POLSKA AKADEMIA NAUK
INSTYTUT GEOGRAFII I PRZESTRZENNEGO ZAGOSPODAROWANIA
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HOLOCENE AND LATE VISTULIAN PALEOGEOGRAPHY AND PALEOHYDROLOGY

Edited by Adam Kotarba

This volume is dedicated to Professor Leszek Starkel to honour his 50 years of scientific achievement

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PROFESSOR LESZEK STARKEL
50 YEARS OF SCIENTIFIC ACTIVITY

In the fall of 1953 the Laboratory of Geomorphology and Hydrology of Mountains and Upland Areas was established in Cracow, as a unit of the Institute of Geography, Polish Academy of Sciences. Leszek Starkel became its first research assistant. At that time, one of the major tasks undertaken by Polish geographers was the elaboration of large-scale geomorphological maps of Poland. This is how Professor Starkel's great scientific adventure, which lasts till today, has begun. Still as a graduate student at Jagellonian University, he carried on geomorphological mapping as part of his master theses, under the advisorship of Professor Mieczysław Klimaszewski, the author of the map concept. The map sheet of Dębica and Trzciana, compiled by Leszek Starkel, proved to be an essential source of information on contemporary relief in the piedmont area of Flysch Carpathians. It was published in a multicoloured form. Since then geomorphological mapping has remained an important instrument used in his subsequent studies which unfolded over a long period. These included the Lesko sheet of the geomorphological map of Poland, as well as numerous other studies on Poland, India (Darjeeling Himalaya) and Mongolia (southern Khangai).

The Carpathian Mountains have always constituted a major realm, and often a point of reference in the field studies which have attracted Professor Starkel to a number of mountainous areas of the globe. His analysis of the contemporary relief of the Polish Carpathians has represented the basis for generalizations concerning the role of the Holocene in shaping observed land morphology. Against the general preoccupation of the international community of geomorphologists with glacial and periglacial processes, Professor Leszek Starkel was the first to emphasize the impact the last 10 000 years have had on the evolution of present landforms. Years later, he wrote: "I felt, I was at a threshold of uncovering a kind of a mystery concerning this period". *(My encounters with the Carpathians, 1999, pp. 33).* An effect of detailed, fieldwork based research, was a comprehensive monograph on *The Evolution of Land Relief in Polish Flysch Carpathians during the Holocene,* which appeared in 1960. This volume set a new strand in geomorphology, one that has been followed by a number of scholars and research teams. Among the most frequently quoted works of Professor Starkel is a synthetic study on: *Post –*

glacial climate and the moulding of European relief, published in London as a chapter of a book which originated from the symposium on global climatic change during the period of 8000–0 years B.P. Another important publication in this domain was a study presenting a model of morphology of alluvial fans in the Carpathian piedmont, included in the fundamental *Encyclopedia of Geomorphology*, edited by R. Fairbridge. These studies generated multiplier effects, as they inspired numerous geomorphologists and Quaternary geologists to develop further our knowledge of the mechanisms of relief transformation, of phases and the scale of processes that are responsible for shaping the natural environment in mountain and upland areas.

The visit to India in 1968, in the framework of the International Geographical Congress, and subsequently the symposium of the IGU Commission on Periglacial Geomorphology in Darjeeling, was the event that precipitated Professor Starkel's long–lasting interest in contemporary geomorphological phenomena of extreme character. His analysis of the effects of catastrophic rainfalls, which was published in *Geographia Polonica* (vol. 21, 1972), is a study widely quoted in the international scientific literature. It has also encouraged a number of younger scholars to develop this strand of research, based on observations and data collected for mountainous regions in various parts of the world. The evaluation of mechanisms and effects of extreme natural events now constitutes an important research theme in the context of studies on global climatic change.

Numerous subsequent visits and study tours in India have brought a series of more recent publications, among them a book on: *Rains, Landslides and Floods in the Darjeeling Himalaya*, edited by L. Starkel and S. Basu, published by the Indian National Science Academy in 2000.

In the 1970s Professor Leszek Starkel inspired a new research programme which brought together a number of geomorphologists, Quaternary geologists, paleobotanists and other specialists, such as archeologists and physicists, who studied changes of Poland's natural environment that occurred in the transitional period, extending from late Pleistocene to Holocene. This programme, entitled: *Paleogeography of the temperate climatic zone over the last 15 000 years*, was formally carried on within the framework of INQUA and the UNESCO IGCP Project No 158. It generated a number of interdisciplinary studies. In Poland, a series of six volumes appeared on: *Evolution of the Vistula River Valley during the last 15 000 years*, edited by L. Starkel. A synthetic volume presenting results of multifaceted research on the evolution of the Vistula valley system was written by L. Starkel and published in 2001.

Professor Leszek Starkel has led and co-ordinated a number of interdisciplinary research, as well as publication projects. One of the latter was a monograph on: *Geography of Poland: the natural environment* (in Polish; 1991), to which L. Starkel provided the general concept and contributed several chapters. It is still the most comprehensive source of knowledge on Poland's nature seen in the spatial perspective.

These are only some of Leszek Starkel's numerous fields of research interest and activity. Their scope is truly global, though in his studies on the functioning of geosystems and ecosystems in various climatic zones, the Carpathian Mountains have occupied a special place. So was also his lasting involvement in the work of the Institute. For more than 33 years he was head of the Department of Geomorphology and Hydrology of Mountains and Uplands in Cracow, the strongest research unit of the Institute of Geography and Spatial Organization. There, as well as at the Institute's Research Station in Szymbark, which he organized in 1966, a series of dissertations were produced, under His guidance, focusing, among others, on the circulation of matter in slope and valley ecosystems; overland flow and slope wash, chemical denudation, slow and rapid mass movements, and fluvial process activity. His numerous travels, and vast international contacts reflect his special interest in comparative research and synthetic studies. One of their outcomes is the book on *Paleogeography of Holocene,* published in 1977.

While carrying on the research work, writing, editing and teaching, Professor Leszek Starkel has still been able to find plenty of time, energy and enthusiasm for organizational activity, both at a national and international level. Among other tasks, he chaired the Polish National Committee of IGBP, the Committee on Quaternary Research of the Polish Academy of Sciences, was Vice Chairman of the Cracow Branch of the Polish Academy of Sciences, member of the Section on Biology and Earth Sciences of Poland's National Committee on Scientific Research. On the international scene, he served as Chairman of INQUA Eurosiberian Subcommission on Holocene, of its Continental Paleohydrology Commission, as well as of UNESCO IGCP 158 Programme. His activity has been vital for the development of comparative studies and map projects in East-Central Europe, within the framework of the Carpathian – Balkan Geomorphological Commission.

The contribution of Professor Leszek Starkel to the progress in science, and to the development of scientific institutions and organizations has been widely acknowledged. In Poland, he is Ordinary Member of the Polish Academy of Sciences and Active Member of the Polish Academy of Sciences and

Arts. Abroad, he is a member, and an honorary member of Academia Europaea, the Academy of Natural Sciences in Sweden (Lund), the British Quaternary Association, the Slovak Geographical Society, the Hungarian Geographical Society, the International Association for Quaternary Research.

Today, with 50 years of experience in scientific work, and a long record of scientific accomplishments, Professor Leszek Starkel is as effective as ever in carrying on research, writing, and participating in field studies. Just when these words are being written, he is with the INQUA Congress in Reno, Nevada. We wish him the continuation of his busy activity and further achievements in disentangling of scientific questions, the new and the old ones.

The present volume on: *Holocene and Late Vistulian Paleogeography and Paleohydrology* brings together a collection of papers by long-time friends, colleagues, and students of Professor Leszek Starkel. A parallel, special issue of *Geographia Polonica* features contributions on: *The Role of Catastrophic Events and Human Activity in the Transformation of Landscapes*, by both Polish and foreign authors, who also dedicate their papers to Professor Leszek Starkel, on the occasion of the 50 Years anniversary of His scientific work.

Piotr Korcelli and Adam Kotarba

CURRICULUM VITAE

Professor Leszek Starkel was born in Starachowice, Poland on September 8, 1931. He graduated in physical geography at Jagellonian University in Cracow in February 1954. He obtained a doctorate in geomorphology at the Institute of Geography, Polish Academy of Sciences in 1959 and later habilitated his doctor's degree in 1964. The title of extraordinary professor was granted to him in 1971 and ordinary professor in 1979. In 1983 he was elected the corresponding member of the Polish Academy of Sciences and 15 years later its full member.

In November 1953, he was appointed as the assistant in the newly formed Department of Geomorphology and Hydrology of the Institute of Geography PAS, located in Cracow. Having attained all the levels of scientific career, he was the head of Department from May 1968 till November 2001. Since January 2002 he has been retired but is still employed for a half-time position in the Department.

Apart from that he established in 1960 the Research Station in the Carpathians in Szymbark and between 1980 and 1990 he was the leader of the state research programme on the Evolution of environment of Poland, which involved the collaboration with all physical geographers in Poland.

The research activity of Professor Starkel has focused on the following fields:

1. Geomorphological mapping (the author of detail maps and editor of general geomorphological map of Poland 1:500 000).
2. Relief evolution of the Polish Carpathians (monographs of the Polish Flysch Carpathians).
3. Holocene palaeogeography of the Polish Carpathians, Poland and the whole Europe.
4. Evolution of the river valleys during the last 15 000 years (the author and editor of several monographs)
5. Palaeohydrology of the late Quaternary encompassing Poland, temperate zone and the globe (the co-editor and co-author of monographs on Temperate Palaeohydrology and Global Palaeohydrology).

6. Contemporaneous geomorphic processes (especially the role of extreme events in the evolution of landscape in the temperate climate and in monsoonal climate, with Darjeeling Himalaya and Cherrapunji region as examples).

7. Typology, regionalization and evaluation of natural environment (including the Polish Carpathians and the whole Polish territory – the editor of great monograph Geography of Poland – natural environment).

8. Latitudinal and vertical zonality of geoecosystems in the continental climate of Asia (Khangai Mts in Mongolia).

ORGANIZATIONAL ACTIVITY BEYOND THE INSTITUTE IN POLAND

Professor Starkel was a chairman of the Quaternary Research Committee of PAS in years 1979–1992 (up to now he has been a chairman of the Polish National Commiittee for INQUA). Between 1989 and 1999 he was a chairman of Polish IGBP Committee (now he is its deputy chairman). In 1990 he was also a deputy chairman of the Committee: Man and Environment PAS. Since 1983 he has been a member of the Executive Committee of Mountain Economy PAS. In 1996 L. Starkel was nominated as the Secretary of the Cracow Branch of Polish Academy of Sciences and later elected its deputy president (1999–2002). Between 1997–2000 he was an elected member of the State Committee for Scientific Research.

INTERNATIONAL SCIENTIFIC ACTIVITY

Prof. Starkel's activity is connected with several international organisations.

Since the late 1950s he has been active in several geomorphological commissions of the International Geographical Union: on Geomorphological Mapping, Periglacial Processes, Slope Evolution, Present Day Geomorphic Processes and GERTEC.

After foundation of the regional Carpatho-Balcan Geomorphological Commission (1962) he was its secretary during the first decade and the editor of Studia Geomorphologica Carpatho-Balcanica (1967–1991).

Most of his international activity is connected with the International Union for Quaternary Research (INQUA). In its Holocene Commission he chaired the Eurosiberian Subcomission (1973–1981) and later the working group on Human impact on soil erosion (1981–1988). In 1991 he was the founder and first president (1991–1995) of the newly formed INQUA Commission on Global Continental Palaeohydrology. During various intercongress periods he also cooperated with other INQUA Commissions on Palaeoclimate and Carbon Cycle.

Between 1978 and 1998 together with B.Berglund from Sweden he organised and chaired the IGCP project 158 on the Palaeohydrology of the temperate zone during the last 15 000 years, leading the fluvial subproject.

In the 1990s he was an active member of ESF programme "Palaeoclimate and Man", led by B.Frenzel.

Finally, he was a Polish delegate to the Advisory Board on Environment and Sustainable Development of the 5-th European Research Framework Programme (1999–2002).

During his 50 years of activity Prof. Starkel participated in more than 100 international congresses and workshops in 37 countries, including 7 IGU, 9 INQUA, 2 Geological and all the 5 International Association of Geomorphologists congresses (invited twice to present a plenary lecture).

He carried out field studies in various parts of Asia (India, Mongolia and China) and Europe (Romania, Bulgaria, Georgia and Ukraine).

DIDACTIC ACTIVITY

Prof. Starkel promoted 9 doctors and his three younger colleagues from the Department received the title of professor.

TITLES AND DISTINCTIONS

Apart from the membership of the Polish Academy of Sciences, Professor Starkel was elected the active member of the Polish Academy of Sciences and Letters in Cracow (1989), foreign member of Royal Physiographic Society in Lund, Sweden (1985), member of Academia Europea (1992). He was given D.Linton Award by the British Geomorphological Research Group (1985) and Loczy Medal by the Hungarian Geographical Society (1995), He is an honorary member of INQUA (1989), Slovak Geographical Society (1999), Hungarian Geographical Society (1993), Polish Geophysical Society (1999) and Association of Polish Geomorphologists (2000). Prof. Starkel was also awarded with the Knight's and Officer's Cross of Polonia Restituta.

Photo 2. Field investigations in Vistula Valley with E. Latrubesse from Bresil and Mrs. E. Czyżowska (2000)
Badania w dolinie Wisły z dr. Latrubesse z Brazylii i mgr E. Czyżowską (2000 r.)

Photo 3. Field symposium in Japan of Commission on Palaeohydrology 1997 (with Prof. A. Yamskich)
Sympozjum terenowe Komisji Paleohydrologii w Japonii (1997 r.) (razem z prof. A. Yamskich)

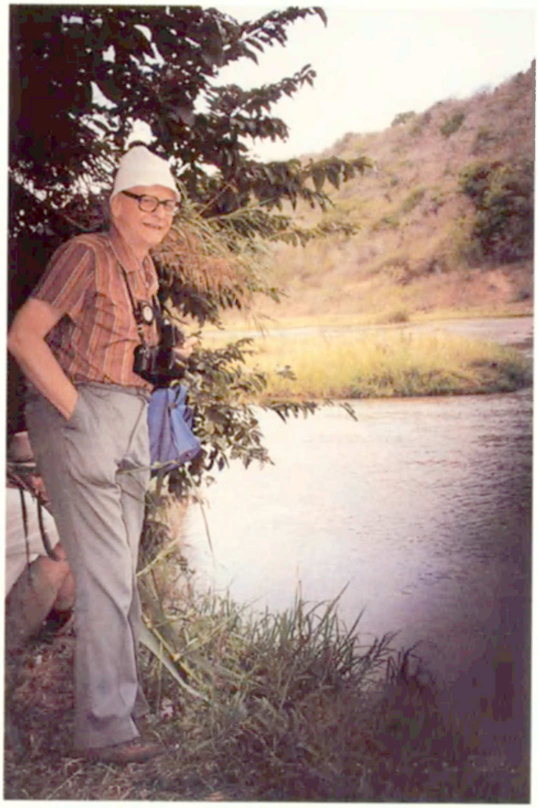

Photo 4. Field symposium in Natalu – Congress INQUA (1999)
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Photo 5. Field investigations on Araguaya river in Bresil (1999)
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Photo 6. Exploring the terrace of Seim river near Kursk (2000)
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Photo 7. Symposium of Commission on Palaeohydrology over Yenisey river (2001)
Sympozjum Komisji Paleohydrologii nad Jenisejem (2001 r.)

SPIS WAŻNIEJSZYCH PUBLIKACJI PROFESORA LESZKA STARKLA

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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Józef Edward Mojski

SELECTED PROBLEMS OF THE INTERGLACIAL RIVER VALLEYS INVESTIGATION IN THE POLISH PLAIN

The following introductory remarks are of principle significance for the study of the origin and evolution of the Pleistocene river valleys in the territory of Poland:

– the study of the greatest interglacial valley is of primary importance;

– it is clear, that we must know the location of the final erosion base i.e. marine sea level;

– in some places the syn- and postsedimentary neotectonic movements are of high importance for the present-day hipsometric state of the interglacial river valleys;

– the average thickness of the Quaternary cover in Poland is mainly from 50 to 200 m. Therefore for the all paleogeomorfic reconstructions a considerable density of boring profiles is necessary.

1. During the each Pleistocene interglacial the erosional base had existed mainly outside of Poland's territory. Such a base had been the North Sea Basin in the west, the Black Sea Basin in the south-east and sometimes the Baltic Sea Basin (Fig. I). As yet we have rather little data available concerning Mazowian (Holsteinian) and Eemian Interglacial Seas, but for older interglacials there is a complete lack of data (Fig. 2, Tab. I). Only the knowledge of the Holocene sea is sufficient (a.o. Mojski, Ed. 1995). The erosional bases of all others (up to 5, Lindner et al. 2002) are as yet completely unknown.

The extent of the Holsteinian Sea in the northern Europe is very well known. During the maximum expansion the sea covered a significant part of German Lowland, from NW up to Berlin. Further to east there existed a widespread limnic basin now called the "Paludina Bank" lake. The eastern boundary of that basin was recently studied in the western part of Polish Plain (Skompski 1994). The top of limnic deposits in this part of Poland lies now at 20 m a.s.l. (Fig. 1). The traces of the Holstein Sea have been found lately in the most NE part of Germany (von Bülow 2002) with the top of adequate deposits at 0 m. A similar hights (–20 m b.s.l.) concern the Holsteinian (Mojski 2002 = Sztum Sea, Makowska 1982) marine deposits at the lower Vistula River

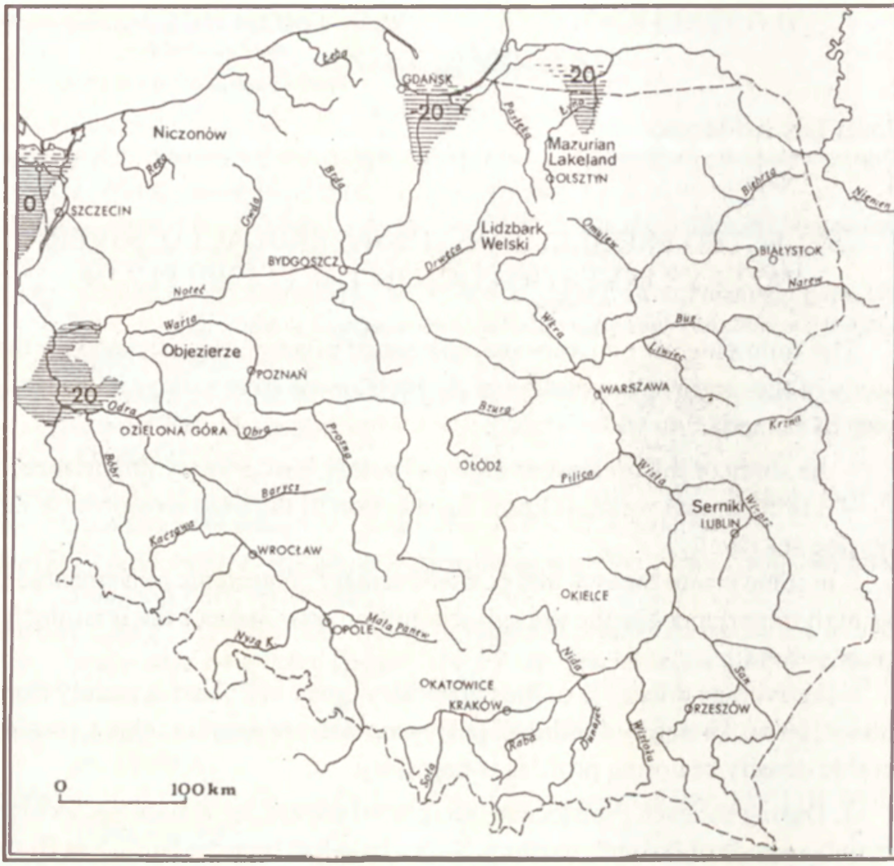

Fig. 1. Occurence of seas as the erosional base for Mazowian (Holsteinian) rivers. (At West – von Bülow 2002; at North, at Lower Vistula river valley – Makowska 1982, Mojski 2002; at Lower Łyna river valley – Kondratiene, Evtuchin 1974)

Występowanie mórz erozyjnego pochodzenia na Mazowieckich (Holsztyńskich) rzekach. (na zachodzie – von Bülow 2002; na północy, w dolinie dolnej Wisły – Makowska 1982, Mojski 2002; w dolinie dolnej Łyny – Kondratien, Evtuchin 1974)

area and the marine Holsteinian (= Likhvinian in Russia) at Domnowo (Kaliningrad District, Kondratiene, Ertuchin 1974, Fig. 1). So, the present day level of the Holsteinian marine deposits top lies between –20 and 20 m in relation to the present sea level, namely higher in the west (North Sea Basin) and lower in the north (Baltic area). It is evidence for a rather small dimension of neotectonic movements in the broad zone of W and N part of Poland, starting from Mazowian Interglacial, i.e. for the last 0,4–0,3 Ma.

The marine deposits of the last, Eemian Interglacial occur very often along the present-day coast of the North Sea and the Baltic Sea (in Poland, Makowska 1982). In both cases their top lies about 0 m, or a little less. It means that neotectonic movements in NW part of Poland during the last ±0,1 Ma were insignificant.

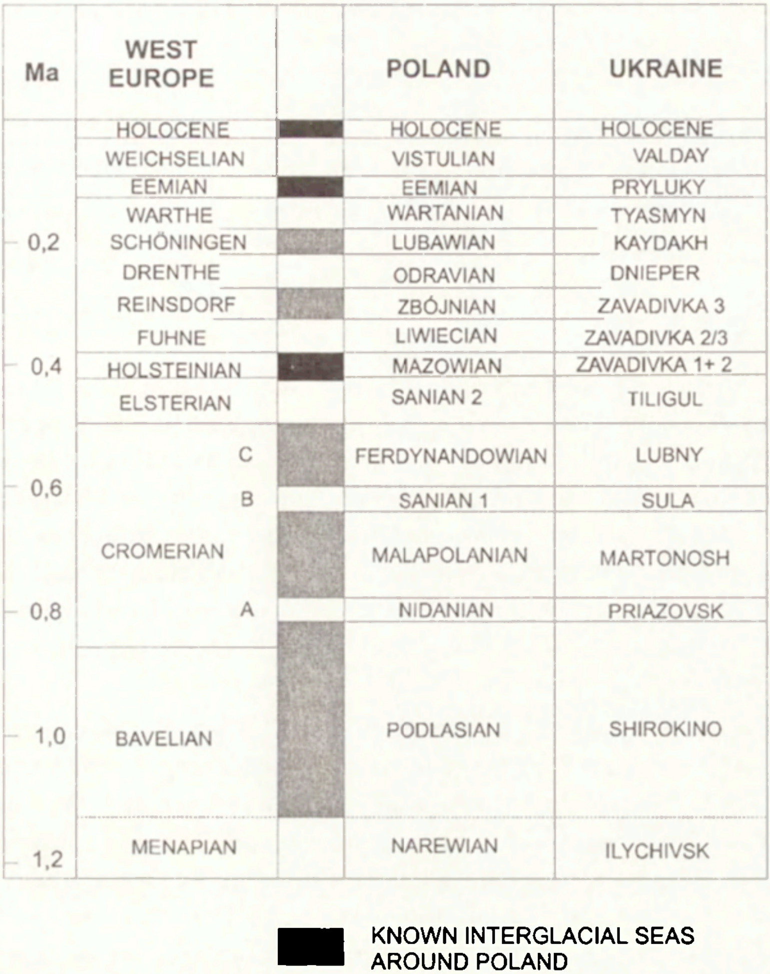

Fig. 2. One of the newest stratigraphic Quaternary subdivision in Poland and adjacent areas (Lindner et al. 2002) and stratigraphic position of interglacial seas around Poland

Nowy podział stratygraficzny Czwartorzędu w Polsce i sąsiednich terenach (Lindner i in. 2002) oraz stratygraficzna pozycja interglacjalu mórz blisko Polski

Diminishing importance of the Black Sea basin as the erosion base for Polish rivers took place during the whole of the Quaternary. The early Mesopleistocene (>0,5 Ma) rivers in Poland drained into the Tchauda Sea (Cromerian, Tab. 1) more than half territory of Poland, but during the Eemian (Karangat Sea) only 10–15 %, and now, in the Holocene (Black Sea) absolutely nothing.

2. The Holocene, namely the interglacial evolution of the Vistula River valley has been studied in detail for the last 20 years by a number of Polish geographers and geologists guided by L. Starkel. Lately L. Starkel (2001)

Table 1. Age correlation of interglacial seas in NW Europe, SE Europe and Baltic area (acc. to Mojski 1993)

NW EUROPE (NORTH SEA BASIN)	AGE INKa	BALTIC SEA BASIN	SE EUROPE (BLACK SEA BASIN)
FLANDRIAN	7,0–0	LITORINA SEA	UPPER EUXINIAN SEA
EEMIAN	125–110	TYCHNOWY SEA	KARANGAT SEA
HOLSTEIN	400–300	SZTUM SEA (DOMNOWO BEDS)	EUXINIAN UZUNLAR SEA
CROMERIAN SEAS	>500	?	TCHAUDA SEA

published a synthesis concerning the origin and evolution of the Vistula River valley during the last 15 ka. It is very important base for the recognition and a determination of lithological and facies features as well as for exemination of the interglacial thickness and fluvial deposits in the country. The results of investigations (Fig. 3) show that the average Holocene deposit thickness is 8,7 m, with increase up to 20 m maximum for small but deep basin-like parts of valley in the former glaciated areas of Poland (e.g. Unisław Basin). The broader parts of valley floor have thickness of Holocene deposits that is only 4–5 m (e.g. Warsaw Basin).

The Holocene fluvial deposits of the Vistula River (Fig. 4) consist of mineral-organic facies. They are mainly finegrained, except for the Carpathians. Initially the meandering type of channel prevailed, and later the deposits of braided river deposits were built up in some places. The above statements are of basic importance for the fluvial deposition in the largest valleys in the country during all interglacials. But, of course the Holocene is not yet finished as an interglacial. Therefore to the average thickness (8,7 m) we must add 2–4 m, as depositional equivalent of the future, younger part of each interglacial. Such a calculation is based on the of fact, that the Last (Eemian) Interglacial lasted 15 up to 20 ka.

The above described features of the Holocene (interglacial) fluvial deposits are a pattern for the recognition of the Pleistocene interglacial river deposits and of corresponding fluvial topography in the Polish Plain.

3. There are possible three main genetic interpretations of the great buried valleys which existed in the Pleistocene cover of Polish Plain

– The fluvial (rivers) valleys.

– The subglacial channel.

– The ice-marginal valleys ("pradolina").

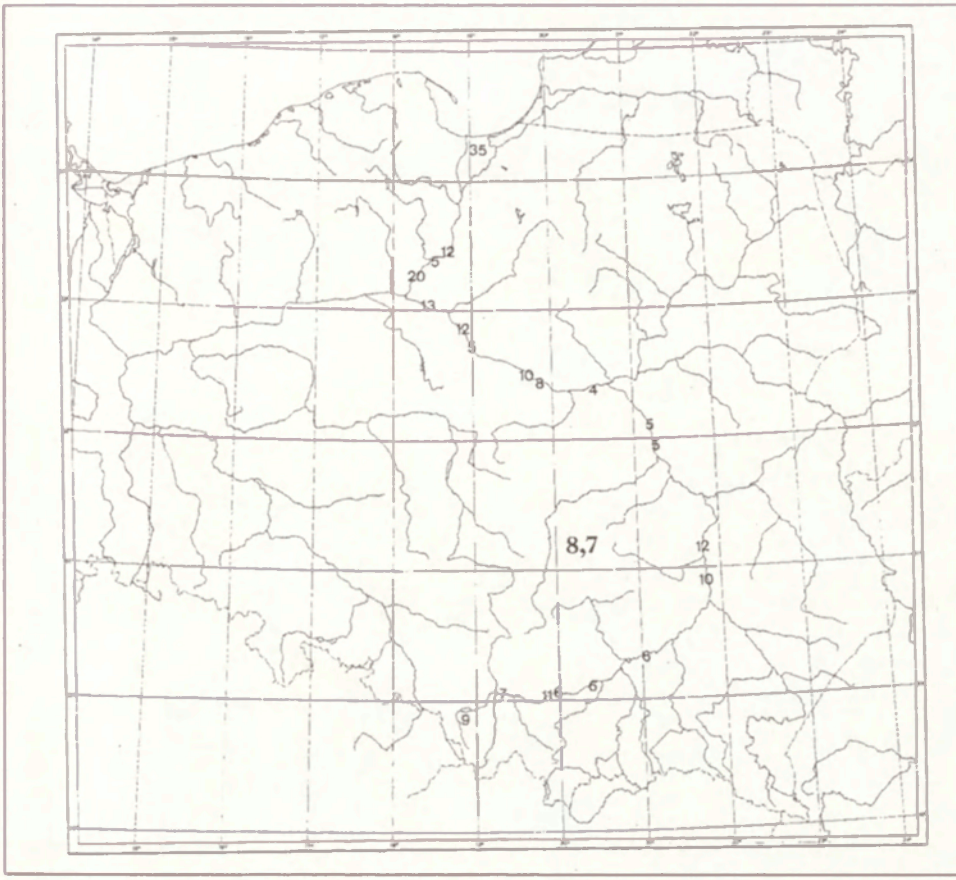

Fig. 3. The thickness (in m) of Holocene deposits in various part of Vistula River valley (acc. to Starkel 2001). 8,7 – average thickness in m. of Holocene deposits in Vistula River valley

Miąższość (w m) holoceńskich osadów w różnych częściach doliny Wisły (wg Starkla 2001). 8,7 – średnia miąższość w metrach holoceńskich osadów w dolinie Wisły

The last possibility is rather a rare case, and very difficult to recognize. The buried ice-marginal valleys are located mainly parallel to ice-sheet front and they have W–E direction. They are broad and infilled with fluvial deposits in part only.

The fluvial valleys and subglacial channels are similar in their direction and depth, and they have very often a similar paleorelief which is visible in various parts of the Pleistocene cover.

But in some cases the differences are significant. Anyway the subglacial channels have longitudinal profiles with uneven bottom, rather small breadth and their infilling consists of glacifluvial mineral deposits, mainly sand and gravel with thin redeposited till lenses. At figure 5 as an example of subgla-

Fig. 4. Holocene valleys and greatest closed basins in Poland
Holocenskie doliny a wielkie zamknięte baseny w Polsce

cial channel there are represented both the Niczonów and Objezierze cross-sections. Niczonów was interpreted (Kopczynska-Zandarska 1970) as a river valley of Mazowian age with the valley floor lying at 150 m b.s.l. However, the author of the paper disagrees with fluvial valley origin interpretation. It is most probably rather a subglacial channel from one of the older glaciations (Sanian 2 ?).

Objezierze (Skompski 1994) is an example of subglacial channel of Vistulian age completely infilled by glacifluvial deposits of the same age, and about 160 m thick! This channel is visible in the present glacial topography. It is a typical subglacial tunnel valley according to description of Kozarski (1966/1967) and Niewiarowski (1995).

Unquestionably major Pleistocene river valleys in the Polish Plain are a prolongation in the northern direction of such a valleys developed on the northern part of the Mid-Polish Uplands, where they are very well recognised and investigated (the valleys of Odra, Warta, Pilica, Vistula, Wieprz and Bug Rivers). A good example is the present-day valley of Wieprz River in the northern foreland of the Lublin Upland (Serniki cross-section, Fig. 5). The age is here documented by limnic sequence of Mazowian age, which covers 120 m thick fluvial and glacifluvial sequence with beds of till older than Mazowian Glaciations (San 1 and San 2 ?). But the shape of the valley at Serniki is fully comparable with the shapes of subglacial channels at Niczonów and Objezierze. Such a comparison (Fig. 5) shows the difficulties with the river valleys reconstructions in the Polish Lowland.

In the same area (foreland of the Lublin Upland!) occur invisible, or poorly visible in the present-day topography buried, deeply and narrowly burried subglacial channels originated probably during the Odranian Glaciation or during the older cold stages. An unequivocal example is subglacial channel in the northern foreland of Lublin Upland (Harasimiuk, Henkiel 1981), 30 km east of Lublin. It is incised 100 m deep in the Upper Cretaceous karst topography. Similar subglacial channels are very often found at the Sudetic Foreland (Michniewicz 1998).

The similar difficulties with the distinction between river valleys and subglacial channels, occuring in some regions, are connected with small amount of bore-holes profile and with later shape of their disturbances of ones. Vertical movements of tectonic origin as well as glaciotectonic processes are of great importance. The first are very common, but they are differenciated in space and time. The records from recently obtained data (Liszkowski 1982) show, that the territory of country have the amplitude between 1and –1 mm (newest data to –2 mm pro year, Kurzawa 2003), namely between 100 m (eastern Poland) and –100 m (western Poland) pro last 100 ka. If so the primary situation of Eemian valley hight has been changed up to 100 or –100 m. But for the Mazowian ones up to 300 or –300 m. The last value is impossible, because presently the level of the Mazowian sea deposits lies between 20 m a.s.l. at the western Poland and –20 m b.s.l. in the northern Poland (Fig. 1). The dimensions of vertical movements were smaller.

4. As an example of the course of the Mazowian river valley is demonstrated the Vistula valley (Fig. 6). In the Carpathians (Zuchiewicz 1995) and in the South Polish Uplands (Pożaryski et al. 1994; Lindner, Marks 1999) the level of the Vistula channel is well documented owing to developed channel deposits visible in the outcrops and present topography. Moving from Warsaw to the north, the course of the Mazowiam Vistula valley is known insufficiently. Fig. 6 shows the various places where suitable deposits have been found.

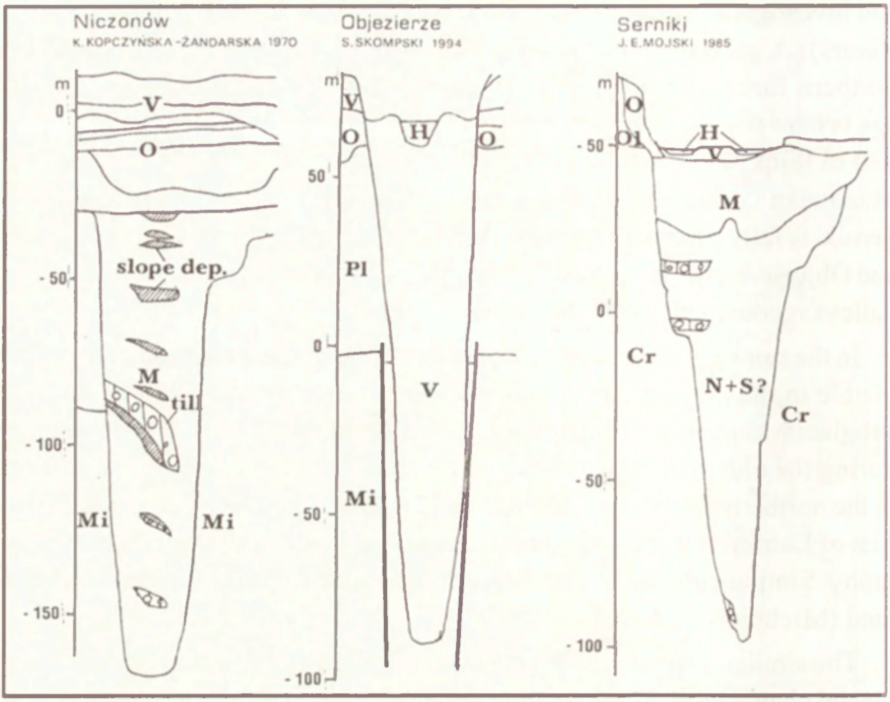

Fig. 5. Cross-sections of the subglacial channels (Niczów in Pomerania Lakeland and Objezierze in Great Poland Lakeland), and Wieprz River valley, north of Lublin (simplified). Cr – Cretaceous, Ol – Oligocene, Mi – Miocene, N+S – Narewian and Sanian cold stages, M – Mazowian, O – Odranian, V – Vistulian, H – Holocene

Przekroje subglacjalnych koryt rzecznych (Niczów na Pojezierzu i Objezierze w Wielkopolsce), dolina Wieprza, na północy Lublina (uproszczone). Cr – Kreda, Ol – Oligocen, Mi – Miocen, – fazy Narwi i Sanu, M – mazowieckie, O – odrzańskie, V – wistuliańskie, H – Holocen

In the Warsaw region it is rather impossible that the Mazowian bottom of the Vistula river occured at 30 m a.s.l. (Sarnacka 1982). A better proposal is a conception (Lindner, Marks 1999), that the Vistula valley bottom occured at that time about 50 m a.s.l. Various proposals were put forward concerning the northern part of the Polish Plain. The latest of them has been described by Niewiarowski, Wysota (1996) and Lisicki (1998). The first suggestion maybe ruled out, because – 40 m b.s.l. in the vicinities of Lidzbark Warmiński, 100 km far from the Sztum Sea is distinctly too deep! The comparison between the value of 0–40 m a.s.l. at the Mazurian Lakeland and – 20 m for Domnowo marine beds is much closer, and therefore Lisicki's (1998) proposal is more reliable (Fig. 6).

The above, very briefly described example, concerning the Mazowian Vistula valley clearly shows the various difficulties with the palaeomorphic interpretation of interglacial fluvial valleys in the Polish Plain, and also illustrate the possible ways of solving them.

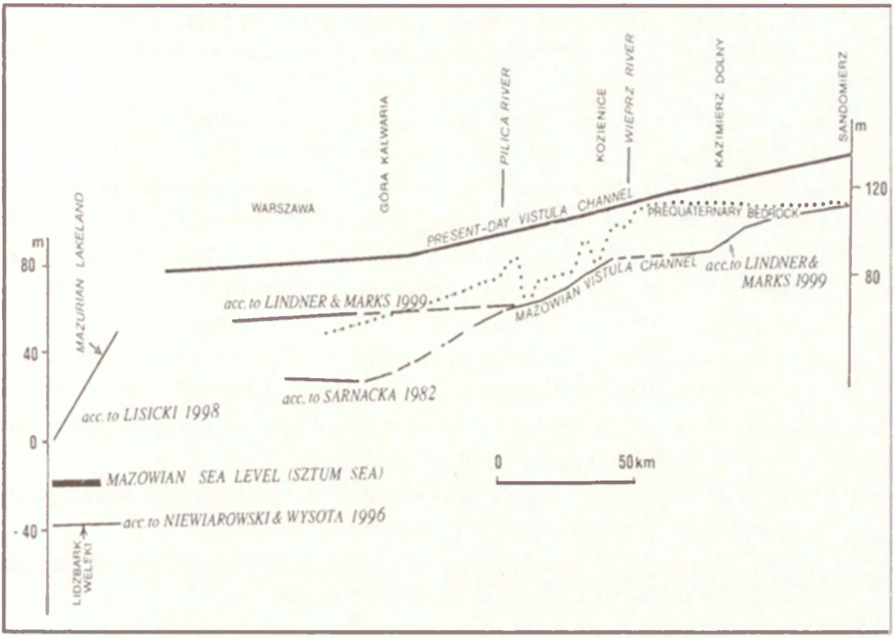

Fig. 6. The Mazowian Vistula River channel between Sandomierz and Northern Poland (simplified and completed)

Mazowieckie koryto Wisły pomiędzy Sandomierzem a północną Polską (uproszczone i uzupełnione)

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WYBRANE ZAGADNIENIA BADAŃ INTERGLACJALNYCH DOLIN RZECZNYCH NA NIŻU POLSKIM

Streszczenie

Badania interglacjalnych dolin rzecznych na Niżu Polskim winny dążyć przede wszystkim do rozpoznania ich baz erozyjnych, a więc ich lokalizacji i położenia w stosunku do poziomu obecnego morza. Obecnie dobre rozpoznanie dotyczy tylko obu młodszych interglacjałów (ryc. 1 i 2, tab. 1), a więc mazowieckiego i eemskiego. Pryncypialną zasadą winno też być stosowanie właściwej definicji interglacjalnych osadów rzecznych. Nie było to i nadal nie jest łatwe. Jednakże bardzo pomocnymi w tej mierze są dobrze rozpoznane rzeczne osady holoceńskie (ryc. 4), a więc również interglacjalne. Rozpoznanie to dotyczy (Starkel 2001) ich przewodnich cech litologicznych, zmian facjalnych i ich miąższości w dolinie Wisły. Wnioski wypływające z tego rozpoznania są równie zaskakujące, jak i oczywiste. Np. miąższość odpowiednich osadów jest kilkakrotnie mniejsza, aniżeli przyjmuje się często dla rzecznych osadów interglacjalnych.

Wiele dobrze rozwiniętych dolin plejstoceńskich na Niżu Polskim i wyżynach cechuje się znaczną ich miąższością, wynoszącą ponad 100 m. Sprawia to, że w wielu sytuacjach są one podobne do rynien subglacjalnych. Bywa to źródłem wielu nieporozumień w interpretacjach genetycznych (ryc. 5, dla form wieku interglacjału mazowieckiego – holsztyńskiego). Nieco ułatwia sytuację fakt, że wiele głębokich dolin rzecznych z wyżyn przedłuża się ku północy w obszar Niżu Polskiego. Najlepszym przykładem jest tu dolina Wisły. Dobrze poznana jest w przełomowym jej odcinku powyżej Kazimierza Dolnego (ryc. 6). Dopiero na północ od Warszawy jej dalszy przebieg nie jest wystarczająco poznany. Można przypuszczać, że w interglacjale mazowieckim biegła ona ku północy.

Konieczne jest też uwzględnianie roli pionowych ruchów skorupy ziemskiej. Wydaje się, że nie była ona tak znaczna, jak niekiedy sugeruje to literatura fachowa. Osady morskie interglacjału mazowieckiego leżą obecnie od około –20, na wschodzie, do około 20 m n.p.m., na zachodzie. Ujście dolin rzecznych musiało więc znajdować się na podobnych wysokościach. Podobne stosunki hipsometryczne mają obecnie miejsce dla interglacjału eemskiego.

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PLENIVISTULIAN DUNES IN POLAND – A NEW VIEW ON THE DEVELOPMENT PHASES OF INLAND DUNES

1. INTRODUCTION

The occurrence of inland dunes in Poland is connected with the broadly understood periglacial zone of the last Pleistocene glaciation, i.e. the Vistulian. In Europe most typical and numerous inland dunes are found in the Middle European Lowland belt, that is in the German Lowland, Polish Lowland, and Polesie. In Poland dunes occur also in the upland belt (Silesian Upland, Little Poland Upland, Lublin Upland) and in the Peri-Carpathian depression belt (Oświęcim Basin, Sandomierz Basin). That is why in the beginning of the 20th century the German and Polish scientists carried out their classic studies of inland dunes just in Central Europe.

Typical inland dunes are usually parabolic ones; transversal and longitudinal dunes are less frequent. Inland dunes, mainly parabolic ones, form dune fields covering large areas, for example in the pradolina of the Warta and Noteć rivers, in the Kampinos forest (westwards of Warsaw), and in several other regions of Poland.

Inland dunes, which have been studied in Poland and in the whole Europe till now, are hills reaching from several to a dozen or so metres (rarely over 20 m), and well visible in the landscape. These dunes are stabilized by podzolic soils and vegetation cover (pine forests).

The last part of the Pleistocene was recognized to be the time of their formation; several dune-forming phases were distinguished, which occurred during cold oscillations of that period since the Oldest Dryas. Recent investigations made in Poland have shifted the beginning of dune formation a little age. In the opinion of B. Manikowska (1991, 1995) the Late Glacial dunes were formed in the Kamion (=Epe) Interstadial (14.5 ka BP). The author of this paper recognized the initial phase of their formation around 16 ka BP on the basis of TL dating (Wojtanowicz 1996).

However, this paper does not concern the Late Vistulian dunes because they have been sufficiently recognized. A question under discussion is the occurrence of older dunes in Poland, i.e. those of the Plenivistulian age. The author has succeeded in discovering and evidencing the occurrence of the Plenivistulian dunes visible in modern relief. This fact throws new light on the problem of the number of dune-forming phases. The results of a research on these dunes provide some new information on the Plenivistulian palaeogeographical conditions, and especially on the prevailing wind directions.

2. PROBLEM OF THE PLENIVISTULIAN DUNES IN THE LIGHT OF LITERATURE

Aeolian processes were very important in periglacial zone during the Plenivistulian when climate was cold and dry. Strong winds caused deflation and corrasion; ventifacts and windpolished boulders were formed, and aeolian deposits (loesses, aeolian coversands, fluvio-aeolian sands) accumulated. In Europe these phenomena were described in many papers (Böse 1991; Christiansen, Swensson 1998; Goździk 1991; Kasse 1999; Kozarski, Nowaczyk 1991; Liedtke 1993; Lindner 1971, 1972, 1976; Maruszczak 1967, 1968, 1983 Schirmer 1999; van Huissteden et al. 2001). A question can be put about dunes. Whether they were not formed during the Plenivistulian or they were destroyed as so little information about them is available? And if they were formed, whether they have survived to the present?

Inland dunes occurring in Georgia and Louisiana (southern part of the United States, North America) originated in the peripheries of periglacial zone of the Wisconsin Glaciation. They were TL dated at several phases of the last glaciation, and mainly at the Upper Pleniglacial, i.e. 30–15 ka BP. These dunes were formed in cold and dry climate (Ivester et al. 2001; Otvos, Price 2001).

In West Europe the Plenivistulian dunes were found only in the Netherlands (van Huissteden et al. 2001). These dunes are low, rather flat ridges.

S.Z. Różycki (1967) was the first who noticed the occurrence of the Plenivistulian dunes in Poland. He considered the Late Glacial systems of parabolic dunes to be accumulated in one of the last stages of dune formation. In his opinion the dune ridges several metres high, which show a W–E orientation and occur in Błonie Plain (central Poland) are the oldest preserved dune generation. He thought that these dunes dated from the Leszno (Brandenburgian) Stadial of the last glaciation so they should be older than 20 ka BP. The subsequent studies indicated that these ridges having a general WNW–ESE orientation consist of flat and irregular sandy hummocks. The TL age

(15.8 ka BP) obtained for one sample taken from one of these hummocks (Gągolin site – sampling depth of about 2 m) is probably too young (Urbaniak-Biernacka 1992).

J. Goździk (1991) also found dunes of the oldest generation in central Poland, in Łódź environs. These low ridges (about 1 m high) of W–S orientation were formed in the Upper Plenivistulian, between 19 and 13.5 ka BP. In the Kielce-Sandomierz Upland M. Barcicki (1997) examined the dune ridge on a W–E alignment, which is 800 m long and about 1.5 m high. TL dating of this dune yielded the following ages: 14.5 ka BP for a sample from the depth of 0.7 m, and 22.5 ka BP for a sample from the depth of 1.25 m. H. Maruszczak (1968, 1983) postulated earlier the possibility of preservation of the Plenivistulian dunes to the present time. The author of this study published the paper about the Plenivistulian dunes in the Sandomierz Basin, SE Poland (Wojtanowicz 1999).

3. PLENIVISTULIAN INLAND DUNES IN POLAND

3.1. DISTRIBUTION

The distribution of the Plenivistulian dunes in Poland is shown in relation to the occurrence of the Late Glacial dunes and the Vistulian ice-sheet extent (Fig. 1). Both the sites described in literature (I, II, III), and examined by the author (Ż, U, K) are marked on the map. The Plenivistulian dunes, which have been found in Poland till now, occur south of the maximum extent of the Vistulian ice sheet. Theoretically, they can occur south of ice-sheet extent during the Pomeranian Stadial (about 15 ka BP) of this glaciation. Therefore, the Wielkopolska region on the Warta and Noteć Rivers, and the Vistula River valley between Warsaw and Bydgoszcz are the areas where the Plenivistulian dunes can also be found.

The Plenivistulian dunes occur in three main physico-geographical regions, in which also most of the Late Glacial dunes are present. From the south, these are the following regions: Sandomierz Basin as a part of the Peri-Carpathian depressions, Central Polish Uplands, and Middle Polish Lowland.

3.2. DESCRIPTION OF THE SELECTED DUNE AREAS

The author of this paper made the research on the Plenivistulian dunes in three areas: Żuków (Sandomierz Basin), Uściąż (Bełżyce Plain), and Kazanów (Radom Plain). The geomorphologic maps of these dune areas were made

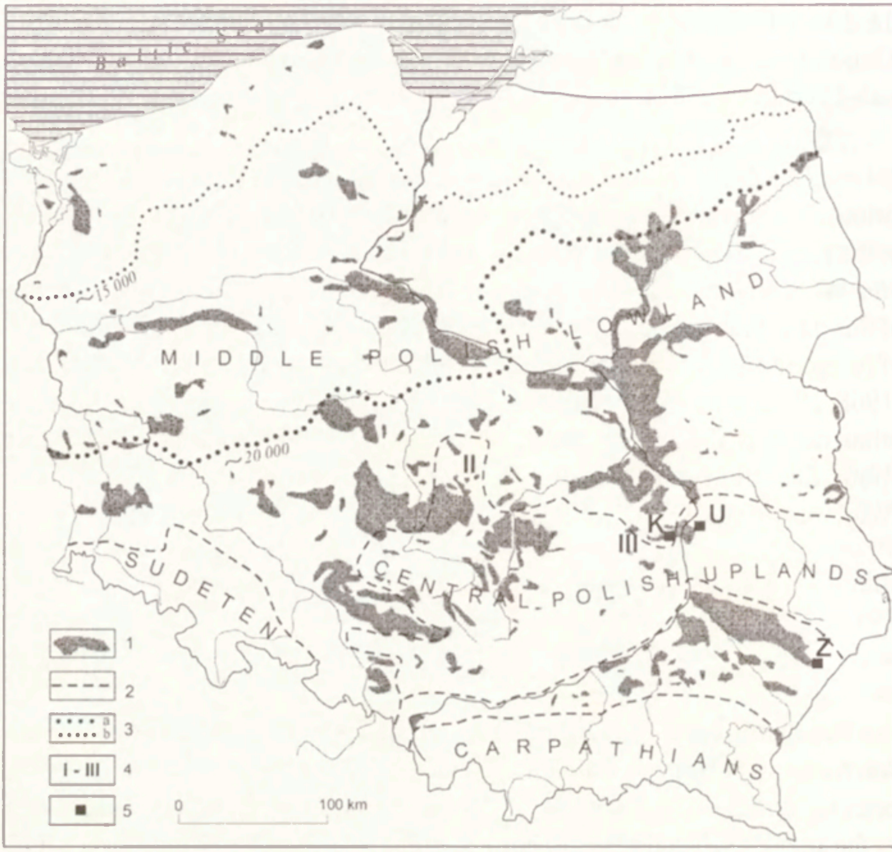


Fig. 1. Distribution of inland dunes in Poland
1 – dunes and sand areas (Galon 1958), 2 – boundaries of some main physico-geographical units, 3 – Vistulian ice-sheet extent: a – maximum extent, b – Pomeranian Stadial, 4 – sites of the Plenivistulian dunes known from literature: I – Blonie Plain (Różycki 1967), II – Łódź environs (Goździk 1991), III – Mirówek, Przedgórze Iłżeckie (Barcicki 1997), 5 – Plenivistulian dunes examined by the author: Ż – Żuków (Sandomierz Basin), U – Uściąż (Bełżyce Plain), K – Kazanów (Radom Plain)

Rozmieszczenie wydm śródlądowych w Polsce
1 – wydmy i obszary piaszczyste (Galon 1958), 2 – granice niektórych głównych jednostek fizjograficznych, 3 – zasięg zlodowacenia Vistulian: a) zasięg maksymalny, b) stadiał pomorski, 4 – znane z literatury stanowiska wydm plenivistulianskich: I – Równina Błońska (Różycki 1967), II – okolice Łodzi (Goździk 1991), III – Mirówek, Przedgórze Iłżeckie (Barcicki 1997), 5 – obiekty wydm plenivistu-lianskich badane przez autora: Ż – Żuków (Kotlina Sandomierska), U – Uściąż (Równina Bełżycka), K – Kazanów (Równina Radomska)

using topographic maps 1:10 000 (Żuków, Uściąż), and 1:25 000 (Kazanów). The dunes were TL dated[1], and standard grain size analysis was made.

3.2.1. ŻUKÓW DUNE AREA (SANDOMIERZ BASIN)

The dune area is situated in the southern Poland, in the eastern part of the Sandomierz Basin, in the Tarnogród Plateau subregion, near Żuków (Fig. 2). The author described this area earlier (Wojtanowicz 1996, 1999). This interesting complex of dune forms differs in character from the Late Glacial dunes investigated by the author in the Sandomierz Basin (Wojtanowicz 1969). The Plenivistulian dunes are less visible in the landscape. These low, WNW–ESE trending ridges are from several hundred metres to several (2.5–6.5) kilometres long, and they often consist of separate hummocks. The ridges are 1–4 m high and 20–60 m wide (Fig. 2). The position of these ridges in relation to other relief forms, and especially their encroachment from river valleys on the western slopes of interfluve areas provide evidence of a general westerly wind direction. The measurements of dune ridge orientations indicate WNW winds from a 290°±20° sector. This fully corresponds with the direction of aeolian sand transport in the Roztocze region and Sandomierz Basin reconstructed by J. Buraczyński (1993, 1994) for the Plenivistulian.

One from the higher hummocks of the Żuków dune area was examined in detail (Fig. 3). Dune sand 4 m thick overlies fluvial sand dated at 60 ka BP. Dune was formed on denudation surface developed on fluvial sands. Sand from the dune base was TL dated at 32 ka BP, and towards the top the successive dates were 29, 28, 22 ka BP. Considering on these results, one can assume that the dune was formed between 32/30 and 20 ka BP. Therefore, this could have been the first stage of inland dune formation in Poland. Two paleosols were found in the dune. The older one developed about 30 ka BP, and probably corresponds to the Denekamp Interstadial (Interphase?). The upper paleosol is older than 22 ka BP, and corresponds to an interphase of the younger Pleniglacial.

Granulometric analysis revealed almost three times greater content of silt fraction, and distinctly more poorly sorted sand in the examined dune in comparison with the Late Glacial dunes in the Sandomierz Basin (Wojtanowicz 1970, 1999).

[1] TL dating was made by mgr Jarosław Kusiak in the Thermoluminescence Laboratory, Department of Physical Geography and Paleogeography, Maria Curie-Sklodowska University, Lublin.

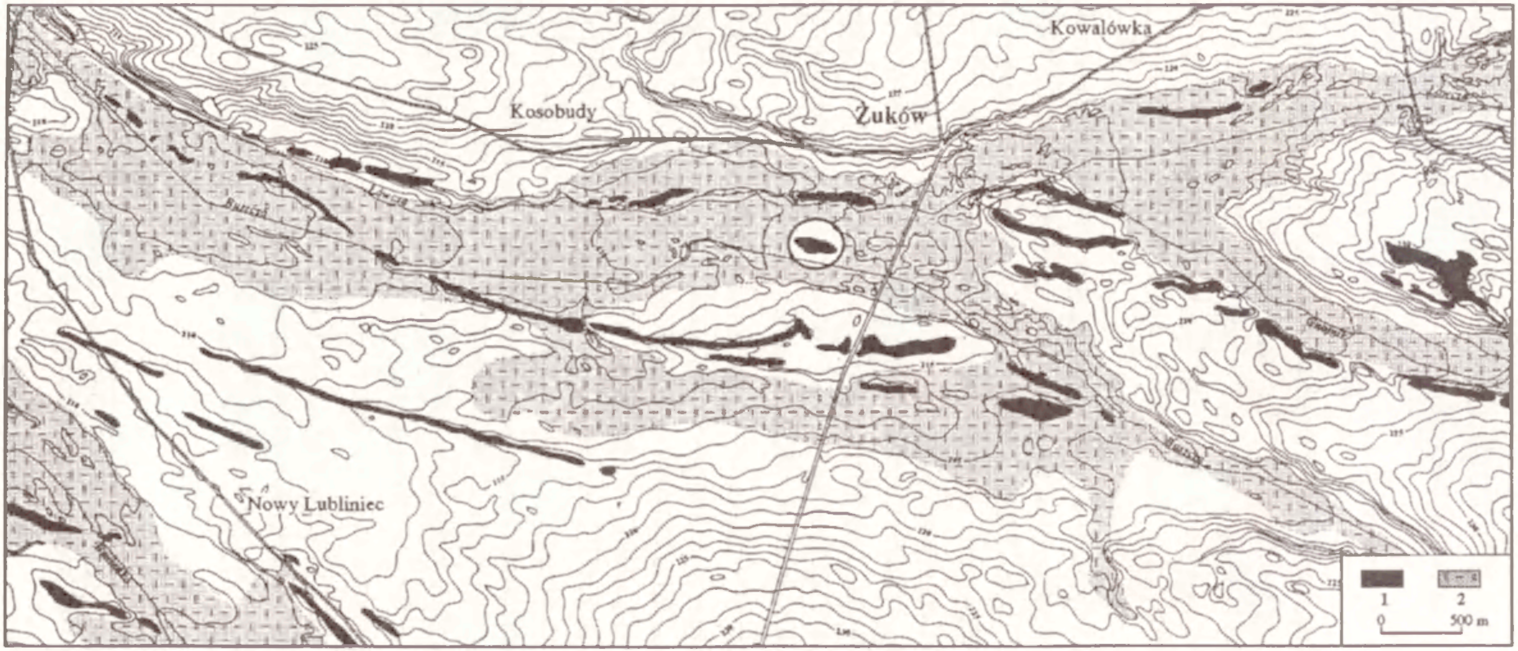


Fig. 2. Situation of dunes in the environs of Żuków (after J. Wojtanowicz 1999)
1 – dunes, 2 – Holocene river accumulation plains

Sytuacja wydm plenivistulianskich w okolicach Żukowa (wg J.Wojtanowicza 1999)
1 – wydmy, 2 – równiny holoceńskie akumulacji rzecznej

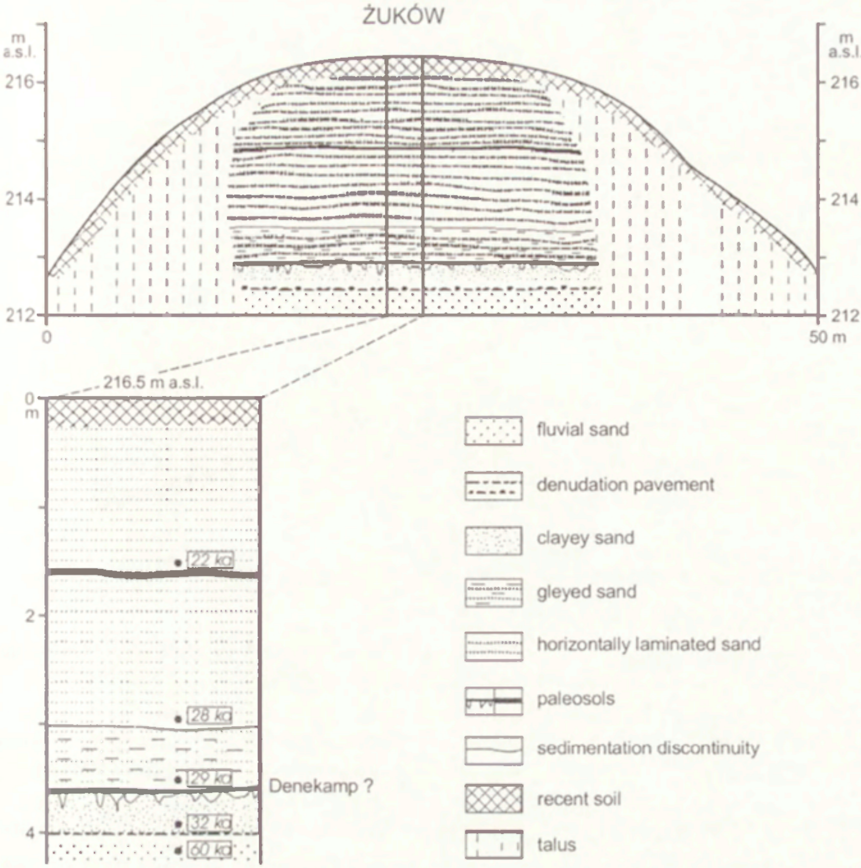


Fig. 3. Cross-section and profile of the Żuków dunc (after J. Wojtanowicz 1999)
Przckrój poprzcczny i profil wydmy Żuków (wg J.Wojtanowicza 1999)

3.2.2. UŚCIĄŻ DUNE AREA (BEŁŻYCE PLAIN)

The Uściąż dune area is situated in the Lublin Upland, in the Bełżyce Plain subregion, in its north-western part adjacent to the Nałęczów Plateau. The dune area is 6 km eastwards of the Vistula River gap trough the South Polish Uplands.

Four dune ridges were found in this dune area. They are 0.75–1.55 km long, 25–70 m wide, and 1–3 m high. These low and flat sandy ridges show a WNW–ESE orientation parallel to the southern edge of the Nałęczów Plateau (Fig. 4). This loess edge is 15–20 m high. It seems that both the loess edge and the dunes were simultaneously formed by the same winds, i.e. probably the winds from a 290°±5° sector. Though rather low, nevertheless some dune ridges are well visible in the landscape, e.g. the dune with outcrop (Photo 1).

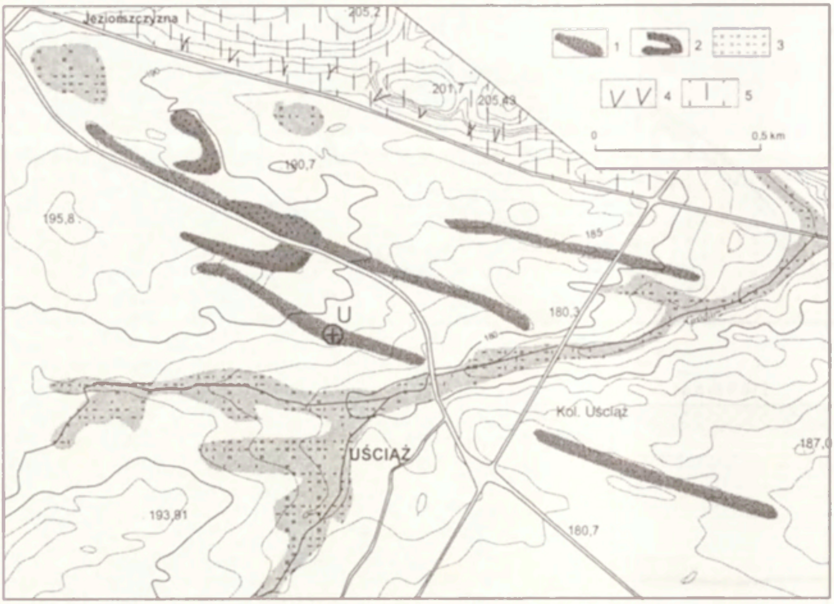


Fig. 4. Situation of dunes in the environs of Uściąż
1 – Plenivistulian dunes, 2 – Late Glacial parabolic dunes, 3 – wetlands with Holocene accumulation, 4 – Nałęczów Plateau edge, 5 – loess cover

Sytuacja wydm w okolicach Uściąża
1 – wydmy plenivistulianskie, 2 – wydmy paraboliczne późnoglacjalne, 3 – strefy podmokłości z akumulacją holoceńską, 4 – krawędź Płaskowyżu Nałęczowskiego, 5 – pokrywa lessowa

The encroachment of parabolic dunes on dune ridges is a very interesting and important phenomenon occurring in the north-western part of the examined area, where one of two parabolic dunes encroaches and covers a part of the dune ridge. The parabolic dune is distinctly higher than the ridge. These two dune types are composed of two distinct groups of sand. Sand from parabolic dunes is typical of the Late Glacial dunes, i.e. loose, medium-grained, well sorted. The older, ridge forms are composed of less loose material, with a higher content of fine sand and silt, and clay interlayers; at the top it is silty sand (Fig. 5, Photo 2).

In the light of the TL dating (14.3 and 15.6 ka BP) the Uściąż sandy ridges should be related to the end of the Plenivistulian. This could have been the second, younger dune-forming phase during the Plenivistulian.

3.2.3. KAZANÓW DUNE AREA (RADOM PLAIN)

The Kazanów dune area is situated in the Polish Lowland belt, in its southern part, in the Radom Plain subregion. Its more detailed topographic and geomorphologic situation is presented in figure 6. Two ridges having a gen-

Photo 1. Uściąż. Dune ridge (ridge with outcrop) clearly visible in the background. Buildings, crops, and trees on the ridge top

Uściąż. Na ostatnim planie wyraźny wał wydmowy (wał z odkrywką). Na nim zabudowa, uprawy, zboża

Photo 2. Uściąż. Profile of the deposits composing the dune ridge (ridge with outcrop)

Uściąż. Profil osadów budujących wał wydmowy (wał z odkrywką)

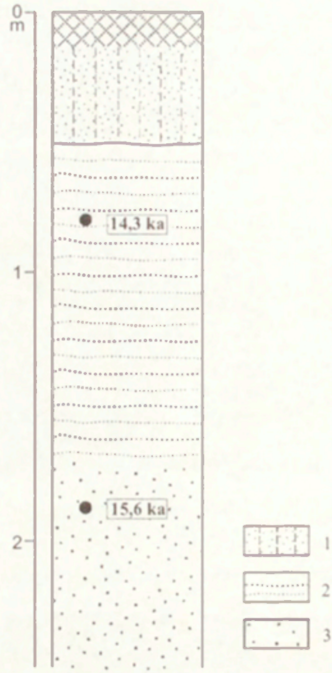


Fig. 5. Profile of the Uściąż dune (situation marked with letter "U" in Fig. 4)
1 – silty sand, 2 – sand with clayey interlayers (traces of periglacial processes), 3 – medium-grained sand

Profil wydmy Uściąż (położenie zaznaczono literą „U" na fig 4)
1 – piasek pylasty, 2 – piasek z przewarstwieniami ilastymi i ślady procesów peryglacjalnych, 3 – piasek średnioziarnisty

eral W–E orientation occur north of the valley of the Iłżanka River (tributary of the Vistula River), on flat plain with relative relief up to 5 m. These flat, low ridges are 1–3 m high and 50–250 m wide. Their length is considerable; the northern ridge is 6.5 km long, and the southern one even 12 km long. This second ridge consists of four parts, which have slightly different (by several degrees) alignments. However, the orientations of both dune ridges, and the thus the direction of dune-forming winds fall within a 281°±8° sector.

Such a slight variable orientation of dune ridges is observed in all three examined dune areas. This phenomenon resulted not from directional variability of dune-forming wind but from different environmental characteristics, i.e. lithology, water conditions, pre-existing relief.

When looking at the detailed topographic map of the Kazanów area, one can notice a very striking fact: villages occupy both dune ridges over all their length. Brzezinki Nowe and Wilczy Ług villages are situated on the northern dune ridge, and on the southern one – Kolonia Zakrzówek, Kolonia Osuchów, Dębnica Nowa, Kroczów Większy, Kroczów Mniejszy, and Ranachów Górny.

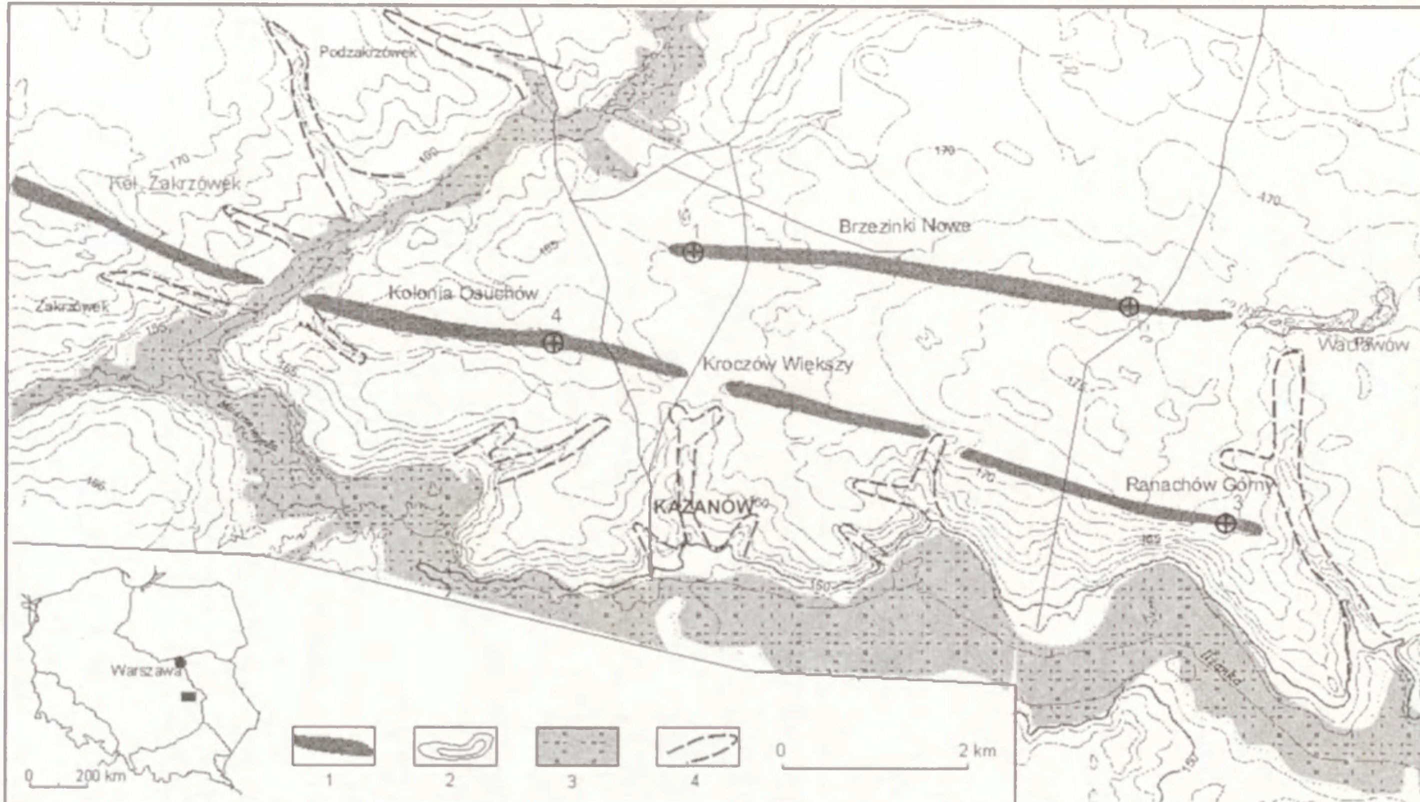


Fig. 6. Situation of dunes in the environs of Kazanów
1 – Plenivistulian dunes, 2 – Late Glacial dunes, 3 – Holocene river accumulation plains, 4 – denudation valleys

Sytuacja wydm okolic Kazanowa

1 – wydmy plenivistulianskie, 2 – wydmy późnoglacjalne, 3 – równiny holoceńskiej akumulacji rzecznej, 4 – doliny denudacyjne

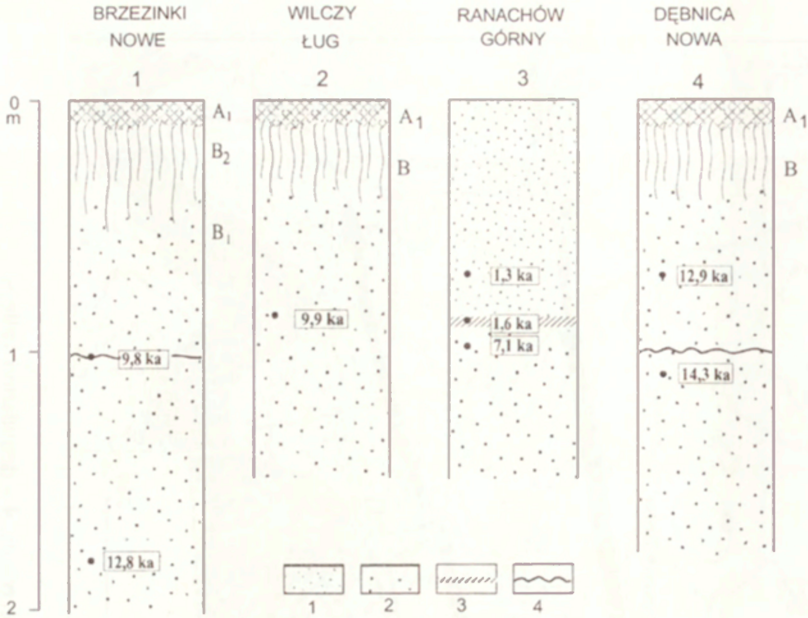


Fig. 7. Profiles of dunes in the environs of Kazanów
1 – grey, loose sand, 2 – fine- and medium-grained sand, 3 – paleosol, 4 – erosion surfaces

Profile w wydmach okolic Kazanowa
1 – piasek luźny szary, 2 – piasek drobno- i średnioziarnisty, 3 – gleba kopalna, 4 – powierzchnie erozyjne

Evidently, these ridges provided good conditions for settlement. They rise over the surrounding, often waterlogged area, so ground surface is dry but groundwater is easily available.

Studies of the Kazanów dune ridges indicate that they were transformed during the Late Glacial and Holocene. This is clearly visible in dune morphology; dune ridges are covered by younger hummocks in some places. Such a situation occurs in the eastern ends of both ridges, i.e. in the environs of Wilczy Lug and Ranachów Górny villages. The northern ridge changes eastwards into the young, Late Glacial parabolic dune. This complex dune-forming aeolian process is reflected in the dune structures and the results of TL dating (Fig. 7). The obtained TL ages seem to be too young. However, it should be noted that almost all samples for dating were taken from the upper parts of the dunes. The most representative site for dune ridge dating is the Dębnica Nowa profile where two most complete sand series occur. The oldest, Plenivistulian one was dated in the top at 14.3 ka BP. The younger, Late Glacial one was dated in its middle part at 12.9 ka BP. I relate the formation of dune ridges in the Kazanów area with the second dune-forming phase in the Upper Pleniglacial.

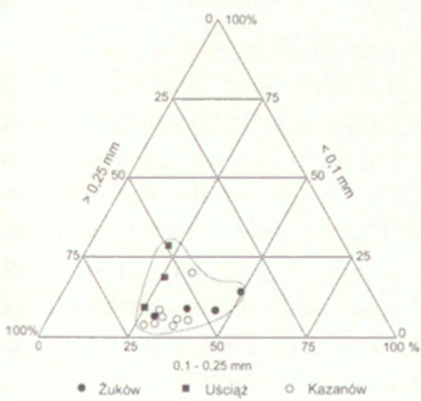
3.3. GRAIN SIZE AND SORTING CHARACTERISTICS OF DUNE SAND

The examined Plenivistulian dunes are composed of medium- and fine-grained sand with silty sand interlayers (Fig. 8). They contain more silt fraction, and sand is more poorly sorted in comparison with the Late Glacial dunes. The values of main granulometric indices (according to Folk and Ward) range in the following intervals: Mz = 1.73 – 2.36, σ_1 = 0.60 – 1.64, Sk_1 = (– 0.03) – 0.55, K_G = 0.80 – 1.69. The lowest values of mean grain size were found in the Żuków dunes, and the highest – in the Kazanów dunes. Most poorly sorted sand occurs in the Uściąż dunes.

It should be noted that the Uściąż dune area adjoins the large loess patch (Lublin Upland), and the Żuków dune area is situated near the loesses of the Roztocze region. These locations could have influenced grain size and sorting parameters of the examined dunes.

3.4. DIRECTIONS OF WINDS

The author considers that the Plenivistulian dunes in Poland were formed by WNW winds (281–290°). The WNW–ESE dune ridges indicate this direction of dune-forming winds. Westerly, and not easterly wind direction is evidenced by the position of dunes in relation to other relief forms (edges, river valleys), the encroachment of dunes on slopes from the west, and also the position of dunes in relation to their potential alimentation areas. J. Goździk (1991, 1995) also postulated the WNW direction of aeolian sand transport in the Upper Plenivistulian (dated by him at 25–14 ka BP) in central Poland.



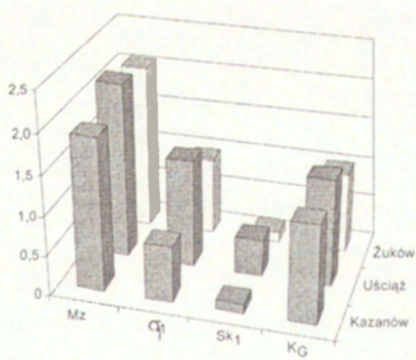


Fig. 8. Grain size and sorting parameters of the Plenivistulian dunes (Żuków, Uściąż, Kazanów)

Parametry uziarnienia wydm plenivistulianskich (Żuków, Uściąż, Kazanów)

The Plenivistulian dunes were formed simultaneously with loesses. It can be supposed that both these types of deposits and aeolian relief forms were formed by the same winds blowing from the same directions. However, in Poland the opinions of loess-forming wind direction are contrasting: either easterlies (Maruszczak 1967) or westerlies were assumed to be loess-forming winds (Lindner 1971, 1972, 1976; Chlebowski, Lindner 1976). Only this latter viewpoint was reliably documented. The investigations made by the mentioned authors in the region of the Holy Cross Mountains indicate that sandy streaks occur in the westernmost parts of loess patches stretched in the W–E direction. Sand came from sandstone crags bearing strong traces of corrasion, which occur in the western side of loess patches. The area of these crags was a kind of "deflation basin" for loesses. Western palaeo-wind direction was also confirmed by mineralogical studies; in one of loess patches the garnet content decreases and the muscovite content increases eastwards.

The opinions of the atmospheric circulation regimes in Europe during the Pleniglacial are different. H.H. Mayer and C. Kottmeier (1989) postulated a western palaeo-wind direction in West Europe, and eastern one in the area eastwards of the Elbe River. Recently H.H. Christiansen and H. Svensson (1998) examined windpolished boulders in Denmark, and inferred that easterly winds (105°±20°) prevailed in the last part of the Plenivistulian (22–17 ka BP). These conclusions aroused a discussion (Vandenberghe et al. 1999). The methodology of those studies was also doubtful. It can be supposed that only the WNW–ESE orientation could be derived from the aeolian features on boulders. The proposed ESE direction of relief-forming winds was based on the assumption of the dominance of anticyclonal winds related to the ice sheet.

The opinion of a great, and even dominant role of anticyclonal winds should be verified in the light of the latest investigations. Katabatic winds were also important though their spatial extent was limited. These local winds were typical of a relatively narrow zone (up to 50 km wide) near the ice front (Vandenberghe et al. 1999). In the opinion of Brodzikowski (1987) during the Vistulian the westerly circulation occurred in West and Central Europe, south of the ice sheet. This wind direction was linked with cyclones, which moved from west to east. Therefore, the Atlantic Ocean conditioned the type of atmospheric circulation in Europe, just as at the present time. The investigations of the Plenivistulian dunes in Poland confirm the opinion about the dominance of westerly circulation.

4. AGE OF THE PLENIVISTULIAN DUNES; DUNE-FORMING PHASES

In the light of the investigations presented in this paper, it is possible to submit a proposition that the role of aeolian processes during the Plenivistulian, and their morphological effects were greater than it has been assumed till now. Not only aeolian coversand plains but also dunes were formed as a result of aeolian activity. These dunes differ in form and structure from the commonly known Late Glacial dunes, and perhaps because of this fact they were not recognized as dunes. They probably occur in the whole, broadly understood periglacial zone of the last glaciation.

The existence of the Plenivistulian dunes changes our opinion about the dune-forming phases of inland dunes. However, it should be stressed that the TL dated dunes examined by the author fit well in the phases of aeolian activity distinguished in the Plenivistulian.

The following development phases of inland dunes can be distinguished (Table 1):

– Phase I /Żuków/ (30–20 ka BP) – between the Denekamp Interstadial and Brandenburgian (Leszno) Stadial, a counterpart of aeolian accumulation of the Older Coversand I. This is the first development phase of the Plenivistulian dunes, which occurred on the turn of the Interplenivistulian and Upper Plenivistulian.

Żuków dune type: flat linear dunes

– Phase II /Uściąż, Kazanów/ (20–15 ka BP) – between the Brandenburgian (Leszno) and Pomeranian Stadials, a counterpart of the aeolian accumulation of the Older Coversand II. This is the second development phase of the Plenivistulian dunes, which occurred in the Upper Plenivistulian.

Kazanów dune type: longer flat linear dunes.

– Phase III (15–8 ka BP) – the Late Glacial dune-forming phase, a counterpart of the aeolian accumulation of the Younger Coversand I and Younger Coversand II.

Dune type: parabolic and ridge dunes.

Table 1. Phases of dune formation in Poland

Age ka BP	Stratigraphy		Dune phases (West-European counterparts)	Dune types
8–10	HOLOCENE		(Younger Coversand II) Phase III (Younger Coversand I)	Late Glacial parabolic and ridge dunes
10–15	PLEISTOCENE	Late Glacial		
15–20		Younger (Upper) Pleniglacial	Phase II [Uściąż, Kazanów] (Older Coversand II)	Plenivistulian flat linear dunes
20–25			Phase I [Żuków] (Older Coversand I)	
25–30		Interpleniglacial		

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WYDMY PLENIVISTULIAŃSKIE W POLSCE – NOWY POGLĄD NA FAZY ROZWOJU WYDM ŚRÓDLĄDOWYCH

Streszczenie

W strefie peryglacjalnej ostatniego zlodowacenia plejstoceńskiego rozwijały się w Europie, w tym i na obszarze Polski, wydmy śródlądowe. W typowym wykształceniu są to przede wszystkim wydmy paraboliczne, w mniejszym stopniu wydmy typu wałów poprzecznych i podłużnych. Wydmy te to wyraziste, czytelne w rzeźbie pagóry o wysokościach od kilku do kilkunastu metrów, dochodzące do powyżej 20 m, tworzące często zwarte kompleksy wydmowe i zajmujące duże po-

wierzchnie. Czas powstawania tych wydm określono na schyłek plejstocenu; początek ich powstawania określono w Polsce na 14,5 ka BP (Manikowska 1991, 1995) lub na 16 ka BP (Wojtanowicz 1996).

Zagadnienie wydm śródlądowych późnoglacjalnych jest dostatecznie dobrze rozpoznane i nie jest przedmiotem niniejszego opracowania. Autor porusza problem występowania w Polsce wydm starszych, z pełni Vistulianu. Udało się odkryć i udokumentować we współczesnej rzeźbie istnienie wydm plenivistuliańskich. Stawia to w nowym świetle zagadnienie faz rozwoju wydm śródlądowych.

Na istnienie w Polsce wydm plenivistuliańskich pierwszy zwrócił uwagę S.Z. Różycki (1967), a później istnienie takich wydm w środkowej Polsce potwierdzali na przykładzie pojedynczych stanowisk J. Goździk (1991) i M. Barcicki (1997). Autor niniejszej pracy opublikował w 1999 roku artykuł o wydmach plenivistuliańskich w Kotlinie Sandomierskiej (Wojtanowicz 1999).

Przebadano trzy obszary występowania wydm plenivistuliańskich (ryc.1). Są to: Żuków (Kotlina Sandomierska), Uściąż (Wyżyna Lubelska; Równina Bełżycka) i Kazanów (pas niżu polskiego; Równina Radomska). Dla wszystkich tych obszarów wykonano mapy geomorfologiczne z zaznaczeniem form wydmowych, przy wykorzystaniu map topograficznych 1:10 000 (Żuków, Uściąż) i 1:25 000 (Kazanów). Wydmy datowano metodą TL oraz wykonano standardową analizę granulometryczną.

Wydmy plenivistuliańskie różnią się swoim charakterem od wydm późnoglacjalnych. Od tych ostatnich są mniej czytelne i mniej wyraźne w rzeźbie. Składają się z wydłużonych, niskich, często płaskich wałów piaszczystych o wysokości 1–4 m, szerokości od 20 do 250 m i długości od kilkuset metrów do kilku kilometrów (2,5–6,5 km) a nawet, jak w Kazanowie do kilkunastu kilometrów (12 km). Wały te wyciągnięte są w kierunku równoleżnikowym; najczęściej jest to kierunek WNW–ESE. Niektóre wały składają się z ciągu całego szeregu oddzielonych pagórków wydmowych.

Na podstawie stosunku do innych form rzeźby, takich jak między innymi wkraczanie od zachodu z dolin rzecznych na stoki wierzchowin – tak jest w Żukowie – można wnosić o ogólnie zachodnim kierunku wiatrów w powstawaniu wydm plenivistuliańskich.

Analiza granulometryczna wydm wykazała zwiększenie w stosunku do wydm późnoglacjalnych zawartości frakcji pylastej (prawie 3-krotnie) i znacznie słabsze wysortowanie.

Przeprowadzone badania upoważniają do postawienia tezy, że rola procesów eolicznych w plenivistulianie była większą niż to się dotąd przyjmuje. Większe były efekty morfologiczne tej działalności. Efektem są nie tylko eoliczne piaski pokrywowe (*Coversand)*, ale także wydmy. Są one inaczej wykształcone, mało podobne do powszechnie znanych wydm późnoglacjalnych i z tego prawdopodobnie względu nie były rozpoznawane, a w każdym razie nie były identyfikowane jako wydmy. Występują prawdopodobnie w całej szeroko pojętej strefie peryglacjalnej ostatniego zlodowacenia.

Istnienie wydm plenivistuliańskich zmienia nasz pogląd na fazy wydmotwórcze wydm śródlądowych. Godnym podkreślenia jest jednak fakt, że wydmy, które badał autor i które wydatowano metodą TL, dobrze wpasowują się w ramy wydzielanych w plenivistulianie faz działalności eolicznej.

Wydzielić więc można następujące fazy rozwoju wydm śródlądowych (tab. 1):

– Faza I (30–20 ka BP) – pomiędzy interstadiałem Denekamp i stadiałem Branderburgian – odpowiednik fazy eolicznej *Older Coversand I*. Jest to pierwsza faza plenivistuliańskich wydm, rozwinięta na pograniczu interpleniglacjału i górnego pleniglacjału.

Żukowski typ wydm: płaskie wydmy linearne.

– Faza II (20–15 ka BP) – pomiędzy stadiałem Branderburgian i stadiałem Pomeranian – odpowiednik fazy eolicznej *Older Coversand II*. Jest to druga faza plenivistuliańskich wydm rozwinięta w górnym plenivistulianie.

Kazanowski typ wydm: długie płaskie wydmy linearne.

– Faza III (15–8 ka BP) – późnoglacjalna faza wydmotwórcza – odpowiednik faz eolicznych *Younger Coversand I i Younger Coversand II*.

Typy wydm: paraboliczna i poprzeczna.

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PLENI– AND LATE VISTULIAN GLACIAL LAKES, THEIR SEDIMENTS AND LANDFORMS: A CASE STUDY FROM THE YOUNG GLACIAL LANDSCAPE OF NORTHERN POLAND

1. INTRODUCTION

Lakes of glacial origin – plus the sediments and landforms they have left behind – play an important role in the young glacial landscape of northern Poland. However, in spite of many years of study, these are not fully known, with the result that discussions continue as regards many of the issues associated with their emergence, as well as the geomorphological consequences of their existence. There are also considerable terminological discrepancies. Leaving aside any in-depth discussion of these I would simply state that the glacigenic lakes on parts of the Polish Lowland influenced by the Vistulian glaciation derive their genesis from glacial ice and its melt waters. To be excluded therefore are the small lakes created from the melting of river ice buried under outwash (Różycki 1957) or icing ice (Kozarski 1975), as well as the thermokarst (thaw and alas) lakes, whose genesis is linked with the melting of ground ice in regions with permafrost (Dylik 1964; Jahn 1975; Klimaszewski 1978; Wojtanowicz 1994).

It is known from the study of today's glaciers, e.g. in Iceland and Alaska, that these may be associated with supra- or subglacial lakes more than once emptied abruptly by catastrophic events. Unfortunately, there is little knowledge or documentation of such types of lakes having existed in connection with the active ice present on the Polish Lowland in the Pleistocene, and even more so in regard to the occurrence of deposits and landforms they should have left behind. What needs therefore to be focused on, as work on the typology of Pleistocene glacigenic lakes is done, is the genetic and spatial linkage between them and the different types of ice-sheet ice, as well as their meltwaters, and the remaining geological and geomorphological traces of their existence in the shape of the relevant deposits and landforms.

Shortfalls in the available data notwithstanding, the young glacial landscape of northern Poland can be said to have included the following types of glacial lake:

a) glacial lakes existing in the period of areal deglaciation in stagnant and dead ice and lying in the forefield of active inland ice, as fed by melting stagnant and dead ice, but also sometimes by water flowing in as active ice melted (attesting to their presence are the sediments and landforms of limnoglacial kames);

b) Plenivistulian and Late Vistulian glacial lakes of the ice-dammed type, which were formed in depressions of varying genesis on the edge of the ice or forefield thereof, but in circumstances in which the outflow of water from lakes was hindered: as these were fed by glacier waters, their sediments are of the classic glacilimnic kind (supraglacial lakes forming in depressions of dead ice are usually also of this type and, while the sediments of the two differ, plains built of glacilacustrine sediments are common to both);

c) Late Vistulian glacial lakes emerging as buried glacial ice melted: to be distinguished among these are the older Pre-Alleröd lakes, which are often supraglacial and were formed as the buried ice melted in the upper part (only), or as the superficially-buried mass of dead ice melted (the sediments in these cases form high lacustrine terraces); as well as younger lakes forming following complete melting of the ice, in which case it is *sensu stricto* the lake basins that are of glacial origin, while the role played by glacial water was confined to the first phase of their emergence.

2. THE PLENIVISTULIAN GLACIAL LAKES EXISTING WITHIN THE ZONES OF STAGNANT AND DEAD ICE, AND THEIR DEPOSITS AND LANDFORMS

The existence of the glacial lakes arising in the Pleistocene within stagnant and dead ice was documented long ago (*inter alia* by Flint 1929 and Markov 1931). What are involved here are those lakes in which deposition has allowed a large mass of glacilimnic sediments to arise, and to be shaped into the protruding or flat landforms known as kames following the melting of the ice. These were recognised to exist in Poland in the 1950s (Bartkowski 1956, 1959; Niewiarowski 1959). In reference to the work of K.K. Markov (1931), the author (Niewiarowski 1959, 1961, 1963) making reference in his typology to the then-known Polish kames, drew a distinction in regard to the so-called "limnoglacial kames" that are built from limnoglacial sediments. It results from the wealth of literature on kames (including the limnological ones) that these last arise in zones of stagnant and dead ice and occur abundantly throughout the peri-Baltic area (Niewiarowski 1965) – which is to say not only where the last glaciation was, but also in the areas of the older ones.

On the basis of a detailed recognition of the morphology of these landforms, as well as the texture, structure and lithofacies of their sediments, it is possible to draw conclusions as to the nature of the glacial lakes that had been present, and the manner in which they appeared. Hence, they are believed to have arisen in already-existing crevasses and clefts in the ice, as well as in different types of depression in the ice surface. The thermal impact of the lake water and air temperature brought about their widening and deepening, something which may have led to the melting of all the ice present beneath their bottoms (Cook 1946; Keller 1952). From the way bottom sediments lie as regards the substratum it is possible to draw conclusions as to whether supraglacial or intraglacial lakes were involved. Where bottom limnoglacial (lacustrine) sediments lie much below the surrounding kame, it can be judged that these sediments were deposited in evorsion depressions (Niewiarowski 1959; Lankauf, Pasierbski 1994).

The glacial lakes under consideration were of very variable sizes, from several hectares to several tens of km^2. The kames arising from their sediments arc mainly built of fine sands and silts, more rarely of clays with horizontal bedding. They are often rhythmic, but distinct varves do not occur in them. Besides these sediments there may be lenses or interbeddings of gravels and diamicton deposits. These last mainly lie on slopes or on the surface of kames. Where a lake was for a period, or from time to time, subject to a flow through, there are layers with ripplemarks or current bedding. Attesting to the formation of limnoglacial kames in glacial lakes within stagnant or dead ice is the fact that the limnoglacial deposits built convex forms standing up to several tens of metres above the moraine plains. These most often take the form of more or less round or oval hills or plateaux. Their slopes are of the ice-contact slope kind, while their sediments often have faults, traces of subsidence and even folds or landslides. These and similar structures arose as ice walls surrounding the sediments laid down in lakes melted.

Kames that arose in supraglacial lakes may have disturbances of their sediments not only on their slopes, but also in their central parts (Kozarski 1960). Squeezed-up structures may be present in them (Keller 1952), but there are no structures reflecting the lateral pressure of active ice. Lakes of this type could occur singly, but were most often present in groups. Besides the glacial lakes existing within the ice, the Polish Lowland also had relatively narrow lakes occurring in contact with dead ice in different types of extensive depression – as for example in terminal basins or broad subglacial channels, where already ice-free slopes stood above the adjacent terrain. In these lakes, the sediments arising were similar to those of the limnoglacial kames, generally of lesser thickness, and after the lake water flowed away emerged kame terraces of limnoglacial types (Niewiarowski 1963, 1988a, b; Andrzejewski 1984; Słowański 1969; Błaszkiewicz 1998; Jaworski 2002). Glacial lakes of this type were fed mainly by waters deriving from the melting of surrounding stagnant and dead ice, but there is also evidence that some had water from melting active ice as their source of alimentation. In these cases the rhythmicity in the sediments is more distinct. These lakes were created zonally in the Pleniglacial, as there was stepwise decline and (shrinkage) of the ice sheet.

3. PLENIVISTULIAN AND LATE VISTULIAN GLACIAL LAKES OF THE ICE-DAMMED TYPE

These lakes were recognised as early as in the second half of the 19th century. They were reported to have occurred in the forefield of, and in contact with, the snout of a glacier or edge of an ice sheet, and were fed by glacial meltwaters. In some cases, they seemed to have arisen in the valleys of extraglacial rivers, in which ice-dammed lakes were present. However, it emerged later that the relationships then in place between the waters of these lakes and glaciers could vary in character. For example, some lakes bordering on to a glacier were not ice-dammed lakes, because the dams on the outflow of waters were on the forefield of a glacier on end moraine ridges (Embleton, King 1969). A very characteristic feature of such lakes are the glacilacustrine sediments laid down in them. These are most commonly rhythmically-laminated clays and silts layered horizontally.

Interest in ice-dammed lake sediments grew markedly after G. de Geer's (1912) discovery that they have paired lamina arising in summer and winter, creating annual sediments known as varves. This discovery offered a means by which geochronologies could be established for the Late Glacial and Holocene, along the so-called Swedish time scale. It emerged later that not all glacilimnic sediments have varves, such that the sediments of smaller, isolated ice-dammed lakes are not suitable for geochronological correlation analysis. This is true, *inter* alia, of the sediments from the ice-dammed lakes once present on Polish territory (Halicki 1932; Krygowski 1950; Różycki 1967).

The existence of ice-dammed lakes in what is today Poland has been most often demonstrated in the former East Prussia, on the Sępopol Lowland and in the vicinity of the Vistula Delta (Jentzsch 187; Klebs 1883; Sonntag 1911, 1919; Kraus 1924). However, varve clays were not reported from them, their place apparently being taken by the so-called "Decton" (cover clays), shaped in the form of massive brown or red clays that are not stratified and without gravel clasts, sometimes with a thin cover of stratified sands or silts. The thickness of these sediments is mostly of 1–2 m. The view that this was a deposit of ice-dammed lakes was questioned by Gagel (1925) and Körnke (1930), who considered it a local variant of till. Discussions as to the genesis of these sediments in fact continue to this day, and thanks to them the matters of the existence and range of ice-dammed lakes here are also the subject of debate. It will be also dealt with in a further part of this article.

In spite of many years of research that have now been done so far, too little is still known about the distribution and range of ice-dammed lakes during the last glaciation, as well as about their types, sediment thicknesses, lithologies and associated landforms.

Among the publications dealing with the whole area of Poland's last glaciations, it is the 1:200 000 scale *Geological Map of Poland* that gives the most complete picture regarding the presence of the sediments of ice-dammed lakes. As this map was devised it was assumed that these sediments were in the main clays and silts (but partly also sands), with horizontal and rhythmical layering resulting from their having been deposited in the lakes existing in the forefield of the glacial snout or edge of ice as they were retreating. Unjustifiably, these sediments were sometimes termed varved clays (since varves are rather rarely present and are often developed only in part of the sediments).

Analysis of the afore-mentioned map shows that the area of the last glaciation had around 70–80 ice-dammed lakes covering more than several km^2. There was a prevalence of rather small lakes, ranging from several to between 10 and 20 km^2. Only around 8–10 could have covered more than 100 km^2, though several of the largest exceeded 300 km^2 in area. The precise determination of the sizes of these lakes is not always possible, however, on account of discussions as to the genesis of some sediments, their ranges and ages. The largest were the Warsaw, Pyrzyce, Złocieniec and Varmia ice-dammed lakes.

Most debatable of all is the existence and size of the Pasłęk ice-dammed lake. L. Roszko (1971, 1983) accepts that, on the backland of end moraines from the Kaszuby–Varmia sub-phase (on the exposed lowlands before the truncating ice sheet), numerous large ice-dammed lakes appeared. One of these that had previously been unknown was the Pasłęk ice-dammed lake (Fig.1).

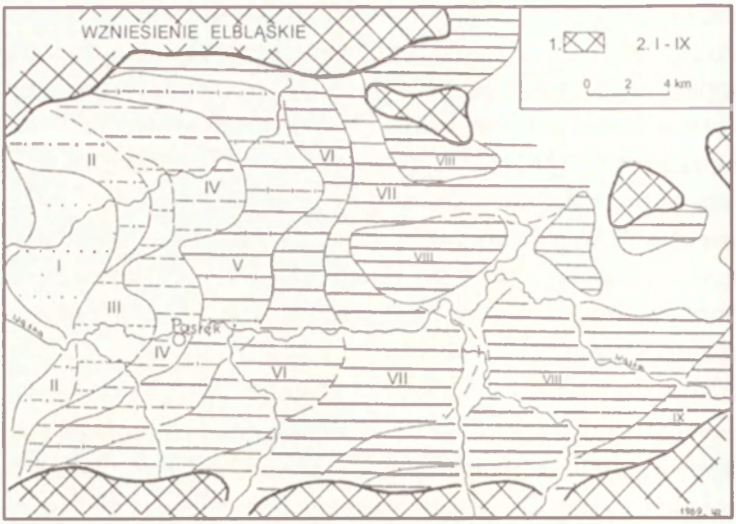

Fig. 1. Morphological sketch of terraces of Pasłęk ice-dammed lake (western part of the Varmia Lowland) after L. Roszko (1971). 1 – moraine plateau, 2 – ice-dammed lake terraces I–IX

Szkic morfologiczny teras zastoiska pasłęckiego (zachodnia część Niziny Warmińskiej) według L. Roszko (1971). 1 – wysoczyzna morenowa, 2 – I–IX terasy zastoiskowe

A trace of the existence of this is provided by the levels (9) sloping down to the west at altitudes of between 84–99 and 16 m a.s.l., as separated more or less distinctly by steps of heights 4–10 m. The genesis here is complex, since some parts have abrasive levels, cutting off the surface of the till, or else abrasive-accumulative levels built on a surface of stratified clayey silts interbedded with fine sands or clays. The sands were laid down by rivers spilling into the lake. Massive clays are present in places. These sediments here overlay till, and attain a maximum thickness of under 4.5 m. North of the Pasłęk Lowland – in the Braniewo Lowland belonging to the Varmia Lowland – L. Dauksza (1972) identified similar limnoglacial deposits, albeit present on only 2 levels. These were linked with the existence of a short-lived local ice-dammed lake on the floor of the Lowland. The prevailing thickness of sediments here is again 1–2 m (maximum 5 m), and there are no varve clays among them. This plain has kames, among other landforms. The Elbląg sheet of the *Geological Map of Poland* at a scale of 1:200 000 indicates both of these areas as till of clay facies, while the explanation to the map has A. Makowska (1979) stating that till is involved, locally with a marked layering of glacial origin. The genesis of this is not explained in detail, however. Reference is thus made to the views expressed by C. Gagel (1925) and B. Körnke (1930). These contradictions as regards points of view result from the fact that the areas in question and their sediments had not been studied in detail at that time.

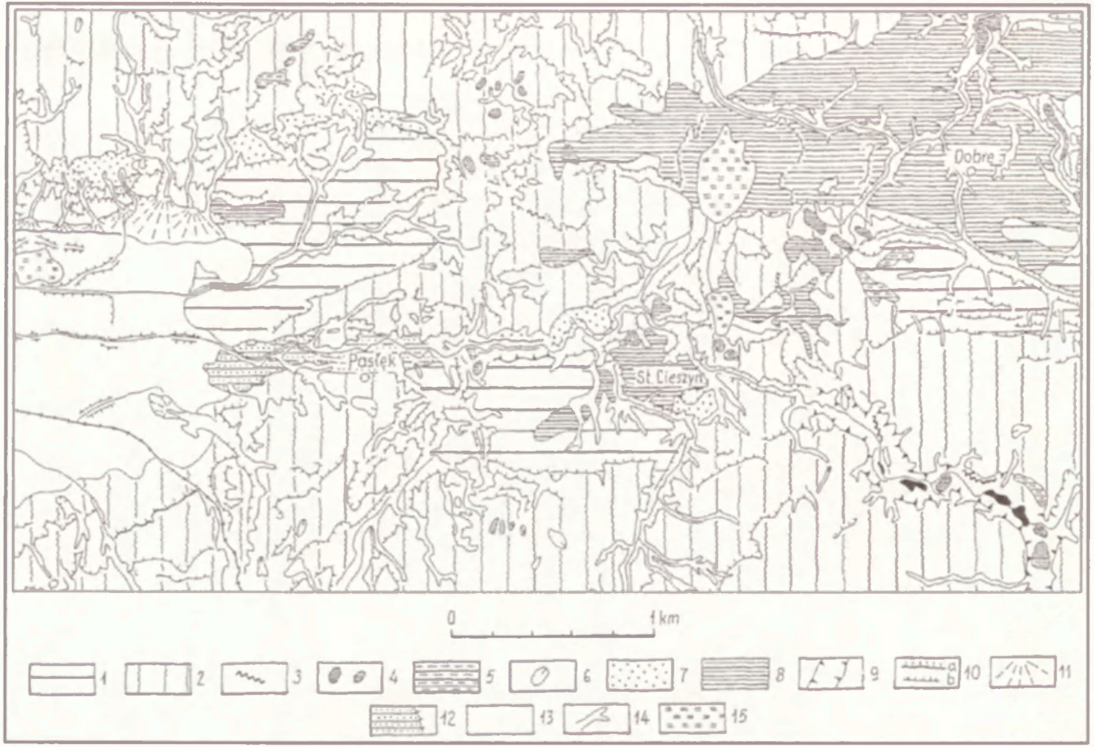

Fig. 2. Geomorphological map of the western part of the Warmia Lowland and its vicinity after K. Petelski (2001): 1 – flat moraine plateau, 2 – undulant moraine plateau, 3 – forms of crevasse accumulation, 4 – kames, 5 – forms of dead ice accumulation, 6 – kettles, 7 – outwash plains, 8 – areas of the ice-dammed lake, 9 – subglacial channels, 10 – edges and slopes: a – plateau, b – terraces, 11 – alluvial cones, 12 - accumulative terraces in fluvial valleys, 13 – the bottom of a fluvial valley, 14 – small valleys, ravines, 15 – peat plains

Mapa geomorfologiczna zachodniej części Niziny Warmińskiej i jej otoczenia według K. Petelskiego (2001): 1 – wysoczyzna morenowa płaska, 2 – wysoczyzna morenowa falista, 3 – formy akumulacji szczelinowej, 4 – kemy, 5 – formy akumulacji martwego lodu, 6 – zagłębienia wytopiskowe, 7 – równiny sandrowe, 8 – obszary zastoiskowe, 9 – rynny subglacjalne, 10 – krawędzie i stoki: a – wysoczyzn, b – tarasów, 11 – stożki napływowe, 12 – tarasy akumulacyjne w dolinach rzecznych, 13 – dna dolin rzecznych, 14 – dolinki, parowy, 15 – równiny torfowe

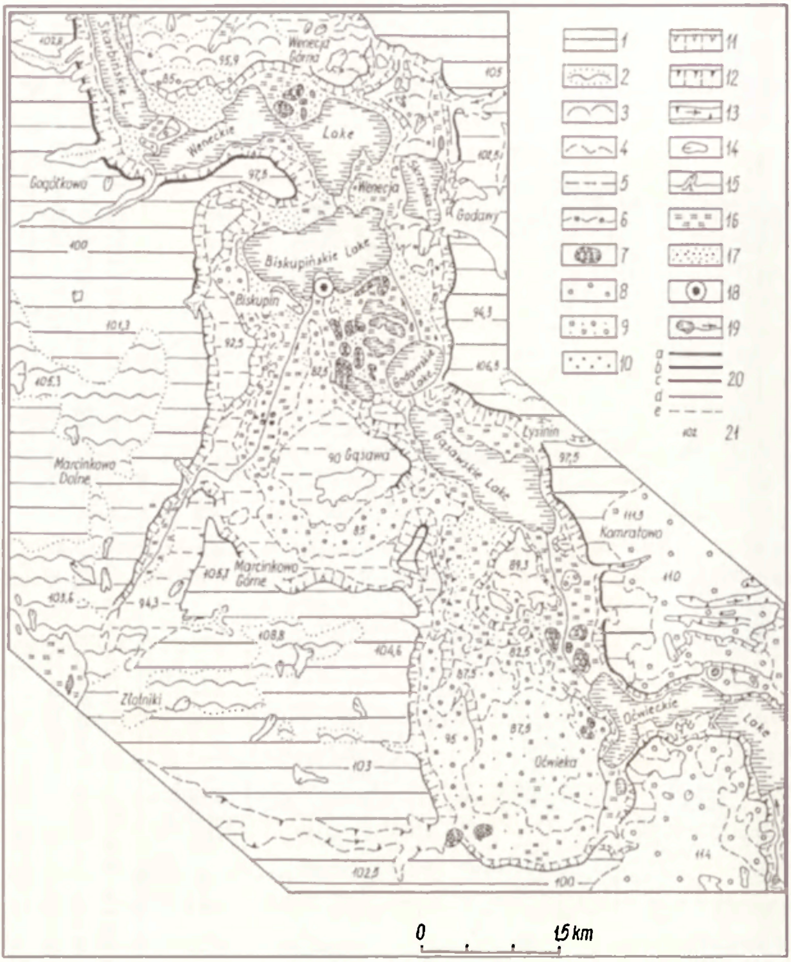

Fig. 3. Detail geomorphological map of southern part of the Żnin subglacial channel and its close vicinity
1 – flat morainic plateau, 2 – undulated morainic plateau, 3 – morainic hummocks, 4 – undulations, 5 – plains, 6 – morainic undulations with fluvioglacial cover within the Żnin subglacial channel, 7 – kames (hillocks, hills and ridges), 8 – outwash plain, 9 – plains built of ice dammed lake sediments (Pleni- Vistulian), 10 – Pleni- or Late Vistulian lacustrine sediments, 11 – slopes of the Żnin channel, 12 – slopes of subglacial fluvioglacial channels, 13 – meltwater valleys, 14 – kettles, 15 – denudative and erosive valleys, periglacial and Holocene, 16 – peat plains, 17 – Subatlantic lake terraces built of minerogenic sediments, 18 – location of fortified settlement (Lusatian Culture), 19 – lakes and rivers, 20 – height of slopes and escarpments: a: >20 m, b: 20–10 m, c: 10–5 m, d: <5 m, e: undistinct, 21 – absolute altitudes

Szczegółowa mapa geomorfologiczna żnińskiej rynny subglacjalnej i jej otoczenia
1 – płaska wysoczyzna morenowa, 2 – falista równina morenowa, 3 – pagórki morenowe, 4 – falistości, 5 – równiny, 6 – falistości morenowe z pokrywą fluwioglacjalną w obrębie żnińskiej rynny subglacjalnej, 7 – kemy (pagórki, wzgórza i wały), 8 – równina sandrowa, 9 – równiny zbudowane z osadów zastoiskowych (plenivistuliańskie), 10 – pleni- lub późnovistuliańskie

Still another picture is that presented by K. Petelski (2001), who states that – at the time the Pasłęk and Dobre sheets of the 1:50 000 *Detailed Geological Map of Poland* were being drawn up – he reported that the area with the Pasłęk ice-dammed lake distinguished by L. Roszko (1971, 1983) did not have distinguished levels of one great lake of this kind, but merely the sediments of small ice-dammed lakes on the area of ground moraine. What is most surprising is the claim that – in the Stary Cieszyn ice-dammed lake identified by him (Fig. 2), a small (40 x 60 m) depression in the till was found to have a 0.3–1.6 m layer of lacustrine sediments (clay and detritus gyttja) and marsh sediments (3 thin layers of peat) beneath a cover of 1.5–2.0 m of massive unlayered clay and silt. Radiocarbon dating showed dates of 13 420±220 years BP for the bottom peat and 11 800±220 years BP for the top peat, while the detailed work based on macrofragments of plants and palynology (Noryśkiewicz *et al.* 2002) reveals that these sediments were deposited partly during the Oldest Dryas and the Bölling and probably partly (the top part) during the Older Dryas. Very important palaeogeographic conclusions are to be drawn from these statements, namely that the lacustrine and mire sediments arose here in an ice-free area and later were drowned by ice dammed lake in which the aforementioned 1.5–2.0 m of clay and silt had been laid down. These sediments are confined to the northern edge, which is in the nature of an ice contact slope bordering with the plain below.

The morphological situation thus points to this last lake having been able to exist only when the neighbouring lowland was still overlain by great patches of dead ice extending above 70 m a.s.l. This lake therefore has nothing in common with the Pleniglacial ice-dammed lake, because it is several thousand years younger. It probably existed in the Alleröd Interstadial, or may be from the Younger Dryas. This example suggest that the Warmia Lowland did not have one huge ice-dammed lake, but a series of lesser ones of different ages.

While glacilimnic sediments were known to exist in many places in northern Poland, the mechanism by which they arose was long poorly known. However, this state of affairs has begun to change in the course of the last 30–40 years, thanks to the results of detailed studies summed up, *inter alia* by P.E. O'Sullivan (1983). In our sedimentological literature, the matter of the

terasy jeziorne, 11 – zbocza rynny żnińskiej, 12 – zbocza subglacjalnych rynien fluwioglacjalnych, 13 – doliny wód roztopowych, 14 – zagłębienia wytopiskowe, 15 – doliny denudacyjne i erozyjne, peryglacjalne i holoceńskie, 16 – równiny torfowe, 17 – subatlantyckie terasy jeziorne zbudowane z osadów minerogenicznych, 18 – lokalizacja grodu biskupińskiego (kultury łużyckiej), 19 – jeziora i rzeki, 20 – wysokość stoków i krawędzi: a: >20 m, b: 20–10 m, c: 10–5 m, d: <5 m, e: niewyraźne, 21 – wysokości bezwzględne glacilimnic sediments is discussed most fully in the work of K. Brodzikowski (1992a, b, c), who – in reference to newer works mainly by American and British workers and on the basis of his own studies from Kleszczów Graben (*i.a* Brodzikowski and Zieliński 1992) – was able to draw distinctions between 11 genetic variants of glacilimnic (ice-dammed lake) sediments, as well as to discuss the mechanisms by which they arose and their lithofacial shaping took place. However, the classification concerned itself with large ice-dammed lakes, while nothing is said about the sediments of the relatively small-sized examples that are prevalent in Poland. Also not taken account of are the ice-dammed lakes existing in dead ice. The concept of ice-dammed lakes is also extended markedly, since there is a separation of morphogenetic types of the following kind, into :

– terminoglacial lakes that are in contact with the ice front or edge,
– supraglacial lakes that developed mainly within dead ice,
– small basins that occurred on the delta plain,
– thermokarst lakes that arose from the melting of dead ice,
– intermorainic lakes that occurred in the forefield of inland ice between end-moraine ridges,
– proglacial lakes that occurred in the forefield far from the ice edge but were fed by ice-sheet waters,
– closed basins of melted-ice origin,
– sandur lakes.

It seems then, that for the above author, ice-dammed lakes were all of those in which glacilimnic sediments had been laid down. In my view, such a division is inconsistent and can not be justified as it ought to be. In several cases, account is taken of the relationship between a lake and glacial ice (i.e. with terminoglacial, supraglacial and proglacial lakes), while in other cases what is important is topographic location – with no definition of links with the ice-sheet ice (as with the intermorainic, sandur and delta lakes) or else processes by which ice melts (closed basins of melting ice origin and thermokarst lakes). In my opinion, the rather widely accepted concept of "ice-dammed lakes" dictates that the only acceptable proposal from K. Brodzikowski is that of the "termino- and supraglacial lakes". This term can probably be linked up with that of the "proglacial lake", which – in spite of being in non-contact with the glacial ice – is fed by the meltwaters from it, which are in turn a consequence of the truncation of glaciers and recession of the ice edge. In turn, the term "thermokarst lake" is not applied correctly here, while the other lakes are glacigenic, but not ice-dammed. In our literature another applied term is that of "extraglacial ice-dammed lake" applied by T. Merta (1978, 1986) in the case of the Warsaw ice-dammed lake. For that author reports – on the basis of study of sedimentary structures (ripplemarks, current) – that, although the north-side dam preventing outflow from this ice-dammed lake was the snout of the ice lobe in the vicinity of Płock, the lake was fed by extraglacial rivers flowing into it from unglaciated areas from the south and east. In Merta's view, there is a lack of evidence for the sediments being supplied by proglacial waters. In spite of this, varves do occur among the sediments of the ice-dammed lake. These mainly vary in thickness between 2 and 10 m, reaching a maximum of 17 m. It is debatable how old this ice-dammed lake is. S.Z. Różycki (1967) accepted that it existed during the Wkra Stage of the Middle Polish Glaciation, but after Eemian organic deposits were found to occur under the ice-dammed lake (Janczyk-Kopikowa 1974; Karaszewski 1974; Sarnacka 1982), it came to be believed that the lake existed later – according to M.D. Baraniecka and K. Konecka-Betley (1987) during the Middle Vistulian Glaciation. However, E. Wiśniewski (1990) was among those to oppose the claim, believing it to have been present during the period of maximal extent of the last (Upper Vistulian) glaciation. T. Merta (1978) allowed that variously-aged sedimentation basis could have existed in the Warsaw ice-dammed lake, even with sediments from them lying at the same level.

Data available to date leave the thickness of sediments from the old ice-dammed lakes of northern Poland as a matter of continuing uncertainty. However, the figures cited most frequently range between several and between 10 and 20 m, with that for the Gniew ice-dammed lake, for example, being 20 m maximum (Błaszkiewicz 2000), while the figure for the Złocieniec ice-dammed lake is 31 m (Maksiak, Mróz 1977). From this lake, only one profile of sediment of thickness 18 m has so far been worked on (Paluszkiewicz 1999). Sediments of greater thickness very often show rhythmic bedding, though varves are only present in some.

Many features point to the ice-dammed lakes of northern Poland having been short-lived, with their waters having flowed away before the younger lakes arose.

Pleni- and Late Vistulian lakes of the ice-dammed lake type emerged within dead ice, without any inflow of glacial waters from active ice. In my view, this type of lake should be taken to include those from which the outflow of water was dammed by dead ice, such that water was ultimately lost only when the ice dam melted. Such lakes most often arose in depressions, often those overlying dead ice at greater depth. In consequence these were in some way supraglacial lakes, whose sources of alimentation were mainly waters from the melting of dead ice. The best known lake of this type was present in the southern part of the Żnin subglacial channel (Niewiarowski 1992, 1994), in which there was also an older glacial lake that generated limnoglacial kames (Fig. 3). This lake existed beyond the dead ice conserving the deeper part of the subglacial channel. These were shallow lakes, and their levels fell in line with the stepwise melting of the dead ice. It is from this that we obtain the 2 or 3 levels present in sediment. Proof of existence is provided by the lacustrine sediments shaped in the form of fine sands with silts, locally calcareous silt or clays. Their thickness is not great, at between 0.5 and 2.5 m.

The intercalations of organic matter yield a radiocarbon date of 17 700±220 years BP. This is the only place in which such an old date has been obtained from lacustrine sediments, and in fact it may not be precise as pollen on secondary place has also been reported. From the indirect evidence, i.e. the mutual relationships between different landforms, it follows that is existed between 17.5 and 16.0 ka BP. The lake in question was a short-lived one, with the water pouring away as ice to the north in the Skarbin subglacial channel melted at least partially.

It is probable that lakes of this type were present in the Pleniglacial in the terminal basins surrounding Lakes Wieczno and Sitno on the Chełmno Morainic Plateau (Niewiarowski 1959). Here too, the lakes developed above deeply-buried dead ice and were even surrounded by it to at least some extent. The sediments of these lakes (clayey silts with an admixture of fine sand, or locally clays) were 2–3 m thick and show horizontal bedding (though no rhythmical bedding). It is highly probable that among the small ice-dammed lakes marked on the 1:200 000 geological maps there are some from Pleni- or Late Glacial times. To be added to the lakes of this type is the example in Stary Cieszyn that was discussed previously (Petelski 2001) (Fig. 2).

4. LANDFORMS ASSOCIATED WITH THE EXISTENCE OF THE ICE-DAMMED LAKE TYPE

Polish geomorphological literature makes the assumption that the most typical landforms associated with the existence of ice-dammed lakes are accumulative plains built of ice-dammed lake sediments. However, the term may be applied to such plains on which the thickness of sediments is great enough, at least 3–4 m, for these to level out uneven areas of the substratum and create plains. An example of such may be the plain of the Gniew ice-dammed lake (Błaszkiewicz 2000). Equally, the use of the term is not justified where the cover of ice-dammed lake sediments reaches only 1–2 m in depth, for then the true plain is the substratum of these sediments – most often a morainic plain built of till with a thin cover of sediments from ice-dammed lakes (as on the Braniewo and Sępopol Lowlands and in the Żnin subglacial channel). Large ice-dammed lake plains have levels (terraces) separated by edges that are more of less gentle and some 4–10 m high. The genesis of these has not yet been fully elucidated. Most often it is assumed that they appeared as a result of the stepwise lowering of a lake surface, though some – like J. Kondracki (1952) have considered them lake cliffs – something that has yet to be documented in the face of a lack of descriptions of geological structure (including that of the sediments lying at the foot of them).

There can nevertheless be no doubt that the edges present in the places of occurrence of short-lived and shallow ice-dammed lakes – with a thin cover of lacustrine sediments – are not lake cliffs, but most often morainic in character, and representative of subglacial relief of the substratum. The large ice-dammed lakes (e.g. the Baltic Ice Lake, American Great Lakes) generated extensive flat platforms (levels) of abrasive origin. The existence of this kind of level in Poland (Fig. 1) was anticipated only by L. Roszko (1971, 1983), with her view being questioned by A. Makowska (1979) and K. Petelski (2001). In a well-documented study on the subject of the Pyrzyce ice-dammed lake, R. Paluszkiewicz (2000) reports the presence of 3 levels that are accumulation levels from lake sediments at least 4--5 m thick, but have a surface of erosive origin since the top part was breached as the lake surface lowered. This is not then an abrasive surface *sensu stricto*. On account of the relatively small areas involved, north Poland has not been found to possess either beach ridges or large submerged deltas. All I could discover in the Polish literature was a claim by W. Słowański (1969) that this type of delta was present in the Biskupiec area of the Mazurian Lakeland.

In some cases – such as that of the Gniew ice-dammed lake of the terminoglacial type (but also those of the supraglacial type) – the accumulated sediments lie higher than the adjacent surface of ground moraine to the south (which was then covered by dead ice). After the lake water had flowed away and the dead ice had melted, an ice contact slope developed on the ground moraine, being formed from lake sediments. This is a case similar to the one of the Stary Cieszyn ice-dammed lake (Petelski 2001). Most likely there are more examples.

The surfaces of plains formed from the sediments of former ice-dammed lakes are usually very diversified. In the Pyrzyce ice-dammed lake, lower kames, eskers and drumlins which had arisen earlier were inundated – something which is attested to by their being covered by lake sediments (Karczewski 1968; Paluszkiewicz 2000). The higher kames formed islands. In northern Poland, the presence of kames and dead ice moraines within or near ice-dammed lakes was a quite widespread phenomenon. Also of widespread occurrence are subglacial channels and kettles, attesting to conservation during the time the lakes existed by dead ice, or else to the fact that a lake was supraglacial in character. Following the disappearance of the water, the aforementioned plains were incised by river valleys and modeled by various relief-forming processes subsequently. In the Warsaw Basin, for example, broad patches of ice-dammed lake sediments were only retained at the foot of the moraine plateau, separated by the broad-terraced valley of the Vistula.

5. LATE GLACIAL LAKES FORMED IN DEPRESSIONS LEFT BY THE MELTING OF BURIED DEAD ICE AND THEIR SEDIMENTS AND LANDFORMS

The Polish Lowland resembles the European Lowland in having a thick cover of Quaternary deposits in which the deglaciation process left buried considerable masses of dead ice in glacial and glacifluvial deposits. Both their thickness and that of the overlying sediments varied markedly, such that the periods taken for the ice to melt were also varied. The result of the ice melting was the creation of depressions of varying areas and depths in which lakes appeared. The discovery in the floor of the lacustrine sediments of thin layers of peat arising at the beginning of the Alleröd interstadial and then sinking to the bottom of even deep lakes created once the ice melted favoured the retention of a view that this warm Interstadial witnessed a total melting of this ice. For example, J. Stasiak (1963, 1967) claimed that – in the Mazurian Lake District – the flow-off of water from ice-dammed lakes led to the presence there up to the Alleröd of an area without lakes. In turn, after studying the sediments of 50 lakes, Więckowski (1978) voiced the opinion that all the lakes on the Polish Lowland appeared between 10 and 12 ka BP. Many new data were supplied by in-depth study within the framework of the IGCP-158 programme concerning hydrological and environmental changes in lakes and marshes (mires) – which was headed and summed up by M. Ralska-Jasiewiczowa (1987); as well as at the simultaneous work done in river valleys headed by Prof. L. Starkel. Among other things, this work showed that trends to the hydrological changes induced by climatic factors in lacustrine, marshland and river environments were similar and synchronous (Ralska-Jasiewiczowa, Starkel 1988). They also confirmed that the melting of the buried ice mainly took place in the Alleröd. However, it emerged later that the area of the Polish Lowland already had lakes in the earlier phases of the Late Vistulian period – in the Oldest Dryas, as well as lakes that only appeared in the Preboreal. This was obviously a result of temporal variations in the time of melting of dead ice (*i.a.* Niewiarowski 1990, 1994, 1998; Nowaczyk 1994; Błaszkiewicz 1998). Bearing in mind the time of appearance of lakes in the Late Vistulian and more rarely as late as in the early phases of the Holocene, as well as a certain differentiation to their sediments and to their relationships with buried ice, it is possible to distinguish between at least 2 generations of Late Vistulian lakes. These concern lakes arising in large depressions of the terminal basin type, as well as those in wide and deep subglacial channels, in which large masses of dead ice melting step by step were buried. The older lakes existed after the melting of nothing more than the upper part of the buried ice. These I have called Pre-Alleröd lakes. The younger lakes were in turn products of the period between the Bolling and the Preboreal. The older ones have been recognised in the Brodnica Lake District (Niewiarowski 1987, 1988a, b), in the Wel subglacial channel whose existence was better documented by T. Jaworski (2002), and later in the Żnin subglacial channel (Niewiarowski 1992). A trace of their presence exists in the form of lake terraces located high above the contemporary water level that are often formed from massive fine and medium sands and silts, locally interbedded by clays, lacustrine marls and chalk, without distinct organic matter and shells of malacofauna. The thickness of sediments rarely exceeds 4–5 m. The lack of organic matter does not allow for a precise dating. Lake terraces in the Ciche-Zbiczno-Bachotek and Wel channels link up with Drwęca terraces VII and VI, that are surely older than the Alleröd, because from that period there is only terrace II or III (Niewiarowski, Noryśkiewicz 1983). Hence, terraces VII and VI were described as terraces from the Oldest Dryas. The Pre-Alleröd lakes were also recognised in many other regions here and in Germany (*i.a.* Chrobok *et. al.* 1988; Nitz 1988). For example, in the Goplo subglacial channel, one of the profiles was found by B. Jankowska (1980) to contain lacustrine sediments from as long ago as the Oldest Dryas, though the later palynological work of B. Noryśkiewicz (Molewski, Noryśkiewicz 2000) revealed that the last melting of ice in the Goplo subglacial channel took place in the Alleröd.

Pre-Alleröd lakes were shallow, but the sediments in them vary. On the Brda outwash plain in the vicinity of Charzykowy, B. Nowaczyk (1994) reported lacustrine sediments from the Oldest Dryas sands silts, and calcareous sediments with numerous mollusc shells. Likewise, M. Błaszkiewicz (1998) reported well-documented lacustrine sediments that had been generated at the end of the Oldest Dryas and beginning of the Bölling in the terminal basin near Stara Kiszewa in Pomerania. These sediments are 6.7 m thick and are shaped in the form of clayey silts, silty clay, marl and massive sandy and clayey silts. In the silty marl there are remains of malacofauna with species typical of cool waters. Lacustrine sediments were laid down on the dead ice, such that as this melted the lakes became deeper. In the Bölling there was a reduction in depth as water passed out via overflow gaps.

The lake sediments from the Pre-Alleröd do not show marked rhythmicity. Lakes of this type may be categorized as glacial because they arose on dead ice and were largely fed by waters deriving from its melting. The remaining features left behind lakes of this kind are lake plains built of minerogenie and partly biogenic sediments. They most often take the form of lake terraces. An important role in the generation of relief was played by the outflow of water from these lakes – often catastrophic in nature – which resulted from the appearance of overflow gaps even several km long and up to 20 or 30 m deep. These gaps are very much behind the shaping of the post-glacial valley system.

More recent in-depth study of lacustrine sediments, in particular that done under Prof. Tobolski in the Lednica Landscape Park area and published in K. Tobolski (ed. 1991, 1998), points to the existence there – by the time of the Oldest Dryas – of many rather shallow (mainly 4-5 m) lakes in which bottoms sediments were mainly developed as sandy silts or calcareous silts with an admixture of organic matter. Detailed palaeobotanical study of plant macrofossils and pollen reveals that there was shrub tundra here at the time. T. Hoffa (1991) accepts that the narrow troughs present are of subglacial origin and were preserved by dead ice that had already melted by the Oldest Dryas. Some of the lakes in question were overgrown, as Lake Jemiolki – the bay of Lednica Lake – already in the had done by end of the Younger Dryas (Tobolski 1998). This research makes it clear that the Oldest Dryas already had climatic conditions favourable to the melting of shallow-buried masses of dead ice, and hence to the appearance of lakes.

Polish literature to date has assumed that the Vistulian Pleniglacial was followed by the first part of the Late Vistulian – the Oldest Dryas about 13,000 (radiocarbon) years BP. There is no precise determination of how long it lasted. At the same time, the accepted situation in Germany is that the Pleniglacial first gave way to the Late Weichselian older than the Bölling and to the Oldest Dryas warming as termed the Maiendorf interstadial by Menke (Bock *et al.* 1985). During this Interstadial, a tundra vegetation with *Betula nana*, *Hippophäe* and *Juniperus* developed in northern Germany. On the basis of varves present in the sediments of some lakes in Schleswig–Holstein and Lower Saxony (Merkt, Müller 1999), as well as in the Eifel region at Meerfelder Maar amongst other places (Litt, Steibich 1999), it proved possible to devise a varve chronology (of calendar years), according to which the Maiendorf interstadial lasted around 635–650 years, between 14 420 BP and 13 800 BP. In contrast, the subsequent Oldest Dryas lasted only 130–140 years. As far as I can determine, the analogous interstadial in Poland has not yet been studied palaeobotanically. However, K. Tobolski (1998) has asserted recently that the Oldest Dryas period identified in the sediments of lakes in Lednica Landscape Park should be treated as a temporary proposition, requiring augmentation with at least three episodes. However, in connection with the calendar chronology established not only in Germany, he allows that the beginnings of the existence of the Lednica Lake and its bay part known as Jemiolki may extend back to around 15,000 calendar years BP. Most important for our considerations is the report from German authors that the lake sediments of the Maiendorf interstadial are also present in lakes in the area under the last glaciation, *inter alia* in Schleswig-Holstein (Bock *et al.* 1995), in Lower Saxony (Merkt, Müller 1999) and in Brandenburg (just across the border from Poland), where I. Schulz and J. Strahl (2001) claim that the deep Kersdorfer subglacial channel of the Maiendorf interstadial witnessed only partial melting of dead ice and the appearance of shallow lakes. The full melting of this ice only took place in Allerод times, when deep lakes finally appeared.

We thus see major similarity with the course to the processes of melting of dead ice in some Polish subglacial chanels. Since there is a well-documented warming of the Maiendorf interstadial in Germany, it must have taken in Poland too, even if it took a rather different course. It was probably at that time that the shallow-lying dead ice melted, in particular where the shallow lakes had only accumulated minerogenic sediments, but also in others assigned in Poland to the Oldest Dryas, where organic matter is present.

A younger generation of Late Glacial lakes of glacigenic origin are those arising following the complete melting of dead ice in the period between the Bölling and the Preboreal. The great majority of these appeared in the Allerөd, which is to say the warmest Interstadial of the Late Glacial. Only the lake basins are of glacial origin *sensu stricto*, and these were only partially supplied by glacial water (from melting dead ice), and then only in their initial period of existence. Later, they were fed via precipitation and groundwater. A characteristic feature in them are the organic sediments that are mainly variants of gyttja and peat. Depending on environmental conditions and depth, the lake troughs were filled to varying extents by lake and mire deposits. Around 2/3 of all of them have now filled to the point of disappearance. Questions regarding these lakes are well-known and will not be addressed here.

6. CONCLUDING REMARKS

It follows from the data given above that, in the part of northern Poland encompassed by the last Vistulian Glaciation, the Plenivistulian and Late Vistulian (Late Glacial) saw numerous glacial lakes present, of varying genesis and age. The oldest had come into existence in the Pleniglacial, in the zones of occurrence of stagnant and dead ice. Arising from their sediments were numerous limnological kames playing an important role in the young glacial landscape. It was at this time, and partly also in the Late Vistulian, that a relatively large number of lakes of the ice-dammed lakes also came and went, being fed by the melt waters of active ice (termino-, supra- and partly proglacial lakes). An exception was the Warsaw extraglacial ice-dammed lake, which, in spite of contact with the edge of the ice, was fed by the waters of extraglacial rivers flowing from unglaciated areas (Merta 1978, 1986).

Younger and much rarer were the ice-dammed lakes arising thanks to the damming of outflow by dead ice, and being fed by the waters melting from it. All the lakes of the ice-dammed type were relatively small and short-lived. Only a few covered several hundred km^2. The sediments of only some of these have been studied in detail, such that both the sediments themselves and the landforms they create require further research. It is reported that, in the older part of the Late Vistulian – probably at the time of the earliest warming and in the oldest Dryas – two types of the so-called Pre-Allerӧd lakes came into being. The first of these were shallow lakes appearing following the total melting of a superficially-buried block of dead ice. The second type in turn comprised larger, but also shallow, lakes appearing on dead ice lying at greater depth and in greater thickness. As a rule, waters from these lakes flowed out via overflow gaps (spillways), while their sediments gave rise to elevated lake terraces. The greatest numbers of these on Polish territory are those formed in different types of depression and appearing after the complete melt of dead ice in the period between the Bӧlling and the Preboreal. Among these, it is only the lake basins themselves that are of glacial origin in the strict sense. This type of lake appeared in the younger part of the Late Glacial and had already disappeared by the Holocene. Only about 1/3 have come through to the present day.

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PLENI- I PÓŹNOVISTULIAŃSKIE JEZIORA GLACJALNE, ICH OSADY I FORMY TERENU WYSTĘPUJĄCE W MLODOGLACJALNEJ RZEŹBIE POLSKI PÓŁNOCNEJ

Streszczenie

W artykule omówiono jedynie te jeziora glacjalne, które były genetycznie związane z lodem lodowcowym i jego wodami roztopowymi, a świadectwem ich istnienia są odpowiednio osady jeziorne i formy terenu odgrywające istotną rolę w rzeźbie młodoglacjalnej Polski Północnej. Były to jeziora o różnej genezie, różnym wieku, o zróżnicowanych rozmiarach i głębokości. Wyróżniono wśród nich następujące typy.

1. Plenivistuliańskie jeziora glacjalne, które powstawały w czasie deglacjacji arealnej w stagnującym i martwym lodzie, w szczelinach, rozpadlinach, przetainach i obniżeniach powierzchni lodowej. Miały one rozmiary od kilku ha do kilkudziesięciu km^2. W jeziorach bezodpływowych zdeponowane były głównie osady drobnych piasków i mułków, warstwowanych horyzontalnie, często rytmicznie, a w częściowo przepływowych, również piaski z warstwowaniem typowym dla wód płynących. Osady te były zdeponowane w lodzie, a po jego wytopieniu wyłoniły się jako wzgórza, stoliwa i plateaux, zwane kemami limnoglacjalnymi. Do tego typu kemów należą też terasy kemowe pochodzenia limnoglacjalnego.

2. Pleni- i późnovistuliańskie jeziora glacjalne typu zastoiskowego. Do plenivistuliańskich jezior zaliczono:

a/ jeziora zastoiskowe terminoglacjalne (Brodzikowski 1992 a, b, c) podparte przez krawędź lodową;

b/ jeziora zastoiskowe proglacjalne występujące na przedpolu krawędzi lodowej, niekoniecznie na kontakcie z nią, ale zasilane wodami proglacjalnymi. Jedynie Merta (1978, 1986) wyróżnia u nas warszawskie zastoisko typu ekstraglacjalnego argumentując, że osady w nim zdeponowane były przynoszone przez rzeki ekstraglacjalne (nielodowcowe). W jeziorach zastoiskowych osadziły się typowe osady glacjolimniczne, głównie mułki i iły, przeważnie horyzontalnie i rytmicznie warstwowanie, ale stosunkowo rzadkie u nas, z występującymi w nich warwami. W jeziorach zastoiskowych krótkotrwałych, o małej miąższości osadów jest większe ich zróżnicowanie, a geneza niektórych z nich (np. niem. *Deckton*) jest nadal dyskusyjna, stąd też zasięg jezior zastoiskowych, w szczególności na Nizinie Warmińskiej (rye. 1 i 2) i Sępopolskiej, jest różnie interpretowany.

Wyróżniono jeszcze jeziora typu zastoiskowego, przeważnie supraglacjalne, powstałe w obrębie martwego lodu w końcowej fazie jego topnienia. Tamę dla spływu ich wód stanowił martwy lód. Istniały one w plenivistulianie (ryc. 3), ale mogły też pojawić się jeszcze w późnym vistulianie (Petelski 2001). Osady tych jezior nie wykazują rytmicznego warstwowania. Wszystkie istniejące u nas jeziora zastoiskowe uległy spływowi. Głównymi formami rzeźby terenu związanymi z jeziorami zastoiskowymi są: akumulacyjne równiny, często sterasowane, zbudowane z osadów o większej miąższości, stopnie terenowe (krawędzie) oddzielające terasy, których gencza nie jest jeszcze w pełni wyjaśniona, oraz wyjątkowo równiny abrazyjne, delty

i stoki kontaktu lodowego. Osady o małej miąższości stanowią jedynie cienkie pokrywy form już istniejących (głównie równin morenowych) i nie można ich nazywać równinami zastoiskowymi.

3. Późnovistuliańskie jeziora glacjalne, powstałe z wytopienia się lodu pogrzebanego. Autor wyróżnia wśród nich jeziora starsze (prealleródzkie), powstałe w czasie pierwszego ocieplenia po pleniglacjale, zwanego w Niemczech Maiendorf i w najstarszym dryasie, po wytopieniu się jedynie górnej części pogrzebanego lodu lub po wytopieniu się płytko pogrzebanych brył martwego lodu. Były to jeziora płytkie zasilane głównie wodami z topnienia tych brył, w których osadziły się głównie osady minerogeniczne, albo z niewielkim udziałem osadów organicznych. Zachowały się one głównie w postaci wysokich teras jeziornych. Spływy wód z tych jezior sprzyjały powstaniu przełomów przelewowych, odgrywających dużą rolę w kształtowaniu się polodowcowej sieci dolinnej. Młodszą generację jezior stanowią jeziora, które powstały po całkowitym wytopieniu się pogrzebanego lodu, w okresie od böllingu po okres preborealny, w których w sensie ścisłym, glacjalnego pochodzenia są niecki jeziorne, a udział wód lodowcowych istniał w nich jedynie w początkowej fazie ich powstawania.

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Małgorzata Gutry-Korycka

THE MORPHOGENESIS OF VALLEYS AND CONDITIONING OF RIVER OUTFLOW

1. THE MORPHOGENESIS OF VALLEYS AND PROGLACIAL CHANNELS – A KEY TO THE ASSESSMENT OF CONTEMPORARY WATER CIRCULATION

The research carried out in relation to numerous river valleys provides evidence that valley development is highly complicated, being characterized by a changing pattern of morphological, lithological and structural forms. According to L. Starkel (1990, 1996), the valley pattern of the East-Central European Lowland is connected with the maximum extent of the most recent glaciation. At that time, water from melting inland ice used to spill over into the Volga and Dniepr rivers, forming channels of marginal outflows along the ice sheet scarp (Fig. 1).

Contemporarily the courses of these wide wetland proglacial valleys run from the old Neman, via the Vistula and Oder to the Elbe (Figs. 1, 2). Over the area distant from inland ice – changes of climate have left their mark in the form of segment erosive dissections and postglacial accumulation. A further impact came with changed forms of valleys and sizes of river channels. In relation to the aggradation of phases, rivers can be divided into sections of considerably diversified genesis, e.g. the polygenetic valleys of the Vistula, Oder, Bug, Narew, Noteć, San and many other rivers (Fig. 1).

L. Starkel (1996) has stated that, before the glaciations, the waters of the Upper Vistula had flowed westward towards the Oder river, as well as partially discharging into the Dniestr (Black Sea basin catchment). After breaking through a gap in the South Polish Upland, the water changed direction towards the north, that is to say towards the pre-Baltic. These old valleys had been covered by sediments of several glaciations which dammed up the flow and forced the process of formation of proglacial valleys, shifting them still more northward along with the backwasting of the ice sheet (Figs. 1,2). Therefore the Vistula valley lowland is formed by alternate sections of basins and gaps, while erosion scarps underline the considerable dissection of the previous

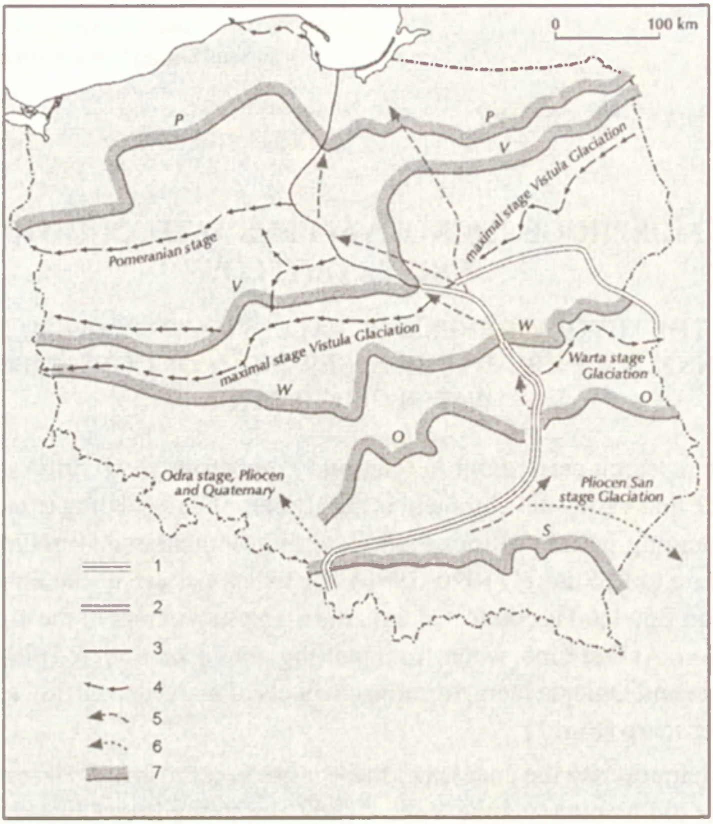

Fig. 1. The age of different sections of the Vistula valley and changes in flow directions (after Starkel et al., 1990)
1 – chain of lower Vistula river courses after retreat of last glaciation, 2 – chain of Vistula and Bug rivers without change from last interglacial, 3 – upper Vistula river valley older than Glaciation (excluding chain valley upstream of Kraków including after San Glaciation), 4 – directions to river flows via proglacial valleys during stay of continental ice, 5 – old part of Vistula flow in interglacial, 6 – old flow of pre-Carpathian glaciations or during maximal glaciation, 7 – limits of glaciations: S – San; O – Oder; W – Warta; V – Vistula; P – Pomeranian

Wiek różnych odcinków Wisły i zmiany kierunków odpływu (wg Starkla i innych, 1990)
1 – bieg dolnej Wisły po ustąpieniu ostatniego lądolodu, 2 – bieg Wisły i Bugu nie zmieniony od ostatniego interglacjału, 3 – górna Wisła o biegu starszym od zlodowaceń (poza odcinkiem powyżej Krakowa włączonym po zlodowaceniu Sanu), 4 – kierunki odpływu pradolinami w czasie postojów lądolodu, 5 – dawne (kopalne odcinki przepływu Wisły w interglacjalach, 6 – dawne przepływy rzek na przedpolu Karpat przed zlodowaceniami lub w czasie zlodowacenia maksymalnego, 7 – zasięgi głównych zlodowaceń: S – Sanu, O – Odry, W – Warty, V – Wisły, P – fazy pomorskiej

postglacial plateau. Over the lowlands, widely-formed erosive platforms are filled by thick layers of alluvial fluvial facies, presenting a clear system of cut and built terraces that is well marked in the landscape.

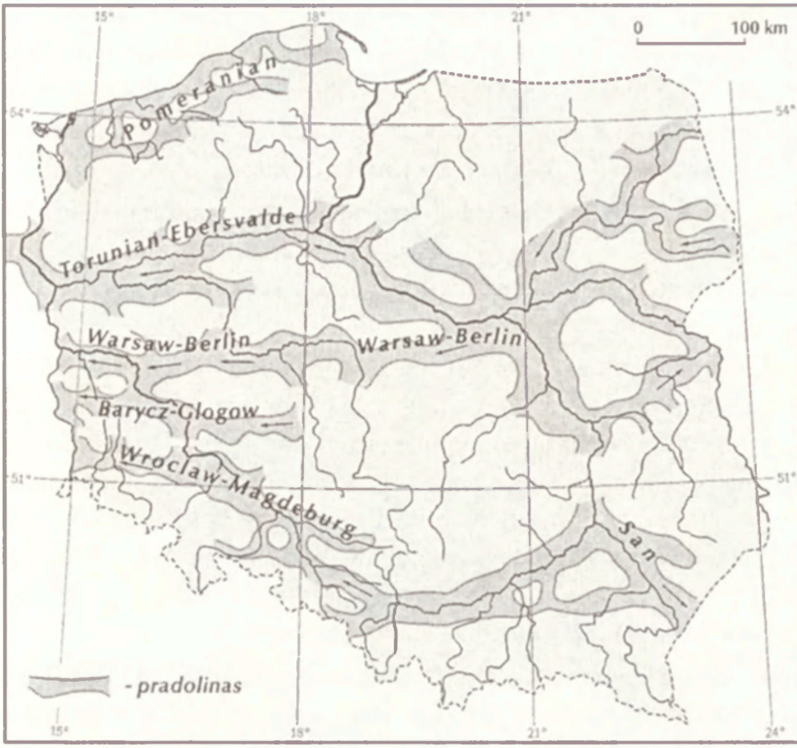

Fig. 2. Main chains of proglacial channels (Barbag, Dylik 1968; Jankowski 1994)
Główne ciągi pradolin (Barbag, Dylik 1968; Jankowski 1994)

The polygenetic and most frequently complex character of valleys comes as a result of past high waters and floods, as was stressed by L. Starkel (1996, 2001), A.T. Jankowski (1994) and others. Valley bottoms are areas affected by high waters flowing out of river channels, beyond a river's bank lines. By dating sediments and materials from the palaeochannels left by a river, and checking off registered levels of flush waters in historical times, it is possible to reconstruct valley changes along with flooding phenomena in the past (Starkel 1998, 2001).

Bottoms of river valleys contain sediments from the Pleistocene or earlier, also those of the period at the end of the glaciation and beginning of the Holocene, i.e. the most recent 10–12 thousand years (Starkel 1998).

Floods in successive millennia have contributed to transformations of valley bottoms, their directions in particular sections of the river flow often being diverse.

After an analysis of the history of river valleys in Holocene, L. Starkel (1998) has come to the conclusion that mountain water flows have transformed rock forms in river beds when water effected a breakthrough into the massive rock. Piedmont rivers are characterized by side shifting beds and an up-lifting of flooded plains covered with mud sediments. Smaller sub-Carpathian river valleys were often enlarged as alluvial materials were deposited into new erosive spillways and alternate beds (Starkel 1998).

The system of river valleys was basically formed in the Pleistocene and Holocene, that is to say during the last two million years. A. Jankowski (1994) expresses the view that the dissections of river accumulation sediments are connected with interglacials.

However, the Holocene is the period of greater stabilization of the river system, though it has at the same time seen the influence of human impacts upon valleys and river channels grow. Several sequences of accumulation and erosion phases brought about polycylic periodical development phases of valleys and river channels (Jankowski 1994), while uplifting movements in the Carpathian and Sudety Mountains along dislocation edges (10–20 m) have caused an erosive reaction on the part of rivers (a breakthrough by the Vistula across the Upland), as well as valley incisions.

However, the beginnings of the network of proglacial valleys (Fig.1) should be linked with maximal extent of the San glaciation (*Geography of Poland* 1991). Confirmation is therefore offered for opinions about links between the formation of these, the time factor and the direction of river outflows during the Warta and Vistula glaciations along with phases corresponding with them. The consequence of the polygenetic character of those late Pleistocene phases linked with the disappearing ice sheet is provided by the course of the main extent of the proglacial valleys, among them the northernmost one called Pradolina Pomorska in Polish (Fig.2).

As L. Starkel (1996), A. Jankowski (1994) and others have stressed, the Holocene is not so much marked in valley genesis as the Pleistocene. The main period of further modification, of what was an already formed river system, is then to be seen in historical times.

The polygenesis of the valley, the mode of channel formation and dynamics of flow have determined the development of the different lithological sediments. Such complex valley and proglacial valley forms are capacious, well renewable water bearing structures (Falkowski 2002).

The alluvia of the proglacial valleys are most frequently composed of two sedimentation cycles: fluvioglacial in the bottom layer and fluvial in the top layer (Blaszyk, Gorski 1978). Fluvioglacial formations are coarse sands, gravels and all-up aggregates of a considerable thickness, as a rule without organic matter. The overlying alluvia forming the contemporary valley bottoms (all together 5% of the area of Poland) are less thick. They present fine sands accumulated and river drift sand collected where water transportation ceases to function (*Geography of Poland* 1991). Along valley sections with a more extreme gradient, stream muds are often accumulated with a predominance of coarse sand strata finely sorted (Bajkiewicz-Grabowska, Mikulski 1993). Facies accumulated in abandoned channels form alternate river, lake, marsh and marsh-peats sediments and are characteristic for flood plains and flooding sheets. Along river sections with moderate flow, sediments contain fine silt and all aggregates (*Geography of Poland* 1991). Valleys frequently flooded have proper mud alluvia accumulated, without a clear differentiation into genetic levels. Where intervals between successive floodings are longer, the characteristic soils are alluvial, developing as gley- or humic muds in relation to the persistence and depth of water.

2. PATHWAYS OF WATER CIRCULATION IN VALLEYS AND PROGLACIAL VALLEYS

The differentiation to pathways of water circulation and infiltration along and across a valley and river channel brings limits to the transformation of high waters and alimentation from elevations around, setting therefore the constant or periodic hydraulic contact with underground waters located in alluvia. Large, widely formed valleys with clearly-marked terraces facilitate complex mechanisms of water circulation inside a frequently well-bounded valley. The above is a consequence of the hydrodynamic relation between the filled channel and the bearing strata. Due to insufficient familiarity with valley morphogenesis, the actual presentation of water circulation and infiltration is based on indirect presumptions.

Alluvial waters flow parallel or at an angle to the river channel as an independent stream, in line with the gravity force conditioned by the incline (Fig. 3), with speed rarely exceeding $10^3 m/s^{-1}$. The flow net is constructed so that ΔW and ΔX are approximately equal (Thomson 1999), where the aquifer thickness is in the square and $\Delta h = h_1 - h_2$ is the contour interval between equi-potential lines. However, since $\Delta X \approx \Delta \approx W$, the flow through any square is simply:

$$q = k \cdot y \cdot \Delta h \qquad 1)$$

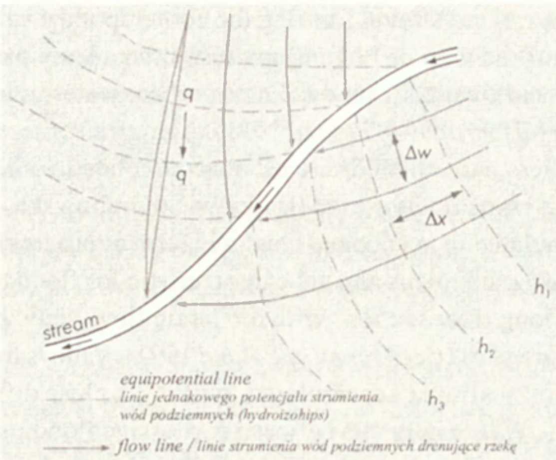

Fig. 3. A planimetric view of groundwater flow network and river (Thompson 1999)
Planimetryczny rzut strumienia dopływu wód podziemnych do rzeki (Thompson 1999)

Groundwater flow through a square is computed from the hydraulic conductivity (k) and the geometry of the flow net, that is spacing the flow lines so that they equal the spacing of the equipotential lines.

The alimentation and circulation of alluvial waters derive from the infiltration of meltwaters and precipitation via hydrogeological windows, from the discharge of lateral waters from the hills, from two dimensional filtration from river beds and also – even perhaps often – from waste removal (under pressure). The proportionality between alimentation of a direct and indirect nature is a consequence of hydrodynamic conditions but also of climate (Fig. 4). Prevailing in a humid climate is a specific equilibrium based on the fact that a river receives underground water via percolation under average or low states of rivers, while high waters spill over the channel and support underground waters damming up alluvial waters (Pietrygowa 1989; Gutry-Korycka, Gąsowska 2000, 2001). In arid and semi-arid regions, waters from seasonal or occasional river corridors percolate beneath the water aquifer (Fig. 4), and in this way make the process still harder to envisage.

3. THE FORMING OF PEAK HIGH WATER IN RIVER VALLEYS

Forms by which flood waters reach their peaks depend on the rate of flooding and elevation of the wave, on the longitudinal gradient to the bed incline, on the crosswise section of the valley, the length of the river and the time at which peak high water occurs. Under the influence of the above factors, the

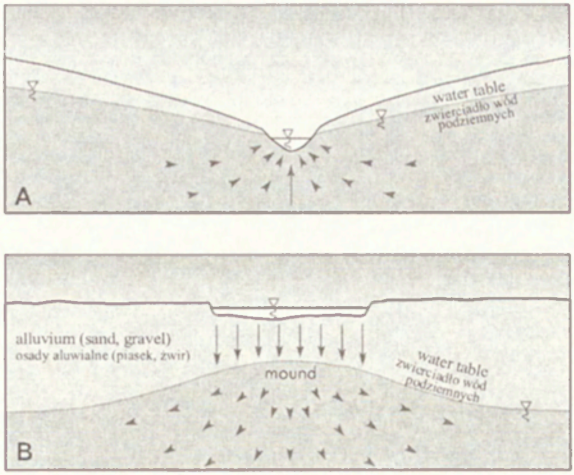

Fig. 4. In humid regions (A) a stream channel receives groundwater through seepage. In arid regions (B) stream water seeps out of the channel and into the water table below (Strahler 1996)

W regionach wilgotnych (A) rzeka otrzymuje wody podziemne w wyniku przesączania. W regionach suchych (B) wody z koryta rzek wsiąkają do wód podziemnych poniżej (Strahler 1996)

form of the spate of water moving along the channel undergoes partial transformation and gradually flattens. Valley indicators of subterranean waters, as W. Chełmicki (1991) showed, are in close hydraulic contact with the river, and their hydrological regime reflects the river rhythm.

The morphogenesis of typical and proglacial valleys may considerably condition the transformation and a form of the peak high water hydrogram, and in this way disturb the hydraulic link between the river and alluvial waters (Guidelines 2002). Regularities and anomalies in their courses in the valleys and channels of the great Polish rivers have been disclosed, among others by Z. Pasławski (1960, 1967), Z. Pasławski and K. Olejnik (1967), L. Skibniewski (1960 a,b, 1961, 1962, 1969) and I. Dynowska (1969, 1970).

L. Skibniewski (1969) is of the opinion that spate waters infiltrate down into the valley and proglacial valley. Capable of serving as illustrations here are the diminishing of outflow along the Oder and Warta. According to Z. Pasławski (1967) and Z. Pasławski and K. Olejnik (1967), losses of waters from river channels are used for the alimentation (filtration) of aquifers in the proglacial valleys. Losses may be irrevocable, causing a negative anomaly of flow, or else be merely apparent. L. Skibniewski (1961, 1969) says that apparent anomalies may be seen as feeding a river during low water or as water filtration losses from channels during the high water period.

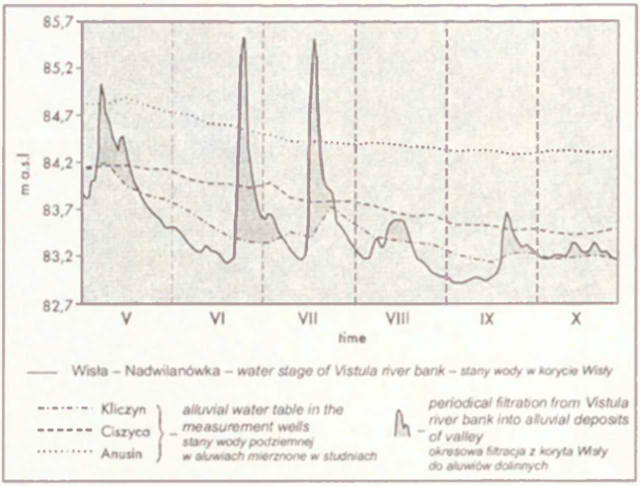

Fig. 5. Water stages of the Vistula river and alluvial groundwater table in the valley during selected floodings with river channel filtration (Gutry-Korycka, Gąsowska 2000 2001)

Przebieg stanów wody Wisły powyżej Warszawy i stanów wód podziemnych podczas wezbrania, ilustrujący filtrację wód korytowych do aluwiów (Gutry-Korycka, Gąsowska 2000 2001)

Some authors suggest that the volume of filtration losses for alimentation of alluvial waters in a valley is in proportion to the scale, duration and extent of river water rise. Also perhaps connected arc the saturation level, as controlled by the water table (Różycki 1996; Pietrygowa 1989; Gutry-Korycka, Gąsowska 2000, 2001). River bank infiltration and drainage of alluvial water occur in turns depending on the river water regime. Especially intensive filtration occurs during outflows from river beds. M. Gutry-Korycka and G.Gąsowska (2001) found that the phenomenon of the supplementation of underground waters in the Vistula valley during flooding frequently takes place in the zone of flooded terraces and above, or even affects the whole valley or proglacial valley (Fig. 5). Filtration losses correspond to an inertia of alluvial waters, whereas a significant hydraulic drop of the water table lessens bank infiltration more distantly from the river bed. The process as related to mountain rivers was evidenced by D. Małecka (1973, 1974) and D. Małecka and A. Witkowski (1981). The arrangement of isohypses of alluvial waters elevation related to the Dunajec river bed indicates the clear drainage character of the valley during base water, as caused by the change in the hydraulic gradient of groundwaters (Fig. 6). A decisive influence on the dynamics and orientation to the circulation of high waters in the valley is also exerted by the balance of water in the corridor. Transformation of a flood wave results in a hydraulic drop of the alluvial water table in gravels,

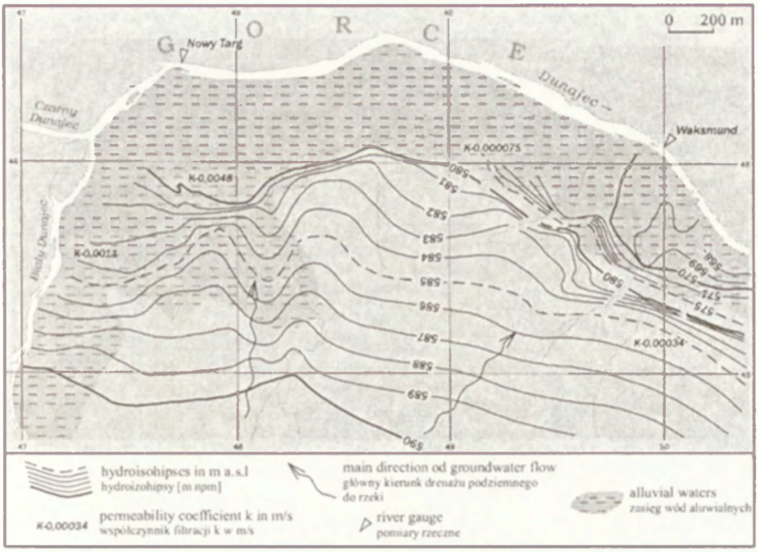

Fig. 6. Hydroisohipses distribution in Dunajec river valley (Małecka 1974)
Rozkład hydroizohips wód podziemnych w dolinie drenowanej przez Dunajec koło Nowego Targu podczas niżówki (Małecka 1974)

or boulders with an admixture of clay. Materials pile up and as there is simultaneous quick alimentation (from the south, in accordance with the gradient and direction of drainage) the intensity of the phenomenon is enhanced. The process repeats during risings in river levels due to melting water, falls of rain or mixtures of the two.

The storage of alluvial waters in a typical or proglacial valley is determined by the character of the hydraulic connection (interactions) with the river. Knowledge of the water-bearing hydrogeological structures in broad proglacial valleys is mostly weak, due to difficulties with drilling in sandy-gravel sediments and registering the results (Błaszyk, Górski 1978). The full cycle of valley sedimentation in lowland rivers as related to recent glaciation consists of fluvioglacial sediments and Holocene accumulation containing sand, gravel, boulders and silt.

A characteristic feature is sedimentation during periglacial periods, with a small admixture of fossil plants.

The scheme of water circulation in a proglacial valley includes two zones: of sub-areal infiltration and of in-flow and out-flow from the river channel and elevations (Fig.7).

The configuration of hydroisopies (lines of location of similar pressurized water tables) in the Oder proglacial valley along the Barycz-Głogów section

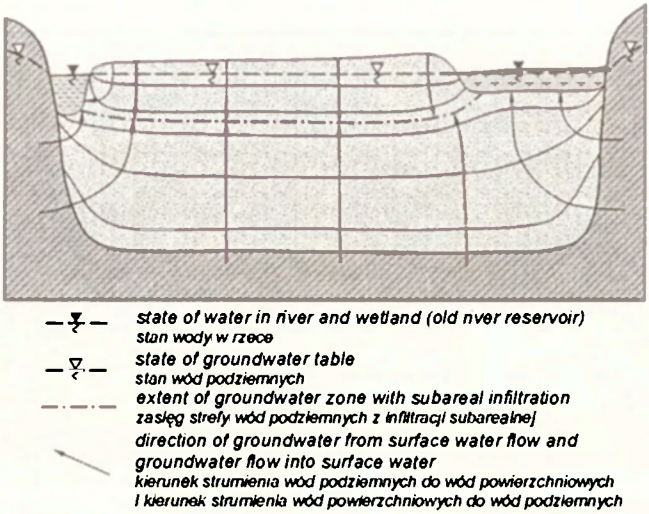

Fig. 7. Scheme for water circulation in proglacial valley system (after Blaszyk, Górski 1978)

Schemat systemu krążenia wód w pradolinie (wg Błaszyka, Górskiego 1978)

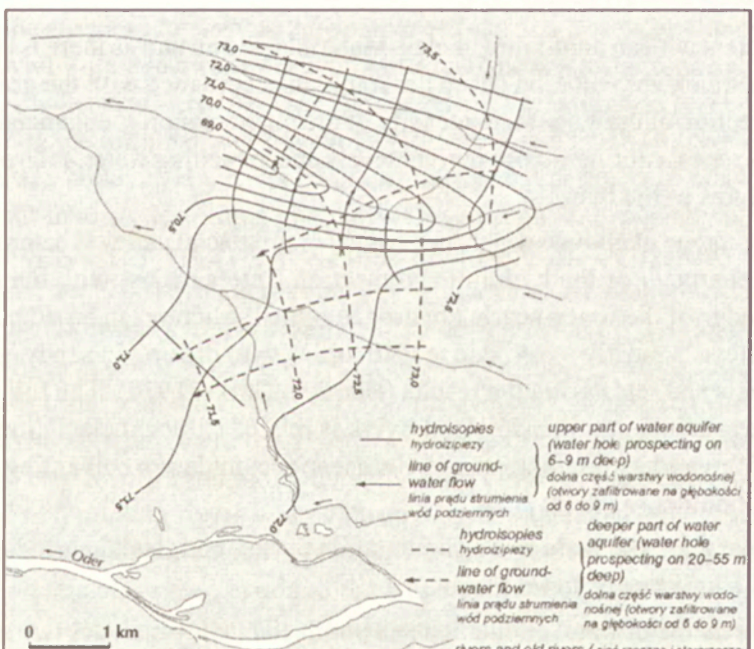

Fig. 8. Piezometric area of upper and deeper part of water aquifer in Barycz-Głogów proglacial valley (Oder river) (after Błaszyk, Górski,1978)

Powierzchnia piezometryczna górnej i dolnej części warstwy wodonośnej w pradolinie barycko-głogowskiej (wg Błaszyka, Górskiego 1978) comprises water circulation in upper and lower zones, having different capacity, direction, rate of filtration and drainage (Fig.8).

The dynamics of subterranean waters in older valleys of hillslope rivers, for example the Pilica and Wolborka – depend on the isolation of aquifers in Jurassic and Cretaceous sediments as related to those of the Pleistocene and Holocene (Maksymiuk 1971, 1979; Moszczyńska 1986). Interaction between the regime of river water and subterranean water in Cretaceous formations is different and clearly asymmetrical, imposed by the geological structure of the karstic substratum. The regime for the alluvial water table and for waters at deeper levels is linked hydraulically with the peak high water level in the river, but with a delay in regard to the water regime in the Pilica river. The reaction of water regimes in weathered and cracked Cretaceous formations to changing high waters in the river is extinguished with growing distance, as is evidenced by regime amplitudes. Similar rules have been observed by J. Moszczyńska (1986) in the valleys of the Wolborka and Moszczenica. The genetic complexity of valleys in the Polish Lowland, as considered in regard to different ways that peak high water forms, as well as the amount of groundwater, was presented by T. Falkowski (2001, 2002).

The above examples have illustrated the complex hydraulic configuration of alluvial waters in morphogenetically dissimilar valley sections. Resemblance in regard to the means and rate of alimentation, filtration and drainage over time is a derivative of valley morphogenesis in the regional and local configuration.

4. DIGITAL SIMULATION OF GROUNDWATER DRAINAGE IN THE PROGLACIAL VALLEYS OF RIVERS

The role of a river in the natural drainage of underground waters was formulated mathematically in the middle of the 19th century by French hydrologists. Among others, Dupoit used an equation to describe the link between the shape of an underground water table and the outflow into a water course.

This equation is valid only where there is good (constant) hydraulic contact between the water bearing layer or layers with a free water table and the river (boundary condition I and II). However, the formula does not hold good in regard to the borderline contact between a river channel and the aquifer layer as a type III boundary condition III, it is determined by the occurrence and limited permeability of bottom sediments filling the channel of a water course.

Mathematical description of relations between hydraulic channel waters and alluvial waters in the valley or proglacial channel becomes yet more complex as a result of the exploitation of intakes of subterranean water (Kostecki 2000). The hydrological measurements and morphogenesis of such complex hydrodynamic configurations allow for an assessment of the exactitude with which the groundwater circulation system has been identified. This can be done as the simulated process under natural conditions, as well as those enforced by the presence of water intakes or damming devices (drops, barrages or sluices) placed in the bottom of a valley or proglacial channel.

Simulation models have been used in regard to the proglacial valleys of great rivers, e.g. the Warta – HYDRYLIB (Kostecki 2000), the Oder (Blaszyk, Górski 1978) and the Biebrza – SIMGRO (*Hydrological System*... 2002).

However, the choice of a simulation programme entails many numerical and measurement-related assumptions (Kostecki 2000; *Hydrological System*... 2002) which allow for:

– the reconstruction of the hydrological structure with the assumed exactitude corresponding to the achieved degree of understanding and the morphogenesis of the valley,

– precise reconstruction of the existing lay-out of surface waters, along with the infrastructure of intakes, radiant wells, piezometers and infiltration ponds,

– an analysis and identification of filtration parameters resulting from the morphogenesis of valley alluvia,

– reconstruction of the courses of river channels and oxbows in relation to the hydrodynamic pattern,

– numerical representation of the hydraulic contact zones (contacts) between the river and underground water table (or lack of same), as corresponding with any one state of valley channel and alluvial waters.

The algorithm bases itself on an iterative method which allows for the mathematical description of the filtration field fixed under conditions of a pressing groundwater table in an agreed period (hour, day and night) in regard to the data block. The computing scheme arises out of an understanding of the hydraulic conductivity of the water-bearing layers taken from drilling test pumpings, as obtained during the exploitation of a water intake.

The numerical simulation entails the resolving of the reverse task, that is to say the reconstruction of the level of the subterranean water table through the correcting of exactly determined filtration parameters and basic boundary conditions in a valley.

The computational algorithm entails a description of the filtration process performed in a constant physical field, in accordance with Darcy's linear law:

$$V = - k \cdot \frac{\Delta H}{\Delta l} \qquad 2)$$

where: V – filtration velocity (m/s^{-1}),

k – coefficient of filtration (m/s^{-1}),

ΔH – difference of piezometric pressures (m)

Δ l – route (distance) of filtration (m)

The process of filtration established as conditioned by strata thrust may be presented in algebraic form as follows:

$$\frac{\delta}{\delta x} = \left(kx \cdot m \frac{\delta H}{\delta x}\right) + \frac{\delta}{\delta x}\left(ky \cdot m \frac{\delta H}{\delta x}\right) + q_{vertical} \qquad 3)$$

on the assumption that

$$\frac{\delta H}{\delta t} = 0$$

Taking into consideration the isotropic character of the water bearing layer the above equation assumes the form:

$$k \cdot m\left[\frac{\delta}{\delta x}\left(\frac{\delta H}{\delta x}\right) + \frac{\delta}{\delta y}\left(\frac{\delta H}{\delta y}\right)\right] = q_{vertical} \qquad 4)$$

$$k \cdot m\left(\frac{\delta^2 H}{\delta^2 x}\right) + \left(\frac{\delta^2 H}{\delta^2 y}\right) = q_{vertical} \qquad 5)$$

where:

$k \cdot m = T$ – the hydraulic conductivity of the layer (m^2/s)

k – also the coefficient of filtration (m/s)

m – thickness of the water-bearing layer (m)

Stationary filtration in a river valley requires that the following assumptions be made (Kostecki 2000):

– type I boundary condition (Dirichlet), where pressure at a given point on the areal border does not change with time, function H'=f(x,y,z) = const; blocks with type I condition are placed on the boundary of the digitized area;

– type II boundary condition (Neumann), with flow on the boundary at any point is constant, such that the function of the form Q=f(x,y,z)=const; blocks of the mathematical model for type II conditions serve e.g. in representing the magnitude of infiltration and also the intensity of well exploitation. In that case blocks are marked as corresponding to type II conditions (infiltration " – ");

– boundary condition III (Dirichlet-Neumann) – interpreted as discharge of an intensity that is a function of a given pressure Q=f(H) for H=const; with a digitized block net with type III conditions – surface waters are presented as rivers or lakes (oxbows). Modelled in such blocks is the effect of filtration (drainage or infiltration) of water through a weakly permeable stratum to the water-bearing stratum.

Parameters of the weakly-permeable stratum are determined by parameter TRP expressing unit permeability through the layer of bottom sediments in the channel. Its interpretation via the mathematical model is the following:

$$TPR = \frac{F}{x^2} \cdot \frac{k'}{m'} \qquad 6)$$

where:

F – active surface or river bed (old river bed) in computation block,

X^2 – area of elementary computation block,

K' – filtration coefficient of weakly permeable sediments,

M' – thickness of weakly permeable stratum.

In adopting a proper digitised network step and number of computational nodes in the simulation model, adjustments should be made to ensure the maximum compatibility between the description of the filtration field and the morphogenesis and measured values of hydraulic heights.

M. Kostecki (2000) performed an example simulation of a water-bearing stratum in a binary layer system for the area of a water intake in the Warta proglacial valley in Krajkowo. Relations with river impact on water filtration system have been considered under several conditions: low water level and bank line level in the river bed (Fig. 9).

Model calibration was effected by the method of successive approximation in line with changes in parameters (effective infiltration, TRP, change in given boundary conditions). The basic assumption was to obtain via numerical calculation a reliable balance (mass) of underground and surface waters, both in natural conditions and those enforced by the exploitation of water intakes.

Another mathematical model, SIMGRO (**Sim**ulation of **Gro**undwater) and surface water levels is distributed physical parameters (nodes of a finite elements network).

In SIMGRO the finite element procedure is applied to approach the flow equation, which describes transient groundwater flow in the saturated zone of the Biebrza valley (*Hydrological System...*, 2002). This model demonstrated that, with groundwater flow in the saturated zone, the direction is mainly

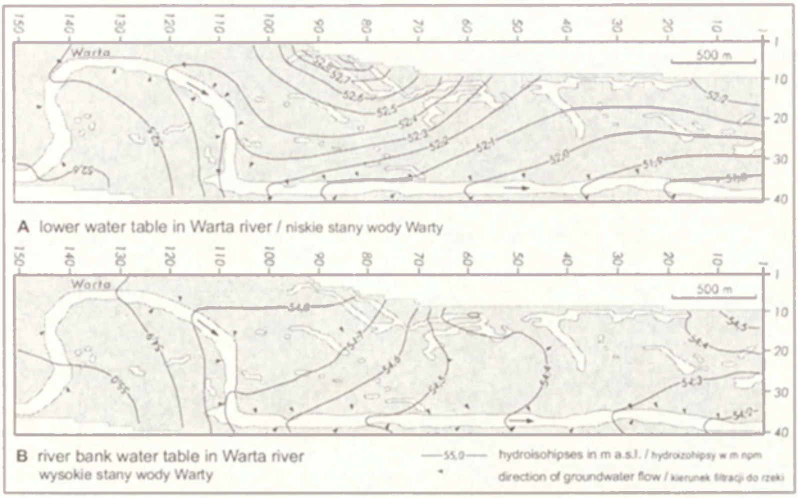

Fig. 9. The hydroisohipses of groundwater flow distribution in the Warsaw-Berlin proglacial valley (Warta river valley) near water supply in Krajkowo (Kostecki 2000)

Rozklad hydroizohips wod podziemnych w pradolinie warszawsko-berlinskiej (dolina Warty). ujęcie wod podziemnych w Debinie (Kostecki 2000)

perpendicular to the Biebrza valley, towards to the river. The mathematical model allows for simulation of effective infiltration (drainage) in an area comprising a narrow belt of the valley along the river. It is made mostly of alluvial soils, as part of the proglacial valley. The SIMGRO model calculates the effective recharge or drainage over the short or long terms.

5. CHANGES IN PROGLACIAL-VALLEY STORAGE AS A RESULT OF HUMAN IMPACT

Changes in the moisture levels of medium alluvial clay during the vegetation period, being under the influence of forced equalization due to the water dam in Brzeg Dolny (Fig. 10) demonstrate clear relations between the level of the Oder and the content of water in different layers of sediments in the valley down as far as 100 cm (Plywaczyk et al. 2001).

An alternate rhythm of fluctuations in water level and moisture content in alluvial clays as seen in the Oder below the damming sees the rhythm more subject to variability (Fig. 10), as dependent on the natural conditions of water circulation in the valley.

Forecasting the changes in retention by the proglacial channel under the influence of taming schemes – and not only at times of low water but also those of high water in the Oder – will require a better recognition of morphogenesis in both the horizontal and perpendicular configurations.

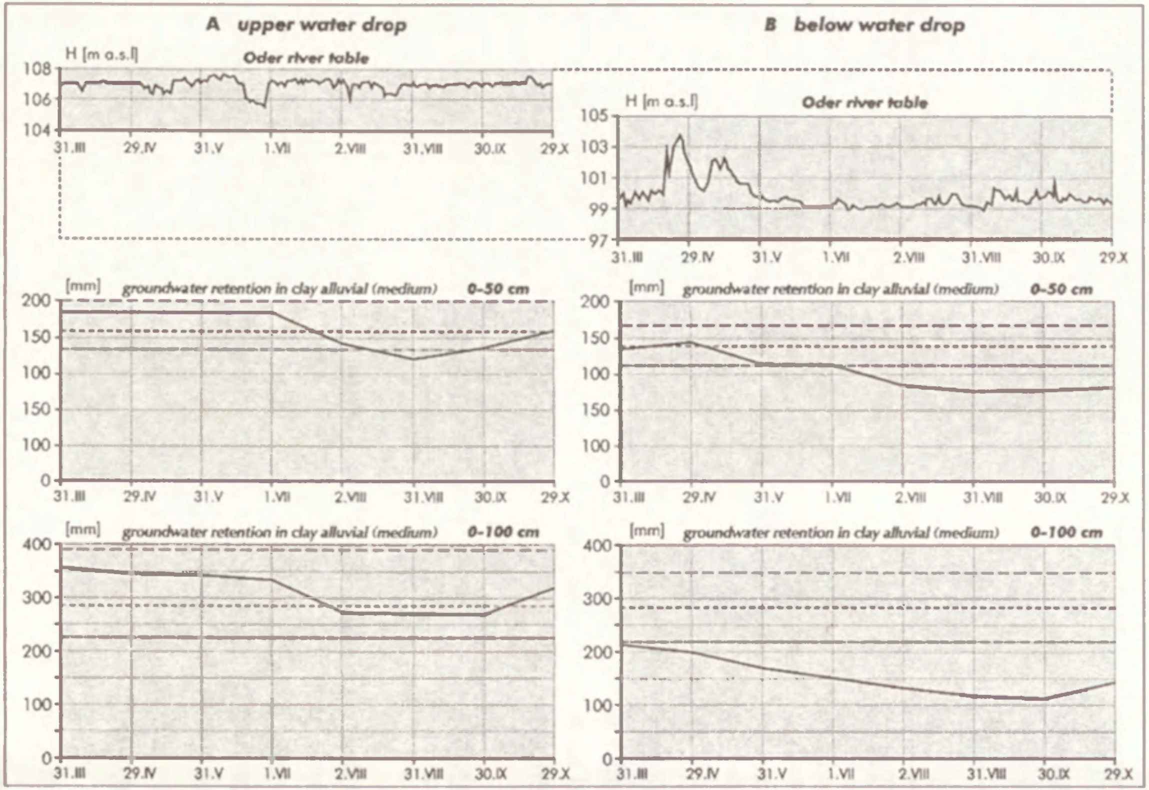

Fig. 10. Change in Oder river water table in Brzeg Dolny and state of groundwater retention in two layers (0–50 cm and 100 cm) (after Pływaczyk *et al.*, 2001)

Przebieg zapasów retencyjnych wód w aluwiach (warstwy gleby – mada średnia na głębokości 0,0 ÷ 0,5 m i na głębokości 0 ÷ 100 cm) w zależności od stanów wody Odry (pradolina wrocławsko-magdeburska) w Brzegu Dolnym (wg Pływaczyka i innych, 2001) A – powyżej stopnia piętrzącego B – poniżej stopnia piętrzącego

6. CONCLUSIONS AND REMARKS

The above examples of polygenetic forms and structures in typical and proglacial valleys do not exhaust the enormous differentiation of bottom valley forms to be found, close to the channel and terrace zones of Polish rivers. The morphogenetic variability of the valleys and channels is the cause of difference in types of transformation of rising water waves, determining the size, kind and losses where the alimentation of alluvian waters is concerned.

The hydrodynamic relationship between a river and waters in the valley is also a derivative of morphogenesis. The identification and specification of the lithological and structural features of valleys and proglacial channels and the sections thereof should be oriented not only towards genesis but above all to the recognition of the complexity of vertical and horizontal structures. It is also important in establishing types of horizontal alluvial sediments resulting from river accumulation and erosion of the past. This is a key to the understanding of the mechanism underpinning contemporary process of water exchange and circulation.

Mathematical modelling in the system of interaction between river waters and groundwater requires the identification of morphogenetic structures of the valley in the context of the hydraulic conductivity within permeable insulating layers and directions of water circulation. The proglacial channels and valleys of rivers remaining connected with groundwaters during low water periods and spates are a vast reservoir of renewable reserves of underground water, though weakly defined as an object of dynamics research, both theoretically and practically.

The morphogenesis of valleys and rivers is a basis upon which to obtain a high level of effectiveness in the mathematical modelling of the hydrological processes within such complex systems. Numerical models and stationary hydrological measurements of rivers and alluvial waters serve in the forecasting of changes in inflow and drainage in regard to natural conditions and under the impact of ever growing anthropopressure. Without a detailed identification of valley or channel morphogenesis, simulations of water exchange will be not possible.

Physico-chemical features of the polygenetic sediments in river valleys could also, as E. Falkowska (2002) notes, be a natural isolation barrier protecting alluvial waters and lower groundwaters against river pollution. Isolation barriers in the form of weakly permeable formations (sand-clay or, organic soils), and their vertical and horizontal continuity condition sorption capacity in valleys protecting underground waters against pollution. The morphogenesis of a valley should thus be a basic element referred to as the analyst seeks to understand a river's water exchange, alimentation dynamics and drainage.

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MORFOGENEZA DOLIN A UWARUNKOWANIE ODPŁYWU RZEK

Streszczenie

Współczesne drogi krążenia wód aluwialnych i korytowych zostały przeanalizowane w relacji z morfogenezą i litologią dolin oraz z dynamiką stanów wody i odpływem rzek. Szczególną uwagę zwrócono na fazy rozwojowe pradolin i dolin, zwłaszcza na ich poligenezę późnoplcjstoceńską i holoceńską.

Złożony układ dolin, z wyraźnie zarysowanymi tarasami, sprzyja skomplikowanym mechanizmom krążenia wód w dolinie często obwałowanej. Wymusza to związek hydrodynamiczny pomiędzy napełnieniem koryta rzeki a warstwami wodonośnymi w aluwiach i w podłożu doliny.

Określono podstawy teoretyczne opisujące dwuwymiarowy ruch wody w dolinie aluwialnej w strefie wezbrania i niżówek rzeki.

Rola dolin i pradolin została zaprezentowana pod kątem kształtowania fal wezbraniowych, anomalii odpływu i spłaszczania fal powodziowych w nawiązaniu do filtracji i podparcia wód aluwialnych i rzecznych.

Przytoczono kilka przykładów relacji pomiędzy wodami podziemnymi a rzeką w strukturach wodonośnych różnych regionów (nizinnych, pradolinnych, wyżynnych i górskich).

Na podstawie polskich doświadczeń przedstawiono opis matematyczny związku hydraulicznego wód korytowych i aluwialnych w pradolinie Warty, podlegającego silnym wymuszeniom antropogenicznym (ujęcia wód podziemnych dla Poznania).

Stwierdzono, że efektywność takich modeli symulacyjnych, stanowiących o dokładności prognozy zmian, nie jest możliwa bez wnikliwego rozpoznania genezy i litologii dolin, rzutujących na ocenę parametrów fizycznych i hydraulicznych. Opis interakcyjny dynamiki rzeki i warstw wodonośnych podczas wezbrania i niżówki jest pochodną morfogenezy doliny.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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Henryk Maruszczak

LATE GLACIAL AND HOLOCENE STAGES OF RELIEF DEVELOPMENT IN LOESS UPLANDS IN SE POLAND

1. INTRODUCTION

A considerable part of the South-Polish Uplands is covered with loess, the largest areas of which occur in the Lublin region, between the middle Vistula and Bug rivers (SE Poland). In this region most of loess covers consist of the Neopleistocene beds (= younger loesses), i.e. eolian carbonate deposits. They are characterised by specific features ("loess relief"), where their thickness exceeds 3–5 m. These relief forms are composed of two generations: a older landforms created during loess silt accumulation, and b/ younger landforms developed when that accumulation was terminated (Maruszczak 1961, 1985).

Younger loesses (= Vistulian loesses = Weichselian loesses) cover various elements of basement relief, from planation surfaces in interfluve areas to terraces in river valleys. The older generation of loess relief is represented by: 1/ accumulation scarps, which are best visible where loess thickness exceeds 5–10 m; 2/ enclosed cup-depressions ("wymoki" in the Polish literature) on planation surfaces of interfluve areas and terraces; 3/ trough-like denudation valleys of dell type developed on slopes. The following forms belong to the younger generation: 4/ dry erosion-denudation valleys; 5/ gullies; 6/ suffosion kettles (forms of underground erosion) connected with gullies – Fig. 1.

Landscape peculiarities of the discussed loess uplands depend mainly on the younger forms developed due to erosion and denudation in the Late Glacial and Holocene.

2. STRUCTURE OF THE NEOPLEISTOCENE LOESS COVER

In poorly diversified upland areas with relative relief not exceeding 30 m, the thickness of the Neopleistocene loesses and loess-like deposits is rather small. These thin covers of silt and silt-sand eolian deposits are undistinguishable in relief (*vide* Maruszczak 1991a). Specific loess relief is only charac-

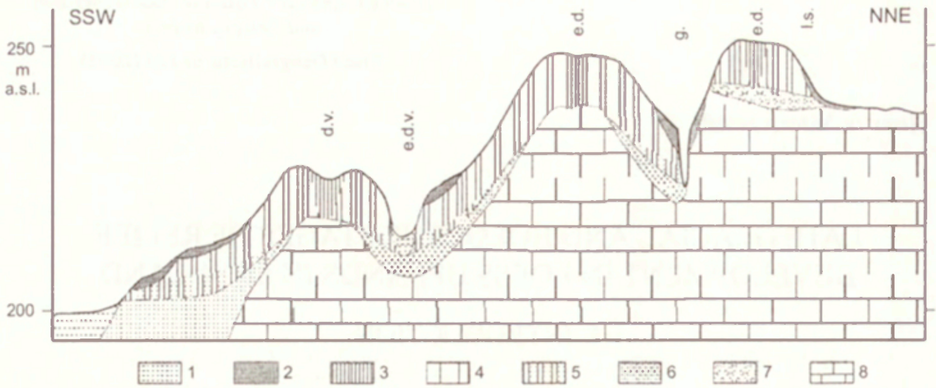

Fig. 1. Relief of the Neopleistocene loess cover and its relation to geologic basement in the westerrn part of the Lublin Upland (Maruszczak 1985)
1 – Holocene alluvia in valley bottoms; 2 – slope/deluvial deposits (Late Glacial and Holocene); 3 – Neopleistocene loesses decalcified in the Holocene; 4 – loesses and loesses with sandy interrlayers (last glacial); 5 – deluvial and solifluction deposits (last glacial); 6 – sandy fluvial deposits (last glacial); 7 – tills and fluvioglacial deposits (Mesopleistocene); 8 – Palaeocene and Upper Cretaceous carbonate rocks; e.d. – enclosed cup-depressions ("wymoki"); d.v. – dell type valleys; e.d.v. – erosion-denudation dry valleys; g. – gullies; l.s. – scarps of loess cover

Ukształtowanie pokrywy lessów neoplejstoceńskich oraz ich stosunek do podłoża geologicznego w zachodniej części Wyżyny Lubelskiej (Maruszczak 1985)
1 – aluwia holoceńskie na dnach dolin; 2 – utwory zboczowe/deluwialne (późny glacjał i holocen); 3 – lessy neoplejstoceńskie odwapnione w holocenie; 4 – lessy i lessy z przewarstwieniami piaszczystymi (ostatnie zlodowacenie); 5 – utwory deluwialne i soliflukcyjne (ostatnie zlodowacenie); 6 – piaszczyste osady rzeczne (ostatnie zlodowacenie); 7 – gliny zwałowe i osady fluwioglacjalne (mezoplejstocen); 8 – węglanowe skały paleoceńskie i górnokredowe; e.d. – zagłębienia bezodpływowe typu wymoków; d.v. – doliny nieckowate typu "dell"; e.d.v. – suche doliny erozyjno-denudacyjne; g. – wąwozy; l.s. – krawędzie pokryw lessowych

teristic of proper loesses over 3-5 m thick. The Neopleistocene loesses with such a thickness occur in the described uplands as separate patches ("islands" or flat ridges = "gredas"), where relative relief exceeds 30 m – Figs. 2 and 4. Distribution of these "islands" and "gredas" enables among other things analysis of the dynamic conditions of loess silt accumulation during the last glacial (Maruszczak 1985).

The thickness of the Neopleistocene proper loesses ranges from 5 to 25 m in the discussed area; the average thickness is about 10 m. Their older beds were accumulated in the lower and middle Vistulian, in the time interval of 80/75 – 28 ka BP; their deposition rate was rather slow, i.e. about 0.1 mm/year. The upper part of loess covers with average thickness of about 5 m was accumulated in the time interval of 28 – 15/12 ka BP, i.e. in the upper Vistulian (these beds are distinguished as upper younger loess LMg – *vide* Fig. 5). The maximum intensity of silt deposition (0.8-1.0 mm/year) occurred in the Last

Fig. 2. Relief of the westerrn part of the Lublin Upland in inclined light from NW (Computer analysis of the topographic map made by dr L. Gawrysiak from Institute of Earth Sciences, Maria Curie-Sklodowska University, Lublin)

Ukształtowanie terenu zachodniej części Wyżyny Lubelskiej przy oświetleniu skośnym od strony NW (komputerowa analiza mapy topograficznej według dra L. Gawrysiaka, Instytut Nauk o Ziemi UMCS, Lublin)

Glacial Maximum, and loess beds accumulated at that time are most typical of periglacial proper loess (Maruszczak 1987).

The following features of our loesses are most important from the geomorphological point of view: a/ distinct domination of grain size 0.05–0.005 mm; b/ porosity over 40%; c/ mean carbonate content about 10%. These features, also very important from the point of view of engineering geology (Grabowska-Olszewska 1985), make loess little resistant to mechanical and chemical action of water, and unstable under load ("collapsing loess"). Thus, they influenced the development of loess relief in the Late Glacial and Holocene.

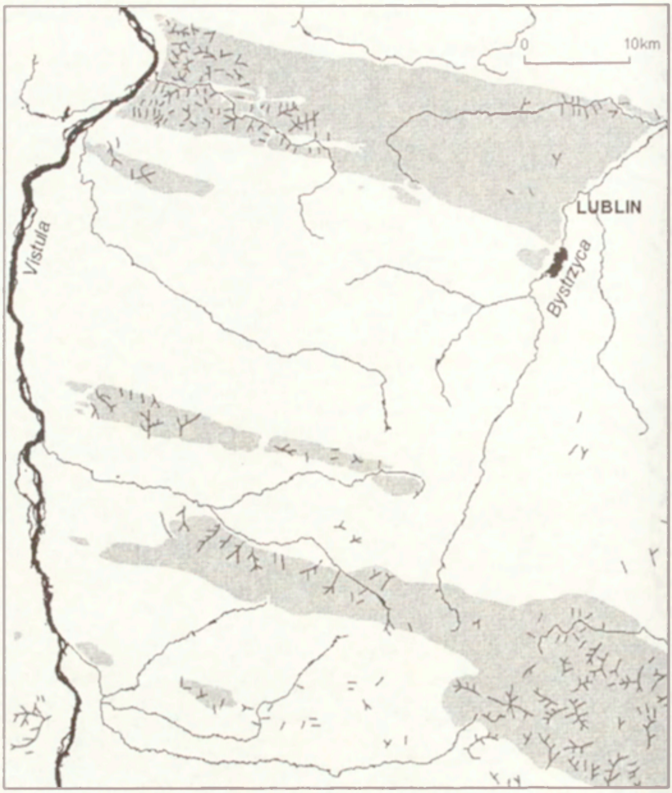

Fig. 3. The extent of the Neopleistocene loess cover over 3–5 m thick (Maruszczak 1961), with the marked net of the Holocene gullies (after *General Geomorphologic Map of Poland* 1:500 000, 1980) in the western part of the Lublin Upland

Zasięg pokrywy lessów neoplejstoceńskich o miąższości ponad 3–5 m (Maruszczak 1961), z oznaczoną siecią wąwozów neoholoceńskich (*Przeglądowa mapa geomorfologiczna Polski* 1:500 000) w zachodniej części Wyżyny Lubelskiej

3. LATE GLACIAL STAGE OF RELIEF DEVELOPMENT IN LOESS UPLANDS

Warming in the Late Vistulian (14/13 – 10 ka BP) caused degradation of periglacial permafrost. Therefore, erosion and thermoerosion strongly influenced relief development of the loess covers during interstadial warmings (Bölling, Allerod); many gullies were formed then. The role of mechanical denudation and solifluction increased during stadial coolings. These processes caused quick transformation of the Late Glacial gullies into erosion-denudation valleys. Many of these valleys cut the whole cover of the Neopleistocene loesses, especially on the terraces in larger river valleys (Fig. 6). In the areas

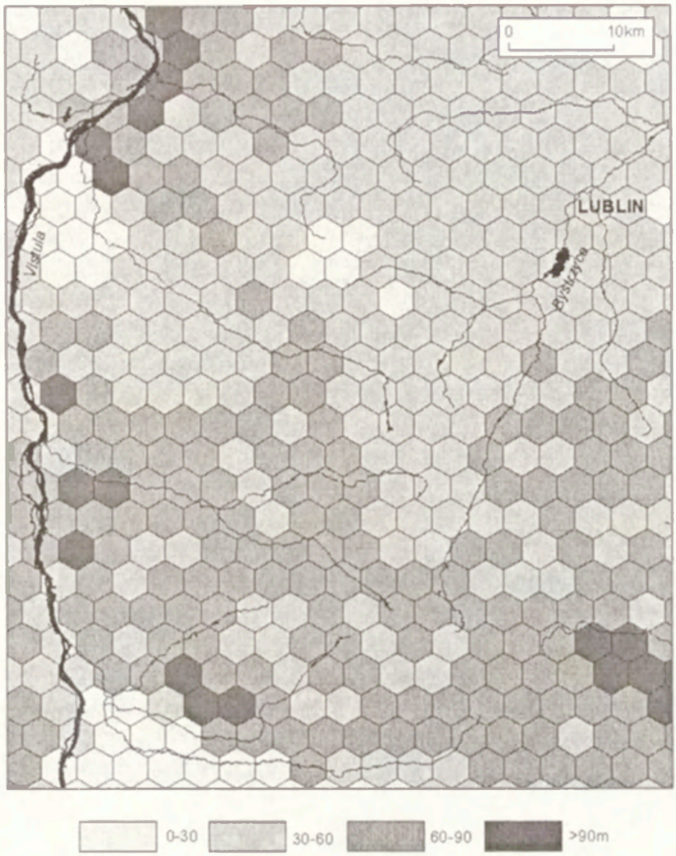

Fig. 4. Relative relief in the western part of the Lublin Upland calculated in hexagonal measurements fields of 10 km² in area (according to the computer analysis of the topographic map as in Fig. 2)

Wysokości względne w zachodniej części Wyżyny Lubelskiej, obliczone w heksagonalnych polach pomiarowych o powierzchni 10 km² (komputerowa analiza mapy topograficznej, jak na rycinie 2)

with relative relief exceeding 60 m, loess deposits were gradually transported downwards in a complex of erosion-denudation valleys, but large fragments of degraded and redeposited loesses remained in places in the Late Glacial dissections of slopes. It is sometimes difficult to distinguish the redeposited material from primary loess beds; it is very useful to study soil catenas occurring on different relief elements in doubtful cases (Fig. 7). It should be stressed that not less than 10–20 % of the Vistulian loess cover was submitted to processes forming erosion-denudation valleys (Fig. 1).

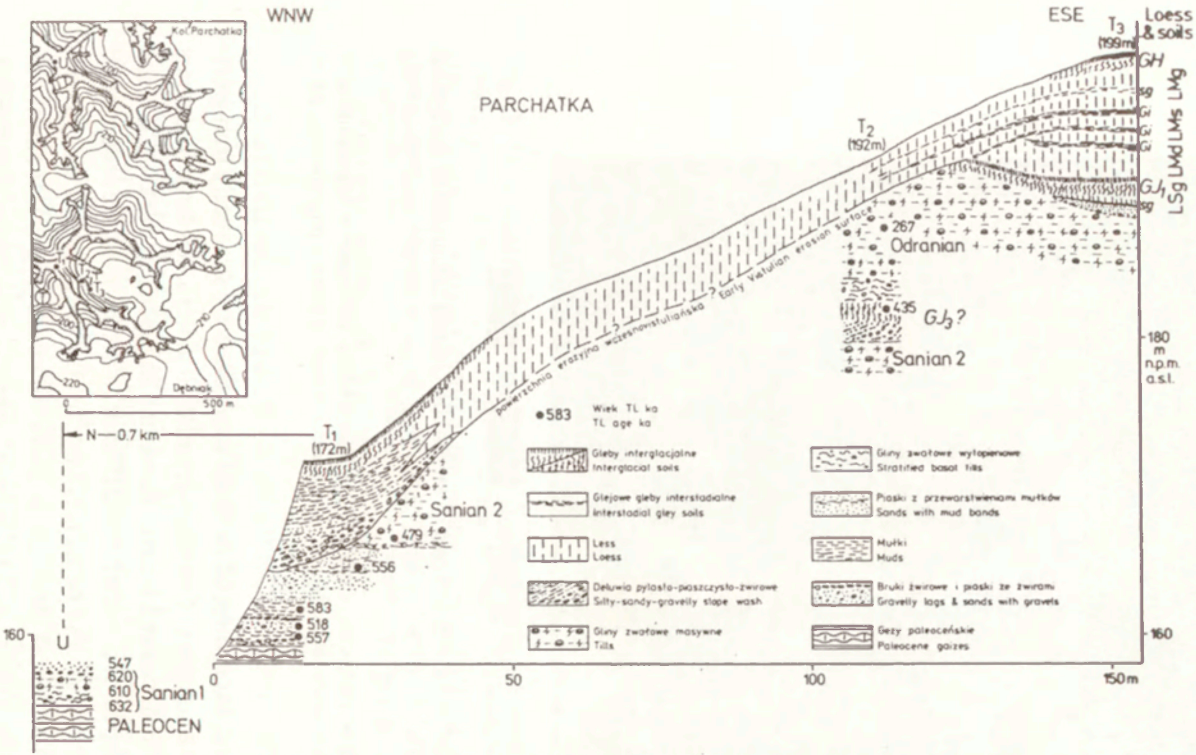

Fig. 5. Geological cross-section of the vicinity of Parchatka village on the Vistula River showing the relation of the Neopleistocene loess cover to the older Quaternary deposits and their basement; worked out by H. Maruszczak in the years 1991–1994 (Pożarski et al. 1994, pp. 27). LM – younger loess (Vistulian), LMd – lower, LMs – middle, LMg – upper, LSg – upper older loess (Wartanian), Gi – interstadial soil, sg – soil sediments, GH – Holocene soil, GJ_I – Eemian soil

Przekrój geologiczny okolic Parchatki nad Wisłą środkową, ilustrujący stosunek lessów neoplejstoceńskich do starszych utworów czwartorzędowych oraz ich podłoża; opracował H. Maruszczak w latach 1991–1994 (Pożarski i in. 1994, s. 27). LM – less młodszy (Vistulian), LMd – dolny, LMs – środkowy, LMg – górny, LSg – less starszy górny (Wartanian), Gi – gleba interstadialna, sg – sedymenty glebowe, GH – gleba holoceńska, GJ_I – gleba z okresu interglacjału eemskiego

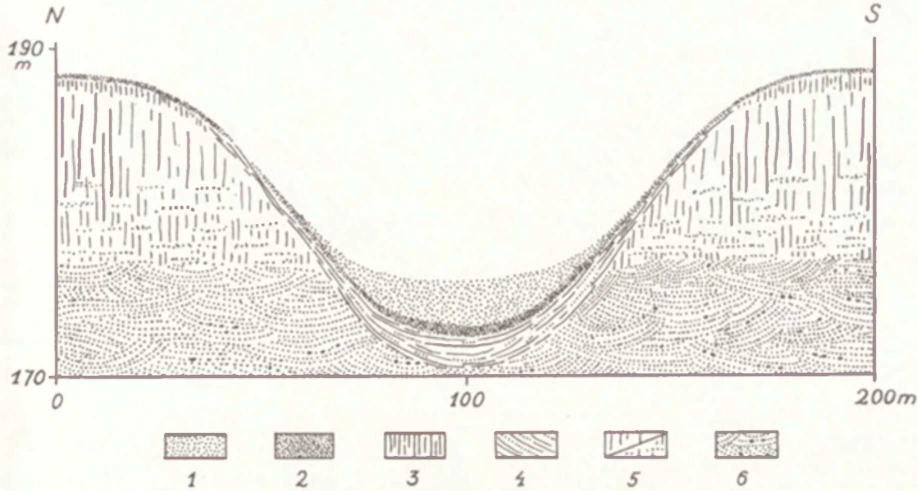

Fig. 6. Geological cross-section of the Late Glacial erosion-denudation valley cutting the Neopleistocene terrace of the Bystrzyca River in Lublin (Maruszczak 1961, pp. 103). 1 – products of recent soil erosion, 2 – humus horizon of Holocene soil, 3 – illuvial horizon of Holocene soil, 4 – Late Glacial slope deposits pedogenetically transformed in the top part, 5 – loesses and loesses with sandy interlayers (last glacial), 6 – fluvial sands and sands with gravels (last glacial)

Przekrój geologiczny późnoglacjalnej doliny erozyjno-denudacyjnej, rozcinającej terasę nadzalewową rzeki Bystrzyca w Lublinie (Maruszczak 1961, s. 103). 1 – produkty erozji gleby (współcześnie), 2 – poziom humusowy gleby holoceńskiej, 3 – poziom iluwialny gleby holoceńskiej, 4 – późnoglacjalne utwory zboczowe w górnej części przekształcowne pedogenetycznie, 5 – lessy i lessy z przewarstwieniami piaszczystymi (ostatnie zlodowacenie), 6 – fluwialne piaski i piaski ze żwirami (ostatnie zlodowacenie)

In the period of 3–4 thousand years a dense net of dry erosion-denudation valleys was formed in the discussed loess uplands. Outside the Neopleistocene loess extent such valleys are considerably less developed as is revealed from comparing Figure 2 and 3. Topographic features are much more differentiated within the extent of loess covers than beyond them. This difference resulted from the fact that loamy, loamy-sandy and sandy-silty covers adjacent to the Neopleistocene loesses are more resistant to erosion and denudation, and moreover they are less thick.

4. HOLOCENE STAGE OF RELIEF DEVELOPMENT IN LOESS UPLANDS

In natural conditions the loess relief was stabilized in the Holocene when the discussed uplands were within the temperate forest zone. It is accounted for by the widespread occurrence of forest soils which developed quickly.

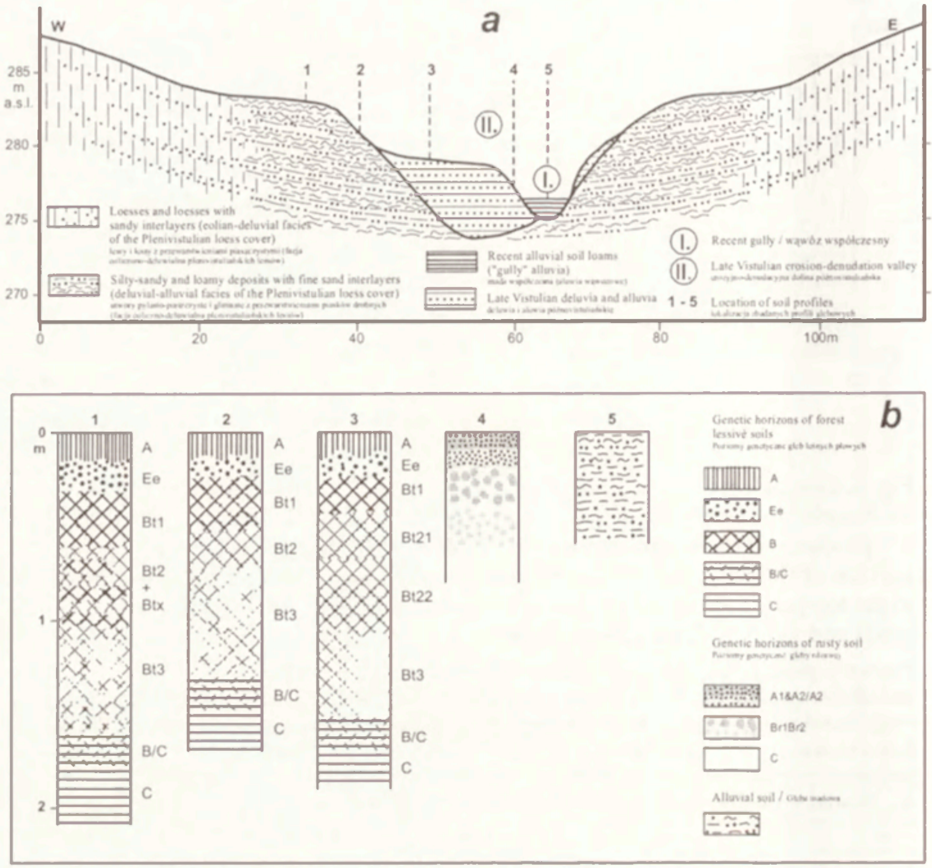

Fig. 7. Geological cross-section of the upper part of the erosion-denudation valley cutting the Neopleistocene loesses in Guciów (20 km to SW of Zamość); (Maruszczak 1998). a – relationships between the Neoholocene gullies, the Late Glacial erosion-denudation valleys and the Plenivistulian loess cover; b – soil profiles recording age differentiation of the presented relief elements

Przekrój geologiczny górnej części doliny erozyjno-denudacyjnej Gucjów (20 km na SW od Zamościa); (Maruszczak 1998). a – relacje wzajemne wąwozów neoholoceńskich i późnoglacjalnych dolin erozyjno-denudacyjnych do pokrywy lessów plevistulianskich; b – profile glebowe dokumentujące zróżnicowanie wiekowe prezentowanych elementów rzeźby

They reached maturity and considerable thickness (1.0-1.5 m) during the first several thousand years. Lessivé soils represented the zonal soil type on loess in the Holocene optimum.

In such rather stable conditions, outside river valleys the transformation of loess covers resulted mainly from chemical denudation, i.e. leaching of carbonates from the Neopleistocene loess beds. Laboratory analyses indicate that carbonates occurring in our loesses are usually very easily dissoluble (Dwucet 2001); therefore, the indices of solute yield are considerably higher in river catchments with loess cover (Janiec 1999). Leaching of carbonates and the resulting "loess collapse" were the strongest in depressions of the topographic surface. In this way the enclosed cup-depressions of "wymok"[1] type and trough-like denudation valleys of dell type were deepened (Fig. 1).

Stronger changes occurred only when human economic activities started. The development of agriculture in the Neolithic was most important. The processes of mechanical denudation and gully erosion, which were so typical of the Late Glacial, developed when plough fields replaced mixed forests. However, these processes were strong only locally, in the regions where settlement developed more intensively. Gullies were then formed in loess landscape (Kruk et al. 1996). The Neolithic settlement was not stable; soil cultivation was usually abandoned in the areas degraded by erosion, so gullies stopped to develop. Such conditions lasted in the discussed uplands till historical times. A thousand years ago the average population density was 1–2 persons/km^2 in the Lublin Upland, and 5–6 persons/km^2 in the loess uplands near Cracow (Maruszczak 1991b).

Population density in the Lublin Upland increased to 10 persons/km^2 only on the turn of the 15^{th} century (in Cracow vicinity this index was 2–3 times higher). The boundaries between arable land and forests were stabilized only at such population density, and farming adapted to cereal cultivation became more important. Thus, soil erosion and gully erosion developed more intensively (Maruszczak 1994, 1997). Large and branched systems of gullies were formed in loess uplands with relative relief exceeding 60 m (*vide* Figs. 3 and 4). Gullies, which were episodically formed in less diversified areas could be improved and included again into arable land.

Therefore, recent gullies, which are so important in the landscape of a considerable part of our loess areas, were mainly developing during the last 500 years. In the areas with relative relief exceeding 60 m, even over 20% of the Neopleistocene loess cover can be within the reach of gully erosion (Fig. 5, hypsometric sketch in the upper left corner). However, it should be stressed that strongly developed gullies cut the bottoms of the Late Glacial erosion-denudation valleys (Fig. 7), so they are subordinate forms. The intensity of soil erosion (tillage erosion) also increased during the last 500 years. This process resulted not only in some relief changes of upland slopes but also in levelling the interfluve and terrace plains; small, enclosed cup-depressions became less visible (Maruszczak 1985, pp. 30–31).

[1] Enclosed cup-depressions collect more rainwater. In consequence the crops in fields get drenched; that is why these depressions are termed by the farmers as "wymoki" (Polish "wymokly" = drenched).

5. FINAL REMARKS

1) Loess uplands in SE Poland are characterized by the occurrence of a dense net of dry erosion-denudation valleys incised into the Neopleistocene eolian loess cover. These valleys developed during periglacial permafrost degradation in the Late Glacial (14/13–10 ka BP); they usually occur in areas with relative relief exceeding 30 m. They were formed by intensive thermoerosion process (periglacial gully erosion), and then solifluction rapidly transforming the slopes of gullies. This "excessive formation of valleys" (*sensu* Büdel 1977, *vide* pp. 82 – "fossilen excessiven Talbildung") was typical of the Late Glacial stage of loess relief development. Therefore, it can be assumed that it was a zonal type of relief-forming process. Its effects were weaker both in West Europe because of limited permafrost extent, and in East Europe due to continental climate with low outflow of atmospheric water.

2) In the Holocene the main features of loess relief remained almost unchanged in natural conditions typical of mixed forests in temperate climatic zone. The depth of enclosed cup-depressions ("wymoki") and trough-like denudation valleys (dells) increased only in places as a result of carbonate leaching (chemical denudation); therefore, their role in the loess landscape became more important. Mechanical denudation and gully erosion developed more intensively only in historical times when the area of arable land increased. In the discussed loess uplands of SE Poland the development of gullies was stronger only in the Little Ice Age when population density exceeded 5-10 persons/km^2. The development of the present gullies was mainly initiated during torrential rainfalls occurring once in several tens of years. Larger gullies are stable forms in our loess uplands, as they do not undergo cultivation. These gullies are usually incised in the bottoms of the Late Glacial erosion-denudation valleys. The gully volume is many times smaller than valley volume; such quantitative evaluation indicates that the rank of gully is lower. Therefore, gullies are subordinate forms, and do not represent zonal features of loess relief in the discussed area.

3) Gullies are natural relief forms in temperate semi-desert and dry steppe zones mainly. J. Tricart (1953) was wrong considering them to be the representative forms of the temperate typical steppe zone. In the opinion of L.S. Berg (1949) "stiepnyie bludtsa" are more frequent, and thus more representative features of steppe zone. They correspond to the Polish "wymoki". However, "wymoki" are not very numerous in the loess landscape of SE Poland due to the limited extent of planation surfaces in the interfluve areas, and on terraces in larger river valleys. It can be also added that "wymoki" in

Poland are usually much smaller than similar steppe forms; they quickly become less and less important in the landscape as a result of soil erosion (tillage erosion).

4) The Late Glacial erosion-denudation valleys cut the Neopleistocene loess cover deeply, and often reach to its bottom. They constitute 10–20% of the surface of the discussed loess uplands. Trough-like denudation valleys of dell type and enclosed cup-depressions ("wymoki") occupy a similar area in which the Neopleistocene loess layers are decalcified and clayey. At present the carbonate layers with porosity exceeding 40%, which are liable to mechanical and chemical denudation, make about two thirds of the Neopleistocene loess cover. It should be stressed that most of them are young, i.e. not older than 25–15 thousand years, so they are potentially most vulnerable to surface and underground erosion. The possibilities of surface erosion development are limited because carbonate loess layers are mainly associated with convex elements of topographic surface. Such an opinion is right only with the assumption that land use, especially soil cultivation, is rational both from the natural and social point of view. However, greater changes of the kind and extent of arable land can result in the transformation of the rainwater drainage pattern, and therefore also in the development of gully erosion in these places in which the natural relief does not predispose them to erosion. Unreasonable locations of greater buildings can cause the development of underground erosion, which is defined in the Russian literature as suffosion. In the West-European literature this term in not used. It corresponds to German "subrosion" (sensu J. Büdel 1977, pp. 152), and to English "piping".

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PÓŹNOGLACJALNY I HOLOCEŃSKI ETAP ROZWOJU RZEŹBY WYŻYNY LUBELSKIEJ W POLSCE SE

Streszczenie

W pokrywie lessowej w Polsce SE dominują warstwy neoplejstoceńskie (lessy młodze, LM na rycinie 5). Na Wyżynie Lubelskiej warstwy LM, o miąższości 5–25 m (przeciętnie 10 m), występują wyspowo, w płatach wyraźnie wyodrębniających się (ryc. 1–3). Cechy odrębności "rzeźby lessowej" zawdzięczamy przede wszystkim gęstej sieci młodych suchych dolin erozyjno-denudacyjnych (Maruszczak 1961, 1985) oraz wąwozów; poza zasięgiem lessów formy te nie odgrywają znaczniejszej roli (ryc. 3 i 4).

Późnoglacjalne doliny erozyjno-denudacyjne najsilniej rozwijały się tam, gdzie lessy występują w terenach o wysokościach względnych ponad 30 m (ryc. 3–5). Doliny te powstały w późnym glacjale (14/13–10 ka BP), tzn., w okresie degradacji zmarzliny wieloletniej, w wyniku intensywnej termoerozji (wąwozowej erozji peryglacjalnej) oraz soliflukcji szybko przekształcającej skarpy wąwozowe (ryc. 1 i 6). Dla późnoglacjalnego etapu rozwoju rzeźby lessowej charakterystyczne było więc "ekscesywne rzeźbienie dolin" (Budel 1977, s. 82 – "fossilen excessive Talbildung"). Był to więc strefowy typ rzeźbienia pokrywy lessowej. Jego znaczenie w Europie zachodniej było mniejsze ze względu na bardziej ograniczony zasięg zmarzliny wieloletniej, a w Europie wschodniej ze względu na kontynentalny klimat z niskimi wskaźnikami odpływu wód atmosferycznych.

W holocenie, poczynając od borealnej fazy fito-klimatycznej, zasadnicze rysy rzeźby lessowej pozostawały bez istotnych zmian. Tylko miejscami, w wyniku ługowania węglanów (denudacji chemicznej lessów), wzrosła głębokość form typu wymoków oraz dolin nieckowatych typu dell (ryc. 1). Dopiero w czasach historycznych, ze wzrostem arealu śródleśnych pól uprawnych, intensywniej rozwinęła się denudacja mechaniczna i erozja wąwozowa. Na wyżynach lessowych w Polsce SE, współczesne wąwozy pojawiły się na większą skalę dopiero w XV–XIX w., tzn. w małej epoce lodowej, gdy gęstość zaludnienia wzrosła powyżej 5–10 mieszk./km^2. Na Wyżynie Lubelskiej i Roztoczu wąwozy lessowe były formami "trwałymi", gdyż wyłączano je z rolniczego użytkowania ziemi. Tych wąwozów, predysponowanych działalnością gospodarczą, nie można więc traktować jako reprezentatywnych, strefowych form rzeźby, tzn. tak jak niesłusznie to sugerował J. Tricart (1953) w odniesieniu do wschodnioeuropejskich stepów umiarkowanych. Należy jeszcze podkreślić, że są to formy "włożone" w dna dolin erozyjno-denudacyjnych (ryc. 1), a ich struktura jest o jeden rząd wielkości mniejsza (ryc. 7). Przy takiej ilościowej, geomorfologicznej ocenie nie można przeceniać roli wąwozów w krajobrazie lessowym.

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THE PROBLEM OF OUTFLOW FROM THE ICE-DAMMED LAKE IN THE WARSAW BASIN DURING THE LESZNO PHASE

1. INTRODUCTION

From among issues in the field of glacial geomorphology, i.e. Quaternary palaeogeography, there are still many that can be seen as open, or the subject of debate, even though they are presented in the literature as canons. For example, almost every Polish school textbook has it that Poland's classic ice-marginal valleys are that of the present Noteć and Warta running between Toruń and Eberswald, as well as the Warsaw-Berlin example. At the same time, however, it is not possible that the latter existed during the Vistulian Glaciation along the whole length from the Warsaw Basin, since a proglacial lake was created in the latter area at that time.

The Warsaw ice-dammed lake and its emergence, age and relationship with the Warsaw-Berlin ice-marginal valley were subjects for many German and Polish workers over a period of 120 years. Some, like G. Berendt (1879), F. Wahnschaffe (1881), K. Keilhack (1898), J. Siemiradzki (1909), J. Samsonowicz (1922) and R. Galon (1972) were supporters of the idea that the Vistula waters flowed west from the Warsaw Basin via the Warsaw-Berlin ice-marginal valley, while others, like L. Henkel (1909), S. Lencewicz (1922), J. Mikołajski (1927) and S. Jewtuchowicz (1967) excluded such a possibility, in spite of what was at that time limited understanding of the geomorphological relations in the Vistula valley downstream and in the vicinity of the Warsaw Basin.

The matter of the relationship between the Warsaw ice-dammed lake and Warsaw-Berlin ice-marginal valley was addressed in 1999 by L. Marks, who judged that: "*While the appearance of the Warsaw Basin in its present shape is indeed linked only with the end of the Warta Stage (Glaciation), a decisive role in its final formation has to have been played by the flow of water via the Warsaw-Berlin ice-marginal valley, and later the Toruń-Eberswald ice-mar-*

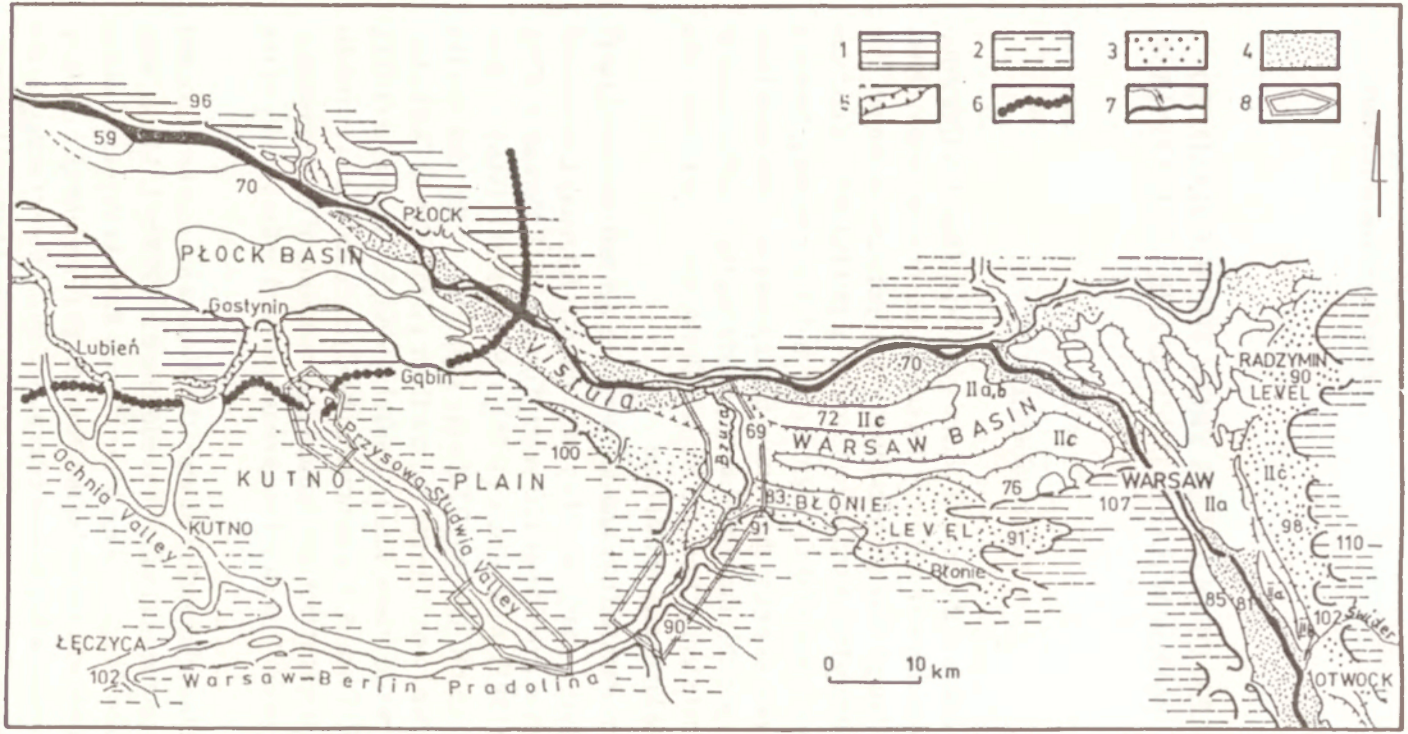


Fig. 1. Geomorphological sketch of the Vistula Valley between the Warsaw and Płock Basins. 1 – morainic elevation from the period of the Vistulian Glaciation; 2 - morainic elevation from the period of the Middle European Glaciation; 3 – Radzymin-Błonie Level; 4 – flood terrace; 5 – troughs; 6 – extent of the Vistulian Ice Sheet; 7 – rivers; 8 – location of areas of detailed geomorphological study referred to in the text

Szkic geomorfologiczny doliny Wisły między Kotliną Warszawską a Kotliną Płocką. 1 – wysoczyzna morenowa z okresu zlodowacenia vistulian; 2 – wysoczyzna morenowa z okresu zlodowacenia środkowopolskiego; 3 – poziom radzymińsko-błoński; 4 – terasa zalewowa; 5 – rynny; 6 – zasięg lądolodu vistulianskiego; 7 – rzeki; 8 – lokalizacja obszarów szczegółowych badań geomorfologicznych omawianych w tekście *ginal valley at the end of the Vistulian Glaciation"*, as well as that... *"The most problematical is the interrelationship between the ice-dammed lake formed in the Warsaw Basin and the ice-marginal valley system that represented an important fragment of E-W drainage on the Central European Lowland"*.

The author of the present article has long associated himself with issues related to the geomorphological development of the Vistula valley between the Warsaw and Toruń Basins, as well as having pursued geomorphological and geological research into the ice-marginal valleys cutting across the Kutno Plain (Fig. 1). These valleys, initiated at the time the ice sheet came to a standstill during the Leszno Phase and leading into the "Warsaw-Berlin ice-marginal valley" are the key to resolving the matter of which direction waters from the ice-dammed lake then existing in the Warsaw Basin flowed – and hence also the matter of the relationship between the lake and the aforementioned channel. Moreover, a valuable augmentation of our knowledge in this regard has come with the results of geomorphological and geological studies carried out in the valley of the Lower Bzura by L. Andrzejewski (1991, 1994).

In 1994, an article co-written by the author and L. Andrzejewski used the results of research they had carried out to question whether the outflow of water via the "Warsaw-Berlin ice-marginal valley" during the Vistulian could have begun from the Warsaw Basin, which must after all have been filled to a certain height by the ice-dammed lake. The question then arising – which at that time was hard to answer – concerned whether the lake could have existed in the more than 2000 years between the maximal extent of the Vistulian Ice Sheet (dated to about 20,000 years BP) and the Kujawy sub-phase dating back to around 17,2000 BP, without a relief channel taking away the excess of water, in conditions in which the Vistula and its tributaries were constantly feeding the lake from the south, while meltwaters were flowing in from the north-east via the Biebrza-Narew ice-marginal valley.

It therefore seemed worthwhile to come back to this question and attempt to answer it in the light of observations commenced with, as well as experiences gained by the present author and other researchers in areas that are still experiencing glaciation.

However, the adoption of any point of view as to the matter in question requires a brief presentation of the current state of knowledge on relief in the Warsaw Basin, as well as on the relationship between the ice-marginal valleys cutting across the Kutno Plain, as well as the Warsaw-Berlin ice-marginal valley.

2. VIEWS REGARDING THE GENESIS AND AGE OF THE LEVELS AND TERRACES OF THE WARSAW BASIN

In the extensive depression that is the Warsaw Basin, S.Z. Różycki (1967) was able to distinguish the following levels and terraces (from top to bottom): the Radzymin-Błonie Level, the Otwock Terrace (IIc), the Falenica Terrace (IIb) and the Praga Terrace (IIa). The Radzymin Level is located on the right side of the Basin, while its equivalent on the left side is the Błonie Level (Fig. 1). The surfaces of these levels have clays. According to M.D. Baraniecka and K. Konecka-Betley (1987), the Radzymin Level near Otwock reaches an elevation of 103 m a.s.l. and descends as one moves northwards to reach an altitude of 88–90 m a.s.l. in the vicinity of Radzymin and Marki. On the left hand side of the Basin, the Błonie Level near the mouth of the Bzura is at an altitude of 81–82 m a.s.l. It is present further on in the gorge section of the Vistula Valley between the Warsaw and Płock Basins at an altitude of 82–83 m a.s.l. (21–22 m above the level of the Vistula).

In the southern part of the basin, the Otwock Terrace (IIc) is at altitudes of between 100 and 103 m a.s.l. (16–19 m above the level of the Vistula), while at the confluence of the Bzura the altitudes are 71–72 m a.s.l. (7–8 m above the Vistula), and near Płock 63–64 m a.s.l.

The Falenica Terrace (IIb) attains heights of 96–97 m a.s.l. near Otwock, before descending into a depression between islands of the Otwock Terrace. Its elevation near the mouth of the Bzura is of 69–72 m a.s.l., while in the eastern part of the Płock Basin it lies at 60 m a.s.l.

The Praga Terrace (IIa), lying above the floodplain, is at altitudes of 90–92 m a.s.l. in the southern part of the Basin (i.e. is 7–9 m above the level of the Vistula), while it is at the same level as the Falenica Terrace close to the confluence point of the Bzura.

According to S.Z. Różycki (1967), the stratified clays covering the Radzymin-Błonie Level were laid down in the course of the Wkra Stage of the Oder Glaciation. The Otwock Level (IIc) was in turn created through the blocking of the course of the River Vistula to the north during the Leszno Phase. The dumping of sand and gravel sediments in the Warsaw Basin was such that the height of the Radzymin-Błonie Level was in parts reached. A period of erosion then followed, prior to a renewed episode of accumulation during the Poznań Phase. It was at this time that the lower Falenica Terrace arose. In the opinion of S.Z. Różycki, it was during the Pomeranian Phase that the Praga Terrace was created.

Later studies have shown that – beneath the stratified clays at the surface of the Radzymin-Blonie Level – there are sediments with organic fragments dating back to the Eemian Interglacial (Janczyk-Kopikowa 1975; Karaszewski, 1975; Sarnacka 1982). For this reason, the sedimentation of the aforementioned clays needs to be linked with the Vistulian Glaciation.

M.D. Baraniecka and K. Konecka-Betley (1987), basing on the results of thermoluminescence dating, were convinced that the above clays were 51,000 and 53,000 years old. This related them to the transgression of the Central Vistulian Ice Sheet, whose extension this far south had not been documented. In the view of Baraniecka and Konecka-Betley, the Upper Vistulian's Leszno Phase was associated with accumulation in the Warsaw Basin – initially of riverine and glaciofluvial sediments, and ultimately also of loams and clays. It is these sediments that form the Basin's highest terrace, which is to say the Otwock Terrace (IIc). It was in the Bölling that the sandy series of the Falenica Terrace (IIb) was laid down, and in the Alleröd that there arose the Praga Terrace (IIa).

3. THE RELATIONSHIP BETWEEN THE "WARSAW-BERLIN ICE-MARGINAL VALLEY" AND THE WARSAW ICE-DAMMED LAKE

If it is accepted that the maximal extension of the Vistulian Glaciation brought with it – in the Warsaw Basin situated some 100 km from the edge of the ice sheet – a deposition of stratified clays covering the point of confluence of the Świder and Vistula at an altitude of c. 102 m a.s.l., the consequent assumption would be that the water level of the large ice-dammed lake must have reached beyond this altitude. With such a level of water, a large area of heights adjacent to the Vistula Valley would have been inundated and there could have been an outflow of the excess water from the Warsaw ice-dammed lake to the west, via the "Warsaw-Berlin ice-marginal valley". For the floor of this spillway at the divide near Łęczyca, some 70 km from the Warsaw Basin, also sits at an altitude of 102 m a.s.l. From the Łęczyca area, the "Warsaw-Berlin ice-marginal valley" declines in altitude to both west and east. The western part of the channel is taken – over considerable lengths – by the Warta and Ner, and by the eastern part of the Bzura, which empties into the Vistula within the Warsaw Basin.

The above considerations as to the likelihood of outflow from the Warsaw ice-dammed lake were embarked upon on the basis of the hypsometric relationships currently holding good for the "Warsaw-Berlin ice-marginal valley". This does not mean that its floor in the Łęczyca area was as it is now in the

Leszno or Poznań Phases. Account needs to be taken of the fact that the Warsaw-Berlin channel is crossed by a tectonic structure known as the Kujawy-Pomeranian Anticline. This is often believed to have been subject to uplift, though unambiguous evidence that this is really the case is so far lacking. In the light of the above remarks, there remains a possibility that the floor of the "Warsaw-Berlin ice-marginal valley" on the present-day divide between the Vistula and Oder catchments was at a different altitude at the time the Vistulian Glaciation attained its maximum extent. It may have lain lower, and then been raised by neo-tectonic uplift, or it may rather have been at a higher altitude, as a consequence of glacial isostatic phenomena raising the land beyond the ice sheet due to the latter's pressure on the mineral substratum it covered. In addition, the possibility that the floor of the ice-marginal valley was deformed by halokinetic movements cannot be excluded.

4. THE RELATIONSHIP BETWEEN THE OCHNIA AND PRZYSOWA-SLUDWIA PROGLACIAL VALLEYS AND THE "WARSAW-BERLIN ICE-MARGINAL VALLEY"

The matters of the elevation of the water surface in the Warsaw ice-dammed lake and the possibility of outflow from it via the "Warsaw-Berlin ice-marginal valley" take on a different perspective when account is taken of the geomorphological studies referred to in the introduction, regarding the two proglacial valleys dissecting the Kutno Plain, via which meltwaters flowed into the ice-marginal channel (Fig. 1).

The first of the above, in which the Ochnia now flows, enters what was the ice-marginal valley ca. 50 km west of the Warsaw Basin, while the second – used by the Przysowa and Sludwia – does so ca. 25 km from the Basin. Both valleys are extensions of sub-glacial channels indicative of the maximal extent of the Vistulian Glaciation. As the literature makes clear, the Ochnia and Przysowa-Sludwia valleys have been known for many years (Lencewicz 1922; Kotarbiński, Urbaniak-Biernacka 1975; Baraniecka 1979; Dylik 1984). There is a view that meltwaters headed south via them, though detailed geomorphological studies to confirm this have not so far been made.

The Ochnia Valley, which is 41 km long and 1–2 km wide, begins ca. 3 km west of Lubień Kujawski. In its upper section, the floor is at 120 m a.s.l., while at the confluence point with the "Warsaw-Berlin ice-marginal valley" it is at 95 m a.s.l. The slopes of the valley are long and inclined gently. 11 km along the course of the Ochnia Valley, contact is made with the proglacial valley of the Lublinianka, which is ca. 7 km long and around 0.5 km wide. This is

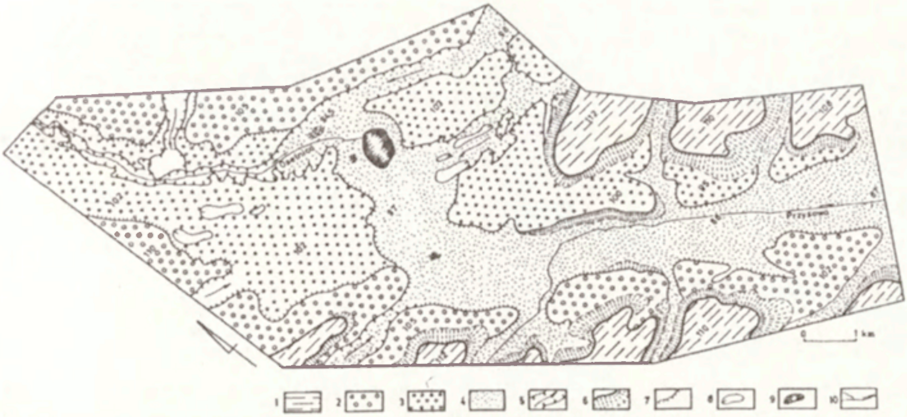


Fig. 2. Geomorphological sketch of the upper section of the Przysowa-Sludwia proglacial channel. 1 – morainic elevation from the period of the Middle Polish Glaciation; 2 – upper outwash level; 3 – lower outwash level; 4 – valley floor; 5 – the Osetnica meltwater channel and valley; 6 – slope; 7 – edge; 8 – cave-in lakes; 9 – lacs; 10 – rivers
Szkic geomorfologiczny górnego odcinka proglacjalnej doliny Przysowy-Sludwi. 1 – wysoczyzna morenowa z okresu zlodowacenia środkowopolskiego; 2 – wyższy poziom sandrowy; 3 – niższy poziom sandrowy; 4 – dno doliny; 5 – rynna Osetnicy i doliny wód roztopowych; 6 – zbocza; 7 – krawędzie; 8 – wytopiska; 9 – jeziora; 10 – rzeki

a continuation of the sub-glacial drainage channel of Lake Lubiński. 2 km north-west of Kutno a second proglacial channel joins that of the Ochnia, i.e. the valley of the Głogowianka, which begins from a sub-glacial drainage channel north-west of Gostynin. This valley is around 13 km long and 1–4 km wide. Along the Ochnia Valley, beyond its floor, there are fragments of a level at a higher elevation that was shaped by meltwaters. The fragments are of differing widths and lengths. In its upper part, this channel is at around 122 m a.s.l., while in the Kutno area it is at 107 m. Below Kutno, this level links up with the floor of a western outbranching of the Ochnia valley that makes contact with the "Warsaw-Berlin ice-marginal valley" at an altitude of c. 98 m a.s.l. The floor of the eastern arm, along which the Ochnia flows, is at 95 m a.s.l. at the point where the river discharges into the Warsaw-Berlin channel. Beyond the valley floor, there is no recognised higher level that would correspond with that of the western arm. Probably any such structure that may have existed was destroyed by the flowing water that later used this part of the valley only.

The second proglacial valley, 35 km long, is that of the Przysowa-Sludwia. It begins at the mouth of the Osetnica sub-glacial channel at an altitude of ca. 97 m a.s.l. and descends steadily to 84 m a.s.l. near the exit into the "War-

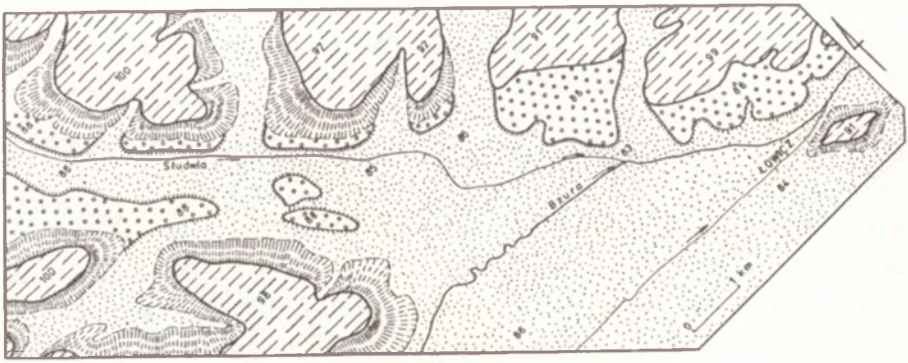


Fig. 3. Geomorphological sketch of the confluence section of the Przysowa-Sludwia proglacial valley and the "Warsaw-Berlin Proglacial Channel". Explications see fig. 2

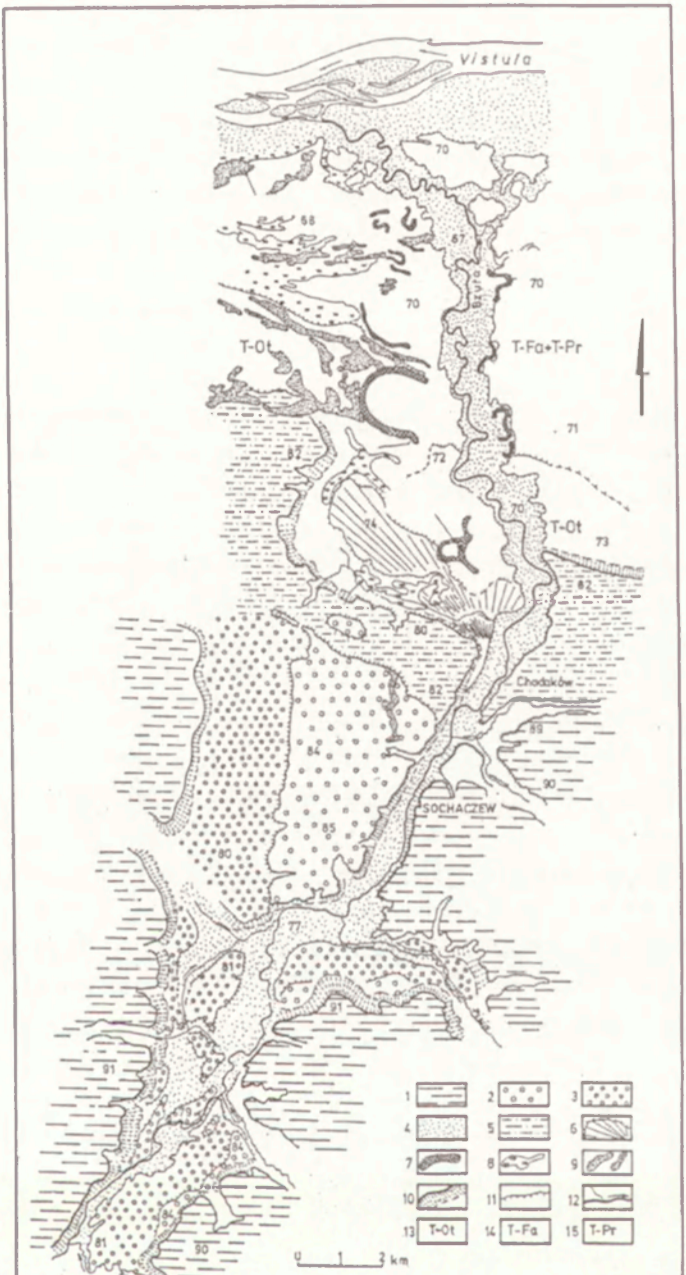
Szkic geomorfologiczny ujściowego odcinka proglacjalnej doliny Przysowy-Słudwi do "pradoliny warszawsko-berlińskiej". Objaśnienia patrz rycina 2

saw-Berlin channel" near Łowicz. It initially has a width of ca. 3.5 km, but by the end this has narrowed to 2 km. From the outset, the Przysowa-Sludwia Valley is characterised – beyond the valley floor area – by two outwash levels formed from glaciofluvial formations (sands and gravels). The first (higher) of these lies initially at an altitude of 105 m a.s.l. on the right hand side of the valley, but disappears after 9 km (Fig. 2). The second (lower) level, on the left hand side of the valley, has an initial altitude of 100 m a.s.l. and runs without interruption to the point where the valley joins the "Warsaw-Berlin Proglacial Channel", at which point it attains 88 m a.s.l. It thus rises 3–5 m above the valley floor here, while attaining a width of around 200 m (Fig. 3).

What was the further fate of the meltwaters that took the aforementioned proglacial valleys in emptying into the "Warsaw-Berlin ice-marginal valley"? This matter has been addressed by geomorphological research carried out in the Bzura Valley by L. Andrzejewski (1991, 1994) (Fig. 4).

Beyond the floor of the Bzura floodplain there are two levels formed from sandy and gravel formations. The first (lower) one is at altitudes of 80–81 m a.s.l. and is the channel floor with a width of some 2 km near the mouth of the Rawka. The second level – at 83-84 m a.s.l. – takes the form of narrow ledges or islands. At a similar altitude between the mouths of the Pisa and Utrata (on the western side of the Bzura), there is an extensive level around

Fig. 4. Geomorphological sketch of the lower Bzura Valley (after L. Andrzejewki, 1991). 1 – morainic elevation from the period of the Middle Polish Glaciation; 2 – upper level of meltwaters; 3 – lower level of meltwaters; 4 – valley floor; 5 – Błonie Level; 6 – alluvial fan of the Bzura; 7 – palaeomeanders; 8 – peat areas; 9 – dunes; 10 – slopes; 11 – edge; 12 – rivers; 13 – the Otwock Terrace; 14 – the Falenica Terrace; 15 – Praga Terrace



Szkic geomorfologiczny doliny dolnej Bzury (wg L. Andrzejewskiego, 1991). 1 – wysoczyzna morenowa z okresu zlodowacenia środkowopolskiego; 2 – wyższy poziom wód roztopowych; 3 – niższy poziom wód roztopowych; 4 – dno doliny; 5 – poziom błoński; 6 – stożek aluwialny Bzury; 7 – paleomeandry; 8 – obszary torfowe; 9 –wydmy; 10 – zbocza; 11 – krawędzie; 12 – rzeki; 13 – terasa otwocka; 14 – terasa falenicka; 15 – terasa praska

6 km long and 3 km wide. This is the equivalent of the aforementioned fragments of the higher level occurring between the confluence points with the Rawka and Pisa. This level is covered by a layer of fine-grained laminated sands ca. 2 m thick, most often overlying clays. It needs to be added that – in many places along the lower Bzura Valley, these clays occur over sandy or gravel formations. Structural studies of the sediments building these levels show them to have been deposited by waters flowing into the Warsaw Basin.

5. THE OUTFLOW OF WATERS FROM THE "WARSAW-BERLIN ICE-MARGINAL VALLEY"

The results of geomorphological research on the proglacial valleys of the Ochnia and Przysowa-Sludwia, as well as the lower section of the Bzura Valley, go a very long way to undermining the view that water from the ice-marginal valley may have flowed from the Warsaw ice-dammed lake to the west, at the time the Vistulian Glaciation attained its maximum. The only way in which such a contention was accommodated was via the deductive assumption that the level of water in the ice-dammed lake was high (on the basis of the previous state of knowledge regarding the relief and geological structure of the Warsaw Basin).

The fact that there was an exit for meltwaters into the "Warsaw-Berlin ice-marginal valley" at such relatively low altitudes as 98 and 88 m a.s.l., followed by a directing of these waters to the east via a section of the aforementioned channel into the Warsaw Basin at an altitude of 83–84 m a.s.l. excludes not only the possible existence at that time of the Warsaw ice-dammed lake (whose water level was above this altitude), but also the existence of an outflow route from the lake westwards, via the "Warsaw-Berlin ice-marginal valley".

Such a standpoint on the part of the author brings further consequences. For example, what were the conditions – and at what time was it possible – for alluvia to be laid down on the Radzymin-Błonie Level, since this is 10–20 m above that at which meltwaters would have flowed into the Warsaw Basin from the east? Even more crucial is the issue raised previously (Wiśniewski, Andrzejewski 1994) regarding the very existence over c. 2000 years of a Warsaw ice-dammed lake with no relief channel serving to take away the excess of water, even though their was constant alimentation via the Vistula from the south and the meltwaters passing along the Biebrza-Narew ice-marginal valley from the north-east?

6. THE RELIEF CHANNEL FROM THE WARSAW ICE-DAMMED LAKE

Similar conditions to those generating the ice-dammed lake in the Vistula Valley 20,000 years ago were put in place by people in 1969, as the Vistula was dammed near Włocławek. A difference between the events lay in the fact that the ice sheet – before it had reached the Gąbin area and blocked off the Vistula valley in the eastern part of the Płock Basin – had for the whole time been hindering or preventing the river from flowing north, with the result that there was probably a permanent ice-dammed lake in front of it as water piled up the Vistula valley (Fig. 5). Did it already have its relief channel at that time? This is hard for anyone now to determine.

The blocking off of the Vistula in 1969 happened quickly. In just one year, the river at the dam site was raised 11m above its previous level. The reservoir formed was 58 km long, with the waters having their own "relief channels" via the turbines of the power plant or weirs. Thanks to these the level of the reservoir has not fluctuated very markedly.

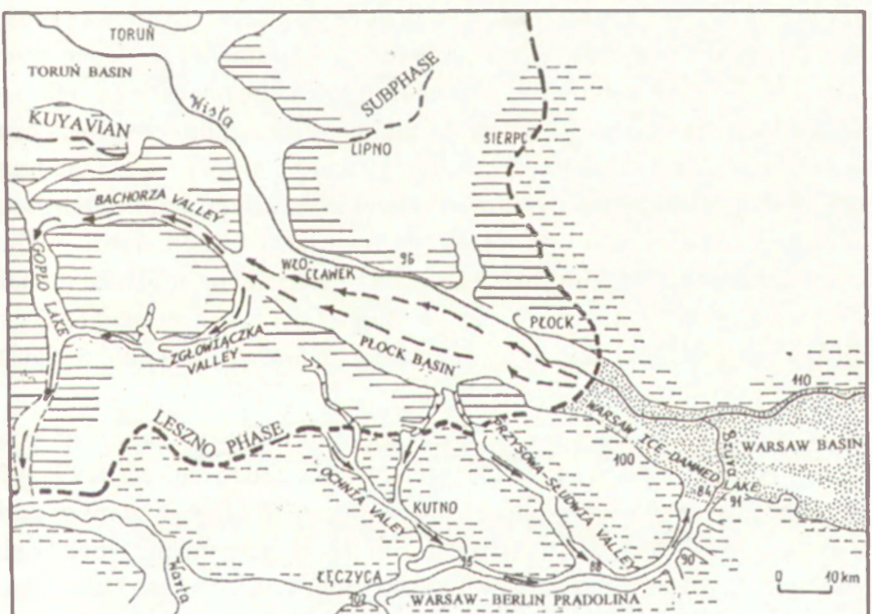


Fig. 5. Probable route of outflow of waters from the ice-dammed lake in the Warsaw Basin

Prawdopodobne drogi odpływu wód z jeziora zaporowego w Kotlinie Warszawskiej

Bearing the above in mind, it is possible to imagine the effects that the ice sheet of the Leszno Phase would have had in blocking the river over a longer period.

The Leszno Phase is dated at 20,000 years BP, while the Kujawy sub-phase – during which the edge of the ice sheet encroached ca. 75 km further to the north-west along the Vistula Valley near today's Nieszawa – dates back to around 17,200 years BP. During this sub-phase, a Vistula fed by meltwaters flowing along the Biebrza-Narew ice-marginal valley had no possibility of flowing west via the Noteć-Warta proglacial channel, since this did not yet exist. In which direction, then, was there an outflow of water from the ice-dammed lake that must have existed in the Warsaw Basin as well as below for over 2000 years, if the route to the west via the "Warsaw-Berlin Proglacial Channel" was rendered impossible for the above reasons? Only one answer to this question is forthcoming – that the waters from the ice-dammed lake most probably penetrated into the ice, just as earlier authorities L.Henkel (1909), F. Wunderlich (1917) and J. Samsonowicz (1922) had suspected.

Left behind from the time when the ice-sheet was in the Płock Basin are quite numerous trough formations whose depressions retain lakes of various sizes to this day, and even eskers, as well as tunnel forms (Skompski 1963). Since the Płock Basin area, as a fragment of the Vistula Valley, sloped down (then as now) to the north-west – which is to say in the opposite direction to that in which the ice sheet was advancing, the aforementioned forms occur in its eastern part at altitudes of between 80 and 82 m a.s.l., while the fragments of channel are retained on lower terraces in the central and western parts. The presence of channel-like forms attests to the fact that there were – within the ice sheet in the Płock Basin – numerous sub-glacial or in-glacial tunnels via which meltwaters flowed in the direction of the ice sheet's edge. These could subsequently be used as escape routes for waters from the proglacial body of water.

The author bases this view on observations of frequent changes of location of meltwater outflows from the glaciers in Iceland in the years 1993–97. This indicates that both sub-glacial and in-glacial drainage systems experience rapid change. Inactive tunnels in the ice sheet covering the Płock Basin could have been used effectively by the water from the ice-dammed lake. Their penetration of the ice sheet and thrust could also have been brought about by a raising of its marginal part.

In 1996, the eruption of a volcano beneath the Vatnajökull ice-cap in Iceland led to the loss of ca. 2 km^3 of water from the nearby Grimsvotn caldera lake lying beneath it. This led to a rising in the level of the glacier, which was followed by an outflow of water in sub-glacial conditions over a distance of some 50 km in the direction of the 23 km wide snout of the Skeidarar glacier. The flow of water in these conditions reached ca. 50 000 m^3/s. This was a volume fifty times greater than the flow of the Vistula near Tczew under average conditions. Similar phenomena – if on smaller scales – occur very widely in Iceland.

The mechanism of outflow from the Goes ice-dammed lake on Spitsbergen has also been described, by M. Grześ and M. Banach (1984). The waters from this lake are created steadily on the Gås glacier, and – as a result of hydrostatic uplift – they rise over the ice dam and flow away in sub- or in-glacial circumstances towards its front.

The result with both of the phenomena described is a runoff of water in line with the incline of both the surface beneath the ice and the glacier itself. However, in the area of the Plock Basin, the conditioning of the outflow of water from the ice-dammed lake was rather different, which is to say in line with the incline of the sub-glacial surface, but not in line with the direction from which the glacier was coming and the growth in its thickness.

The author met with conditions of this kind in Iceland in 1995, at the Tungnaár glacier which is part of the Vatnajökull ice wall 28 km long. In the course of a charge by this part of the Vatna glacier in late 1994 and early 1995, as a result of which it shifted by about 1200 m, there was a covering up of a certain section of the marginal valley along which a river was flowing. It is difficult to judge just how far from the steep, high and cracked front of the glacier it found itself, beneath the ice. Nevertheless the river continued to take the same course, flowing under the ice.

A fundamental problem when it comes to the possible tunnel-mediated penetration of the ice sheet in the Plock Basin by waters from the ice-dammed lake is that concerning just how far upwards in the ice sheet they could have penetrated and what was their further fate? The morainic elevation adjacent to the Płock Basin are at 90–115 m a.s.l. If the level of waters in the Warsaw ice-dammed lake could not have risen above 84 m – as has already been indicated – then their escape within the ice could have taken place at this altitude at most. In this situation, a role draining the ice sheet could have been played by the still-mysterious valley of the Bachorza, which begins from a clearly lowered area within the Kujawy Heights north of Brześć Kujawski, ca. 60 km down the Vistula Valley from the line of maximal extent of the Leszno Phase (Fig. 6). Running through the centre of this area is an E–W trough that the Zgłowiączka has adapted for its course (Andrzejewski 1984).

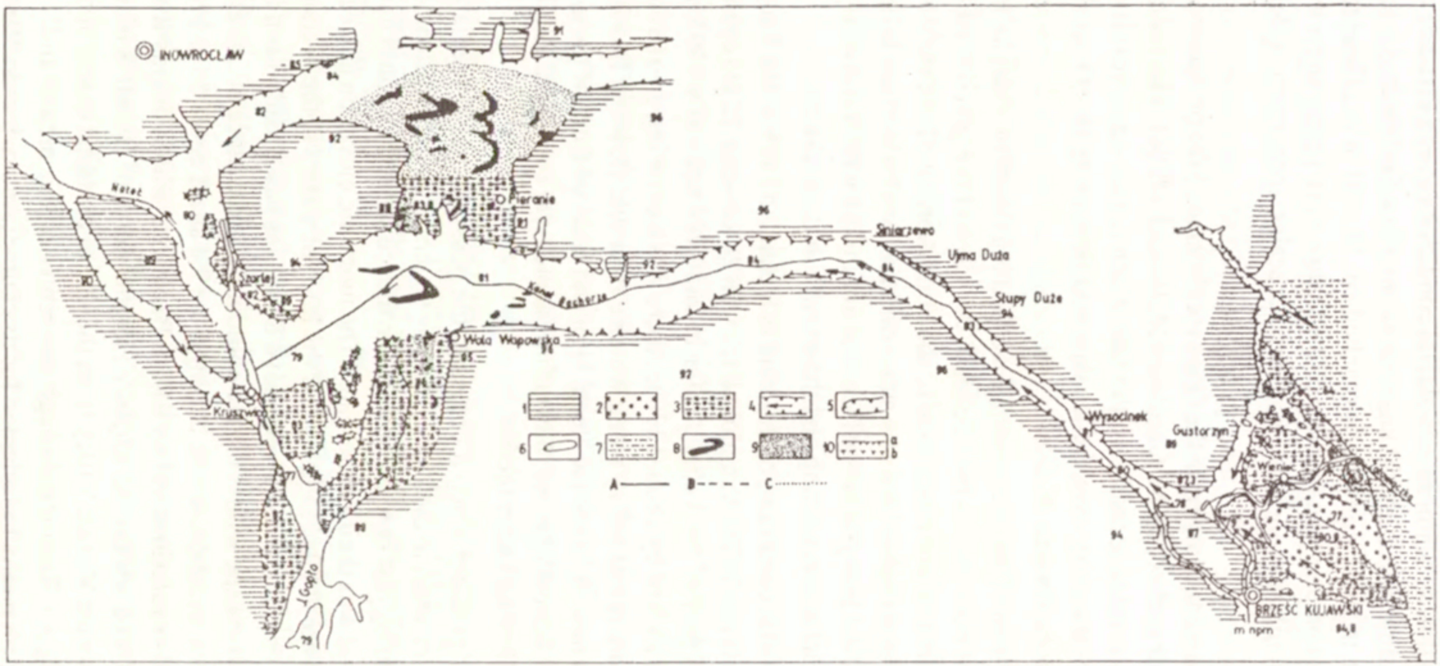


Fig. 6. Geomorphological sketch of the Bachorza Valley. 1 – ground moraine; 2 – levels with a cover of sandy and gravel sediments; 3 – erosive meltwater levels; 4 – meltwater valleys; 5 – troughs; 6 – cave-in; 7 – Vistula erosive terrace; 8 – dunes; 9 – plains of windblown sands; 10 – edges: a. distinct, b. indistinct.

Szkic geomorfologiczny doliny Bachorzy. 1 – morena denna; 2 – poziomy z pokrywą osadów piaszczysto-żwirowych; 3 – erozyjne poziomy wód roztopowych; 4 – doliny wód roztopowych; 5 – rynny; 6 – wytopiska; 7 – terasa erozyjna Wisły; 8 – wydmy; 9 – równiny piasków przewianych; 10 – krawędzie: a. wyraźne, b. niewyraźne.

In this area the author identified two levels, of altitudes 80–82 and 75–77 m a.s.l. (Wiśniewski 1974, 1976, 1987, 1990). The higher level, most often formed from morainie clay, has been recognised as an erosive level. Incised in it, to the south of the Zgłowiączka channel, is the lower level at 75–77 m, which takes the form of two tracks. Covering it to depths of between 1 and 4 m are sands and gravels that contain shelly fragments of species present within the Eemian marine fauna. However, the greatest surprise is that the remains of the Eemian fauna incorporated at the foot of the ice sheet in the area today occupied by the Baltic are here found in sand-gravel formations deposited by meltwaters flowing from the north or north-east, which is to say from the area of the Płock Basin. Attesting unequivocally to this are the results of measurements of the dip of their lamina. From this it needs to be concluded that the waters responsible for the deposition of sands and gravels with Eemian faunal remains flowed in sub-glacial conditions, thereby being able to wash out the basal sediments of the ice sheet.

The further course of these waters probably led to the west via the Bachorza Valley. This valley, which is 42 km long, cuts the Kujawy Elevation by way of a loop extending to the north, before emptying into the Gopło channel near Kruszwica. While its initial width is of some 800 m, this has increased to 4 km by the time the confluence point is reached. In the first section the valley floor reaches altitudes of 76–77 m a.s.l., thereby corresponding with the lower tracks left by flowing water in the Brześć Kujawski area.

However, further along the course, the valley bottom rises (near Ujma Duża and Siniarzewo) to 84 m a.s.l. From this locality on, a downward slope to the west is observed, such that where contact is made with Lake Gopło the altitude is of 79 m a.s.l. The profile of the floor of the Bachorza Valley is thus markedly convex, with its confluence section at a higher elevation than its first section. The result of this convexity near Siniarzewo is that a bifurcation takes place: water in the Bachorza channel flows NE from this place to the Zgłowiączka, as well as west to Gopło. It is here that the divide between the Vistula and Oder drainage basins runs.

In spite of the existence of such a situation – in non-compliance with the fundamental laws of hydrodynamics – it has been documented that the waters (*inter alia* from the Płock Basin area) are directed to the west along the Bachorza Valley. This has been reported by way of textural measurements made in the initial section of the valley. In earlier works in which topics included the role of the Bachorza Valley in the geomorphological development of the Vistula Valley, the author looked for an explanation of the elevation of the valley floor in neo-tectonic or glacio-isostatic movements of the Earth's crust (Wiśniewski 1974, 1976, 1990).

However, bearing in mind the facts detailed above, as well as the geomorphological conditioning allowing for analysis of the route of outflow of waters from the ice-dammed lake in the Vistula Valley, it is worth considering the possibility that the Bachorza Valley initially conveyed waters in sub-glacial conditions. After linking up in the west with the Gopło channel, whose genesis was last a topic of interest for P. Molewski (1999), these could in the same conditions have flowed along that channel, and then via the trough of today's Lake Ślesińskie, southwards to the ice-marginal valley of the Warta in the Konin area, skirting the divide between the Vistula and Oder drainage basins near Łęczyca. Moreover, from the vicinity of what is today Brześć Kujawski, these waters could also have taken another outflow route in sub-glacial conditions, which is to say that leading via the Zgłowiączka channel – the subject of research by L. Andrzejewski (1984), and then via the Głuszyńska Valley again to the Gopło trough (Fig. 5).

7. FINAL REMARKS

The author is aware that there are many issues for discussion, including problematic ones, in the considerations offered here regarding a relief channel from the ice-dammed lake in the Warsaw Basin during the last glaciation, as based on:

– the results of detailed geomorphological and geological research in the Vistula Valley between the Płock and Toruń Basins, and in the proglacial valleys on the Kutno Plain and in the lower Bzura Valley,

– experience gained in the course of work on the marginal zones of today's ice sheets. Nevertheless, these should inspire a revision of views held hitherto in regard to the age of clays situated at various elevations in the Warsaw Basin between Otwock and the confluence point of the Bzura and Vistula. For a doubt arises as to whether these could have been laid down during one glacial episode.

What has been documented is the non-existence during the Vistulian Glaciation of the "Warsaw-Berlin ice-marginal valley" along the whole length beyond the then-arising proglacial lake in the Warsaw Basin. The beginning would therefore seem to be the section now used by the Ner as it flows into the Warta, with the consequence that the name should be the Warta-Oder ice-marginal valley – one that in fact can be met with already in certain publications.

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PROBLEM DROGI ODPŁYWU WÓD Z JEZIORA ZAPOROWEGO W KOTLINIE WARSZAWSKIEJ PODCZAS FAZY LESZCZYŃSKIEJ

Streszczenie

Konsekwencją transgresji lądolodu vistuliańskiego na obszar Polski było blokowanie odpływu rzek płynących na północ. W takiej sytuacji była także Wisła. Gdy czoło lądolodu, podczas fazy leszczyńskiej przebiegało w poprzek wschodniej części Kotliny Płockiej, w jej dolinie utworzyło się jezioro zaporowe. Wypełniło ono Kotlinę Warszawską oraz przewężenie doliny Wisły w dół od tej kotliny (ryc. 1).

W literaturze geomorfologicznej panuje pogląd, iż wody z zastoiska warszawskiego płynęły wówczas na zachód przez pradolinę "warszawsko-berlińską". Wyrażane były też poglądy wykluczające taką możliwość ich odpływu.

Autor we wspólnym artykule z L. Andrzejewskim (1994) zakwestionował możliwość odpływu wód z jeziora zaporowego w Kotlinie Warszawskiej na zachód "pradoliną warszawsko-berlińską", na podstawie wyników badań geomorfologicznych i geologicznych doliny Wisły między Kotliną Warszawską a Kotliną Toruńską oraz dolin proglacjalnych rozcinających Równinę Kutnowską, a także ujściowego odcinka doliny Bzury.

Równinę Kutnowską, położoną na przedpolu fazy leszczyńskiej, rozcinają doliny proglacjalne z nią związane, z których największymi są: Ochni i Przysowy-Słudwi. Odpływały nimi wody roztopowe do "pradoliny warszawsko-berlińskiej". Wody te, płynące doliną Ochni, uchodziły do "pradoliny" około 50 km na zachód od Kotliny Warszawskiej na wysokości około 98 m n.p.m., natomiast wody płynące doliną Przysowy-Słudwi kontaktowały się z nią w odległości około 25 km od kotliny w poziomie 89 m n.p.m. (ryc. 2, 3). Dalsza droga wód roztopowych wiodła na wschód do Kotliny Warszawskiej, do której wlewały się na poziomie 83–84 m n.p.m. (ryc. 4). Dokumentują to jednoznacznie badania strukturalno-teksturalne osadów L. Andrzejewskiego (1991, 1994) w ujściowym odcinku doliny Bzury.

Do połowy lat 70. utrzymywał się pogląd S.Z. Różyckiego (1967), że iły warwowe leżące w Kotlinie Warszawskiej na poziomie radzymińsko-błońskim o wysokości 103 m n.p.m. w okolicy Otwocka i obniżającym się do 82–83 m n.p.m. między Kotliną Warszawką a Kotliną Płocką (19–20 m nad poziomem Wisły), osadziły się w zastoisku podczas stadium Wkry zlodowacenia środkowopolskiego. Późniejsze badania dowiodły, że pod wspomnianymi iłami, po prawej i lewej stronie Wisły, zalegają osady eemskie (Janczyk-Kopikowa 1975; Karaszewski 1975; Sarnacka 1982). Zatem iły zastoiskowe poziomu radzymińsko-błońskiego pochodzą z okresu zlodowacenia vistulian.

Odpływ wód z jeziora zaporowego w Kotlinie Warszawskiej na zachód "pradoliną warszawsko-berlińska" mógłby odbywać się tylko wtedy, gdyby poziom jego wód sięgał wysokości ponad 102 m n.p.m., bowiem dno "pradoliny" w okolicy Łęczycy, w miejscu obecnego działu wodnego między Wartą płynącą na zachód a Bzurą na wschód do Kotliny Warszawskiej ma właśnie taką wysokość.

Jednak stwierdzenie, oparte na podstawie wyników badań proglacjalnych dolin Równiny Kutnowskiej oraz doliny dolnej Bzury, iż płynące nimi wody roztopowe podczas fazy leszczyńskiej i poznańskiej nie kierowały się na zachód lecz na wschód, i osiągały Kotlinę Warszawską w poziomie 83–84 m n.p.m., zaprzecza możliwości istnienia tak wysokiego poziomu wód w zastoisku warszawskim. Był on zatem znacznie niższy, aniżeli można sądzić po wysokości zalegania iłów.

Ponieważ jezioro zaporowe w dolinie Wisły mogło utrzymywać się przez ponad 2 tys. lat i było przez cały czas zasilane przez wody części dorzecza Wisły nie objętego ostatnim zlodowaceniem, ich nadmiar z tego jeziora miał zapewne swoją drogę odpływu. Wobec przedstawionych wyżej faktów jednoznacznie zaprzeczających możliwość kierowania się wód z jeziora zaporowego na zachód przez "pradolinę warszawsko-berlińską", mogły one jedynie ginąć w lądolodzie, który przykrywał Kotlinę Płocką (ryc. 5). Ta część lądolodu obfitowała w liczne sub- i inglacjalne tunele, o czym świadczy występowanie tu wielu jezior rynnowych oraz ozów, którymi wody mogły migrować w obręb czaszy lodowej. Ponadto lądolód przykrywał obszar pochylający się w kierunku z którego się nasunął, co ułatwiać mogło także w warun kach subglacjalnych penetrację wód z jeziora zaporowego. Dalsza ich droga mogła prowadzić w warunkach subglacjalnych doliną Bachorzy i rynną Zgłowiączki na zachód do rynny Gopła, a następnie rynną Jeziora Ślesińskiego na południe do pradoliny Warty pod Koninem (ryc. 5, 6).

Zaprezentowana hipoteza dotycząca możliwości odpływu wód z zastoiska warszawskiego jest zapewne dyskusyjna, lecz należy wspomnieć, że już wiele lat wcześniej m.in. L. Henkel (1909), E. Wunderlich (1917) i J. Samsonowicz (1922) przyjmowali odpływ Wisły na północ w warunkach subglacjalnych. Hipotezie tej nie brakuje także podstaw, za które można uważać współcześnie zachodzące zjawiska mniej lub bardziej katastrofalnych spływów wód roztopowych z jezior zaporowych w warunkach sub- i inglacjalnych, znane i opisane już w literaturze (Grześ, Banach 1984). Spotkał się z nimi także autor w czasie badań na Islandii w latach 90.

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MORPHOLOGICAL FORMS CONNECTED WITH EXPLOITATION OF MINERAL RAW MATERIALS IN POLAND

1. INTRODUCTION

The earth surface is modified to great extent as the result of human activity. Morphological forms connected with agricultural and building activity are most widely spread. They are common in the areas of human settlements. However, they are hardly noticeable because they are small on the vertical scale and are formed relatively slowly. More spectacular are such forms as earthworks, channels or water reservoirs. Especially important are forms created as the result of the excavation of mineral resources. They occur in rela-

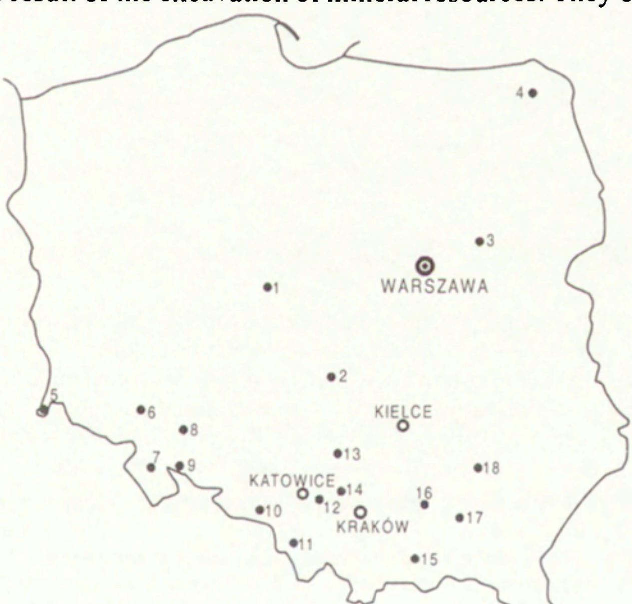

Fig. 1. Sites discussed in text
Stanowiska omawiane w tekście
1 – Konin, 2 – Bełchatów, 3 – Sadowe, 4 – Suwałki, 5 – Turoszów, 6 – Strzegom, 7 – Kłodzko, 8 – Strzelin, 9 – Paczków, 10 – Buk, 11 – Skoczów, 12 – Szczakowa, 13 – Niegowa, 14 – Olkusz, 15 – Nowy Sącz, 16 – Tarnów, 17 – Pilzno, 18 – Machów

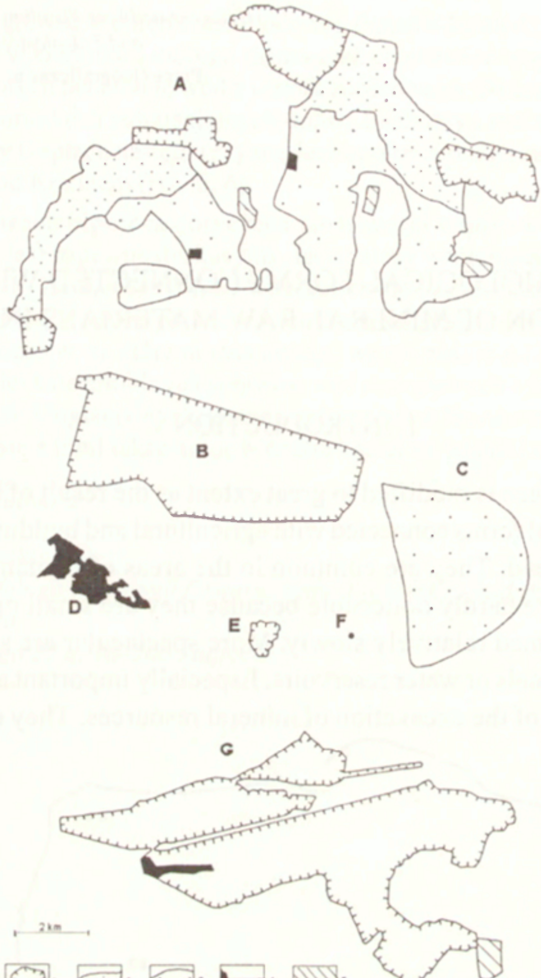

Fig. 2. The size of some antropogenie morphological forms connected with open mining 1 – pits, 2 – new heaps, 3 – reclaimed heaps, 4 – water reservoirs, 5 – ponds. A – pits and heaps of brown coal mines north of Konin, B – pit of brown coal mine in Bełchatów, C – external heap in Bełchatów mine, D – post-exploitation pools of gravel pit in Paczków, E – limestone quarry in Czatkowice, F – marble quarry – Zygmuntówka near Kielce, G – pits of filling sands between Szczakowa and Bukowno

Wielkość niektórych form antropogenicznych w Polsce związanych z eksploatacją odkrywkową 1 – wyrobiska, 2 – hałdy świeże, 3 – hałdy rekultywowane, 4 – zbiorniki wodne, 5 – stawy osadnikowe. A – wyrobiska i hałdy kopalń węgla brunatnego na północ od Konina, B – wyrobisko kopalni węgla brunatnego w Bełchatowie, C – hałda zewnętrzna kopalni w Bełchatowie, D – stawy poeksploatacyjne żwirowni w Paczkowie, E – kamieniołom wapieni w Czatkowicach, F – kamieniołom marmuru Zygmuntówka k. Kielc, G – kopalnie piasków podsadzkowych między Szczakową a Bukownem tively short periods of time (dozen or so years), are considerable in size (Fig. 2) and play an important role in public reception (Rutkowski 2001a). These problems are well known to miners and geologists dealing with mineral row deposits but less known to geomorphologists. This paper concerns some morphological forms found in Poland, in moderate climate, which were formed mostly as the result of surface and hole exploitation and sometimes underground mining. All the sites described in this paper, in which such forms were found, are marked on the map of Poland (Fig. 1).

2. FORMS CREATED AS THE RESULT OF OPEN MINING

Stony, sandy, gravel and clay pits are formed as the results of surface exploitation. Different in character are the brown coal mines in Bełchatów, Konin (Fig. 2 A, B), Turoszów and sulphur mine in Machów, which are very big and with high overlay. Pits can be on slopes and on flat areas. Slope pits (Fig. 3 I, e.g. sandstone quarry in Radków near Kłodzko, dolomite quarry in Dubie near Kraków) generally do not require artificial drainage. It is necessary in the case of deeply cut pits situated mostly in the flat areas and on small hills (Fig. 3 II, III), e.g. granite quarry in Strzelin, or limestone quarry at Zakrzówek in Kraków. Sometimes quarries situated on a slope of a hill change with the progressing exploitation into deeply cut pits. An example of this is Ostrówka quarry, near Kielce, which has now bottom at 190 m a.s.l. and was originally situated on a slope of a hill 303 m high a.s.l. Sporadically deeply cut pits are found to be situated near top of a hill, e.g. at Pychowice in Kraków (Fig. 3 III). When exploitation is terminated slope pits remain usually dry, while deeply cut pits, which were drained during the exploitation, fill with water, e.g. some quarries of granite in Strzegom, limestone in Chęciny near Kielce or at Zakrzówek in Kraków (Fig. 3 II B).

Stability of pits and their walls depends on litology of rocks, size of pit and on litology and thickness of the overlay and climate. It also depends on the age of a pit. In the moderate climate in Poland, walls made of slightly cracked granite are practically stable. It can be seen in granite quarry in Strzelin, where there are vertical walls 110 meters high, with the width of pit 100–200 m and its length 650 m. Also limestone walls of Skały Twardowskiego in Kraków are stable. They are overgrown with bushes and trees, which make them similar to natural rocks (Fig. 3 I B a). If non-resistant rocks are present in pits then, when exploitation ceases, the walls crumble and their morphology is less visible. (Fig. 3 I B b). Small pits are quickly filled, especially if overlay consists of clay-rocks, loams, loesses or sands.

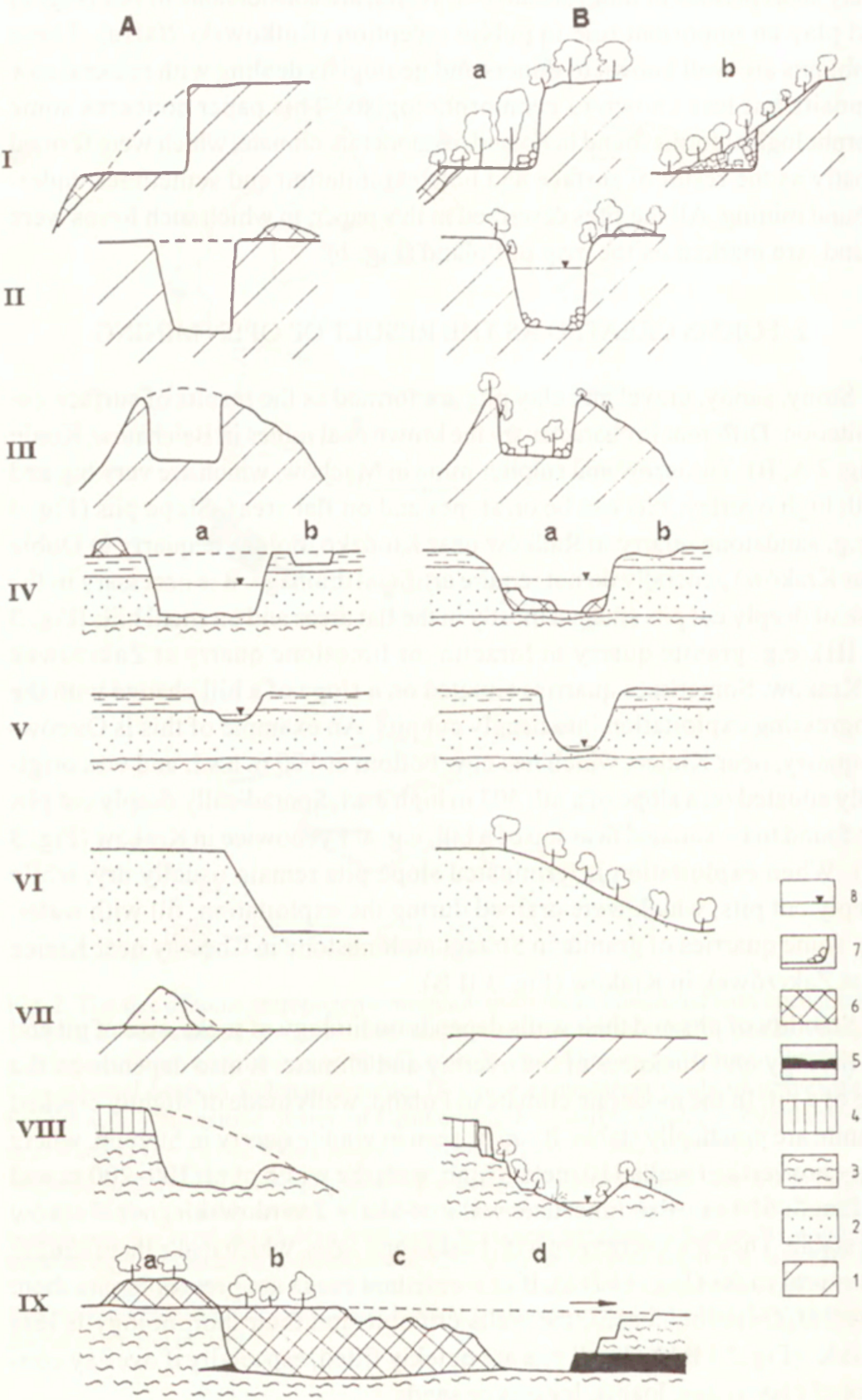
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Fig. 3. Pits and heaps when excavation was stopped (A) and their transformation some scores years later (B). (Rutkowski 1997, supplemented)
I – slope pit, a – in resistant rocks, b – in non-resistant rocks, II – deeply cut pit, III – deeply cut pit on the top of the hill, IV – pits on river terrase, a – gravel pit, b – muds pit, V – sand exploatation in the river trought, VI – sand pit on the slope, VII – sand pit in dune, VIII – slope pit of loam, IX – brown coal pit: a – external heap, b – reclaimed internal heap, c – new heap, d – direction of exploitation. 1 – resistant rocks, 2 – sands and gravels, 3 – clays, muds and loams, 4 – loesses, 5 – brown coal, 6 – heaps, 7 – crumbled rocks, 8 – water level. Dashed line – original land morphology

Wyrobiska i hałdy w chwili zaprzestania wydobycia (A) i ich dalsze przekształcanie się (B) (Rutkowski 1997, uzupełnione)
I – wyrobisko stokowe, a – w skałach odpornych, b – w skałach mało odpornych, II – wyrobisko wgłębne, III – wyrobisko wgłębne na szczycie pagórka, IV – wyrobiska na terasie rzecznej, a – żwirownia, b – punkt eksploatacji mad, V – eksploatacja piasku w korycie rzeki, VI – piaskownia na stoku, VII – piaskownia w wydmie, VIII – wyrobisko stokowe (eksploatacja glin), IX – hałdy: a – zewnętrzna, b – zrekultywowana hałda wewnętrzna, c – hałda świeża, d – kierunek eksploatacji. 1 – skały zwięzłe podłoża, 2 – piaski i żwiry, 3 – iły, mułki i gliny, 4 – lessy, 5 – węgiel brunatny, 6 – hałdy, 7 – osypiska, 8 – poziom wody. Linia przerywana – pierwotna morfologia terenu

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If we ignore the excavations done by local people the size of pits depends on the purpose of exploitation. The pits of brown coal, sulphur and sand for filling have the largest area (Fig. 2 A, B, G). Also limestone and marls pits producing cement or lime are nuge (e.g. Trzuskawica near Kielce, 0.9 km^2). Similar in size are limestone pits providing material for furnace flux (Czatkowice near Kraków, 0.6 km^2, Fig. 2 E).

Similar or smaller are aggregate pits providing material for building and roads, e.g. sandstone quarry in Klęczany near Nowy Sącz (60.9 ha) and quartzite quarry in Wiśniówka near Kielce (35.2 ha), porphyry quarry in Zalas near Kraków (27.2 ha) and granite quarry in Graniczna near Strzegom (12.5 ha).

The smallest are quarries excavating stone blocks for decoration purposes (Fig. 2 F), eg. marble (Dębnik near Kraków, 1.9 ha; and Zygmuntówka near Kielce, 1.5 ha) and sandstone (Szczytna near Kłodzko, 2.1 ha, Głębiec near Skoczów, 1.6 ha). Small pits composed of resistant rocks, with area of several square meters, if intact can be morphologically stable.

2.1. SANDS AND GRAVELS

In sand and gravel mines exploitation can be carried out above the ground water level and below it in pits filled with water (Fig. 3 IV, V). Dune sands, esker sands and gravels are excavated above ground level, leading to flattened area (Sadowne near Brok and Tarnów region, Fig. 3 VII). Sands and gravels of sandur origin can be excavated from dry deposits (Potasznia near Suwałki) and also from partly filled with water deposits (Sobolewo near Suwałki). When exploitation ceases, edges of pits are smoothed to some extent. Sometimes big water reservoirs are formed. The most useful fraction of sandy-gravel sediments consists of gravels, while demand for sand is smaller. Big amounts of sand obtained during sorting out are used to fill in old pits. This sand is transported by water in pipes forming at their outlet alluvial fans, similar to glaciofluvial deposition forms (Zieliński 1982).

Sands for filling are excavated above the ground water level. In the biggest sand mine in Poland, between Szczakowa and Bukowno, near Olkusz (Fig. 2 G) sands of fluvial and profluvial origin are excavated. The total area of pits is 32.9 km^2, and their depth is up to 30–40 m. In the years 1954–1996 as much as 617×10^6 m^3 of sand were excavated, with a maximum of 22×10^6 m^3 in the year 1970. When exploitation ceases the edges of pits are smoothed to some extent and the whole area is afforested (Fig. 3 VI). After 40 years part of the old pits differ from the intact areas only on the detailed topographic maps. Some old pits are filled with water (e.g. reservoirs Dziećkowice near Katowice with an area of ca. 6.6 km^2).

Exploitation from underneath the water is carried out on a big scale on low river terraces in the Carpathians, Sudetes, in their foreland and in the valley of Odra river. (Fig. 3 IV). It starts with the removal of loamy-muddy sediments (thick up to a few meters), which are deposited on heaps. Then, sands and gravels (thick a few to a dozen or so meters), which are below those deposits, are excavated, leading to the reservoirs filled most often with water. They are covered in part by the overlay and filled with sand separated from gravels, and their edges become smoothed out. When exploitation ceases reservoirs remain. When pits are within the flooded areas they are filled with loams. When they are left intact their banks become afforested during a dozen or so years and resemble natural forms. Numerous pits can be found in the valley of Nysa Kłodzka west from Paczków where they spread over 3.5 km covering the area of 1.3 km^2 (Fig. 2 D) while in Odra valley they spread over ca. 10 km between Olza and Buków, and they are also in Dunajec valley near Tarnów or Wisłoka valley near Dębica. In old days excavation of sands and gravels was also carried out in river channels (Fig. 3 V B). As the result of this, and due to channelization, rivers incised into substratum, e.g. up to 4 m in Kraków during the 19th and 20th century.

2.2. PEATS

Outcrops remaining after peat exploitation can have either irregular or elongated shape (Fig. 4). They can be often seen in lower parts of the area,

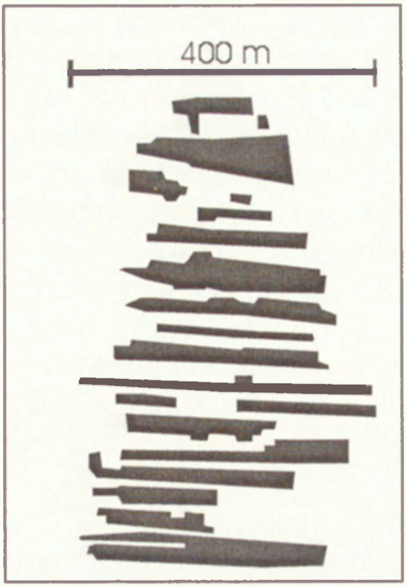

Fig. 4. Elongated outcrops after peat exploitation near Konin
Wydłużone wyrobiska pozostałe po wydobyciu torfu w pobliżu Konina

where the boundaries of outcrops correspond to the boundaries of individual proprieties, which were most often very long and narrow pieces of land. Hence, many outcrops are elongated. They are filled with water or overgrown with swamp vegetation.

2.3. CLAYS

Pits remaining after exploitation of clayey rocks are similar to those described above. The basic difference is that they are quickly levelled as the result of slumping (Fig. 3 VIII B), deeply cut pits are quickly filled with water. Pits differ in size. One of the biggest are those of refractory clays in Jaroszów near Strzegom (30–76 ha). Clay pits exploited for ceramic industry are often small (Zielonki near Kraków, ca. 5 ha).

2.4. BROWN COAL AND SULPHUR

Somewhat different in character are great pits and heaps connected with the exploitation of brown coal, e.g. Konin and Bełchatów (Fig. 2 A, B) and Turoszów, or sulphur in Machów. These deposits have very thick overlay which during the first stage of exploitation is deposited on external heaps (Bełchatów, Machów) and later on internal heaps.

In Konin region brown coal is present in zones taking narrow and irregular forms (Fig. 2 A). Exploitation is carried out in 4 outcrops with surface ranging from 0.7 to 3.9 km^2, giving a total of 12.0 km^2. New heaps of rough surface (ca. 10.5 km^2), are of internal type. Others (19.6 km^2) were reclaimed by levelling, afforesting or turning into agricultural land. It concerned both external and internal heaps, which sometimes can be distinguished only on the detailed topographical maps or aerial photographs.

If Konin region is taken as 218 km^2, we get high degree of the land surface transformation, of the order of 19.6%. This value does not include settlers of the power stations (2.3 km^2), fish pools (2.1 km^2) and peat pits (0.3 km^2). Generally speaking Konin region is highly transformed and it is characteristic that its pits are quickly filled with heaps and reclaimed (Fig. 3 IX).

Brown coal is also excavated in Bełchatów region and there is the biggest pit and heap in Poland (Fig. 2 B, C). The removal of the overlay started in 1977. In the beginning of eighties the pit had an area of ca. 5.4 km^2, and now of 21.9 km^2. The pit lies in a fault trough oriented WWN–EES and is over 8 km long, 2–3 km wide and ca. 200 m deep (maximum 300 m). The external heap formed in 1977–1993 covers the area of nearly 14.8 km^2 and is ca. 170 m high above the ground. They are the deepest pit and the highest artificial mountain in Poland.

Big dormant pit and a heap are in Machów near Tarnobrzeg, where limestones with sulphur covered with Miocene clays and Quaternary sands were excavated. The pit is up to 80 m deep and has an area of ca. 5.9 km^2. When the bottom is made water-tight the pit will be turned into the lake. The heap made of clay has an area of 8.3 km^2 and is about 60 m high. It was built on the substratum consisting in its upper layer of plastic clays, which were gradually squeezed out. Here, under the sod cover, slices more than 10 m high were formed. Also horizontal shifting took place. In the places where heap was on non-plastic substratum deformations did not occur. Phenomena observed here can be regarded as anthropogenic model of glaciotectonic processes occurring at foreland of the continental ice shield (Dadlez, Jaroszewski 1994). Described here forms do not today exist and the heap is now covered with wood.

3. FORMS CREATED AS THE RESULT OF HOLE EXPLOITATION

Hole exploitation of salt is done by leaching salt with water through boreholes from the surface. As the result caverns are formed. They collapse, in turn, leading to sinkholes in the surface. Such forms were found near Krakow (Barycz near Wieliczka and Łężkowice near Bochnia – Garlicki 1993; Rutkowski 1994–95). In Barycz the salt was excavated at the depth of 230–280 m. It started in 1923 and during 70 years obtained about 10^{10} kg (Garlicki 1993). Since the specific weight of halite is 2.1–2.2 kg/m$^3\cdot10^3$, it corresponds to the volume of about 4.6x10^6 m^3. As the result of the exploitation the surface subsidence was observed and since 1931 formation of sinkholes (Fig. 5). 33 sinkholes of total area of ca. 5 ha were found by 1974. Most of them were filled with brine. The biggest had an area of ca. 2 ha and depth of up to 27 m. In 1974, in the north part of the area shown in Fig. 5, a sinkhole of area 1.9 ha, depth 7 m and volume 35 000 m^3 was suddenly formed and it got gradually larger. Its formation was accompanied by sudden outflow of brine carrying big blocks of sandstone. Now, all these sinkholes are covered with municipal waste and can not be seen.

Similar surface subsidence and formation of sinkholes take place at present (year 2002) in the area of dormant salt mine in Łężkowice near Bochnia (Garlicki 1993).

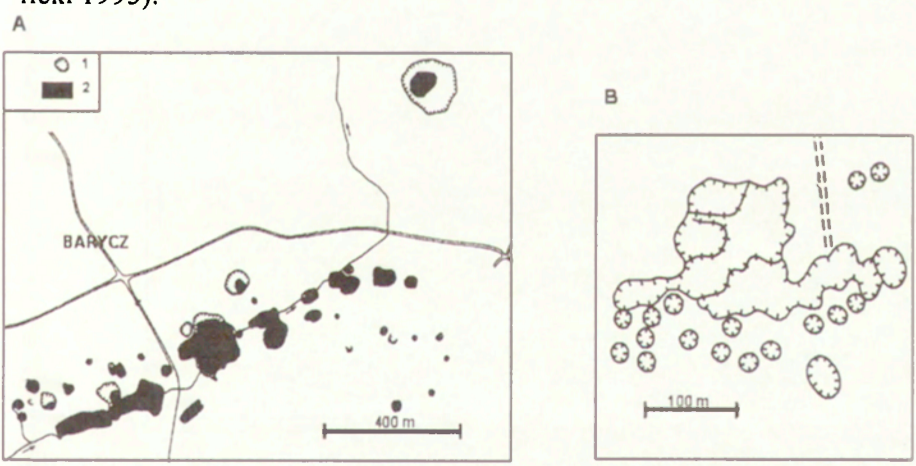

Fig. 5. Anthropogenic karst forms.
A – karst sinkholes in Barycz near Kraków (Rutkowski 1994–1995). 1 – sinkholes, 2 – sinkholes filled with water. B – anthropogenic karst funnels (pipping) west from Olkusz

Antropogeniczne formy krasowe.
A – leje krasowe w Baryczy k. Krakowa (Rutkowski 1994–1995). 1 – zapadliska, 2 – zapadliska wypełnione wodą. B – antropogeniczne zapadliska krasowe (sufozyjne) na zachód od Olkusza

4. FORMS CREATED AS THE RESULT OF UNDERGROUND MINING

Underground exploitation is carried out in Poland mainly in Upper Silesia (Katowice region) region, on its border and in Lublin region; excavated are: hard coal, zinc, lead and copper oars. Hard coal excavation in Upper Silesia leads to the formation of numerous heaps of huge area, sometimes of many hectares and also mining induced subsidence basin of considerable size and depth, and sometimes sinkholes. These forms are often filled with water and covered with material coming out of heaps. They are most numerous on the topographic maps done on the turn of 70s and 80s. Now, in the year 2002, when intensive reclaiming by levelling, filling in and building is carried out, the number of these forms is much lower and the morphology of those which remained is not clear. As the result of extensive mining, highly developed industry and building nearly all over the area of Upper Silesia is transformed by man activity.

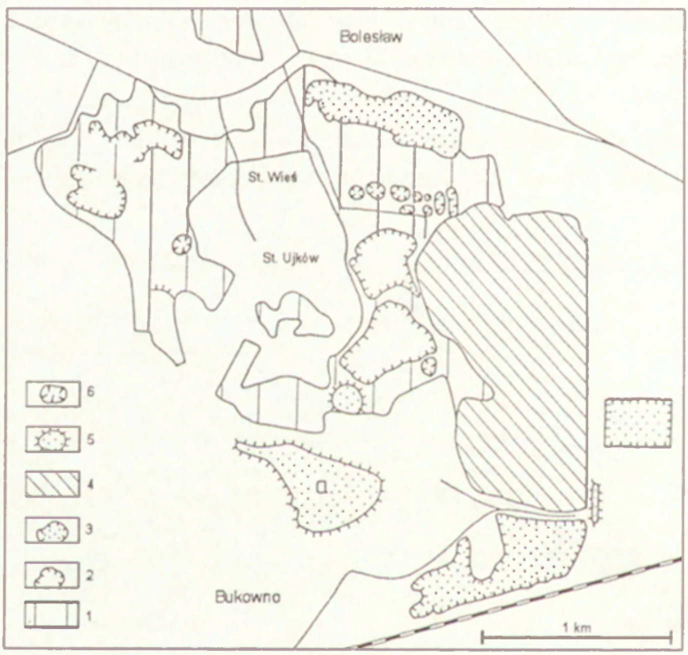

Fig. 6. Anthropogenic land form connected with mining and smelting in Bolesław near Olkusz (situation from 1996, after Rutkowski 2001b)

1 – areas transformed by mining and smelting, 2 – quarries, 3 – sand pits, 4 – flotation ponds, 5 – waste heaps, 6 – sinkholes, a – smelting heaps

Antropogeniczne formy terenu związane z górnictwem i hutnictwem w rejonie Bolesławia k. Olkusza (stan z 1996r., Rutkowski 2001b)

1 – tereny przeobrażone przez górnictwo i hutnictwo, 2 – wyrobiska skał zwięzłych, 3 – pias-kownie, 4 – stawy osadnikowe, 5 – hałdy i zwałowiska, 6 – leje zapadliskowe, a – hałda huty

Forms connected with underground mining in Bolesław near Olkusz (Fig. 6) are preserved much better. In XIV century, and probably even in XI or XII century, silver, and lead and now zinc ores were found there in Triassic dolomites and exploited. In Medieval ages excavation was carried out in small shafts several dozen meters apart. These places are now marked by small heaps, a few meters high, and sometimes shallow sinkholes (1–2 m).The shafts can be still found in the areas covered by woods. Contemporary underground mining in a one of the outcrops of Triassic beds leads sometimes to collapsing of underground rooms which were without backfill, and formation of sinkholes, sometimes of several dozen meters in diameter (Wilk et al. 1973).

Specially interesting phenomena in Olkusz region occurred in those places where triasic dolomites are covered with Quarternary sands, thick up to 60 m. Collapsing of post-exploitation rooms and drainage of substratum, caused flow of waters from Quaternary to triasic sediments. It caused translocation of sand to old mines and karst channels, giving rise to the formation of sinkholes on the surface. Sinkholes are either single or built of a few adjacent ones (Fig. 5). They are isometric or elongated, located along the line or scattered. The biggest have length of up to 300 m and depth of up to 40 m. At first the walls of sinkholes were steep and then the inclination decreased to the angle of order of 30–32° (angle of natural slip). In the years 1978–1983 Tyc (1977) found 111 sinkholes of total volume of $970 \cdot 10^3$ m^3. Described above forms, found in the sands of Olkusz region can be treated as anthropogenic piping funells or anthropogenic karst sinkholes reproduced in non-karst rocks. They occured as the result of exploitation or drainage of the area. The sinkholes are filled in by mine authorities.

5. THE DEGREE OF THE LAND TRANSFORMATION

The size of the area transformed by exploitation (pits and heaps) expressed as the percentage of the total area of a region, or given in square kilometres, can be an indication of the degree of this transformation.

It can be determined on the basis of the detailed topographic maps, e.g. 1:10 000, or/and aerial photographs. Data obtained from the maps are slightly underestimated because old pits and heaps could be taken on the maps as natural forms. Aerial photographs solve this problem for the areas not covered by plants. In order to calculate the area of the region, its boundaries have to be taken arbitrarily. They can correspond to the limits of a sheet or to the administrative boundaries of a town. It is also possible to use pits which are

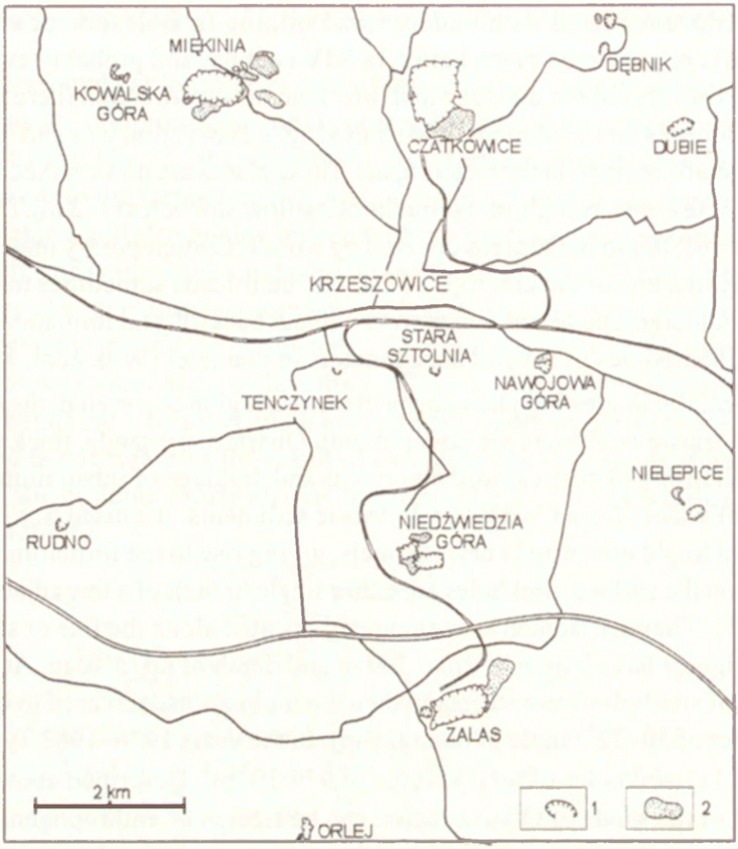

Fig. 7. Pits and heaps connected with an open exploitation in Krzeszowice near Kraków region (state about 1980, after Rutkowski 1997). 1 – pits, 2 – heaps

Przekształcenie powierzchni terenu związane z eksploatacją odkrywkową w rejonie Krzeszowic (stan ok. 1980 r., Rutkowski 1997). 1 – wyrobiska, 2 – hałdy

furthest apart to describe the approximate rectangle, which is taken as an area of the region. This approach is always used by the author. It should be mentioned, that for big, isolated mines, such as e.g. Bełchatów, the calculation of the transformed area is inappropriate.

The degree of transformation of Jura Krakowsko-Częstochowska between Niegowa and Olkusz is relatively small. It was calculated sampling the area by fields of 4 km^2 and found to be between 0 to above 2.5% (Szczypek, Trembaczowski 1987), while for the whole Jura it is 0.14% on average. Calculations done by the author confirm this (Rutkowski 1997). In Krzeszowice region near Kraków, regarded as highly changed by exploitation, only 1.95% of the total area of 115 km^2 was transformed about the year 1980 (Fig. 7). It includes 0.64% of area covered by heaps. The area transformed by exploita-

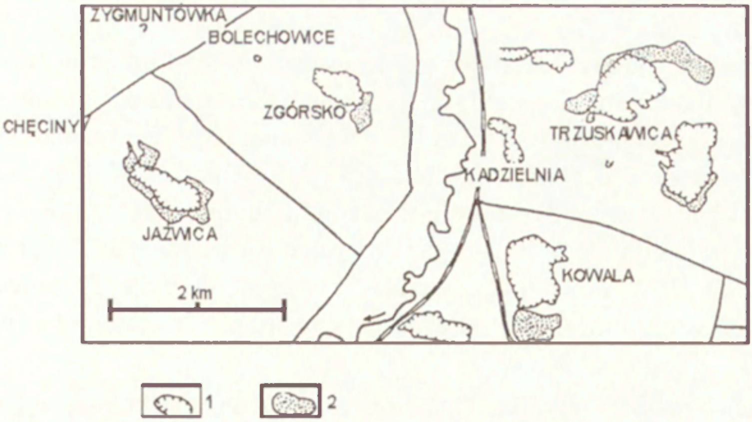

Fig. 8. Pits and heaps connected with an open exploitation in Chęciny near Kielce region (state about 1980, after Rutkowski 1997). 1 – pits, 2 – heaps

Przekształcenie powierzchni terenu związane z eksploatacją odkrywkową w rejonie Chęcin (stan ok. 1980 r., Rutkowski 1997). 1 – wyrobiska, 2 – hałdy

tion is only twice as big as that occupied by asphalt roads, embankments and other earthworks. Corresponding values for Strzegom region are 2.44 % (including 0.62% for heaps) of total area of 93 km^2, and for Nowy Sącz region 2.68% (including 1.07% for pits filled with water) of total area of 44 km^2. Similarly small degree of transformation is found for Kraków, 1.03% (including 0.37% pits filled with water) of the total area of 327 km^2 (Rutkowski 1994–1995). The highest degree of transformation was found for Chęciny region (Fig. 8) – 7.75% (including 2.00% for heaps), of the total area of 32 km^2. In 1996 it was even higher and equal to 9.78%.

In the regions where extensive open and underground mining and metallurgy industry takes place the degree of transformation can be much higher (Rutkowski 2001b), e.g. in the region of Bolesław near Olkusz (Fig. 6) it mounts to 28.90% (including 11.98% for sedimentation pools and foundery heaps) of the total area of 11.5 km^2.

The other measure of the transformation could be also anthropogenic mechanical denudation (Jania 1983) i.e. average lowering of an area in a given period of time, or since a certain point of time. It is calculated by dividing the volume of the excavated row materials by the area on which the surface exploitation takes place.

6. FINAL REMARKS

Exploitation of mineral raw materials leads to the formation of positive and negative morphological forms. Although they are often considerable in size the percentage of the area of Poland occupied by them is small. Stability of those forms depends on their size, litology, and first of all on their susceptibility to water and on the climatic conditions and their duration in time. In moderate climate in Poland the described forms are stable over at least several dozen of years. The most stable are edges of pits built of granites, some sandstones and limestones. The least stable are pits in clay rocks and their heaps.

In drier climate described here forms would be more stable, with the exception of edges of pits formed in sands, which could be blown away. On the other hand such forms would deteriorate quicker if the humidity were higher and/or in periglacial climate.

Antropogenic forms connected with the mineral row material exploitation occupy usually insignificant area, considerably smaller than that occupied by the forms connected with agriculture, building and communication. In public reception the influence of those forms on landscape is exaggerated, particularly when viewed negatively. It is one of the reasons for which they are eliminated by filling in or levelling. It should be remembered , however, that a heap which becomes an artificial mountain in a flat area, a lake in an old gravel pit or an amphitheatre in an old quarry could make landscape more picturesque.

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FORMY MORFOLOGICZNE ZWIĄZANE Z WYDOBYCIEM SUROWCÓW MINERALNYCH W POLSCE

Streszczenie

Praca dotyczy form morfologicznych powstających w klimacie umiarkowanym, związanych z wydobyciem surowców mineralnych w Polsce prowadzonym metodą odkrywkową i otworową, w mniejszym stopniu podziemną (ryc. 1).

Wyrobiska w których wydobywa się skały zwięzłe mają rozmiary od ok. 1 ha w przypadku dekoracyjnych marmurów, do prawie 1 km^2 w przypadku wapieni stosowanych w produkcji materiałów wiążących (ryc. 2). Głębokość może przekraczać 100 m. Ściany wyrobisk założonych w skałach zwięzłych i nie spękanych mogą być trwałe. Stan zachowania hałd zależy głównie od ich litologii budujących je utworów.

Piaski i żwiry wydobywa się ze złóż zawodnionych i suchych. W pierwszym przypadku pozostają po tym stawy poeksploatacyjne (ryc. 1, 2). W przypadku złóż suchych pozostają wyrobiska o znacznych nieraz rozmiarach (np. kopalnia piasków podsadzkowych w Szczakowej o powierzchni 33 km^2). Jeśli wyrobiska są dostatecznie duże, stanowią trwały element w morfologii.

Wyrobiska i hałdy związane z eksploatacją skał ilastych są na ogół niewielkie i rzadko dochodzą do kilkudziesięciu hektarów. Ulegają one szybko zasnuciu na skutek spełzywania (ryc. 3).

Odmienny ze względu na wielkie wymiary charakter mają formy związane z wydobyciem węgla brunatnego i siarki. Wyrobiska posiadają powierzchnię kilku km^2 (maksymalnie 22 km^2) i głębokość do ok. 300 m (ryc. 2, 3). Hałdy o powierzchni

do kilkunastu km^2 mają wysokość do 170 m. Największym obiektem jest kopalnia węgla brunatnego w Bełchatowie i związana z nią hałda. Charakterystyczny kształt niewielkich wyrobisk potorfowych (ryc. 4) wiąże się z własnością gruntów.

Eksploatacja soli kamiennej otworami wiertniczymi powoduje powstawanie zapadlisk (ryc. 5) i osiadanie powierzchni. Jest to klasyczny przykład antropogenicznych form krasowych wywołanych zapadaniem się komór powstałych na skutek wyługowania soli na dużej głębokości, które reprodukują się w nie krasowiejącym nadkładzie.

Górnictwo podziemne powoduje osiadanie terenu, powstawanie zapadlisk i hałd. Formy takie, typowe przed kilkudziesięcioma laty szczególnie dla Górnego Śląska, są obecnie w dużym stopniu zrekultywowane. Najlepiej są one zachowane w rejonie Olkusza (ryc. 6).

Stopień przekształcenia powierzchni terenu obliczony dla badanych rejonów (ryc. 6–8) jest zazwyczaj niewielki 1–2%, rzadziej bardziej znaczące 8–29%. Rzeczywista wielkość form jest zasadniczo mniejsza niż funkcjonująca w odczuciu społecznym. Stan zachowania i trwałość form (ryc. 3) zależy od ich charakteru, wieku i wielkości, litologii oraz klimatu.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
Prace Geograficzne nr 189 (2003)

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LITHOFACIES DIVERSIFICATION OF THE ALLUVIA IN THE AREA OF KĘPA DZIKOWSKA, KĘPA BAZAROWA, KĘPA STROŃSKA AND OF THE VISTULA FLOODPLAIN NEAR TORUŃ

1. INTRODUCTION

The dynamics and tendencies of the early and contemporary fluvial processes in the lower Vistula valley floor have been registered in the morphological and morphometric parameters of the channel and the floodplain, as well as in the structure and texture of the alluvia in terms of the vertical and spatial alterations therein. The evolution of the lower Vistula valley has been the subject of many papers, especially of the geomorphology (Galon 1934; Niewiarowski 1987; Tomczak 1982, 1987; Wiśniewski 1976, 1985, 1987, 1990; Drozdowski 1982; Drozdowski, Berglund 1976; and others), which make it possible to reconstruct numerous palaeoenvironmental elements which determine the major trends in the fluvial processes in the valley. A summary of many years research, set against the backcloth of palaeogeographic changes in the whole Vistula Basin since the last glaciation up to the present time, is offered by Starkel's work (2001).

Clearly visible against that background is a reduced interest in detailed sedimentological research into the Vistula's alluvia in terms of lithofacies. The issue is relatively young, which may be evidenced by papers on the lithofacies record of the alluvia in the floodplains of the Vistula or its tributaries, in terms of changing palaeohydrological conditions (Andrzejewski 1994, 1995, 1996; Andrzejewski, Szmańda 2000; Szmańda 2000; Kordowski 1997, 1999, 2001; Kordowski, Szmańda 2001; Myślińska 1980; Myślińska et al. 1982).

Another factor significantly determining fluvial processes during the past several hundred years has been man's continued pressure. In the case of the Vistula valley this factor is of particular importance, as not only the Vistula channel but also the whole river system has been subject to gradual anthropogenic alteration since the early Middle Ages at least (Babiński 1984, 1985,

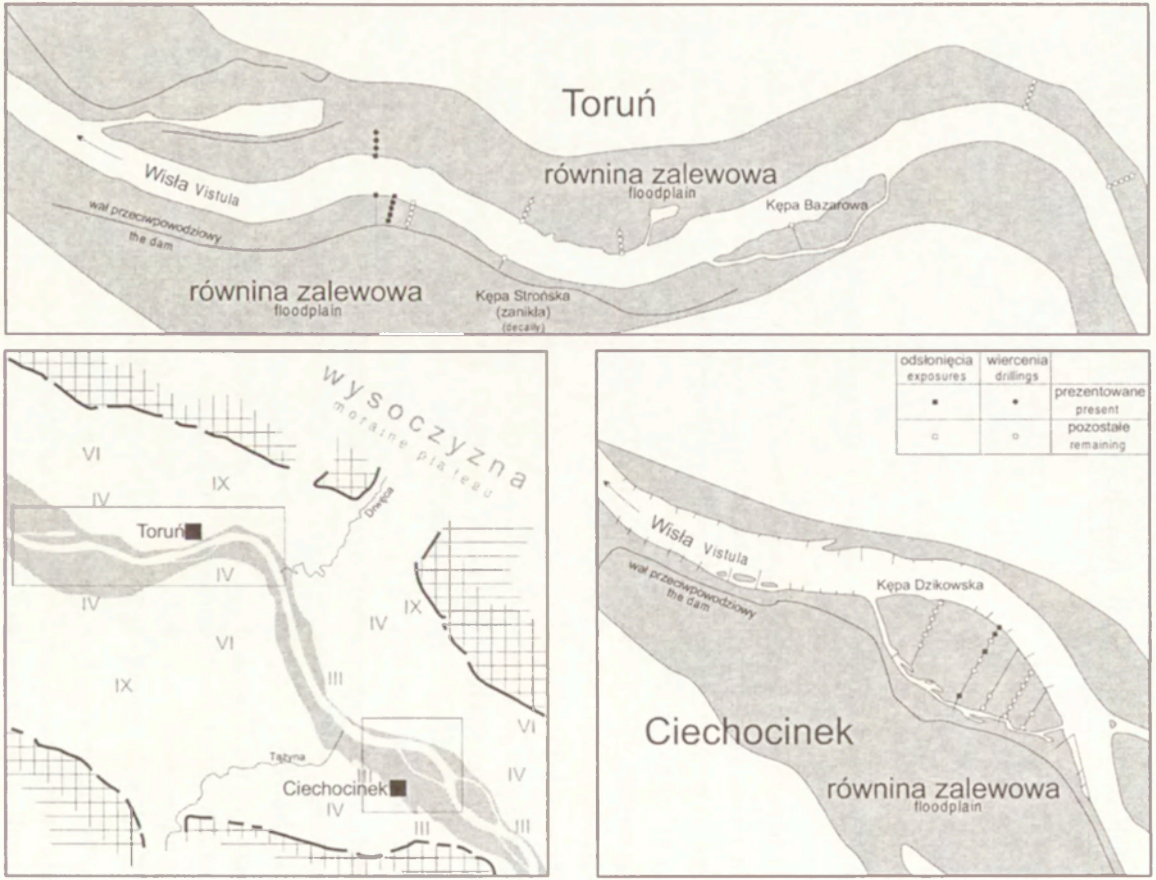


Fig. 1. Site sketch of the location of the researched exposures and drillings near Ciechocinek and Toruń

Szkic sytuacyjny lokalizacji analizowanych odsłonięć i wierceń w rejonie Ciechocinka i Torunia

1992; Makowski 1998). Next to its substantial theoretical significance, research into the Vistula's alluvia, which shape the river valley bed, also has practical value, as it enables the forecasting of the dynamics of fluvial processes in the face of more and more frequent floods, which turn out disastrous at times.

The research comprised parts of the bed of the lower Vistula valley near Ciechocinek and Toruń (Fig. 1). It mainly consisted in the comparative analysis of the lithofacies formation in the layer top sequences of the Vistula's alluvia in the area of three "channel islands" (Kępa Dzikowska, Kępa Bazarowa and Kępa Strońska), which are set solid and partially integrated in the structure the surrounding floodplain. The palaeohydrologic environment of the channel mesoforms of the islands, which had previously undergone dynamic transformations, undoubtedly differed in a large extent from the relatively stable surfaces of the Vistula's floodplains. This view of the issue may turn out to be significant to the correct interpretation of the sedimentological record to be found in the bed of the lower Vistula valley, understood very broadly. It is a known fact that the bed was shaped in the past by variable types of river channel systems, and mainly by braided and anastomosing (Andrzejewski 1994; Niewiarowski 1987; Tomczak 1987; Teisseyre 1992; Babiński 1992).

This paper stems from the recent morphological and sedimentological research into the bed of the Vistula valley near Toruń which was carried out by the authors, as well as from a few M.Sc. theses, supervised by L. Andrzejewski (Janusz 2000; Jarosz 2000; Szmidt 2001; Juśkiewicz 2002).

2. RESEARCH AREA AND METHODS

The issue of the lithofacies formation of the layer top parts of the Vistula's alluvia was undertaken in a relatively small fragment of the Vistula valley floor near Ciechocinek and Toruń. 22 exposures were researched, each 2.5 to 5 metres deep, and located in the islands subject to analysis and in the floodplain, mainly at the sites of contemporary undercuts in the Vistula channel. Also researched were 42 drillings of comparable depth. The location of the exposures and drillings is related to the 13 cross-sections of the valley bed (Fig. 1). An aggregate number of 853 deposit samples were collected for sedimentological analyses.

All the samples underwent a combined, i.e. laser-sieve, analysis of the texture, in which the samples were sifted through a set of sieves with mesh diameters distributed at 0.5 φ, and in the case of finer fractions, i.e. from

4.25 φ, the laser method was applied, with a Fritsch LPS grain size meter – Analysette 22, C version. The statistical analysis was based on the generally applied formulas by R.C. Folk, W.C. Ward (1957), where the extrapolation of the extreme segments of the grain-size distribution curve was related to the actual distributions. The results of the analysis were processed with the Jo.S.Ek. software, which made it possible to create a database, and to conduct a thorough statistical analysis combined with graphic presentation. In the researched exposures, vertical variability of the texture characteristics were presented, based on Miall's code of lithofacies (Miall 1978, 1980, 1996; Zielinski 1998), and on the mean diameter (GSS). When working out the log of lithofacies, the fraction classification was applied according to the (CUG) of the Polish Geological Institute. In the classification, 13 grades were distinguished, ranging from 2 mm to 0.01 mm. Essential data on the frequency and range of changes in the textural parameters of the vertical profile are contained in two curves showing all the changes in the tested characteristics in 0.5 m of the vertical profile, from the exposure. The number of changes in the textural features followed from the number of appropriate grades distinguished in the 0.5 m section of the profile, whereas the range of changes was indicated by the difference between the respective ranks of the extreme grades. Organising data in this manner seems to make possible objective determination and comparison of the texture formation and of the variation dynamics in alluvial sequences of 0.5 m. The figures (3, 4, 5) show both the vertical and the spatial locations of the researched exposures and drillings. An essential stage in the analysis was when the afore mentioned curves were placed on joint graphs, one for each islands analysed and one for the Vistula's floodplain near Torun (Fig. 6). The objective was to find differences or similarities in the vertical texture changes in alluvia in different fluvial sub-environments. The rank evaluation of the dynamics of the stream sedimentation environments for the samples collected in the analysed islands and the Vistula's floodplain was based on (Passega 1964; Passega, Byramjee 1969) CM chart (Fig. 7).

The lithofacies analysis of the roof sequences of the alluvia in the researched islands and floodplain was set against the background of the dynamic changes in the Vistula's bed between the beginning of the 19th century and the present (Fig. 2). The evolution of the islands was thus evidenced with respect to their spatial alterations, and the time was indicated when the clumps became stable and joined in the structure the floodplain, which occurred at the turn of the 19th century, as the Vistula channel improvement began.

3. THE PALAEOGEOGRAPHIC CHARACTERISTICS OF THE RESEARCHED FRAGMENT OF THE VISTULA VALLEY FLOOR

Like most rivers of the Polish Lowland, the Vistula reached at least the meadow terrace level as early as the beginning of late Vistulian period. This phenomenon resulted from the northward recession of the last ice-sheet as well as from the quickly established and low base-level of erosion and accumulation for the Vistula, the base-level first having been in the Baltic ice lake, and then in the pre-boreal yoldia sea. Therefore, the balance of erosion and accumulation processes in the lower Vistual valley during the recession of the last ice-sheet is definitely negative (Andrzejewski 1994). The issue is much more complex and disputable in terms of Holocene changes in the level of the Vistula's channel, as well as erosion and accumulation. E. Drozdowski (1982) is of the opinion that the formation of the floodplain in Grudziądz Basin had two stages, i.e. in early Holocene until the Atlantic period the valley bed was being filled with vast amounts of alluvia, after which vertical stabilisation of the Vistula's channel occurred. A similar view of the matter was offered by W. Niewiarowski (1987), who, in his research on Unisław Basin, found a marked increase in the thickness of alluvia, which began from late pre-boreal period, and which was interrupted with short erosion stages. In the past 3–4 thousand years the increase in the thickness of alluvia resulted primarily from the vertical growth of flood lithofacies. In contrast, A. Tomczak (1987), doing research on Toruń Basin, pointed to a deep erosion tendency which lasted at least until the boreal period, and which was followed by an accumulation stage, lasting up to our times. E. Wiśniewski (1976), in his analysis of the structure of the Vistula's floodplain near Ciechocinek, concludes that all through Holocene the Vistula actually showed deep erosion tendencies, which ended in the sub-atlantic accumulation stage.

The contemporary morphology of the floodplain shows significant diversification. R. Galon (1934) distinguished two levels there – the higher level, 3–5 metres in height, and the lower one, 2–3 metres above the level of the river. Whereas W. Niewiarowski (1987), doing research in Unisław Basin, divided the higher level into three grades, distinguished by the nature of their respective alluvial soil covers. More recent research (Babiński 1992; Kordowski 1997, 1999, 2001; Szmańda 2002) indicates that a yet lower level may be singled out in the area directly surrounding the Vistula's channel, whose creation and contemporary transformation are conditioned by anthropogenic factors.

When analysing tendencies in the fluvial processes in the bed of the lower Vistula valley, one ought also to consider the question of alluvia thickness in the long profile. Basing on the current findings it may be stated that there are differences in alluvia thickness between certain sections of the valley floor. E. Wiśniewski (1976) discovers a deep incision of the Vistula near Ciechocinek, where, at certain spots, the river's alluvial floor drops up to 17 metres with respect to the current channel floor. In the area of Toruń Basin the maximum alluvia thickness is estimated by A. Tomczak (1982, 1987) at about 15–16 m. Such estimates recently found confirmation in geological cross-sections made through the Vistula valley floor near Toruń, as part of the construction work on a temporary road bridge and on the motorway bridge near Brzoza and Grabowiec. The floodplain reaches the height of approx. 38–39 m a.s.l., while the floor of the Vistula's alluvia underlying the contemporary bed in Toruń drops to 23.5 m a.s.l. The thickness of the Vistula's alluvia in Unisław Basin was estimated by W. Niewiarowski (1987) at 8–20 mm which is equivalent to the thickness of alluvia in Grudziądz Basin (Drozdowski 1982). In summary of the question of alluvia thickness in the floor of the lower Vistula valley, in accordance with the river's course, the authors believe there is no substantiation for any significant variations in the river's flow level during Holocene, with the exception of its mouth section, i.e. the delta, as differences in alluvia thickness may be caused by local overdeepenings in the channel, whose morphology is mainly shaped during bankfull-channel stages. This fact was noted by several researchers (Rotnicki, Młynarczyk 1989; Babiński 1992; Andrzejewski 1994).

The question of the lower Vistula valley development, aside from its lithological formation, ought to be considered in the context of the evolution of the river's channel systems. The rather numerous works done to date suggest that in its bed formation period, i.e. the floodplain and the overflood terrace, or during the past 12–13 thousand years, the lower Vistula never reached the status of a mature meandering river, despite its local tendencies to meander in basin widenings. The complex development of the Vistula channel system followed from the polygenetic nature of the individual elements in the valley floor, and the resulting limitations in free bed development. As the extra-channel alluvia accumulated gradually, as their cohesiveness as well as the share of plants and biogenic deposits grew, the Vistula, once a braided multi-current river, transformed into an anastomosing one. As shown by (Smith 1987; Smith, Smith 1980; Brown 1987), among others, the evolution of braided rivers towards anastomosing is a result of several physiographic parameters, and in particular by local changes in the valley floor descent. However, this process was not synchronised along the whole profile of the lower Vistula channel, which makes the palaeohydrological reconstruction, and the interpretation of their sediment record, markedly more difficult. These are some of the factors justifying the need to do further research into the Vistula's alluvia and the vertical and spatial variability thereof.

4. STRUCTURAL AND TEXTURAL ANALYSIS OF THE ALLUVIA WITH A PALAEOHYDROLOGICAL INTERPRETATION

The lithofacies research only concerned the layer top (upper) sequences of the Vistula's alluvia near Ciechocinek and Toruń, 3 to 5 m in thickness. As stated before, the total thickness of channel and overbank alluvia which form the Vistula's floodplain in the bed of Kotlina Toruńsko-Bydgoska is not stable, and it varies from 3 m to 16 m, with the average of approx. 10–12 m. It is difficult to unambiguously define the borderline between channel and overbank alluvia deposits, and thus to determine the thickness of deposits formed during floods. The author's research in the mouth sections of the floors of lateral valleys showed that the thickness of channel alluvia in plains formed by meandering rivers is variable, with a tendency to grow from about 1.5 m to 4.0 m in central parts of valley floors (Andrzejewski 1994). The maximum thickness of the channel alluvia near Ciechocinek and Toruń is estimated at 4–5 m (Tomczak 1971; Myślińska 1980; Wrotek 1990; Szmańda 2000). An assumption can therefore be made that the analyses dealt with this very facies of the alluvia. In its vertical formation, the authors mentioned above generally point to two sequences: the upper one, or the silt-sand layer, 0.5 – 2.0 m thick; and the lower one, resting immediately on the channel alluvia, and consisting of dust and clay deposits, with the thickness of 1 to 2.5 m. The lithology of overbank deposits in the western part of Kotlina Toruńska was recently studied by J. Kordowski (2001), who, basing on an analysis of the textural features, showed their considerable diversification in terms of the spatial shares of individual lithofacies.

The currently available identification of the structural and textural features of the alluvia in the Vistula valley bed does not really solve the question of their spatial and vertical variability in relation to the channel system types, as changing over at least the past 12 thousand years. The problem is particularly essential in the case of the lower Vistula valley floor, which, as said above, developed differently in gap sections and in basin widenings. (Falkowski 1982; Tomczak 1971, 1982, 1987; Mycielska-Dowgiałło 1987; Niewiarowski 1987; Babiński 1992 and others).

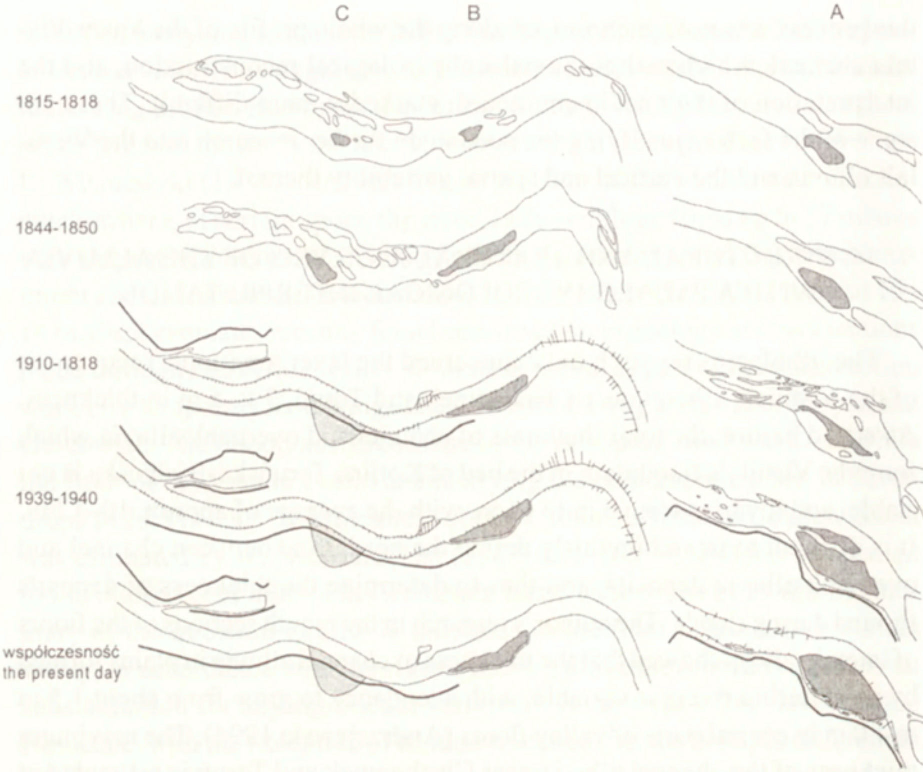


Fig. 2. Transformation dynamics of the Vistula's channel shape and its "channel islands". A – Kępa Dzikowska, B – Kępa Bazarowa, C – Kępa Strońska
Dynamika przekształceń zarysu koryta Wisły i mezoform korytowych.

In an analysis of the vertical alluvia formation in the researched "channels islands" and in the Vistula floodplain, significant differences were discovered. Generally, much greater changes of textural parameters can be found in the vertical profiles of the islands, in the structure of their roof sequences. Compared with the floodplain, the changes are more obvious in respect of both their frequency and range. When the vertical diversification of the alluvia in Kępa Dzikowska, Kępa Bazarowa and Kępa Strońska was analysed, down to 4–5 metres, a considerable texture variety was found between the gravel and clay fractions. These facts are a sign of rather frequent and extreme fluctuations in the energetic environment of the river while the deposits were formed, which is supported by a diagram of a rank evaluation of the current environments dynamics for the samples collected (Fig. 7A). A selected example may be the section of Kępa Dzikowska (Fig. 3), in which three exposures are presented. Exposure (I) shows the structure of the oldest and highest part of the island (Fig. 2). In the upper part, there are fine deposits of silts and silt sands, towards

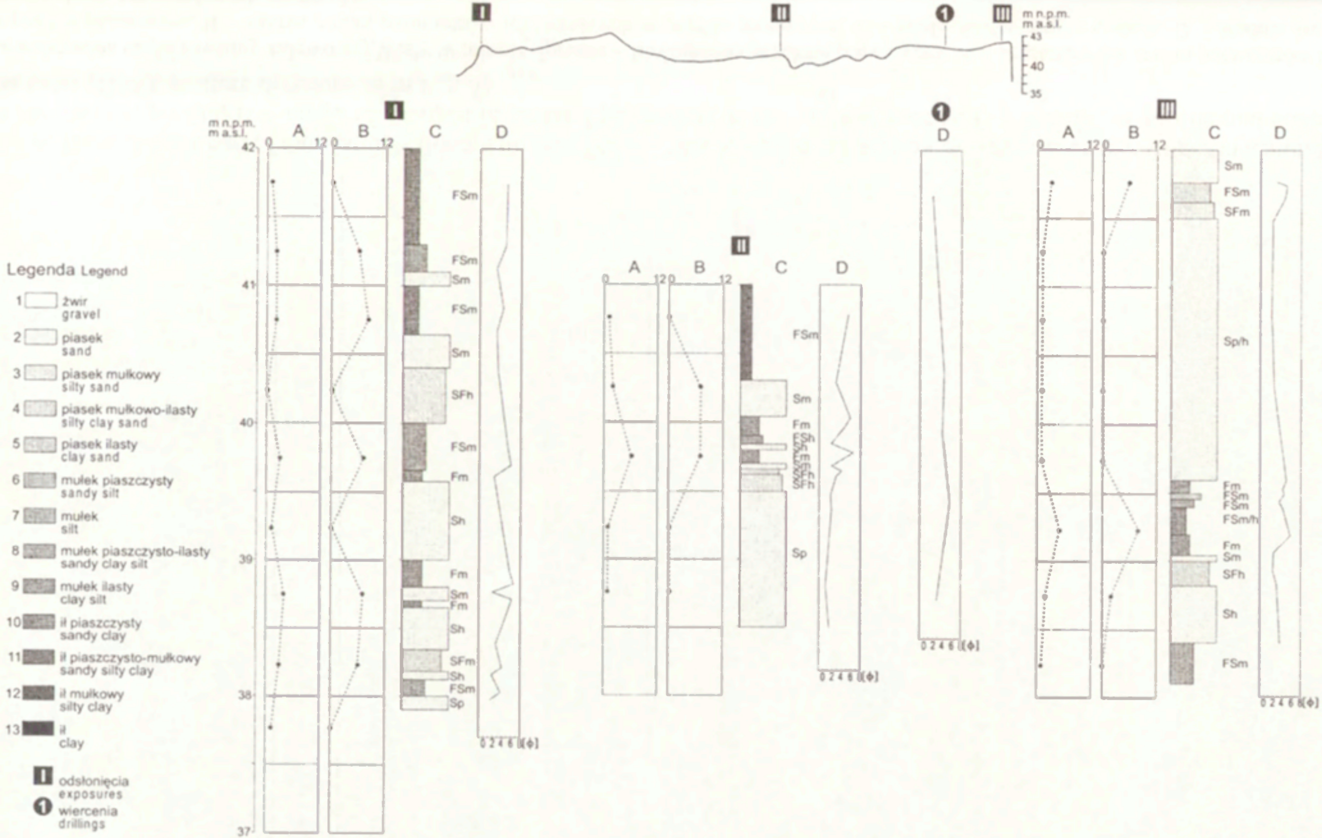


Fig. 3. Kępa Dzikowska – lithological cross-section: A – frequency of changes in textural parameters in the vertical profile, B – range of changes in textural parameters in the vertical profile, C – lithofacies profile and code, D – mean diameter (GSS)

Kępa Dzikowska – litologiczny przekrój poprzeczny. A – częstotliwość zmian parametrów teksturalnych w profilu pionowym, B – zakres zmian parametrów teksturalnych w profilu pionowym, C – profil litofacjalny z kodem, D – średnia średnica (GSS)

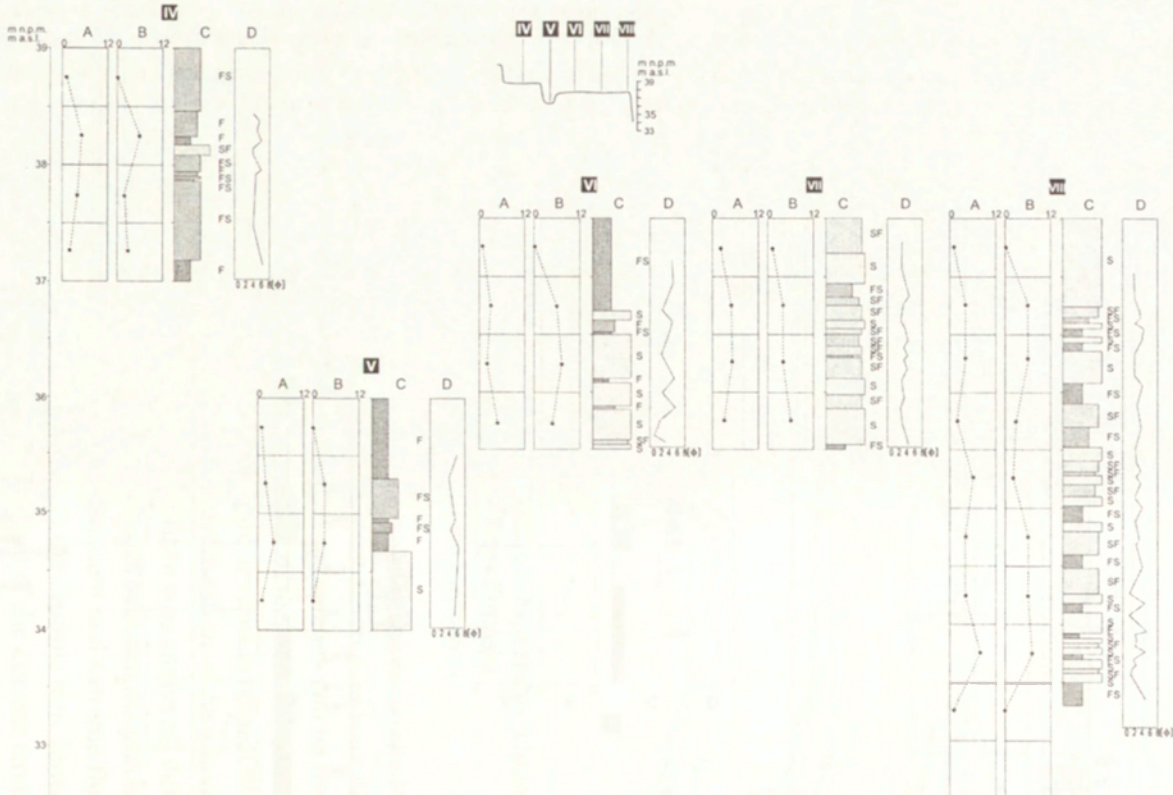


Fig. 4. The left-bank part of the Vistula's floodplain near Toruń – lithological cross-section. A – frequency of changes in textural parameters in the vertical profile, B – range of changes in textural parameters in the vertical profile, C – lithofacies profile and code, D – mean diameter (GSS); (texture divisions as in Fig. 3)

Lewobrzeżna część równiny zalewowej Wisły w rejonie Torunia – litologiczny przekrój poprzeczny. A – częstotliwość zmian parametrów teksturalnych w profilu pionowym, B – zakres zmian parametrów teksturalnych w profilu pionowym, C – profil litofacjalny z kodem, D – średnia średnica (GSS),

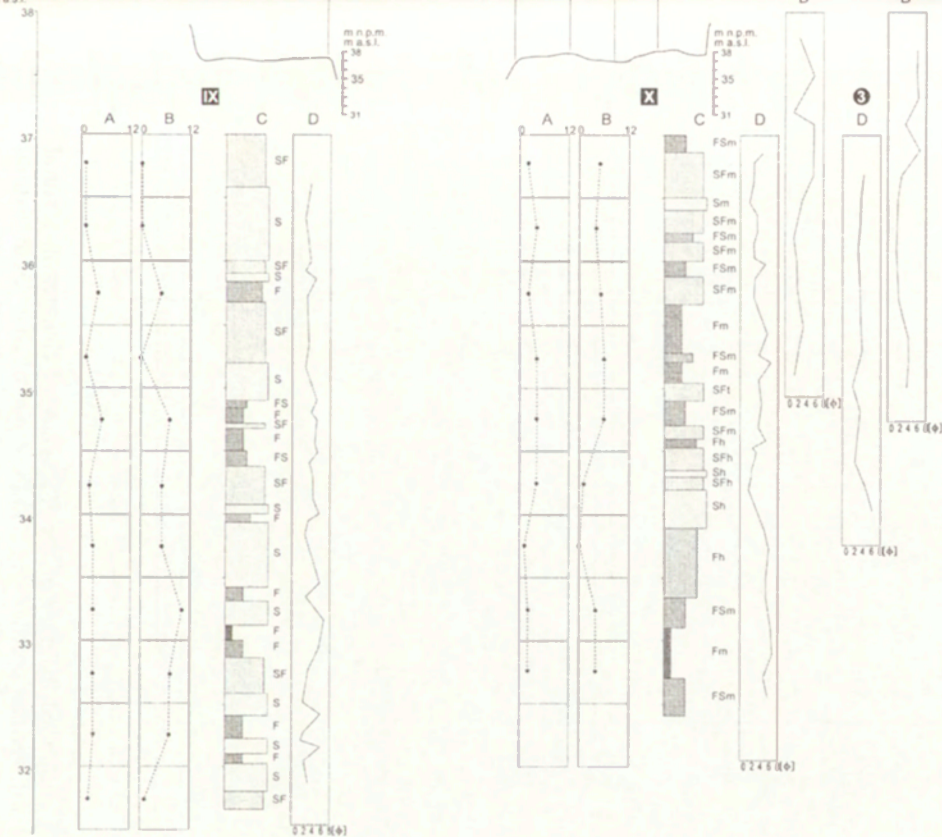


Fig. 5. The Vistula's floodplain near Toruń – lithological cross-section. A – frequency of changes in textural parameters in the vertical profile, B – range of changes in textural parameters in the vertical profile, C – lithofacies profile and code, D – mean diameter (GSS); (texture divisions as in Fig. 3)

Równina zalewowa Wisły w rejonie Torunia – litologiczny przekrój poprzeczny. A – częstotliwość zmian parametrów teksturalnych w profilu pionowym, B – zakres zmian parametrów teksturalnych w profilu pionowym, C – profil litofacjalny z kodem, D – średnia średnica (GSS), (wydzielenia teksturalne jak na rycinie 3)

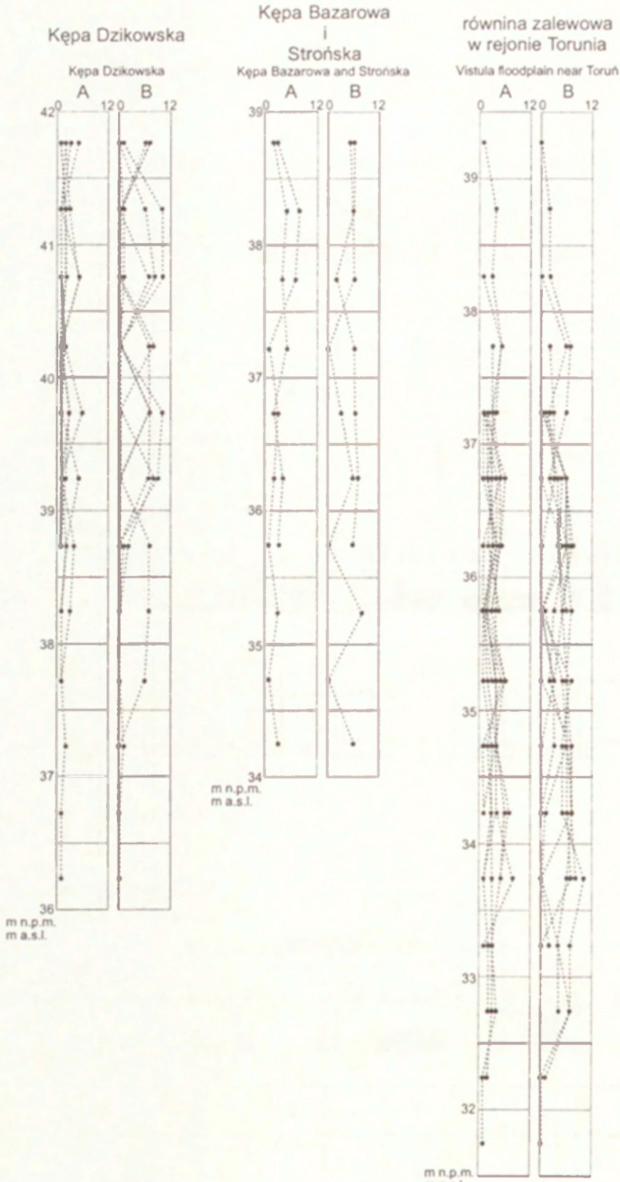
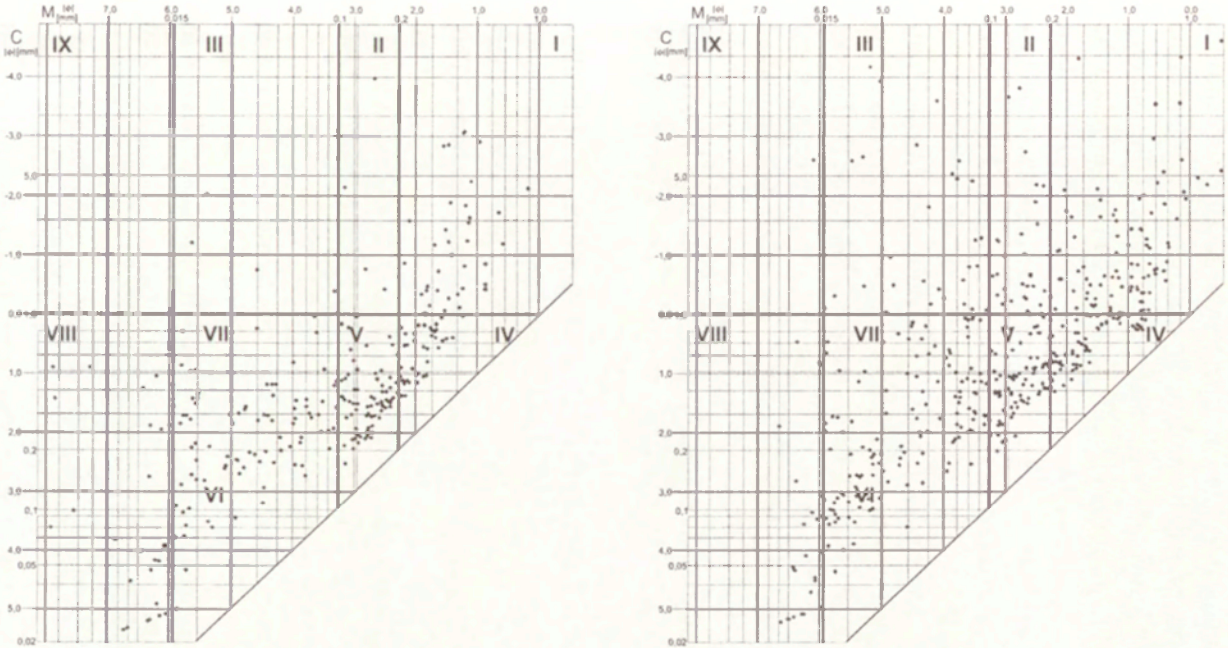


Fig. 6. A combined diagram of the frequency of changes in textural parameters in vertical profiles (A), and the range of changes in textural parameters in vertical profiles (B) in Kępa Dzikowska, Kępa Bazarowa and Kępa Strońska, and in the floodplain near Toruń

Wykres zbiorczy częstotliwości zmian parametrów teksturalnych w profilach pionowych (A) oraz zakres zmian parametrów teksturalnych w profilach pionowych (B) w obrębie Kępy Dzikowskiej, Bazarowej i Strońskiej oraz równiny zalewowej w rejonie Torunia



Kępy Dzikowska, Bazarowa i Strońska

Kępa Dzikowska, Bazarowa and Strońska

I - 13,43%, II - 1,41%, III - 1,77%, IX - 0,00%
IV - 14,49%, V - 26,86%, VI-VII - 31,45%, VIII - 10,59%

Równina zalewowa w rejonie Torunia

Vistula floodplain near Toruń

I - 20,78%, II - 5,88%, III - 6,86%, IX - 0,39%
IV - 15,69%, V - 16,86%, VI-VII - 27,06%, VIII - 6,48%

Fig. 7. Rank evaluation of the dynamics of stream sedimentation environments for all the samples collected, and split by the islands and floodplain

Rangowa ocena dynamiki prądowych środowisk sedymentacyjnych dla wszystkich pobranych próbek z podziałem dla kęp oraz równiny zalewowej the bottom turning into fine and medium sands, very frequently separated by thin layers of very fine deposits. A similar rhythm of textural changes can be observed in exposure (II), situated at the opposite rim of the same (age-wise) part of the island. A dramatically different structure was revealed in exposure (III), where a thicker series of massive and cross-layered sands is found at the layer top parts. Below the level of about 2.5 metres, sequences of fine and very fine alluvia begin, with multiple sand interbeddings. One ought relate the different vertical alluvia formation system in the youngest part of the island to the different deposit environment. The younger part of the researched island became incorporated in the older one in mid-20th century (Fig. 2). Earlier, that part belonged to the dynamic mid-channel bar, which appeared at the turn of the 19th century. The layer top parts is thus built with channel fractions representing the rather variable energetic environment of the river. This phenomenon also finds evidence in the other exposures in Kępa Dzikowska and in the Toruń islands. The analysis of the spatial changes in the structural and textural parameters of the alluvia in the clumps showed their marked variability even at short sections of several dozen metres. It is reflected in large amplitudes of the percentage shares of samples located in fields I, II, III and IX, i.e. characteristic of significant dynamic activity of deposit environments (Fig. 7).

The substantial vertical and spatial lithofacies diversification of the alluvia in the researched "channel islands" should therefore be interpreted as an effect of parts of early mid-channel bars, differing as to their age and energetic conditions, being incorporated in the structure of a single formation. This condition was characteristic of the period preceding the regulation of the Vistual's channel, which, in Toruń Basin, had features of a braided multi-current river, with an evident tendency to create more stable systems, typical of anastomosing rivers.

An analysis of the vertical overbank alluvia of the Vistula near Toruń and below Kępa Bazarowa and Kępa Strońska (Fig. 1) showed that the dynamics of textural changes to the alluvia in the floodplain is lesser in relation to the situation in the islands, as described above. On the one hand, it is reflected in a narrower range of changes in the textural parameters, and on the other – in a greater frequency of changes, especially at depths ranging from approx. 2.0 m to 4.5 m (Fig. 4, 5, 6B). In this view, the duality of the vertical diversification of the overbank alluvia becomes apparent. It is visible in increased shares of relatively thick sand and silt lithofacies in the layer top parts, down to the depth of about 2–2.5 m. The bottom sequences of the overbank alluvia, down to the depth of approx. 4.5 m, demonstrate a markedly greater frequency of textural changes, along with a simultaneous tendency for the range of those changes to decrease. Thus, the lower alluvia sequence mainly consists of finer deposits in the form of fine sands, as well as silts and clays.

In a sedimentological analysis of several dozen drillings and exposures it was also found that the share of the sand fraction in the whole structure of flood alluvia becomes reduced down the Vistula's course, in the section below the mouth of the Drwęca, i.e. where the river's channel takes its characteristic westward turn. It seems that the phenomenon can be explained by the fact that the river's energy is decreased as it suddenly changes its flow direction.

The Vistula's floodplain, as shaped in the past several thousand years, before the channel was regulated, was characterised by a relative dynamic balance compared with the "channel islands" under research. Its vertical lithofacies diversification, however, does not correspond with the classic model of a floodplain shaped by a meandering river (Allen 1965, 1970a, 1970b; Schumm 1977, 1981). Conversely, the effect is evident of the anthropogenic transformation of the fluvial system in the lower Vistula valley. It can be seen in the nature of the layer top parts sequences of the alluvia, and it is reflected in increased shares of more coarse fractions, formed in much thicker and homogeneous series.

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LITOFACJALNE ZRÓŻNICOWANIE ALUWIÓW W OBRĘBIE KĘPY DZIKOWSKIEJ, BAZAROWEJ I STROŃSKIEJ ORAZ RÓWNINY ZALEWOWEJ WISŁY W REJONIE TORUNIA

Streszczenie

W oparciu o szereg odsłonięć i kilkadziesiąt wierceń dokonano analizy litofacjalnej stropowych sekwencji aluwiów w obrębie trzech kęp (Kępy Dzikowskiej, Bazarowej i Strońskiej) oraz równiny zalewowej Wisły w rejonie Ciechocinka i Torunia. Główne zagadnienie dotyczyło pionowego i przestrzennego zróżnicowania parametrów teksturalnych i strukturalnych aluwiów. Analizę litofacjalną aluwiów ujęto na tle dynamicznych zmian koryta Wisły od początku XIX w do okresu współczesnego. Wykazano istotne różnice w sposobie wykształcenia aluwiów w obrębie utrwalonych i częściowo zespolonych z równiną zalewową kęp w stosunku do budowy otaczającej równiny zalewowej. Generalnie w budowie stropowych sekwencji kęp zauważyć można zdecydowanie większe zmiany parametrów teksturalnych wyrażające się częstością zmian w ujęciu pionowym oraz ich zakresem w stosunku do równiny zalewowej. Różnice te wynikają ze zdecydowanie odmiennych środowisk energetycznych, dynamicznie przekształcanych łach śródkorytowych, w stosunku do bardziej stabilnych warunków paleohydrologicznych panujących w trakcie kształtowania się osadów pozakorytowych na równinach zalewowych Wisły. Analiza wykształcenia pionowego aluwiów pozakorytowych Wisły w rejonie Torunia, powyżej i poniżej Kępy Bazarowei i Strońskiej, wykazała, że dynamika zmian teksturalnych aluwiów równiny zalewowej jest mniejsza w stosunku do sytuacji w obrębie kęp. Wyraża się to z jednej strony mniejszym zakresem zmian parametrów teksturalnych, a z drugiej większą częstotkliwością zmian, szczególnie przy głębokości od ok. 2,0 do 4,5 m. Czytelna staje się dwudzielność pionowego zróżnicowania aluwiów pozakorytowych, wyrażająca się w częściach stropowych do głębokości 2–2,5 m większym udziałem litofacji piaszczysto-mułkowych o stosunkowo dużej miąższości. W spągowych sekwencjach aluwiów pozakorytowych do głębokości ok. 4,5 m zarysowuje się zdecydowanie większa częstotliwość zmian teksturalnych przy jednoczesnej tendencji do zmniejszania się zakresu tych zmian. Dolną sekwencję aluwiów budują zatem głównie utwory drobniejsze wykształcone w postaci piasków drobnych, mułków i iłów.

Analiza sedymentologiczna kilkudziesięciu wierceń oraz odsłonięć pozwoliła także na stwierdzenie, że z biegiem Wisły na odcinku poniżej ujścia Drwęcy, a więc charakterystycznego skrętu koryta w kierunku zachodnim w całej strukturze aluwiów powodziowych stopniowo maleje udział frakcji piaszczystej. Zjawisko to można, jak się wydaje, wyjaśnić zmniejszeniem się energii rzeki w wyniku nagłej zmiany kierunku jej płynięcia.

Równina zalewowa Wisły kształtowana w okresie ostatnich kilku tysięcy lat, przed regulacją koryta, charakteryzowała się względną równowagą dynamiczną w stosunku do analizowanych kęp. Jej pionowe zróżnicowanie litologiczne nie odpowiada jednak klasycznemu modelowi dla równiny zalewowej kształtowanej przez rzekę meandrujacą (Allen 1965, 1970; Schumm 1977, 1981). Czytelny jest natomiast wpływ antropogenicznego przekształcenia systemu fluwialnego doliny

dolnej Wisły w postaci stropowych sekwencji aluwiów, wyrażający się zwiększeniem udziału frakcji grubszych wykształconych w zdecydowanie miąższych i jednorodnych seriach.

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HISTORICAL CHANGES OF THE VISTULA CHANNEL AND ITS REFLECTION IN THE FLOOD PLAIN BETWEEN JÓZEFÓW AND KAZIMIERZ DOLNY

1. INTRODUCTION

The river reach here studied, between Józefów and Kazimierz Dolny, of the length of about 50 km, is located in the Vistula gap through the southern Polish Uplands. The antecedent gap is situated on the Upper Cretaceous rocks dipping to the north – the gaizes, marls and chalks. The nature and the width of the Vistula gap vary, depending on the substratum resistance. The valley narrows even to 1.2 km, dissecting the hardest layers of the Uppermost Maastricht near the town of Kazimierz Dolny. On the other hand, the valley widens to 10–12 km on the outcrops of soft chalk in the area of the Chodelka Basin.

Until now, the results of the previous studies allowed us the establishment of the main stages of the valley evolution during the Pleistocene (Pożaryski 1953; Pożaryski et al. 1993, 1994) and the Holocene (Pożaryski 1955; Falkowski 1967, 1982; Pożaryski, Kalicki 1995). Some Late Quaternary alluvial series, differing as to their age, could be distinguished in the 30 m thick valley fill. At their top there are fluvial sediments associated with the Vistulian, forming the high IIIrd (11.0–18.5 m) and medium IInd (3–9 m) sandy terraces. In the Holocene the sandy floodplain I, covered with silty overbank deposits, developed (1.0–3.5 m). W. Pożaryski (1955) distinguished the old and the medium floodplain, covered with old overbank deposits and young overbank deposits, respectively, and the youngest sand-ovcrbank deposits floodplain.

However, the structure of the Vistula floodplain differs among the narrow and wide reaches. In the narrow, gap-type reaches, there are segments of the flood plain of various ages. Here, the palaeomeanders are preserved only sporadically. The limited width of the valley restrained free meandering, and so the river had likely a permanent tendency to anastomosing, as well as to hindered preservation of the older series of channel alluvia. Yet, the records of changes in the type and rate of sedimentation on the flood plain are found in the few preserved older fragments. The evidence of changes is provided by subsequent covers of overbank deposits, separated by the buried soils. In the younger inserts, on the other hand, the overbank deposits have been laid down directly on the channel deposits. In these sections the overbank deposits of the same age are therefore facially differentiated and occur at various levels. The structure of the flood plain in the wide sections, such as Chodelka Basin, is different. A number of inserts of alluvia of various ages, associated with the meandering river, occur here one beside another at the same level. Moreover, overbank deposits show here a facial differentiation, related to the floodplain morphology and to the distance from the active channel (Pożaryski, Kalicki 1995; Kalicki, Pożaryski 1996).

The authors referred to have taken no notice of the most recent evolution of the Vistula river valley. It was only Falkowski (1967, 1982), who, when studying the southern part of the gap, applied the genetic approach to the analysis of the floodplain structure and distinguished the fragments formed by the meandering and braided river. He stated that the rate of overbank sedimentation increased rapidly in the Young Holocene, simultaneously with the increase of human impact. W. Pożaryski (1955) distinguished two components of the overbank deposits. The old component, clay sediment, was deposited in the Subboreal, because the artifacts of the Lusatian culture had been found at its top on the fossil soil. The sandy intercalation between the two overbank components was deposited in the period of 2500–2000 BP. The young overbank component, the sandy-silty sediment, was deposited between the beginning of our era and 1000 AD. This component was covered with sands accumulated during the last millenium. A recent study showed, though, that the grain size composition of the overbank deposits is only an indicator of the facial differentiation and is mainly associated with the distance from the active channel (Pożaryski, Kalicki 1995; Kalicki, Pożaryski 1996). The anthropogenic factor is responsible for the river turning wild since the 15th century and for the change of the river pattern into the braided one in the 19th century (Falkowski 1982). Hovewer, lack of detailed documentation in the Falkowski's paper (except for the profile at Basonia) causes that these opinions should be considered as not fully evidenced, especially as the author has not applied any of the chronostratigraphic methods.

In this situation, analysis of old maps might turn out helpful in verification of the earlier ideas and in establishing the timing of changes in the river pattern. This analysis, supported by the geological profile, allows us tracing the reflections of these changes in the sedimentation pattern on the flood plain.

2. THE CARTOGRAPHIC PART

2.1. THE METHOD OF HISTORICAL ANALYSIS AND THE CARTOGRAPHIC SOURCES

The purpose of tracing the changes in the Vistula river channel during the recent centuries was served by collecting the historical data, and for the last 230 years the cartographic sources were made use of. Thus, eight map series were gathered, encompassing the entire area or the fragments of this area. The scales of all maps were photographically unified, and the cartographic method was applied to bring them all to the unified co-ordinate system and to identical projection. The *Military Topographical Map* issued in 1999 provided the basis for further work.

The oldest maps are charged with significant location errors, resulting from the lack of mathematical and geodesic extensive knowledge. These maps were based upon the few localities, for which the astronomical measurements of latitude and longitude existed. Initially, these maps were being elaborated without the use of precise instruments (e.g. lack of precise chronometers caused errors in longitude). In the maps elaborated later on a lot of imprecision was brought about by the insufficient density of the triangulation grid, or by the ignorance of the magnitude of the magnetic deviation. Until as late as the middle of the 19^{th} century each side of the river was charted separately.

In order to unify the source materials it was necessary to re-interpret them. This consisted in identification of the common points on the old and contemporary maps, followed by calculation and reconstruction of the location of other elements of the map. Numerous characteristic points indicated on old maps do not exist nowadays any more, and the biggest share of them ceased to exist exactly in the river valley. Alas, the older the documents, the less common points wc have, and, consequently, the precision and the level of detail of the processed map decrease. For the maps elaborated on the basis of (even most modest) triangulation grid, the spot heights served as the primary points of reference. Effort was made to verify the cartographic material gathered and to complement it with the information from other historical documents. The analysis was carried out starting with the contemporary maps, with gradual movement backward in time, towards the oldest maps (that is – in the direction opposite to the description provided in the present paper).

Initially, the reliability of the reconstructed course of the Vistula river was checked through the comparison of the detailed relief, geological or geomorphological maps. Yet, the ultimate verification of the results of historical analysis has to be carried out in the field with geological and geomorphological methods.

The following maps were made use of in the cartographic analysis:

1. Rizzi-Zannoni's "*Carte de la Pologne divisée par provinces et palatinats et subdivisée par districts*" of 1772, printed on the scale of 1:692,000. This map formed the background for analysis.

2. F. Czaykowski's "*Województwo sandomierskie na powiaty i parafie podzielone r. MDCCLXXXVI*" ["*The province of Sandomierz divided into counties and parishes, year MDCCLXXXVI*", in Polish], published in 1786, but charted earlier, on the scale of 1:185,000.

3. Maps of the atlas of the Crown Provinces of the Polish Commonwealth by Karol de Perthees, "elaborated in the last years of peace", published on the scale of 1:225,000. "*Mappa szczegulna woiewodztwa sandomierskiego z 1788–1791*" ["*Detailed map of the province of Sandomierz of 1788–1791*"] and of "*Mappa szczegulna wojewodztwa lubelskiego z 1786*" ["*Detailed map of the province of Lublin of 1786*"] were made use of.

4. "*Carte von West-Gallizien*", elaborated on the basis of a sparse and not too precise triangulation grid under the supervision of the colonel of the Austrian army, Mayer von Heldensfeld, in the years 1797–1803, on the scale of 1:28,800. After generalization, the map was published in Vienna in 1808 on the scale of 1:172,800. The mathematical foundations of the map neglect the curvature of the Earth, so that errors appear in the meridional orientation of the map sheets. This map contains detailed information on the location of islands, abandoned loops, degree of overgrowing of the valley, etc. These materials were used by L. Sawicki (1928), who also copied the hand-drawn maps from the Viennese archives, and presented in his work the examples of changes in the Vistula river channel during the 19th century.

5. The Map of the Head Quartermaster's Department of the Polish Army was much more precise and contained more details. The Bessel ellipsoid provided the basis for the grid calculation. Twisting of many fragments of the map is observed, resulting from the lack of knowledge of magnetic declination. The images of the southern and southeastern parts of the so-called Congress Kingdom originate from 1832. Table charting was carried out on the scale of 1:42,000, and the map was engraved on the scale of 1:126,000.

6. "*Novaia topograficheskaia karta zapadnoi Rosii*" ["*New topographic map of western Russia*", in Russian], called "two-verst" map, published on the scale of 1:84,000 at the turn of the 20th century. The cartographic projection of the charting was the polyhedral Muffling projection. The basis for the

instrumental charting was provided by the triangulation grid. Field charting was carried out on the scale of 1:21,000. On the basis of the "two-verst" map the Germans published in the years 1914–15 "*Karte des westlichen Russland*" on the scale of 1:100,000, which was made use of in the present study.

7. The battlefield map on the scale of 1:100,000, elaborated by the Military Geographical Institute. The source of reference was the polyhedral Muffling projection and the Bessel ellipsoid. This battlefield map is very detailed and precise, with an extensive description. The sheets of interest to our study were developed in 1937.

8. The *Military Topographical Map* of 1:100,000, published by the Topographical Service of the Polish Army, and elaborated by the State Geodesic-Cartographic Enterprise. The universal transversal Mercator projection and the ellipsoid, WGS–84, were assumed.

The analysis did not account for the map of G.D. Reyman from the middle of the 19^{th} century, since the Vistula river channel – the primary subject of considerations – was copied on this map from the earlier "*Carte von West-Gallizien*" of Mayer's.

3. THE CHANGES OF THE VISTULA CHANNEL DURING THE LAST CENTURIES

On the basis of the written historical sources it is possible to trace the old river channel only in rough approximation. In the middle Vistula valley the main river current constituted already in the Middle Ages the boundary of numerous political, administrative and property units. The invariability of these boundaries was more persistent than the river's current. When analyzing the administrative divisions, the parish grid, the boundaries of deaneries and dioceses, we can indirectly identify the course of Vistula in the period of Renaissance.

The maps of the Sandomierz and Lublin Provinces, dating from the second half of the 16^{th} century, elaborated for the purposes of the *Historical Atlas of Poland*, present the re-established boundaries of provinces, counties and parishes. Over centuries, the boundary between the two voivodships mentioned went on the stretch of the valley analyzed here almost exclusively along the winding main channel of Vistula. The exceptions were constituted by two small enclaves (corresponding to roughly 1–1.5 km of the channel length) around the river crossings between Kazimierz and Wojszyn, and between Solec and Kamień, which were subordinated, including the right of toll collection, to, respectively, Kazimierz and Solec. The location of crossings in these particular places is documented since the early Middle Ages, but the ford near Kazimierz started to be intensively utilized only from the 13th century onwards, when, after a change of riverbed, the crossing near to Świeciechów was abandoned (Wąsowiczówna 1957).

Over the stretch from Świeciechów almost down to Józefów, Vistula flowed along the right bank, then turned towards the East where it rebounded from the cliff in Józefów and formed a wide bend. The village of Wola Pawłowska was situated at that time directly on the riverside. Then, Vistula flowed along the eastern edge of the valley, washing away the cliff at Kaliszany, Piotrowin and Kamień. Further downstream the channel went some 1.5–2.5 km to the East of the contemporary channel. Thus, Kępa Gostecka, and the grounds of the villages of Braciejowice, Zakrzów and Jarnutowice were situated in the 16th century on the left-hand side of the river. The course of the next segment of Vistula was very similar to the contemporary one, the river, though, much more winding and forming over the stretch from Brzeście to Zastów numerous small bends (the grounds of the Janowiec Parish stretched somewhat farther to the East than the present channel of Vistula). On the analyzed segment of the Vistula valley, its flood plain as well as an important part of the higher terraces were deforested since the Middle Ages and then during the entire period when the river was used as the major transport route.

The detailed changes in the course of Vistula over the last 230 years can be reconstructed on the basis of the cartographic materials. There are as many as four temporal cross-sections concerning the first 100 years, when, owing to the coincidence of numerous historical events, wars and political changes, shifts occurred in the settlement system, land use in the valley, as well as curbing, and the abandonment of the river transport. During the same period many elementary catastrophes took place on the analyzed segment of the valley – the epidemics, hunger, and, first of all, catastrophic floods (the largest ones being indicated by boldface numbers), in the years 1749, 1774, 1783, 1813, 1838, 1839, 1840, 1844–45, 1850, 1852, 1854, 1855, 1867 and 1871 (Kurzyp 1989). The non-walled river frequently changed its channel, washed out shores, destroyed cultivated fields, roads, and settlements.

The cartographic sources allowed us the reconstruction of the course of the Vistula and some of its tributaries as early as already in the 1770s (the map is a compilation, several elaborates have been made use of, primarily, though, the map of the Sandomierz province by F. Czaykowski). The course of the river is simplified, which is the consequence of the scales of the original materials (Fig. 1). Within the gap near to Kazimierz, the Vistula flowed in one chan-

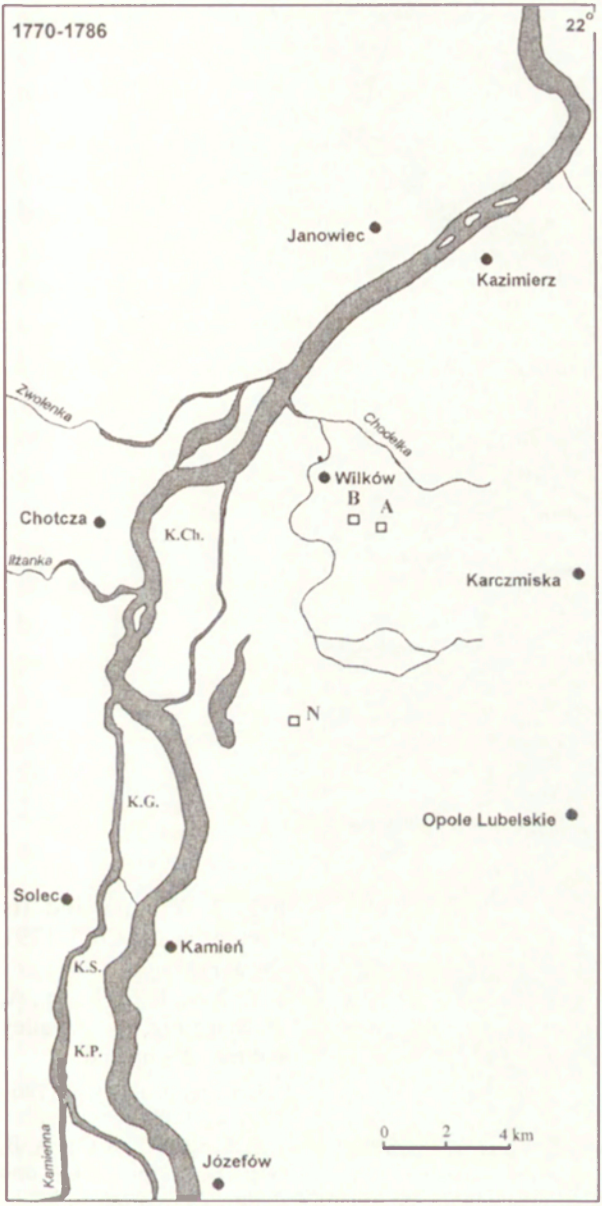

Fig. 1. Vistula and its tributaries in 1770–1786 (by J. Plit)

Wisła i jej dopływy w 1770–1786 (wg J. Plit)

K.P. – Kępa Piotrowińska; K.S. – Kępa Solecka; K.G. – Kępa Gostecka; K.Ch. – Kępa Chotecka; location of geological profiles: A – Szczekarków A; B – Szczekarków B; N – Niedźwiada

nel, the sand bars appearing in the channel, witnessing to the tendency of the river to turn wild. Upstream from this stretch Vistula flowed in two parallel branches, joining in a couple of places. The main channel was wider and had a slightly winding course, while the secondary channel, situated either on the left or on the right hand side, was over long stretches parallel to the main one. Attention is attracted by a large old riverbed area to the northeast of Zakrzew.

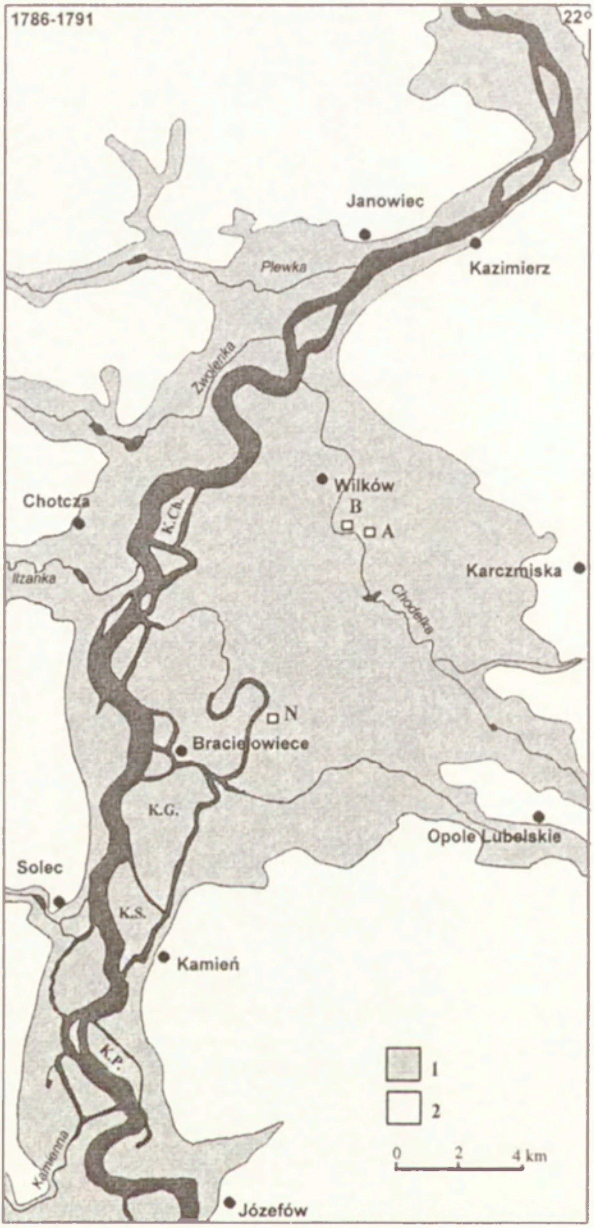

Fig. 2. Vistula and its tributaries in 1786–1791 (by J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N see Fig. 1; 1 – valley bottom; 2 – upland

Wisła i jej dopływy w 1786–1791(wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N patrz rycina 1; 1 – dno doliny; 2 – wyżyna

The outlet stretches of the Vistula tributaries, Zwolenka and Kamienna, were situated differently than today, but the course of these rivers was similar. The sources for the highly significant divergences as to the shape and course of Chodelka river with respect to the map of Czaykowski's should most probably be sought not in the actual change of the river channel, but in the lack of precise cartographic information. Yet, the channel of this river was located far (some 3 km) to the West of the palaeomeanders in Szczekarków, and close to them a smaller flow existed, which is no longer visible on the later maps.

The reconstruction of the course of the Vistula in the last years of the first Commonwealth is based upon the map of Karol de Perthees (Fig. 2). In the background of the map the reach of the bottoms of the Vistula valley and the valleys of its tributaries are shown. The main river channel is meandering, and the numerous developing small bends demonstrate the tendency towards the increase of the river winding. This tendency was not interrupted by the two large floods, which took place between making these two cartographic charts. The traces of these floods may, however, be identified by the numerous branches, joining both each other and the main channel, appearing mainly in the southern part of the study area. The arms of the river close to a straight course could be interpreted as the relief channels, while those of the winding course correspond most probably to the "inundated" older riverbeds. That is why the floodplain of the the Vistula has on this area the character of islands separated by numerous river arms. One of such activated old riverbeds is the system of palaeomeanders directly cutting into the site in Niedźwiada. In the Chodelka Basin the creek of Leonka flowed to the South of Opole Lubelskie and joined the Vistula at Braciejowice. Chodelka had a more winding course and flowed into the Vistula by Zastów. At the same time, its channel moved some 2 km to the East and the river was already roughly 1 km from the palaeomeanders in Szczekarków.

At the turn of the 19th century a turnaround in the tendencies of river course development took place, with the reasons of these changes apparently lying in the plexus of the natural and historical conditions. Very significant changes in the riverbed took place in spite of the absence of large floods in this period. The braiding of the Vistula can perhaps be seen through the increased linearity of its channel as compared to 10–12 years earlier, as well as through the numerous central bars, especially on the stretch Braciejowice-Chotcza (Fig. 3). The sole preserved well-developed winding existed a bit to the North of Józefów. Yet, even this bend has a broken meander neck. Downstream, important channel changes took place, since over this stretch the previous secondary arms became the main channel, while the traces of the old course are constituted by the sequences of oxbow lakes and narrow arms. The existing depressions were made use of also by the tributaries of the Vistula, like the mouth stretch of Kępianka. Owing to the shift of the main current, the Vistula flowed to the East of Kępa Piotrowińska, Kępa Solecka and Kępa Gostecka. In the Chodelka Basin, downstream of Braciejowice, the river split up

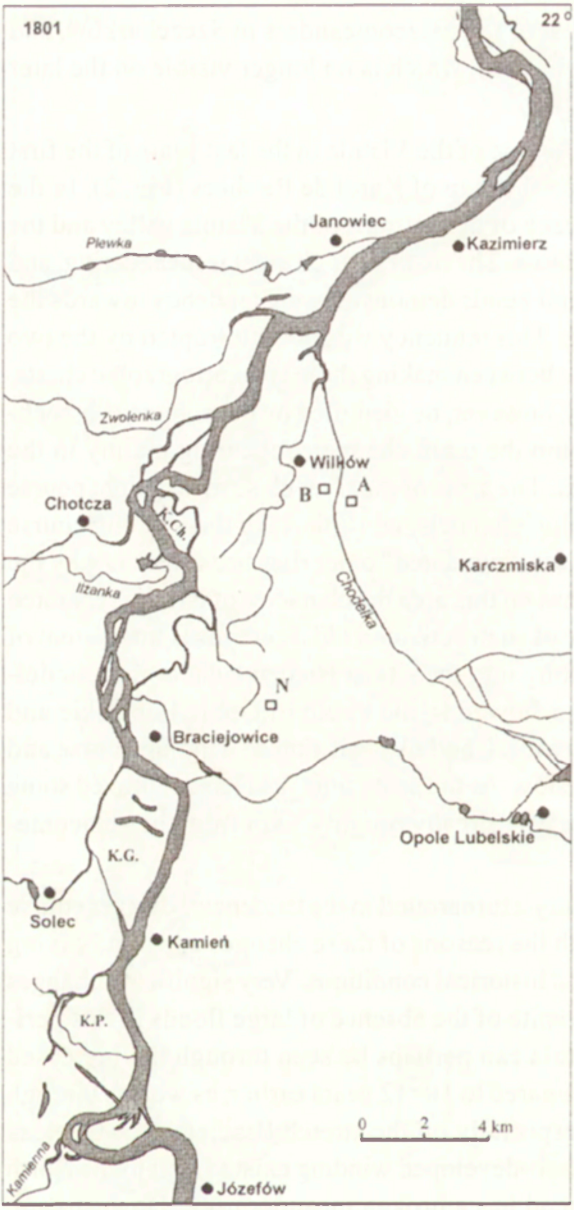

Fig. 3. Vistula and its tributaries in 1801 (by J.Plit)
K.P., K.S., K.G., K.Ch., A, B, N (see Fig. 1)
Wisła i jej dopływy w 1801 (wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (patrz rycina 1)

into arms, forming numerous islands (such as Kępa Chotecka). Yet, their spatial distribution was entirely different from the one documented on the map of Perthees. During the high discharges a valley got eroded between the villages of Głodno and Zastów, parallel to the main channel, made use of by a small flow (described on the map as Wiśliska). The palaeomeander system by Niedźwiada lost the direct contact with the channel of the Vistula. At the end

Fig. 4. Vistula and its tributaries in 1830-1840 (by J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (see Fig. 1)
Wisła i jej dopływy w 1830-1840 (wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (patrz rycina 1)

of the 18th century Chodelka was straightened and canalised, but its distance from the abandoned channels in Szczekarków has not undergone change until the inter-war period. Its downstream segment got connected with the newly developed side arm of the Vistula, and the mouth shifted towards the very escarpment.

In the years 1800–1840 four catastrophic floods occurred, bringing about significant changes in the valley bottom (Fig. 4). The Vistula was a multiple channel river, with numerous arms of different width and the winding or straight line course. On the segment Józefów-Kamień the Vistula flowed along the eastern edge of the valley. Kępa Gostecka was an island, but the main current concentrated in the eastern arm. Between Braciejowice and the mouth of Chodelka the river split into three-four parallel arms, joining together and splitting up again. Hence, the zone of active riverbed ranged between 0.9 and 3.5 km. Numerous short-lived bars appeared in the main channel, shown on the quartermastership map as devoid of vegetation. Side by side with the impermanent bars there were also the overgrown islands, which are indicated in Fig. 4. The floods activated again the ancient palacomeanders, like the sinusoidal system downstream of Braciejowice, appearing already on the map dating from the end of the 18th century (see Figs. 2 and 4). Yet, this system, washing away the palaeochannel in Niedźwiada, had not had a direct contact with the Vistula. Other palaeomenaders are well legible, as well, in the Chodelka Basin, probably owing to their inundation and stagnation of the flood waters in them. On the quartermastership map numerous sand bars and the series of boggy areas were designated. Likewise, the tributaries of the Vistula, such as, for instance, Kępianka, changed their channels, as demonstrated by the palaeomeander of Kępianka, cut out in the western escarpment of Kępa Solecka. The floods entailed as well significant changes in the land use of the floodplain. The villages of Czupel and Kempa were entirely destroyed, while Kludzie, Wynisłów (called later on Kolonia), Niwa and Gniadków – partly. The inhabitants moved most often their homes onto the edges of the uplands, felling the forests nearby for the new cultivated fields. Yet, new villages emerged also on the floodplain, like, e.g., Kępa Gostecka, located on the higher fragments of Kępa Gostecka area, built-up by the fen soils. The frequent upwellings caused a decrease of intensity and changes in land use of the valley bottom. Meadows and pastures dominated, and on some areas the process of natural vegetation succession has started. The floodplain forest entered the old, inactive arms of Vistula, like to the West of Kępa Piotrowińska (where Vistula flowed in the years 1786–91) and in the northern part of the depression within Kępa Gostecka (river channel of 1801).

During the subsequent 75 years the braiding process of the Vistula riverbed was in progress. That is why at the beginning of the 20th century Vistula was a typical braided river, with numerous, linear secondary arms of varying width (Fig. 5). The chains of abandoned loops existed along the Vistula river channel, activated only during high discharges and filled with sediments.

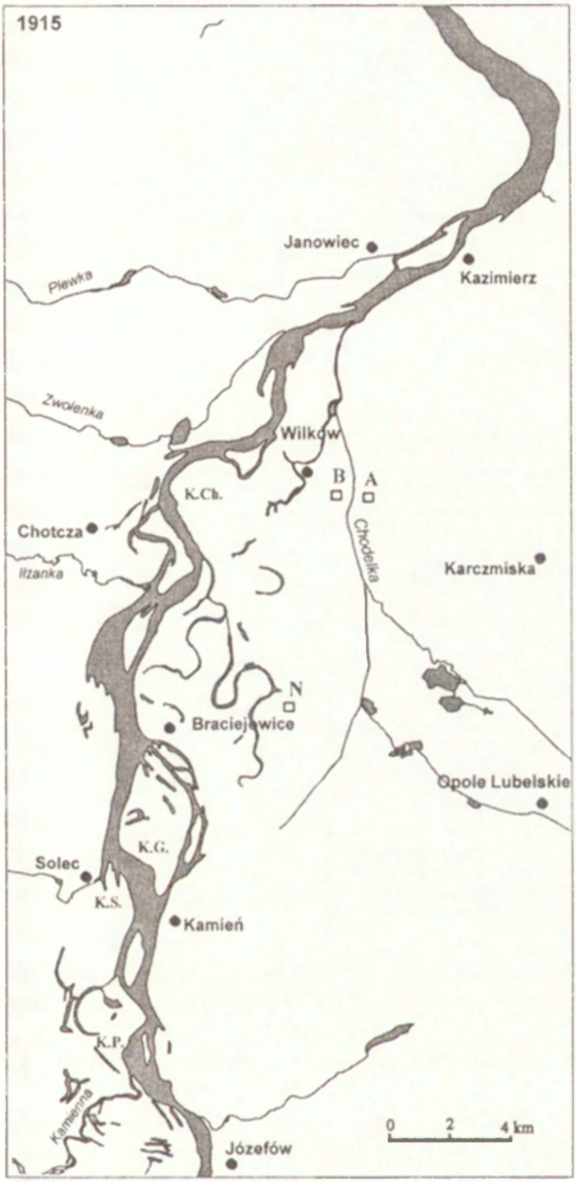

Fig. 5. Vistula and its tributaries in 1915 (by J. Plit) K.P., K.S., K.G., K.Ch., A, B, N (see Fig. 1)

Wisła i jej dopływy w 1915 (wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (patrz rycina 1)

The shape of the channel changed considerably depending upon the water level, and so on the map "*Karte des Westlichen Russlands*" the river banks are designated with the solid or broken line, and sometimes these banks were not shown at all. The figure 5 shows the simplified reach of the Vistula, solely with the reference to the solid and broken line boundaries. The width of the active zone ranged from 0.4 km to the South of Piotrowin to 4.5 km at Kępa Gostecka, most often, however, 1.0–1.5 km. Numerous overgrown islands and

Fig. 6. Vistula and its tributaries in 1937 (by J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (see Fig. 1)
Wisła i jej dopływy w 1937 (wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N (patrz rycina 1)

non-stabilized sand bars appeared within the river channel. In 1915 the Vistula flowed to the East of Kępa Piotrawińska. Kępa Gostecka was still an island, but the main current of the river flowed on its western side. In the Chodelka basin the palaeomeanders situated somewhat farther away from the Vistula were being slowly filled with the organic material and the silty sediments. Yet, the bigger floods made use of these old channels, somewhat changing their shapes, as this can be observed between Głodno, Barciejowice, Zakrzów and Majdan. The palaeomeander system in the area of Niedźwiada was clearly visible, but did not have a direct contact with the Vistula. Similarly as in the preceding period, there were also the changes in the land use of the valley bottom going on. The numerous floods in the second half of the 19th century caused the complete or partial destruction of roads and of such villages as: Gniadków, Ostrów, Kolonia, Przeprawa, Leśne Chałupy. Significant areas, acquired through deforestation and drainage works, were taken for fields and meadows. Still, in many depressions the chains of bogs and wetlands persisted. Local population, trying to defend themselves against the frequent floods, started to build the flood protection walls. These walls surrounded the selected localities, less frequently – the fragments of fields and orchards – which formed the systems of polders. Locally, construction of protection infrastructure was forced by the intensive lateral migration of the Vistula (some 500 m to the West), when the river started to directly threaten the village of Lucimia. Upper surfaces of the walls were often used for establishment of roads (Plit 2000).

The map of the Military Geographical Institute of 1937 documents the significant anthropogenic changes of the channel. The river got to an important degree straightened up and the majority of the parallel side arms were cut off (Fig. 6). The petrification of the channel course took place owing to the construction of groynes and flood protection walls. During the inter-war period the continuous system of flood protection structures was not yet created, but the majority of the segment length was already protected by them. Older walls were made higher. In many places the new structures were constructed, closer to the channel, thereby narrowing down the area of the active alluvial accumulation plain. On the map, the non-overgrown alluvia and bars occupy a distinctly smaller area than 30 years earlier, concentrating along the main channel between the walls. The narrowing of the active accumulation plain zone caused the decay of the palaeomeander systems, including the one that used to wash away the site in Niedźwiada. In addition, almost all wetlands of the valley bottom were drained, and so a number of the dried out old riverbeds disappeared from the map. Further, the floodplain forests were cut out, and their surface on the terraces was significantly limited. The settlement system entered even the area of high flood hazard.

Until the end of the 20th century the river got even more straightened up and had shallow, broad channel (Fig. 7). The active zone stretched only between the walls. In this period, uniquely during the extreme floods the walls get broken or water flows over them. In such cases old arms get activated, like,

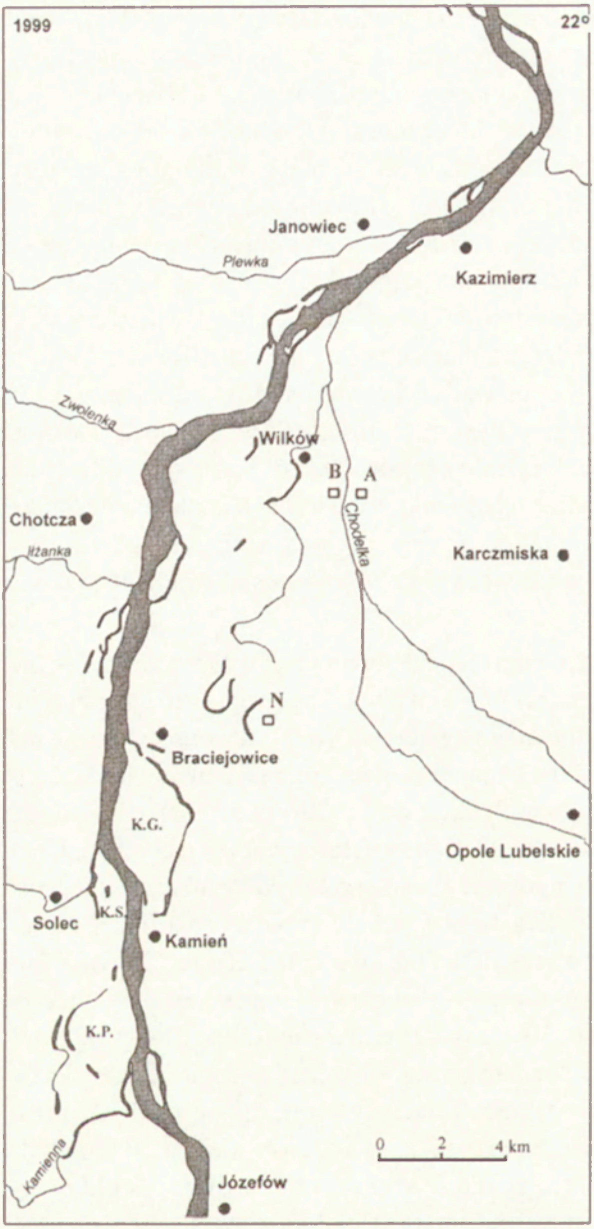

Fig. 7. Vistula and its affluents in 1999 (by J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N see Fig. 1

Wisła i jej dopływy w 1999 (wg J. Plit)
K.P., K.S., K.G., K.Ch., A, B, N patrz ryc. 1

for instance, the old riverbeds on the eastern side of Kępa Gostecka in 2001. The drainage work was continued after the World War II and numerous abandoned channels were filled in. This led to an almost complete draining of the palaeomeander system near Niedźwiada. The recent regulation works moved the channel of Chodelka into the area of palaeomenders in Szczekarków. The changes in the land use within the valley bottom took place as well. Horticulture, along with the hop and legume growing, developed (Plit 2000). The very limited human intervention on the area between the walls allows for the development of the natural vegetation succession – regeneration of the floodplain willow-and-alder forests and the willow brushwood. Due to planting, almost exclusively of the Scots pine, the area of forests on the sandy terraces and uplands increased as well.

4. THE GEOLOGICAL PART

The changes of the course and pattern of the Vistula River, reported before, must have been recorded in the geological profile of the valley bottom. Therefore, some profiles from the river segment studied, reflecting the most recent history of the valley, were analyzed. However, the chronostratigraphy of the youngest part of these geological profiles is the primary problem, because only few of these profiles were dated. The grain size analyses with the sieve method and the laser method with the help of Fritsch "Analysette-22" were performed. The grain size distribution parameters were calculated from cumulative curves according to the Folk-Ward formulae.

Young buried soils occurred in some profiles, which were located near to the limits (Ciszyca Przewozowa) or outside (Nieszawa, Basonia) of the cartographic study area. Therefore, our attempts were concentrated on paleomeander fills located on the flood plain called by W. Pożaryski (1955) "old overbank deposits flood plain". Both paleomeanders at Szczekarków and the paleochannel at Niedźwiada are the farthest and the nearest from the active bed of the Vistula river, respectively.

The paleomeander at Szczekarków A, with the radius of roughly 500 m and 250–300 m of width, was cut off during the Atlantic/Subboreal transition (4500±110 BP; Gd–6956). Its organic fill was covered with clayey silt (Mz=7,6–7,0φ), 0.5 m thick, with the coarser upward sequence. The bottom of the clastic member was dated at 200±100 BP (Gd–6957) cal. 1640–1880 AD (Fig. 8). The type of sediments demonstrated a stable environment of deposition in the oxbow lake and, thereafter, on the peat bog.

The paleomeander at Szczekarków B, with the radius of roughly 400 m and 250–300 m of width, did undercut from the West the paleomeander at Szczekarków A. However, its paleochannel fill is entirely different from the one of the paleomeander at Szczekarków A. High diversity of deposits were witnesses to a significant variety of the sedimentation environment. Three members could be distinguished (Fig. 9). Alternation of silts and silty sands in the lower part of profile (2 m) reflected the connection between the oxbow

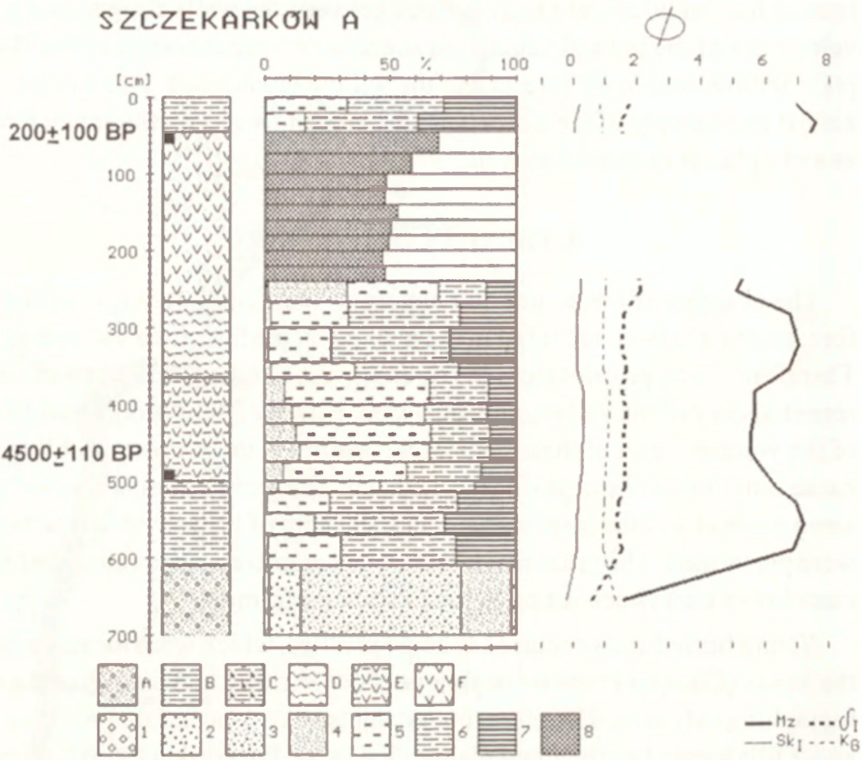

Fig. 8. Filling of palaeomeander Szczekarków A, grain size composition, content of organic matter and Folk-Ward's grain size distribution parameters (by T. Kalicki). Sediments: A – gravels with sands, B – clayey silts, C – organic clayey silts, D – gyttja with silts, E – peaty silts, F – peats. Fractions: 1 – coarse gravel (below -4φ), 2 – medium and fine gravel (-4 to -1φ), 3 – coarse sand (-1 to 1φ), 4 – medium sand (1 to 2φ), 5 – fine sand (2 to 4φ), 6 – coarse and medium dust (4 to 6φ), 7 – fine dust (6 to 8φ), 8 – clay (above 8φ), 9 – content of organic matter

Wypełnienie paleomeandra Szczekarków A, skład granulometryczny i wskaźniki Folka-Warda osadów, zawartość materii organicznej (oprac. T. Kalicki). Osady: A – piaski ze żwirami, B – mułki ilaste, C – mułki ilaste organiczne, D – mułki gytiowate, E – mułki torfiaste, F – torfy. Frakcje: 1 – grube żwiry (poniżej -4φ), 2 – średnie i drobne żwiry (-4 do -1φ), 3 – piasek gruby (-1 do 1φ), 4 – piasek średni (1 do 2φ), 5 – piasek drobny (2 do 4φ), 6 – pył gruby i średni (4 do 6φ), 7 – pył drobny (6 do 8φ), 8 – ił (powyżej 8φ); 9 – zawartość substancji organicznej

lake and the river. A layer of peaty silts from this member, roughly 1 m above the bot-tom, was dated at 4010±140 BP (Gd–9404). The upper 2 m of the fill is similar to the sediments of the Szczekarków A profile. There are peaty silts and peats with sandy silt intercalations dated post 3240±80 BP (Gd–10488). The unstable character of sedimentation could be connected with the vicini-

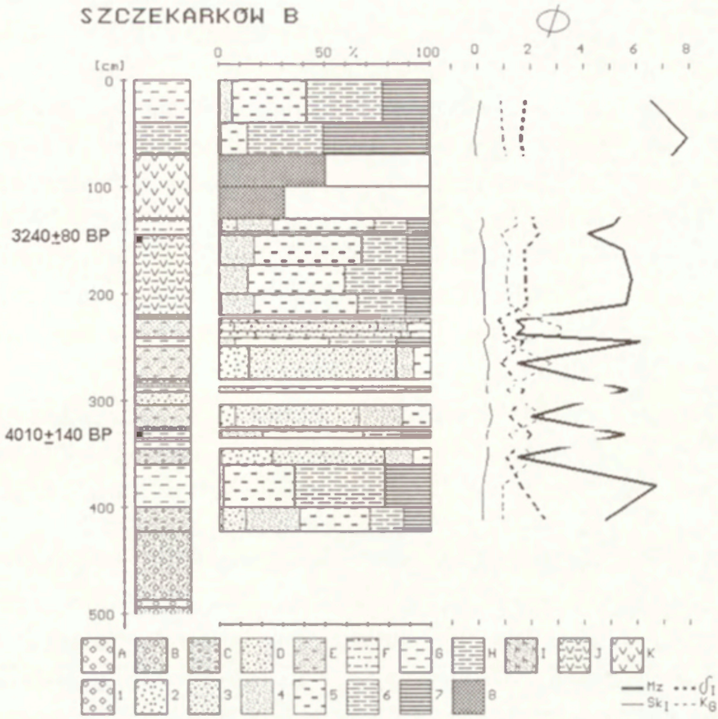

Fig. 9. Filling of palaeomeander Szczekarków B, grain size composition, content of organic matter and Folk-Ward's grain size distribution parameters (by T. Kalicki). Sediments: A – gravels, B – gravels with sands, C – clayey gravels with sands, D – sands, E – silty sands, F – sandy silts, G – silts, H – clayey silts, I – organic sands, J – peaty silts, K – peats. Fractions: 1 – medium and fine gravel (-4 to -1φ), 2 – coarse sand (-1 to 1φ), 3 – medium sand (1 to 2φ), 4 – fine sand (2 to 4φ), 5 – coarse and medium dust (4 to 6φ), 6 – fine dust (6 to 8φ), 7 – clay (above 8φ), 8 – content of organic matter

Wypełnienie paleomeandra Szczekarków B, skład granulometryczny i wskaźniki Folka-Warda osadów, zawartość materii organicznej (oprac. T. Kalicki). Osady: A – żwiry, B – piaski ze żwirami, C – piaski ze żwirami, zaglinione, D – piaski, E – piaski zaglinione, F – mułki piaszczyste, G – mułki, H – mulki ilaste, I – piaski organiczne, J – mułki torfiaste, K – torfy. Frakcje: 1 – średnie i drobne żwiry (-4 do -1φ), 2 – piasek gruby (-1 do 1φ), 3 – piasek średni (1 do 2φ), 4 – piasek drobny (2 do 4φ), 5 – pył gruby i średni (4 do 6φ), 6 – pył drobny (6 do 8φ), 7 – ił (powyżej 8φ); 8 – zawartość substancji organicznej

ty of the Vistula river at that time. The member of the organic sediments covered with clay (Mz=7,9φ), 70 cm thick, changes upward into the clayey silts (Mz=6,6φ).

The change of the sedimentation environment in both paleomeanders occurred just in the last centuries, when the peats were covered with overbank deposits (Pożaryski, Kalicki 1995). If the older changes of the sedimentation pattern were connected directly with the Vistula river, the last one (after the 2^{nd} half of the 17^{th} century) could be caused by the Chodelka river. As shown previously (Figs. 1–7), Chodelka flowed more than 2 km away from both sites at the end of the 18^{th} century, and then its channel moved to the vicinity of the paleomeanders. Chodelka river cut both paleomeanders after the water engineering works and the canalisation of the bed in the early 19^{th} century. Due to the damming of the tributaries by the Vistula river during floods, the clayey overbank deposits buried the peats in the paleomeanders and then, when Chodelka was closer, the silty-clayey overbank sediments were deposited in the late 18^{th} century.

The paleomeander at Niedźwiada (Fig. 10), with the radius of 500–600 m and 300 m of width, undercut the terrace IIa, with dunes (Pożaryski 1955). The point bars are some 2 m above and the paleochannel roughly 0.5 m below the river level. The paleomeander is located near to the Vistula river and is undercut from the West by the young system of paleomeanders, which had periodically a direct connection with the Vistula river. Some definite members can be distinguished in the palaeomeander fill, 4 m thick (Fig. 4). The silty sands, bad sorted, changing upward into sandy silts, overlie the sandy-gravel channel deposits. This member accumulated in the abandoned channel, with a significant connection with the river, recorded by gravels (up to 0.5 cm) at the top of this unit as the "lag layer". Upward, a member with a simple sequence occurred, the silts (Mz=6,1–7,2φ) changing into clayey silts with organic intercalations, reflecting the stable conditions of sedimentation. The bottom of silts and organic intercalation in the upper part of this member were dated at 2550±150 BP (Gd–9159) and at 2090±110 BP (Gd–10487), respectively. The third member recorded high variety of sedimentation pattern in the paleomeander. The layers of sand with gravels up to 1 cm began and ended this unit, and between these layers the peaty silts, incorporated by sands, with a layer of organic clayey silts (Mz=7,1φ), occurred. These data evidenced flood, which transported and accumulated coarser sediments in the gradually overgrown paleochannel during the inundations. These floods could be correlated with the flood phase in the Roman time or, likely, with the period when the vicinity of the active meander belt, currently the paleochannel system, cut the paleomeander at Niedźwiada from the West, causing flooding of the abandoned channel. Avulsion of the river led to the stabilisation of the sedimentation conditions in the paleomeander and to accumulation of the uppermost member since 1670±130 BP (Gd–9631). There are clayey silts, 2 m thick, upwards peaty silts, 0.70 m thick. The vicinity of the Vistula river and

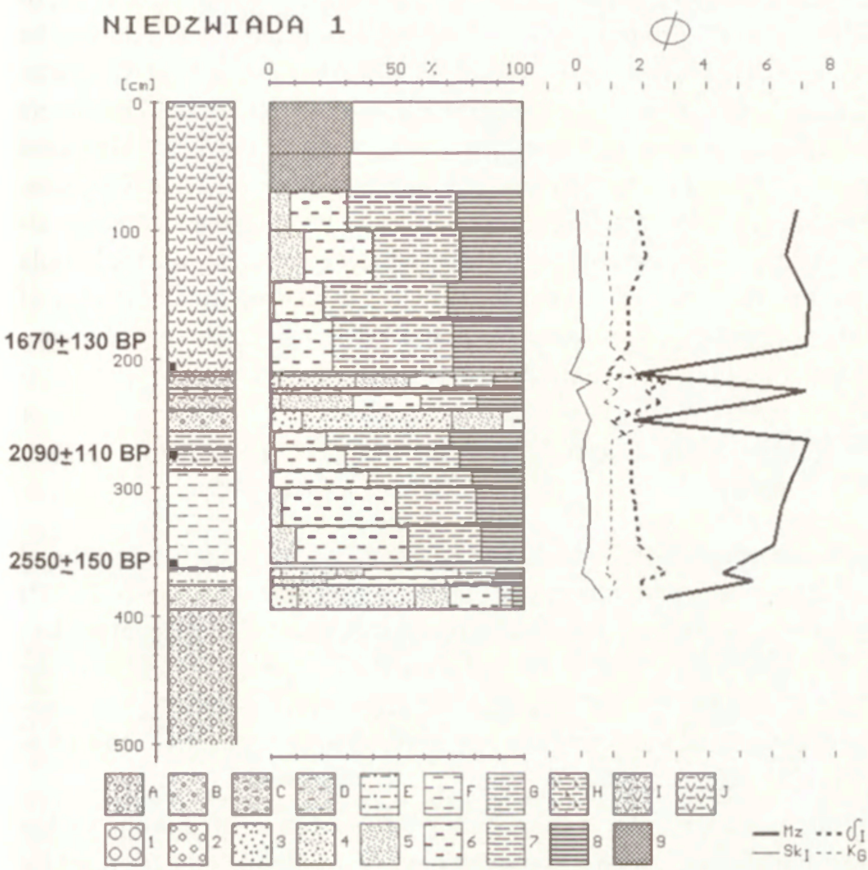

Fig. 10. Filling of palaeomeander Niedźwiada, grain size composition, content of organic matter and Folk-Ward's grain size distribution parameters (by T. Kalicki). Sediments: A – gravels with sands, B – sands with single gravels, C – clayey gravels with sands, D – silty sands, E – sandy silts, F – silts, G – clayey silts, H – organic clayey silts, I – peaty-silty sands, J – peaty silts. Fractions: 1 – coarse gravel (below -4φ), 2 – medium and fine gravel (-4 to -1φ), 3 – coarse sand (-1 to 1φ), 4 – medium sand (1 to 2φ), 5 – fine sand (2 to 4φ), 6 – coarse and medium dust (4 to 6φ), 7 – fine dust (6 to 8φ), 8 – clay (above 8φ), 9 – content of organic matter

Wypełnienie paleomcandra Niedźwiada, skład granulomctryczny i wskaźniki Folka-Warda osadów, zawartość materii organicznej (oprac. T. Kalicki). Osady: A – piaski ze żwirami, B – piaski z pojedynczymi żwirami, C – piaski ze żwirami, zaglinione, D – piaski zaglinione, E – mułki piaszczyste, F – mułki, G – mułki ilaste, H – mułki ilaste organiczne, I – piaski zaglinione, organiczne, J – mułki torfiaste. Frakcje: 1 – grube żwiry (poniżej -4φ), 2 – średnie i drobne żwiry (-4 do -1φ), 3 – piasek gruby (-1 do 1φ), 4 – piasek średni (1 do 2φ), 5 – piasek drobny (2 do 4φ), 6 – pył gruby i średni (4 do 6φ), 7 – pył drobny (6 do 8φ), 8 – ił (powyżej 8φ); 9 – zawartość substancji organicznej the low position of the paleochannel caused, respectively, high rate of deposition and possibility of organic accumulation in the paleomeander during the last centuries. However, the clay content in the organic sediments at Niedźwiada is much higher than in both paleomeanders at Szczekarków. The clastic sediments were transported from the Vistula river by the Early Medieval paleomeander system, which cut from the West the paleomeander at Niedźwiada. According to old maps (Figs. 1–7), this system was repeatedly reactivated and flooded in the last centuries, but it lost direct connection with the Vistula river at the turn of the 19th century. Consequently, a drop in the delivery of overbank deposits to the paleomeander took place and the content of organic matter in the uppermost part of the peaty silts rose up to 30%.

5. THE SUBBOREAL AND THE SUBATLANTIC HISTORY OF THE VALLEY

Fast sedimentation of silty-clayey overbank deposits (Mz=6,1–6,7φ) buried the soil on the older part of the flood plain (Ciszyca Przewozowa – 5170 BP) in the narrow reaches, whereas channel changes, reflected by paleomeanders at Szczekarków, cut off at 4500 BP and at 4010 BP in the wider reaches of the study area, occurred during the period of increased river activity, previously well documented and reported from the Vistula valley near Cracow (Kalicki 1991; Starkel et al. 1991).

In the Subboreal, the Lusatian culture sites were located in the valley bottom, though on the higher and older parts of the flood plain, accreted by the Atlantic and the Subboreal overbank deposits (Ciszyca Przewozowa), or on the parts adjacent to the slope of the valley (Basonia). The layer of the clastic deposits, found in the organic fill of the paleomeander at Szczekarków B (ca 3240 BP) correlate very well with the flood phase reported from the region near Cracow (Kalicki, Krąpiec 1991). The increase of human activity on the loess areas (Kruk 1988) caused formation of the alluvial fans at the outlets of erosion valleys and of the upland tributaries, which changed the type of sedimentation in the back-swamps near the valley scarps (Ciszyca Górna) (Pożaryski, Kalicki 1995).

At the beginning of the Subatlantic (2550 BP), some changes of the Vistula channel in the Chodelka Basin took place and the meander at Niedźwiada was cut off. Sandy intercalations, sometimes with gravels, in the fill of this paleo-meander, were accumulated in the Roman time (between 2090 and 1670 BP). The changes of the channel and of the sedimentation pattern are likely the traces of lateral migration and floods, which were very frequent in the upper Vistula drainage basin (Kalicki 1996; Kalicki, Krąpiec 1996). However, as we know due to the graves of the Cloche Grave Culture (3rd century BC), of Early Roman time (1st half of the 1st century AD) and of the Late Roman Przeworsk culture (2nd/3rd century AD), found on the valley bottom (Biernacki 1968), people settled the flood plain also during the phase of increase river activity.

The following phase of the intensified accumulation of the overbank deposits was the younger Medieval (700 BP), which resulted in the formation of the subsequent buried soil (Nieszawa, Basonia, Swieciechow), covered with silts (Mz=5,8-6,5φ) (Pozaryski, Kalicki 1995). Numerous sub-fossil oaks of this period, reported from the alluvia in the Sandomierz Basin (Kalicki, Krąpiec 1996), provide evidence for the increased river activity, not only in the gap section. Development of the age-distinct layers of soil in the overbank deposits of the Vistula flood plain is to the contrary of the Falkowski's (1982) hypothesis that the accumulation rate of overbank deposits increased in a stable and rapid manner since the Atlantic.

A common change in the sedimentation conditions was observed on the whole flood plain exactly in the last centuries (Pozaryski, Kalicki 1995). In the vicinity of the river there is a change in the grain size composition of the overbank deposits, silts being transformed into sandy silts (Mz=2,9-3,8φ) of levee (e.g. Parchatka). In the narrower stretches of the valley the youngest overbank deposits covered the whole bottom and overlaid the Medieval silty sediments without any buried soil, which has not developed yet. An increased frequency, and the likely important magnitude of floods resulted in fossilization of the youngest soil having developed on the clayey overbank deposits (Ciszyca Przewozowa) in the uppermost fragments of the flood plain. This soil has been covered with sandy silts (Mz=5,4φ). In the wider fragments of the valley the youngest overbank deposits of the levee facies were deposited along the Vistula river whereas clayey silts (Mz=7,0φ), similar to the Atlantic overbank sediments with respect to the grain size, were accumulated on the organic deposits of the peat bogs (Lucimia) and of the paleomeanders (Szczekarkow A), which were situated at a far distance from the channel and in the low-lying areas (to 1.5 m above the river level). All these changes were probably caused by the increasing anthropogenic modification of the Vistula's drainage basin. At first, the river channel showed a tendency to turn wild (from the 15th century onwards), and then, from the 19th century, Vistula river changed its pattern to the braided one (Falkowski 1982). The changes of the river channel arc very well documented by the old maps from the last 230 years. During this period the zone of active alluvial plain of the Vistula river was relatively narrow (2–4 km wide) and in many cases it corresponds well to the youngest part of the flood plain, called by W. Pożaryski (1955) "silty-sandy terrace". However, according to the cartographic and geological data, older fragments of the flood plain occurred also in this belt. According to the buried soil sequences in the overbank deposits (e.g. Ciszyca Przewozowa), these rests of the older flood plain ("kępy") persisted in this zone from the Mezoholocene. Since the end of the 18th century the channel of Vistula has been getting constantly shorter, and the biggest changes occurred in the period between the World Wars (Table).

Map from the year	1791	1840	1915	1937	1999
Length of the main channel (km)	63	58.5	60.5	55.5	54.5

The primary reason of the changes in the Vistula river and its tributaries in the area of Chodelka Basin was the anthropogenic impact, and especially construction of embankments along the channels. These works caused the change in the Vistula river pattern from the multi-channel (until the end of the 19th century) to single-channel. Episodic sand bars occur in the riverbed only during the period with lowest discharges.

6. CONCLUSIONS

The cartographic-geomorphological method has an important research potential, which until now has been taken advantage of to an only very limited degree. The cartographic analysis of the bottom of Vistula river valley allowed for the establishment of the reach of the youngest fragments of the floodplain. This enables the determination of the portions of the valley bottom that are older than the last centuries, the ones, within which we can search for the traces of the more ancient history of the valley bottom.

The determination of the changes in the river course over the last centuries on the lessre cognized areas may constitute the first step in the study of the structure of the floodplain. In the valley of Vistula, whose Holocene history has been well identified over several stretches, this constitutes a significant complement, allowing for the tracing of the most recent stages, frequently neglected or treated marginally by the geomorphologists. It enables, as well, the verification of opinions that were based upon the geological data. Simultaneously, this type of study constitutes the bridge and fills the gap between the palaeogeographic research and the study of the contemporary morphogenetic processes.

The comprehensive cartographic-geomorphological analysis allows for the grasping and correct interpretation of the records found in the geological profiles at various sedimentation environments of the valley bottom. By applying the geographical actualism we can transfer this situation backwards into the older periods, for which we do not dispose any more of the cartographic image.

The documentation of the channel changes, and of the settlement and historical changes in the river valley, may constitute the foundation for the further physical-geographic and historical-economic studies.

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HISTORYCZNE ZMIANY KORYTA WISŁY I ICH ODBICIE NA RÓWNINIE ZALEWOWEJ POMIĘDZY JÓZEFOWEM A KAZIMIERZEM DOLNYM

Streszczenie

Przebieg i szerokość przełomu Wisły przez wyżyny jest zmienna (1,2–12,0 km) i uwarunkowana odpornością podłoża. Aluwia wypełniają głęboką na 30 m dolinę, która powstała w interglacjale małopolskim (Pożaryski 1953; Pożaryski i in 1993, 1994). Pozornie jednolita równina zalewowa jest różnowiekowa, a jej skomplikowaną budowę maskuje miąższa pokrywa mad (Pożaryski 1955; Pożaryski, Kalicki 1995; Kalicki, Pożaryski 1996). W wąskich odcinkach, gdzie rzeka nie mogła swobodnie

meandrować, na najstarszych fragmentach równiny zalewowej znajdujemy zapis zmian szybkości pionowej akrecji osadów pozakorytowych. Kolejne pokrywy madowe rozdzielają gleby kopalne. W rozszerzeniu Kotliny Chodelskiej na równinie zalewowej występuje obok siebie, na jednym poziomie morfologicznym, szereg różnowiekowych włożeń aluwiów związanych ze swobodnie meandrującą rzeką.

Celem zweryfikowania dotychczasowych poglądów na temat najmłodszej historii doliny i ustalenia czasu zmiany rozwinięcia rzeki (Pożaryski 1955; Falkowski 1967, 1982) przeanalizowano dokumenty historyczne oraz kartograficzne od końca XVIII wieku w powiązaniu z profilami geologicznymi, w których zapisane są subatlantyckie zdarzenia. W badaniach kartograficznych wykorzystano 7 horyzontów czasowych: 1770–1786 (mapa Czaykowskiego, Rizzi-Zannoniego), 1786–1791 (mapa Pertheesa), 1801 (mapa Heldensfelda), 1830–1840 (mapa kwatermistrzostwa), 1915 (mapa "dwuwiorstówka"), 1937 (mapa WIG), 1999 (Wojskowa Mapa Topograficzna). W analizie geologicznej wykorzystano trzy profile osadów wypełniających paleomeandry położone na starszych fragmentach równiny zalewowej i zlokalizowane w różnej odległości od aktywnego koryta Wisły, w jego pobliżu (Niedźwiada) oraz w znacznej odległości, pod krawędzią terasy (Szczekarków A i B).

W oparciu o historyczne źródła pisane można odtworzyć (w ogólnych zarysach) przebieg koryta Wisły z drugiej połowy XVI w., gdyż już od średniowiecza rzeka stanowiła granicę wielu jednostek politycznych, administracyjnych oraz własnościowych.

Szczegółowe zmiany biegu Wisły w ciągu ostatnich 230 lat można odtworzyć w oparciu o materiały kartograficzne. W siedemdziesiątych latach XVIII wieku, w przełomie koło Kazimierza, Wisła płynęła jednym korytem, a w jego obrębie występowały łachy, świadczące o tendencji do dziczenia. Powyżej tego odcinka, Wisła płynęła dwoma równoległymi ramionami łączącymi się ze sobą w paru miejscach (ryc.1). U schyłku pierwszej Rzeczypospolitej główne koryto rzeki ma charakter meandrowy, a liczne rozwijające się niewielkie zakola wskazują na tendencję do zwiększania krętości rzeki. Tendencji tej nie zakłóciły dwie duże powodzie, które wystąpiły pomiędzy tymi dwoma zdjęciami kartograficznymi (ryc. 2). Na przełomie XVIII i XIX w. nastąpiło odwrócenie tendencji w rozwoju rozwinięcia rzeki, a przyczyny tych zmian należy upatrywać w splocie warunków przyrodniczych i historycznych. Bardzo duże zmiany koryta wystąpiły pomimo braku dużych powodzi w tym okresie. Na roztokowanie Wisły może wskazywać większa prostolinijność jej koryta niż 10–12 lat wcześniej oraz liczne łachy śródkorytowe, szczególnie na odcinku Braciejowice-Chotcza (ryc. 3). W latach 1800–1840 wystąpiły cztery duże powodzie, które doprowadziły do dużych zmian w dnie doliny. Wisła miała charakter rzeki wielokorytowej, z licznymi ramionami o różnej szerokości i przebiegu zakolowym lub prostolinijnym. Strefa aktywnego łożyska rzeki wahała się od 0,9 do 3,5 km, a w głównym korycie występowały liczne krótkookresowe łachy i zarośnięte wyspy. Powodzie ponownie uruchomiły niektóre stare paleomeandry, a inne uległy podtopienia. Całkowitemu zniszczeniu uległo kilka wsi. Nastąpiły też zmiany w użytkowaniu dna doliny. Spadła intensywność jego użytkowania, zaczęły dominować łąki i pastwiska, a na niektórych obszarach rozpoczął się proces naturalnej sukcesji roślinności (ryc. 4). W okresie następnych 75 lat postępował proces roztokowania koryta i na początku XX wieku Wisła była rzeką typowo roztokową z licznymi, prostolinijnymi, drugorzędnymi ramionami o różnej szerokości. Liczne powodzie w drugiej połowie XIX wieku spowodowały całkowite lub częściowe zniszczenie dróg i wsi. Znaczne obszary zostały zmeliorowane i odlesione. Rozpoczęto budowę wałów przeciwpowodziowych (ryc. 5). W okresie międzywojennym nastąpiły duże antropogeniczne zmiany koryta. Rzeka została w znacznym stopniu wyprostowana i odcięto większość bocznych równoległych ramion. Nie stworzono jeszcze ciągłego systemu umocnień przeciwpowodziowych, ale większość odcinka była już nimi zabezpieczona. Dno doliny zostało zmeliorowane i odlesione (ryc. 6). Do końca XX wieku rzeka została niemal całkowicie wyprostowana i ma płytkie, szerokie koryto. Aktywną strefą jest jedynie obszar międzywala, gdzie rozwija się naturalna sukcesja roślinności (ryc. 7).

Przedstawione powyżej zmiany przebiegu i rozwinięcia koryta Wisły zostały częściowo zapisane w profilach geologicznych dna doliny. W subborealnych paleomeandrach w Szczekarkowie A i B ich organiczne wypełnienia zostały przykryte ponad półmetrową warstwą mułków ilastych (Mz=7,6-7,9φ) przechodzących ku górze w mułki pylasto-ilaste (Mz=6,6-7,0φ), których spąg był datowany na 200±100 BP (Gd-6957) tj. 1680, 1760 lub 1800 cal. AD (ryc. 8 i 9). O ile starsze zmiany warunków sedymentacji w tych starorzeczach były związane bezpośrednio z korytem Wisły, o tyle ta ostatnia wywołana została prawdopodobnie przez Chodelkę. Z końcem XVIII w. Chodelka płynęła jeszcze ponad 2 km na zachód od Szczekarkowa, a później jej koryto przemieściło się w pobliże paleomeandrów, by w końcu XX w. przecinać obydwa starorzecza. Podpieranie dopływów przez Wisłę w okresie wezbrań spowodowało, że na torfach akumulowały mady, które stały się bardziej pylaste po zbliżeniu się Chodelki do obu paleomeandrów. Subatlantycki paleomeander w Niedźwiadzie położony jest znacznie bliżej aktywnego koryta Wisły i jego wypełnienie wskazuje na dużą szybkość i zmienność warunków sedymentacji. Najwyższe ogniwo wypełnienia, akumulowane po 1670 BP, tworzą mułki organiczne i torfiaste (ryc. 10). Stosunkowo niewielka odległość od Wisły oraz sąsiedztwo prawdopodobnie wczesnośredniowiecznego systemu paleomeandrów, który był wielokrotnie uaktywniany i podtapiany w ostatnich stuleciach (ryc. 1–7), spowodowała dużą dostawę materiału klastycznego. W efekcie osady organiczne w Niedźwiadzie są znacznie bardziej zailone od torfów ze starorzeczy w Szczekarkowie, a ilość substancji organicznej wzrosła w nich do 30% dopiero po tym jak system paleomeandrów utracił bezpośredni kontakt z korytem Wisły na przełomie XVIII i XIX w.

W metodzie kartograficzno-geomorfologicznej tkwi duży potencjał badawczy wykorzystywany dotąd w minimalnym stopniu. Analiza dna doliny Wisły pozwoliła ustalić zasięg najmłodszych fragmentów równiny zalewowej i umożliwia wytypowanie fragmentów, na których możemy szukać zapisu starszej historii dna doliny. Ustalenie zmian biegu rzeki w ostatnich stuleciach na obszarach słabiej poznanych może stanowić pierwszy krok w badaniach budowy równiny zalewowej. W dolinie Wisły, której holoceńska historia została na wielu odcinkach dobrze rozpoznana, stanowi istotne uzupełnienie pozwalające prześledzić najmłodsze etapy, często pomijane lub traktowane marginesowo przez geomorfologów. Umożliwia też zweryfikowanie poglądów, które były wyrażone w oparciu o dane geologiczne. Kompleksowa analiza kartograficzno-geomorfologiczna pozwala na uchwycenie i prawidłową interpretacje zapisu znajdowanego w profilach geologicznych w różnych środowiskach sedymentacyjnych dna doliny. Stosując aktualizm geograficzny możemy te sytuacje przenosić w starsze okresy, dla których brak jest już obrazu kartograficznego. Dokumentacja zmian koryta, zmian osadniczych i historycznych w dolinie rzecznej może stanowić podstawę do studiów fizycznogeograficznych i historyczno-gospodarczych.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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VISTULIAN AND HOLOCENE EVOLUTION OF THE WISŁOK RIVER IN THE NORTHERN MARGIN OF THE SUB-CARPATHIAN TROUGH

1. INTRODUCTION

The geological study of the Wisłok river valley in the Sub-Carpathian Trough is 100 years old and was initiated by W. Friedberg (1903) as part of research for the Geological Atlas of the former Austrian Galicia. W. Friedberg (1903) distinguished in Terliczka and Łukawiec bluish-grey young dilluvial (=young Pleistocene) clays covered by alluvial (= Holocene) deposits containing a fossil flora. The structure and age of the terrace levels in Rzeszów was determined by A. Jahn (1957), who dated the 20-metre-high loess terrace at the Vistulian, and a lower rendzina terrace, partly inserted in the loess terrace – at the Holocene. Within the rendzina terrace in the Rzeszów area, several insertions of alluvia dated at the Late Glacial and the Holocene were distinguished by L. Starkel (1960). L. Starkel (1972) associated the dune-covered sandy terrace situated a dozen metres above the Wisłok channel with the Middle Polish (Saalian) Glaciation. The peat which he found in Łukawiec was radiocarbon-dated by M. Geyh in 1975 at 43.9 ± 2.1 ka BP. The date indicates that the peat was formed during the Interpleniglacial period of the Vistulian Glaciation (Starkel 1980). W. Laskowska-Wysoczańska (1971) refers to organic clays in the profile of the Terliczka borehole, lying on gravel at the depth of 12 metres. After her opinion the clays come from the Eemian Interglacial period. The preparation of the *Detailed Geological Map of Poland* 1:50, 000 several years ago involved the publication of material on the geological structure and stratigraphy of the Quaternary deposits in the Rzeszów area (Zimnal 1999) and the alluvial plains at the Wisłok and San confluence (Wójcik et al. 1999). Recently, the preliminary results of geomorphological study of the Wisłok river valley between Łąka and Czarna were published (Gębica et al. 2002). It has been found in Czarna-Podbór that the sandy terrace 8–11 m above river level is built from overbank deposits with peaty mud yielded an age > 36.6 ka BP covered by fluvial sands dune-shaped on the top. In the profile of a 7-metre-high rendzina terrace in Łukawiec, loams with tundra flora have been dated >38.5 ka BP (Gębica et al. 2002).

Between 2001–2002, 16 manual probes were carried out in Łukawiec and Medynia Łańcucka in order to determine the extent of the erosional base of the rendzina terrace, built from Vistulian deposits. During the geomorphological mapping a sandy hummock in the shape of an erosion remnant was discovered and several new sites of Vistulian and Holocene deposits were located. The organic remains which had been found were palinologically analysed by K. Szczepanek from the Botany Institute of the Jagiellonian University in Cracow and radiocarbon-dated by A. Pazdur from the Department of Radioisotopes at the Silesian University of Technology in Gliwice.

2. GEOMORPHOLOGICAL SETTING

Below Rzeszów, a 40-kilometres-long stretch of the Wisłok river flows within an erosional depression called the Sub-Carpathian Trough Valley (Starkel, 1972). To the south and to the north, the trough is enclosed by high plateaus 210–270 m a.s.l. (Fig. 1). The plateaux are built from Miocene clays covered by preglacial alluvia, glacial and fluvio-glacial deposits from the Sanian 2 (Elsterian) Glaciation and Vistulian loesses (Laskowska-Wysoczańska 1971; Zimnal 1999). In the Sub-Carpathian Trough, glacifluvial deposits occur in places, covering interstadial organic deposits found in Jasionka (Laskowska-Wysoczańska 1967) and recently dated at the Ferdynandów Interglacial period by L. Lindner (2001). The valley-floor of the Wisłok, carved out of Miocene clays, is filled by alluvia from the Vistulian and Holocene periods. At the outlet of the valley from the Carpathian mountains, a terrace covered by loess extends at the altitude of 20 metres above the Wisłok channel. At a greater distance from the mountain edge, at the foot of the Kolbuszowa Plateau, stretches a dune-covered sandy terrace, 0.5 to over 2 km wide and 8–11 m high above the Wisłok level. A broader extent of the terrace is indicated by a sandy erosion remnant 2 km long and 1 km wide, found in Łąka in the central part of the valley, with a preserved system of braided channels.

The present-day bottom of the Wisłok valley, 4–6 km wide, is occupied by a Holocene rendzina terrace dissected to the depth of 10 m in Rzeszów and 7–8 m in the Łukawiec area. It forms a higher level of a floodplain 205–190 m a.s.l., which was partly water-logged during the 1934 flood (Lewakowski 1935). At the outlet from the Carpathian mountains, it is overbuilt by an alluvial fan with several abandoned paleochannel systems of the Wisłok and its tributary, the Czarna. A narrow and sinuous system of paleomeanders 20–25 m

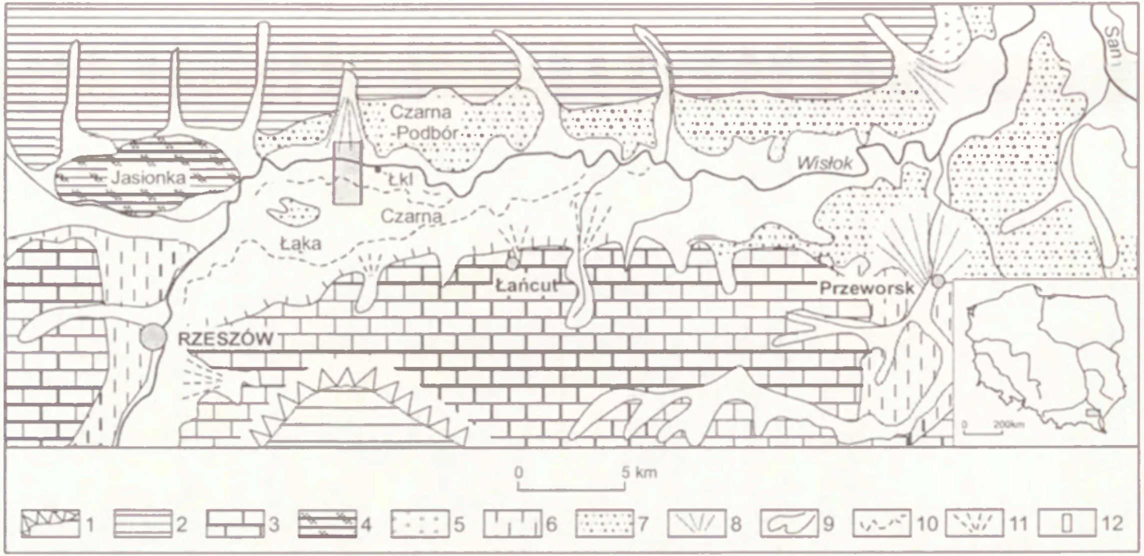


Fig.1. Geomorphological sketch of the Wisłok river valley in the Sub–Carpathian Trough (according to *Geomorphological Map of Poland 1:500 000*, partly modified)

1 – escarpment of Pogórze Dynowskie Foothills; 2 – the Kolbuszowa Plateau; 3 – the Kańczuga Plateau with loessial cover; 4 – fluvioglacial plain from the Sanian 2 Glaciation; 5 – fluvial terrace (13–18 m high above the Wisłok river) from Wartanian Glaciation (?); 6 – Vistulian terrace with loess cover (20 m high above Wisłok river level); 7 – dune–covered sandy terrace from Vistulian Glaciation (8–12 m high above the river level); 8 – alluvial fans (Vistulian); 9 – erosional–accumulated valleys; 10 – rendzina terrace (7–10 m high above the river level) with Holocene oxbow-lake system; 11 – alluvial fans (Holocene); 12 – location of study area

Szkic geomorfologiczny doliny Wisłoka w Rynnie Podkarpackiej (według: *Geomorfologiczna mapa Polski 1:500000*, częściowo zmieniony)

1 – Próg Pogórza Dynowskiego, 2 – Wysoczyzna Kolbuszowska, 3 – Wysoczyzna Kańczucka z pokrywą lessów, 4 – równina fluwioglacjalna z okresu zlodowacenia Sanu 2, 5 – terasy rzeczne (13–18 m nad poziom rzeki) z okresu zlodowacenia Warty?, 6 – Terasa z pokrywą lessów (20 m nad poziom rzeki) z okresu zlodowacenia wisły (vistulianu), 7 – Terasa piaszczysta w stropie zwydmiona (8–12 m nad poziom rzeki) z okresu zlodowacenia wisły, 8 – stożek aluwialny z vistulianu, 9 – doliny erozyjno-akumulacyjne, 10 – terasa rędzinna (7–10 m nad poziom rzeki) z systemami holoceńskich starorzeczy, 11 – stożki napływowe z holocenu, 12 – położenie obszaru badań

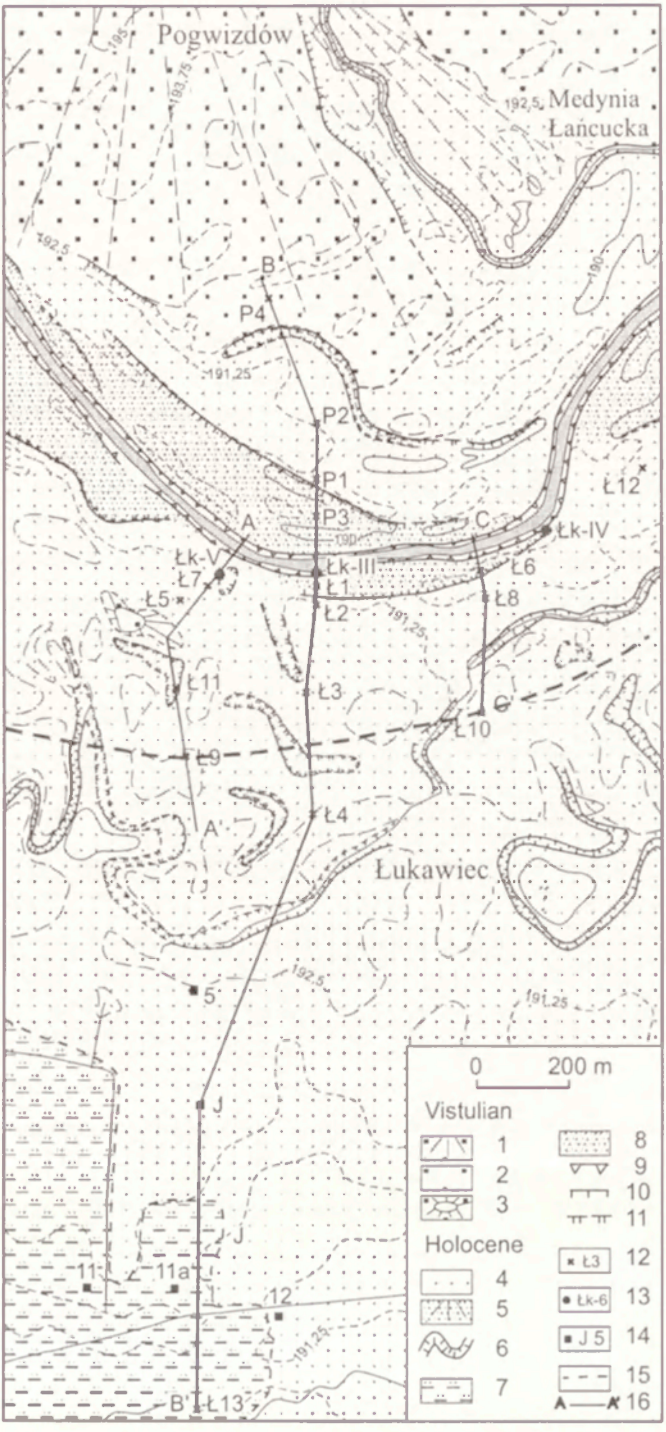
wide is best-known. Partly filled with water, the channel called the Old Wisłok was abandoned during one of the catastrophic flood in mid-18th century (Strzelecka 1958). The lower level of the floodplain stretches in the shape of a ledge several dozen to 500 m wide and 3–6 m above the Wisłok channel. The present-day channel is 25–35 m wide and, despite partial regulation, many of its sections are natural in character. There are no embankments along the Wisłok river.

3. FLUVIAL FORMS AND DEPOSITS IN MEDYNIA ŁAŃCUCKA AND ŁUKAWIEC

In the north, the valley-floor of the Wisłok forms a broad fan of an alluvial cone 0.8 km^2, dissected by a small Pogwizdówka stream (Fig. 2). The plain of the fan gently inclines between 197 and 191 m a.s.l. (13–7 m above Wisłok level). The fan is built from sands and loams with thin interbeddings of gravels 1–2 cm in diameter, which form rhythmic sequences throughout the 5-metre profile of bore P4 (Fig. 3). The fan and a sandy terrace are undercut by rectilinear erosion scarps and by a pronounced bow of the old river-bed, whose edge is 2–3 m high. In the studied part of the valley, the rendzina plain is only 100–140 m wide. It is built from a loamy-organic series in the floor, a middle series of sands and channel gravel and alluvial loams at the top (bores P2, P1). A bottom series of ashen-brown loams, situated on ferruginized gravel, was identified in an erosion undercutting in Medynia Łańcucka. The loams,

Fig. 2. Geomorphological map of the Wisłok river valley in Łukawiec and Medynia Łańcucka (Gębica 2003)
Vistulian: 1 – alluvial fan (7–13 m high above Wisłok level); 2 – sandy terrace (8–11 m above the river level hight); 3 – erosional remnant (9–10 m high above the river level); Holocene: 4 – higher floodplain (rendzina terrace, 7–8,5 m high above the river level); 5 – alluvial fan; 6 – paleochannels; 7 – floodbasin; 8 – lower floodplain (3–6 m high above the river level); 9 – the erosion scarps and cuts of high above 4 m; 10 – erosion edges and cuts of high to 4 m, well preserved; 11 – smoothed edges and cuts of high to 4 m; 12 – boreholes; 13 – exposures; 14 – archival boreholes; 15 – southern extend of fossil erosional socle built of Plenivistulian deposits; 16 – lines of geological cross-sections

Mapa geomorfologiczna doliny Wisłoka rejonu Łukawca i Medyni Łańcuckiej (Gębica 2003)
Vistulian: 1– stożek aluwialny (7–13 m nad poziom rzeki), 2 – terasa piaszczysta (8–11 m), 3 – ostaniec erozyjny (9–10 m nad poziom rzeki); Holocen: 4 – wyższy stopień równiny zalewowej (terasa rędzinna, 7–8,5 m nad poziom rzeki), 5 – stożek napływowy, 6 – starorzecza, 7 – basen popowodziowy, 8 – niższy stopień równiny zalewowej (terasa łęgowa, 3–6 m nad poziom rzeki), 9 – krawędzie i podcięcia erozyjne o wysokości powyżej 4 m, 10 – krawędzie i podcięcia erozyjne o wysokości do 4 m, 12 – wiercenia, 13 – odsłonięcia, 14 – wiercenia archiwalne, 15 – zasięg kopalnego cokołu erozyjnego utworzonego z osadów pleniglacjalnych, 16 – linie przekrojów geologicznych


Pogwizdów
Medynia Łańcucka
Łukawiec
B
P4
P2
P1
P3
A
C
Łk-V
Ł7
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Łk-III
Ł1
Ł2
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Vistulian
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eroded at the top, form a 2-metre-high socle of a rendzina terrace. The lower level of the floodplain (meadow terrace) 100–200 m wide is formed by organic silts and clays overbuilt by channel deposits (sands with gravel and pebbles) and overbank alluvial sands and loams (bore P3).

In Łukawicc the whole breadth of the valley is taken up by the rendzina terrace elevated to 7–8.5 m above river level (Fig. 2). The terrace comprises poorly marked, single meander scrolls belonging to an older paleomeander system, and a younger, sinous and narrow paleochannel. Its width (10–15 m) and a small radius of curvature (30–40 to 60 m) indicate that it is an abandoned system of old river-beds of the Czarna stream which formerly (before the avulsion of the Old Wisłok to the north) ran to the Wisłok and discharged itself into it in a village of Czarna, to the west of Łańcut. Nearer the present-day channel, the rendzina plain is partly built from an alluvial loam (madas) cover 2–4 m thick. It contains insertions of peat and mouldy clay layers lying directly on sands with loam interbeddings (bores Ł2, Ł3, Ł8–Ł11, Fig. 3). The socle of the terrace is built from cohesive organic silts and clays. On the erosional socle a series of sands with gravel occurs. The abandoned system of paleomeanders is accompanied by a 0.6–0.8-km-wide zone of the levee built from sandy loams (bore Ł4). The levee gradually passes into a flat waterlogged plain cut by unpronounced paleochannels and drained by ditches. The plain is built from clays and silts overlying a series of sands (Fig. 3, profile B-B'). In places, peat occurs and fills paleochannels (identified in archival bores). At a distance of 300 m away from the present-day channel of the Wisłok, an isolated hummock rises 1–2 m above the plain – a remnant of a sandy terrace plain of the height of 193.5 m a.s.l (Fig. 3, profile A–A'). In the north, it gently slopes towards the flat plain covered by a thin layer of alluvial loams 0.5–1 m thick. In the south, the hummock is undercut by an old river-bed 10–13 m wide. Digging for a "wild" landfill allowed an insight into the structure of the hummock. Under the alluvial loams, there is a 4-metre-thick cover of layered fine-grained sands with laminae of silts covering organic muds with peat lenses. Organic deposits are disturbed by involutions.

The contemporary Wisłok channel is accompanied in places by ledges of the lower level of the floodplain (meadow terrace) 40–150 m wide and 5–6 m above river level. The plain is built from channel deposits (sands with gravel and sands) covered by a thin layer of sandy alluvial loams (boring Ł1).

4. STRATIGRAPHY AND AGE OF ALLUVIAL FILLS

The oldest dated series consists of loams with tundra flora forming a 2-metre-high socle of the rendzina terrace excavated in the Lukawiec I site (Fig. 1). The radiocarbon age of the deposits was determined to be older than 38 500 BP (Gębica et al. 2002). A younger alluvial fill is built from layers of highly disturbed organic loams which crop out in profiles Lk-III and Lk-V (Fig. 3). The dating of the organic layer sheared in the top at the height of about 3 m above channel level pointed to an age of 30 000± 1 500 BP (Gd-12404), which corresponds to the Denekamp Interstadial. The above results of radiocarbon dating of the deposits, including the date 43900±2100 BP (Starkel 1980), imply that the erosional socle of the rendzina terrace contains very complex series of sediments dated from the Middle PleniVistulian.

Deposits dating back to the Upper Pleniglacial are represented by sands with gravel pavement (the bottom part of series II in Fig. 3). They are situated on disturbed interpleniglacial loams (series Ib), and are covered by rhythmically alternating sands and silts (middle and upper part of series II). In exposure Lk-IV, sands interbedded with silts are truncated at the height of 4 m above the channel level and are covered by a younger series of sediments (serie III). It is formed by a 0.7-metre-thick layer of channel sands gravels and covered by sandy loams and cohesive peat 0.3 m thick. A peat sample from the depth of 2.9–3.0 m contained numerous pollen grains of *Pinus* (102 pollen grains) and *Betula* (116) trees, including the *Betula nana* and probably *Pinus cembra* types, as well as single grains of *Ulmus* and *Alnus*. Among herbaceous plants, grasses (*Graminae* – 25 grains), sedges (*Cyperacae*) and *Filicales* prevailed. Such a plant composition indicates, according to K. Szczepanek, the turn of a Late Glacial – the beginning of the Holocene. It is confirmed by radiocarbon dating which has determined the age of the peat layer at 10 150 ± 140 BP (Gd-15380), and thus the channel sands deposited under the peat come from Younger Dryas. The 2.7-m-thick clay cover overlying the peat (series IV) represents the Early Holocene – Atlantic periods.

Closer to the present-day channel, in several probe profiles, peat or clayey humus layers covered by alluvial loams (silts and clays- series IV) lie immediately on sands from the Upper PleniVistulian (series III). The peat insertion was sampled at the depth of 2.5–2.7 m (Probe L3) and was found by K. Szczepanek to contain numerous pollens of *Pinus* (114 grains), *Betula* (73), *Alnus* (40), *Corylus* (15) trees, as well as several grains of each *Quercus*, *Picea*, *Tilia*, *Ulmus*. Apart from tree pollen, also herbaceous (*Graminae* – 69 grains), sedge (*Cyperacae* – 10) and fern (*Filicales* – 33) pollen was found.

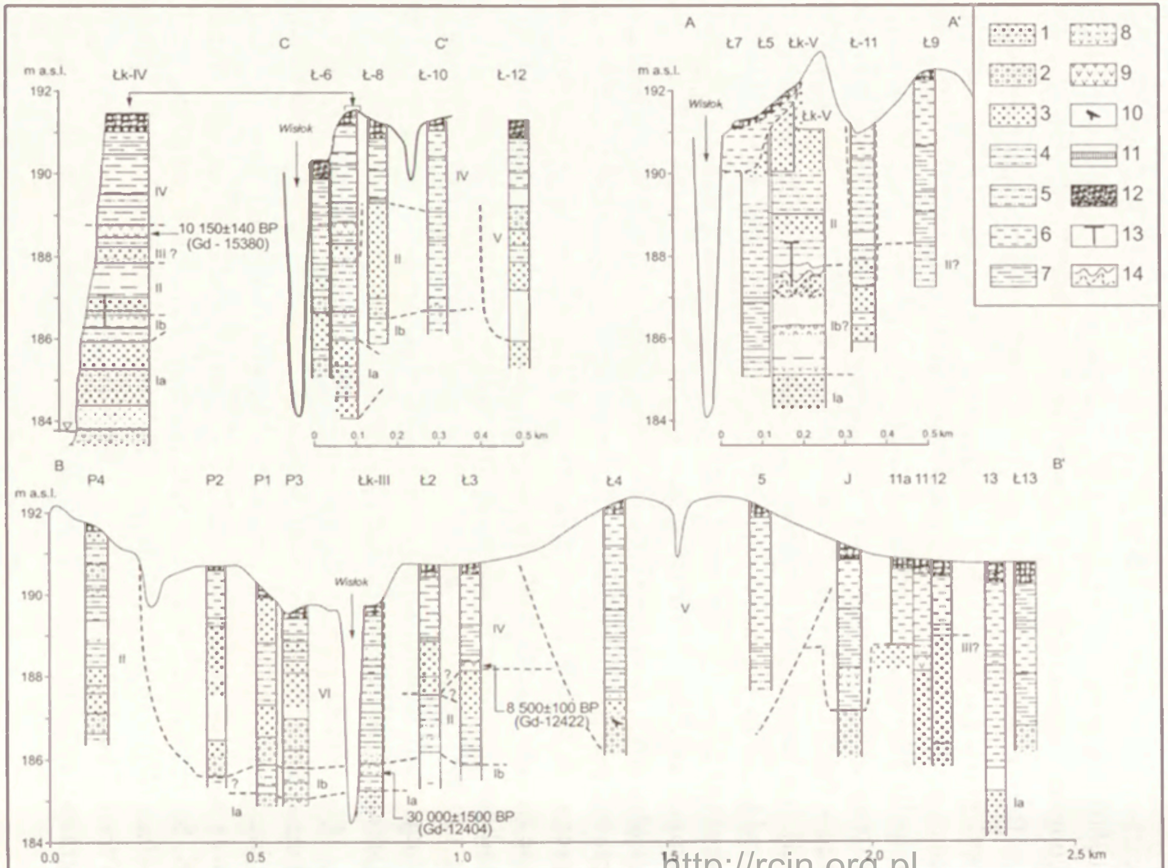

Fig. 3. Geological cross–sections in the northern margin of Wisłok river valley in Łukawiec (after P. Gębica 2002)

1 – sandy gravel; 2 – sand with gravel; 3 – sand; 4 – silty sand; 5 – sandy silt; 6 – silt; 7 – clay, clayey silt; 8 – organic silt; 9 – peat; 10 – wood debris; 11 – fossil soils; 12 – present-day soils; 13 – ice–wegde casts; 14 – involutions; Ia, Ib, II – series of Plenivistulian age; III – series of Late Vistulian age; IV, V, VI – series of Holocene age

Szczegółowe profile geologiczne północnego obrzeżenia doliny Wisłoka w rejonie Łukawca

1 – żwir piaszczysty, 2 – piasek ze żwirem, 3 – piasek, 4 – piasek mułkowaty, 5 – mułek piaszczysty, 6 – mułek, 7 – ił, mułek ilasty, 8 – mułek organiczny, 9 – torf, 10 – fragmenty drewna, 11 – gleby kopalne, 12 – gleba współczesna, 13 – kliny mrozowe?, 14 – inwolucje; Ia, Ib, II – ogniwa osadów plenivistulianskich; III – ogniwa osadów z późnego vistulianu; IV, V, VI – ogniwa osadów holoceńskich

Such a composition of the spectrum is characteristic of forest communities of the Holocene climatic optimum (beginning of the Atlantic period). It has been fully confirmed by the results of radiocarbon dating of the analysed peat sample, which points to 8 500 ± 100 BP (Gd-12422). Thus, the thick layer of clays covering the peat came from Atlantic period.

5. THE COURSE OF FLUVIAL ACCUMULATIONS

The northern part of the Wisłok valley in the sub-Carpathian Trough is filled with fluvial deposits which provide a record of fluvial accumulation and erosion phases from the PleniVistulian to the Holocene. The 4–5-metre-thick cover of rewashed river gravel in the fossil trough (series la) was probably accumulated in periglacial conditions of the Lower Pleniglacial (Jahn 1957) or is much older. The sandy bars which formed on the gravel and pass into loams manifest a marked decrease in transport dynamics, gradual decline of the river flow and channel abandoning.

The abandoned paleochannels or shallow thermokarst lakes (Kasse et al. 1995) were filled by organic silts, clays and peaty muds yielded the radiocarbon age >38.5 ka BP (Gębica et al. 2002). It was truncated by sands with graveles at the top. A younger accumulation phase is represented by gyttja loams with peat lenses. They are probably deposits of wide shallow oxbow-lakes (profile Lk-V). At the same time, loams with organic matter indicating pedogenesis were formed in the profile Łk-IV. The deposits lie 3.5–4 m above channel level and are sheared by a pronounced erosional surface. In places, they are strongly disturbed by sandy load casted megastructures and involutions. Some of the structures resemble cryoturbations found in Vistulian periglacial deposits (Vandenberghe, Broek 1982). The dating of organic loams cropping out in the Wisłok channel in profile Lk-III in the same horizon indicates the decline of the Middle PleniVistulian (30 ± 1.5 ka BP). Thus, the dissection of the plain surface took place after 30 ka BP.

Accumulation of a 4–5 metre thick sand series, which is one of the components of the remnant hummock in Łukawiec, took place as early as the descending phase of the Upper Pleniglacial. Initially, it was dominated by the deposition of sandy channel and extra-channel facies of the braided river, well-developed in the bottom part of the alluvial series. The deposition occurred during the period of maximum cold, which is indicated by the structures of frost-cracks (Mol 1997). The deposits in the top part reveal a marked change of sedimentation type – from channel to overbank (flood sheet) deposition.

The described course of fluvial processes in the Inter- and Upper PleniVistulian periods was of panregional significance as similar processes and sediments were identified in the valleys of the Sandomierz Basin (Mamakowa, Starkel 1974; Mamakowa et al. 1997), Lublin Upland, Roztocze (Superson 1996), Silesian Upland (Jersak, Sendobry 1991) and Central Poland (Manikowska 1996).

The dissection of the sandy terrace is believed to have taken place in the bölling. It is manifest in the occurrence of peats situated on channel sands under alluvial loams, dated at 11.82 ± 0.25 ka BP in Łąka (Gębica et al. 2002). In San valley near Stubno the deep incision before the Late Vistulian is marked by paleochannel with organic filling dated at 15200 ± 500 years BP (Klimek 1992). In Łukawiec, a younger insertion of channel deposits within the Pleniglacial series was found at the height of several metres above the Wisłok level. The deposits in the insertion probably come from Younger Dryas, since they are covered by peat dated 10.15 ± 0.14 ka BP. Thus, the abandoning of the alluvial plain and the deepening of the Wisłok channel took place at the beginning of t he Eoholocene, which is consistent with research in other valleys of the upper Vistula basin (Starkel, Gębica 1995; Starkel 2001). At the beginning of the Holocene, the alluvial plain was covered by peat, and subsequently, as a result of frequent floods, the peat was covered by clays. The development of meander channels of the Wisłok and its tributary, the Czarna, led to the dissection of the sandy plain and its peat-covering at the beginning of the Atlantic period (peat floor dated 8.5 ± 0.1 ka BP). A new alluvial fill was inserted. Gradual aggradation in the valley floor and an increase in flood frequences during the wet phase of Atlantic period brought about sedimentation of clays on peat in flood basins situated further from active meander channels. Nearer the active channels, within the zone of the levees, silty and sandy loams, overlying clays were deposited.

It is difficult to date the final abandoning of the Czarna paleochannel channel. The avulsion of the paleochannel system of the Old Wisłok from the southern part of the valley to the northern trough shortened the Czarna course. Since the mid-18th century, due to a coincidence of natural and anthropogenic factors (increase in farming and deforestation) a gradual widening of the Wisłok channel caused by a large rock-debris supply took place. During the regulation process, at the beginning of the 20th century, the Wisłok bed was narrowed and deepened by 2–3 m. As a result, a lower plain level was formed (meadow terrace), flooded during annual rise of water. Incision of the bed into the cohesive organic loams caused irregular dissection of the loams and the

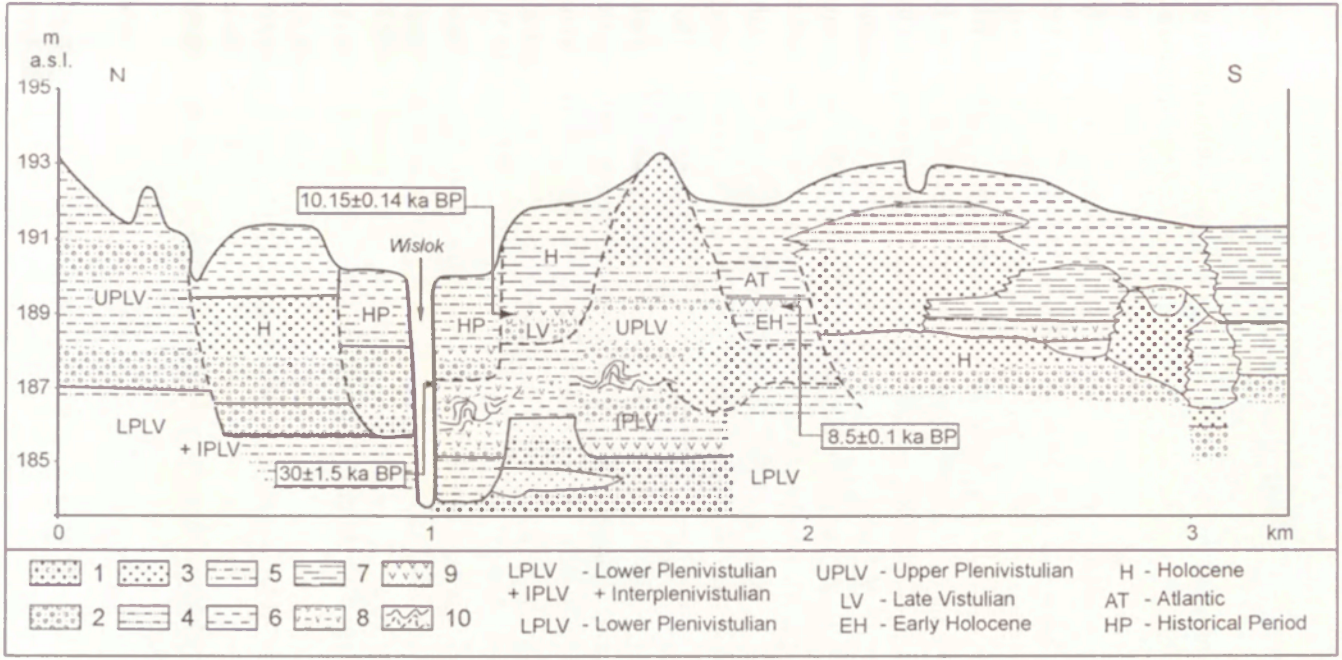

Fig. 4. Synthetic morphological-geological section of Wisłok river valley in Łukawiec and age of alluvial fills. Explanation of lithological symbols see fig. 3

Syntetyczny przekrój dna doliny Wisłoka w północnej strefie Rynny Podkarpackiej w rejonie Łukawca oraz wiek datowanych włożeń aluwialnych. Objaśnienia litologicznych symboli, patrz rycina 3
LPLV+ IPLV – dolny i środkowy plenivistulian, LPLV – dolny plenivistulian, UPLV – górny plenivistulian, LV – późny vistulian, EH – wczesny holocen, H – holocen, AT – atlantyk, HP – okres historyczny

formation of local steps in the Wisłok channel. They were gradually dissected due to headward erosion, which led to the exposure of basal gravel.

6. FINAL REMARKS

The study of a parallel course of the Wisłok River in the northern part of the Sub-Carpathian Trough, allowed us to distinguish a complex sequence of Pleniglacial deposits and a system of alluvial inserts and paleochannels from the Late Glacial and the Holocene (Fig. 4). Pleniglacial deposits can mainly be found in the left-side sandy terrace covered by dunes in the top and in the alluvial fans of side tributaries. In Lukawiec, in the rendzina terrace, a fossil erosion level was discovered comprising deposits dated >38,5 and 30 ka BP (Middle Pleniglacial) and an insertion from the Upper Pleniglacial period. It resembles the structure of the fossil outliers buried by Holocene deposits in the Brzeźnica site on the Wisłoka river (Mamakowa, Starkel 1974) and the marginal zone of the San river valley in the area of Walawa and Barycz, where loams with Pleniglacial flora were found on both sides of the river under Holocene deposits (Klimaszewski 1948; Starkel 1960). In the northern trough of the Wisłok, systems of insertions and paleochannels were also found, dated at the turn of the Late Vistulian (Younger Dryas), the beginning of the Holocene and the Atlantic period. This changes of fluvial processes and high sedimentological rate of channel and overbank deposition is due to climatic and vegetational changes during the Upper Vistulian and Holocene. They were earlier recorded in the meridional course of the Wisłok valley in the area of Rzeszów and the San valley (Starkel 1960; Szumański 1986; Klimek 1992). In the Holocene the alluvia were reworked as deep as 5–6 metres. Beyond these zones, the Holocene alluvial loams lies directly on the sands from the Upper Pleniglacial. It cannot be excluded that the southern trough, used by a system of paleomeanders of the Old Wisłok, also registers a number of Late Glacial and Holocene alluvial fills of different ages. The parallel occurrence of a system of two fossil trough valleys separated from the active floor by a sandy Vistulian terrace is possible in wide valley-stretches and on alluvial fans, where the river course often changes due to flooding and the whole channel systems are abandoned by avulsion.

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EWOLUCJA DOLINY WISŁOKA NA PÓŁNOCNYM OBRZEŻENIU RYNNY PODKARPACKIEJ W VISTULIANIE I HOLOCENIE

Streszczenie

Wisłok– najdłuższy dopływ Sanu płynie obecnie północnym skrajem równoleżnikowego obniżenia zwanego Rynną Podkarpacką, którą przed i w czasie zlodowaceń południowopolskich odpływały wody na wschód do doliny Dniestru.

Badania równoleżnikowego odcinka doliny Wisłoka w Rynnie Podkarpackiej pozwoliły na wyróżnienie w profilach w Łukawcu złożonej sekwencji form i osadów pochodzących z Vistulianu i Holocenu. Z ostatniego piętra chłodnego pochodzą: terasa piaszczysta (8–12 m) w stropie zwydmiona, stożek napływowy oraz ostańce terasy piaszczystej na prawym brzegu rzeki. W obrębie równiny rędzinnej (7–8 m nad poziom Wisłoka) został stwierdzony kopalny cokół erozyjny utworzony z osadów środkowego plenivistulianu (daty radiowęglowe 43 ka i 30 ka BP) oraz leżącej na nim pokrywy piasków fluwialnych pochodzących z górnego plenivistulianu. W rozcięcia pokywy pleniglacjalnej zostały włożone aluwia budujące

przedallerodzką równinę roztokową (11,82 ka BP), paleokoryto z młodszego dryasu (10,15 ka BP) oraz pokrywa mad atlantyckich (8,5 ka BP). Wyróżnione graniczne daty dobrze korelują z wydzielanymi przez L. Starkla fazami zwiększonej geomorfologicznej aktywności rzek w Polsce i Europie, związanymi ze zmianami klimatu. Młodsze aluwia budują m.in. fragment równiny z wałem przykorytowym krętego systemu starorzeczy rzeki Czarnej – dopływu Wisłoka oraz system starorzeczy Starego Wisłoka zajmujący do połowy XVIII południową rynnę. Podczas je-dnej z katastrofalnych powodzi nastąpił przerzut Wisłoka na północ i rzeka wkroczyła do północnej rynny dolinnej. W okresie sprzed regulacji dno Wisłoka było szersze i miało układ zbliżony do roztokowego, później nastąpiło rozcięcie równiny i pogłebienie koryta. W ten sposób powstał niższy stopień równiny zalewowej (3–6 m nad poziom rzeki) zalewany w czasie corocznych wezbrań.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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DEPOSITION AND DESTRUCTION OF HOLOCENE CALCAREOUS TUFA CASCADES IN THE BOHEMIAN KARST (CZECH REPUBLIC)

1. INTRODUCTION

Over large parts of Europe from the British Isles to the Mediterranean and from Spain to the Czech Republic, Slovakia and Poland, the rates of calcareous tufa deposition were high in the Early and Mid Holocene, but declined markedly thereafter (Weisrock 1986). Over much of Europe, it has been postulated that in the Late Holocene (since ~ 2.500 years B.P.) there was a sharp decline in the deposition of tufa (Goudie et al. 1993). There has been a considerable debate about the causes of this phenomenon, with some authors stressing the importance of natural climatic changes, and others asserting that miscellaneous human activities (e.g., deforestation connected with agricultural activities) were crucial.

Several tens of springs in the Bohemian Karst deposited calcareous tufa from their waters during the Early and Mid Holocene. Tufa deposition continues at a number of localities at present, too (Kadlecova, Žak 1998). As shown by V. Lozek (1992), most of the calcareous tufa bodies of the Bohemian Karst show very similar lithological and biostratigraphic pattern, reflecting Holocene climatic, hydrological and biotic conditions.

2. GEOGRAPHICAL, GEOLOGICAL AND HYDROGEOLOGICAL SETTINGS

Bohemian Karst extends between Praha and Zdice over an area of ca 200 km^2. With its NE margin, this karst area reaches to the territory of Prague. With its climatic conditions, the Bohemian Karst is a warm and dry area lying at altitudes of 210–500 m a.s.l. Total annual precipitation is 500 mm with an average annual temperature of 8–9 °C (Quitt 1971).

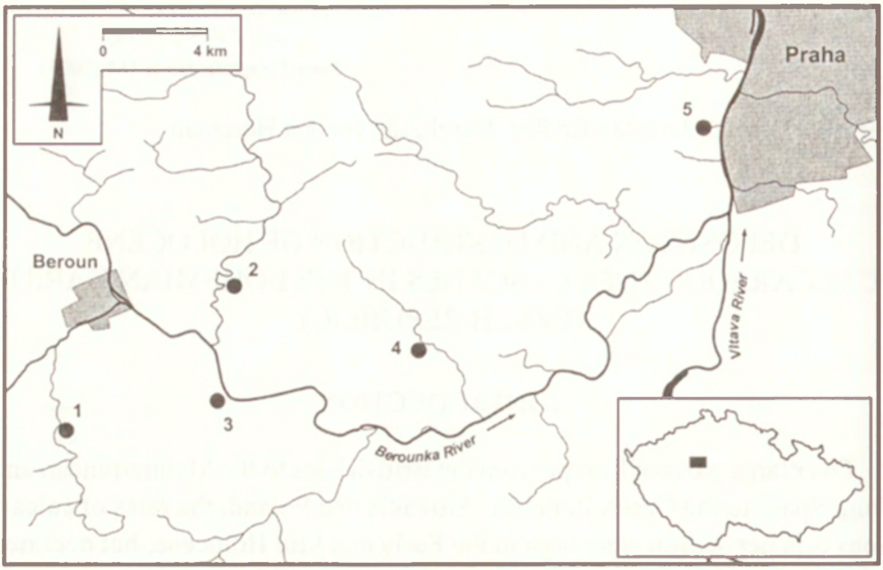

Fig. 1. Schematic map of the geographic positions of selected tufa cascades studied
1 – tufa cascade at Kotyz; 2 – tufa cascade at Svaty Jan pod Skalou; 3 – tufa cascades in the Cisarska Gorge; 4 – tufa cascade "Petranka" in the Karlicke Valley; 5 – tufa cascades in the Čertova Gorge at Mala Chuchle

Schematyczna mapa rozmieszczenia badanych trawetynów kaskadowych
1 – trawertyny kaskadowe w Kotyz; 2 – trawertyny kaskadowe w Svaty Jan pod Skalou; 3 – trawertyny kaskadowe w Wąwozie Cisarska; 4 – trawertyny kaskadowe „Petranka" w Dolinie Karlické; 5 – trawertyny kaskadowe w Wąwozie Čertova w Mala Chuchle

The Bohemian Karst is located in the central part of the Prague Basin. This basin contains a continuous succession of Ordovician to Middle Devonian deposits accompanied by volcanic activity (Chlupáč et al. 1998). The Bohemian Karst itself, located SW of Prague, is composed of Upper Silurian shales, limestones and basalt lava flows (diabases) plus volcaniclastic rocks (tuffs) and Lower to Middle Devonian limestones and shales. During the Late Paleozoic Variscan Orogeny, sediments of the Prague Basin were folded into large anticlinal and synclinal structures and faulted (Chlupáč et al. 1998).

Groundwaters of the Paleozoic basin fill are issued on many places in the Bohemian Karst: along exposed boundaries between limestones and non-carbonate rocks near bottoms of karstic valleys, and at fault intersections. Most of the karst springs are characterized by relatively stable discharge (between 0.1 and 20 l/s) and temperature, somewhat exceeding average annual temperature in the area in some cases. This is indicative of a deeper karst-water circulation (Kadlecova, Žak 1998; Žak et al. 2001). A typical feature of most springs of the Bohemian Karst is the deposition of calcareous tufa in their proximity, precipitating from the issued karst waters. About 70 localities where calcareous tufa precipitates or was deposited in the past have been described from the Bohemian Karst (Kovanda 1971; Lozek 1992; Kadlecova, Zák 1998). The most common forms are tufa barriers and cascades, which were growing near karst springs during the Holocene, barring the bottoms of karstic valleys.

Five tufa sections with characteristic development were selected for the correlation of lithological features, variations in molluscan assemblages and time successions of accumulation and subsequent destruction of tufa bodies in the Bohemian Karst. These include (from SW to NE): the cascade at Kotyz, the cascade at Svaty Jan pod Skalou, cascades in the Cisarska Gorge near Srbsko, the cascade of "Petránka" in the Karlické Valley, and the cascade on the Čertova Gorge at Mala Chuchle (Fig. 1).

3. METHODS USED

3.1. MALACOZOOLOGICAL ANALYSES

Acquisition of samples for paleomalacological and biostratigraphic analyses followed the unified methodology of Lozek (1964). If permitted so by the character of the sediment, samples of the matrix 3–4 l in volume were taken from all macroscopically distinguishable beds within the studied tufa sections. These were later, after careful drying, washed in laboratory conditions. Individual shells and their fragments were hand-picked from the washing residue under a stereomicroscope and determined. Fragments of shells were recalculated to whole individuals following the standard methodology. For a better orientation in fossil malacocoenoses, the proportions of the registered mollusc species in selected tufa sections were statistically processed into the form of histograms (malacospectra). The malacospectrum of species (MSS) expresses relative proportions of the individual species within the principal ecological groups. Attribution of molluscan species to the principal ecological groups in the histograms follows Lozek (1964).

3.2. DATING ANALYSES

3.2.1. DATING BY THE U-SERIES METHOD (U-SERIES LABORATORY OF THE GEOLOGICAL INSTITUTE OF THE POLISH ACADEMY OF SCIENCES, POLAND)

Samples ca. 1–2 kg in weight were collected in the field and subjected to a careful separation of carbonate cement in the laboratory. Preference was given to portions with pure, crystalline cement with no substantial admixture of clay minerals and other detrital components. Dating of carbonates employed the $^{230}Th/^{234}U$ method. Uranium and thorium were separated from the carbonate using a standard chemical procedure (Ivanovich and Harmon, 1992). The samples were dissolved in 6 M nitric acid, and uranium and thorium were separated by a chromatographic method using the DOWEX 1 x 8 ion exchanger. The efficiency of chemical separation was controlled by addition of spike $^{228}Th/^{232}U$. Activity measurements (α spectrometry) were taken on the OCTETE PC device of the EG&G ORTEC company. Spectral analysis and age calculation were performed with the use of "URANOTHOR 2.5" software (Gorka, Hercman 2002). It was necessary to perform correction of samples for an admixture of detrital thorium. The assumed initial $^{230}Th/^{232}Th$ activity ratio of detrital contamination was 1.5 ± 0.5. Uncertainty in this ratio results in a larger error of the calculated corrected age.

3.2.2. RADIOCARBON DATING BY THE AMS METHOD (POZNAN RADIOCARBON LABORATORY, POLAND)

Small pieces of charcoal were (after treatment with hot solutions of acid, alkali, and acid to remove all carbonate carbon and easily soluble organics) measured by the AMS method. The ^{14}C data on organic matter have been calibrated for variable initial ^{14}C concentration using the OxCal v3.5 calibration program (Bronk 2001).

4. CALCAREOUS TUFA CASCADES

4.1. TUFA CASCADE AT KOTYZ

Tufa cascade at Kotyz (No. 1 in Fig. 1) is preserved in the slope 15 m above the present karst spring. The tufa sequence was exposed by a test hole.

Lithology

Tufa body is underlain by slope sediments composed of fragments of dark Silurian shales. The section can be subdivided into 16 beds (Fig. 2).

Malacozoology

Although the basal interval (Beds 16–10) is of very low species diversity due to the poor fossilization ability, it occasionally yielded species like *Vertigo substriata*, *Perpolita petronella* and *Vitrea crystallina*, characteristic for early Holocene. This interval completely lacks typical glacial elements, however, the accompanying species *Vallonia costata* and *V. pulchella* indicate the presence of open habitats of dry to mesic character.

The overlying beds 9–1 markedly differ from the above mentioned underlying interval in the fossil malacofauna they contain. The bed of solid structural tufa (Bed 9) already shows a gradual onset of forest species, but the species of open habitats are still maintaining a markedly higher abundance. It was only in this bed that the early Holocene element *Discus ruderatus* was rarely encountered. From Bed 8 upwards, a rapid boom of forest species occurs at the expense of open-landscape elements. Molluscan species in this bed document closed woodland conditions, also persisting in the overlying interval (Beds 7–4). Fossil molluscs from Bed 8 and from the interval of Beds 7–4 indicate environmental conditions of the Holocene climatic optimum, i.e. a massive expansion of woodland molluscan species such as *Bulgarica cana*, *Vitrea diaphana*, *Platyla polita*, *Vertigo pusilla* etc., and high abundance of the hygrophilous element *Carychium tridentatum*, whereas the open-landscape elements and the majority of mesophiles disappeared or became very rare (e.g., *Vallonia pulchella* and *Truncatellina cylindrica* are completely absent from Beds 6–4).

The abundance of open-landscape species increases and that of moist woodland species decreases within Beds 2 and 3. After termination of tufa deposition the cascade was covered by dark brown humic soil sediment with increasing number of rock fragments redeposited from the upper part of the slope. The appearance of modern immigrants (e.g., *Oxychilus cellarius*) is characteristic for this uppermost bed in the section.

Chronology of tufa deposition

Charcoal from Beds 8 and 4 was subjected to ^{14}C dating (Fig. 2). Given on the obtained data and the occurrence of fossil molluscan assemblages, horizons of different age can be distinguished in the tufa accumulation. Tufa deposition started in the Boreal period. Solid tufa with rare shale fragments (Bed 9) contains molluscan fauna indicating moist, light forest environment with decreasing abundance of species *Discus ruderatus*, *Trichia sericea* and *Perpolita petronella*. The underlying Beds 16–10 can be thus generally dated to the late Glacial to Preboreal because of the sporadic occurrence of undemanding molluscan species and the prevalence of Early Holocene elements.

The interval between Beds 8 and 4 is characterized by massive development of woodland communities; the abundance of hygrophilous elements dominated by *Carychium tridentatum* indicates humid climate. This interval can be attributed to the Holocene climatic optimum – Atlantic and Epiatlantic, as was also confirmed by ^{14}C dating to the period 5,070–3,090 years BC

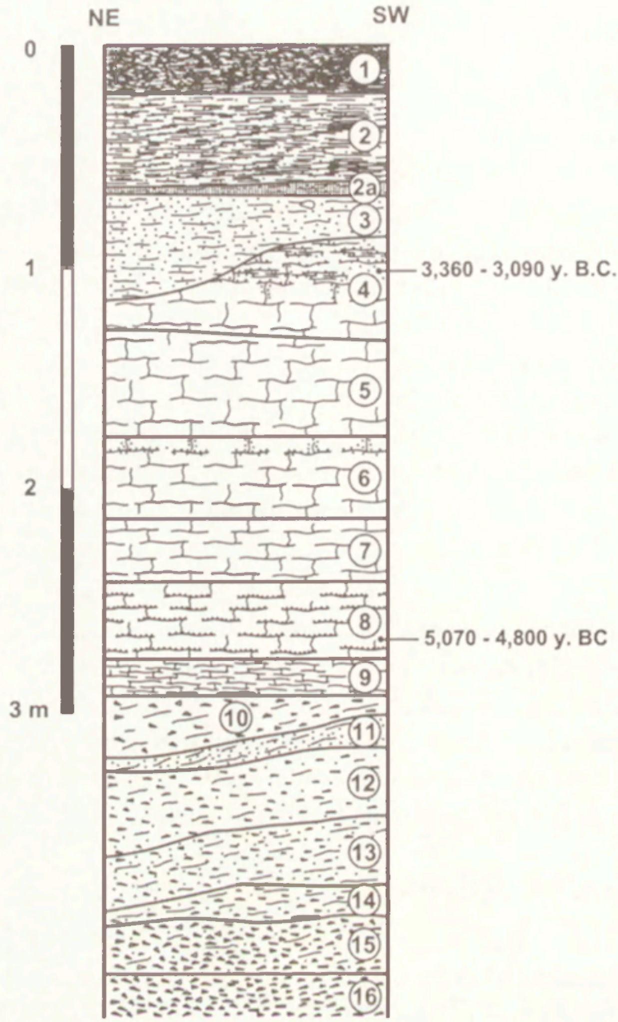

Fig. 2. A section through the spring tufa deposit at Kotyz
1 – dark brown humic Rendzic Leptosol with sporadic limestone and diabase clasts (4 cm in diameter); 2 – dark yellowish brown slightly humic soil sediment with rare limestone clasts (0.5 cm); 2a – very dark greyish brown slightly humic soil sediment; 3 – dark brown slightly humic soil sediment with abundant tufa encrustations; 4 – pale brown fine-grained tufa with coarse encrustations in the upper portion of the bed and dark stains coloured with Fe and Mn oxides; 5 – light yellowish brown tufa; 6 – yellowish brown tufa with coarse encrustations in the upper portion of the bed; 7 – very dark greyish brown tufa with rare, very dark brown Silurian shale clasts; 8 – very dark brown mouldered slightly loamy tufa with rare shale clasts; 9 – dark greyish brown solid tufa with shale clasts; 10 – very dark brown shales (clasts 1 cm in diameter), dark greyish brown loamy matrix; 11 – very dark brown shales (clasts 2 cm in diameter) with tufa encrustations (30%), dark brown loamy matrix; 12 – very dark brown shales (clasts 2 cm in diameter) with sporadic tufa encrustations, very dark grey matrix;

13 – very dark brown shales (clasts 2 cm in diameter) with sporadic tufa encrustations, very dark greyish brown matrix; 14 – very dark brown shales (clasts 3 cm in diameter) with tufa encrustations (20%), dark stains coloured with Fe and Mn oxides, dark yellowish brown matrix; 15 – very dark brown shales (clasts 4 cm in diameter) with sporadic tufa encrustations, very dark greyish brown loamy matrix; 16 – very dark brown shales (clasts 5 cm in diameter), very dark greyish brown loamy matrix.

Przekrój przez trawertyn źródliskowy w Kotyz

1 – gleba typu rędziny ciemnobrązowa, humusowa ze sporadycznymi okruchami wapiennymi i diabazowymi (o średnicy 4 cm); 2 – gleba ciemno żółtobrązowa, lekko humusowa z rzadkimi okruchami wapiennymi (0,5 cm); 2a – gleba ciemno szarobrązowa, lekko humusowa; 3 – gleba ciemnobrązowa lekko humusowa z licznymi inkrustracjami trawertynowymi; 4 – trawertyn drobnoziarnisty jasnoszary, z inkrustacjami w górnej części warstwy i ciemnymi plamami zabarwionymi tlenkami żelaza i manganu; 5 – trawertyn jasno żółtobrązowy; 6 – trawertyn żółtobrązowy z inkrustacjami w górnej części warstwy; 7 – trawertyn ciemno szarobrązowy z rzadkimi okruchami ciemno brązowych łupków sylurskich; 8 – zwietrzały trawertyn gliniasty, ciemnobrązowy z nielicznymi okruchami łupków; 9 – trawertyn masywny, ciemno szarobrązowy z okruchami łupków; 10 – łupki ciemnobrązowe (okruchy o średnicy 1 cm), ciemny szarobrązowy gliniasty matrix; 11 – łupki ciemno brązowe (okruchy o średnicy 2 cm) z inkrustacją trawertynową (30%), ciemnobrązowy gliniasty matrix; 12 – łupki ciemnobrązowe (okruchy o średnicy 2 cm), ze sporadycznymi inkrustacjami trawertynowymi, bardzo ciemny szary matrix; 13 – łupki ciemnobrązowe (okruchy o średnicy 2 cm), ze sporadycznymi inkrustacjami trawertynowymi, bardzo ciemny szarobrązowy matrix; 14 – łupki ciemnobrązowe (okruchy o średnicy 3 cm) z inkrustracją trawertynową (20%), ciemne plamy zabarwione tlenkami żelaza i manganu, ciemnożółty matrix; 15 – łupki ciemnobrązowe okruchy o średnicy 4 cm)), ze sporadycznymi inkrustacjami trawertynowymi, bardzo ciemny szarobrązowy gliniasty matrix; 16 – łupki ciemnobrązowe (klasy o średnicy 5 cm), bardzo ciemny szarobrązowy matrix

(Fig. 2). In Bed 3, the accumulation of tufa is terminated, and the recovered molluscan assemblages from this bed point to a prominent drying and a massive retreat of forests contrasting with a prominent spread of open-landscape elements such as species *Truncatellina cylindrica*, *Pupilla muscorum*, *Vertigo pygmaea* and common representants of the genus *Vallonia*. This trend continues in the overlying strata. Beds 3–1 must be therefore placed to the late Holocene (Subboreal, Subatlantic and Subrecent – Fig. 10), which is also indicated by incursions of modern immigrants of steppe mollusc *Xerolenta obvia* in Bed 2 and ecologically indifferent *Oxychilus cellarius* in Bed 1.

4.2. TUFA CASCADE AT SVATY JAN POD SKALOU

A section in calcareous tufa cascade at Svaty Jan pod Skalou (No. 2 in Fig. 1) represents a significant site for the Holocene stratigraphy and a unique archive of local climate and nature development (Lozek 1967; Kovanda 1971).

The tufa body was partly eroded and quarried in the past. The lower portion of the cascade can be studied in a test-pit excavated in the early 1960s and in small cellars excavated in tufa, while the basal portion of the cascade

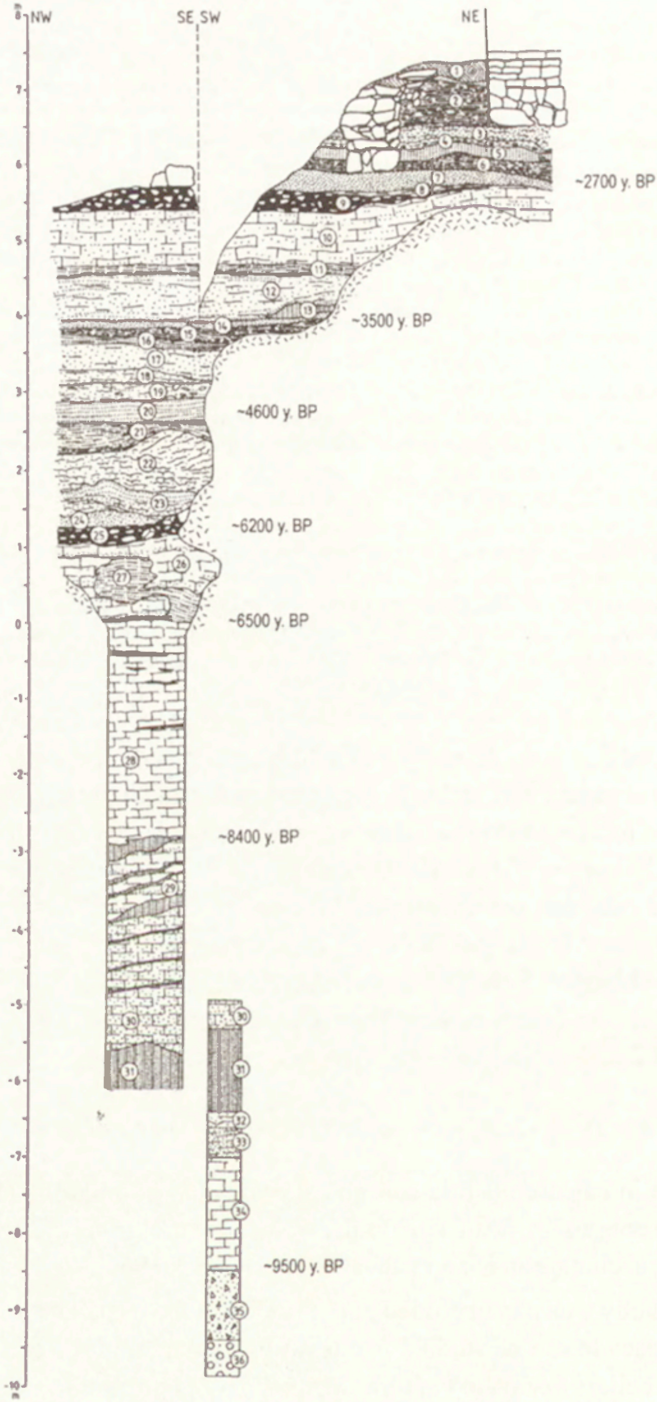
NW
SE SW
NE
~2700 y. BP
~3500 y. BP
~4600 y. BP
~6200 y. BP
~6500 y. BP
~8400 y. BP
~9500 y. BP

Fig. 3. Vertical section in the calcareous tufa accumulation at Svaty Jan pod Skalou based on drilling, excavated pit S1 and outcrop above (Žak et al. 2002)
1 – dark brown humic soil with small limestone and tufa clasts, a bed of tufa debris at the base; 2 – brown humic soil with limestone clasts and a bed of limestone scree; 3 – light brown, extremely loose tufa to brown soil with frequent tufa clasts; 4 – fine- to medium-grained limestone clasts with loamy or tufa matrix; 5 – brown sandy soil, with tufa debris and limestone clasts at the base; 6 – brown sandy soil with tufa clasts; 7 – brown-yellow to yellow loose tufa, locally irregular relics of solid tufa; 8 – yellow solid tufa; 9 – medium- to coarse-grained limestone clasts with loamy and sandy matrix; 10 – grey-yellow partly loose tufa; 11 – grey-yellow extremely loose tufa with brown loamy and sandy bed, with a thin bed of limestone clasts at the base; 12 – light grey solid tufa, in places loose; 13 – brown sandy soil with tufa debris; 14 – light grey solid tufa, in places loose; 15 – dark brown sandy and clayey soil; 16 – grey to brown-grey loose loamy tufa with rusty limonite streaks; 17 – grey to brown-grey strongly loose tufa with limonite streaks, 18 – light grey solid tufa, in places loose; 19 – white grey solid tufa with conspicuous limonite streak on the surface; 20 – light grey fine sandy lacustrine tufa; 21 – light grey, locally rusty brown solid tufa; 22 – light grey to yellow, locally rusty brown solid tufa; 23 – light grey strongly loose tufa; 24 – light brown loose tufa with a lenticular bed of sandy gravel in the upper part; 25 – medium- to coarse-grained limestone clasts with sandy matrix; 26 – light brown loose tufa, locally with loamy matrix; 27 – yellow white solid tufa; 28 – grey-yellow to light grey massive solid tufa; 29 – light grey loose tufa, locally with loam admixture; 30 – greyish yellow to light grey massive solid tufa; 31 – light grey to brown loose tufa, locally with loam admixture; 32 – grey-yellow loose tufa; 33 – loose grey-brown tufa; 34 – grey-yellow to brown-yellow massive solid tufa, more porous at the base; 35 –medium-grained limestone clasts with tufa fragments and encrustations in the upper part of the bed, rounded pebbles of Paleozoic and Proterozoic rocks in the lower part of the bed; 36 – fluvial sandy gravel with well rounded pebbles of Paleozoic and Proterozoic rocks and minor presence of limestone clasts; tufa absent

Pionowy przekrój przez osady trawertynowe w Svaty Jan pod Skalou oparty o wiercenia, odkrywkę S1 i szurf powyżej (Žak i in. 2002)
1 – gleba humusowa, ciemnobrązowa z małymi okruchami wapieni i trawertynów, warstwa okruchów trawertynów w spągu; 2 – gleba humusowa, brązowa z okruchami wapiennymi i warstwą zwietrzeliny wapiennej; 3 – trawertyn bardzo luźny, jasnobrązy; gleba brązowa z licznymi okruchami trawertynu; 4 – okruchy wapienia, drobno- do średnioziarnistych w gliniastym lub trawertynowym matrix; 5 – gleba brązowa, piaszczysta z okruchami trawertynu i wapienia w spągu; 6 – gleba brązowa, piaszczysta z okruchami trawertynu; 7 – trawertyn brązowożółty do żółtego, luźny, lokalnie nieregularne relikty zwięzłego trawertynu; 8 – trawertyn żółty, zwięzły; 9 – okruchy wapienne średnio- do gruboziarnistych z gliniastym i piaszczystym matrix; 10 – trawertyn szarożółty, częściowo luźny; 11 – trawertyn szarożółty, bardzo luźny z brązową gliniasto-piaszczystą warstwą i z cienką warstwą okruchów wapiennych w spągu; 12 – trawertyn jasnoszary, zwięzły, miejscami luźny; 13 – gleba piaszczysta brązowa z okruchami trawertynu; 14 – trawertyn jasnoszary, zwarty, miejscami luźny; 15 – gleba piaszczysta i ilasta, ciemnobrązowa; 16 – trawertyn gliniasty szary do brązowoszarego, luźny z rdzawymi smugami limonitu; 17 – trawertyn gliniasty szary do brązowoszarego, bardzo luźny ze smugami limonitu; 18 – trawertyn jasnoszary, zwięzły, miejscami luźny; 19 – trawertyn, białoszary, zwarty ze smugami limonitu; 20 – trawertyn jeziorny, drobno piaszczysty, jasnoszary; 21 – trawertyn jasnoszary, miejscami rdzawobrązowy, zwarty; 22 – trawertyn jasnoszary do żółtego, miejscami rdzawobrazowy, zwarty; 23 – trawertyn jasnoszary, bardzo luźny; 24 – trawertyn jasnobrązowy, luźny, z soczewkowatą warstwą piaszczystego żwiru w górnej części; 25 – okruchy wapienne średnie do gruboziarnistych w piaszczystym matrix; 26 – trawertyn jasnobrązowy, luźny, miejscami z gliniastym matrix; 27 – trawertyn żółtobiały, zwięzły; 28 – trawertyn szarożółty do

jasnoszarego, masywny, zwięzły; 29 – trawertyn jasnoszary, luźny, miejscami z domieszką gliny; 30 – trawertyn szarożółty do jasnoszarego, masywny, zwięzły; 31 – trawertyn jasnoszary do brązowego, luźny, lokalnie z gliniastymi domieszkami; 32 – trawertyn szarożółty, luźny; 33 – trawertyn szarobrązowy, luźny; 34 – trawertyn szarożółty do brązowożółtego, masywny, zwięzły, w spągu bardziej porowaty; 35 – okruchy wapienne, średnioziarniste z fragmentami trawertynu i inkrustracjami w górnej części warstwy, w dolnej części warstwy otoczaki skał paleozoicznych i proterozoicznych; 36 – piaszczysty żwir rzeczny z dobrze otoczonymi fragmentami skał paleozoicznych i proterozoicznych oraz niewielką ilością okruchów wapiennych, brak trawertynu

below the groundwater table has been studied by drilling in 1996 (Žák et al. 2001).

Lithology

The calcareous tufa cascade consists of a complex of various types of tufas up to 17 m thick (Fig. 3). The tufa cascade is underlain by a sandy gravel fluvial terrace of the near Kačák Stream, characterized by a mixture of rounded pebbles of Upper Proterozoic and Lower Paleozoic rocks and angular limestone clasts. The base of the tufa accumulation lies about 1 m below the present level of the Kačák Stream.

Malacozoology

Malacofauna from the tufa accumulation was synoptically described already by Petrbok (1923). The most recent extensive inter-disciplinary study was provided by Žák et al. (2001). This study gives a complex review of molluscan fauna systematically processed since the 1960s. Complete fossil malacocoenoses from this tufa body were also presented by Žák et al. (2002), who – besides the list of species – also provided a careful description of malacocoenology and the reconstruction of palaeoenvironmental changes.

Chronology

The lower part of the section between the base of solid tufa (~ -9 m) and the test pit bank (0 m) was deposited between ~ 9500 cal. years B.P. and ~ 6500 cal. years B.P. and corresponds to the Late Boreal and most of the Atlantic. Lithologically, most of this section is characterized by massive structural tufa almost free of clastic material. The fossil molluscan assemblage is characterized by high numbers of open-landscape species (open forest with parklands in some places). Nevertheless, already Bed 30 (-5 m, Fig. 3) provided a demanding thermophilous element *Truncatellina claustralis* with Early Holocene elements, such as *Discus ruderatus* and *Perpolita petronella*, occurring in very low numbers.

Loose beds between the solid tufa and the lower scree horizon (Bed 25) were formed, given on geochronological data, between ~ 6500 and ~ 6200 cal. years B.P. This complex includes very rich malacofauna characterized by the appearance of sensitive woodland species and some aquatic species. The lower scree horizon is characterized by the maximum species abundance (46 molluscan species), high number of aquatic and wetland elements.

The thick complex between the lower and upper scree horizons shows a rather wide variety of lithofacies, but its molluscan assemblages show lower species diversity. Aquatic and wetland species remain important and their diversity reaches its maximum in the intercalation of fine-grained grey limnic tufa (Bed 20). The richest woodland assemblage was recorded in Bed 15 (32 species). This part of the section contains several horizons richer in organic matter (up to 0.5% org. C), which represent fossil soils. The best developed soil horizon corresponds to ~ 4200 cal. years B.P. Red-yellow coloured and narrow black zones paralleling fossil soils are a result of post-depositional mobility of Fe and Mn and their redeposition on redox boundaries.

The upper scree horizon (~ 2800 cal. years B.P.) with humic rendzina matrix contains prehistoric pottery fragments belonging to the Late Bronze Age, late stages of the so-called Knovíz Culture (Ložek 1967). Woodland snails remain dominant.

The woodland character of the malacofauna persists in the uppermost, predominantly clastic intervals. Bed 3 also yielded archaeological artefacts attributed to Early Iron Age, the so-called Bylany Culture, dated to the 7th–6th century BC. The tufa formation was terminated at ~ 2400 cal. years B.P., either because climate changed or the spring was relocated to the base of the tufa deposit and started to erode its base. This could have been supported by erosive activity of the Kačák Stream. As a result of these processes, a cave (called Ivan's Cave) was formed in the tufa body. An original presence of small cavities and caves cannot be ruled out. Unfortunately, the cave was artificially enlarged several times and its original walls were mostly destroyed.

4.3. TUFA CASCADES IN THE CÍSARSKÁ GORGE

The Císarská Gorge is a valley 700 m long and over 100 m deep (Kadlecová, Žák 1998). Tufa is intensively precipitated from waters of a spring in the upper part of the valley. Tufa forms three cascades, marked by numbers I, II and III in the upstream direction. Only Cascades I and II were subjects of this study (No. 3 in Fig. 1).

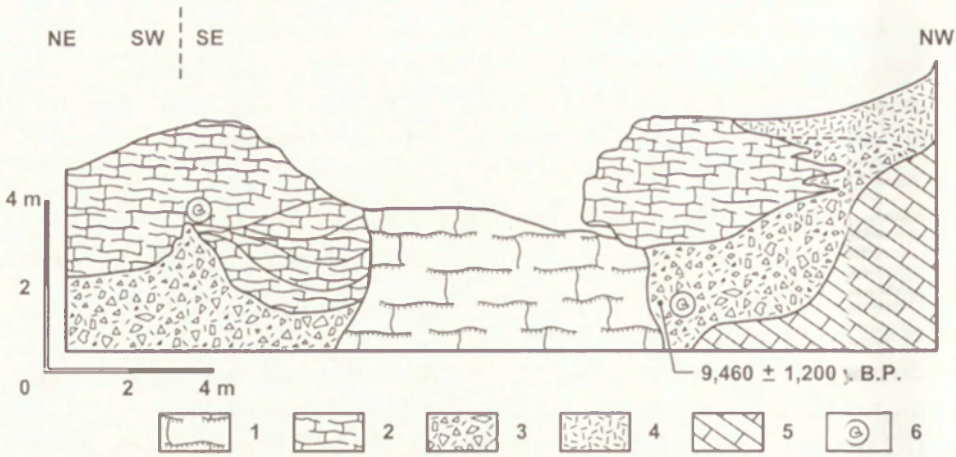

Fig. 4. Tufa cascade I in the Císarska Gorge. 1 – the youngest calcareous tufa; 2 – massive porous brown yellowish calcareous tufa; 3 – limestone scree cemented by brown porous carbonate, angular clasts with average size of 4 cm; clast size up to 10 cm in the upper part; 4 – grey loose limestone scree; 5 – limestone bedrock; 6 – sampling point on fossil molluscs

Trawertyn kaskadowy I z Wąwozu Cisarska. 1 – najmłodszy trawertyn węglanowy; 2 – trawertyn węglanowy brązowożółty masywny, porowaty; 3 – zwietrzelina wapienna scementowana przez okruchy węglanowe brązowe, porowate, ostrokrawędziste, o średnicy 4 cm; w górnej części okruchy o średnicy do 10 cm; 4 – zwietrzelina wapienna szara, luźna; 5 – podłoże wapienne; 6 – miejsca poboru próbek kopalnych mięczaków

Lithology

Limestone scree cemented by carbonate is deposited on the base of the thickest cascade – Cascade I. Above the scree, a bed of solid structural tufa was deposited, interfingering with slope sediments laterally. A young tufa accumulation is presently growing in the erosive channel transecting the cascade (Figs 4 and 5). Cascade II, lying ca. 130 m from here, has an analogous lithological character (Fig. 6).

Malacozoology

In both cascades, samples corresponding in their level with radiometrically dated beds were selected for malacozoological and malacostratigraphic analysis (see sampling points in Fig. 4 and Fig. 5). Fossil molluscs from the bed of massive porous brown yellowish calcareous tufa from Cascade I (Fig. 4) are markedly dominated by woodland species. Gastropods *Sphyradium doliolum*, *Platyla polita*, *Vertigo pusilla*, *Aegopinela pura*, *Discus rotundatus*, *Macrogastra ventricosa*, *Alinda biplicata*, *Clausilia pumila* and other species were found to be very abundant. These are accompanied by woodland species of moderate size (*Monachoides incarnatus*, *Urticicola umbrosus*, *Heli-*

Fig. 5. Tufa cascade I in the Císarská Gorge – the youngest tufa accumulation growing in the erosional channel formed in an older cascade. Photo by P. Zajiček

Trawertyn kaskadowy I z Wąwozu Císarská – akumulacja najmłodszego trawertynu w kanale erozyjnym utworzonym w starszej kaskadzie. Fot. P Zajiček

codonta obvoluta, *Cepaea hortensis*), again highly abundant. Gastropods with non-specific ecological affinity (ecologic group C) reach higher abundance than open-landscape and forest-free elements, which are considerably less frequent and restricted only to species *Vallonia costata* and rare *Truncatell-ina cylindrica*, probably redeposited from the ambient steep slopes. Fossil malacocoenosis clearly indicates a moist, continuous forest covering valley areas, which was getting lighter towards higher elevations.

A poor molluscan assemblage clearly dominated by *Vallonia costata*, *Punctum pygmaeum* and *Vitrea crystallina* was ascertained in the underlying bed of limestone scree cemented by brown porous carbonate from Cascade I (Fig. 4). These undemanding species were occasionally accompanied by the Early Holocene element of *Discus ruderatus* and ecologically indifferent *Cochlicopa lubrica* together with *Euconulus fulvus*. The proportion of woodland species in this bed was very low: they were limited only to the species of *Monachoides incarnatus* and *Platyla polita*. A similar composition of fossil malacofauna was also observed in the sample collected from the bed of limestone scree cemented by brown porous carbonate from Cascade II (Fig. 6). The common *Discus ruderatus* and *Perpolita petronella* are followed by gastropods *Vertigo alpestris*, *Discus rotundatus*, *Aegopinella minor*, *Trichia hisp-*

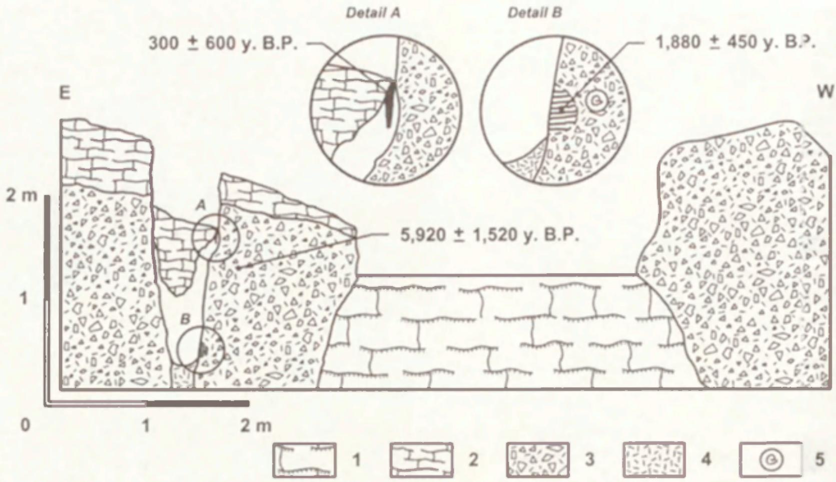

Fig. 6. Tufa cascade II in the Cisarska Gorge. 1 – the youngest calcareous tufa; 2 – massive porous yellowish brown calcareous tufa; 3 – limestone scree cemented by brown porous carbonate, angular clasts with average size of 4 cm; clast size up to 10 cm in the upper part; 4 – grey loose limestone scree; 5 – sampling point or fossil molluscs. Detail "A" shows a bed of dated carbonate precipitated on plant roots. Detail "B" shows the position of carbonate laminae precipitated in a small cavity developed in limestone scree

Trawertyn kaskadowy II z Wąwozu Cisarska. 1 – najmłodszy trawertyn węglanowy; 2 – trawertyn węglanowy masywny, porowaty żółtobrązowy; 3 – zwietrzelina wapienna scementowana przez brązowe, porowate, ostrokrawędziste węglanowe okruchy o średnicy 4 cm; w górne części okruchy o średnicy do 10 cm; 4 – szara luźna zwietrzelina wapienna; 5 – miejsca poboru próbek kopalnych mięczaków. „A" przedstawia warstwę datowanych wytrąceń węglanowych na korzeniach roślinnych. „B" przedstawia położenie warstwy węglanowej wytrąconej w małej pustce rozwiniętej w zwietrzelinie wapiennej

ida and *Ena montana*, accompanied by indifferent species *Punctum pygmaeum*, *Cochlicopa lubrica* and *Euconulus fulvus*. Both of these beds are completely lacking any of the prominent elements of continuous forest characteristic for the Holocene climatic optimum or the youngest Holocene.

Chronology

As indicated by molluscan remains in the beds of limestone scree cemented by brown porous carbonate from Cascades I and II, their fill can be dated to the Early Holocene. This is suggested not only by the common occurrence of Early Holocene species *Discus ruderatus*, *Perpolita petronella*, *Vitrea crystallina* and even *Vertigo alpestris*, but also by the notable absence of elements indicative of a continuous forest of the climatic optimum. The Early Holocene age of the scree is also confirmed by a radiometric age of its carbonate matrix (Żak et al. 2003) – 9,460 ± 1,200 years BP (sample TMS-1) – Table 1 and

Table 1. U-series age data of carbonates from tufa cascades in the Císařská Gorge.

Location	Sample	Lab. No.	U content (ppm)	Error (ppm)	$^{234}U/^{238}U$	Error	$^{230}Th/^{234}U$	Error	$^{230}Th/^{232}Th$	Age (without correction) ka	Error ka	Corrected age ka	Error ka
Cascade I	TMS-1	W 743	2,363	±0.042	1,367	±0.018	0,097	±0.002	9,7	11,09	±0.25	**9,46**	±1.20
Cascade II	TMS-5	W 744	0,786	±0.015	1,985	±0.032	0,020	±0.001	10	2,21	±0.15	**1,88**	±0.45
Cascade II	TMS-28	W 914	1,162	±0.021	1,876	±0.034	0,068	±0.003	6,4	7,65	±0.34	**5,92**	±1.52
Cascade II	TMS-29	W 917	1,789	±0.032	1,735	±0.028	0,008	±0.001	2,3	0,86	±0.09	**0,30**	±0.60

Fig. 4. The encountered assemblages rather indicate the Boreal period (sample from Cascade I), or possibly a transition between the Boreal and early phases of the Atlantic (sample from Cascade II – with carbonate matrix dated at 5,920 ± 1,520 years BP – sample TMS-28, see Tabule 1 and Fig. 6), when the landscape had the character of open spaces alternating with small areas of light poor deciduous forests having, however, a different composition than the present ones. Moisture conditions were more favourable already but still did not permit a development of more continuous woodland complexes, as documented by the notable absence of demanding woodland molluscan species or only a low proportion of hydrophilous elements.

In contrast with the underlying beds, malacostratigraphic analysis of a sample from the bed of massive porous brown yellowish calcareous tufa from Cascade I (Fig. 4) yielded malacofauna indicating the Holocene climatic optimum – Atlantic and Epiatlantic (*sensu* Jäger, 1969). This is evidenced by almost a complete absence of open-landscape elements as well as by the absence of Early Holocene elements or elements from the youngest phases of the Holocene.

In a time succession, the Císařská Gorge was first filled with talus with angular scree fragments in the Early Holocene (Preboreal, Boreal to early phases of the Atlantic). The scree was later overlain by accumulations of structural tufa precipitated during the Holocene climatic optimum (Atlantic, Epiatlantic).

4.4. PETRÁNKA TUFA CASCADE IN THE KARLICKÉ VALLEY

The tufa accumulations (No. 4 in Fig. 1) form relics in the banks of a small gorge 12 m deep, with a karst spring in its closure (Kadlecová, Žák 1998).

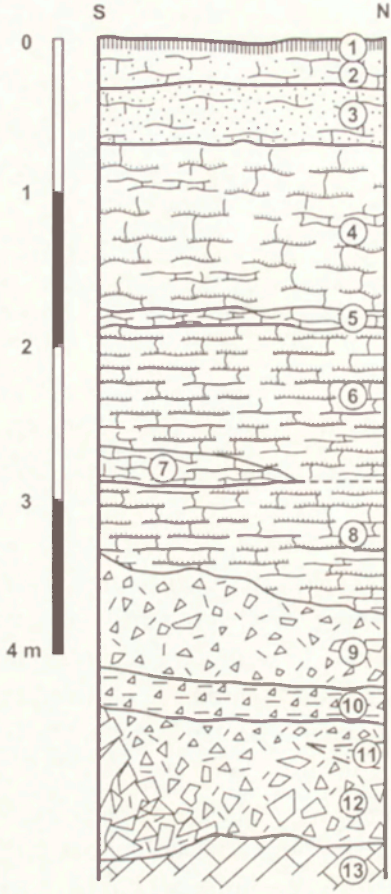

Fig. 7. A relic of the tufa cascade exposed near the Petránka Spring in the Karlické Valley. 1 – dark brown humic Rendzic Leptosol; 2 – loose grey tufa, slightly humic; 3 – loose grey tufa; 4 – a complex of light brown loose tufa with thin beds of solid tufa; 5 – lens-like beds of solid tufa with dark stains coloured with Fe and Mn oxides; 6 – loose, light yellow-brown tufa; 7 – a lens-like bed of solid tufa; 8 – loose, light yellow-brown tufa with sporadic limestone and shale clasts (0.5 cm in diameter); 9 – scree horizon formed by shale plates and limestone clasts up to 20 cm large in a silt-sand matrix; 10 – brown sandy silt with rare limestone clasts; 11 – yellow-brown silt; 12 – coarse scree horizon formed by limestone blocks up to 70 cm large in a silt-sand matrix; 13 – limestone bedrock

Relikt trawertynu kaskadowego odsłonięty w pobliżu Źródła Petranka w Dolinie Karlicke.
1 – rędzina ciemno brązowa; 2 – trawertyn lekko humusowy, szary, luźny; 3 – trawertyn, szary luźny; 4 – kompleks trawertynu szarobrązowego, luźnego z cienkimi warstwami trawertynu zwartego; 5 – soczewkowate warstwy trawertynu zwartego z ciemnymi plamami zabarwionymi tlenkami żelaza i manganu; 6 – trawertyn jasno żółtobrązowy, luźny; 7 – soczewkowata warstwa trawertynu zwartego; 8 – trawertyn jasno żółtobrązowy, luźny z pojedynczymi okruchami wapieni i łupków o średnicy 0,5 cm; 9 – warstwa zwietrzeliny utworzona przez płyty łupków i okruchy wapieni o wielkości do 20 cm w pylasto-piaszczystym matrix; 10 – mułek piaszczysty, brązowy z rzadkimi okruchami wapiennymi; 11 – mułek żółto-brązowy; 12 – warstwa zwietrzeliny utworzona przez bloki o wielkości do 70 cm w pylasto-piaszczystym matrix; 13 – podłoże wapienne

Lithology

The basal part of the tufa succession is formed by a bed of coarse limestone clasts, filling a small depression in the underlying dark Paleozoic limestone. The scree is stratified by beds of dark silt with frequent shale clasts. The scree is overlain by a complex of loose tufa with beds and lenses of yellow-brown massive structural tufa. The upper portion of the section is formed by loose grey tufa with Rendzic Leptosol on the surface. For a detailed description of the section – Fig. 7.

Malacozoology

The basal part of the section (Beds 12, 11 and 9) contains no molluscan fauna.

The first fauna (*Trichia sericea*) was identified in a bed formed by brown sandy silt (Bed 10) overlain by a bed of slope scree (Ložek, Kadlec 2000). Bed 8 (loose light yellow-brown tufa with sporadic limestone and shale clasts) yielded rich fossil malacofauna dominated especially by *Carychium tridentatum*, *Sphyradium doliolum*, *Aegopinella pura*, *Aegopinella minor*, *Discus rotundatus* and *Columella edentula*, accompanied by frequent *Cochlicopa lubricella* and by aquatic gastropods *Galba truncatula* and *Anisus leucostoma*. Altogether 23 molluscan species were identified in this horizon. Similar malacofauna was also encountered in the overlying tufa horizons, showing only very slight differences in the compositions of fossil malacocoenoses. While Beds 7–5 are characterized by absolute absence of open-habitat and steppe elements, *Vallonia costata* appears in Bed 4 and belongs among the dominant species in Bed 3 already. This indicates partial drying and deforestation during the deposition of this horizon. The uppermost beds are marked by the appearance of *Truncatellina cylindrica* (Bed 2) and the modern immigrant *Oxychilus cellarius* (Bed 1).

The whole tufa interval (Beds 8–2) yielded relatively rich woodland malacofauna, which has not been impoverished to a higher degree but has survived until the present days with only slight species modifications (e.g., gastropods *Sphyradium doliolum*, *Aegopinella pura*, *Macrogastra ventricosa*, *Cochlodina laminata*, *Alinda biplicata*, *Ena montana*, *Urticicola umbrosus*). This molluscan assemblage indicates warm and light forest environment.

Chronology

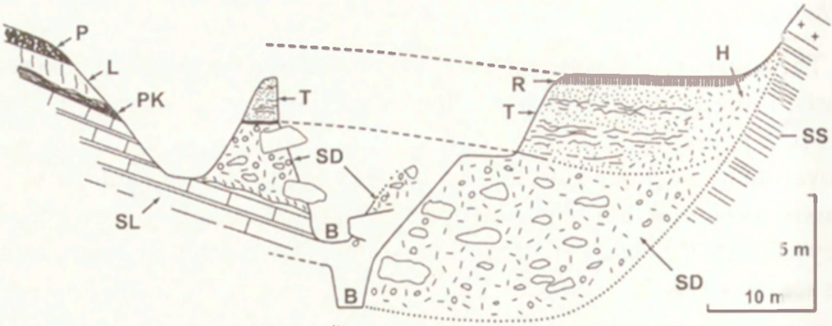

Fig. 8. Cross-sections through the Čertova Gorge at Mala Chuchle (Ložek, Jag:r 1968). SL – Silurian limestones, SS – Silurian shales and basaltic volcanic rocks, 'D – last Glacial slope deposits, P – Holocene Haplic Luvisol, L – last Glacial loess, Pk – Interglacial soil sediment, R – Holocene Rendzic Leptosol, T – Holocene tufa accunulation, H – Holocene talus, B – valley bottom

Przekrój przez Wąwóz Čertova w Mala Chuchle (Ložek, Jäger 1968). SL – wapiene syluru, SS – lupki i wulkanity syluru, SD – osady stokowe ostatniego zlodowacenia, P – hooceńskie luwisole, L – lessy ostatniego zlodowacenia, PK – interglaojalna gleba kopalna, R - rędziny holoceńskie, T – trawertyny holoceńskic, H – zbocze holoceńskie, B – dno doliny

As indicated by the molluscan assemblages, tufa deposition probably started during the Boreal period and continued to the Atlantic and Epiatlaıtic climatic optimum. As indicated by fossil malacocoenoses in Beds 3 ard 2, the existing woodland complexes maintaining a closed character up to this level increased the proportion of lighter habitats at this boundary.

The basal beds of limestone scree may represent the Preboreal coller conditions with intensive mass wasting. After the termination of tufa deposition, the cascade was incised by running water down to the underlying linestone.

4.5. TUFA CASDADES IN THE ČERTOVA GORGE AT MALA CHUCHLE

A tufa cascade (No. 5 in Fig. 1) was originally barring the Čertovı Gorge in the width of 30–40 m. As a result of later erosion, tufa relics now le at the height of up to 12 m above the present gorge bottom.

Lithology

The bottom of the gorge is filled with slope deposits – coarse linestone scree with sandy and silty matrix transported by water and mass movements from the slopes probably during the end of the last glacial stage (Fig.8). The tufa cascade was deposited on these slope deposits. The studied section is exposed in the northern slope of the Čertova Gorge. Beds of grey tobrown-

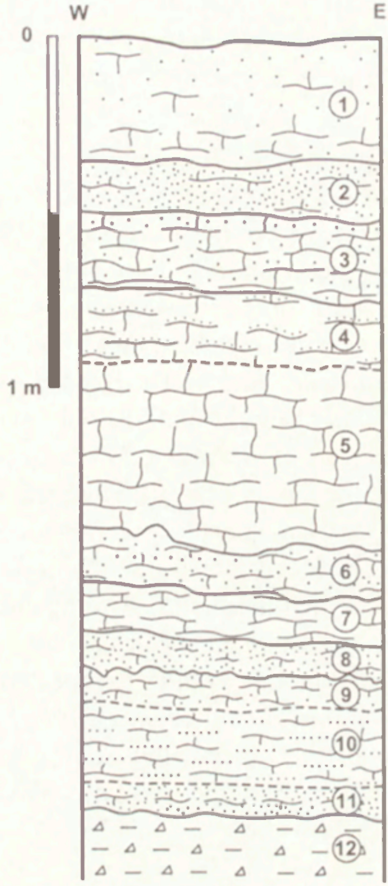

Fig. 9. A section in the Epiatlantic to Subboreal calcareous tufa exposed in the northern slope of the Čertova Gorge. 1– brownish grey loose tufa with rusty limonitic stains in the lower portion of the bed; 2 – brownish grey loose fine tufa; 3 – greyish brown partly loose leaf tufa; 4 – light grey medium-grained loose tufa; 5 – yellowish grey, almost solid structural porous tufa with horizons with leaf encrustations; 6 – brownish grey loose tufa with encrustations; 7 – brownish grey, partly loose leaf tufa; 8 – light grey fine-grained loose tufa; 9 – greyish yellow tufa with a rusty limonitic band on the surface; 10 – light grey fine-grained loose tufa with a rusty limonitic band; 11 – dark grey fine-grained loose tufa with dark stains; 12 – dark brown clayey silt with pebbles and rock fragments – slope deposits

Przekrój przez trawertyn węglanowy (późny Atlantyk – Subboreal), odsłonięty w północnym stoku Wąwozu Čertova. 1 – trawertyn brązowawoszary, luźny z rdzawymi plamami limonitu w dolnej części warstwy; 2 – trawertyn drobnoziarnisty, brązowawoszary, luźny; 3 – trawertyn szarobrązowy, częściowo luźny; 4 – trawertyn średnioziarnisty, jasnoszary, luźny; 5 – trawertyn żółtawoszary, prawie zwarty, porowaty z warstwami z inkrustowanymi liśćmi 6 – trawertyn brązowawoszary, luźny z inkrustracją; 7 – trawertyn brązowawoszary, częściowo luźny; 8 – trawertyn drobnoziarnisty, jasnoszary, luźny; 9 – trawertyn szarawożółty z rdzawymi nalotami limonitu na powierzchni; 10 – trawertyn drobnoziarnisty, jasnoszary, luźny z rdzwym nalotem limonitu; 11 – trawertyn drobnoziarnisty, ciemnoszary, luźny z ciemnymi plamami; 12 – mulek ilasty ciemno brązowy z otoczakami i fragmentami skał – osady stokowe ish grey, mostly loose tufa form a tufa succession underlain by slope deposits. For a detailed description of section – Fig. 9.

Malacozoology

A tufa section on the left valley bank provided malacozoological material of extremely high species diversity consisting of species of broad ecologic range. Woodland species *Monachoides incarnatus*, *Discus rotundatus*, *Sphyradium doliolum* and *Acanthinula aculeata* occur almost continuously in the whole interval, being accompanied by hydrophilous species *Carychium tridentatum* and *Zonitoides nitidus*. The finds of wetland species *Vertigo angustior* and *Vertigo antivertigo* well correspond with the occurrences of aquatic to paludal species *Galba truncatula*, *Radix peregra* and *Succinea putris*, which indicates local presence of limited wetlands and small pools. These were surrounded to a large degree by continuous woodland formations of small areal extent, as evidenced by the common occurrence of woodland and wetland species. On the other hand, also the common occurrences of woodland species with open-habitat species should be taken into account. Among them, gastropod *Vallonia costata* was repeatedly encountered, being accompanied by *Vallonia pulchella*, *Truncatellina cylindrica* and *Pupilla muscorum*. Malacocoenological spectra thus show many indications of alternating changes in the apperance of the environment, which is typical for the Epiatlantic and Subboreal periods.

Chronology

The oldest portion of the tufa body is preserved in a small relic in the southern slope of the valley as indicated by the molluscan assemblages (Ložek, Jäger 1968). The molluscan assemblage dates tufa deposition to the younger stage of the Holocene – the Epiatlantic and Subboreal (*sensu* Jäger, 1969), dominated by wet forest environment. Younger parts of the tufa cascade are not preserved, maybe due to younger erosion, when running water eroded the tufa cascade. The youngest Holocene incision of the Čertova Gorge reached the underlying Late Pleistocene slope sediments. The spring now resurges near the base of the tufa body due to subrosion.

4. DISCUSSION ON THE DEVELOPMENT AND DESTRUCTION OF TUFA CASCADES IN THE BOHEMIAN KARST

4.1. CASCADE DEVELOPMENT

Holocenc development of tufa cascades of the Bohemian Karst shares many common features conditioned by climatic and hydrologic influence. The formation of cascades, which were building on the bottoms of karstic valleys, must have postdated the end of intensive fluvial erosion characteristic for the Glacial/Holocene boundary (Vandenberghe 1993; Vandenberghe et al. 1994, a.o.). The tufa is usually underlain by limestone talus derived from the ambient slopes. Thicknesses of preserved talus depends largely on the character of source Paleozoic rocks forming valley slopes (e.g., stratification and tectonic deformation) and on the erosive potential of the stream at the site the cascade started to develop. Molluscan assemblages found in talus beneath the tufa evidence Boreal age of these colluvia.

Talus in the Cisarska Gorge is secondarily cemented by carbonate precipitated from karst waters. Radiometric age of the cement was determined at 9,460 ± 1,200 years BP and 5,920 ± 1,520 years BP – see Table 1 and Figs 4 and 6. The tufa cascade at Svaty Jan pod Skalou also started to accumulate in the Boreal (Žak et al. 2001, 2002). Massive structural tufa with minimum clastic component were deposited on all cascades under climatically favourable condition of the Atlantic. Accumulations of these structural tufas reach up to 2–5 m in thickness (Kovanda 1971; Lozek 1992). Deposition of tufa continued in the late period of the Holocene climatic optimum (Epiatlantic). Short-term climatic oscillations, however, occurred in the Epiatlantic, being characterized by alternation of periods of relative aridity and humidity. These short-term climatic oscillations are indicated by horizons of initial carbonate soils and by talus intercalations in the tufa successions. Climatic oscillations culminated by markedly dry and warm period in the Subboreal (Jager 1969). As a result, sedimentary record in younger portions of the tufa cascades is lithologically more variable, with alternating beds of friable and compact tufa, colluvio-fluvial sediments and soil horizons (Zak et al. 2001). Principal accumulation stages of tufa cascades in the Bohemian Karst are terminated in the late Subboreal at 2,500 BP (Zak et al. 2002).

Molluscan assemblages detected in the tufa bodies show many common features. Malacospectral analyses of tufas at Kotyz, Svaty Jan pod Skalou and Petranka in the Karlicke Valley revealed the proportions of molluscs of the main ecologic groups (Ecologic group: A – woodland species in general, B

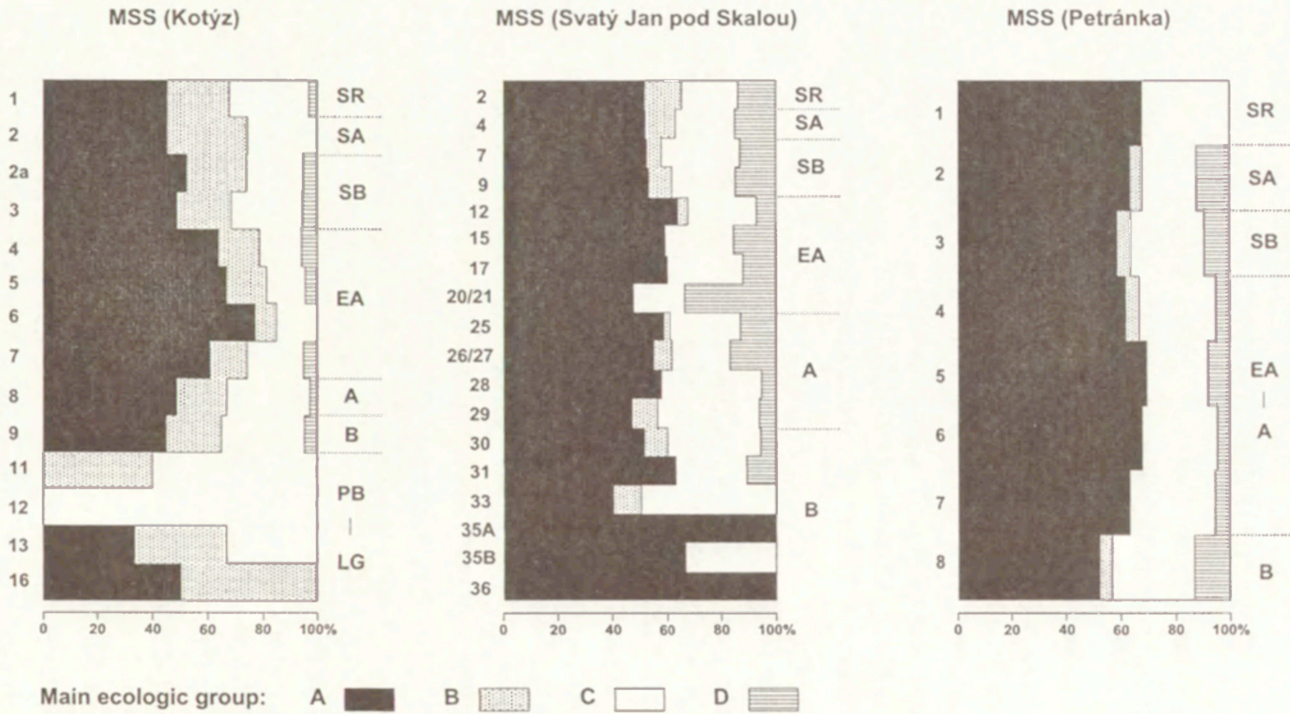

Fig. 10. Malacospectra of species (MSS) showing the proportions of the main ecologic groups based on Ložck (1964)
A – woodland species (in general), B – open ground species (in general), C – indifferent species, D – marsh and aquatic species. Numbering of beds on the left, chronology (based on molluscan assemblages and radiometric data on the right). Chronology: LG – Late Glacial, PR – Preborcal, B – Boreal, A – Atlantic, EA – Epiatlantic, SB – Subboreal, SA – Subatlantic, SR – Subrecent

Malacospcktra gatunków (MSS) przedstawiające wzajemny stosunek głównych grup ekologicznych na podstawie Ložka (1964)

A – gatunki leśne (ogólnie), B – gatunki terenów otwartych (ogólnie), C – gatunki obojętne, D – gatunki torfowiskowe i wodne. Numeracja warstw z lewej, chronologia (na podstawie zespołów mięczaków i danych radiometrycznych) z prawej. Chronologia: LG – późny glacjał, PR – Preboreał, B – Boreał, A – Atlantyk, EA – Epiatlantyk (późny Atlantyk), SB – Subboreał, SA – Subatlantyk, SR – współczesność

– open-landscape species in general, C – indifferent species, D – marsh and aquatic species) – Fig. 10. As the underlying intervals of the tufa bodies at Kotyz (Beds 16–11) and Svaty Jan pod Skalou (Beds 36–31) contain very poor malacozoological material, the ratio between the woodland malacofauna component and open-habitat component or indifferent species is biased to a considerable degree. In contrast, the overlying beds contain high numbers of molluscan species already and are fully eligible for statistical evaluation. As shown by the malacospectra, the overlying beds of the Atlantic–Epiatlantic climatic optimum marked by intensive tufa accumulation are dominated by woodland species in the malacocoenoses, while open-habitat species are scarcely represented (Svaty Jan pod Skalou, Kotyz) or present in very low proportions (Petranka). This indicates continuous woodland formations with prevalence of hydrophilous molluscan species. In the Epiatlantic phase, climatic oscillations culminate with the Subboreal period characterized by suppressed woodland component and the onset of open-habitat elements. This is particularly obvious from the malacospectra of tufas from Svaty Jan pod Skalou and Kotyz. In the Petranka tufa in the Karlicke Valley, the Subboreal period is characterized by the abundant occurrence of gastropod *Vallonia costata*, a typical open-habitat species; however, the woodland component still maintains a prominent dominance. Molluscan assemblages from the youngest Holocene phases often share the features of modern communities as evidenced by molluscan occurrences in the neighbourhood of the tufa bodies (Lozek 1974; Hlavač 2002).

4.2. TERMINATION OF CASCADE DEVELOPMENT

Termination of cascade development was due to the coming dry period, probably manifested in a discharge reduction of the karst springs. Reduced discharges of the issued water could have resulted in subrosion, which created new paths in the basal portions of the tufa bodies. Karst waters ceased to flow on top of the cascade surfaces. The onset of this dry period dates to the Late Bronze Age; it resulted in dramatic changes in vegetation and in molluscan communities, but also in a decrease in the activity of river systems. These phenomena have been documented from many regions of the Czech Republic. A major environmental change dating to the Late Bronze Age has been documented from central and northern Bohemia on the basis of multidisciplinary study of sediments deposited under sandstone rockshelters. Changes in molluscan communities indicate deforestation associated with more intensive agricultural use of land (Svoboda et al. 1996; Cilek et al.1996).

A sudden vegetational change in the Late Bronze Age is also indicated by pollen spectra from sediments filling an ox-bow lake of the Labe River near Tisice (15 km N of Prague). Such a change probably reflects increased anthropogenic stress on the environment (deforestation connected with agricultural activity) controlled by changing climatic conditions (Pokorny, in print). Given on archaeological evidence of settlement on the Klejnarka Stream flood plain (60 km SE of Prague), Pavlu (2002) identified a period of severe floods (3,500–2,500 BP), followed by a period of relative fluvial stability.

4.3. DESTRUCTION OF CASCADES

Erosion responsible for the destruction of all large tufa cascades in the Bohemian Karst must have been associated with a period extremely rich in flash precipitation events. Such period was characterized by increased discharges of streams but also of karst springs. The onset of erosion and cascade destruction is constrained by radiometric ages of carbonate laminae precipitated in a small cavity in limestone talus of Cascade II in the Cisarska Gorge. These laminae were formed at 1,880 ± 450 years BP (sample TMS-5, see Table 1 and Fig. 6), prior to erosion, when the cavity was closed. Only a part of the carbonate filling of this cavity has been preserved: on the surface of a dilated fracture formed by detachment of a part of the cascade due to erosion and destruction of the body (Fig. 6).

Tufa cascades of the Bohemian Karst were destructed in the period of anomalous precipitation and increased fluvial activity. This is a characteristic obviously fitting the Little Ice Age (LIA) – one of the coldest periods in the whole Holocene (Bradley et al. 2003). Climate deterioration signaling the onset of the LIA in Bohemia was marked by a lowering of average winter temperatures and an increase of precipitation in the half of the 15th century (Brazdil 1996), and consequently by an increased incidence of flood events as documented by higher incidence of floods on the Berounka River (Ninger, Zelinka 1872), which flows across the Bohemian Karst area (Fig. 1). Increased flood activity of the Labe River in central Bohemia has been documented by radiometric dating of a stump of a tree growing on top of fluvial sediments of the Labe River near Lzovice. The age of the tree was set at 541 years BP (Šilar et al. 1994). It is significant that the tree stump is overlain by flood sands up to 3 m thick dating to the younger period of the LIA (Ruzickova, pers. comm. 2003). The period of intensive fluvial erosion, deepening the beds of minor streams by as much as several metres in the LIA and in the 20th century, has been well documented from Great Britain and from Crete (Macklin et al. 1992; Maas et al. 1998). In both studied catchments, the authors give considerable significance not only to the climate deterioration but also to the anthropogenic land-use effect resulting in a weakened ability of land to retain water at extreme precipitation events.

It can be assumed that erosion and destruction of tufa cascades in the Bohemian Karst occurred during several flash precipitation events in the LIA and due to increased precipitation and severe floods in the 20th century. Larger volumes of water flowing through the Cisarska Gorge during the LIA is also indicate by the radiometric age of the carbonate precipitated on plant roots in Cascade II (sample TMS-29, see Table 1 and Fig. 6). The carbonate was precipitated on the root at 300 ± 600 years BP, i.e., at the time of undoubted existence of the fracture destructing the cascade (see above). At that time, however, water must have been still flowing across the upper part of the cascade.

The rate of erosion of sedimentary fill of a valley by running water can be observed on several examples. In 1995, extremely high precipitation resulted in the destruction of tufa layers up to 1 m thick in the valley of the Bubovicky Creek 2 km NE of the Cisarska Gorge (Zak et al. 1996). Second example of modern erosion is located in a small side valley at Dolni Roblin Village some 300 m N of the spring and the Petranka Cascade. This side valley was barred by an earth dam more than 80 years ago. A small reservoir formed upstream of the dam holding water to run the local mill. The dam was destructed during a severe flood of 1953, and a new channel was formed by headward erosion in the loamy fill of the valley. The new channel was considerably deepened due to the anomalous precipitations preceding the flood of 2002. Now, an erosive cut 3.8 deep is present at the site of the former dam (the stream incised by 2 m in August 2002). Strong erosion associated with the flood of 1953 also altered Cascade I in the Cisarska Gorge: the erosive cut in the cascade as shown on a photograph of Petrbok (1955) taken before 1950 has a markedly narrower shape than it has today. Moreover, the young accumulation of tufa now depositing in the streambed (Fig. 5) is missing on the old photograph.

At present, accumulation of tufa prevails over its erosion at most localities in the Bohemian Karst area. New tufa bodies are being formed on eroded cascades in the Cisarska Gorge and in the Čertova Gorge at Mala Chuchle. Present tufa formation has been reported from many other places in the Bohemian Karst (Kadlecova, Zak 1998). The rate of present tufa deposition can be observed at Mala Chuchle. An artificial streambed was constructed here, the walls of which were fixed with stone blocks in the late 1980s. Tufa is precipitated from karst water flowing through this channel. Since the time of this reconstruction, a tufa accumulation up to 0.6 m thick was built in the channel. It is, therefore, highly probable that the youngest tufa in the Cisařska Gorge cascades was deposited during the last 50 years, thus representing a period of weaker fluvial and erosive activity.

5. CONCLUSIONS

1. Holocene tufa cascades of the Bohemian Karst were growing in the Boreal to Late Subboreal period (9,500–2,500 years B.P.). The end of their formation is marked by the onset of a dry period in the Late Bronze Age. This climatic change accelerated agricultural activity associated with vegetational changes (deforestation). These processes have been documented at many sites in central Bohemia.

2. Erosion and destruction of tufa cascades continued in a period of frequent flash precipitations. The local erosion of some cascades in the Bohemian Karst in the last years makes us to assume that most cascades of the Bohemian Karst were destructed in the recent past – probably during the LIA. In this perspective, LIA seems to be the period of maximum erosion intensity within the whole of the Holocene.

3. The last 50 years (before 1997) in Bohemia were marked by the absence of floods and by a relatively low intensity of fluvial erosion. As a result, new tufa accumulations grew on some of the cascades.

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TWORZENIE I NISZCZENIE HOLOCEŃSKICH KASKAD TRAWERTYNOWYCH W KRASIE CZESKIM (REPUBLIKA CZESKA)

Streszczenie

Badaniami objęto pięć holoceńskich kaskad trawertynowych położonych w Krasie czeskim. Stwierdzona zmienność litologiczna trawertynów oraz zmienność zespołów malakologicznych pozwoliły na określenie warunków powstawania i późniejszego niszczenia osadów. Skala czasu dla zapisanych w osadach zdarzeń oparta została na badaniu zespołów malakologicznych oraz izotopowym datowaniu materii organicznej metodą radiowęglową i węglanu wapnia metodą uranowo-torową w trzech stanowiskach. Badane kaskady trawertynowe powstawały w okresie boreal – późny subboreał (9 500–2 500 lat BP). W profilach osadów zapisały się wahania klimatyczne. W okresie optimum klimatycznego powstawały masywne trawertyny z nieznacznymi jedynie domieszkami materiału klastycznego. Litologia młodszych części profili jest zdecydowanie bardziej zróżnicowana. Okresowo powstawały warstwy ze zdecydowanie większą zawartością domieszek klastycznych lub poziomy glebowe. Zmiany klimatyczne zapisały się także w zmienności zespołów malakologicznych. W epoce późnego brązu, w efekcie zmian klimatycznych powodujących spadek wydajności źródeł, tworzenie trawertynów zostało zahamowane. Osuszenie klimatu widoczne w wielu profilach Środkowych Czech przyśpieszone było przez wzrost aktywności rolniczej i przez związane z nim odlesienie terenu. Niszczenie kaskad rozpoczęło się w okresie po 2 000 lat BP. Związane było ono z okresami anomalnie wysokich i gwałtownych opadów i powodowaną przez nie wzmożoną aktywność rzek. Najprawdopodobniej w okresie małej epoki lodowej niszczenie kaskad osiągnęło największą intensywność. W okresie ostatnich 50 lat (przed 1997) w Czechach nie były notowane istotne powodzie i intensywność erozji rzecznej była stosunkowo niska. W rezultacie, lokalnie, nastąpiło wznowienie depozycji trawertynów w badanych stanowiskach. Jednak w wielu stanowiskach obserwować można ślady niszczenia trawertynów w okresach wzmożonych opadów i powodzi z ostatnich lat.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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HOLOCENE FACIES DEVELOPMENT IN MID-EUROPEAN UPLANDS

1. INTRODUCTION

The name "Holocene" or "recent whole" for the post-glacial epoch of last 12 thousand years seems to have been proposed for the first time by Sir Charles Lyell in 1833, and adopted by the International Geological Congress in Bologna in 1855. However Lyell himself used the term "Recent" which was by U.S. Geological Service abandoned as late as in 1968. The Holocene stratigraphy was based on the succession of vegetation that was originally studied on the area of Denmark and adjacent Southern Scandinavia (Iversen 1973; Farrand 1990). The changes of vegetation were initially recorded on the basis of macroscopical remnants preserved in peatbogs as a sequence of four distinct climatic periods, namely dry Boreal, humid Atlantic, dry and warm Subboreal and more humid and colder Subatlantic concluded by Recent or Presence. This division established by Norvegian botanist A. Blytt and Sweden scientist R. Sernander was as early as at the end of 19th century amended by transitional phase between the Late Glacial and Boreal that was called the Preboreal. The pollen analysis developed by L. v. Post in 1916 served as a principal tool of paleobotanical research not only in Northern Europe but virtually everywhere where peat and marsh pollen bearing sediments could be found. The Mid-European region pollen pattern was described in monographs by F. Firbas (1949, 1952), E. Krippel (1986) and in number of later studies (e.g. Pokorny and Jankovska 2000). Besides a number of local stratigraphies N. Roberts (1998) suggested that the Holocene represents a series of climatic shifts which should not be handled as individual climatic periods (Wright et al. 1993). Newly P.J. Crutzen and E.F. Stoermer (2000) proposed that "recent epoch" starting at the onset of Industrial revolution of latter part of the 18th century could be called "Anthropocene" to emphasize the central role of human activities in forming new global environment.

Peat and marsh sediments are in the Czech Republic, Slovakia and in neighbouring countries restricted to some, often small scale areas such as humid montane and submontane regions, lowland oxbow lakes or around some major resurgences of Artesian springs. There are the whole landscapes where due to the entirely different environmental conditions pollen analysis cannot be applied or where it yields only stratigraphically insignificant fragments of evidence (see maps in Firbas 1949 or Succow and Jeshke 1986). However even in these regions we may found Holocene sedimentary sequences reflecting temporal and paleoenvironmental changes that could be studied by different methods. The limnic and palustrine environments tend to be conservative ones and usually less sensitive to such aspects as humidity oscillations. The pollen analysis reflects pollen influx from larger areas. On the other hand rock screes, sandy sediments under sandstone rockshelters, tufa bodies, cave fills and other Holocene sediments may provide a complementary information about actual function of different landscapes or even habitats. We propose in this article a "holistic" approach towards understanding of Holocene environments, where pollen analysis represents just one aspect in the whole array of possible approaches. We focus on problems as follows: (1) to identify the problems that can be solved by study of different Holocene facies, (2) to summarize the principal findings and (3) to compare these results with the paleobotanical studies.

2. CONCEPT OF HOLOCENE STRATIGRAPHICAL PROVINCES

K.D. Jager (2000, pers. com.) proposed that the Holocene sequences can be in some areas characterized by almost uniform sedimentational pattern while other areas are more or less different ones. The Scandinavian stratigraphy had become the base for international recognition and division of the Holocene, however the further to south we are going the bigger contrasts we may observe – e.g. in Palestine The Late Glacial represents the strikingly humid period while the climatic optimum of the Late Atlantic of the European division is there one of driest climatic phases (Niemi et al. 1997). The denomination "climatic optimum" has thus in Near East paradoxical meaning of an environmental crisis! But even in European temperate zone we found sharp distinctions – in Western Europe where relatively humid and warm oceanic climate prevailed throughout the whole Holocene we observe a number of rather short-lived climatic "ups and "downs" probably dependent on the function of the North Atlantic Oscillation (Roberts 1998), while in Central Europe

Fig. 1. Labske piskovce (Elbe Sandsteine, Northern Bohemia) – one of the large sandstone rockshelters developed in Upper Creataceous quartzose sandstones. The upper part of Holocene infillings is usually acidic and does not contain any paleobotanical evidence with the exception of scattered charcoals, while lower part of the profiles may display calcareous content and abundant finds of molluscs and Vertebraete remnants. These finds have recently substantially enlarged the areas where Holocene paleoenvironmental reconstructions are possible. (Photo: H. Rysova)

Łabskie piaskowce w północnych Czechach – jedna z większych nisz skalnych utworzona w górno kredowych piaskowcach kwarcytowych. Górna część wypełnienia holoceńskiego ma na ogół odczyn kwaśny i nie zawiera żadnych dowodów paleobotanicznych z wyjątkiem nielicznych węgielków, podczas gdy dolna część profilów wykazuje zawartość węglanów i liczne ślimaki oraz szczątki kręgowców. Te materiały pozwoliły znacznie rozszerzyć zasięg obszarów badanych pod kątem holoceńskich rekonstrukcji paleośrodowiskowych (Fot.: H. Rysova)

Fig. 2. The "Sweden Hole" in Labske piskovce (Elbe Sandsteine). Polycultural site inhabited in Mesolithic, Neolithic, Eneolithic, Bronze age and 19-20th century period. (Photo: H. Rysová)

Stanowisko "Sweden Hole" w łabskich piaskowcach, Wielokulturowe stanowisko zamieszkałe w okresach mezolitu, neolitu, eneolitu, bronzu i okresu 19-20 wieku. (Fot.: H. Rysová)

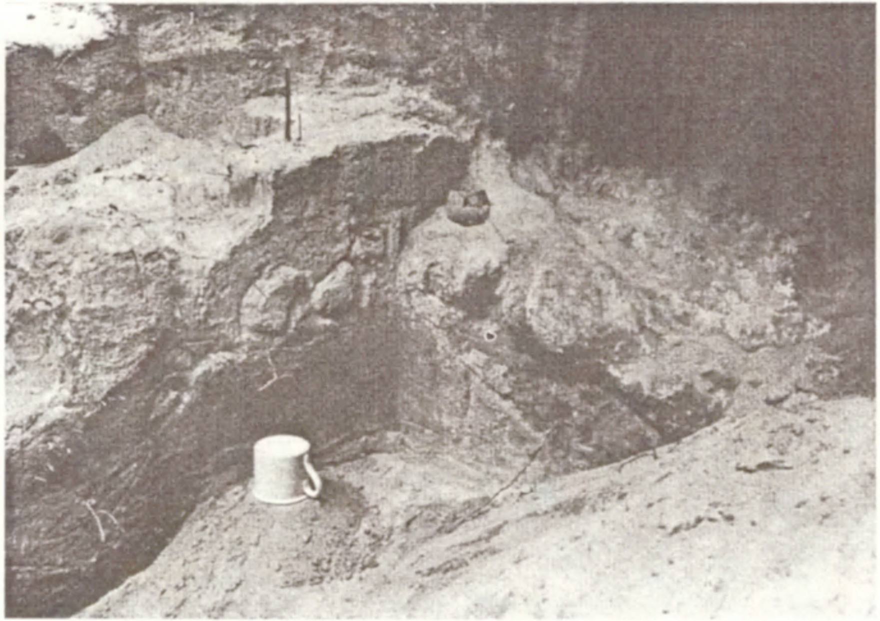

Fig. 3. The complex Holocene profile in sand infilling of small but significant rockshelter Stará skála (Old Rock) in Česká Lípa district, where ritual Neolithic bowl was discovered at the place where modern, white cup is now located, but its original place probably was at place rocky bottom of rockshelter some 40 cm above (where original Neolithic bowl lies). Archaeological finds are dating rock collapses, accumulation rates and buried horizons of various origins (Photo: H. Rysová)

Kompleksowy profil holoceński w piaskach wypełniających małą lecz ważną niszę skalną Stara skala (Old Rock) w Česka Lipa, gdzie został znaleziony rytualny neolityczny puchar, w miejscu tym obecnie jest zlokalizowana biała miska, pierwotnie leżąca na dnie niszy, około 40 cm wyżej. Znaleziska archeologiczne datują, różnego pochodzenia, odpadanie skał, wielkość akumulacji i kopalne horyzonty (Fot.: H. Rysova)

Fig. 4. Neovulcanic boulder field above Sebuzin in Elbe Valley (Northern Bohemia). Scree and boulder fields were for decades considered to be periglacial phenomena, however they are at some places underlain by pure loess and field evidence points to dominant Holocene age of the main mass of scree fields (Photo: H. Rysova)

Neowulkaniczne pole blokowe nad Sebuzinem w dolinie Łaby (północne Czechy). Piargi i pola blokowe przez dziesiątki lat były uważane za zjawiska peryglacjalne, jednak w kilku miejscach spoczywają na czystych lessach, a dowody terenowe wskazują na holoceński wiek piargów (Fot.: H. Rysova) where continentality-oceanity shifts associated with more frequent Siberian Height episodes we find several main specific Holocene layers corresponding to the Mid-European stratigraphical province defined by K.D. Jager (1969) on the basis of tufa stratigraphy.

3. HOLOCENE ENVIRONMENTS

Two central types of the contrasting natural environments are playing the principal role in Holocene studies. Both of them need different approaches and methods, and provide different answers. Besides these sources other methods and lines of evidence are emerging – e.g. speleothem and tree rings studies (Roberts 1998).

1. **Aquatic sequences:** Classical Holocene system is established mostly on the basis of paleobotanical analysis in limnic and palustrine aquatic environments such as lakes, marshes and peatbogs. Such environments are common in glaciated landscapes, but rare in a periglacial zone especially in lowlands. We can find on the area of Czech and Slovak Republics only limited number of such Holocene profiles – especially in frontier mountains, Treboň Basin, marshes in Česká Lipa district in Czech Republic and mostly individual and scattered sites such as Orava Basin in Slovakia. The greater part of Mid-European area including almost all Thermofyticum remains basically uncovered. Pollen analysis reveals general features of vegetational development in the broader area but usually does not allow the study of geological processes – e.g. intensity of pedogenesis, slope processes and erosional events. The archaeological finds are less common. The aquatic environments are dependent, but at the same time distinct from adjacent "average" landscape matrix. Peatlands and wetlands are linked to areas that are very different from warm-dry and fertile zones characterized by intensive prehistoric land-use. The correlation with Vertebrate and Molluscan fauna is due to the natural acidity almost impossible, but on the other hand the European network of pollen profiles provides basis for large scale, trans-continental correlations.

2. **Subaeric terrestrial sequences:** are omnipresent at different altitudes of diversified range of landscape types. They are often continuous and can be dated on the basis of radiocarbon, molluscan assemblages and archaeological finds. They must contain some quantities of carbonate to preserve organic remnants such as bones but on the other hand the presence of carbonate usually excludes due to oxidation the preservation of pollen spectra. The most relevant types of terrestrial sequences are as follows:

– Slope sequences, mostly deposited under rock cliffs display changes in rock scree size, humus and carbonate content. The enhanced dynamics of scree formation is conveyed by coarse sedimentation, often without matrix. On the other hand the standstill phase can be determined on the basis of soil horizons evolution. Some special cases like Early-Holocene cemented screes (9–12 ka BP) are product of more humid episodes (Žák et al. 2001), while coarse grained screes of Subboreal (700–1250 BC) containing dark, humic material correspond to a major continentality shift (Ložek 1988).

– Fine grained wash-down sediments deposited in dellens display the alike depositional pattern in dependence on anthropogenic deforestration, tillage or pasture. They may represent the products of flash-floods and other hydrological events (Ložek 1976).

– Karst and pseudokarst infillings of cave and rockshelter entrances represent continuation of slope sedimentation but other aspects such as flowstone (humid phase) and breakdown horizons (usually continental or oscillating climate) are present. The soil sediments washed down into dark cave environment are not influenced by pedogenesis and thus conserved for further research. The human settlement is common and the correlation between natural conditions and human cultures is being possible in such aspects as hunting activities, Mesolithic hazelnut gathering, deforestration and many other ones (Kukla, Ložek 1958; Svoboda et al. 1998).

– Floodplain deposits represent key transitional members between aquatic and terrestrial sequences. They reflect erosional and depositional processes of individual catchments. Gravel and sand correspond to the high flood activities, while Fluvisol formation is dependent on stand-still phases. The buried archaeological finds provide evidence for morphological development of floodplains in different time slices. The sediments of oxbow lakes and abandoned meanders are in some areas of Bohemian Createceous Basin calcareous and thus they may enable correlation between faunal and floral evidence.

– Tufa deposits have the exceptional and often dominant position among other Holocene sequences because their architecture reflects several important structures: intensity of carbonate metabolism, hydrological changes, rock scree formation episodes and most significantly prolonged drought episodes. There exists an almost uniform pattern of tufa sedimentation in Mid-European context in a range starting in Thuringia (Germany), covering the whole area of Czech Republic and reaching as far as to Southern Moravia and South-Eastern Slovakia. Tufa bodies often contain intercalations of several (2–4) humic fossil soils of Epiatlantic to Subboreal age corresponding to dry periods. The thickness of deposits is often between 5–10m (some 17m in Svatý Jan pod Skalou in Bohemian Karst), it usually covers the interval 9500 to about 2200 years BP. The deposits can be studied by a number of complementary methods including stable isotope studies and correlations between archaeological, biostratigraphical and geochemical methods (Jäger 1982, Žák et al. 2002). The tufa outcrops located around resurgences in lower parts of karst valleys indicate slope retreat, downcutting and standstill phases in slope dynamics.

3. BIOSTRATIGRAPHICAL METHODS OF TERRESTRIAL SEQUENCES

The Holocene terrestrial biostratigraphy is founded on two groups of animals – Molluscs and Vertebrate, even when some other groups like Insecta or Ostracoda may provide additional evidence (Roberts 1998). Malacological analysis is, due to the carbonate nature of mollusk shells, limited to calcareous districts or sites where the shells are not dissolved by acidic solutions. These districts are covering surprisingly large areas of Central Europe – all karst areas, sites located on small outcrops of metamorphic carbonates and some volcanoclastic strata, some Artesian springs and thermal water deposits and travertines, calcareous layers – mostly sandstones and marls – of Mesozoic and Kenozoic (Paleogene) age, some floodplain sediments and Holocene soils derived from loessic substrate. The number of Mollusc species can reach in one profile (or even individual layers) some 20–40 species and the total number of all shells can be measured in an order of thousands to tens thousands individuals and expressed as histograms that resemble pollen graphs. Molluscs are environmentally sensitive and they tend to form assemblages closely connected to basic phytocenological units. Therefore the fossil assemblage of an individual time slice may be directly attributed to given habitats with the exception of acidophilous biotops such as heaths or pine woods.

Molluscan evidence is at least twofold: besides a strong relation to a type of vegetation we find another important relation to the soil cover (Ložek 1988). Their thanatocoenoses are composed of species that originally lived at the site in question and the transported species. While out-washed floodplain sediments may represent an "average" sample of few kilometers of the river course, the horizontal transport of shells in rocky habitats is restricted to few meters and the vertical transport corresponds to the fall-out from above lying terrain. This situation together with well-known ecological demands of individual, often highly specialized mollusc species enables to reconstruct not only the general type of vegetation but even the presence of small steppe patches or miniature marshes in otherwise different environment e.g. in forests.

Vertebrate, especially mammals need the alike calcareous conditions of fossilization, but because of the bone size they are less sensitive to post-depositional changes such as dissolution. The big bones of large animals like deer that cannot be directly linked to one specific environment represent the most common type of finds. The number of Vertebrate species is almost always lower than number of Mollusc species and the number of individuals can be lesser by two or three orders, because we may find per one square meter as many as several hundred molluscs. Therefore the places of secondary concentration of Vertebrate remnants like predator burrows or owl´s castings represent the most welcomed evidence. One must be, however, aware that these places – often located in diversified karst terrain where small cavities are abundant – cover the whole predator territory. We often find e.g. under nests of large owls a thanotocoenosis that is composed of the local mollusc assemblage mixed together with Vertebrate remnants of animals collected in various environments, frequently in the vicinity of large river and its floodplain. Such places may yield a surprisingly valuable complementary evidence of the local conditions and larger landscape background.

Archaeological finds are abundant especially in karst and sandstone caves and rockshelters and around tufa bodies. They provide the cheapest dating technique and they enable studies of human induced impact in the vicinity of site where e.g. phases of deforestation or sheet erosion associated with Neolithic farming can be deciphered. The lines of sedimentological, pedological, geochemical and biostratigraphical evidence may thus intersect in complex intepretation of given Holocene strata – e.g. on the area of Bohemian Karst (130 km^2) some 50 Holocene sites were excavated to obtain plastic picture of Holocene environments from lowest points in floodplain of Berounka River, over valley sides and plateau relief to the tops of highest hills. The Holocene research in the Slovak Carpathians provided large scale landscape interpretation of environmental and climatic changes from floodplains, to adjoining forests of hilly regions, xerothermic patches of isolated rocks at different altitudes to the montane marshes, plateaus and high peaks (Ložek 1992).

The advantage of several rather independent lines of evidence lies not only in the complexity of paleoenvironmental interpretation, but in their mutual verification – e.g. chalky loose sinter (foam sinter) of flowstone in limestone rockshelters prove high humidity, that is verified by the presence of molluscs preferring very humid places. This layer never contains archaeological finds

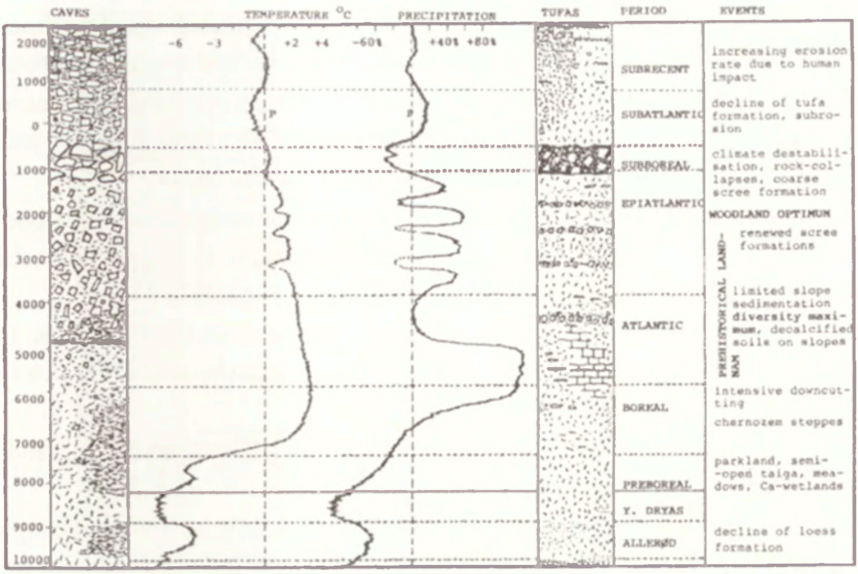

Fig. 5. The division of Holocene as proposed by K.D. Jäger (1969). The age is given according to non-calibrated radiocarbon data. Temperature and precipitation is expressed as the function of sedimentological character and paleoenvironmental molluscan analysis compared with contemporary conditions (Ložek, Cilek 1995)

Podział holocenu według K.D. Jägera (1969). Wiek określono według nickalibrowanych dat radiowęglowych. Temperatura i opady są wyznaczone jako funkcja cech osadów i analiz ślimaków w stosunku do warunków współczesnych (Ložek, Cilek 1995)

preferring very humid places. This layer never contains archaeological finds but next overlying horizon often harbours Early Neolithic pottery. We may interpret this situation recorded at several places as the Neolithic advance connected to dry oscillation that succeeded immediately after important humid phase. The paleobotanical evidence such as charcoals, seeds, fossil leaves in carbonate horizons or pollen bearing humic strata is usually fragmented and restricted to sporadic stratigraphical horizons but it provides a welcomed supplementary environmental information (Draxler 1972; Svobodová, Svoboda 1988).

4. PRINCIPAL ELEMENTS OF PALEOENVIRONMENTAL ANALYSIS

The colourful, "rainbow" spectrum of the gathered observations of Holocene depositional sequences embodies some basic aspects of paleoenvironmental and paleoclimatic information that may be listed as follows:

– Standstill phases in slope and floodplain sediments are indicated by mature soil horizons, by tufa and flowstone deposition without clastics in karst valleys and caves and generally by the absence of coarse-grained sediments. These phases are characterized by low accumulation and negligible erosion.

– Desintegration of rock surfaces, retreat of valley slopes and cave entrances is indicated by coarse screes and boulder fields. The interstices are usually free or filled with dark humic matrix (Cilek 2000a, b).

– Extremely humid phases are indicated by carbonate horizons or carbonate cemented screes. Carbonate deposition is associated with intensive leaching mostly of brown forest soils even at steep slopes (Ložek 1984).

– Prolonged dry periods are indicated by buried soils in tufa bodies. Due to the natural mobility of springs several outcrops must be studied to filter out the site-specific signal (Žak et al. 2001).

– Important erosional phases are indicated by gully downcutting and by formation of wash-down fans.

– Vegetational changes are indirectly indicated mostly by malacozoological analysis. Mollusc are sensitive to environment along these three major axes: 1. Dry-humid climate, 2. Woodland – open grounds, 3. Natural – anthropogenic migration.

– Patches or small enclave recognition, e.g. small steppe patches within forested regions are indicated by malacological analysis, because even the area of several square meters is capable to sustain a relict assemblage of molluscs.

All these processes can be studied as the sequence of Holocene environments or succession of landscapes that can be correlated with onsets and declines of different prehistoric civilizations and with postglacial development of vegetation (Ložek 1982).

5. EVIDENCE: SIGNIFICANT EXAMPLES

The series of observations gathered mostly by V. Ložek in aprox. 200 various Holocene sites in few paragraphs. However we may try to point to several significant examples listed as follows:

– The maximum migration/deposition of calcium carbonate takes place before the Neolithic colonization. It represents the most humid phase of the whole Holocene and it ends abruptly. Carbonate layers and flowstones can be found even in contemporary dry and warm conditions of the Thermofyticum (e.g. in chernozem area of Southern Moravia). The fast and intensive formation of soils happens in Pre-Neolithic carbonate phase due to the enhanced intensity of chemical weathering.

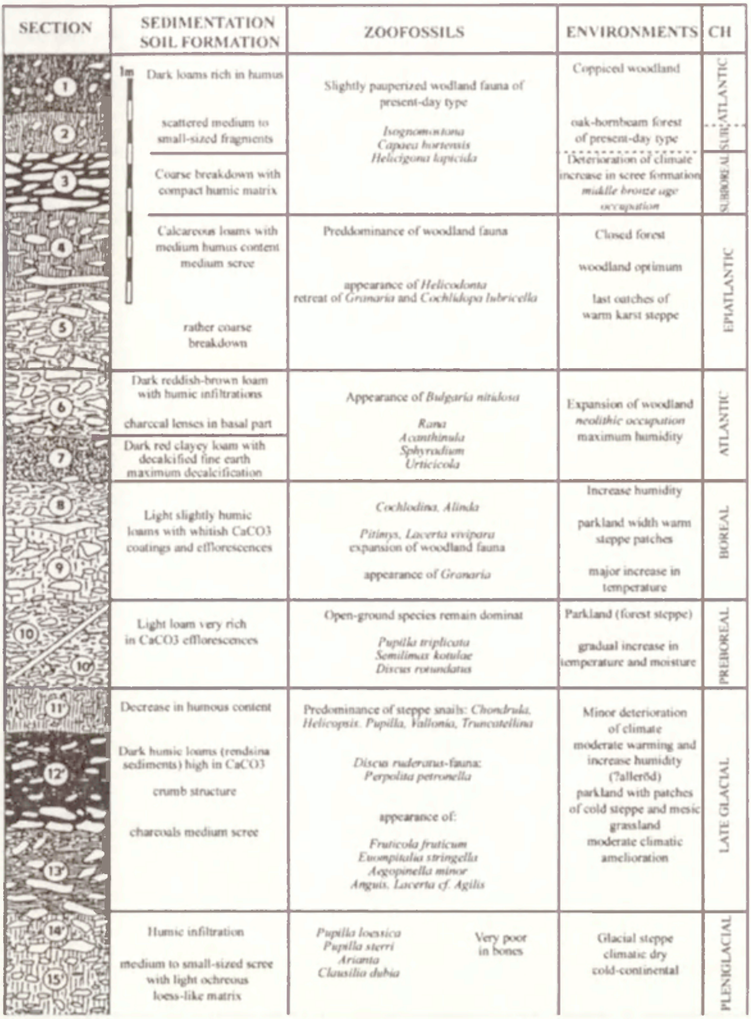

Fig. 6. Depositional sequence in the cave entrance of Martina Cave near Tetin in Bohemian Karst, Central Bohemia (altitude 360 m, mean annual temperature 8.2°C, rainfall 520 mm). Situation: upper part of the east-facing slope of the Koda Plateau, small rocks, oak-hornbeam woodland. Important horizons: 12 – humic rendzina sediments with Late Glacial malacofauna, 7 – dark red loam with high content of decalciefied clay immediately underlying the Neolithic horizon – it corresponds to the Postglacial humidity maximum, 3 – stony horizon with black humic matrix (rendzina sediment) and Middle Bronze Age pottery in its basal part. Drawn by V. Lozck.

Sekwencja osadów przy otworze jaskini Martina koło Tetina w Krasie Czeskim, (wysokość 360 m npm, średnia roczna temperatura 8,2°C, opady 520 mm). Górna część stoku płaskowyżu Koda eksponowanego ku wschodowi, male formy skalne, las dębowo-grabowy. Ważne poziomy: 12 – osad rendzinowy z późnoglacjalną malakofauną, 7 – ciemno czerwony ił z wysoką zawartością odwapnionych glin ułożonych bezposrednio pod poziomem neolitycznym – odpowiada postglacjalnemu maksimum wilgotnościowemu, 3 – poziom kamienisty z czarnym humusowym wypełnieniem (rendzina) oraz wyroby garncarskie z epoki środkowego bronzu u ich podstawy.

SECTION	SEDIMENTATION SOIL FORMATION	ZOOFOSSILS	ENVIRONMENTS	CH
1	Loose soil structure, decreasing humus content	*Cecilicides acicula*, maximum of *Oxychilus inopinatus*, *Truncatellina cylindrica*, *Cochlicopa lubricella*, *Balea perversa* in high amounts Citellus	Steppe grassland middle ages grazing soil erosion	LATE SUBATLANTIC – SUBRECENT
2	Dark rendsina sediments rich in medium to fine rock debris	Modern fauna dominated by open-ground species; appearance of *Xerolenta obvia*	Lat'ene steppe grassland shrubland	SUBATLANTIC
3	Coarse bouldery breakdown with black rendsina matrix B-giant block separated from the rock wall	Dramatic retreat of woodland fauna *Eliomys*; gradual decline of *Carychium tridentatum*, *Vitrea contratcta* and *Clausilia pumila*; re-expansion of open-ground species	Moisture minimum dry interval bronze/hallstatt hillforts of Pálava summits; retreat of woodland	SUBBOREAL
4	Bouldery breakdown with brownish grey humic matrix	Decreasing of number woodland species; minimum of *Helicopsis striata*, *Pupilla sterri*, *Chondrula tridens* and *Vallonia costata*; *Carychium tridentattum maximum*	Predominantly mesic woodland, late climatic optimum; gradual increase in moisture	EPIATLANTIC
5	Maximum CaCO3 - precipitation foam sinter	Woodland fauna culmination *Bulgaria cana*, *Ruthenica*, *Sphyradium doliolum*, *Macrogastra ventricosa*, *Crociduramicrotus cf gregalis*, *cricetus*; retreat of steppe species *Chondrina maximum*	Damp closed forest, moisture maximum, early climatic optimum	ATLANTIC
6	Increasing CaCO3 - precipitation angular debris	Expansion of *Arianta*, *Discus ruderatus*, *Vitrea crystallina*; invasion of woodland fauna *V. costata*, still abundant *Microtus cf gregalis*, *Cricetus*; appearance of *Carychium tridentatum*	Expansion of woodland, parkland, steppe; rapid increase in	PREBOREAL BOREAL
7	Fine angular debris loess-like matrix scattered coarser fragments	Tolerant open-ground malacofauna dominated by *Vallonia costata microtus cf oeconomus*	Temperature and moisture epipaleolithic xeric grassland moderate increase inmoisture	LATE GLACIAL
8	Typical loess	Typical loess malacofauna refugial occurrence of *Citochondria clienta*	Loess steppe severe dry-continental climate	PLENIGLACIAL

Fig. 7. Depositional sequence of slope sediments including abundant screes and fallen blocks in Soutěska Pass, Pálava Hills in Southern Moravia (altitude 375 m, mean annual temperature 9.0°C, rainfall 550 mm). Situation: foot of a high, slightly overhanging west-facing rock wall, xerothermic karst grassland and shrubland. Important horizons: 8 – typical yellow loess of the Vistulian (Weichsel) Pleniglacial, 7 – limestone rubble with Epipalaeolithic artifacts, 5 – foam sinter indicating the moisture maximum, 3 – very coarse breakdown blocks with rendzina-like matrix and Late Bronze Age pottery. Drawn by V. Ložek.

Sekwencja osadów stokowych zawierających liczne piargi i odpadnięte bloki skalne na przełęczy Soutěska, Wzgórza Pálava w południowych Morawach (wysokość 375 m npm, średnia roczna temperatura 9,0°C, opady 550 mm). Podnóże wysokiej, lekko przewieszonej ściany skalnej eksponowanej ku zachodowi, kserotermiczne murawy krasowe i zarośla. Ważne poziomy: 8 – typowy żółty less z pleniglacjału vistulianskiego, 7 – wapienny rumosz z epipaleolitycznymi artefaktami, 5 – martwica wapienna wskazująca maksimum wilgotności, 3 – bloki bardzo grubego rozpadu z wypełnieniem przypominającym rendzinę oraz wyroby garncarskie z epoki późnego brązu

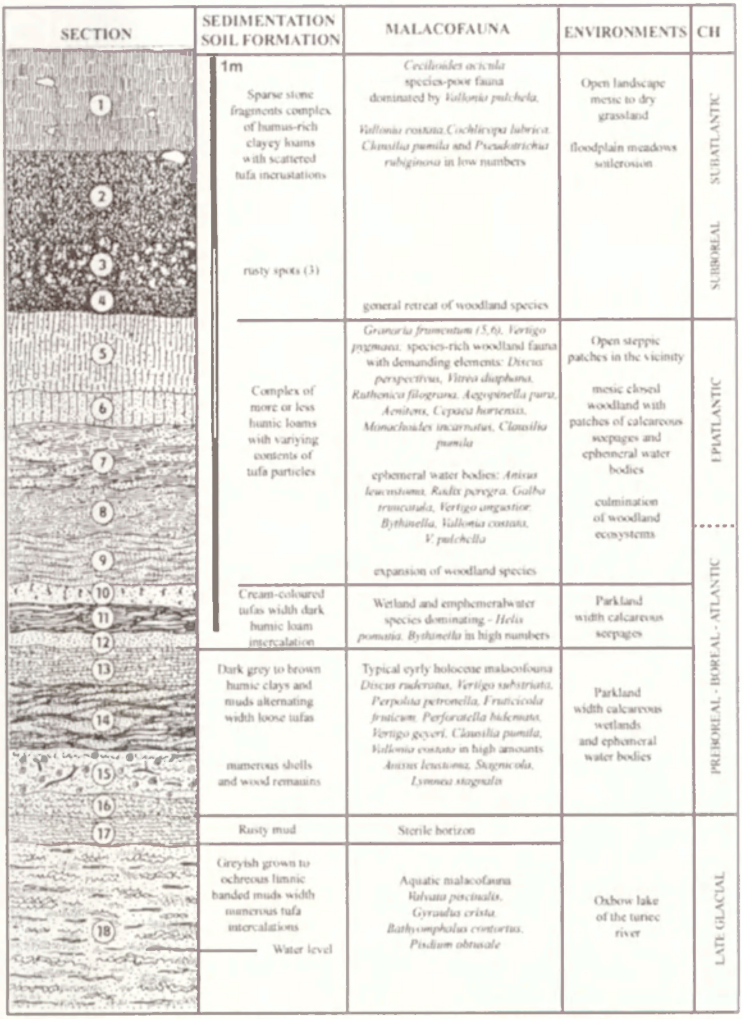

Fig. 8. Depositional sequence of alluvial sediments with colluvial admixture of cut-bank of Turiec River at Laskar, Turčianska kotlina Basin, Central Slovakia (altitude 440 m, temperature 7.3°C, rainfall 765 mm). Situation: marginal zone of the Turiec River floodplain at the foot of a gentle hillside slope, open country, pastures and fields. Important horizons: 18 – oxbow-lake sediments dominatede by aquatic malacofauna, 16 to 5 – complex of more or less tufaceous strata of Early to Middle Holocene age, 4 to 1 – increased colluvial sedimentation in deforested farming landscape of Late Holocene. Drawn by V. Ložek.

Sekwencja osadów aluwialnych z domieszkami deluwiów w podcięciu erozyjnym rzeki Turiec w miejscowości Laskar, Kotlina Turczańska, środkowa Słowacja (wysokość 440 m npm, średnia roczna temperatura 7,3°C, opady 765 mm). Strefa marginalna równiny zalewowej rzeki Turiec u podnóża łagodnych zboczy, teren otwarty, pastwiska i pola uprawne. Ważne poziomy: 18 – osady starorzecza z dominacją malakofauny wodnej, 16 do 5 – zespół warstw tufowych wieku wczesno i środkowo holoceńskiego, 4 do 1 – osady koluwialne krajobrazu rolniczego, wylesionego, późnoholoceńskiego

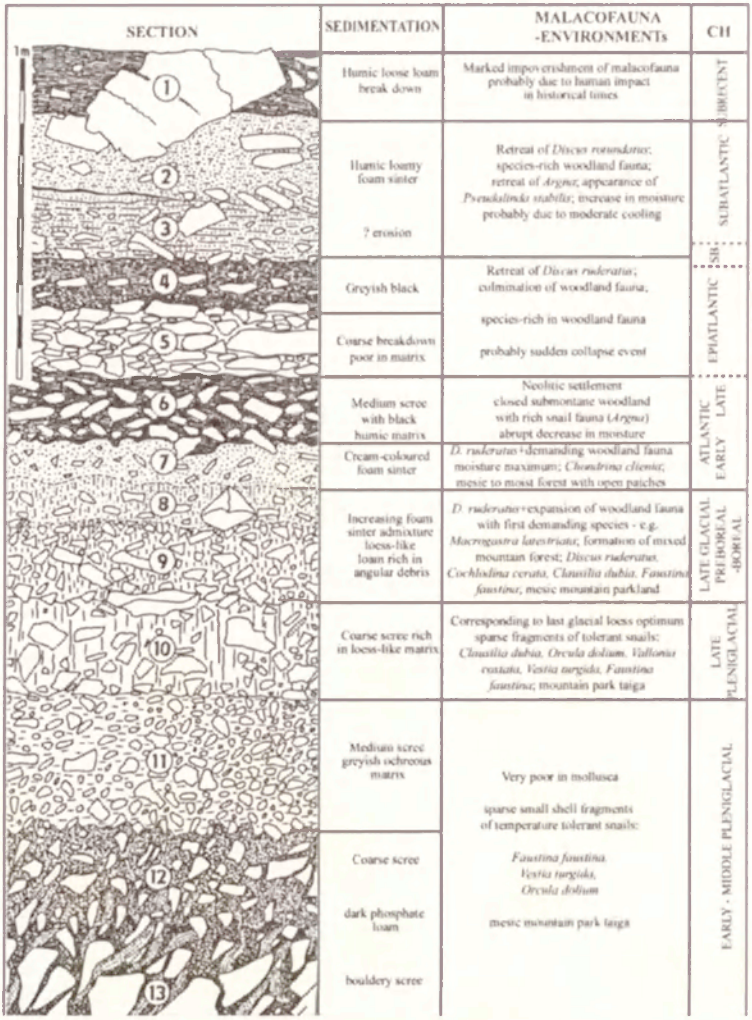

Fig. 9. Depositional sequence in the cave entrance of Velka Ruzinska Cave, Ružinske Mts., Eastern Slovakia (altitude 614 m, mean annual temperature 6.0°C, rainfall 700 mm). Situation: large cave situated in the northeast-facing slope of the Maly Ružinek Gorge in the mountainous karstland, extensive submontane beech and mixed scree forests. Important horizons: 13 to 12 – dark phosphate loams of Early Vistulian Pleniglacial, 7 – light foam sinter horizon with marked upper boundary indicating the moisture maximum, 6 – scree with black matrix and abundant Neolithic pottery (by V. Ložek).

Sekwencja osadów przy wejściu do jaskini Velka Ružínska, Ružínské Góry, Wschodnia Słowacja (wysokość 614 m npm, średnia roczna temperatura 6,0°C, opady 700 mm). Usytuowanie: ujście jaskini Mały Ružinek w górzystym krasie, rozległe lasy bukowe i mieszane na rumowiskach skalnych (piargach). Ważne poziomy: 13 do 12 – ciemno fosfatowe gliny z wczesnego pleniglacjału vistulianskiego, 7 – lekko piankowa skała, poziom z zaznaczoną górną granicą świadczącą o maksymalnym zawilgoceniu, 6 – rumowisko z czarnymi skałami macierzystymi i dużą ilością wyrobów garncarskich (V. Ložek).

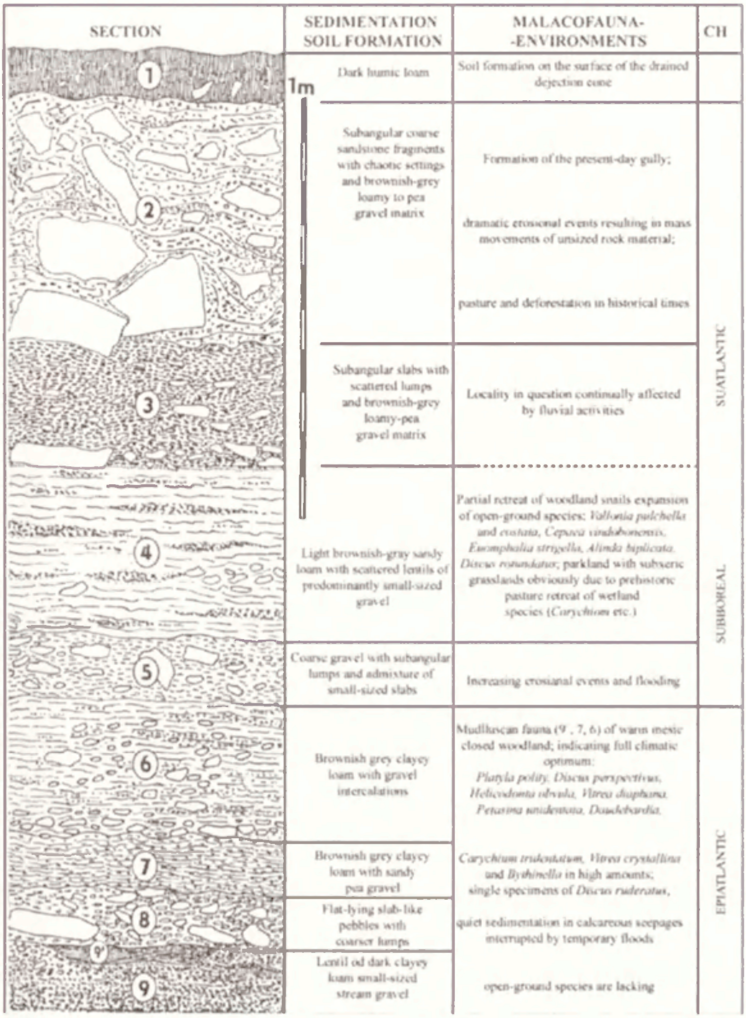

Fig. 10. Depositional sequence of a dejection cone in National Nature Reserve near Velka nad Velickou in White Carpathians, Southeastern Moravia (altitude 350 m, mean annual temperature 7.8°C, rainfall 700 mm). Situation: mouth of the ravine Velky Jazevci jarok into Velička Valley, remnants of oak-hornbeam stands within extensive meadows and pastures on Flysh substrate. Important horizons: 9 to 6 – comparatively quiet deposition under forest cover, 5 to 3 – gradual deforestration accompanied by soil erosion, 2 – intensive erosion of bedrock, torrential floods transporting very coarse rock fragments, 1 – downcutting event and the formation of surface soil horizon (by V. Ložek).

Sekwencja osadów na stożku w rezerwacie kolo miejscowości Velka nad Veličkou w Białych Karpatach, południowe Morawy (wysokość 350 m npm, średnia roczna temperatura 7,8°C, opady 700 mm). Ujście jaru Velky Jazevči do doliny Velička, pozostałości dębowo-grabowego lasu z rozległymi łąkami i pastwiskami na podłożu fliszowym. Ważne poziomy: 9 do 6 – stosunkowo spokojna depozycja pod pokrywą leśną, 5 do 3 – stopniowe wylesienie i towarzysząca mu erozja gleb, 2 – intensywna erozja podłoża skalnego, powodzie transportujące grubo okruchowe fragmenty skał, 1 – rozcinanie i formowanie poziomu glebowego (V. Ložek).

– Tufa deposits and some other sites reflect massive humid phase, that is later succeeded by alternating sequence of carbonate and scree horizons set in by probable oceanity-continentality climatic shifts (Lozek 1984,1997).

– The maximal migration/deposition of calcium carbonate is accompanied by rapidly growing number of woodland molluscs species and thus by evolution of closed forest. The upper tree limit is the highest during late climatic optimum (Lozek 1978).

– The Subboreal (sensu Jäger 1969, 700–1250 BP) represents the most critical period of the Holocene in relationship to unprecedented intensity of rock falls and collapses and coarse scree formation. The Subboreal sediments are often accompanied by abundant Late Bronze Age ceramic fragments. Harsh continental climate, deforestration, droughts and expanded landnam (expansion of settlement area, grazing) are to be expected (Jäger, Lozek 1983).

– Short but vigorous erosional event takes place at the beginning of peak of Pre-Neolithic carbonate phase (Jäger, Lozek 1983).

– Erosion, subrosion and the end of tufa formation takes place at the majority of deposits during 2–5th century B.C. (Zák et al. 2002; Kadlec et al. this volume).

– The grassland and steppe relicts including chernozem soils were preserved in areas of continuous human impact settled during Neolithic colonization.

– Anthropic influences lead to the spreading of modern molluscan immigrants that never lived in Central Europe before.

– The first half of the Holocene can be characterized by presence of soils containing some quantitites of calcium carbonate, while later we observe leaching of carbonate, acidification and oligotrophisation of soils that may lead to profound vegetational changes and environmental collapses in large areas. This phenomenon is especially characteristic for Bohemian Cretaceous sandstone areas (Lozek 1998a, b).

6. LANDSCAPE EVOLUTION OF MID-EUROPEAN UPLANDS

On the basis of gathered evidence we can compile the Holocene evolution in hilly and submontane areas including warm and dry regions such as karstlands, where paleobotanical methods do not provide the necessary framework for evaluation of actual environmental changes. The initial point of Holocene sequences is rooted in the **Late Pleniglacial**. Loess sedimentation prevails at altitudes 150–300 (350)m above sea level. Northern and arcto-alpine molluscan fauna is characterized by presence of *Vertigo parcedentata, Columella columella*; Central-Asian and North-Asian species are *Vallonia tenuilabris, Pupilla loessica, Vertigo pseudosubstriata,* but steppe species such as *Pupilla sterri* and *Helicopsis striata*, xerothermic species such as *Pupilla triplicata,* indifferent species such as *Succinella oblonga* or *Trichia hispida* are present as well. This assemblage and loess as the substrate do not have any analogy in contemporary Europe. They reflect a very specific conditions of cold continental loess grassland, where the demanding forest species of latter Holocene could not survive (Lozek 1991a, b).

However in more humid hilly regions or at the periphery of loess area some species could have survived the Ice age as witnessed by some finds in Slovakian Karst (*Faustina faustina, Laciniaria plicata, Cochlodina cerata* and others), and from the southern foot of the Low Tatra Mts. near Farkasovo in the Hron valley.

The Late Glacial is characterized by the decline of loess sedimentation, development of chernozem soils in dry loessic landscapes and by expansion of taiga species such as *Discus ruderatus,* and parkland species such as *Fruticicola fruticum.* The Late Glacial is gradually transformed into parklands patchwork where mesic meadows, numerous marshes and woods are interwoven in delicate, high diversity pattern. Soils remain almost undeveloped, but we may observe at some places initial chernozems (Zernoseky in Northern Bohemia – Lozek, Šibrava, 1982) or rendzinas (Cave Martina in Bohemian Karst – Horacek et al. 2002).

The Preboreal displays at the beginning picture like the Late Glacial, but a rapid expansion of thermophilous forest species of fauna and flora takes place. Of particular interest is the peculiar combination of forest species (*Discus ruderatus, Vertigo substriata, Perpolita petronella, Clausilia cruciata*) and steppe species (*Chondrula tridens, Pupilla triplicata, Pupilla sterri, Helicopsis striata*), some of the latter species are surviving together with small mammals such as: *Microtus gregalis, Ochotona sp., Sicista sp.* from the Pleistocene.

The Boreal represents almost smooth continuation of the Preboreal characterized by prevailing dry, continental climate, but a new phase of progressive warming leads to the fast development of chernozem grasslands, where *Chondrula tridens, Cochlicopa lubricella, Truncatellina cylindrical* are spreading. Warm parklands develop in some hilly regions, especially in karst areas. Southern faunal element *Granaria frumentum* appears there. Hazelnut and oak woods and groves are occupied by woodland species like *Aegopinella* *minor, Helix pomatia, Ena obscura* and others. *Fruticicola fruticum* and *Euomphalia strigela* participate in ecotone formations.

Fauna with index species *Discus ruderatus* is still common, but more demanding mostly forest species disseminate fastly during the concluding humid phase documented by flowstone precipitation in caves and limestone rockshelters. The notable feature is short but intensive downcutting event (Jäger, Ložek 1983). The mature carbonate chernozems were developed in dry, warm areas even outside their contemporary occurrences (Bohemian Karst, Srbsko-Vodopády – Ložek 1998a).

The original division of the Holocene (Iversen 1973; Firbas 1949 – zone VI)) assigned warm and humid phase termed as "Atlantic". However the Mid-European tufa bodies and some other terrestrial sequences display in Atlantic position two distinct sedimentary phases. The earlier one was proposed by K.D. Jäger (1969) to be called the **Atlantic**, the latter part was denominated as the Epiatlantic. The main differences between these two phases is stable, first very humid, then humid climatic regime of the Atlantic and oscillating, uneven climatic regime of the Epiatlantic, where the Atlantic type of climate alternates with prolonged drought periods. The Epiatlantic sensu K.D. Jäger (1969) was extended over part of the former Subboreal that became restricted to pronounced dry and continental period of the Late Bronze Age cultures.

The Pre-Neolithic Atlantic displays the maximal humidity of the whole postglacial period as documented by foam sinters in otherwise clastic strata of otherwise warm and dry karstlands (Ložek 1984). The latter part of the Atlantic is drier, intense soil formation started and carbonate sedimentation was ended. We can still find several humidity demanding species such as *Carychium tridentatum*. Very humid early phase is documented by the expansion and later withdrawal of many hygrophiles even into otherwise xerothermic landscapes such as Pálava in Southern Moravia (Ložek 1985).

The Epiatlantic represented a "newly" proposed term by K.D. Jäger (1969) that became justified by a number of further explorations in different ecosystems of former Czechoslovakia and Germany. The humid climate is prevailing, forests had reached their maximal diversity and extent. Some hygrophilous montane species such as *Macrogastra densestriata* in Bohemia and Middle Germany, or the Carpathian element *Macrogastra latestriata* penetrate into lower regions. The characteristic *Discus ruderatus* fauna (Late Glacial-Boreal) completely disappears in hilly regions and lowlands. The forest molluscan diversity culminates but penetration of some xerotherm species from south takes place (*Truncatellina claustralis*). Substantial feature is culmination of hygrophiles *Carychium tridentatum* and almost complete disappearence of grassland species including *Vallonia costata*, that was very common in the Atlantic. While the development of molluscan fauna seems to be gradual and smooth one, the tufa sedimentation displays sudden changes. In complete tufa series several, usually four intercalations of soils, screes and wash-down soil sediments appear. Given the accumulation rates and the length of Epiatlantic these dry or unstable oscillations were not longer than few decades, maximally 200 years. They had not left any profound or in fossil record visible impact on vegetation and molluscan assemblages but they altered sedimentational pattern (Wittislingen in Middle Germany – Seitz 1951; several deposits of Bohemian Karst – Zák et al. 2002; Slovak karst – Ložek 1991b). While the slope deposits of Atlantic age are indicated by intensive pedogenesis leading to formation of brown decalcified soils, Epiatlantic sedimentation consists of small to medium sized screes with mostly humic, carbonate rich matrix. Epiatlantic and Atlantic correspond to the postglacial climatic optimum, but the highest floral and faunal diversity culminates during Epiatlantic, when both successions have reached their climax stages (Ložek 1982).

The significant factor in landscape development is represented by the beginnings of farming. Wild, ineffective accessory farming activities of indigenous Mesolithic population have met with the migrational wave of Danubian farmers at the beginning of the Neolithic era. The dry and warm, loess-chernozem areas with scattered forests, shrublands and grasslands were the primary targets for farming and they have remained continuously settled ever for the last several thousands years. The forest and soil evolution was stopped due to the transformation to cultural steppe. Therefore the Epiatlantic culmination of the woodland malacofauna diversity is missing in these areas. The farming land expansion significantly continues during the Eneolithic and Bronze Age eras to reach its peak during succeeding the **Subboreal** sensu K.D. Jäger (1969).

The Subboreal of the Mid-European Holocene stratigraphic province corresponds to the Late Bronze Age (1400) 1250–700 B.C. The most striking demonstration of this mostly dry and climatically contrasting continental weather period is the marked horizon of humic soils and screes buried in tufa bodies. This similar pattern can be found all over Middle Germany to Bohemia and as far as South-Eastern Slovakia. The age of soil horizons can be independently dated by abundant finds of local Late Bronze Age pottery. The Subboreal climate lead to the unprecendented formation of coarse screes due to the rock collapses and cave entrances retreat (Ložek 1988). Interstices in coarse screes often remain unfilled by fine grained matrix and they contain autochtonous forest molluscan fauna. These circumstances prove (1) rapid, sometimes “catastrophic” formation of screes due to the rock falls, (2) these changes were taking place still in forested environments and thus without significant anthropic impact. Some of the more demanding species like *Macrogastra densestriata, Macrogastra latestriata, Truncatellina claustralis* and at some places *Bulgarica cana, Carychium tridentatum* are retreating, but on the other hand some species of grasslands including modern elements that have never lived in Central Europe before(*Oxychilus inopinatus* – since Neolithic, *Cecilioides acicula* – undeground species). For these reasons we consider the Subboreal from the point of Holocene biostratigraphy as the initial phase of the Late Holocene diversity depletion (Lozek 1982).

The Subboreal may be regarded as a kind of a boundary in some Mid-European regions or landscape elements, because of the irreversible relief and soil changes taking place. Severe erosion leads to formation of flat floodplains because of the enhanced accumulation of floodplain loams (Auelehm). Tufa bodies were affected by downcutting, several meters deep gullies developed at many different places. The dramatic change that may be depicted as “Lusatia crisis” was taking place in sandstone areas (Lusatia culture represents one of the most widespread cultures of the Late Bronze Age). The older Pre-Subboreal sand strata are calcareous and may contain as many as 40 molluscan species including demanding forest species, while younger sediments are acidic and oligotrophic – the diversity is significantly reduced to 5–7 environmentally indifferent species that have survived in these areas till now. The leaching of calcium carbonate had a profound effect on the vegetation – the original mostly deciduous beech and oak woods were transformed into prevailing oligotrophic pine forests. Fortified settlements were for the first time located in mountainous regions close to the tree limit, sometimes 1200–1500 m above sea level (Poludnica in Nízke Tatry Mts., Slovakia – Pieta 1981).

The further biodiversity deterioration continued through the **Subatlantic** when some more demanding molluscan species disappeared. Last centuries before our era are characterized by more humid, colder climate and partial reforestration of hilly regions. Great migration took place during the 1-6th century when it was concluded by the Slavic arrival around 540 and then 590 (Třeštík 1997). The Slavs occupied at the beginning the old loess-chernozem settlement area, but during the 10th century and later they spread over most of the still virgin landscape including mountainous regions. Mining colonization played an important role in the Bohemian mountains, but pastoral Valachian colonization was even more intensive in the Carpathian mountain circle (Häufler 1955). The significant expansion of southern or indifferent species (*Pupilla muscorum, Vallonia sp.*) have taken place and at some regions still continues. The humankind has become during at least two last industrial centuries the principal geological factor and biological restraint.

7. CONCLUSIONS

Every medium, every line of fossil evidence is bringing not only an individual but also interconnected message about studied paleoenvironments. The paleobotanical, mostly pollen analysis studies, document just one, however basic and unique, source of Holocene information. Large areas of Central Europe, including some of the most sensitive landscapes such as sandstone areas, karstlands and loess-chernozem districts, are lacking suitable, long and continuous profiles in peatbogs and associated aquatic environments, but they contain a number of other sedimentary sequences such as tufa bodies, slope series, calcareous marshes and floodplain sediments that document different aspects of environmental change. These aspects, like intensity of carbonate metabolism, relation of humid and dry phases to prehistoric civilizations, slope dynamics and other ones, are hardly traceable by pollen methods. Marshes and peatbogs can be in Central Europe mostly found in marginal and obscure environments such as peneplainised mountains or basins that significantly differ from interior formations – e.g. it is virtually impossible to generalize the results obtained in the Trebon Basin (Southern Bohemia) to a nearby location of the Palava Mts. in Southern Moravia. We may thus treat Holocene studies as a compilation of evidence coming from individual Holocene facies. These facies may be ordered in certain patterns that correspond to Holocene stratigraphic provinces.

North-South Holocene transect is likely to intersect the Scandinavian province between Scandinavia, Northern Germany and Poland, Mid-European province between Middle Germany and Alps; and the Mediterranean province dependent more on the position of intratropical zone of convergence than on North Atlantic Oscillation. The West-East transect in area north of Alps and east of Pyrenees is dependent on the function of North Atlantic Oscillation and thus on prevailing oceanic or continental climates (Hurrell 1995; Broecker 1997). While in oceanic climate the Holocene changes can be mostly described in terms – more humid/less humid, the continental climates are due to the enhanced climatic differences more often trespassing the limits of natural systems. Droughts are more common and thus forest-grassland changes may take place. Such change may have relatively minor consequences in some regions like humid mountains, but may lead to environmental collapse in other, generally drier landscapes. The continental climate may bring phases of slope destabilization, solifluction, downcutting, rock collapses, floodplain planation and aggradation and other phenomena that are recorded in subaeric sequences.

The complementary study of various Holocene facies of Mid-Europe led K.D. Jäger (1969) to define the Holocene stratigraphy that differs from classical division by pointing out a new phase between the Atlantic and Subboreal termed as the Epiatlantic. Numerous studies carried out in recent years have proved the validity of this division. The other important aspect of Holocene studies is the relationship between biostratigraphic and geochemical methods. The study of C and O stable isotopes in the tufa body at Svaty Jan pod Skalou (Bohemian Karst) provided paleotemperature curve that could be compared to Greenland Ice Core oxygen isotopes (Zák et al. 2002), but besides temperature proxies it did not add much evidence about landscape evolution. The nearby Holocene profiles excavated in slope sediments under rocks or in cave entrances yield on the other hand a uniform pattern of sedimentation including rich molluscan assemblages that enabled to decipher Holocene history in different time slices. The field sedimentological evidence e.g. clearly demonstrates that Pre-Neolithic foam sinter horizon represents the most humid Holocene event, while buried Subboreal soil represents the driest Holocene event (Lozek 1997).

The other significant contribution of molluscan analysis is the recognition of Mid-European chernozem areas as specific facial districts where due to the prehistoric landnam the natural course of two thirds of the Holocene was altered towards steppe environment. The complete development of forest systems that is so typical of the Mid-European Holocene has never happened here (Lozek 1982). The main purpose of this article was to demonstrate that paleoecological methods combined with archaeology and sedimentology are usually not capable to reconstruct a detailed paleotemperature curve but may grant something even more valuable – the information about general or site specific landscape changes recorded as series of time slices.

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HOLOCEŃSKIE FACJE ROZWOJU WYŻYN ŚRODKOWO-EUROPEJSKICH

Streszczenie

Na rozległych obszarach Europy Środkowej zbudowanych z piaskowców, skał krasowiejących i w krajobrazach lessowo–czarnoziemnych, brak stanowisk zawierających długie i kompletne profile osadów torfowiskowych oraz innych środowisk związanych z obfitością wody. Jednak na tych obszarach występują inne środowiska sedymentacyjne, takie jak pokrywy tufowe, osady stokowe, bagna węglanowe, które dokumentują różne aspekty przemian środowiska. Przemiany w utworach węglanowych, relacje wilgotnych i suchych faz klimatu do przejawów działalności cywilizacji prehistorycznych i dynamika stoków są przedmiotem rozważań tej pracy. Artykuł podsumowuje rezultaty analiz paleoekologicznych, w tym holoceńskich zespołów ślimaków. Analizy te w powiązaniu z wynikami badań archeologicznych i sedymentologicznych poszerzają wiedzę o przemianach krajobrazów wyżynnych w poszczególnych okresach holocenu, pokazując równocześnie relacje środowisk prowincji skandynawskiej, przez północne Niemcy i Polskę, Karpaty i Alpy, do prowincji śródziemnomorskiej.

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HYPOTHESIS ON GLACIAL MORPHOLOGY ON AMBA ARADAM MOUNTAIN (NORTHERN ETHIOPIA)

In the Ethiopia the presence of glacial landforms is reported only in the mountain areas at much higher elevations (Mts. Simien, 4620 m a.s.l.; Arsi, 4180 m; Bale, 4357 m) (Nyssen et al. 2002).

Some Authors, such as E. Nilsson (1940) and J. Hovermann (1954), described glacial traces found at lower altitudes, but subsequent researchers (Semmel 1963; Potter 1976; Hastenrath 1977; Messerli, Rognon 1980) refused this hypothesis with different arguments (Nyssen et al. 2002). Moreover Kuls and A. Semmel (1962) consider that some deposits found as low as 1800 m a.s.l. result from differential weathering of basalt, rather than from glacial activity.

In particular, H. Hurni (1981, 1982, 1989) points out that in the last cold period, dated between 20.000 and 12.000 years BP, on Mts. Simien the moraine deposits are found only at altitudes over 3780 m a.s.l. and glacial cirques at altitudes over 4250 m a.s.l.

In December 2002, during an excursion to Amba Aradam Mountain, I had the opportunity to observe some very interesting geomorphologic features which seem to contradict the above mentioned data (Panizza 2002).

The area surveyed is located SW of Makale, in the Tigray region, in northern Ethiopia (Fig. 1), at an altitude of 2500 to 2780 m a.s.l. and latitude of about 13°30' N. It corresponds to an E–W stretching valley, whose head is formed by two small tributary valleys with a NE to SW arrangement. From the geologic standpoint, it is made up of a siliciclastic sandstone sequence of continental facies with quartz conglomerate, shaly and laterite levels, known as "Amba Aradam Formation" (Shumburo 1968; Beyth 1972; Dramis, Coltorti, Pieruccini 2002), also known as "Upper Sandstone" (Merla, Minucci 1938; Mohr 1962). The age of this formation is ascribable to the Cretaceous (Dramis, Coltorti, Pieruccini 2002; Nyssen et al. 2002). From the structural viewpoint, the valley's lower portion seems to be affected by an E–W oriented tectonic line which might have conditioned the original modelling. This tectonic feature does not seem to be linked to the main fault systems described by Y. Arkin et al. (1971) and by M. Beyth (1972), which are NNW–SSE and NNE–SSW oriented (Dramis, Coltorti, Pieruccini 2002).

Starting from the double head of the valley, the geomorphologic features observed are as follows (Fig.1).

– The head of the little northern valley (1a in the geomorphologic sketch of Fig. 1) is modeled within a semi-circular cavity with high and steep flanks, partially affected by rock falls which form debris accumulation at the foot of the rocky slopes. Also the head of the little NE valley (1b in the geomorphological sketch of Fig. 1) has a similar form, although it is not as clearly defined and looks more degraded by erosion and accumulation processes. Theese landforms seem to be related to glacial cirques, in a more or less good state of conservation.

– Immediately downstream of the confluence of the two small valleys, a debris deposit is found (Fig. 2 and point 2 in the geomorphological sketch of Fig. 1), made up of lithologically homogeneous elements (from the Amba Aradam Formation), though with quite a varied grain-size distribution (from blocks to sand and silt). These elements are mixed, non stratified or smoothed, with no apparent signs of friction. This deposit could be ascribed to a moraine deposit from glacial confluence.

– On the flanks of the main valley some sandstone outcrops appear to have been modeled in the form of convex and smoothed surfaces, with traces of sub-horizontal grooves which subsequently were partially sectioned by rock shattering processes (Fig. 3 and points 3 in the geomorphological sketch of Fig. 1). Theese landforms look like "*roches moutonnées*", more or less degraded.

– On the floor of the main valley a palustrine deposit is found (Fig. 4A and point 4 in the geomorphological sketch of Fig. 1); its origin is evidently due to the damming of the valley by debris accumulation.

– Two detrital bodies developed on the two sides of the valley (Fig. 4), one facing south (B) and the other north (C), appear to be responsible for the valley damming and the aforementioned palustrine deposit. On the whole these two deposits make up an arc-shaped form of varying clarity; the right-hand side deposit (B in Fig. 4 and point 5a in the geomorphological sketch of Fig. 1) has been partially terraced by anthropogenetic processes. From the grain-size viewpoint, they are both made up of very heterogeneous materials – up to boulders of about ten cubic meters in volume – resulting from the Amba Aradam Formation. Outcrop of the detrital body (Fig. 5) (C in Fig. 4 and 5b

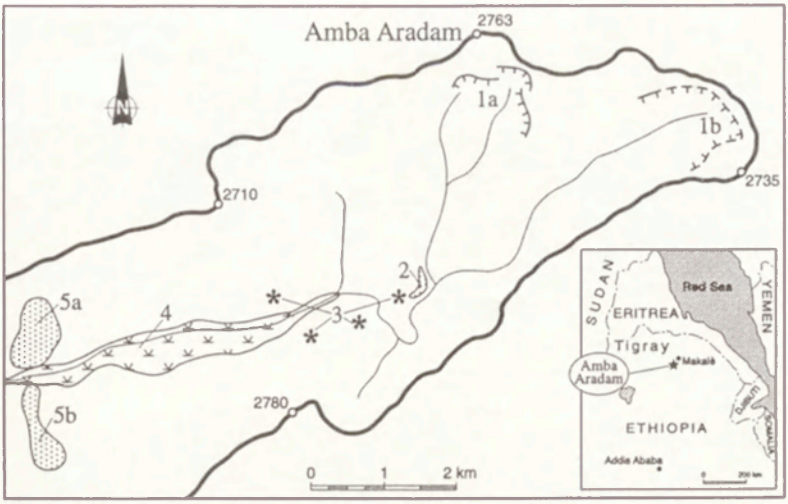

Fig. 1. Geomorphologic sketch of the Amba.Aradam south-western slope and location of area surveyed. 1 – semi-circular cavities that seem to be related to glacial cirques; 2 – debris deposit that could be ascribed to a moraine; 3 – rocks modeled in form of *roches moutonnées*; 4 – palustrine deposits; 5 – debris deposits that seem to correspond to a small moraine arc

Szkic geomorfologiczny stoków poludniowo-zachodnich Amba Aradam oraz lokalizacja badanych obszarów. 1 – półkoliste formy wklęsłe przypominające cyrki polodowcowe, 2 – osady gruzowe, mogące być morenami, 3 – powierzchnie skalne przypominające garby mutonowe, 4 – osady zastoiskowe, 5 – osady gruzowe przypominające małe luki morenowe

Fig. 2. Semi-circular cavity (point 1a in the sketch of figure 1) that seems to be related to glacial cirque in a more or less good state of conservation

Półkolista forma wklęsła (punkt 1 a na rycinie 1) przypominająca cyrk polodowcowy dość dobrze zachowany

Fig. 3. Arenaceous rocks modeled in form of *roches moutonnées* (point 3 in the sketch of figure 1)

Piaskowce modelowane na kształt garbów mutonowych (punkt 3 na rycinie 1)

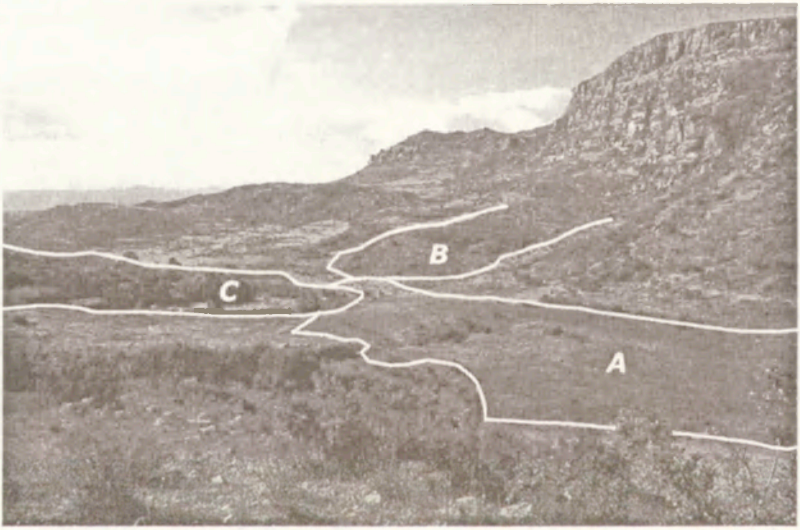

Fig. 4. Palustrine deposits (A: point 4 in the sketch of figure 1) and debris accumulations (B and C: points 5a and 5b in the sketch of fig. 1), the latter seem to correspond to a small moraine arc, with frontal moraine deposits

Osady zastoiskowe (A: punkt 4 na rycinie 1) i akumulacja gruzowa (B i C: punkty 5a i 5b na rycinie 1). Te ostatnie wydają się korespondować z małymi łukami morenowymi i osadami moreny czołowej

Fig. 5. Diamicton from deposit 5b of the geomorphological sketch of figure 1, that could be ascribed to a moraine deposit
Gruz tkwiący w drobnofrakcyjnym materiale (punkt 5b na rycinie 1) uważany za osad morenowy

in the geomorphological sketch of Fig. 1) is a diamicton-type deposit, that is unsorted with sand and coarse particles dispersed through a mud matrix. It is not stratified and some elements show a certain degree of smoothing. This set of landforms and deposits seems to correspond to a small moraine arc, with frontal moraine deposits.

This framework seems to be completed by the presence of less inclined portions of slope, similar to the so called "glacial shoulders".

The explanation of the Amba Aradam morphology, though, is in contrast with the fact that in Ethiopia the presence of glacial cirques and moraines was reported in mountains at much higher elevations. Certainly the conformation of this E–W stretching narrow and deep valley, sheltered from the wind, could have favored the persistence of snow and ice during a cold Pleistocene period. This possibility, though, needs to be further investigated by means of particularly detailed geomorphologic surveys, correlations with other similar traces in other parts of Africa placed at the same latitude, and a precise reconstruction of the climatic conditions (temperature, precipitation and wind regimes etc.) existing in the Tigray during the Pleistocene.

As an alternative hypothesis, all these features might constitute a very singular case of "geomorphologic convergence", that is, landforms that have the same shape and appearance but different genesis. In this case, the area would be a very good educational example to alert the onlooker to simplistic deductions based mainly on exterior appearances, that is, on prevalently descriptive characteristics (Panizza 1996).

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HIPOTEZA O MORFOLOGII GLACJALNEJ GÓR AMBA ARADAM W PÓŁNOCNEJ ETIOPII

Streszczenie

W górach Amba Aradam osiągających wysokości 2500–2780 m npm i położonych na szerokości geograficznej 13°30'N autor rozpoznał formy charakterystyczne dla obszarów poddanych morfogenezie glacjalnej, takie jak cyrki, wygłady lodowcowe, barki lodowcowe, osady i formy morenowe oraz osady zastoiskowe. Ten zespół form i osadów nie był dotychczas rozpoznany, chociaż jest znany z obszarów górskich Etiopii wznoszących się ponad 4000 m npm.

Jest to hipoteza wymagająca dalszych studiów, gdyż formy rzeźby o takich samych lub podobnych cechach mogą mieć inną genezę. Dlatego jest przewidywane szczegółowe kartowanie geomorfologiczne oraz dalsze prace nad rekonstrukcją warunków klimatycznych północnej Afryki w plejstocenie.

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Holocene and Late Vistulian Paleogeography and Paleohydrology
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THE DRIVING FORCE BEHIND THE COLD CLIMATE SPELLS DURING THE HOLOCENE – A WORKING HYPOTHESIS

1. DUST STORMS AND GLACIATION EVENTS DURING THE PLEISTOCENE

There is a general agreement between scientists that the Milankovitch theory of orbital forcing does explain in principle the mechanism causing glaciation and deglaciation phenomena during the Pleistocene. Yet, H.H. Lamb (1982) argued that the orbital variations, which are of a slow trend, can not explain the abruptness of the climate changes even during the Pleistocene. As triggering mechanisms having an abrupt effect, Lamb suggested either variations in the sun's output or in the amount of volcanic dust in the atmosphere. He brought as an example the massive volcanic explosions of Mount St Helens in May 1980 poured myriads of submicroscopic-sized rock into the stratosphere, where they are beyond the reach of the rain, which washes such impurities out of the lower atmosphere. The volcanic matter travels round the Earth in ten days to a few weeks. Convection and turbulence surges transfer some of the material to higher and lower heights, spreading the material into a global shadowing layer causing a cooling effect. The higher the exploded material is thrown, the longer will the cooling effect persist. The particles are liable to stay for one to seven years, or more, in the stratosphere. The cooling effect is due to their partial interception of mainly the short-wave solar radiation, while the Earth's mainly long-wave back-radiation passes nearly unhindered. Thus while, the dust layer warms up, the Earth's surface temperatures' fall. The cooling is at its maximum in the first year. In the past this ranged between 0.1 to around 1.0 °C. In 1783, when there were two very great eruptions – in Iceland and in Japan – in the same year, the combined effect may have been a cooling of the northern hemisphere by 1.3 °C, gradually tailing away to zero over the following four or five years. Indeed data from ice cores demonstrate a correlation between enhanced dust concentrations reduced carbon dioxide in the atmosphere and glacial periods (Petit et al. 1999). The effect of the dust is considered by J.H. Martin (1990) and

M.B. Aberlin (1996) to be due to the fact that it contains iron oxide, which is a fertilizing agent. This causes a bloom of the phyto-plankton the fixing of carbon dioxide and its reduction in the atmosphere. Indeed iron was found to be the limiting nutrient in the central Pacific Ocean and dust from the Gobi desert in China has been shown to induce productivity increases near Hawaii. Experiments of spreading iron containing fertilizers conducted near the Galapagos Islands induced phytoplankton blooms (Watson et al. 2000).

2. CLIMATE CHANGE FORCING MECHANISMS DURING THE HOLOCENE

While the Milankovitch forcing mechanism coupled with dust storms providest an explanation to the mechanism causing climate changes during the Pleistocene, when it comes to the Holocene, its periodicity of changes, lasting only a few centuries (maximum a millennium) and even a few decades, the question of mechanism is still debated.

J.E. Kutzbach and F.A. Street-Perrott (1985) still maintain that the Milankovitch forcing mechanism was responsible for the climate changes, which affected the lake levels in the Northern Hemisphere, tropics and subtropics since the Last Glaciation. They claim to have arrived at a good correlation between an atmospheric general-circulation model (GCM), simulating the climates of January and July at 3,000 years intervals, and the evidence regarding palaeo-levels in the Northern Hemisphere tropics and sub tropics. Both model and observations show high levels between 6,000 and 9,000 BP. However, Kutzbach and Street Perrott admit that the response of the monsoon system to the amplified seasonal cycle of solar radiation as a function of the variations in earth's orbit, are phenomena, which occur on a too long time scale to account for the millennia, not to speak on the century, long climatic changes.

F.A. Street-Perrott and R.A. Perrott (1990) found a correlation between the droughts in the Sahel and tropical Mexico since 14,000 years ago and the injections of fresh water into the northern part of the North Atlantic. The decrease in salinity reduces the formation of North Atlantic Deep Water circulation, which is an important factor in the regulation of the global oceanic circulation. This results in the cooling and a change in the distribution of sea surface temperatures, affecting the rainfall in the northern tropics. Still the reasons for the fresh water injection phenomena are not yet clear.

W. Dansgaard et al. (1984) claim that there exists a quasi periodicity of 2550 years in the oxygen-18 oscillations of the deep Greenland ice cores, which brought them to the conclusion that the forcing of these cycles is connected with the cycles of solar radiation. Since the Milankovitch effect is too slow for these periodicity, they suggest to look for the reason in solar productivity. They point on the evidence that the carbon-14 concentration in atmospheric carbon dioxide has varied in a non-random way in the past and that its Holocene variations are in antiphase with climatic temperature changes including the 2550 year cycle. They also point out that the Berilium-10 concentration, which is a cosmogenic isotope, in the ice is in antiphase with that of the oxygen-18. They thus speculate that low solar activity in terms of particle emissions, is connected with cold climates.

H. Oeschger et al. (1984) suggest that the variations in the Berilium-10 concentrations mainly reflect changes of the precipitation rate, although they note that solar induced variations of this isotope by a factor of 1.5, have been found for the Maunder minimum of a quiet sun lasting from 1640 to 1710 AD.

The cold period, which occurred between ca 850 and 760 BC was caused according to B. Van Geel and H. Renssen (1998) by reduced solar activity, evidenced by the higher levels of ^{14}C in the atmosphere. This can be concluded from the divergence between the dates arrived by carbon 14 dating methods and those calculated by dendro-chronological methods, namely the number of tree rings. The reduced solar activity, which means reduced ultraviolet radiation, may also lead to a decline in ozone production in the lower stratosphere. This could have resulted in a decrease of the latitudinal extension of the Hadley Cell circulation, which may lead to a weakening of the monsoons, as well as to an expansion of the Polar Cells and the displacement towards the south of the cyclonic tracks.

The rapid carbon dioxide changes and the fact that the oceans form the main carbon dioxide reservoirs affecting climate changes on the time scale of ten thousand of years, brought H. Oeschger et al. (1984) to suggest that the correlation between carbon dioxide concentration and palaeo-temperatures are connected with the upwelling phenomena which regulates the enrichment of the surface water by phosphate. They suggest that during periods of more rapid vertical mixing, there is enrichment in biological productivity, which results in higher carbon dioxide concentration. On the other hand if the upwelling of deep water is reduced the organisms will have time to consume all the phosphate, which will result in the lowering of carbon dioxide concentration.

E.T. Degens et al. (1984) suggested that there might be a few simultaneous driving forces, which act as triggers, mostly effective when they happen together with an orbital insolation signal, for example sun-spots. D.J. Schove (1984) claimed that the relationship of sunspots and climate depends on the magnitude of the sunspot cycles. In the case of the medieval Nile floods, he found a reversed relationship.

3. VOLCANIC DUST AS A TRIGGERING FACTOR

S.C. Porter (1986) demonstrated a close relation between the pattern of the northern hemisphere glacier variations during the last millennium and the volcanic aerosol production, as found in the Greenland ice cores. He thus maintains that sulfur-rich aerosols generated by volcanic activity are a primary factor in forcing climate change on the decade level. He found that glaciers advance whenever acidity levels rise sharply from background values to reach concentrations > 1.2 meq. H+/kg above background. The glaciation lags by about 10–15 years behind the beginning of the acidity increase.

R.S. Bradley et. al. (1987) found extremely sharp drops in temperature, particularly in fall months, occurred after several major volcanic eruptions. High temperatures are sometimes associated with major El-Nino years. When the two occur more or less simultaneously their influence is minimized

R.A. Bryson and R.U. Bryson (1996, 1998) developed a methodology of archaeo-climatic modeling of palaeo-climates based on a series of hierarchical steps, each of which relies on certain underlying assumptions. Accordingly they calculated various parameters such as the rate of loading by volcanic dust, impact on optical depth and insolation of dated volcanic events. These they used to predict ice volume, glacial area and albedo, the values of which were combined to model seasonal hemispheric mean temperatures, which could then be used to ascertain the timing of major atmospheric circulation features. Synoptic climatology was then used to determine the relationship between contemporary monthly precipitation or temperature and pertinent circulation features. This was applied to the modeled past monthly positions of the relevant circulation features in order to calculate past precipitation or temperature at specific intervals. The Brysons' general conclusion is that volcanic eruptions were the main triggering factor for climate changes during the Holocene. They suggest that the explosions, strong enough to load the atmosphere with a quantity of aerosols, which affected atmospheric optical depth, modulated also insolation. By a series of calculations they arrived at time-series for atmospheric circulation features at different locations.

K.B. Taylor and B.F. Molnia (1993) have measured the electrical conductivity (ECM) of the layers of a core in the summit of the Greenland ice sheet. This is affected by their dust content, which by neutralizes their acidity. The oxygen 18 composition a palaeo-climate indicator and the ECM records were found to be synchronous, which shows that the ECM is related to climate changes. Layers with heavier oxygen–18 compositions, which correspond to warmer climates show also higher ECM values, which is an evidence of lower dust content. The fluctuations in the ice conductivities were found to be abrupt, which means that conditions of dust transport also changed abruptly. This means that the periods of climate change were also periods of rapid changes in the atmospheric circulation and changes of wind speeds due to changes in the temperature gradients between high and mid northern latitudes.

L.A. Scuderi (1990) correlated tree-ring variations in the Sierra Nevada and the sulfur rich aerosols recorded in the Greenland ice cores on the decadal level. The good correlation brought him to the conclusion that indeed volcanic eruptions are significant forcing mechanism on the time scale of tens to hundreds of years of climate changes. He maintains that while single volcanic events may only lead to short colder periods a cluster of significant events may lead to glaciation.

4. THE COMBINED MECHANISM – VOLCANISM AND DUST STORMS

These observations has brought the present author to suggest that indeed volcanic activity was an important if not major cause for the periods of cold climates during the Holocene (Issar 2003). In other words, if one assumes that due to the Milankovitch cycles the Holocene is a post glacial and thus a warm period, then one needs a cooling mechanism for the discontinuation of the steady state warm phase. Yet, the volcanic dust veil as a cooling mechanism may work only as a very short-term effect but does not answer the long duration phenomena lasting for a few centuries to more than a thousand years. Thus one needs some kind of an additional mechanism that will produce a carbon dioxide sequestering sink, which will keep the global cooling process going on for a longer period.

Such a process might occur once the "volcanic effect" is viewed as a triggering factor of high-pressure systems over the Gobi and Sahara deserts, which generate dust storms like those which occurred during the glacial periods These occurred also during historical periods as shown by the data collected by A. Bücher (1986). He found a correlation between contemporaneous cyclonic storms entering the Sahara and "red rains" over southern Europe. Moreover, his survey of Roman and Monastery chronicles brought him to produce a curve of periods of high frequencies of such events during historical periods. The peak was found around ±0 B.C.E/C.E, which according to A.S. Issar (2003) was a cold period.

The combined model will be the following:

Since after the Last Glacial period, the relatively steady state of global climate is that of an interglacial, namely warm. Once there are volcanic euptions (especially when they converge with a minimum of solar radiation the atmosphere is being loaded with enough aerosols to affect atmosphericoptical depth and thus to modulate insolation, causing the cooling of the oceans' surface, causing lower average temperatures of the atmosphere. As a result of this short period of cooling, the cold continental high pressures zone over Siberia (the Okhotsk high) extending over the Gobi desert is strengthened. In the same time the movement of the northern jet stream and the ITCZ southward will enable cyclonic storms to enter over the Sahara and produce dust storms (Bücher op.cit.) which will export big quantities of dust, similar to what happened during the Pleistocene glacial periods (and Younger Dryas but on a smaller scale). This will further reduce the insolation but even more important, it will add fertilizers especially iron (which was found to be the limiting nutrient in the central Pacific Ocean) and phosphates to the oceans. This process will lead to a bloom of the phyto-plankton, as a short term CO^2 sink, which will cause a bloom of the zoo-plankton which will forming a long term sink, due to the sequestering by the carbonate skeletons. This will continue until certain equilibrium is reached, and once there is a period of less volcanic activity the CO^2 will return slowly to the atmosphere and a stage of warm steady-state will be reached again.

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PRZYCZYNA WYSTĘPOWANIA OKRESÓW OCHŁODZENIA KLIMATU W HOLOCENIE – HIPOTEZA ROBOCZA

Streszczenie

Autor przedstawia hipotezę sugerującą, że przyczyną zmian klimatu w holocenie była działalność wulkaniczna. Wzmożona aktywność wulkanów powodowała szybki efekt oziębienia, który prowadził do formowania układów wysokiego ciśnienia nad pustyniami Gobi i Sahary. Jest udowodnione, że aerozole ograniczają efekt insolacji a przez to powodują ochłodzenie powierzchni oceanów i obniżenie średniej temperatury atmosfery. Burze pyłowe charakteryzowały okresy glacjalne i ochłodzenia w okresie postglacjalnym. Efektem burz pyłowych były "czerwone opady" bogate w tlenki żelaza i fosforany, działające jako czynniki użyźniające środowisko biologiczne oceanów. To z kolei wywoływało zakwit fitoplanktonu i formowanie krótkoterminowego obniżenia poziomu CO_2. Dalszą konsekwencją był zakwit zooplanktonu i usunięcie CO_2 w węglanowych skorupach. Gdy został osiągnięty pewien poziom równowagi, przy niższej aktywności wulkanicznej, następował powrót do stadium równowagi stałej charakterystycznej dla warunków cieplejszych.

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