance to the development of the edges of loessy lobes. In 1963 the author summarized his arguments, proving that the dust was mainly deposited in places where orographic obstacles impaired the velocity of the wind [13]. By assuming such a thesis we get the best explanation of the absence of loesses in plains and their connection with strongly sculptured highland areas. Other arguments prove that the prevailing amount of the dust was accumulated — as suggested by A. Malicki [11] — owing to transport in low layers of the atmosphere and on a small distance. This kind of transport and the substratum relief were decisive for the unequal distribution of loess. It was accumulated in lobes extending in the direction of prevailing winds (i.e. mainly east winds, and near the end of loess accumulation — west winds [13]). Accumulation edges of such lobes probably developed in places where disturbances occurred in the system of atmospheric currents [8, pp. 447-448].

The distribution of loess in the form of separate lobes has also been observed in other regions. B. A. Fedorovich has established that along the border of Asiatic moderate deserts loess occurs as "manes" resembling longitudinal dunes, or as large "beds" [4]. The beds reach 30-50 km of width and are divided by 15 km wide depressions filled with sand. They extend in the direction of the prevailing winds and have clear-cut, almost straight-lined edges. The mentioned author associates their origin with the turbulent system of wind streams, conditioned by orographic obstacles.

The conditions in which undrained depressions and troughs developed may be much easier and more exactly determined. Original forms of undrained depressions were mainly modelled by uneven dust accumulation. This occurred primarily within the terrace and near watershed levels characterized by small surface inclination. Since highland areas where loess accumulation occurred were rather strongly dissected, the extension of such levels and hence of depressions was limited [12]. Instead, these relief features favoured the development of a dense network of trough-like valleys formed by solifluction, ablation and nivation processes characteristic for the periglacial zone.

At the time of loess accumulation the Danubian areas lay in the forest-steppe and steppe zone. Dynamical conditions of dust accumulation were different here than on the Vistula. It is clearly proved by the loess distribution in the form of a rather close cover, mainly connected with the plains lying on both sides of the lower Danube. Since accumulation occurred there on plains, it may be concluded that the velocity of winds transporting the dust was smaller than on the Vistula [14]. Besides, it may be supposed that the land features did not favour the arising of stronger disturbances in the system of atmospheric currents. The above emphasized features of atmospherical dynamics enable to explain the compactness of the loess cover and the resulting absence of loessy edges. Only in places where, in the immediate vicinity of the dust source (alluvium of the Danube valley), there existed distinct orographic obstacles, increased accumulation occurred (Fig. 3). Thus mighty loess humps were formed along the high, Bulgarian banks of the Danube. The largest humps were mainly formed as a result of accumulation by winds parallel to the river bank [14]. That is why it may be supposed that the dynamics of atmospheric currents causing the formation of humps was the same as of the winds forming edges on the Vistula basin.

Other relief forms of the Danubian regions synchronous with loess accumulation were also modelled by processes resembling those on the Vistula. Thus undrained depressions were mainly caused by an uneven dust accumulation. It is probably due to plain relief rather than to accumulation peculiarities that the role of that process was more important here than on the Vistula. For these regions are characterized by a large part of flat surfaces, and depressions within them were left outside the draining system (Fig. 3). At the same time these relief features were unfavourable to the formation of trough-like valleys. Among processes which formed such valleys in moderate climatic conditions ablation came first. But its results were limited for the quantity of water flowing off superficially was relatively small owing to the plain relief and the lack of permafrost in the substratum.

4. Conditions of relief development after the end of loess accumulation

The comparison of forms sculptured in loessy covers after the end of dust accumulation shows that processes of linear (concentrated) erosion played a main part in their formation. The development of these processes resulted in the formation of dry valleys and recent erosion cuts.

Dry valleys cutting the loessy cover were formed in the Vistula basin mainly at the end of the last glacial period and at the beginning of the Holocene [12]. It was a transition period when periglacial climatic conditions changed into a moderate climate. The disappearance of the permafrost and the increase of humidity favoured the development of concentrated erosion and thus the ravine-type forms. However, those forms underwent a rapid evolution, especially in periods of recurring cold. During such periods solifluction and ablation processes transformed the steep slopes of ravines. Thus a dense network of dry valleys was formed. Their development was set back in the Holocene when rich forest vegetation prevailed in loessy regions [12]. On the Danubian plains no radical change of climatic conditions occurred after the end of loess accumulation, because, in spite of growing warmth and humidity, the reciprocal relation of temperature and atmospheric precipitation remained the same. This was best reflected in the fact that in the Holocene those areas still belonged to the forest-steppe and moderate steppe zone. There was no transition period particularly favourable for the formation of dry valleys. That is why the network of these forms is much less dense here than in the Vistula basin.

Recent erosion cuts (gullies) arise currently and are swiftly developing, which makes analysis of their developmental conditions easy. An essential condition of their appearance was the destruction of natural vegetation and its replacing with the now dominating cultivated plants. The natural forest and steppe vegetation protects the loess relatively well against the destructive action of water erosion. The possible differences in the protective values of natural vegetation were however obliterated, when it was replaced by cultivated plants. The type of the latter is the same in both compared areas; it is characterized by a predominance of cereal cultures. In the forest zone, beside grain cultures, also root crops (potatoes, sugar beet) play an important role, intensifying the erosion. In the steppe zone a similar role is played by the culture of corn and sunflower.

Gullies of the Vistula basin may be considered as the most characteristic form of loess relief. Locally on loessy uplands their network reaches 8-9 km/km^2 and the average index in areas of 500 km^2 surface amounts to 2 km/km^2 [2]. In areas where the surface reaches several thousand km^2 the average index may amount to 0·5 km/km^2. The best developed gullies occur in areas where relative altitudes are above 60-70 m. They form then strongly branched systems in which the main forms attain a length of several kilometres. If the altitudes do not exceed 50-40 m gullies occur sporadically and do not form branched systems.

On the contrary, the Danubian gullies are usually short and only exceptionally form branched systems. They occur almost exclusively on the more steep and high slopes of larger river valleys [5]. The poor development of these forms is rather closely connected with the general features and geological structure of the area. A great part of plains in the lower Danube basin belongs to the Ghetic Depression which sank as early as in the Quaternary. That is why we find here vast, young and very weakly cut surfaces of the Pleistocene lacustrine accumulation. But beside them there are highlands belonging to the Moesic Plateau; their relief age and relative altitudes are the same as those of the Polish loessy uplands. However, in such uplands gullies are also poorly developed. It is not the relief then that can elucidate the different density of gullies in the areas compared. Those differences are primarily connected with hydroclimatic developmental conditions of the periodical stream erosion.

The Danubian loess areas are characterized by a drier climate than that of the Vistula region. The average yearly precipitation amounts there to about 500 mm and the temperature to 10-11°C. whereas on the Vistula region the respective figures are about 600 mm and 7-8°C. This is followed by a difference in the quantity of water drained off superficially. The yearly average index of the run-off within the Danube basin does not, as a rule, reach 1 l and even 0·5 l/sec./km^2, while in the Vistula basin it amounts to at least 3-4 l/sec./km^2 [3, 7, 20]. The quantitative relation of water drained off superficially in both areas is approximately 1:5. This relation must have been similar in the Pleistocene. This is best proved by the differences in the density of the valley network. In the Danubian loessy upland the index of this density only exceptionally reaches 2·0-2·5 km/km^2, whereas on the Vistula it rises up to 4·0-4·5 km/km^2. Within the loessy plains this index amounts to less than 0·5 km/km^2 on the Danube and to about 1 km/km^2 in the most poorly sculptured areas in the Vistula region. Hence the relation of the two indices for both compared areas is 1:2. It could be noted marginally that the poor density of the valley network is at least partly responsible for the weeker development of gullies in the former of the two areas.

The determination of the quantitative relation of water flowing off superficially (1:5) and of the valley network density (1:2) suggest further reflections. As there are no reasons to determine the index relation of gully network density, it will be assumed that it is the same as in the case of the valley network (1:2). The comparison of the quoted relations leads to the conclusion that the morphogenetic efficiency of a water unit flowing off superficially in the steppe zone is greater than in forests. It results undoubtedly from the fact that loess of the steppe zone is worse protected against the eroding action of water than loess from the forest zone. A decisive factor is the drying up of steppe vegetation during late summer and autumn. This concerns not only natural vegetation but also cultivated plants.

The development of gullies in the Danubian steppe can be then connected mainly with summer conditions. The frequency of abundant precipitation in the compared areas is, however, the same. It may be illustrated by the fact that the number of days with a daily precipitation higher than 10 mm reaches in both areas an average of 12-18 yearly. For summer and autumn months (August-October) the index is 4-6 days [18, 22]. Absolute extension values of summer erosion may then be similar in both areas. It is true that superficial drainage on the Danube comprises the smaller part of the precipitation but its effects are more important. Consequently a weaker development of the Danubian gullies than of those in the Vistula basin should be explained by developmental conditions of these forms in other seasons.

The period of winter and spring thaw is of decisive importance to the formation of gullies in Poland's loessy upland [12]. Instead, on the lower Danube the extent of erosion is much smaller in this season owing to: a) a shorter duration of the snow cover and a smaller reserve of stored water; b) a weaker development of seasonal frozen ground. The snow cover on the Danube banks remains to 30-60 days, and in the Vistula region 50-60 days [3, 7, 18]. The average temperatures of the coldest month which could be considered as the indirect index of ground congelation intensity are respectively from 0 to -3°C and from -2 to -5°C. Thus the steppe soil, less congelated and dried during the previous autumn season, is able to absorb much more thaw water than that of the forest zone. Consequently we have a lower drainage index and a limited erosion.

5. Conclusions

1. Owing to the specific qualities of loess, regions covered with it have a different relief from the surrounding areas which induces to apply the term "loess relief". This relief is characterized by a complex of the following main forms: a) larger forms (loessy edges and humps), and b) smaller forms (closed depressions — pans, trough-like valleys, dry valleys and gullies and, closely connected with the latter, pits or caved kettles). They can be found in almost all the studied parts of Europe, but the role of single forms changes according to the place.

2. Developmental conditions of larger forms were primarily determined by the dynamics of winds, transporting and accumulating the loessy dust. On the Vistula those winds had generally a greater velocity and were more differentiated according to the land features. Loess accumulation in the form of separate lobes with distinct edges (loessy edges) was connected with those winds. Instead, on the Danube the winds velocity was lower and their structure more regular. That is why the loess was more uniformly accumulated. Only along the high Danubian bank which caused more serious disturbances in the system of atmospheric currents, in the immediate vicinity of the dust source (alluvium) an exceptionally strong accumulation occurred. This caused the formation of loessy humps.

3. Among smaller forms the most characteristic of Poland's loessy regions — now belonging to the moderate mixed forest zone and at the time of loess accumulation to the periglacial tundra zone — are dry valleys and gullies. The network of dry valleys and gullies, i.e. of forms shaped mainly by erosion, is about twice as dense here as in regions belonging to the steppe zone. It is due to a relatively damp climate characteristic of which is a fairly high index of superficial drainage of atmospheric water. Owing to a strong dissection of the land the extent of flat surfaces (with low inclination) is limited and the spread of closed depressions, accompanying such surfaces — smaller.

4. In the loessy regions of Rumania and Bulgaria belonging presently, as well as during the loess accumulation, to the forest-steppe and moderate steppe zone, the most characteristic, beside peculiar loess humps, are closed depressions (*stiepnyie bludtsa*). A weaker development and a smaller density of the valley network (dry valleys and gullies) may be explained by a drier climate. Owing to a poorer land dissection the share of flat surfaces is here much larger. This favours the infiltration of atmospheric water deep into the loess cover and thus the development of chemical suffusion (leaching of carbonates) which forms closed depressions.

5. It can be thus stated that, under conditions modified by Man's economical activity, in areas belonging to the mixed forest zone and characterized by a transition from maritime to continental climate (Poland), the erosive intensity of intermittent streams (the so-called "gully erosion") is greater than in regions belonging to the forest-steppe and steppe zone, characterized by a moderate continental climate (plains of the lower Danube)

The quoted facts indicate that opinions of a very strong development of erosion in the moderate steppe zone are not justified, at least a generalization of such opinions onto the whole zone is unfounded. Among other authors such opinions were expressed by J. Tricart and by M. Gornung and D. Timofeev. In 1953 J. Tricart wrote that erosive cuttings of the gully type represent original forms, exactly corresponding to the climatic zone of the moderate steppe [21]. Instead M. Gornung and D. Timofeev, in a tabular setting of sculpturing processes in the main morpho-climatic zones, published in 1958, assumed that gully erosion throughout the globe is most richly developed in the moderate steppe zone [6]. These views are susceptible of criticism as: a) in this zone gullies do not represent a wide-spread form in natural conditions and developed largely only after the steppe had been brought under cultivation; b) the network of these forms may be much more strongly and largely developed — in conditions modified by Man's economical activity — in the mixed forest zone. That is why B. Kosov's opinion seems to be better grounded. Analysing in 1960 the intensity and developmental conditions of gully erosion in the natural zones of the Soviet Union he concluded that this process is best developed in the forest-steppe zone [10]. Undoubtedly similar conclusions were still earlier drawn by L. Berg. In his monograph on the geographical zones of the Soviet Union this author presented a comprehensive description of gullies in a chapter devoted to the forest-steppe zone [1], and in a chapter concerning steppes he characterized the closed depressions (*stiepnyie bludtsa*).

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MUTUAL RELATION BETWEEN LOESS AND DUNE ACCUMULATION IN SOUTHERN POLAND

Elżbieta Mycielska-Dowgiałło

It may be sometimes observed in the southern regions of Poland that dune accumulation invades summit surfaces covered with the loess mantle. This phenomenon has been noticed at several points of the southern area of the Sandomierz Upland (Fig. 1). Its southern part is formed of Cambrian rocks, covered on their SE side by Miocene formations. The thickness of pre-Quaternary rocks increases towards the South.

On the older substratum lie remnants of boulder clay, originating from the Mindel glaciation and from fluvioglacials of that and the following glaciation. The whole of this formation is covered by a thick mantle of loess (up to 30 m thick), within which may be distinguished 3 horizons, divided by 2 fossil soils (Brörup, Paudorf, [13]). The loess covering is dissected deep erosive valleys sometimes 30 m deep [13].

In the ceiling of loess profiles on flat watershed areas, intercalations of sand are fairly often detected. It appears from the analysis of grain rounding that they are eolian sands. It is characteristic that intercalations were not found anywhere in middle or lower parts of loess profiles. A series of outcrops were observed, in which the dunes invaded the loess surface (their lower layers are then indented into the loess) or eolian sands in the form of slight layers interstratified with loess. It occurs sometimes, that the dune form (from loess-sand material) is covered in its peripheral parts by decalcified loess dust. Within the upper plane between two tributaries of the Vistula — the Koprzywianka and the Gorzyczanka — these formations have a circular, irregular shape, or are elongated in a NW-SE direction.

In one of them of circular shape, a whole series of excavations and borings has been performed, of which four are presented here (Fig. 1, pt 1; Fig. 2). It may be clearly seen that eolian sands, lying on the loess, underlie in the peripheral parts of the formation a thin layer of decalcified upper-loess dust. The amount of loess intercalations increases generally towards the bottom.. They react strongly with hydrochloric acid.

The sands determined in the figure as stratified are as a rule of a light colour, with strongly rusty horizontal streaks. These are zones of intense washing of ferric

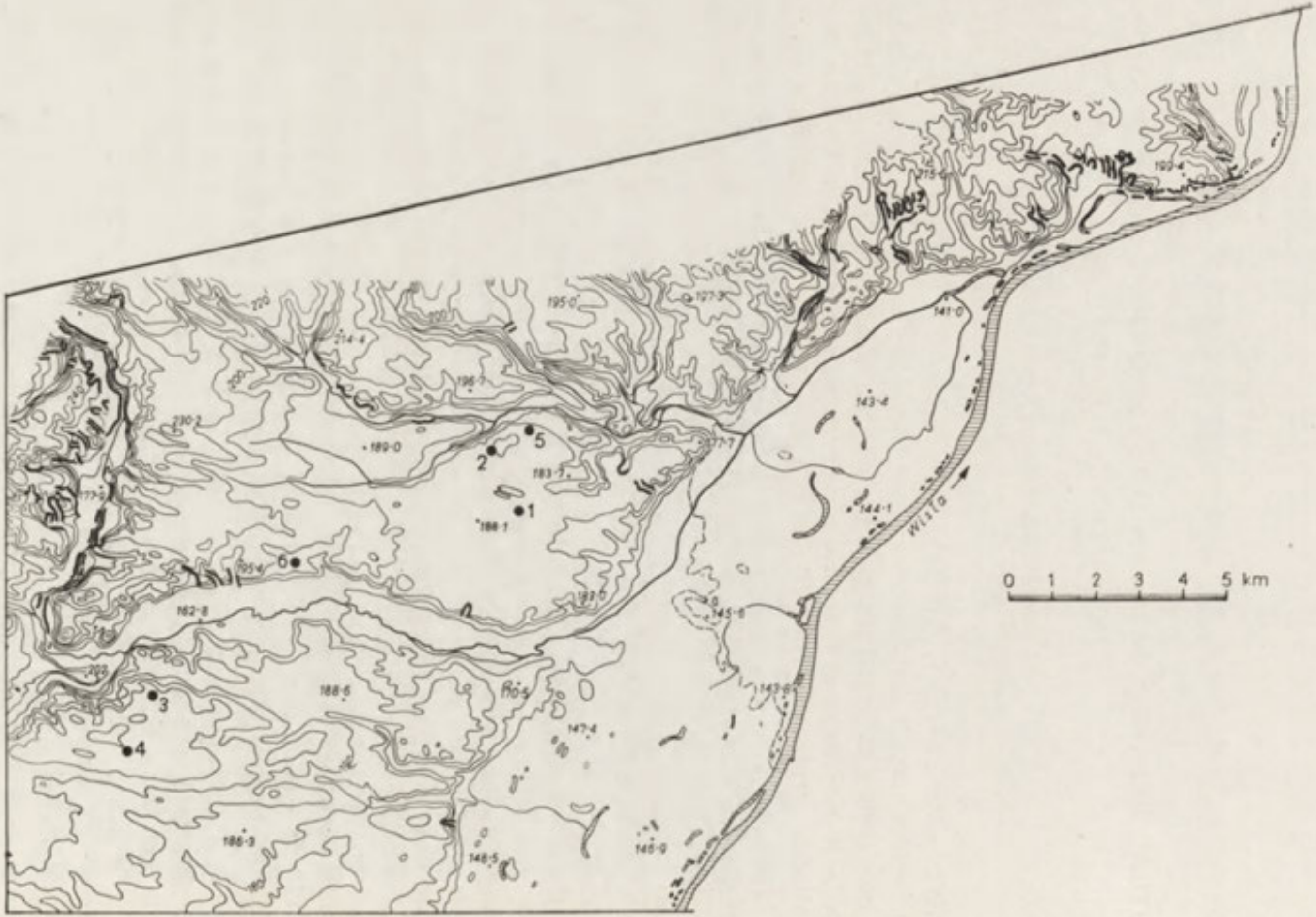

Fig. 1. Map of the investigated area; 1–6 — investigation points

compounds. Inside of convex formations, the ferruginous streaks reach almost up to the present surface (Fig. 2, excav. 1).

The analysis of rounding of sand grains in the mentioned formation presents a very high index, typical for eolian sands. The degree of rounding was measured by means of the method of Mihaltz, Ungar and David [3, 12].

Within the sandy loess layers traces of structures may be seen similar to periglacial ones. They particularly disturb the thin dust layers, whereas the sandy layers are sectioned by vertical or near-vertical fissures.

A similar profile is seen in the walls of a sand-pit in a wide, convex dune-hill, at a 1 km distance N from thefirst one (Fig. 1, pt 2). The amount of dust intercalations increases towards the bottom. However, the bottom of the sandy-loess zone has not been reached in spite of the depth of the open workings (4 m) from the surface. The thin dust layers react with hydrochloric acid. The thin layers of sand and loess dust are inclined to NE.

Another interesting profile may be seen in the Nietuja dune, south of the Koprzywianka valley (Fig. 1, pt 3; Fig. 3, 4, 5). Below a light coloured, close sand, sands may be found, cemented by ferrous compounds. Within this complex of layers traces of deep fissures and heaves are noticeable. It is difficult to establish whether they are of frost character, or merely a result of mechanical slides of part of the material. Below these layers which reach up to above 2 m under the present surface, the admixture of loess dusty material is increasing in the form of streaks.

Similarly as found by Kadar [6] in Hungary, rusty streaks may be connected with the infiltration of ferrous compounds from neighbouring soil. These compounds probably settled above the level of ground water. The waviness of thin loess layers should be linked with the eolian accumulation process. The rounding of sand grains, similarly as in the preceding cases is of a high degree and typically eolian. The whole described dune formation is of elongated shape from WNW to ESE, the wind-exposed side being slightly inclined WNW-wards.

A similar dune form may be found within the loess summit surface, but of less regular shape, to the west of the village of Jeziory, about one km south of the previously described form (Fig. 1, pt 4; Fig. 6). A whole series of excavations have been dug into that dune. Two of these excavations as the most characteristic ones have been presented here. The ceiling of the dune (excav. 1) is built of sand without interbedding, with only rusty, irregular stains. Lower, at a depth of 1·5 m occur dusty intercalations, the number of which increases downwards, quite distinctly. At a depth of 4·9 m, the boring pierced a thin humic layer. The author is inclined to consider it as a trace of fossil soil. Still lower pure loess is found, without admixture of sandy streaks. Within the thin sandy dust layers appear incidentally ferruginous streaks horizontally distributed.

In the second excavation, very much as in the first described form, sands from the dune phase lying on the loess form the substratum of decalcified, upper-loess dust. There are interesting traces of solifluxial flow in the bottom of this upper horizon. Within the light-colored sands, horizontally streaked, traces of fissures and wedges may be seen, supposedly of glacial origin. In this layer occur also the above mentioned ferruginous streaks.

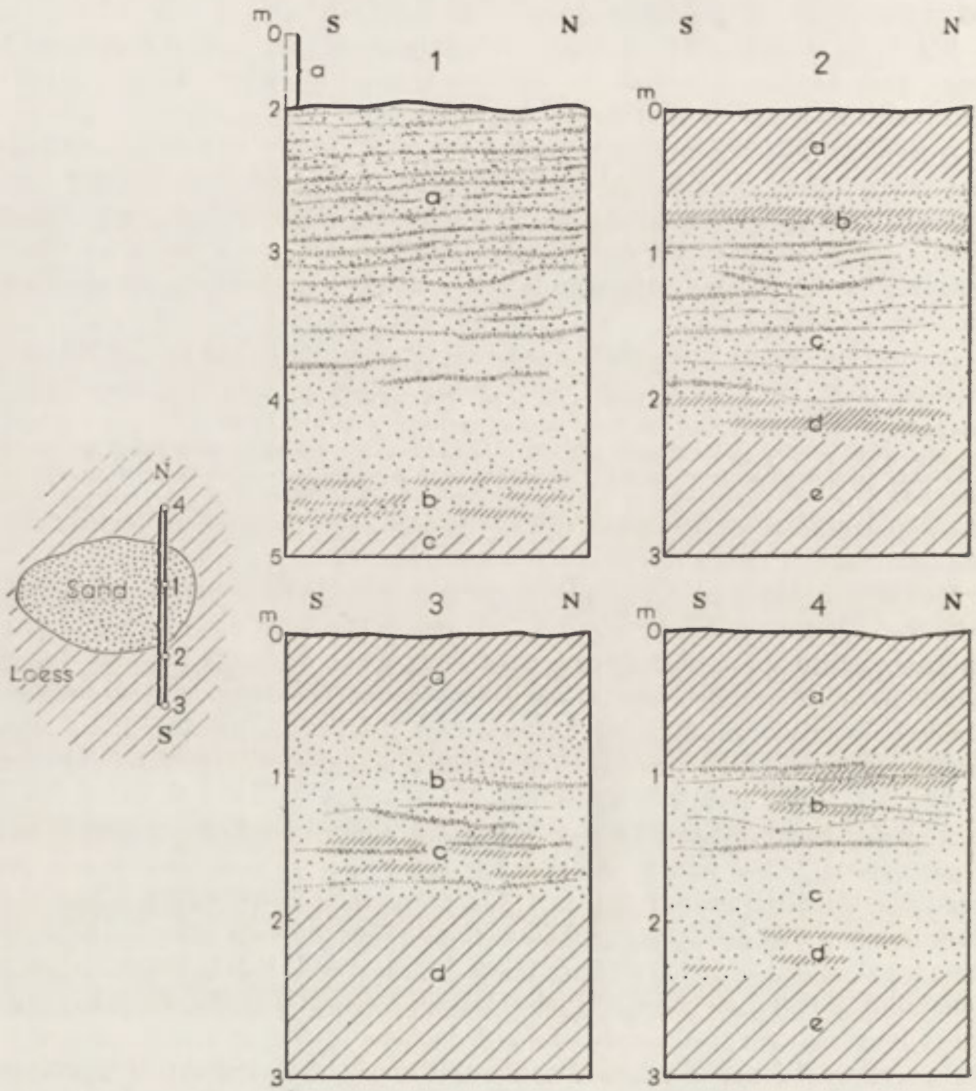

Fig. 2. Profiles of excavations in dune formation, in the watershed of Koprzywianka and Gorzyczanka rivers

Excav. 1: a — stratified sands, b — stratified sands with loess dust intercalations, c — loess, Excav. 2: a — decalcified loess dust, b — loess dust with sand intercalations, c — stratified sands, d — loess dust with sand intercalations, e — loess. Excav. 3: a — decalcified loess dust, b — stratified sands, c — sands with loess dust intercalation, d — loess, Excav. 4: a — decalcified loess dust, b — loess dust with sand intercalations, c — light-colored sands d — sands with loess intercalations, e — loess

Fig. 3. Dune formation in Nietuja. Photo E. Mycielska-Dowgiałło

Fig. 4. Transverse excavation within the Nietuja dune. In the upper part ferruginous layers are visible in the sand, at the bottom loess dust intercalations. Photo E. Mycielska-Dowgiałło

Fig. 5. Thin strata of loess dust in the lower part of the Nietuja disclosure. Photo E. Mycielska-Dowgiałło

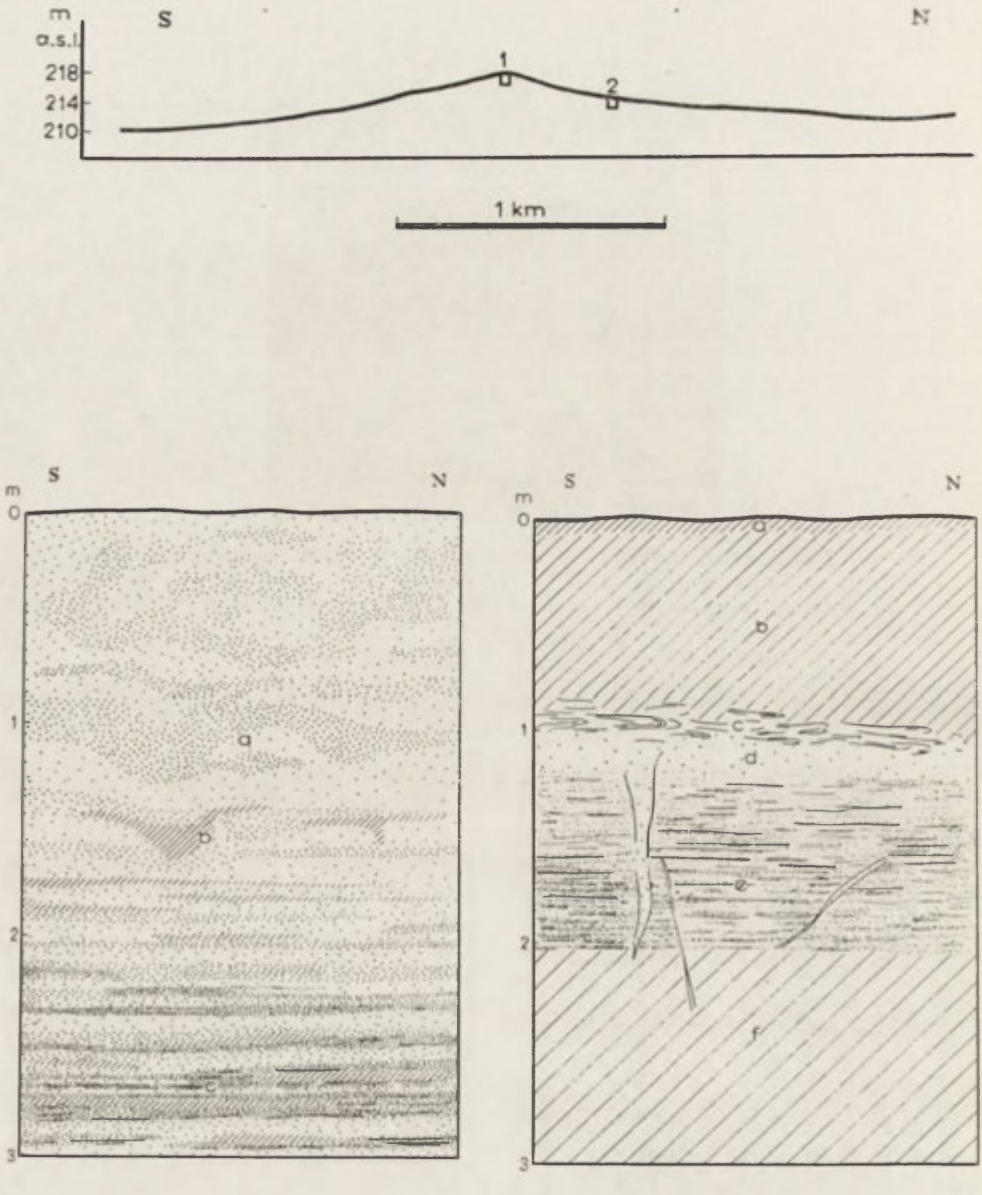

Fig. 6. Profile through the dune at Jeziory with cross-sections of two excavations

Excav. 1: a — variegated sands, b — dust intercalations in light-color and brown sands, c — rusty sands, with ferruginous intercalations, interstratified with dust and brown sand

Boring below excavation: 3·0-4·5 m — brown sand with dust intercalations, 4·5-4·9 m — grey-clayey dust, 4·9-5 m — remnants of fossil soil, 5-7·5 m — loess

Excav. 2: a — present humic soil, b — light-colored decalcified loess dust, c — solifluxal disturbances, including the bottom of the loess formation, d — light-colored, close sands, filling out the wedge and fissures, e — brown and light-colored stratified sands, with ferrous compound streaks, f — loess. Boring below excavation: 3-5 m — loess

A series of smaller or larger sand-islands could be noted also along the south loess edge. The profiles of excavations present here often in the ceiling loess streaks interbedded with sand of a high index of grain rounding and of thin layers arranged into wavy ripplemarks. A similar wave system may be noted on the north slope of the loess embankment, south of the Koprzywianka valley.

The wind blowing from the lowlands blew sands on to the loess which deposited on the slopes, where the wind lost its carrying power. The loess was not yet stabilized at the time, which may be seen from streaks of this material inside of the sands.

A noteworthy profile appears in the road tunnel, south of the Gorzyczanka valley, on a flat, wide summit plane. Above the loess are lying streaked sands, of high grain rounding, cut by small faults. They present ferruginous streaks. Still

Fig. 7. Road tunnel south of the Gorzyczanka valley
Photo K. Straszewska

A — light-colored, decalcified loess, B — streak of fossil humic soil filling deep pouches, C — light-colored sand, D — light-colored, streaked sands cut by numerous small faults, maybe of frost origin, with ferruginous thin strata, E — light-colored loess, reacting with hydrochloric acid

higher lies light-colored sand, covered by a stripe of fossil soil, which fills wide pouches. The present surface is formed of a layer of decalcified loess (Fig. 7).

The wall of the ravine near Zbigniewice (Fig. 1, pt 6; Fig. 8) presents another interesting profile. This disclosure occurs within the summit surface, directly above the Koprzywianka valley slope. Above the light grey, decalcified loess lies a complex of middle-grained, light sands, with ferruginous streaks and thin layers of dusty sand. This complex is sectioned by a deep form of glacial wedge, filled with light-colored, thick-grained sand. The rounding measurements of samples taken showed a high index of rounding.

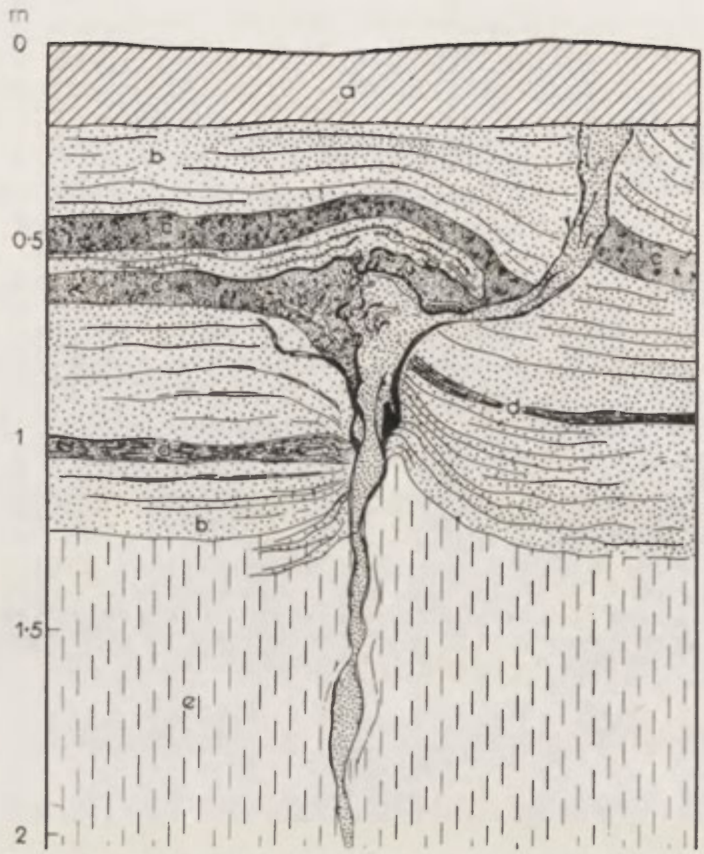

Fig. 8. Profile of the ravine near Zbigniewice

a — contemporaneous soil, b — light-colored, middle-grained sands, c — grey, dust sands, d — sand strongly cemented by ferrous compounds, e — decalcified, yellowish-grey loess

Observations performed in the Sandomierz Upland enable to suppose, that the eolian dust-sandy layers developed in climatic conditions different from loessogenic ones. The air temperature supposedly increased, which facilitated the summer drying of the sand and its blowing over, though the climate might have been still rough, which is witnessed by small glacial structures, especially within dusty layers, and the lack of vegetal covering or its insignificant expansion, which favoured a secondary overblowing of loess dusts.

Owing to the discovery in the Jeziory drilling of fossil soil at the border-line between loess and dust-sand layers (this soil supposedly originating from the Bölling), the layers would correspond to the earlier Dryas. At this period took place the first dune forming phase, distinguished by Chmielewska and Chmielewski from Witów [2]. According to Wasylikowa the earlier Dryas was characterized by poor vegetation, similar to the tundra [19].

The dune forming phase is connected, according to these authors with winds blowing from N to W. In the area discussed, the dune formations connected with the above phase have gentle wind-ward slopes, in W or NW directions.

Poser in his characteristics of winds in the late glacial for European regions also mentions the prevailing NW winds in Poland [15].

From countries beyond Poland, the intercalation of sand-dune formations with lower lying loess are quoted from Bulgaria by Maruszczak and Trembaczowski [11]. Maruszczak and Mishew connect that formation with the last phase of the last glaciation [10].

The probable formation corresponding to sand-loess layers in western Europe is the second horizon of cover-sands described by Maarleveld [8].

On the discussed territory the existence of older dune phases prior to the Bolling, distinguished by Kobendzina [7] from the Kampinoska Forest is hardly to be expected.

While dunes were accumulating in the north of Poland, in the south loess was setting.

The warmer climate in the Allerõd produced again a horizon of fossil soil. It was probably detected in the road tunnel, south of the Gorzyczanka valley. In other profiles, the only remainder it left were well formed, ferruginous horizons. Later processes of erosion, denudation and also of eolian deflation destroyed the humic soil, disclosing the lower lying strata. It may be assumed that the old dune forms thus survived to the present day, cemented by the ferruginous layers.

In the same way, thick ferruginous layers preserved from destruction the dunes of the eastern Hungarian regions (Fig. 9). Ferruginous streaks are also noted in the lower dune series distinguished by Stankowski in SW Poland [18].

Fig. 9. Ferruginous layers in eolian sands of NE Hungary. Photo Z. Borsy

A drop of temperature in the younger Dryas (the period of tundra-steppe according to Firbas [4], steppe, according to Wasylikowa [19]) induced the formation of the last generation of glacial forms and the secondary blowing over of loess dust (formation of the upper layer of loess dust). This would supposedly correspond to the "covering loess" of Pozaryski [16]. The thickness of such a formation hardly ever exceeds 1 m.

South of the loess edge, on a substratum of fluvio-glacial sands and destroyed boulder clay of the Mindel glaciation, a series of parabolic dunes and blown-in embankments have developed.

Their shape and construction is distinctly different from the above described forms. They are constituted of middle and coarse-grained sands, without dusty material intercalations. In the transverse profiles, no ferruginous layers could be detected. The shape of these formations do not present moreover any trace of secondary destruction. They indicate a prevailing WSW-ENE and W to E direction of winds. The dune phase during which they developed is undoubtedly younger than the formation period of dunes described above.

Basing on investigations of the Lublin Upland, Jahn connects the parabolic dunes with the dry phase of the later Dryas [5]. Schönhals as well, basing on investigations of western Germany mentions eolian sedimentations carried by western winds and setting in the younger Dryas [17]. Maruszczak distinguished only one dune forming period, corresponding to the Boreal period of the Holocene [9]. Borówko-Dlużakowa considers that during the early Holocene only the Preboreal period was dry [1]. Stankowski, basing on investigations performed in NW Poland distinguishes four dune phases in the Holocene [18].

It is not easy to establish to which of the mentioned periods belong the young, well shaped parabolic dune formation. They are doubtlessly younger than the previously described forms, and probably connected with the largest and most generally extended dune forming phase in the region. It is difficult, however, to establish if it belongs to the late glacial or to the beginning of the Holocene.

The comparatively small amount of dune forming phases of this region as contrasted with Northern Poland could be explained by a more rapid expansion of vegetation on southern territories, at the end of the glacial and the beginning of the Holocene. Only distinctly drier periods and cold climate could induce a renewal of eolian activity.

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SOME PLEISTOCENE AND HOLOCENE SEDIMENTARY ENVIRONMENTS IN THE LIGHT OF THE MECHANICAL GRANIFORMAMETRY

BOGUMIŁ KRYGOWSKI

The mechanical graniformametry was initiated 27 years ago [3] when the first apparatus for the mechanical selection of sand grains according to their abrasion grade was constructed. Further experiments in this direction, conducted for several years [4, 5, 6, 9, 11, 14, 15] allowed such a machine to be built, which works fast and correctly, as many hundreds of analyses made on this newest model (Fig. 1) have shown.

In order to demonstrate the usefulness of this graniformameter, which means the application of the mechanical graniformametry, some materials (samples) of the Pleistocene and Holocene sedimentary environments have been chosen to be determined by means of this method.

The characteristics include, in this case, only one element (of the so-called graniformametry) namely the abrasion grade of quartz-grains in diameter 1·0-1·25 mm of separate sediments. Based on the thousands of analyses made on the bulldozer graniformameter, we could determine indices of the abrasion grade — W_0 and its heterogeneity — N_m according to the formulas [10]

$$W_0 = \frac{\sum [n \cdot k] \cdot 100}{N}$$

where: W_0 index of abrasion,
n number of grains in angle classes,
k mean angle characterizing a given angle class,
N number of grains in sample examined.

$$N_m = Q_3 - Q_1$$

where: N_m index of heterogeneity (dissimilarity),
Q_1 and Q_3 quartiles.

The enclosed graphics and tables show in the range of the grain abrasion some features of separate sedimentary environments, discussed in this paper, so we can sketch the exact characteristic for each environment respectively.

A general property of all sedimentary rocks is usually a remarkable variability of the abrasion grade (W_0 — values) and of its heterogeneity index (N_m — values).

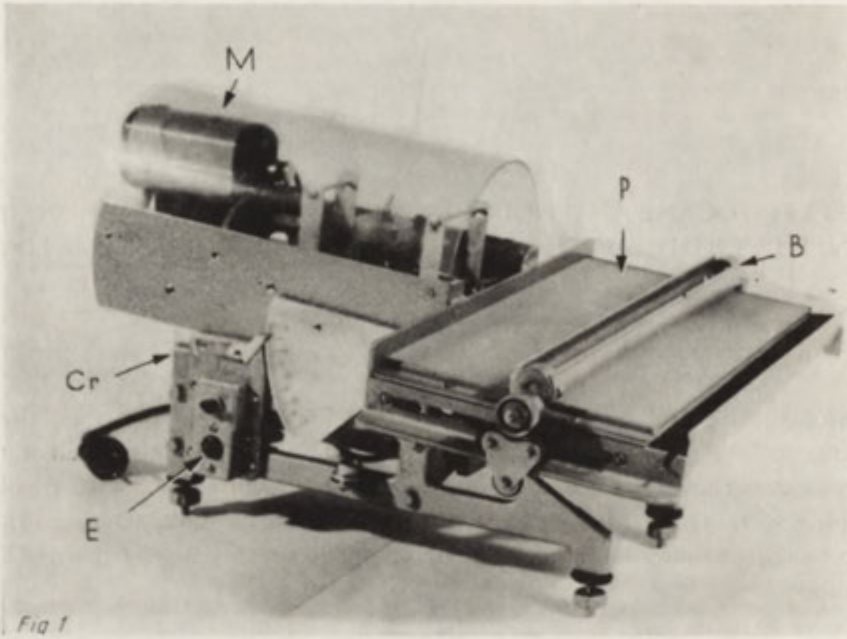

Fig. 1. Bulldozer graniformameter-apparatus to the mechanical selection of sand grains according to their abrasion grade

Main elements: P — a mechanically matted glass plate — inclined plane, B — bulldozer (two lintals), M — motor moving the bulldozer, Cr — crank to lift the inclined plane to the required angle, E — switching on (details of the acting of the machine are described in B. Krygowski's paper [11])

We do not know such sedimentary environments in which the W_0 and N_m values of separate samples of quartz-grains would be the same.

Till environment. It is clearly seen in Fig. 2 — I, representing the W_0 and N_m values for the till wall, 14 m high from Rewal (Baltic Sea shore-line zone). A very characteristic feature is the strong jumping of these index values, determined for 66 samples taken every 15 cm of the till wall, on the vertical line (profile). Based only upon this property it is possible to divide the Baltic till layer into three distinct sections:

A — with the low W_0 values and a strong jumping of the W_0 curve (probably a feature of disturbed sedimentation at the end of the accumulation period of the till layer),

B — with a little higher W_0 values and weak jumping (probably a feature of stabilized accumulation process),

C — with a little higher W_0 values than this in B and a strong jumping too (indicates that the accumulation process was disturbed at the beginning of the formation of the till layer and the grains were probably taken from the older till, that is why they are more abraded than the B layer grain).

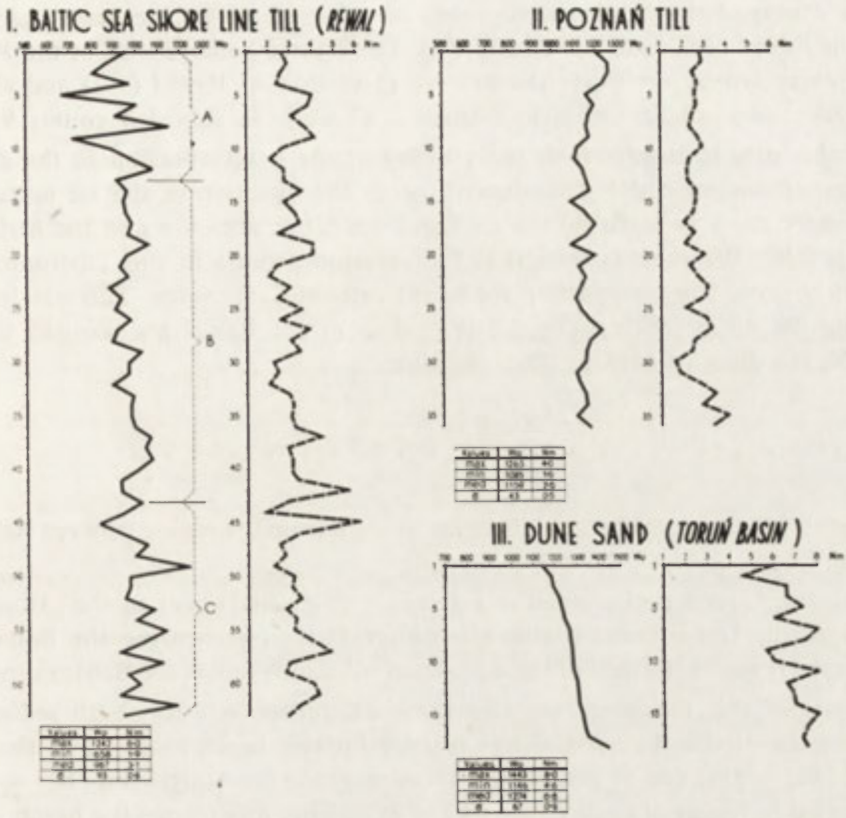

Fig. 2. Some sedimentary environments in the light of the W_0 and N_m curves (abrasion grade and its heterogeneity)

A similar picture is seen in Fig. 2-II representing the abrasion grade of the Poznań till with this difference that the jumping of the W_0 curve is here more quiet in general and the W_0 values higher than in Rewal (Fig. 2-I).

All these properties of the till environment are presented also numerically in table 1.

TABLE 1

SOME DATA OF THE ABRASION GRADE OF THE TILLS BELONGING TO THE BALTIC GLACIATION

	Rewal — Baltic Sea shore-line zone		Poznań — 250 km southward from the Baltic shore-line [1]	
values	W_0	N_m	W_0	N_m
max	1242	6·0	1263	4·0
min	658	1·9	1085	1·6
average	967	3·1	1159	2·6
standard deviation	93	0·8	43	0·5

The average values show the stronger abrasion of the quartz grain in Poznań than in Rewal (difference about 200 W_0). The greater homogeneity of the Poznań till is expressed by the lower average N_m (2·6) than in Rewal (5·1) and also by the lower standard deviation, in Poznań — 43 while in Rewal it counts 93.

All this data distinctly show the growing of the grain abrasion in the ground moraine environment going southward, i.e. in the direction of the ice movement. The longer the way made by the ice sheet the better abrasion and the higher its homogeneity. We should stress that the abrasion process in this environment is extremely slow. The transporting coefficient determines its value. This obtained by dividing the difference between the W_0 values of the Rewal and Poznań tills by the 250 km distance between these localities:

$$\frac{200\ W_0}{250\ \text{km}} = 0{\cdot}8/1\ \text{km}$$

This result is provisional because there are no data from the region between Rewal—Poznań.

Baltic beach sand environment is graphically (Fig. 3-III) near to the glacial tills and especially from Rewal (Baltic shore-line). It is clear because the Baltic Sea beach sand is mainly formed of the denudated till, which builds the Baltic shore-line.

In spite of this fact there are also some differences between both sediments. The jumping of the W_0 curve is less marked for this beach ($N_m = 2{\cdot}4$) than for the till ($N_m = 5{\cdot}1$) and in general the abrasion grade for beach sand ($W_0 = 1212$) is remarkably higher than for tills ($W_0 = 967$). This difference (W_0 beach — W_0 till $= 245$) probably determines the size of the abrasion of the till sand in the beach environment. Unfortunately it has been impossible to establish the length of transport (time and distance) [11].

River environment (Fig. 3-I) according to the example of the Kwisa river (Sudety Mts). The general features of the abrasion of this river sand are:

1. a strong jumping of the W_0 curve on the whole length of the river (as a result of the following: abrasion, selection, disintegration, participation in the river sediment of the new slope cover materials, the sediment of the tributaries etc.),

2. a permanent rising of the W_0 curve downstream expressed more or less by the same N_m value in the enclosed table 2.

TABLE 2

THE W_0 AND N_m VALUES FOR GRAINS (1·0-1·25 mm) OF SEPARATE SECTIONS OF THE KWISA RIVER

	Sections of the river	W_0	N_m	W_0/km
A	mountains	826	2·8	2·3
B	highland	915	2·6	
C	lowland	1110	3·0	

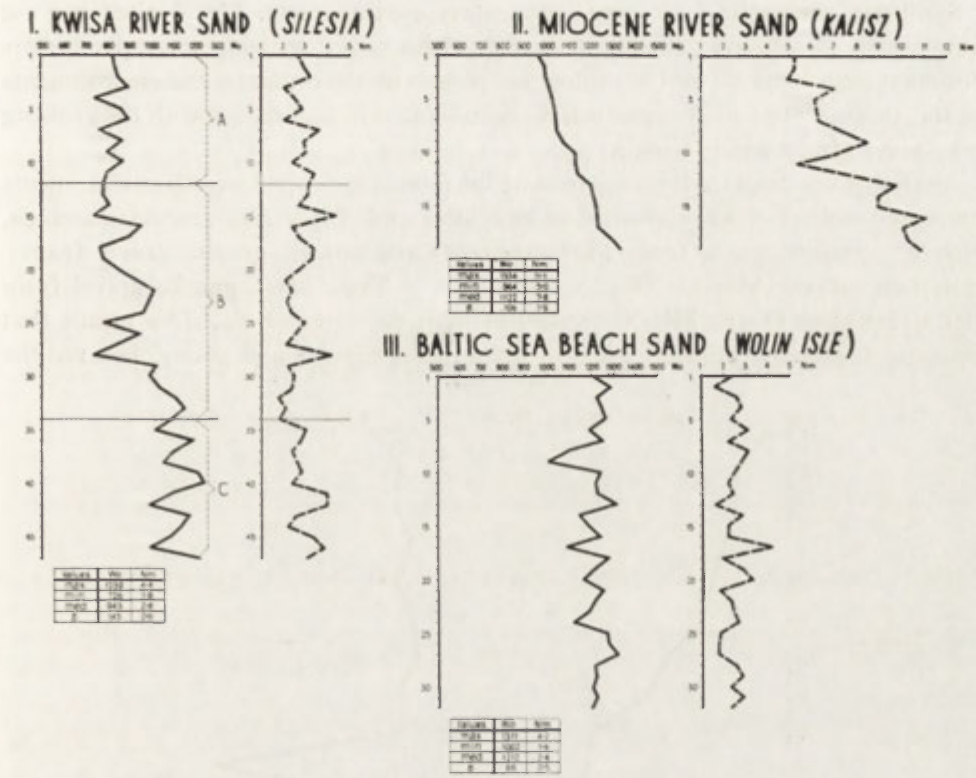

Fig. 3. Some sedimentary environments in the light of the W_0 and N_m curves (abrasion grade and its heterogeneity)

These data also indicate that the abrasion process probably took place along the whole length of the river with, more or less, similar selection conditions (similar hydrological conditions) expressed by almost the same N_m value in three Parts of the Kwisa river. The $\overline{W}_0$ data allow to divide the Kwisa river environment into three sub-environments: A — mountainous, B — highland and C — lowland. The characteristic of these subunits is as follows:

A — 1. the lowest $\overline{W}_0$ (826) typical for fresh angular grains, 2. weak jumping of the W_0 curve, 3. no rising of the W_0 downstream.

B — 1. higher W_0 (915), 2. strong jumping of the W_0 curve, 3. distinct rising of the W_0 downstream,

C — 1. highest $\overline{W}_0$ (1110), 2. a little weaker jumping of the W_0 than in B, 3. distinct rising of the W_0 downstream.

All this shows that the river discussed does not represent a homogenous environment, and it is desireable to mention at least three sub-units [10, 11, 16].

Synthetic comparison of some sedimentary environments. Fig. 4 gives a good comparison of all environments discussed in this paper showing some parameters simultaneously. The N_m and W_0 values are plotted on the ordinate, and environments on the abscissa. The environments have been plotted in accordance with the growing W_0 values (from left to right).

Analysing the diagram it is easy to state the following properties: All environments connected with the Karkonosze (Sudety Mts) and Tatra Mts granite materials, such as: granite waste from Michałowice (Karkonosze), granite gravel from mountain gullies (Morskie Oko, Czarny Staw — Tatra Mts), granite gravel from mountains talus (Tatra Mts) show low indices; the W_0 and N_m. This means that grains of these environments are still quite fresh, angular and young, because the

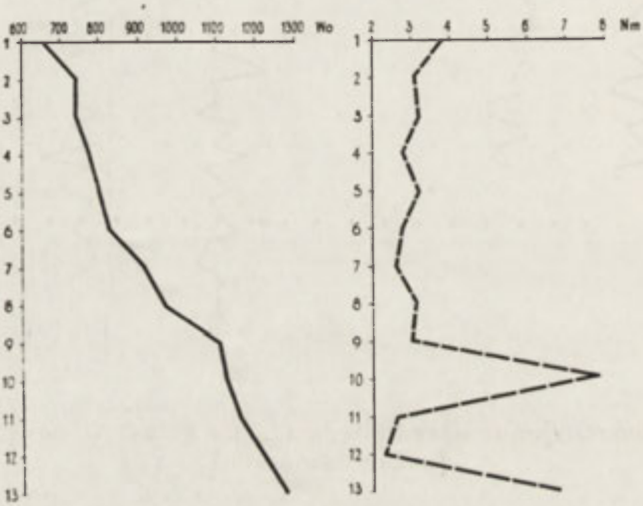

Fig. 4. A comparison of the sedimentary environments according to two parameters: W_0 and N_m (at the growing to the W_0 to right)

Environments (sands, gravels, tills): 1 — Morskie Oko gully (Tatra Mts), 2 — Czarny Staw gully, 3 — Czarny Staw talus, 4 — Morskie Oko talus, 5 — granite waste, 6 — mountainous Kwisa river section (Sudety Mts), 7 — highland Kwisa river section, 8 — Rewal till (Baltic Sea shore-line), 9 — lowland Kwisa river section, 10 — miocene river (Kalisz), 11 — Poznań till, 12 — Baltic Sea beach (Wolin Isle), 13 — dunes (Toruń Basin) according to the diploma work under the direction of R. Galon

transport, which usually abrades grains, was to short or hardly existed, as in the case of the granite waste. The mountainous section of the Kwisa river belongs, with its low W_0 and N_m values, to the group of environments with angular, young grains and a good selection (low N_m). It is worth mentioning that the N_m value of this mountain river environment is the lowest in the group of the mountainous environments, due to longer transport in comparison with such environments as: mountain talus, mountain gullies or mountain slope waste covers, in which the distance transported is usually very short.

Based on these facts one principle can be stated: the nearer the separate environments become to rivers, the lower is their heterogeneity (N_m), and this means the selection process is better during the water transport. It is unusually interesting that the initial N_m (heterogeneity) is relatively low. It is probably connected with the texture of the granites. The other environments on the diagram such as: highland river — 7, till — 8, 11, South Baltic beach — 12, are characterized by low N_m and high W_0.

The better abrasion of the grains of the Poznań till than of the Baltic Sea is due to the longer transport of the first, as has been mentioned above.

The fact that the lowland Kwisa river environment — 9 is located in the classification of environments above between the Baltic Sea till and the Poznań till can be explained by the fact that the Kwisa river flows in the lowland zone through the Pleistocene sediments, known for their "rounded" grains (typical of the Pleistocene material). Thus the river sediment of this river section is a mixture of abraded Pleistocene sand and angular fresh mountainous (granite) sand. That is why it is placed so near the Poznań till.

The Baltic Sea beach (Wolin Isle) has a special position in the classification order. Its W_0 is very high (strongly abraded grain) at the lowest N_m, i.e. the best selection. Two agents of this environment are responsible for these properties: 1. a very strong abrasion and 2. very strong selection.

Only the dune sands from Toruń Basin have a little higher W_0 than the beach sands, but a very high N_m. It indicates that also in the dune environment the abrasion was very strong but the selection process was extremely weak. Those are just typical features for this sedimentary environment.

The Miocene river from Kalisz represents a similar type with the exception that the abrasion of these sands is remarkably lower.

These facts suggest that the accumulation process in the Miocene environments (lowland rivers and basins) was very fast.

The mixture of already well abraded grains (perhaps in the dune environment) and more weakly rounded ones was accumulated chaotically. The distinct similarity between the dune environment and Toruń Basin supports the conception of the remarkable role played by dune processes in the continental Miocene river and basin sedimentation in south Great Poland (Kalisz).

Sedimentary environments in the light of the rectangular set W_0 and N_m. The position as well as the differences between the separate environments is well shown by the rectangular set (Fig. 5). The separate environments have not only different positions in the rectangular set but also different shapes of their fields. The Poznań till field, for instance, lies more to the right of the Baltic Sea till and is quite round because the homogeneity of this till is high, due to the long transport. The Baltic Sea till lying more to the left (more weakly abraded grain than in the Poznań till) has also an oval shape but with characteristic long peninsulas (arms) which represent the foreign material in reference to the basic mass of grains included in the main

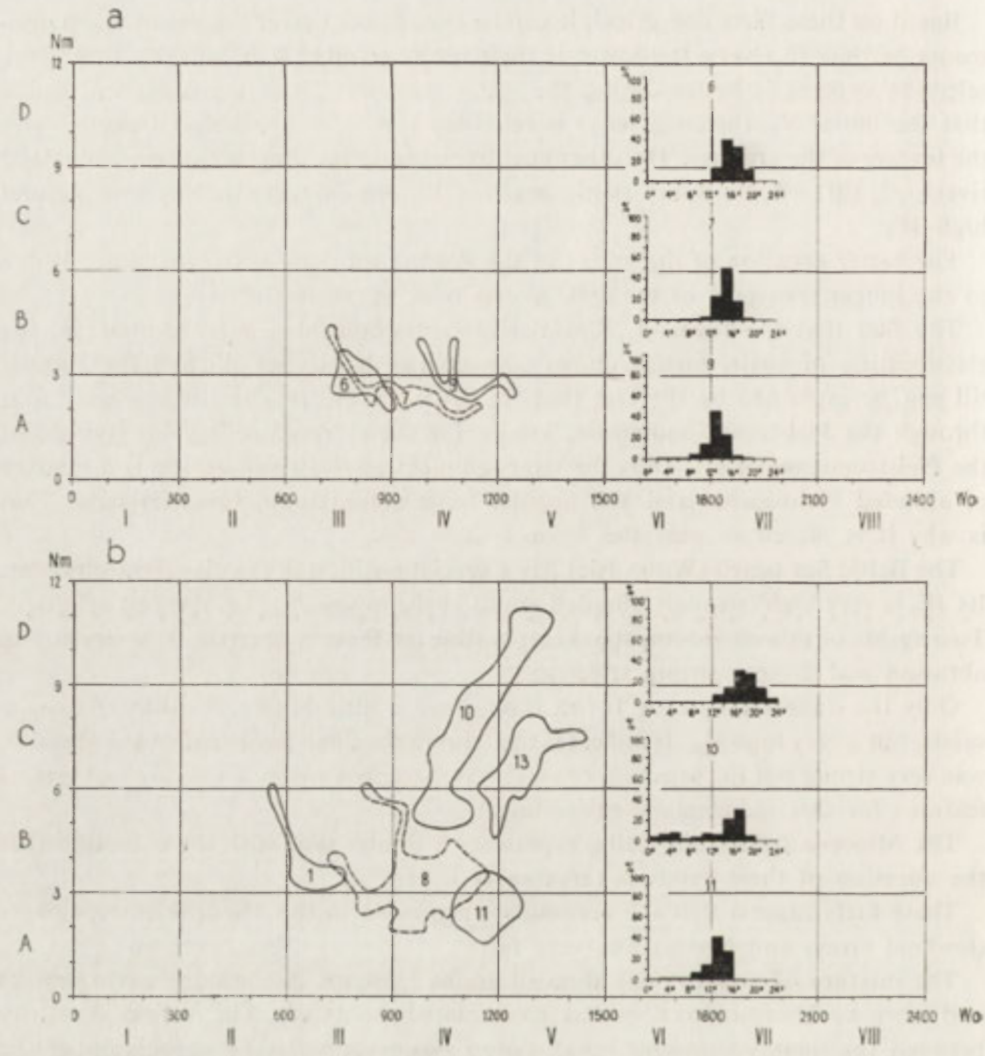

Fig. 5. Shape and position of separate environments in the rectangular set W_0 and N_m

(a) Environments of Kwisa river: 6 — mountainous, 7 — highland, 9 — lowland. The histograms representing the most typical abrasion in some environments have also been located in the rectangular sets. (b) Environments: 1 — Morskie Oko gully (Tatra Mts), 8 — Rewal till (Baltic Sea), 10 — miocene river (Kalisz), 11 — Poznań till, 13 — dunes (Toruń Basin)

field. That is just an example of the joining process of the new, foreign material with the till mass sedimentologically already mature. We see here some possibilities of calculating the participation of these foreign materials in "old" till material.

For the mountainous environments, too, (gully, talus, mountain river sections and river environments in general) such peninsulas are very characteristic and are connected with rapid and frequent admixtures of foreign material (tributaries, slope granite waste etc.) with the old (parent) sand material of the given environments. All the mountainous environments discussed in this paper are of this type, fields with peninsulas.

The Kwisa river consists of three fields with similar peninsulas described above. Long peninsulas of the highland and lowland fields going up and to the right, especially in the last field, reveal the participation of the Pleistocene material (rounded grain, greater heterogeneity) in the parent river material. This influence is seen also in the highland field.

The Baltic Sea beach environment (field left) represents sand which has a lower N_m (heterogeneity) the more it is beach-like and the higher W_0 (abrasion). Peninsulas going up and to the left show a new admixture of fresh till grains with less abraded grains and a little greater heterogeneity.

Dune and Miocene river environments represent highly typical fields. Their histograms, too, showing at least two culminations support the idea that these environments primarily belong to some different environments.

It has been shown that the mechanical graniformametry gives very typical characteristics of separate sedimentary environments and therefore it can be used in the sedimentological research of the modern geomorphological and geological investigations, in general.

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ESKERS AND KAMES IN THE SPITSBERGEN AREA

JAN SZUPRYCZYŃSKI

For the first time, eskers and kames were distinguished in the areas of Pleistocene glaciations of Europe and North America. Only much later these landforms were also discovered in areas of contemporary glaciers [2, 8, 16]. Within older Pleistocene glaciations, eskers show lengths from a dozen to several score kilometers and heights reaching from a dozen to well over 50 m (the greatest esker height, up to 140 m, was found at Uppsala, Sweden). In contrast with eskers formed in the Pleistocene inland ice period, eskers found in contemporary glaciated areas attain dimensions very much smaller, being only several hundred meters long and but a few or, at the most, a dozen meters high [2, 14, 15]. In a similar way, kames found in areas of present-day glaciation are of much smaller size. A further fact should be stressed here: in areas of contemporary glaciation, esker and kame forms are found but very rarely. This must probably be ascribed to the considerably smaller areas of today's glaciation, compared with those of the Pleistocene period.

In Antarctic, an inland ice area closest approaching Pleistocene conditions, the climate is unfavourable to the development of crevasse forms. The severely cold climate precludes a downflow of meltwater in open glacier crevasses, as well as drainage within the glacier, thus prohibiting an accumulation of fluvioglacial material within the inland ice. Further, virtually everywhere the Antarctic inland ice protrudes beyond the land shelf and ends as cliffs in the sea — a fact that anyway would preclude any esker forming.

So far, but few eskers have been discovered along the margin of Greenland's inland ice [2, 11]. Towards the close of the 19th century, A. Kornerup observed at the western edge of the Greenland inland ice a sandy-gravelly ridge, up to 6 m high and some 80 m long, extending parallel with a glacier. On the eastern shores of Greenland, in the forefield of the inland ice, M. Vahl discovered in 1928 a sandy-gravelly ridge, up to 5 m high and up to 30 m long. It is probable that, in both instances, these forms were eskers.

In Alaska, eskers have been described from the forefield of the Vakuta and Woodworth glaciers [3]. The forming of eskers *in statu nascendi* has also been observed at the Malaspina glacier [1, 2].

Very rarely small forms of eskers and kames are discovered in glacier forefields or on mountain glaciers of the Alps or the Caucasus Mts, of Pamir and of Scandinavia.

In areas of contemporary glaciation, the greatest variety of esker forms has, so far, been determined in Iceland, in the marginal zone of the Vatnajokull glacier [5, 14, 15, 16]. These eskers occur in both the northern and the southern forefield of this glacier; they are found here in considerable agglomeration in lengths from a dozen meters to three kilometers [15]. In Iceland, their height is from 1 to 10 m. They consist of sandy-gravelly material; often the bottom of these forms contains an ice base, whereas the fluvioglacial material is only up to 4 m thick. Alongside of well developed large forms, there are also found smaller forms, not more than 100 m long and 1 to 5 m high.

In the marginal zone of Vatnajokull eskers of englacial and subglacial type have been observed [5, 15]

Distribution of eskers in Spitsbergen

In recent years, a considerable number of eskers and kames have been discovered in the marginal zones of, or lying on top of, Spitsbergen glaciers. Within the scope of an expedition organized by the Polish Polar Association (*Polskie Kolo Polarne*), M. Klimaszewski undertook in 1938 investigations in the area between Kongsfiord and Eidembukta. Apart from other glacial and glaciofluvial forms, this scientist found here esker and kame forms also (Fig. 1, area 1).

During research work carried out in 1959 and 1960 by the Polish Spitsbergen Expedition of the III International Geophysical Year, headed by S. Siedlecki, S. Jewtuchowicz and J. Szupryczyński investigated in detail the esker and kame forms discovered in southern Spitsbergen (Fig. 1, areas 2, 3, 4). In 1963, the present author made, within the scope of a research group of the expedition of the Norwegian Polar Institute, comparative studies in northern Spitsbergen. On the shore of Bockfjord, on top of a small lateral glacier, he discovered a small esker form *in statu nascendi* (Fig. 1, area 5).

In his summary of the results of his geomorphological studies in north-western Spitsbergen, M. Klimaszewski [9] gave very brief data on esker forms observed in this region. In 1938, he had observed short esker ridges in the forefield of the Comfortles and Aavatsmark glaciers. The esker ridge on the Comfortles forefield, built of sands and silts, was 3 m high and had a slope angle of up to 35°. The short esker ridge in the forefield of Aavatsmark, consisting of fine gravel, showed a height of 1·5 m. In both instances, records on the length of these forms, as well as further morphometric data are lacking.

In 1959, detailed studies of the morphology and the internal structure of the eskers occurring in southern Spitsbergen were made by S. Jewtuchowicz [7]. On the southern shore of Hornsund fjord, in the forefield of the Gås glacier, he discovered

Fig. 1. Sketch of distribution of eskers and kames in Spitsbergen, valid up to 1963

A. — kames, B. — eskers. These eskers and kames were investigated by: 1 — M. Klimaszewski, 2, 3 — S. Jewtuchowicz 4, 5 — J. Szupryczynski

a number of short esker forms (Fig. 1, area 2). The largest is 186 m long, oriented obliquely to the present glacier margin, and running from SE to NW. This esker is connected with other eskers of smaller size, of 32 to 90 m lengths, thus forming a pattern resembling a river with its tributaries [7, p. 52] (Fig. 1). Near this esker system there occur solitary eskers extending in straight lines. The height of these eskers of the Gås forefield is small, 2 m at the most, counting from their base up. These eskers are built of material of various fractions. Their structure comprises sand, gravel, and pebbles of up to 20 cm diameter. In all the eskers investigated, S. Jewtuchowicz noticed stratified material; but there also were esker sectors in which the material lacked segregation. Their internal structure indicated their origin in subglacial tunnels.

On the Bunge glacier (Fig. 1, area 3), S. Jewtuchowicz [7, p. 53] discovered an esker *in statu nascendi*, melting from the ice due to ablation. The morphological axis of this esker runs parallel with the direction of the glacier. This esker rests on the surface of the glacier; its height, measured from the ice surface, was 6 m. In 1959, the esker was 80 m long and extended in a straight line. Its crest was very sharp and narrow. The position in which the form of this esker appeared, indicates its englacial origin.

In the marginal zone of the Werenskiold glacier, on Wedel Jarlsberg-Land four esker forms were investigated in 1960 (Fig. 1, area 4). Detailed descriptions of their morphology and their geological structure were given in a paper written in Polish [13]. One of these eskers lies at the base of the distal slopes of ice-morainic ramparts, situated some 1·5 km from the glacier margin's 1960 position. The remaining three eskers lie near the glacier margin, and next to the glacier snout.

The esker adjoining the slopes of the ice-morainic ramparts is of minor size; its total length is 128 m, its maximum height 4·4 m. The morphological axis of this esker extends from N 310° to N 325°, that is, perpendicular to the glacier snout descending into the valley in a north-western direction. At its base it is 5-10 m wide, while the crest width is 0·5-2·0 m. This esker is developed in the shape of a ridge, with its surface gradually subsiding westwards. It consists of stratified fluvioglacial deposits: sandy silts, sands and gravels. At its base, the strata dip westwards at an angle of 6° to 16°. The geological pattern and the interior structure of this esker reveal its supraglacial origin; that is, it must have developed in an open glacier crevasse. The material building the esker is fluvioglacial from base to top. Structural measurements indicate that in the crevasse, where the esker originated, the flow was linear, from SE to NW.

The further three eskers are situated next to the glacier margin and on the surface of the glacier. Of particular interest is the esker on top of the glacier, in close vicinity to its snout. Its length, measured with a surveyor's tape along its morphological axis, was exactly 125 m on July 20, 1960; six weeks later, on Nov. 1, it was only 107 m. The morphometric data of this esker were as follows:

length — 125 m, later 107 m
direction of morphological axis — N 240° to N 295°
height — 1·6 m to 8·4 m
width at base — 1·3 m to 25·0 m
width of crest — 0·1 m to 2·2 m
slope inclination — exceeding 30° throughout.

This entire form consists of four sectors, separated by depressions in the undulating crest line (Fig. 2). The contours of this esker are angular; for a considerable length, the crest width does not exceed 0·5 m (Fig. 2). The upper part of the esker contains distinctly stratified fluvioglacial material (Fig. 3), resting on ice which here constitutes the base of this form. The thickness of the fluvioglacial material is from 0·6 to 4·2 m. In general, fine-grained material prevails, as shown in table 1.

Fig. 2. Esker on surface of Werenskiold glacier. Photo J. Szupryczyński, July 1960

Fig. 3. Inner structure of esker lying on surface of Werenskiold glacier. Photo J. Szupryczyński, July 1960

TABLE 1

STRUCTURE OF THE ESKER'S FLUVIOGLACIAL MATERIAL

Fraction, in mm	Weight, in g	Per cent
. 10	17,200	4·2
10 — 5	19,700	5·0
5 — 2	36,600	9·2
5 — 1	70,000	17·5
1 — 0·5	77,700	19·4
0·5 — 0·2	100,600	25·1
0·2 — 0·1	57,100	14·3
0·1 — 0·06	10,600	2·7
. 0·06	10,500	2·6
Total	400,000	100.0

This fluvioglacial material is angular; its grains are not rounded at all — proof of a very short distance of transport. The directions of dip and strike of the strata of this esker are shown in the diagrams attached (Fig. 4). Structural measurements were made solely in the parts indicating original accumulation. The esker rests on the glacier, and its slopes are being undermined by meltwater streams flowing over the glacier surface. In melting, the ice causes heavy disturbances in the material originally accumulated, as revealed, among other facts, by changes in the dip of the strata. This esker developed englacially, in a tunnel in the glacier interior. Inside the esker a distinct stratification has been observed, and on its surface lies a thin

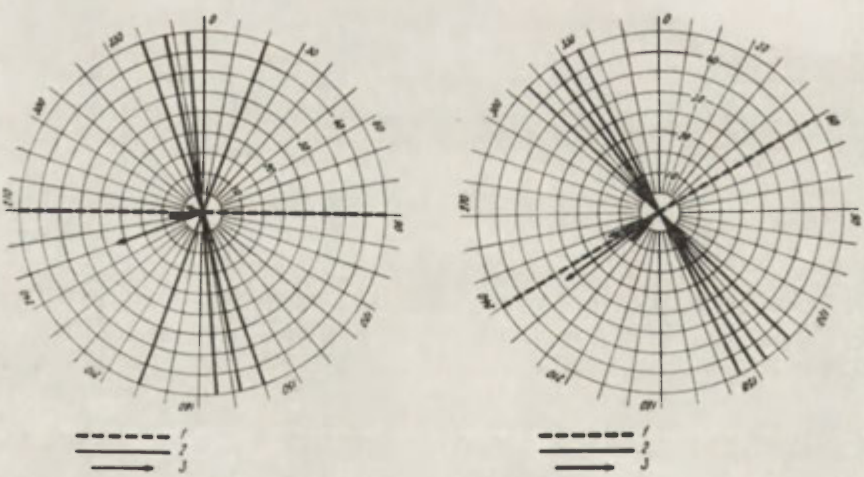

Fig. 4. Diagram of strikes, directions, and angles of dip of strata in esker on surface of Werenskiold glacier

1 — morphological axis of esker, 2 — strike of strata, 3 — directions and dip angle of strata

layer of an ablation moraine (Fig. 3). The structure of the strata discloses a linear westward water flow within the tunnel; the dip of the stratification is within a range from 4° to 22°.

This esker, formed in an englacial tunnel, gradually became uncovered with ablation proceeding at the glacier's snout. The ice forming the base of the esker constituted the floor of the ice tunnel. While a cover of fluvioglacial material protected initially the ice forming the base of the esker against melting, later on the entire form collapsed and underwent flushing away. In a period of 40 days during the 1960 season of research work, meltwater destroyed in the eastern part of this esker a sector of 18 m length. In this part, the esker dimensions were rather scant: height up to 1·60 m, base width up to 2·80 m.

North of the form described above, there lie two further eskers. The larger one shows the following features:

length along morphological axis — 213 m
morphological axis — varying from N 225° to N 340°
height — up to 1·80 m
width at base — 3·0 to 7·5 m
width of crest — 0·4 to 4·0 m
slope inclination — 30° to 34°, at places undercut by meltwater — up to 60°.

This esker lies at the outlet of a tunnel of an englacial stream; its eastern side is badly damaged by meltwater flowing down the glacier. The surface of this esker shows distinctly a westward inclination of 4° to 6°. From the base upwards to the top, distinctly stratified material has been observed, consisting of silts, sands and gravels. The continuity of stratification can be studied over the full length of this esker.

The material building the esker lacks rounding — a fact indicating a short distance of transport. Structural measurements carried out in the interior of this form show a westward flow of water, parallel with the run of the morphological axis. The strata dip at 8° to 12°. This esker developed supraglacially at the glacier snout, in a very narrow crevasse, and was deposited directly on bedrock. Although of small size, it shows very distinct morphological contours and a distinct interior structure. Unless it suffers destruction by meltwater, it may persist for a considerable time.

South of the form described, there lies a short esker ridge of 46 m length; this is a form bent in the shape of a crescent. The morphological axis of its eastern part, 21 m long, runs approximately at N 260°; the remaining part, 26 m long, shows a N 310° direction of its morphological axis. The height of this form is from 1·2 m to 1·4 m. Its base width is 4·4 m, with up to 2·0 m width of the crest surface. Over its entire length the crest is flat, with a 5° slant westwards. The interior structure of this esker corresponds to that of its neighbour. In many places the slopes, inclined up to 30°, are undercut by meltwater from the glacier. The origin of this esker is the same as the one previously described.

Numerous minor eskers have been discovered in the marginal zone of the Nann glacier on Torell Land (Fig. 1, area 4). A picture of the esker system in the southern region of this glacier is shown in Fig. 5. Table 2 gives the morphological data for the individual esker sectors.

TABLE 2

MORPHOLOGICAL DATA OF THE INDIVIDUAL ESKER SECTORS

Esker sectors	Direction of morphological axis	Length of crest m	Width of crest m	Width of base m (ca)	Maximum height m	Maximum angle of slope inclination m
I	N 15°	9·0	0·1	7·0	4·2	40°
II	N 270°	8·0	0·1	7·0	4·2	40°
III	N 200°	15·0	0·1	7·0	4·2	40°
IV	N 290°	36·0	1·0	5·0	2·6	36°
V	N 310°	18·0	3·0	6·0	1·5	36°
IV	N 280°	14·0	3·0	6·0	1·6	36°
VII a)	N 290°	17·0	2·5	8·0	3·8	50°
b)		23·0	2·0	4·5	2·6	30°
VIII	N 260°	8·0	1·5	3·0	1·0	30°
IX	N 290°	10·0	6·0	8·5	4·2	55°
X	N 270°	16·0	0·5	5·0	3·0	60°
XI	N 270°	14·0	6·0	8·0	3·0	32°
XII	N 300°	55·0	2·0	6·0	3·0	30°
XIII	N 315°	33.0	0.1	6.0	3.5	40°

The total length of all twelve esker parts is 233 m. In numerous places, the interior of this esker is exposed. The esker is built of fluvioglacial material in a variety of fractions, from silts to sands and coarse gravels. The strata within the esker dip at 6° to 14°, parallel with the axis of the form. At many places of the crest, a mantle of till with blocks up to 0·3 m diameter is found — traces of an ablation moraine. The esker sectors farest to the east rest on the glacier snout. Ice forming the esker base has also been found in sectors X and XI.

The origin of this esker form is complex. Its eastern part has been deposited in a tunnel inside the glacier, as implied by the following features: 1. an ice forming the base, 2. the ablation moraine cover. Thus, sectors X, XI and XII developed englacially.

On the other hand, the western sectors originated supraglacially in an ice canyon cleaving the ice to its very base; evidence to this assumption are: 1. the lack of ice at the esker base, 2. a distinct stratification extending from top to bottom of the form, 3. here a covering by an ablation moraine is lacking. Therefore, sectors I to VIII must have developed independently of sectors X, XI and XII, as implied by the morphological evidence mentioned.

There are also several eskers in the northern region of the Nann glacier, the largest of them some 140 m long. Its course is linear, the morphological axis running in a N 250° direction. The height of this esker, including its ice-socle, is 16 to 18 m. The fluvioglacial material lying on its surface is up to 2·5 m thick. At its base, this

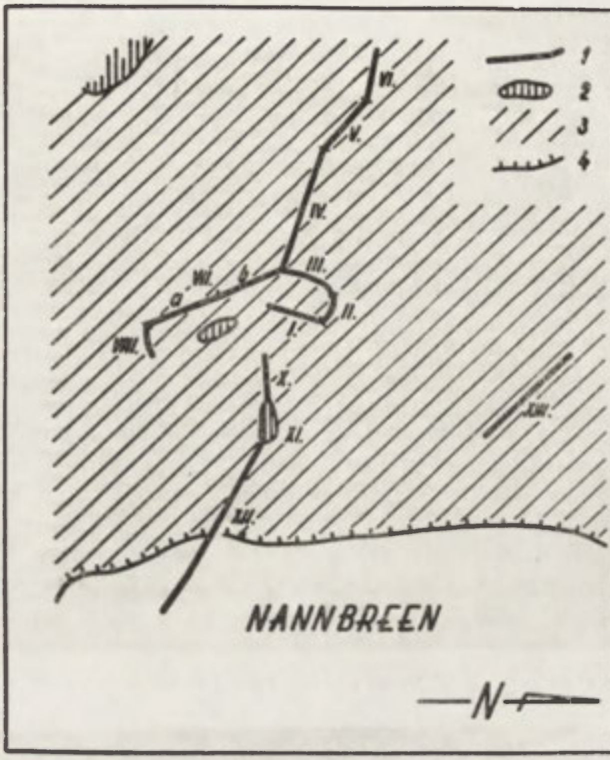

Fig. 5. Map of eskers lying in front of snout of Nann glacier

1 — esker sector, 2 — kames (with exception of sector XI showing widening of esker), 3 — ablation moraine, 4 — glacier margin

form reaches a width of as much as 45 m. The ridge is developed in the shape of a sharp crest 5 to 20 cm wide (Fig. 6); the slope inclination is 38° to 40°. The fluvioglacial material covering this esker is distinctly stratifield; here and there, this fluvioglacial mantle is overlain by an ablation moraine. The stratification of the fluvioglacial deposit extends at 6° to 16° north-westwards. The geological structure and the profile of this esker indicate its englacial origin. Principal evidence to this assertion is the covering of the esker by an ablation moraine, and its having developed on an ice base. It is conceivable, that this esker form is actually of much greater length, and that, in time, its eastern sectors may emerge yet from the glacier, provided deglaciation continues in this region.

The remaining eskers occurring in this region of the Nann glacier are of much smaller size, reaching lengths up to 40 m only, and heights of 8 m, inclusive their ice bases. These forms also originated in englacial tunnels.

Numerous minor esker forms were discovered on top of the southern flank of the Torell glacier (Fig. 7); however, due to the expedition's seasonal work being completed, no detailed investigations were carried out here.

Fig. 6. Esker on surface of Nann glacier. Photo J. Szupryczynski, August 1960

Fig. 7. Area of "dead ice" forms at southern flank of Torell glacier. In middle part distinctly visible is a crescent-shaped esker, up to 8 m high. Photo J. Szupryczyński, August 1960

Fig. 8. Esker in front of glacier on Bockfjord. Photo J. Szupryczyński, July 1963

A small esker was found in northern Spitsbergen, in the forefield of a minor glacier on the shore of Bockfjord (Fig. 1, area 5). So far, a sector 80 m long has come into view (Fig. 8), since this esker is gradually emerging from the glacier. At the base, the width is up to 10 m; the narrow crest forms a sharp ridge 5 to 20 cm wide. The slope inclination is as much as 45°. In the part that so far has been uncovered, this esker is up to 10 m high, consisting of fine sand from base to top. Thus, the geological structure reveals this to be a supraglacial form, developed in a glacier crevasse.

Kames and kame terraces

In the regions containing eskers, kames also have been observed (Fig. 1). This must be ascribed to the formation of both these forms in dead ice, and to their connection with areal deglaciation as now commonly is taking place in the Spitsbergen area [9, 13]. The kame forms discovered in Spitsbergen are of smaller size. In the forefield of the Comfortles glacier (Fig. 1, area 1), M. Klimaszewski found two kame forms of 3 to 4 m height. In their lower part, they were built of gravels in delta stratification; thus they must have developed in a basin of stagnant water. The upper part consisted of normally stratified gravel, gradually accumulated by water flowing through the crevasse. On the margin of the Aavatsmark glacier, M. Klimaszewski discovered a kame ridge of some 350 m length, 3 to 6 m wide and about 3 m high. This form contained stratified coarse gravel.

In the flanks of the following glaciers: Uversbrae, Comfortles, Aavatsmark, Osborne, Vestgota and Eidem, well developed kame terraces have been observed [9, 10]. They constitute unilateral terraces built of gravels and sands, deposited

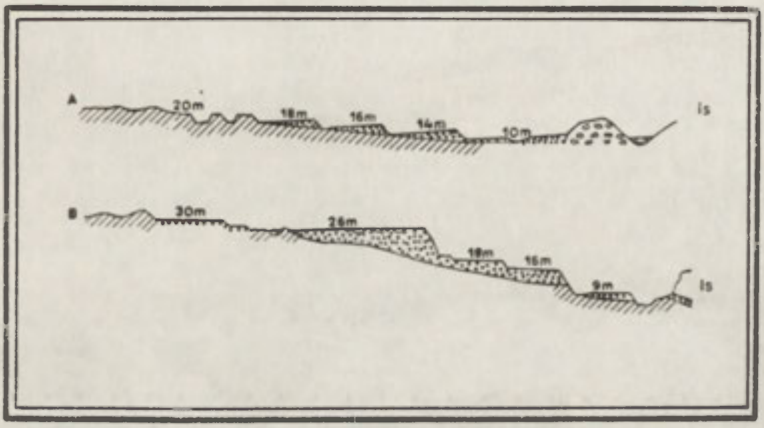

Fig. 9. Kame terraces in forefield of Eidem glacier, after M. Klimaszewski. Is-Ice sheet

between a mountain slope and the glacier by the action of marginal flow of meltwater streams (Fig. 9).

A few minor kames occur in the forefield of the Gås glacier [7]. In this region, the kames are spatially connected with eskers (Fig. 1, area 2). They form regular hillocks up to 2 m high, built of stratified material.

Kames *in statu nascendi* were observed by S. Jewtuchowicz on the Bunge glacier (Fig. 1, area 3), rising some 3 m above the glacier surface. The kame forms in the forefield of the Gås glacier, as well as those on the Bunge glacier, were formed in depressions covered over, that is, in ice caves with water drainage. For the kames of the Gås forefield, S. Jewtuchowicz assumes a subglacial origin, whereas, in my opinion, those on the Bunge glacier developed englacially. The interior structure of these kames reveals them to have been formed by material accumulating in rather inert water.

I also observed numerous smaller kame forms in the marginal zone of the Nann glacier [13]. As to their maximum dimensions, they were up to 4·0 m high, some 22 m long and up to 15 m wide (Fig. 10). Built of silts, fine-grained sands and gravels, they were stratified from bottom to top. The individual layers were deposited nearly horizontally being inclined but slightly, up to 8°. These kames developed supraglacially in depressions of dead ice, reaching down to the glacier base (in the very thin ice sheet of the dead glacier snout).

Fig. 10. Kame in front of Nann glacier. Photo J. Szupryczyński, August 1960

On Nordaustlandet Island, situated at the north-eastern shores of Wahlenberg fiord (Fig. 1), K. Sandfort [12] found six splendidly conserved kames at an altitude of close on 200 feet (some 65 m a.s.l.). These cone-shaped kames with flat crests were up to 50 feet (some 16·5 m) high, built of sands and gravels. It must be assumed, however, that these kames are not connected with contemporary glaciation, but rather developed during postglacial glaciation [6].

Conclusions

On Spitsbergen, eskers have so far been discovered sporadically only in its southern and northern part; none were found in the central part of Spitsbergen [4]. These eskers are of small size, up to 200 m long and 18 m high. They are built of fluvioglacial material, that is, of silts, sands and gravels. Some of these forms were observed *in statu nascendi*. The origin of the Spitsbergen eskers is diverse. In part, they originated subaquatically (Comfortles and Aavatsmark glaciers), others subglacially (marginal zone of Gås glacier) or englacially (Bunge, Werenskiold, Nann and Torell glaciers) or, finally, supraglacially (Nann and Werenskiold glaciers, and marginal zone of small glacier at Bockfjord). Englacial eskers predominate;

these are ephemeral forms, disintegrating upon melting of the ice forming the esker base.

As a rule, eskers are accompanied by kames. These are very small forms, up to 4 m high, with widths of a dozen or so meters; they are built of silts, sands and gravels. They develop in depressions of the glacier surface, or within the glacier body. Thus we distinguish: supraglacial kames (marginal zones of Comfortles and Nann glaciers), englacial (Bunge glacier), and subglacial kames (marginal zone of Gas glacier).

Apart from kame hillocks, kame terraces have also been discovered in Spitsbergen. It may be expected, that future investigations of the marginal zones of Spitsbergen glaciers will yield further discoveries of esker and kame forms, and make possible studies of their origin. It seems probable, that forms of this kind will prove as common in Spitsbergen as they are in Iceland.

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ACTIVITY OF RUNNING WATER IN SOUTH-WESTERN SPITSBERGEN

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Observations concerning the activity of running water under arctic conditions being sporadic, done in various places and at different times fail to give a sound basis for a proper evaluation of the morphological importance of that agent. Weather conditions especially in a maritime type of arctic climate, vary from year to year deviating greatly from the mean values [5, 11]. Observations collected during a comparatively dry summer may suggest that running water is scarce and its morphologic importance is negligible, while those done during a wet summer will suggest just the reverse. Moreover, the aquatic processes show a distinct annual periodicity with the maximum in spring. Much of it or all may escape observations carried out during the summer period only. Nevertheless water is sometimes mentioned in the existing literature concerning the arctic morphology as a morphogenetic agent [2, 5, 6, 7, 9, 10, 13, 15, 16].

Blanck, Rieser and Mortensen [2] write about transportation of fines from between the coarse debris by meltwater running over and under the ground surface. In Mortensen's opinion [14] the importance of running water in the creation of the relief in Spitsbergen is undervalued. Some small forms regarded as solifluctional are created by water. According to him and Büdel [3] the rows of rills on steep slopes (over 15° of declivity) cause the parallel retreat of these slopes without altering their general form. The flat-bottomed, steep-sided river beds owe their form to the erosion and transportation of gravel occurring on the whole breadth of the river bed during the spring thaw [3]. Surface wash received more attention from Jahn [10]. According to his measurements and calculations this process alone can lower the surface of slope by about 1 mm per 150 to 170 years. Some slopes show a positive denudation balance [8]. Wash and solifluction can work on the same slope [10].

The whole year observations carried out by the present author at the Polish I.G.Y. Station at Hornsund, West Spitsbergen, show that water may remain there in a liquid state for about 6 months per year. The total measured precipitation for the year 1957—1958 reached 340 mm. Of that figure 133 mm was rain, 84·5 mm was snow and 129 mm was rain with snow. 91·4 mm of the latter fell in October and November. At that time the low temperature of air and ground as well as the lack of free outflow of the rivers (their mouths being sealed up by storm ridges) prevents water from joining the sea. Therefore that part of precipitation can be added to one which is stored in the form of snow and ice. The whole stored precipitation amounts then to 220·4 mm. One ought to add to it that part of autumnal rain water which saturates the soil and vegetation cover dried up in late summer. It can be taken that at least 2/3 of the annual precipitation remains stored through the winter to become liquid during the spring thaw period. Its volume is probably much greater than recorded by the standard methods [1, 4, 5]. Kuziemski [12] shows that of the 5·1 to 6·6 million m³ of water discharged during the 50 days of the summer of 1958 from the river basin of Revelva (Hornsund) only 0·9 to 1·3 million m³ came directly from precipitation. On account of the volume ratio itself it can be taken that thaw water must play a much more important part in the creation of the relief than does the rainwater. The more so that the meltwater runs off mostly in a short period of about 4 weeks.

Melting of snow on steep, southern slopes may begin at noon as early as the second part of May. That water, however, is imbibed by the lower lying snow covers. The intensive melting starts by the end of May and lasts till the end of June. The snow covering steep, south and south-west exposed slopes disappears first. Then it melts from the less inclined and less favourably exposed slopes. In the upper, steep and rocky parts of slopes dispersed flow occurs. Slightly lower on still steep but even slope the flow acquires a linear form and produces parallel grooves running down the slope. If the slope is covered with thick debris water flows along the spaces

Fig. 1. Hornsund. Surface of the Fuglebergsletta colluvial fan in spring 1959. Desiccation polygons in the background are still covered with a fresh accumulation. Water occupies all the hollows. Photo Z. Czeppe.

Fig. 2. Hornsund. Fuglebergsletta. Erosion proceeding along the fissures. Photo Z. Czeppe

free from snow and ice between or under the blocks. In these channels water may sometimes reach a considerable velocity. Forced to the surface by some sort of blockade it is strong enough to displace large blocks around the outlet. Below it an ejection cone is built. The subsurface flow is observed also on slopes covered with vegetation. Here too devastation around the outlet of the underground channel suggests the great energy of the water stream. The underground channels cleared of the smaller rock fragments become an easy way of flow and serve from year to year. Their position is often marked by a slight depression of the surface of the slope. In spring these depressions are filled up with long and narrow patches of firn which secures them from the surface flow erosion, but subsequently nivation processes tend to sharpen their outlines. Among the numerous parallel grooves etching the mountain slopes between Rotjesfjellet and Gulliksenfjellet there are also those created by suffosion (subcutaneous erosion) in various stages of development. Neither these, however, nor the surface erosion rills develop into valleys. It seems that vertical erosion in all of them is possible only in early spring, when there is a sufficient quantity of meltwater. But as soon as they achieve a certain depth further deepening is stopped by the accumulated snow, which at the crucial time keeps them well frozen and filled up. Later on they carry some water from their own snow fields, but there is not much of it. Nivation processes steepen their sides and widen them slowly. Büdel [3] states that rill-erosion (*Runsenabspülung*) causes the parallel retreat of slopes. In that case one would expect to find sharp-edged interfluves between the rills. They are found only on the naked, very steep slopes built of the rocks that weather into fine debris. They are anything but common.

The retreat of slopes is therefore rather dependent on the surface and subsurface wash of a less organized type.

On the lower part of the slope, especially where it is covered with fine waste or vegetation, the flow assumes a sheet form. The sheet flow is also bound to the mildly inclined hillsides and nearly flat surfaces of colluvial cones and marine terraces. The flow occurs here while the ground surface is still frozen. Water brings and deposits a fair amount of fine waste washed off the higher parts of slopes as

Fig. 3. Hornsund. Fuglebergsletta. Stones and packets of soil are falling into a fissure. Photo Z. Czeppe

well as the material blown off the snowless rocky surfaces, terrace edges and hill tops by the autumnal and winter gales and imbedded in the snow cover. In melted snow samples there is up to 2 g of mineral sediment per one litre of water. All these deposits are spread and left on top of various, often coarse materials building the surface of the ground. In that way a sedimentary soil profile is built in reverse to that of the frost sorting type, where the coarse material rests on top of the fine.

Such a soil profile is found on the large colluvial fan at the foot of Fugleberget. Spring sheet-flood ends with water occupying all the indentures of the surface. These are often situated around bigger blocks. Wandering along the maze of

Fig. 4. Hornsund. Fuglebergsletta. A sorted net streatching into sorted stripes. Photo Z. Czeppe

hollows water carries off the fine sediment and therefore its ways are marked by thicker rubble. As unfreezing progresses downwards in the soil profile, water deepens the channels. This is done very rapidly along the fissures which are often found running from one protruding block to another. These are soon changed into narrow rills into which packets of mixed material fell from the banks. Fine materials are soon carried away and stones collect in the rill. The irregular patches of higher ground built mainly of fine sediments give out an excess of water and contract, breaking into smaller and more regular roundish forms. The resulting fissures are soon occupied by water and enlarged. The water net acquires a more regular pattern which changes slowly into a sorted net.

Where the declivity of slope achieves about 2° the water action is clearly stronger in the direction of slope than along the contour lines. The whole water net tends to stretch out in the direction of stronger flow. With a slightly greater declivity it changes into parallel streams creating the pattern closely resembling some sorted stripes. These essentially water-created forms undergo subsequent retouching by autumnal multigelation.

Forms similar to sorted circles of erosional origin were found on the left side terrace of the river Revelva below the lake. There are flat-bottomed gaps cut into the terrace edge. They were probably used some time ago by the streams coming down from Skoddefjellet. The flat bottom of one of these gaps, about 8 m broad, is dissected by a net of criss-crossing channels, less than 1 m broad. Between them there occur "islands" up to 0·5 m high, circular on level ground, oval on inclined slopes, with a maximum breadth of about 1 m and length of about 2 m. They were cut out of the terrace gravel and sand. Usually their surfaces are covered by gravel for the finer material is washed away. This gravel cover slides down the slopes of the hillocks and builds a ring around the core, which is much more sandy. The whole form is often very similar to the sorted circle. On the lower terrace of the river these forms join the swarm of "normal" sorted circles and it is difficult to draw a dividing line between them.

When by the end of the spring thaw sheet-flood changes into linear flow all sorts of hollows, polygonal fissures, sorted nets, stripes and similar lines attract and carry water. Some of them are changed into permanent water courses. On the moss covered surfaces of marine terraces one can see little brooks flowing along a zigzag course following the fissures of the tundra polygons. Their beds are usually incised into pit and only on the edges of terraces is erosion more active.

Water deposition may create necessary conditions for the action of frost processes. A layer of torn off vegetation deposited on an overgrown ejection cone of a torrent flowing down from the glacial corries of Torbjornsenfjellet, became an additional insulator and caused the growth of a hydrolaccolith. The deposition of layers of fine sediments on the surface of old solifluction tongues and sorted steps takes place commonly every spring.

Thaw water on the slightly inclined, naked surfaces collaborates with frost in the creation of micro-stripes. Grooves filled with small stones are 3 to 4 cm broad and about 5 cm deep. The separating ridges are often slightly broader. They are arranged into parallel, discontinuous lines running down the slope but curving round obstacles. They are found on short slopes inclined at about 15° which have a very thin snow cover in winter. In spring at noon the surface thaws and soon freezes again. Fine soil between the little stones swells up achieving a spongy structure and a cauliflower-like appearance. It does not settle down on thawing because there is too little water, which tends to run down towards the small depressions occupied by stones. It directs itself from one to another down the slope. The slight depth of thaw, small amount of water, and short periods of thawing do not allow anything but little grooves to develop in which the thicker rock fragments collect. The tiny streams of water pass by even small obstacles and often disappear after a short run. Therefore the micro-stripes are deprived of the continuity and straightness of the big sorted stripes.

Large quantities of meltwater find their way to the river valleys covered at the time with snow and ice. Along the valleys water runs down but it cannot join the sea because the outlet has been completely blocked up by a storm ridge built up in autumn. Meltwater collects behind the bar flooding the immediate neighbourhood. Its level rises up till the frozen barrier is overflowed with numerous cascades. The one that first cuts into the ridge and drains the lake becomes the outlet of the river for the coming season. The opening of a narrow way in the storm ridge results in a rapid discharge of the barred water and a scour along the line of greatest velocity. The channel thus eroded in the broad and generally flat river bottom can be traced for some distance up stream. The higher stage of the river holds on till the end of the main thaw period, that is untill about the end of June or beginning

Fig. 5. Hornsund. A little brook crossing the moss-covered surface of a marine terrace. Photo Z. Czeppe

Fig. 6. Hornsund. Micro-stripes created by the collaboration of micro-multigelation and micro-erosion. Photo Z. Czeppe

Fig. 7. Hornsund. One of the barred-lake cascades overflowing the snow-flanked storm ridge. Photo Z. Czeppe

of July. At that period gravel is carried along the line or lines of current. Lateral erosion is visible only in the concave sections of meanders. It is helped by nivation processes connected with the firn patches lying in the shadow of the river banks. These processes act slowly and do not alter the cross-section of the river bed to a considerable degree for some years at least. The motor vehicles of the Polish Expedition used the same crossings for four summers in succession.

In autumn the river is insufficiently supplied with water and slowly freezes thus having no power to fight with the rapidly built storm ridges and its mouth becomes sealed again. Next spring the river may cut an outflow channel in quite a different place. The annual changes of the position of the river outlet result in the broadening of the lowest section of the river valley. It remains, however, separated from the sea by a storm ridge and connected to it by a narrow channel only. All the rivers of south-west Spitsbergen, from river Vimsa north of Hornsund to the river Luktvassebreia in Sörkappland show this type of "bottle-neck" outlet.

The facts described above do not agree with the thesis according to which gravel in periglacial rivers is transported on the whole breadth of the river bottom [3]. It may be true for the big rivers which have enough power to wipe out the barrier completely. But even the biggest river of the described section of the coast, Bungebreia, in late October still had an underwater bar checking its outflow.

Activity of rain water is much less important for the development of the relief han that of the meltwater on the account of both its much smaller total amount

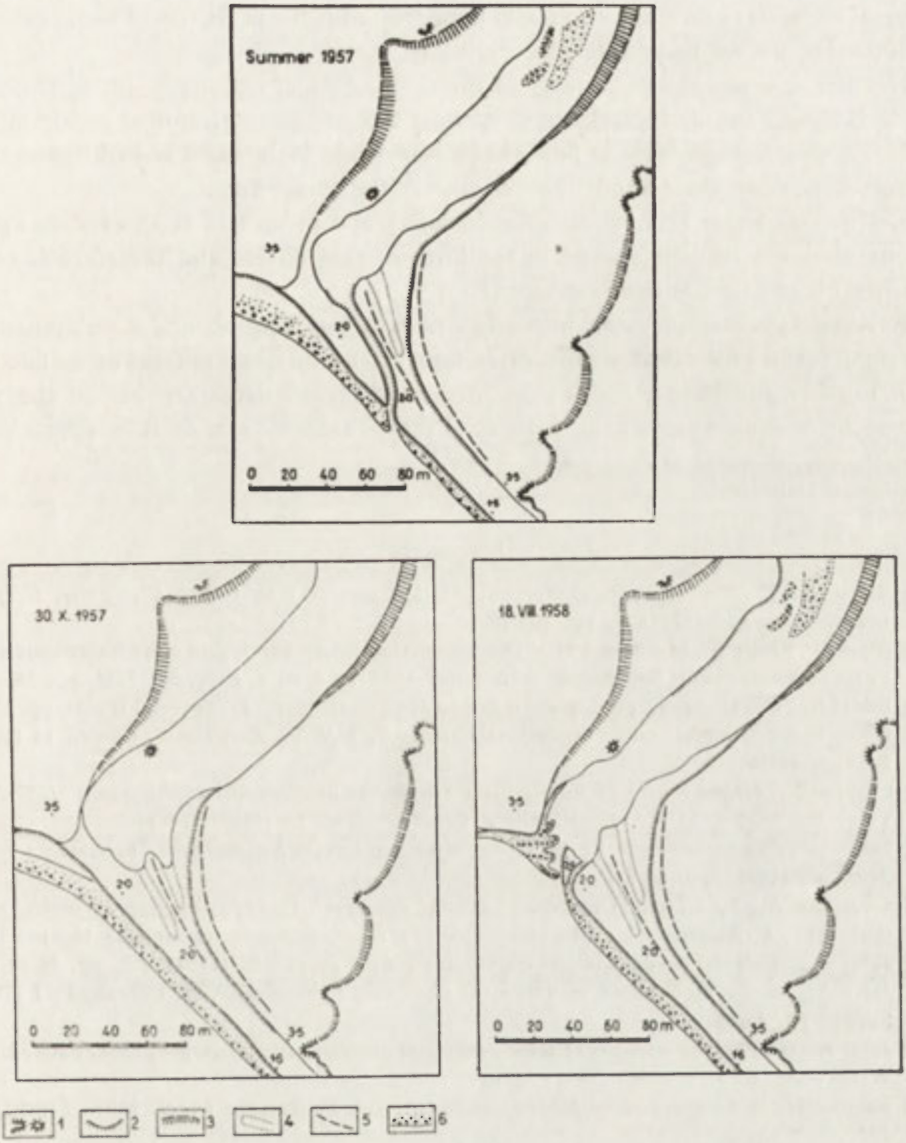

Fig. 8. Hornsund. Changes in the outlet of the river Revelva

1 — rocks, 2 — undercut edges, 3 — steep slopes, 4 — traces of old channels, 5 — older strom ridges, 6 — fresh accumulation of gravel

and its distribution over time and supply in small quantities. It is mainly limited to washing off the rock surfaces the sediment left by the vanishing snow. It may carry away fine material leaving coarse material on the surface of the ground where the frost segregation could not have achieved this. Rain of longer duration may wash some waste down the slopes, but it does not seriously influence the water stage of the rivers and it has no greater importance for the processes of linear erosion.

Summing up we may state the following:

1. There is a marked periodicity of the running water activity resulting from the winter storage and spring release of at least 2/3 of the total annual precipitation.

2. The greatest intensity of flow is achieved while the ground is still frozen. The dispersed flow or sheet-flood dominates over the linear flow.

3. Running water acts mainly as a denuding and much less as an eroding agent. Accumulation is usually spread in the form of thin covers and therefore it easily escapes observation and measurement.

4. Running water plays an important part in the creation and development of the small forms of the tundra relief often regarded as the result of frost or solifluction.

5. Further qualitative and quantitative studies are necessary before the part played by flowing water in the periglacial morphogenesis can be fully appreciated.

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