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CHANGES IN THE GEOGRAPHICAL ENVIRONMENT OF POLAND

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CONTENTS

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PROGRESS IN RESEARCH ON THE EVOLUTION OF THE GEOGRAPHICAL ENVIRONMENT OF POLAND

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AIM AND ORGANIZATION OF THE RESEARCH

A new research programme for the period of 1981–1985 entitled "Changes of the geographical environment of Poland" and registered under the symbol MR I–25 was set up in 1980. The programme was coordinated by the Institute of Geography and Spatial Organization, Polish Academy of Sciences.

The aim of the programme was to identify the origin and mechanisms of the geographical environment evolution as well as the stages and tendencies of current changes which occur under an increasing human impact. It will be followed by a forecast of the changes of the environment in the future on the basis of the inherited features and the current transformations (Fig. 1).

The studies concentrated under this project differ from other programmes dealing with natural resources and nature protection in so far that the other programmes focus on the identification of the degradation due to pollution of the air, water and soil while the project in question is addressed to recognize the longlasting tendencies in the functioning and transformations of geosystems.

The present-day geographical environment of Poland consists of the elements of the nature (geological structure, relief, climate, water, plant cover and animals) which have been substantially modified already and which are still being modelled by man. These elements are interlinked, though they are not of the same age and importance in their functions in the present-day environment. When some of these elements are usually a background for the occurring phenomena (geological structure, relief) the other ones take an active part in the circulation of matter (soil, vegetation) or even are the major carriers of energy (water). The substratum rocks, which supply the mineral substances taking part in the circulation, are the oldest elements of the environment and their age in Poland varies from 2 billion years to a few thousand years. The relief developed on them is of various ages from the Paleogene to the young glacial relief of the last 20 thousand years (Klimaszewski 1980). The foundations of the soils of Poland are vari-aged as well. The last cold stage can be most frequently recognized in the present-day soils. The plant cover is the effect of the evolution of the forest in the Holocene and that of the following anthropogenic modifications. The same is the case of the animal kingdom. Climate is the most autonomous element of the present, and is related to the atmospheric circulation, although local conditions substantially modify it. Water circulation of the atmospheric origin is similarly differentiated by other elements of the environment.

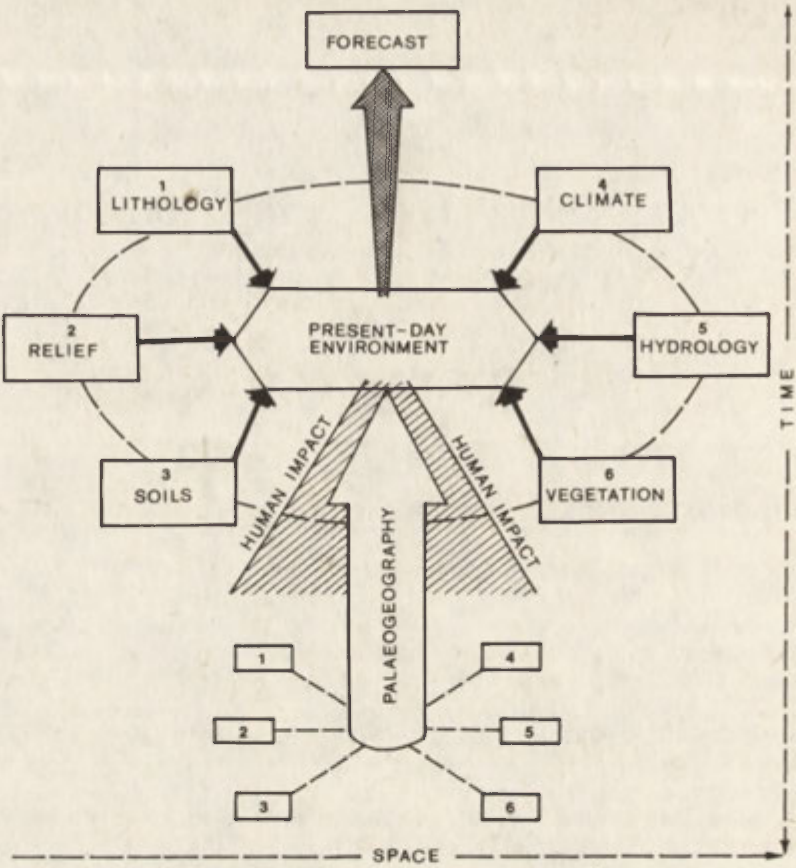

Fig. 1. Ideal model of the organization of research in the national project MR I—25 "Changes of the geographical environment of Poland" (showing changes in space and time)

Thus, the present-day environment is the coexistence of the vari-aged elements whose foundations are related to the distant or recent past. Therefore, the identification of the age of these foundations and that of the rate of changes seems to be very important in order to understand the restoration of the natural resources. The restoration is, in principle, proportional to the time and rate of formation (Fig. 2). Degradation of mineral resources, relief, soils and deep reservoirs of groundwater which require many thousand or even millions of years, and which were often formed under climatic or tectonic conditions different from the present-day ones, can be considered as the irreversible changes.

The project MR I—25 involved the departments of the Institute of Geography and Spatial Organization, Polish Academy of Sciences (Departments of Geomorphology and Hydrology in Cracow and Toruń, and Warsaw departments of Climatology, Organization of the Environment, and of Biogeography), the university geographical centres (Poznań, Warsaw and to the less extent Toruń, Lublin, Cracow, Łódź), schools of pedagogy (Kielce, Słupsk), geological departments of the University of Mining and Metallurgy in Cracow and that of the Warsaw University, Radiocarbon Laboratory of the Silesian Polytechnical University, Department of the Natural Resources and Nature

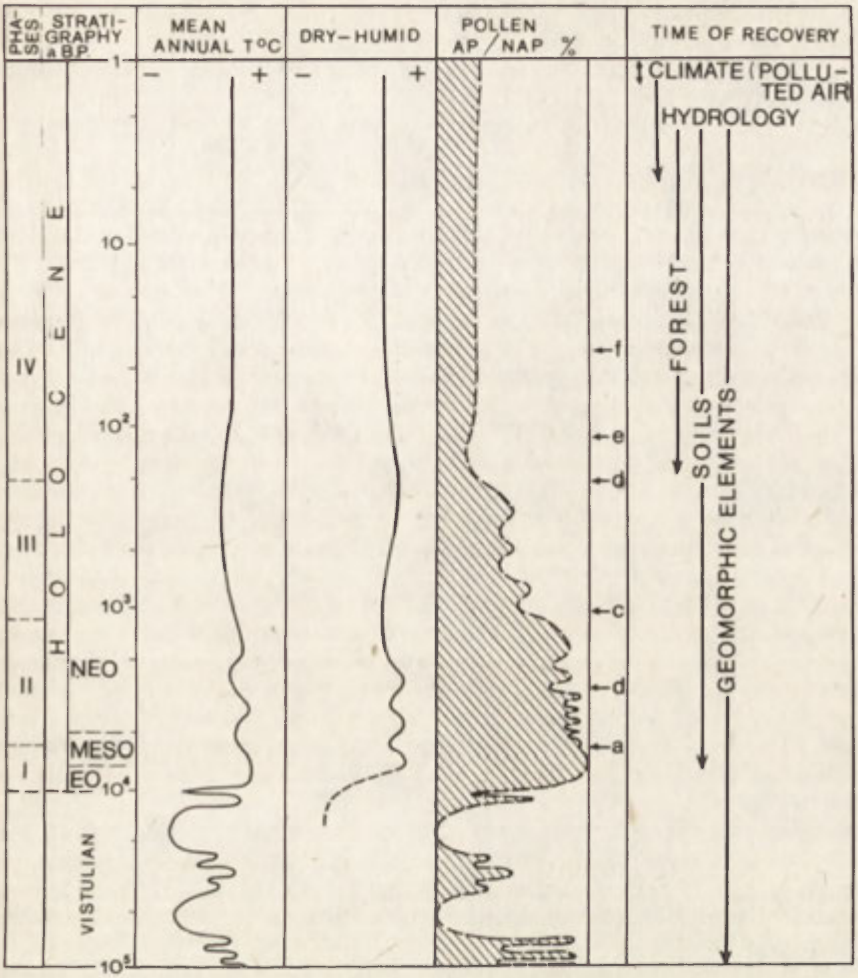

Fig. 2. Environmental changes in southern Poland (in logarithmic scale). Main phases during Holocene: I — before deforestation, II — pre-historic agriculture, III — historical (intensified agriculture), IV — industrialization. Important episodes in environmental changes: a — first Neolithic landnam phases, b — increased deforestation in late Bronze age, c — mediaeval deforestation, d — introduction of potatoes, beginning of braided channels, e — regulation of river channels and lowering of ground water level, f — construction of water reservoirs

Protection and the small teams or individual persons of the national institutes: the Research Institute of Forestry, the Institute of Cultivation, Fertilization and Pedology and the Institute for Land Reclamation and Grassland Farming.

The studies were concentrated in 6 research groups and were carried out at various scales: national, regional and the detail ones – especially the latters were based on research stations (Fig. 3). One of the research groups aimed at the paleogeographical reconstruction of the last cold stage and the Holocene, the second group – at the selected topics of the evolution of the valleys during the last 15000 years. The task of the next two groups was the study of the mechanism of the changes and contemporary processes in the area of mountains, uplands and lowlands. The macro-scale investigations of the elements of the environment of the country and studies of the regions being under diverse human impacts, were the goals of separate groups.

PALEOGEOGRAPHY OF THE VISTULIAN AND HOLOCENE

The last cold stage, from which originate the features of the glacial and periglacial relief and glacial, fluvial, slope and eolian (loess, dunes) deposits, as well as the Holocene — the period of a still lasting warm forest climate are most substantially marked in the Polish landscape up-to-date. Therefore, the studies of the project MR 1—25 concentrated on the history of the nature in the last decades. The fundamental transformations are shown in the logarithmic scale in order to emphasize the changes which have taken place during the period of human activity.

The results were based on the radiocarbon datings. Two hundred and fifty datings were made for the project in the laboratory of Gliwice (M. Pazdur et al. in press). The changes of climate and vegetation have been recognized in the early, middle and late Vistulian (Stankowska and Stankowski 1985; Tobolski 1984) and the double glaciations in the last cold stage have been proven (Drozdowski, 1986). The detail facial analysis of the deposits of the marginal zones of the Last Glaciation in the western Poland became the basis for a reconstruction of the deglaciation (Kasprzak and Kozarski 1984; Kozarski et al. 1985). The melting of the dead ice blocks turned out to be older than the Alleröd (Niewiarowski, in press). The concept of an abstract post-glaciotectonic surface of the period of the older glaciation has been suggested for the periglacial zone (Rotnicki 1985). The grain size analysis and that of the quartz abrasion enabled one to delimit the extent of the glaciofluvial and extraglacial waters at the foreland of the last ice-sheet (Rotnicki, in press). Among the stages of the transformation of the periglacial zone of the valleys older phase of the Vistulian has been learned more precisely (Klatkowa and Turkowska 1984).

The particular attention has been paid to the valley evolution during the last 15000 years as a result of the cooperation with the IGCP—158 Programme entitled "Paleogeography of the temperate zone during the 15000 years". The studies of ca. 20 reaches of the Vistula valley and its tributaries (Starkel, ed., 1981, 1982, in press), on the sections of the Prosna valley (Rotnicki 1983), those of the Warta river (Kozarski 1983; Kozarski and Tobolski, eds, 1981; Kozarski et al. 1984) and those of the Słupia river (Florek 1984) have been completed. The role of the climatic changes, that of vegetation and of the human impact have been evidenced, and Falkowski's hypothesis (1975) on the changes in the nature of the river channels from the braided into the meandering ones in the Lateglacial and, in turn, into the braided ones during the recent centuries has been generally confirmed. The change of the braided river channels into the meandering ones at the threshold of the Bölling has been proven in the middle Warta valley (Antczak 1985; Kozarski et al. 1984). However, the transformation of the large meanders towards the small ones has been first recognized by Szumański (1983) in the San valley. The phase of the anthropogenic aggradation in the last millenium was recognized at the best in the Wisłoka valley (Starkel, ed. 1981) and in the upper Vistula valley (Rutkowski and Starkel, eds. 1984). The role of the eustatic changes in the Baltic base level (Drozdowski, Tomczak, Wiśniewski — in: Starkel ed. 1982) and that in tectonic movements (Niedziałkowska et al. 1985) have been determined in the Vistula valley as well. The studies on the inserts of the channel alluvia facies, channel parameters and channel shifts became the basis to state the more stable periods in the Holocene alternated with humid phases of a large frequency of floods (Starkel 1983) which correspond to the cooling and increased humidity, the latters registered in the fluctuations of glaciers, landslide activity, and changes in the lake levels and those in the groundwater table. Incidentally, the hypothesis about the achievement of the threshold values during the Holocene phases of a large intensity of floods and about the reestablishment of the equilibrium latest of all after several hundred of years, has been put forward and partially proven (Starkel 1983).

The analysis of the relation between the channel parameters and deposits, and the hydrological parameters of the Prosna river allowed to reconstruct the discharges in the past based on the deposits and forms of the paleochannels (Rotnicki 1983; Rotnicki and Borówka 1985). The evolution of the chosen closed systems were simultaneously studied stating the oscillations of the lake levels and changes of the type of denudation of the dominance of chemical or mechanical ones (Borówka 1985; Wojciechowski 1985). Recently, K. Więckowski has discovered the deposits of the laminated gyttja, with a recorded annual rhythm, in the lake Gościąż near Płock in the zone of the maximum extent of the last ice-sheet. The above will enable to calibrate many stratigraphic methods. In all the cases, a substantial change in the type and rate of sedimentation after deforestation have been stated. The washing process was initiated simultaneously with the beginning of the Neolithic agriculture (Wasylikowa et al. 1985). Based on numerous datings of the profiles with the fossil soils, the original model of the multi-phase development of soils in Poland has been elaborated, indicating the role of humid phases in a maturation of the soil profiles (Kowalkowski 1984). The changes in the sedimentation rate of the peatlands have been summarized by S. Żurek (1984) parallely with the attempts of a large team of paleobotanists working on the history of the lakes and peatlands, the team acting outside this project (under the supervision of M. Ralska–Jasiewiczowa). The studies on the malacofauna history allowed to reconstruct the thermal and water changes as well those in forestation (Alexandrowicz 1983). The molusca turned out to be a very sensitive indicator of the local changes.

The following have been stated based on the entire studies of the increasing human impact on the environment: – due to deforestation and cultivation of the soils a general acceleration of the energy and matter circulation has been achieved which resulted in aggradation in depressions, sometimes directly at the foot of the slope, when the removal of the material lag behind the supply.

PRESENT-DAY PHYSICO-GEOGRAPHICAL PROCESSES IN VARIOUS TYPES OF ENVIRONMENT

The field studies were aimed at determination of the type, direction and rate of the processes of the exchange of energy and matter under the natural conditions and those of the human activity. The studies were concentrated in the Carpathians, in the belt of the uplands, in the region of the Włocławek reservoir and on the Pomeranian Morainic Plateau (Fig. 3).

The studies on the radiation balance and exchange of the energy between the atmosphere and ground were continued (Paszynski et al. 1983; Grzybowski 1984). The mechanism of the transformation of precipitation into the overland flow and throughflow and transfer into the runoff from a catchment has been recognized in the Carpathians (Słupik 1981; Froehlich and Słupik 1986). The role of the type and changes of the land use in the runoff modelling have been stated in the Ropa catchment (Soja 1981). The oscillations of the ground water table and ground water temperatures were investigated on the flat bank in the surrounding of the Włocławek reservoir (Glazik 1986). In relation to the catastrophic ice-jam and flood in the zone of the backwater of the Włocławek reservoir the influence of the depth and the type of channels onto the ice cover and ice-jam formation was determined (Grześ 1986).

Numerous works were devoted to the present-day morphogenetic processes. The relations between the magnitude of leaching and the carbonate content in the substratum rocks – loesses, marls (Janiec 1985), flysh rocks (Welc 1985) and glacial

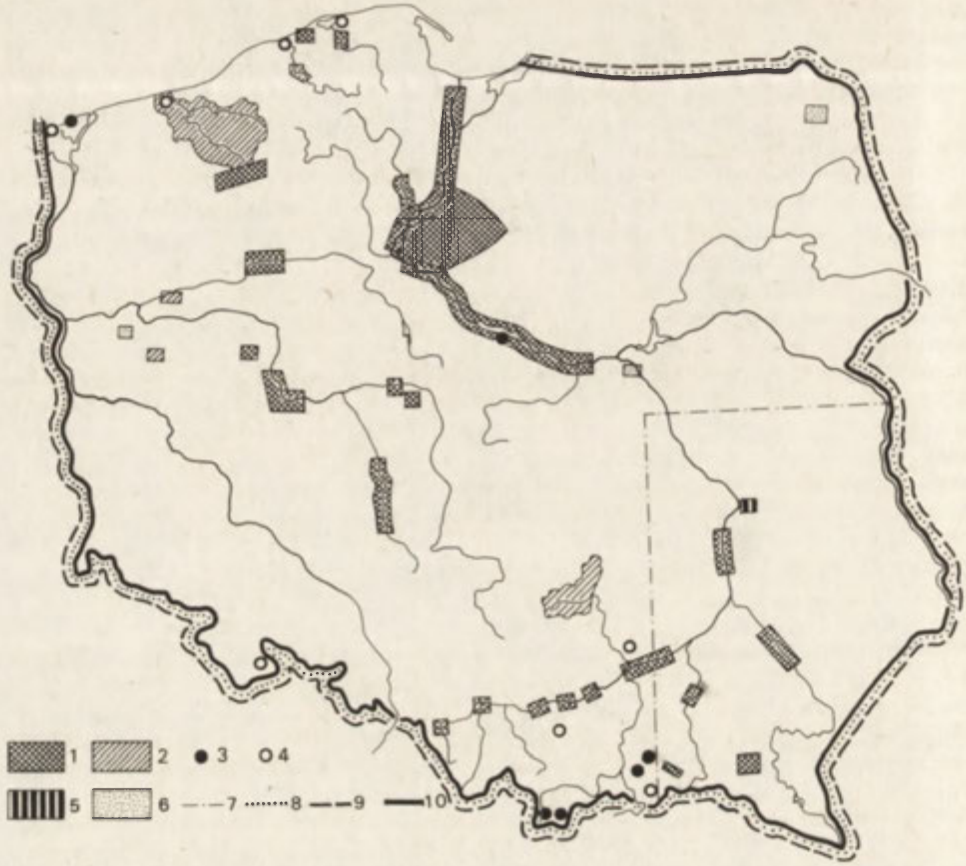

Fig. 3. Areas investigated and surveyed in the project MR I—25: 1 — paleogeographic research, 2 — studies of present-day processes, 3 — field stations with monitoring of hydrologic and geomorphic processes, 4 — detailed topoclimatic studies, 5 — detailed phytogeographic studies, 6 — detailed applied environmental studies, 7 — areas covered by general maps of geomorphic processes, 8 — areas covered by general climatological maps, 9 — areas covered by general mapping of potential vegetation, 10 — areas covered by general mapping of geosystems (physico-geographic typology)

deposits (Kostrzewski and Zwoliński 1984) as well as the rate of inter-soil circulation (Froehlich 1982) have been recognized. The measurements on the cart roads and the splash experiment enabled one to state that 80—90% of the suspended loads in the channel originate from the roads in many regions (Froehlich and Slupik 1985; Froehlich — in press).

The gravitational processes were subject to the detail studies of E. Gil and A. Kotarba. These studies were continued in a cooperation with E. Thiel of the Institute of the Hydroengineering, Polish Academy of Sciences. Based on the measurements of the rockfalls and of the transfer on the alluvial cones in the Tatra Mts several models of the modifications of the high-mountain slope have been elaborated (Kotarba et al. 1983; Kotarba et al. — in press). The photogrametric studies of the cliff on the Wolin Island allowed to recognize the role of the lithology on the type of the cliff modelling

(Kostrzewski and Zwoliński — in press). The directions of the evolution and maturation of the erosional edge at the water reservoir were also investigated (Banach 1986).

The studies of the relation between the slope and channel systems are of particular importance for the methods of the determination of the denudation magnitude to be more precise. The wider the valley floors (with an increase of the catchment area) the more independent slope systems (Froehlich 1982; Krzemień 1985, Biernat 1985). The detailed studies referred to the transportation mechanism, modelling of channels and floodplains of the meandering lowland rivers (Rotnicki and Borówka 1985; Młynarczyk 1985) as well as to the mountain and fore-mountain braided rivers (Baumgart—Kotarba 1983, 1985). Whereas the former ones served as a basis for a reconstruction of discharges (Rotnicki 1983), the letter ones were used to learn the role of floods of various magnitudes by means of the aerial photographs (Baumgart—Kotarba 1985).

The works on the zonality of the relief-forming processes in the Tatra Mts (Kotarba et al. 1987) and in the Khangai Mts in Mongolia (Kowalkowski and Starkel 1984) were of synthetic character.

The detail studies of the distribution of the organic matter in the coniferous forest (Grabińska 1985), corresponding to those on the productivity of the forest and meadow ecosystems (Trampler et al., 1983; Żurek 1984) were carried out independently. The monograph of the coniferous forests of Poland (Breymeyer, ed. in press) has been submitted for publication.

The investigations of the physical, chemical and biological processes have indicated that different types of land use may bring negative effects for the magnitude and quality of the resources. Water circulation and transportation of the mineral material are subjected to a permanent acceleration. The importance of various forms of protection practices like channelization, dam constructing, drainage of fields, intensified forestation had been also investigated and the negative results of these practicies, which make a chain of transformations causing the final decrease of water, soil and other resources in our country, had been indicated.

MACROSCALE STUDIES OF THE STATUS AND CHANGES OF THE ENVIRONMENT

These studies were carried out parallely to the detailed ones and concentrated on the elaboration of the series of the general maps of Poland in the scale of 1:300000 (maps were or will be printed in the scale of 1:500000). Among these maps the geomorphological (Starkel, ed. 1980) and hydrological ones (Galon, ed. 1980) were elaborated before 1980. The following two maps were completed in 1981—1985. The map of the potential vegetation shows the actual possibilities for a restoration of the natural habitats (W. Matuszkiewicz 1985). The map was accompanied by the detailed studies of the chosen forest communities (Matuszkiewicz and Kozłowska 1984) and the field weeds. The map of the landscape types of Poland was elaborated by the group as well, and it is mainly based on the geomorphological-lithological criteria (Kondracki and Richling, eds. 1984). This map is a continuation of the research initiated by Kondracki (1978). The elaboration of the maps of the present-day processes (Radłowska, Bogacki and the team) presenting the spatial distribution of the intensity and conditioning of the chemical and physical processes was started from the sheet Lublin. The concept of the synthetic map stressing the differentiation of the areas with the dominance of degradation and agradation (Bogacki and Zgorzelski 1985) was also elaborated. The map of the changes in the chemism of the present-day soils of Poland indicating the acidity of the soils even on the calcareous substratum (Kern 1984), is of substantial value.

The group of J. Paszyński (Grzybowski et al. 1985) is preoccupied with the elaboration of the map of the components of the radiation balance in another scale. The above attempts are accompanied by the studies on the typology of topoclimates (Grzybowski 1985). The bioclimatological atlas of Poland has been prepared in the less detailed scale (Kozłowska—Szczęsna, ed. 1986) as well. The atlas is based on the detailed studies of the optimum conditions for work and rest of man concentrated in numerous resorts (Kozłowska—Szczęsna, ed. 1981, 1985; Krawczyk 1985; Błażejczyk 1983). The investigations on the productivity of the forest and meadow ecosystems have been carried out in the nation-wide scale under the supervision of A. Breymeyer (see above).

APPLIED ASPECTS OF THE STUDIES OF THE ENVIRONMENT

Although the problems of the forecast of the environmental changes will be emphasized in 1986—1990, numerous investigations in the detailed, regional as well as in the nation-wide scales have dealt with applied and forecast aspects.

The most prominent among these works are those of the group of A. S. Kostrowicki which concentrated on the evaluation of the environmental resources, determination of the human impact and recommendations for the most rational utilization of these regions, the latter being frequently the areas of the conflicting development of various branches of the economy. The monograph of the surrounding of the Wigry Lake has been elaborated recognizing the conditions for the development of forestry, agriculture, recreation and tourism and providing the proposal for the rational utilization of this region (Kostrowicki, ed. 1987). Simultaneously, the studies are carried out in the urbanized suburb-areas of Warsaw (Łomianki district) as well as in the agricultural--industrial region of Pińczów where, after approximately 25 years since the first survey, the changes in the elements of the environment and in the pattern of land use are being registered.

All the elaborations of the general map are of the applied nature, indicating either the direction of the transformations (the studies of the ecosystem productivity, of the relief-forming processes) or the possibilities of the utilization of resources for a reconstruction of the plant communities (the map of the potential vegetation) or for the health protection (the bioclimatological atlas or the development of the resort network).

However, the practical conclusions of equal importance arise, from the detailed studies of the present-day processes which are carried out in the field research stations and which are coupled with the analysis of the changes in a historical period. The intensification of the landslides on the scarp at the Włocławek reservoir or the danger of the ice-jam floods (cf. Szupryczyński ed. 1986) indicate the necessity for changes in the nature of the land use and the water management. The studies in the research stations in Szymbark and Homerka (Frycowa) enabled one to formulate the detail proposals for the protection against erosion and floods in the mountain part of the Vistula catchment, such as elimination of potato crops, limitation of arable fields and of the cart-road network, limitation of the water management structures on the streams and others (cf. Słupik 1981; Froehlich 1982, 1985; Gil 1985; Starkel 1985).

It has to be emphasized that numerous crucial proposals have been directed to the regional authorities (Suwałki, Włocławek, Nowy Sącz) as well as to the appropriate departments responsible for the water management and to research councils (e.g. Council for the Mountain Area Management).

METHODOLOGICAL PROGRESS IN 1981—1985

One of the aims of the investigation was to make the existing methods in the environmental studies of the past and of the present-day acting processes to be more precise as well as to elaborate the new ones.

In the reconstruction of the past transformations, the most prominent is the international recognition of the Radiocarbon Laboratory of the Institute of Physics of the Polytechnical University of Gliwice, following the improvement of both the method of measurements (new, more precise counters, automatic records) and of the preparation and combustion of samples (A. Pazdur et al. 1986; M. Pazdur et al . 1986). The sedimentological-facial methods have been successfully applied for the reconstruction of the glacial, glacio-fluvial and fluvial environments (Kasprzak and Kozarski 1984; Rotnicki 1985; Starkel, ed. 1981). K. Rotnicki (1983) and Rotnicki and Borówka (1985) have based the reconstruction of discharges not only on an analysis of the channel and sediment properties but also on the comparison with the relations between the present-day hydrological regime and sediment load transport. The malacological method with the application of the numerical techniques has been successfully used (Alexandrowicz 1983) in order to reconstruct the habitats. The analysis of heavy metal content (Klimek and Zawilińska 1985) and coal gravels (Rutkowski 1986) has been applied to differentiate the young alluvia.

Some original methods have been introduced in the studies of the present-day relief-forming processes, e.g. in the cognition of the influence of the gradient and the substratum type on the splash intensity (Froehlich in press), while the station in Homerka took part in the international programme of investigations of the supply sources to the channel using the radioactive Cs method. The complex studies on the landslides have been extended by the introduction of geophysical methods. The repeated serial photographs have served for measurements of the channel changes after floods while the space images — for the identification of the tectonic structures in the intramontaneous Orawa—Nowy Targ Basin (Baumgart—Kotarba 1983). The general research programme for the network of physico-geographical stations after 1985 has been prepared as well (Kotarba, Froehlich, Kostrzewski and others).

In the field of the bioclimatological studies, the methodological textbook evaluating the methods applied both abroad and in the country, including many proposed by the authors themselves, has been elaborated (Kozłowska—Szczęsna, ed. 1985). The programme of the ecological monitoring prepared under the supervision of A. Breymeyer (1984), is applied in the institutions of the national network. New techniques and methods of the identification and evaluation of various geoecological elements, functioning under the conditions of human impact, have been elaborated as well (Kostrowicki 1981; Kostrowicki, ed. in press).

PERSPECTIVES OF THE STUDIES

During the 5-year long existence of the programme "Changes of the geographical environment of Poland" numerous synthetic works have been completed on the level of research subjects and themes. Unfortunately, the synthesis has not been achieved. Numerous problems are currently summarized in the "Natural environment" section of the first volume of "Geography of Poland". Among the others, S. Kozarski has presented the paleogeographical changes of Poland during the last cold stage while A. Kowalkowski, H. Maruszczak and M. Ralska-Jasiewiczowa, K. Rotnicki and L. Starkel have shown the changes of the geographical environment in the Holocene due to the climatic changes and the superimposed increasing human impact. The present-day climate of Poland is presented by J. Paszyński and T. Niedźwiedź, the relief-forming processes by the group under the supervision of M. Bogacki and L. Starkel, the vegetation by W. Matuszkiewicz, and the productivity of ecosystems by A. Breymeyer. The results of the studies have been presented in a series of lectures given at the Polish Academy of Sciences during the academic year 1983/1984. The majority of the texts of the above lectures are currently in press edited by L. Starkel.

The series of syntheses and monographs, including the synthesis of the paleogeography of the Holocene of Poland, the map of the present-day relief-forming processes with the commentary, the monograph of the hydrological conditions and others, are expected to be prepared by the end of 1990.

FINAL REMARKS

In the above described MR I—25 Programme coordinated by the Institute of Geography and Spatial Organization, Polish Academy of Sciences attended about 200 research and technical workers (including 50 professors and associate professors) who published over 150 dissertations and papers, and submitted for publication ca 70 papers. The six research group leaders coordinated the work of the groups and organized the annual report-meetings as well as the scientific conferences: professor Karol Rotnicki, associate professor Teresa Kozłowska—Szczęsna, professor Adam Kotarba, professor Jan Szupryczyński, professor Edmund Falkowski (working until the end of 1983) and professor Andrzej S. Kostrowicki.

The central office was located in the Department of Geomorphology and Hydrology of the Mountains and Uplands in Cracow. The main organizational work was done by the late associate professor January Slupik (until the end of 1982) and his successor dr Roman Soja with technical assistants: Mrs Teresa Mrozek and Miss Joanna Jankosz.

REFERENCES

Alexandrowicz S. W., 1983, Malacofauna of Holocene calcareous sediments of the Cracow Upland, *Acta Geologica Polonica* 33, 1—4, 117—158.

Antczak B., 1985, Rhythmites on lower terraces of the Warta river, Poland, and their paleohydrologic implications, *Quaestiones Geographicae, Special Issue* 1, 31—43.

Banach M., 1986, Przekształcania brzegów zbiornika włocławskiego (Transformation of the banks of the Włocławek reservoir), *Dokumentacja Geograficzna* 5.

Baumgart—Kotarba M., 1983, *Kształtowanie koryt i teras rzecznych w warunkach zróżnicowanych ruchów tektonicznych (na przykładzie Podhala)*. (Channel and terrace formation due to differential tectonic movements with the Eastern Podhale basin as example), Prace Geograficzne IGiPZ PAN 145.

Baumgart—Kotarba M., 1985, Different timescales of examining the river bed and valley floor evolution (braided river Białka Tatrzańska as example), *Studia Geomorphologica Carpatho-Balcanica* 19, 61—76.

Biernat T., 1985, Areas supplying weathered material and the mechanism of transportation of the suspended material in the Łososina catchment area (Holy Cross Mountains, Central Poland), *Quaestiones Geographicae, Special Issue* 1, 61—73.

Błażejczyk K., 1983, Bioklimatyczna ocena i typologia uzdrowisk Polski (Bioclimatic evaluation and typology of Polish health resorts), *Dokumentacja Geograficzna* 3.

Bogacki M., Zgorzelski M., 1985, *Przestrzenne zróżnicowanie natężenia współczesnej denudacji fizycznej w Polsce południowo-wschodniej* (Spatial differentiation of the intensity of the present-day physical denudation in the SE Poland), Manuscript in IGiPZ PAN, Cracow.

Borówka R. K., 1985, Późnoglacjalne i holoceńskie rozmiary denudacji młodoglacjalnych obszarów wysoczyznowych w strefie maksymalnego zasięgu ostatniego zlodowacenia. Studium metodologiczne (Magnitude of the Late Glacial and Holocene denudation of the young glacial upland areas in the zone of maximum extent of the Last Glaciation. Methodological Study), *Quaestiones Geographicae.*

Breymeyer A., 1984, Ecological monitoring as a method of land evaluation *Geographia Polonica* 50, 371—384.

Breymeyer A., ed., in press, Lasy iglaste na obszarze Polski, geografia, ekologia, użytkowanie (Coniferous forests in Poland, geography, ecology, utilization), *Dokumentacja Geograficzna.*

Drozdowski E., 1986, *Stratygrafia i geneza osadów zlodowacenia wistulianu w północnej części Dolnego Powiśla* (Stratigraphy and origin deposits of the Vistulian Glaciation in the northern part of the Lower Powiśle), Prace Geograficzne IGiPZ PAN 146.

Falkowski E., 1975, Variability of channel processes of lowland rivers in Poland and change of the valley floors during the Holocene, *Biuletyn Geologiczny* 19, 45—78.

Florek W., 1984, *Ewolucja i mechanizm transformacji dolin Słupi i Łupawy (syntetyczne opracowanie wyników badań przeprowadzonych w latach 1981—1984)* Evolution and transformation mechanism of the Słupia and Łupawa valleys (synthetic study of the results of investigations carried out in 1981—1984), Manuscript in IGiPZ PAN, Cracow.

Froehlich W., 1982, *Mechanizm transportu fluwialnego i dostawy zwietrzelin do koryta w górskiej zlewni fliszowej* (The mechanism of fluvial transport and waste supply into the stream channel in a mountainous flysch catchment), Prace Geograficzne IGiPZ PAN, 143.

Froehlich W., 1985, Zabudowa koryt i użytkowanie den dolin pod kątem ochrony przeciwpowodziowej i przeciwerozyjnej (Channel management and valley floor utilization versus anti-flood and anti-erosional protection practicies), *Problemy Zagospodarowania Ziem Górskich.*

Froehlich W., in press, Influence of slope gradient and supply area on splash, *Zeitschrift für Geomorphologie.*

Froehlich W., Słupik J., 1986, Rola dróg w kształtowaniu spływu i erozji w karpackich zlewniach fliszowych (The role of roads in the generation of flow erosion in the Carpathian flysch drainage basins), *Przegląd Geograficzny* 58, 1—2.

Galon R., ed., 1980, *Przeglądowa mapa hydrograficzna Polski 1:500000* (Hydrographic map of Poland 1:500000), IGiPZ PAN.

Gil E., 1985, Racjonalne użytkowanie ziemi na stokach pod kątem ochrony przeciwpowodziowej i erozyjnej (Rational land use on slopes versus anti-flood and anti-erosional practicies), *Problemy Zagospodarowania Ziem Górskich.*

Glazik R., 1986, Dynamika temperatury płytkich wód podziemnych na przykładzie pradoliny Wisły w rejonie stopnia wodnego "Włocławek" (Dynamics of the shallow underground water with the Vistula pradolina in the area of the water gate "Włocławek" as example), *Dokumentacja geograficzna* 5.

Grabińska B., 1985, *Szybkość rozkładu materii organicznej a właściwości środowiska w borach środkowej Polski* (Rate of the organic matter decomposition and the environment properties in forests of the middle Poland), Ph.D. Thesis, IGiPZ PAN, Warsaw.

Grześ M., 1986, Wpływ zbiornika Włocławskiego na zmianę ustroju lodowego Wisły (Influence of the Włocławek reservoir on the change in the Vistula ice regime), *Dokumentacja Geograficzna* 5.

Grzybowski J., 1984, Essai d'etablissemt d'un modele de l'influence des elements du milieu geographique sur le bilan d'energie de la surface active de la terre, *Geographia Polonica* 50, 225—231.

Grzybowski J., 1985, *Podstawy wyróżnienia i klasyfikacji topoklimatów na podstawie analizy struktury bilansu cieplnego powierzchni czynnej* (Fundamentals of differentiation and classification of the topoclimates on the base of the analysis of the heat budget of the active surface), Manuscript in IGiPZ PAN, Cracow.

Grzybowski J., Miara K., Paszyński J., 1985, *Rozkład przestrzenny i przebieg roczny bilansu promieniowania w Polsce* (Spatial distribution and annual pattern of the radiation budget in Poland), Manuscript in IGiPZ PAN, Cracow.

Janiec B., 1985, *Rola wód infiltracyjnych w odwapnianiu lessów z pogranicza Roztocza Zachodniego i Wzniesień Urzędowskich (zlewnia Górnej Sanny)*, (Role of the infiltration water in the loess decalcification of the boundary area between the Western Roztocze and Wzniesienia Urzędowskie (the upper Sanna catchment), Manuscript in IGiPZ PAN, Cracow.

Kasprzak L., Kozarski S., 1984, Analiza facjalna osadów strefy marginalnej fazy poznańskiej ostatniego zlodowacenia w środkowej Wielkopolsce (Sum.: Facies analysis of marginal zone deposits produced by the Poznań Phase of the last glaciation in Middle Greast Poland), *Uniwersytet Adama Mickiewicza, Seria Geografia 29.*

Kern H., 1984, Some problems of geochemical evolution of Polish soils, *Geographia Polonica* 50, 315—327.

Klatkowa H., Turkowska K., 1984, *Rozwój i funkcjonowanie dolin denudacyjnych. Raport końcowy* (Development and functioning of the denudational valleys. Final Report), Manuscript in IGiPZ PAN, Cracow.

Klimaszewski M., 1980, Bezwzględny wiek rzeźby terytorium Polski (Sum.: The absolute age of the relief of the territory of Poland), *Studia Geomorphologica Carpatho-Balcanica* 10, 3—16.

Klimek K., Zawilińska L., 1985, Trace elements in alluvia of the upper Vistula as indicators of paleohydrology, *Earth Surface Processes and Landforms* 10, 273—280.

Kondracki J., 1978, *Geografia fizyczna Polski* (Physical geography of Poland), PWN, Warszawa.

Kondracki J., Richling A., eds, 1984, *Typologia krajobrazu Polski w skali przeglądowej 1:500000* (Typology of the Polish landscape in the scale 1:500000), manuscript map.

Kostrowicki A. S., 1981, Transformation of natural environment under the impact of human activities — a historical aspect, in: *Interaction of the prehistoric man and his environment*, V Symposium of the Commission of Environmental Problems, IGU, Mexico 1981.

Kostrowicki A. S., ed., in press, *Studium geoekologiczne rejonu jezior wigierskich* (Geoecological study of the Wigry Lake region), Prace Geograficzne IGiPZ PAN, 147.

Kostrzewski A., Zwoliński Z., 1984, Zastosowanie metody hydrometrycznej do analizy tempa denudacji chemicznej na obszarze nizinnym Pomorza Zachodniego: Wstępne opracowanie metodyczne (Application of the hydrometric method for the analysis of chemical denudation rate in the lowland area of the Western Pomerania. Primary methodological study), *Sprawozdania Poznańskiego Towarzystwa Przyjaciół Nauk nr 100 za 1982 r.*, 150—158.

Kostrzewski A., Zwoliński Z., in press, Operation and morphologic effects of present-day morphogenetic processes modelling the cliffed coast of Wolin Island, northwestern Poland, *Earth Surface Processes and Landforms.*

Kotarba A., Kaszowski L., Krzemień K., in press, *High-mountain denudation system*, Geographical Studies, Special Issue 3.

Kotarba A., Kłapa M., Rączkowska Z., 1983, Procesy morfogenetyczne kształtujące stoki Tatr Wysokich (Present-day transformation of alpine granite slopes in the Polish Tatra Mountains), *Dokumentacja Geograficzna* 1.

Kowalkowski A., 1984, *Ewolucja gleb w holocenie* (Soil Evolution in the Holocene), Manuscript in IGiPZ PAN, Cracow.

Kowalkowski A., Starkel L., 1984, Altitudinal belts of geomorphic processes in the southern Khangai Mts (Mongolia), *Studia Geomorphologica Carpatho-Balcanica* 18, 95—116.

Kozarski S., 1983, River channel adjustment to climatic change in west central Poland, in: K. J. Gregory, ed., *Background to paleohydrology*, Wiley and Sons Ltd.

Kozarski S., Tobolski S., eds, 1981, Symposium "Paleohydrology of the temperature zone", Poznań, Poland' 81, September 22—28, vol. 1. *Guide-book of excursions*, Adam Mickiewicz University, Poznań.

Kozarski S., Antczak B., Gonera P., 1984, *Przemiany dna doliny środkowej Warty podczas późnego wistulianu i holocenu* (Changes of the middle Warta valley floor during the Late Vistulian and Holocene), Manuscript in IGiPZ PAN, Cracow.

Kozarski S., Kasprzak L., Nowaczyk B., 1985, *Sedymentologiczne podstawy paleogeograficznej analizy zlodowacenia wistulianu w Polsce środkowo-zachodniej* (Sedimentological basis of paleogeographical analysis of the Vistulian glaciation in the Middle Western Poland), Manuscript in IGiPZ PAN, Cracow.

Kozłowska—Szczęsna T., ed., 1981, Problemy bioklimatologii uzdrowiskowej, cz. 4 (Problems of bioclimatology of health resorts, part 4), *Dokumentacja Geograficzna* 2.

Kozłowska—Szczęsna T., ed., 1984, Problemy bioklimatologii uzdrowiskowej, cz. 5 (Problems of bioclimatology of health resorts, part 5), *Dokumentacja Geograficzna* 1—2.

Kozłowska—Szczęsna T., ed., 1985, Metody badań bioklimatu człowieka (The methods of man's bioclimate research), *Problemy Uzdrowiskowe* 1/2 (207/208).

Kozłowska—Szczęsna T., ed., 1986, Wyniki badań bioklimatu Polski, cz. 1 (Results of the bioclimatic research of Poland, part 1), *Dokumentacja Geograficzna* 3.
Krawczyk B., 1985, Związek między strukturą bilansu cieplnego człowieka a niektórymi elementami meteorologicznymi (Relation between the structure of the heat balance of man and some meteorological elements), *Problemy uzdrowiskowe* 5/6 (211/212).
Krzemień K., 1985, Współczesne modelowanie koryta w dolinie glacjalnej na przykładzie Doliny Starobociańskiej w Tatrach (Present-day modelling of the stream-bed glacial valley on the example of Starobociańska Valley in the Tatra Mountains), *Studia Ośrodka Dokumentacji Fizjograficznej* 13, 299—335.
Matuszkiewicz J. M., Kozłowska A., 1984, *Przegląd fitosocjologiczny zbiorowisk leśnych Polski* (Phytosociological review of the Polish forest communities. Part VII. oak-forests), Manuscript in IGiPZ PAN, Cracow.
Matuszkiewicz W., 1985, *Mapa roślinności potencjalnej* (Map of potential vegetation), unpublished.
Młynarczyk Z., 1985, Rola wielkości i kształtu ziarna w transporcie fluwialnym (Role of grain size and shape in fluvial transport), *Poznańskie Towarzystwo Przyjaciół Nauk, Prace Komisji Geograficzno-Geologicznej* 21.
Niedziałkowska E., Gilot E., Pazdur M. F., Szczepanek K., 1985, The Upper Vistula valley near Drogomyśl in the Late Glacial and Holocene, *Folia Quaternaria* 56, 101—132.
Niewiarowski W., in press, The phases of transformation of subglacial channels into river valleys: a case study of the lower Vistula region, *Acta Universitatis Nicolai Copernici, Seria Nauki Matematyczno-Przyrodnicze, Geografia* 20.
Paszyński J., Miara K., Grzybowski J., 1983, Bilans promieniowania w Polsce (Radiation budget in Poland), in: *Materiały Zjazdu Geografów Polskich. Toruń 15—18 IX 1983. Streszczenia referatów*, Uniwersytet Mikołaja Kopernika, Toruń, s. 168.
Pazdur A., Awsiuk R., Goslar T., Pazdur M., 1986, Spalanie próbek organicznych i oczyszczanie CO_2 przeznaczonego do pomiarów radiowęglowych (Burning of organic samples and cleaning of CO_2 used in the radiocarbon measurements), *Zeszyty Naukowe Politechniki Śląskiej*, Geochronometria 2.
Pazdur M., 1982, Badanie dokładności datowania metodą C^{14} późnoplejstoceńskich i holoceńskich osadów organogenicznych (Studies on exactness of radiocarbon datings of the late Pleistocene and Holocene organogenic deposits), *Zeszyty Naukowe Politechniki Śląskiej*, Seria matematyczno-fizyczna 41.
Pazdur M., Pazdur A., Awsiuk R., Goslar T., 1986, Metody preparatyki wstępnej próbek przeznaczonych do pomiarów radiowęgla (Methods of the primary preparation of samples for radiocarbon datings), *Zeszyty Naukowe Politechniki Śląskiej*, Geochronometria 2.
Rotnicki K., 1983, Modelling past discharge of meandering rivers, in: K. J. Gregory, ed., *Background to Paleohydrology*, Wiley and Sons Ltd.
Rotnicki K., 1985, Postglacitectonic surface as a basis of analysis of fluvial relief development in areas of glacial deposition, *Quaestiones Geographicae, Special Issue* 1, 147—158.
Rotnicki K., 1986, Granica zasięgu zlodowacenia wistulian granicą obszarów o różnym stopniu eolizacji osadów ostatniego okresu zimnego (The border of the Vistulian extent as the border of areas of a differentiated eolization of deposits of the last cold period), *Sprawozdania Poznańskiego Towarzystwa Przyjaciół Nauk nr 103 za* 1984 r.
Rotnicki K., Borówka R. K., 1985, Definition of subfossil meandering paleochannels, *Earth Surface Processes and Landforms* 10, 215—225.
Rutkowski J., 1986, The occurrence of carboniferous coal of anthropogenic origin in the contemporaneous Vistula River sediments near Cracow, Southern Poland, *Earth Surface Processes and Landforms.*
Rutkowski J., Starkel L., eds., 1984, *Holocen okolic Krakowa. Materiały Sympozjum* (Holocene of the surroundings of Cracow), Wydawnictwo AGH, Kraków.
Słupik J., 1981, *Rola stoku w kształtowaniu odpływu w Karpatach fliszowych* (Sum.: The role of slope in generation of runoff in the Flysch Carpathians), Prace Geograficzne IGiPZ PAN, 142.
Soja R., 1981, Analiza odpływu z fliszowych zlewni Bystrzanki i Ropy (Beskid Niski), (Analysis of

runoff from the Bystrzanka and Ropa flysch catchments (the Beskid Niski Mts.), *Dokumentacja Geograficzna* 1.

Stankowska A., Stankowski W., 1985, Litostratygrafia i zasięg fazy leszczyńskiej oraz fazy poznańskiej zlodowacenia Vistulian w okolicach Konina (Litostratigraphy and extent of the Leszno and Poznań stages of the Vistulian Glaciation in the vicinity of Konin), *Sprawozdania Poznańskiego Towarzystwa Przyjaciół Nauk* nr 101 za 1983 r.

Starkel L., ed., 1980, *Przeglądowa mapa geomorfologiczna Polski 1:500000* (Geomorphological Map of Poland), IGiPZ PAN.

Starkel L., ed., 1981, The evolution of the Wisłoka valley near Dębica during the Late Glacial and Holocene, *Folia Quaternaria* 53.

Starkel L., ed., 1982, *Evolution of the Vistula river valley during the last 15000 years, part 1*, Geographical Studies, Special Issue 1.

Starkel L., 1983, Reflection of the hydrologic changes in the fluvial environment of the temperate zone during the last 15000 years, in: K.J. Gregory, ed., *Background to paleohydrology*, Wiley and Sons Ltd.

Starkel L., 1985, Zróżnicowanie przestrzenne środowiska Karpat i potrzeby zmian użytkowania ziemi (Spatial differentiation of the Carpathian environment and the need of changes of the land use), *Problemy Zagospodarowania Ziem Górskich.*

Starkel L., ed., 1987, *Evolution of the Vistula river valley during the last 15000 years*, part 2, Geographical Studies, Special Issue 4.

Szumański A., 1983, Paleochannels of large meanders in the river valleys of the Polish Lowland, *Quaternary Studies in Poland* 4, 207–216.

Szupryczyński J., ed., 1986, Procesy hydrologiczne i geomorfologiczne w otoczeniu zbiorników zaporowych na dolnej Wiśle (Hydrological and geomorphological processes in the vicinity of water reservoirs on the Lower Vistula river), *Dokumentacja Geograficzna* 5.

Tobolski K., 1984, *Badania paleobotaniczne interglacjału eemskiego i wczesnego wistulianu z Władysławowa koło Turka* (Paleobotanical studies of the Eemian interglacial and Early Vistulian in the area of Władysławów near Turek), Manuscript in IGiPZ PAN, Cracow.

Trampler T., Giržda A., Dmytrko E., 1983, *Regiony ekologiczno-fizjograficzne i ich rozmieszczenie w makroregionach fizycznogeograficznych* (Ecological-physiographical regions and their distribution within the physical-geographical macroregions), Manuscript in IGiPZ PAN, Cracow.

Wasylikowa K., Starkel L., Niedziałkowska E., Skiba J., Stworzewicz E., 1985, Environmental changes in the Vistula valley at Pleszów caused by Neolithic man, *Przegląd Archeologiczny* 33, 19–54.

Welc A., 1985, Zmienność denudacji chemicznej w Karpatach fliszowych (na przykładzie zlewni potoku Bystrzanka), (Variability of chemical denudation in the Flysch Carpathians (with Bystrzanka catchment as example)), *Dokumentacja Geograficzna* 5.

Wojciechowski A., 1985, *Analiza litofacjalna osadów Jeziora Gardno i jej znaczenie dla paleogeografii obszaru* (Lithofacial analysis of the Lake Gardno deposits and its significance for a paleogeography of the area), Ph.D. Thesis, Adam Mickiewicz University, Poznań.

Żurek S., 1984, *Akumulacja i decesja materii organicznej w torfowiskach na obszarze Polski. Synteza* (Accumulation and decay of the organie matter in peatlands of Poland. Synthesis), Manuscript in IGiPZ PAN, Cracow.

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THE TERTIARY ENVIRONMENT OF POLAND

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It is almost thirty years since M. Klimaszewski published his work *Rozwój geomorfologiczny terytorium Polski w okresie przedczwartorzędowym* (1958a). Since that time geological and geophysical explorations for coal, gas and oil have revealed the existence of a buried Tertiary relief in the Polish Lowlands and sub-Carpathian basins, and even beneath the Carpathians. Palaeobotanical studies also provided new data on the evolution of the Tertiary floras. Thus, it is possible to reconstruct the climates of the Tertiary. In 1967 there appeared the *Plejstocen Polski Środkowej na tle przeszłości w górnym trzeciorzędzie* by S. Z. Różycki, and in 1972 the first volume of *Geomorfologia Polski* edited by M. Klimaszewski. These books included regional syntheses of the geomorphological evolution of the lowlands, uplands and mountains in Tertiary and Quaternary times.

At the beginning of the Tertiary there came into being the three principal palaeogeographical units of Poland's territory which showed different developmental tendencies. The vast Tethys Ocean prevailed in the south throughout the Lower Tertiary. In upper Tertiary times, emergence of the Carpathians took place there. In the north, the present Polish Lowlands showed dominant tendencies towards subsidence causing repeated transgression of the shallow Lower Tertiary epicontinental seas. In upper Tertiary times, here extended the central inland basin and later on either a periodically flooded area or a residual lake of marine origin. The above units were divided by a land-bridge named the meta-Carpathian elevation (J. Nowak 1927). The remaining portion of it is the present belt of both ancient orogens and uplands marked by a diversity of structure-controlled relief types. The history of relief will be described against the background of vegetation and climatic conditions prevailing in the different phases of the Tertiary.

Since most of Poland's territory has been buried in Quaternary times by the Fennoscandian inland-ice, the state of preservation of the old surfaces differs widely. Furthermore, the ice-free areas were subjected to intense periglacial processes. As a consequence, the Tertiary relief had to be reconstructed in southern Poland (L. Starkel 1965). The various erosional-denudational relief types surround frequently, in a series of steps, the great tectonic elevations reflecting both the vertical zonation of features and successive stages in the relief evolution. The latter may be inferred from a complex analysis of the different genetic-chronologic relief types, associated smaller landforms, and of waste- and allogenic sheets. The text is illustrated by four palaeogeomorphological maps which are based on the published data cited under "References".

During the 63 million years which followed from the recession of the Cretaceous sea till the Quaternary (G.S. Odin 1982 in: K. Pożaryska and W. Pożaryski 1984) the environment of Poland underwent many transformations. These were due to changes of the major palaeogeographic features, relief, climate, hydrographic conditions, flora, fauna and nature of both weathering and morphogenetic processes. The opening of the Atlantic furrow and eastward flow of moist air from the Atlantic Ocean, the crumbling of the Tethys Ocean, changes in the shapes of land-masses, the uplift of the great mountain chains of the Alpine system, and the gradual dessiccation of the African continent (van Steenis 1962) brought about essential changes in the general atmospheric circulation over Europe. In Upper Tertiary times, the climate of Central Europe became cooler and, locally, even drier (M. Łańcucka – Środoniowa 1957; W. Szafer 1961, and others). Environmental changes are reflected in the type of weathered materials, deposits and landforms which have been inherited from the pre-Quaternary period.

THE PALAEOGENE

Relief of the oldest foundation persisted in the Karkonosze block, Eastern Sudetes Mts and the Sudetic Foreland. Throughout the Mesozoic era relief planation took place on the Cretaceous islands (H. Teisseyre 1960) leading to the destruction of both

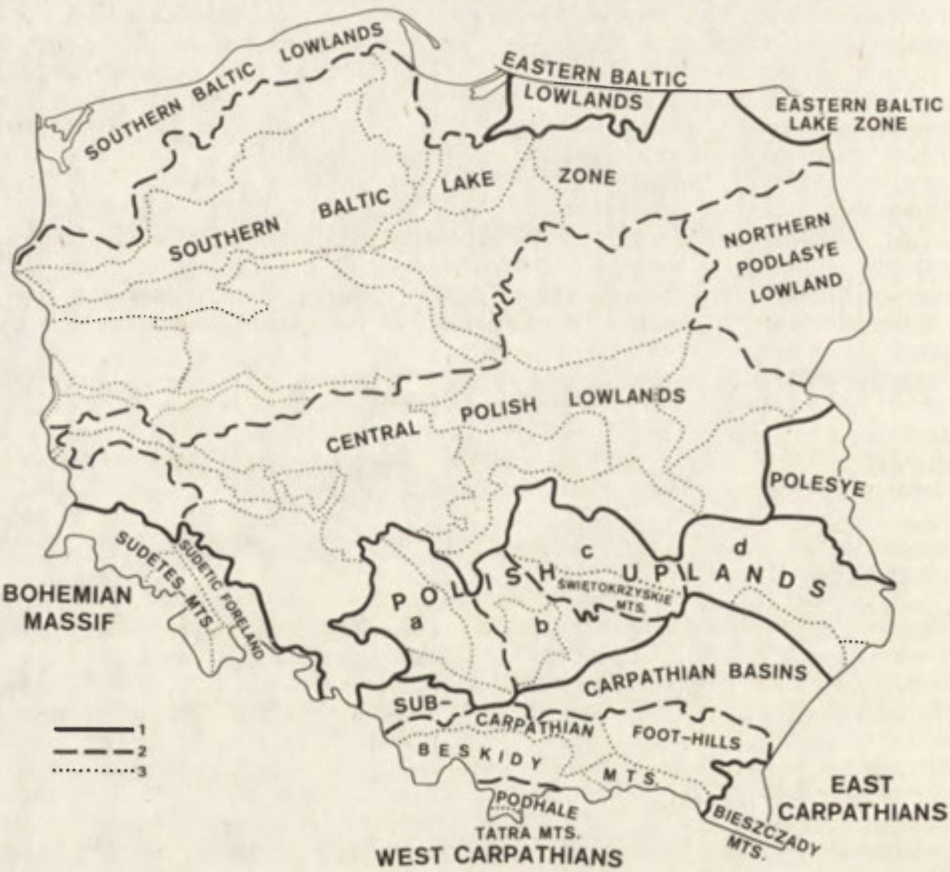

Fig. 1. Major geomorphic units of Poland: 1 — province, 2 — subprovince, 3 — macro-region, a — Silesian-Cracow Upland, b — Nida Depression, c — Kielce Upland, d — Lublin Upland

Variscan and older orogens. Although the post-Hercynian planation surface has been deformed by Laramide and later tectonic movements, it seems likely that surfaces in the Sudetes Mts might also preserve elements of a Cretaceous landscape. The Mesozoic-Palaeogene peneplain occurs in central and western Europe (C. Embleton and J. Demek 1984).

The Laramide orogeny resulted in ą "rise" and depression topography in which the Kuyavy – Pomeranian and meta-Carpathian elevations were the largest upwarps. The former elevation was accompanied by the Szczecin – Miechów and Mazowsze – Lublin synclinoria (M. Książkiewicz et al. 1965).

These orogenic movements also included faulting. Grabens were initiated between Poznań and Brzeg (P. Karnkowski 1980; Z. Deczkowski and I. Gajewska 1980), near Rawicz (M. Piwocki 1975) and in the southern part of the Cracow Upland (S. Dżułyński 1953).

By the Middle Palaeocene the whole area of Poland was a denuded land-mass (E.Odrzywolska – Bieńkowa et al. 1979), only in the south the vast Tethys Ocean continued to exist. Throughout the Lower Tertiary, lasting ca 40 million years, crustal stability and climatically controlled factors led to the formation of an extensive degradational surface named the "Palaeogene planation surface" (L. Sawicki 1925). This surface truncated different Laramide structures.

CLIMATE AND VEGETATION

In Poland phases of cooler climate corresponded with phases of orogenic activity (W. Szafer 1961). A distinct climatic change expressed in the drop of the mean annual temperature occurred at the onset of the Palaeogene (van der Hammen 1957 in: W. Szafer 1961). In Eocene times, there was a warm-temperate to slightly subtropical climate having alternating wetter and drier seasons (W. Kuźniar 1910; J. Lilpop 1957; W. Szafer 1958; I. Grabowska 1983). Similar climatic conditions prevailed in Poland till the middle of the Miocene (J. Lilpop 1957).

Environmental factors affected the plant communities. Since the Upper Palaeocene north-western Poland has been occupied by swamps. The vegetation consisted then of lycopods, mosses and ferns, and, locally, of deciduous shrubs and trees (*Fagaceae*, *Juglandaceae*, *Myricaceae*). During the early Eocene there appeared *Taxodiaceae*, *Palmae* and *Ulmaceae*. In Middle Oligocene times, the majority of trees was coniferous (*Pinus*, *Picea*, *Abies*, *Tsuga* and *Taxodiaceae-Cupressaceae*). Deciduous trees (*Fagaceae*, *Quercoidites*, *Betula*, *Corylus*, *Alnus*, *Carpinus*, *Myricaceae*) also covered the above area (I. Grabowska 1983).

The Eocene Tatric flora including *Lauraceae*, *Cinnamonum*, *Palmae*, *Ficus*, *Cornaceae* and *Myrtaceae* (W. Kuźniar 1910; W. Szafer 1958) resembled the famous subtropical flora at Geiseltal near Halle (GDR) (G. Krumbiegel 1959). A similar flora thrived in Poland also later, till the Middle Miocene (J. Lilpop 1957; E. Odrzywolska – Bieńkowa et al. 1979). Reconstruction of the climatic-ecological conditions on the basis of palaeobotanical data shows subtropical conditions which controlled the type of both weathering and morphogenetic processes.

WEATHERING AND MORPHOGENETIC PROCESSES

Warm-temperate to subtropical climatic conditions favoured the intense karstification of both limestones and ore-bearing dolomites (K. Bogacz et al. 1973, and others). The calcareous Cretaceous rocks have been deeply decalcified, giving rise to the decalcified pumice-like *opoki* and gaizes in the Lublin Upland and north-eastern foreland of the Świętokrzyskie Mts (W. Pożaryski 1951, and others). On the outcrops of the sandy Cretaceous rocks there developed the ferrous "moulding" sands containing fragments of gaizes, Cretaceous sandstones and silicified Jurassic limestones and flints (M. Blaszak 1970, and others). The above residues survived in different karst depressions and crevices in the Cracow Upland. Insoluble clayey residues of *terra rossa* type, either of Palaeogene or of Tertiary age in general, are to be found on the outcrops of soluble Upper Jurassic, Triassic and Palaeozoic rocks (S. Gilewska in: M. Klimaszewski 1972). The Liassic sandstones were affected by silicification (S. Z. Różycki 1967, 1972).

Fragments of the fossil largely Palaeogene weathering profile, more than 100 m deep, are preserved in the Carpathian foredeep. Epigenetic changes of the Carboniferous siltstones and sandstones also proceeded beneath the thick sheet of Miocene marine deposits (W. M. Kowalski 1979, and others).

Throughout the Lower Tertiary there have been formed the exogenetic uranium minerals in the granitic Karkonosze block (K. Mochnacka 1979) as well as nickel, chrysoprase and magnesite veins on serpentine in the present Sudetic Foreland (M. Klimaszewski 1958a, and others). Recent mineralogic studies revealed that typical laterite is absent from the area (S. Dyjor 1985a).

THE ORIGINAL "PALAEOGENE PLANATION SURFACE"

This surface "was not a peneplain, but it connected with each other areas at different stages of advanced development" (M. Klimaszewski 1958a). It included low mountains and hills, plateaus, rolling surfaces and plains of varying origin. The original surface persisted only as a buried feature below the Miocene fill of the Carpathian foredeep and below the Oligocene marine deposits in the present Polish Lowlands.

Within the Carpathian foredeep the original pre-Badenian surface (Table 1) slopes southward beneath the Carpathians. It is preserved as an extensive smooth summit plane being dissected by deep valleys and tectonic scarps in the west. The low gradients (0.5 – 5%) indicate advanced planation (N. Oszczypko and A. Tomaś 1976; W. M. Kowalski 1979; Z. Buła and D. Jura 1983; E. Jawor 1983; W. Bogacz et al. 1984; S. Jucha 1985a, b). The surface truncated the Carboniferous, Jurassic and Cretaceous rocks which are weathered to great depths and, locally, silicified (N. Oszczypko and A. Tomaś 1976; W. M. Kowalski 1980 – 1981). In the south-west the Mesozoic-Palaeogene peneplain was attacked by the waves of the Tethys Ocean. Materials derived from the denuded land accumulated in the Carpathian geosyncline (M. Książkiewicz et al. 1965; J. Kotlarczyk 1979; S. Traczyk 1978, and others).

In the Tethys basin, extending northward of the present Pieniny Klippen Belt, deposition of the flysch series prevailed throughout the Palaeogene. In the south there existed a denuded land-mass which at the middle of the Eocene became slowly drowned giving rise to the Tatric island. This had a luxuriant evergreen vegetation. Although the island was dissected, its relief was rather low showing both flat and cliffed coasts and tors (E. Passendorfer 1959; P. Roniewicz 1969).

During the Palaeogene the majority of northern Poland was a denudational-depositional plain. The extremely low gradient on the land was indicated by sedimentation of the Lower- and Middle Oligocene marine deposits in the tidal zone being similar to the present North Sea watten (J. Liszkowski and J. Stochlak 1969; I. Grabowska and M. Piwocki 1975). Depressions buried beneath a sheet of early Tertiary marine, brackish and terrestrial sediments show that the Palaeogene relief developed in several stages (K. Matl and T. Śmigielska 1977) (Annexe A). A radiometric date from brown coal in north-western Poland suggests that scarplands and tectonic depressions can be traced back to 50.1 + 0.7 million years ago (T. Uberna 1972 in: E. Ciuk 1975).

THE TRANSFORMED PLANATION SURFACE

Within the axis of the meta-Carpathian elevation a surface of a Palaeogene foundation is thought to be represented in the tops of ranges, plateaus and ridges. In the tectonically deformed Proto-Sudetes (including the present Sudetes Mts and Sudetic Foreland) it is difficult to reconstruct. A typical feature of the regolith-covered Proto-Sudetes were hard-rock residuals which resembled the present subtropical inselbergs (W. Walczak in: M. Klimaszewski 1972).

In the Sudetes Mts traces of the "Palaeogene planation surface" are preserved as flat summit planes at between 700 m and 1000 m, and above 1000 m a.s.l. These correspond to the first relief horizon which has been distinguished by A. Jahn (1953). The summit planes are overlooked in the south and west by massive monadnocks such as Mt Śnieżka (1603 m) and by tors of a Tertiary foundation (M. Klimaszewski 1958a, and others). In Oligocene – Lower Miocene times, stripping of the regolith due to climatic deterioration and destruction of the close subtropical forest in the present Sudetes Mts has resulted in a probable etchplain which reproduces the low relief of the original surface (also compare T. Czudek 1971). In the downthrown Sudetic Foreland inselbergs are still rising above thick kaolinitic weathering mantles.

A. Jahn (1980) expressed the view that the major feature of the Palaeogene Proto-Sudetes was inversion of relief. Under the hot and humid climatic conditions both Cretaceous rocks and broad synclines and crystalline massifs which resisted chemical decomposition formed synclinal table mountains and broad ranges. On the flanks of synclines that consisted of deeply weathered Palaeozoic rocks (largely schists) the wide intra-Sudetic depressions developed already at that stage.

In the upland belt the upper time limit of formation of the "Palaeogene planation surface" is the Badenian phase of faulting. In the Silesian-Cracow Upland the former extensive planation surface of a pre-Badenian foundation cut across the folded Palaeozoic sandstones and shales and the tilted Mesozoic sandstones, shales, marls, limestones and dolomites. On the fringe of the Upper Silesian Hercynian massif this surface developed by the intersection of older surfaces, Lower Triassic, Upper Triassic-Lower Jurassic and Lower Cretaceous in age (S. Gilewska in: M. Klimaszewski 1972).

In the Cracow Upland, which consists of Upper Jurassic limestones, the highest surface is expressed as a flat or gently rolling summit plane containing numerous buried karst hollows. The surface is locally studded with residual hills and tors of biothermal limestone (e.g., at Ogrodzieniec 504 m a.s.l.). The surface lies highest in the Myszków – Cracow anticline and slopes gradually northward from 460 to 390 m. The surface discussed is believed to represent a humid-tropical karst-plain above which mogotes of a Palaeogene foundation are rising (M. Klimaszewski 1958b; J. Polichtówna 1962, and others). The distribution of the residuals suggests that in the south these may be well post-Badenian, karst planation reaching the former level of the water-proof Badenian marine clays (S. Dżułyński et al. 1966).

In the neighbouring Miechów Upland the plateaus and ridges consist of nearly flat lying Cretaceous rocks (mostly marls). The planation surface of a pre-Badenian foundation here is preserved mostly on the isolated residual hills (414 m a.s.l.) showing traces of decalcified gaizes and *opoki* (J. Rutkowski 1965). It merges in an east- and south-east direction beneath the sheet of Miocene marine deposits (S. Gilewska 1958).

In the Nida Depression, a monotonous plain at 230–250 m a.s.l. originated throughout the Tertiary on the horizontal Cretaceous marls within the Miechów synclinorium.

The heart of the Kielce Upland lying within the Świętokrzyskie anticlinorium are the folded and uplifted Świętokrzyskie Mts. This Hercynian orogen which has been buried during the Mesozoic era was reexposed in pre-Badenian times (A. Radwański 1969) as low mountains of ridge-and-valley type. The existence of a former degradational surface is suggested by the radial arrangement of stream valleys which break the ridges in a series of gaps, and by traces of surfaces at 400 m and 360 m a.s.l. These surfaces of a Palaeogene foundation, overlooked by the highest hard-rock ridges (Łysica 612 m), truncated Palaeozoic quartzites, sandstones and calcareous rocks, and Mesozoic sandstones (S. Lencewicz 1934, and others). S. Z. Różycki (1967, 1972), however, believes that the oldest planation surface was founded in Miocene times, since early Tertiary residues are missing in the mountains.

The eastern foreland of the Świętokrzyskie Mts consists of weaker Cambrian and Palaeozoic rocks and tilted Mesozoic strata of varying resistance. The foreland ascends in three steps at 160 – 180 m, 200 – 240 m and 260 – 300 m a.s.l. towards the Świętokrzyskie Mts. Explanations of this vertical differentiation of the "Palaeogene planation surface" include a scissor-like arrangement of diverse planations (C. Radłowska 1963) and lithological control of the dynamics of planation (D. Kosmowska – Suffczyńska 1966). The other view is that the middle surface was founded at the furthest in the Miocene since it postdates the Oligocene inundation, whereas the foundation of the highest surface may be Oligocene – Miocene or earlier (S. Gilewska in: M. Klimaszewski 1972).

The Lublin Proto-Upland, made up mostly of calcareous Upper Cretaceous rocks, was a rolling land with a relief of 50 m. The unevenness of the sub-Oligocene surface was due to karst processes and denudation (A. Jahn 1956; M. Harasimiuk 1975, and others).

THE LATE EOCENE AND OLIGOCENE

The development of the "Palaeogene planation surface" was interrupted by a great marine transgression which extended during the Late Eocene across much of nothern Poland, except the Kuyavy-Pomeranian elevation, and lasted till the Upper Oligocene (Annexe A) (E. Odrzywolska – Bieńkowa et al. 1979).

Postdating the Upper Eocene were weak *en bloc* movements accompanied by erosion so that the Oligocene marine deposits rest on a highly varied substratum (K. Pożaryska and E. Odrzywolska – Bieńkowa 1982). These movements resulted in periodical changes of both extent and depth of the marine and inland basins, local recession of the sea, changes of facies of the Oligocene sediments (J. Łyczewska 1958, and others) and in the opening of a broad gate towards the west, along the "central European furrow" (S. Z. Różycki 1967, 1972). At the later stage of recession of the Oligocene sea brackish deposits have been laid down and then swampy and lacustrine sediments. The latter attain the greatest thickness in the karst depression at Rogoźno (H. Klatkowa 1972), in the grabens of the Poznań – Oleśnica dislocation system, and in the grabens between Chruścina – Nowa Wieś and Chobienia – Rawicz which revived already towards the end of the Eocene (Z. Deczkowski and I. Gajewska 1980; E. Ciuk 1977; A. Dąbrowski 1980).

The meta-Carpathian elevation escaped submergence, with exception of the Lublin Proto-Upland. Thus, the degradational period prevailed throughout the Palaeogene, resulting in the "Palaeogene planation surface". In the south there extended the Tethys Ocean, the northern limit of which is not definitely known.

THE MIOCENE

During the Miocene the principal morphostructures of Poland came into being. Tectonic movements in the Alpine-Carpathian geosyncline, and finally the far reaching displacement of the Carpathian nappes led to the fracturing of Poland's territory. These movements also brought about the development of both geobotanical provinces and altitudinal belts of climate and vegetation as well as changes in hydrological and soil conditions, the nature of weathering and the course of morphogenetic processes.

CLIMATE AND VEGETATION

The best evidence of climatic changes during the Miocene is forthcoming from the rich floras that have been identified from the brown coal seams. A cool episode at about 24.4–21 million years ago is indicated by the low collagen values of a fossil vertebrate fauna found in a karst pit at Przeworno I in the Sudetic Foreland (J. Głazek et al. 1977). The climate of the Karpatian (Table 1) was warm-temperate to subtropical and uniform over the whole area of Poland which received abundant rainfall. The vegetation consisted of typical Miocene species (e.g., *Carya*, *Castanea-Castanopsis*, *Celtis*, *Engelhardtia*, *Myrica*, *Platycarya)*. The flat and poorly drained areas were occupied by cypress swamps having many features in common with those in Florida and on the Gulf Coast. The major components of the swamps were *Taxodium-Cupressaceae* and *Nyssa* (J. Raniecka – Bobrowska 1959, and others). Beaver and tortoise were living there (J. Głazek et al. 1971, 1977).

The type of climatic conditions that prevailed during the Badenian is controversial. Palaeobotanical record seems to show a warm-temperate climate somewhat warmer and moister than the present one (M. Łańcucka – Srodoniowa 1966; L. Stuchlik 1980). W. Szafer (1954) and J. P. Bakker (1965), however, suggested that in the Carpathians a warm and humid, almost subtropical climate prevailed till the end of the Pliocene. Marine fauna of Indo-Pacific type also indicates either tropical or subropical conditions (U. Radwańska and A. Radwański 1984, and others). In drier habitats cervine, cats and small forest pigs were abundant (J. Głazek et al. 1971).

The uplift of both the Carpathians and Sudetes Mts resulted in the gradual development of three geobotanical provinces. In the south there extended the mountainous province (the Carpathians), on the northern shore of the Paratethys Sea – the lowland province, in the west – the western European province reaching as far as the tectonic Moravian Gate (L. Stuchlik 1980).

The Lower Badenian flora was mostly of warm-demanding type (e.g., *Lauraceae* and *Mastixioideae*), which became extinct during the Upper Badenian. The Carpathians were clothed with forests dominated by conifers. At the Upper Badenian stage there increased the share of *Cyperaceae* and *Graminae*, and of Arctic-Tertiary trees and shrubs which shed their leaves. On the northern shore of the Paratethys Sea *Angiospermae* dominated over *Gymnospermae*.

The climate of the Sarmatian was warm-temperate although drier than that of the earlier periods. Common trees of the deciduous forest were *Acer*, *Alnus*, *Betula*, *Carya*, *Castanea*, *Fagus*, *Liquidambar*, *Parrotia*, *Populus*, *Ulmus* and *Zelkova*. *Angiospermae* dominated over *Gymnospermae*, whereas *Lauraceae* and *Symplocaceae* were missing (L. Stuchlik 1980).

During the Miocene the extensive lowland had swamps and bogs. The vegetation of the drier habitats was composed of *Cyrillaceae*, *Ilex*, *Ericaceae*, *Leguminosae*, *Rhus*, *Araliaceae* and *Cornaceae*. Tortoises (*Emydinae*) lived in the freshwater lakes (M. Młynarski 1955).

The successive changes of both climate and vegetation affected the nature of weathering and the course of morphogenetic processes.

WEATHERING AND MORPHOGENETIC PROCESSES

The gradual deterioration of climate was associated with the destruction of the close subtropical forest cover and the subsequent wash and removal of the clayey waste which accumulated in the foreland of the rising Tatra Mts (M. Klimaszewski 1987) and the

Sudetes Mts (A. Jahn 1984). In the upland belt the "Palaeogene planation surface" consisting of soluble calcareous rocks was fashioned by karst processes. These led to the formation of karren, sink holes, uvalas, poljes and caves being buried beneath different regolith, clays and sand now (S. Biernat 1960, 1968; J. Głazek et al. 1971, 1977; S. Z. Różycki 1972). On the northern shore of the Paratethys Sea a karst relief of subtropical type continued to develop.

Periodical changes of the groundwater level due to changes in precipitation amount and local subsidence rates are documented by brown coal and lignite accumulations in both karst and tectonic depressions occurring in the Polish Lowlands, Silesian Upland, Świętokrzyskie Mts and in the eastern foreland of the latter (J. Czarnocki 1932; J. Samsonowicz 1934; T. Uberna 1962; S. Biernat 1960, and others).

Towards the end of the Badenian and in Lower Sarmatian time silicification of gypsum, sulphur-bearing limestone and sandstone took place in southern Poland (J. Gołąb 1931, and others).

THE EROSIONAL AND TECTONIC DISMEMBERMENT OF THE "PALAEOGENE PLANATION SURFACE"

Towards the end of the Oligocene the meta-Carpathian elevation became disturbed by tectonic movements. Tectonic deformations are evident in the western part of the elevation, where faults can be observed displacing the "Palaeogene planation surface" in three major phases (K. Birkenmajer et al. 1977).

During the first phase, $28+3$ million years ago, the Sudetic boundary fault (NW – SE) divided up the Proto-Sudetes into the Sudetes Mts and the downthrown Sudetic Foreland (Annexe B). This fault revived in later periods. Somewhat younger ($26.5+3$ and $27+3$ million years ago) is the Odra fault zone (NW – SE in the west, NWN – ESE in the east). With the deep fracture zones basaltic lava outpourings (e.g., Mt St Anna) were associated. Locally, these outpourings brought to an end the development of the sub-basaltic "Palaeogene planation surface". On the lava sheets there developed younger (post-basaltic) surfaces (L. Pernarowski 1963).

At the Oligocene /Miocene transition the "Palaeogene planation surface" became eroded in the rising Isera Mts (and, probably, in the Karkonosze Mts) as indicated by sand and gravel intercalations in the brown coal seams between Bolesławiec and Węgliniec (J. Milewicz and A. Grocholski 1960; K. Chmura and S. Lewowicki 1962; J. Oberc and S. Dyjor 1969). In Oligocene and Lower Miocene times long lasting denudation resulted in the formation of a surface at 600 – 800 m a.s.l. which corresponds to the second relief horizon (A. Jahn 1953, 1980; W. Walczak in: M. Klimaszewski 1972).

During the second phase, $2.5+3$ million years ago, being analogous to the Savian orogenic phase in the Carpathians, melabasanite poured out from the SW – NE trending fractures at Gracze near Niemodlin. $21+3$ million years ago either the crater or caldera of Mt St Anna became filled with lava (K. Birkenmajer 1974; K. Birkenmajer et al. 1977).

During the third phase, $17+2$ to 15 ± 2.5 million years ago, tectonic movements corresponding to the Styrian orogenic phase in the Carpathians caused further division of both the Sudetes Mts and Sudetic Foreland into smaller blocks and subsidence basins. The formation of the WNW – ESE fracture zones was accompanied by basalt outpourings between Legnica and Chojnów, Jawor and Złotoryja, and around Lubań, Zgorzelec and Zawidów (K. Birkenmajer et al. 1977).

It is to the Middle Miocene phase of tectonic movements that we must attribute the final isolation of the easternmost member of Hercynian Europe (S. Dyjor 1983), this extends over southern Poland as the Polish Uplands, except the Lublin Upland being of a transitional nature.

In late Oligocene and Miocene times the negative movements along the front of the Carpathian overthrust were compensated by positive movements of the meta-Carpathian elevation. Height differences might have been responsible for the incision of deep valleys into the Mesozoic-Palaeogene peneplain (A. Michalski 1884; A. Stahl 1933; W. Krach 1947, and others). The buried V-shaped valleys that survived within the Carpathian foredeep (Annexe B) are generally running southwards beneath the present Carpathians. Some valleys are joining broad erosional-denudational depressions which are disturbed tectonically in the west. These depressions are separated by broad ridges trending eastwards (Z. Buła and D. Jura 1983; W. Bogacz et al. 1984; S. Jucha 1985a, b). The upper time limit of the buried valleys is the drowning of the erosional-tectonic relief. In the south-west — such as in the Nizky Jesenik in Czechoslovakia (T. Czudek 1971) — the Paratethys Sea has penetrated to the Carpathian foredeep either in Karpatian or in Ottnangian times (A. Ślączka 1977; Z. Buła and D. Jura 1983) producing islands which disappeared during the Badenian (S. Alexandrowicz 1963; S. Połtowicz 1974; W. Bogacz et al. 1984, and others).

Orogenic movements of the Carpathians also led to the formation of numerous grabens and faults along which the meta-Carpathian elevation descended in a series of steps towards the Carpathian foredeep in the south (J. Czarnocki 1929; S. Dżułyński 1953; S. Alexandrowicz 1958, and others) and the central depression in the north (S. Z. Różycki 1937; Z. Deczkowski 1963; S. Biernat 1968, and others). The renewal of uplift introduced a marked erosion phase in the Nida Basin, the Świętokrzyskie Mts, the Lublin Proto-Upland and in the Carpathians, thus increasing sandy sedimentation in the Paratethys Sea. In Upper Badenian and Lower Sarmatian times extensive deltas originated there (W. Krach 1962; S. Połtowicz and A. Starczewska—Popow 1973, and others). The Miocene marine transgression was brought to an end by uplift and emergence of the Carpathian foreland during the Upper Sarmatian (M. Książkiewicz 1972) leaving a flat depositional surface which was drained in a south-eastward direction.

Subsequent erosion and degradation of the uplifted Świętokrzyskie Mts produced a distinct planation surface of pediment type at 290—300 m a.s.l. above which steep-sided ridges are rising. This surface passes eastward into the levelled surface of the mountain foreland (M. Klimaszewski 1958a, and others).

In Miocene times, cuestas formed on outcrops of the tilted Middle- and Upper Triassic, Middle- and Upper Jurassic, Cretaceous and Badenian rocks (S. Gilewska in: M. Klimaszewski 1972).

The structure of the Carpathians is due to several events. At the Oligocene/Miocene transition (Savian phase) the Magura nappe, together with the folded Dukla and Silesian units came into being. The occurrence of both redeposited fossils and flysch pebbles in various strata shows that the folding flysch ranges were already eroded (M. Książkiewicz 1931, 1972; and others). In Ottnangian time (older Styrian phase), the major folding of the flysch Carpathians took place and the Carpathian foredeep began to form. During the Badenian and Lower Sarmatian (and even in later times) the folded flysch masses were thrown over the northern Carpathian foreland. Displacements were on a wide scale, probably of the order of 60—80 km (S. Połtowicz 1974; R. Ney et al. 1974). Both foredeep fills and structures composing the Carpathians of Poland become younger eastwards (M. Książkiewicz 1972; A. Połtowicz 1974, 1978; J. Kotlarczyk 1986).

The Carpathians were also affected by warping. Already in Badenian time positive movements of the Tatra Mts and Beskidy Mts were compensated by subsidence of the Orawa — Nowy Targ and Nowy Sącz Basins, and of the Jasło — Krosno Depression (M. Klimaszewski 1965, 1987). During the Sarmatian uplift affected the Carpathians, together with the Carpathian foredeep and the upland belt (M. Książkiewicz 1972; S. Połtowicz and A. Starczewska—Popow 1973). Phases of both uplift and intense erosion alternated with phases of stillstand to which the formation of planation surfaces in the Tatra Mts and in the flysch Carpathians is usually related. In Badenian and Lower Sarmatian times the intramontane planation surface came into being. This surface is better developed in the foot-hills, in the north, than in the Beskidy Mts and in the Tatra (L. Sawicki 1909; M. Klimaszewski 1958a, 1965, 1987; L. Starkel in: M. Klimaszewski 1972).

During the Miocene, along a line from Zielona Góra — Poznań — Warsaw to Brześć there existed a great inland sedimentary basin to which the term central depression has been applied by S. Z. Różycki (1967, 1972). It included two overdeepened basins (the Wielkopolska and Mazowsze basins) being separated by the Kuyavy — Pomeranian elevation. At this time the Baltic depression did not exist (S. Z. Różycki 1967, 1972). It is likely that the central depression was not closed as may be evidenced by remnants of a marine fauna in the Miocene brown coal formation at Dobrzyń on Vistula River (M. Brykczyński 1985; E. Odrzywolska—Bieńkowa et al. 1979).

Both revival of ancient faults and formation of new ones in the present lowlands is attributed either to the Pyreneean or to the Savian phase of tectonic movements. The extensive dislocation systems of Poznań — Kalisz and Poznań — Oleśnica, together with the grabens at Sulmierzyce, Sieroszewice, Kępno, Złoczew and Kleszczów near Bełchatów were then reconstructed (Annexe C). Similar processes occurred in the Sudetic area (S. Biernat 1968; E. Ciuk 1977; Z. Deczkowski and I. Gajewska 1980; J. R. Kasiński 1984). Towards the end of the Middle Miocene the lowlands were occupied by extensive swamps and bogs with small waterbodies which took in detrital materials from Scandinavia, the Sudetes Mts and the uplands. Remains of the swampy, aquatic and terrestrial plants gave rise to brown coal seams (S. Z. Różycki 1967, 1972).

The brown coal sedimentary basins were shallow in the north-east, within the rigid pre-Cambrian platform, and both great and very deep (exceeding 200 m) in the west, within the mobile west-central European platform affected by intense epeirogenetic movements (J. R. Kasiński 1984). The occurrence of the older brown coal seams only in the west indicates that the East-European Platform rim subsided in Pliocene times (S. Z. Różycki 1972).

Multiphase post-Laramide tectonic movements are reflected in the cyclic sedimentation of the brown coal formation in the Sudetic area, for instance, in the Żytawa depression. The occurrence of coarse deposits at the base of each sedimentary series of the Miocene — Pliocene cycle indicates uplift and intense erosion of the basin margin, whereas the accumulation of organic matter suggests a rather slow downward movement of the basin bottom. The Pliocene series is developed as the alluvial fan facies indicating further erosion of the Sudetes Mts (J. R. Kasiński 1981, 1983, 1984).

The development of a Tertiary graben in the central depression is illustrated by example of the Kleszczów graben near Bełchatów. The graben fill is tripartite. The lowermost series includes clay and silt with thin brown coal seams. These deposits accumulated in a waterbody which occupied the quickly subsiding graben bottom. Rockfalls occurred at this time on the steep backing limestone scarps. At the Middle Miocene stage slow and uniform subsidence took place, whereby the lake changed into a bog. Peat that accumulated during at least 300000 years gave rise to a brown coal seam attaining a mean thickness of 60 m. It includes unique Tertiary inland deposits, namely lacustrine limestone, gyttja, silt and sand with a mollusc fauna. Peat growth terminated in renewed quick subsidence of the graben bottom. It was followed by deposition of the Upper Miocene clay, silt and sand in a probable lake. In the surroundings of the graben — polje the purity of the limestone was favourable for the formation of caves (S. Biernat 1968, 1971; M. D. Baraniecka and Z. Sarnacka 1971; and others).

Concealed beneath the Tertiary and Quaternary deposits, valleys that follow a number of joints and faults extended towards the central depression in the north. These valleys attained a depth of 40 – 50 m, for instance, the Gosławice – Pątnów valley with a stepped floor. It developed in four cycles alternating with phases of increased lateral erosion. Both the earlier cycles terminated in sedimentation of sand brought in from the south. The third cycle began with the formation of further terraces which subsequently were buried beneath the gravel, sand and Middle Miocene brown coal. Towards the end of the Miocene, the shallow and broad valleys became filled with horizontally bedded clay and silt, with thin brown coal seams and plant remains which correlated with the Poznań series. Its rhytmical stratification reflects cyclic lower-order changes of climate (J. Czarnik 1972).

A brief review of events that took place in the Żytawa depression, the Kleszczów graben and in the Gosławice – Pątnów valley reveals the synsedimentary formation of the tectonic depressions. The accumulation of deposits belonging to three basic sedimentary cycles was due to subsidence proceeding at different rates. Superimposed were the climatically morphogenetic processes (wash, fluvial erosion and accumulation, lacustrine accumulation).

It is likely that towards the end of the Miocene the whole area of Poland was a land-mass (Annexe C). In the south, the rising Carpathians were separated by an escarpment formed of both folded flysch and Miocene rocks from a plain of depositional origin which occupied the former Carpathian foredeep. This submontane plain was drained both south-eastwards into the Proto-Dniester and north-westwards into the central depression. The watershed was at that stage in the Cracow Gate.

In the west, uplift of the Sudetic block mountains has taken place along deep fracture zones. Uplift was followed by intense denudation. Widespread deposition of the clayey waste, and less frequently of sand and gravels derived from the mountains, occurred in the downthrown Sudetic Foreland. Inselbergs were rising above the poorly drained plains. Both the western Sudetes Mts and Sudetic Foreland were remarkable in having at this time a large number of points of active volcanic eruption.

The undulating meta-Carpathian elevation was a denuded "peninsula" being surrounded by extensive depositional plains. Its narrow sole had little relief. The remaining part was marked by the occurrence of diverse relief types including low mountains, table- and scarplands as well as low plateaus with residual hills and tors. Into the plateaus broad valleys were incised.

Northern Poland was occupied by extensive swampy plains, thus accomodating further sedimentation of clayey waste products derived from Scandinavia, the Sudetes Mts and the upland belt.

THE PLIOCENE

In the Pliocene, the progressive cooling and continentalization of the climate took place. It is possible that the whole area of Poland was then a land-mass. During the 12 million years that followed from the Badenian till the end of the Pliocene processes of degradation favoured the formation of stepwise arranged epicyclic planation surfaces on the weak rocks in southern Poland. Destruction and planation affected largely areas that have been raised in Miocene times, especially the Carpathians, the Sudetes Mts and the Polish Uplands.

CLIMATE AND VEGETATION

The vegetation of the Podhale in the Carpathians (W. Szafer 1946 – 1947, 1954; J. Oszast 1970, 1973; J. Oszast and L. Stuchlik 1977; L. Stuchlik 1980), of the Sudetic Foreland and Sudetes Mts (A. Stachurska et al. 1973; A. Sadowska 1985; A. Sadowska in: A. Jahn 1984; M. Łańcucka – Środoniowa in: A. Jahn 1984) and of the central depression (J. Doktorowicz—Hrebnicka 1957; S. Z. Różycki 1972) indicates the gradual cooling and dessiccation of the Pliocene climates.

During the Mio-Pliocene (Pannonian) oscillations of climate took place. Numerous Miocene plants became extinct, and cypress swamps were less important then.

In Lower Pliocene time (Pontian), the climate was temperate. Quaternary plants dominated over the Tertiary ones (about 30%). Spruce and pine were common in the Carpathians. In the Sudetic Foreland the drier habitats bore deciduous forests with numerous climbing plants and an undergrowth being dense shrubs. Herbaceous plants occurred less frequently there. The poorly drained areas and the periodically flooded lowlands were still occupied by cypress swamps which suggest a warm-temperate climate.

The Middle-Pliocene (Dacian) flora of the Carpathians indicates a distinct deterioration of the climate named the intra-Pliocene cooling by W. Szafer. The extinction of Tertiary elements was associated with an increase in herbaceous plants (40%). Spruce and pine with *Tsuga* and *Sciadopitys* were the dominant trees there. In the Sudetes Mts the role of deciduous trees increased, indicating a warm-temperate to temperate climate. The lowlands had *Cyperaceae* swamps with alder and numerous species of herbaceous (mostly swampy and aquatic) plants.

In Upper Pliocene time (Romanian) the more warm-demanding elements reappeared in the Carpathians. The vegetation consisted of open woodland (with Crataegus, Cornus, Prunus, Rosa and Vitis) resembling the present Pontian scrubs. At higher altitudes there thrived coniferous forests. The climate was warm-temperate to temperate and fairly dry. Mixed forests were widespread in the Kłodzko Basin (the Sudetes Mts). Here the climate was temperate, although warmer than the present one. Quaternary elements dominated over the Tertiary ones, and the role of herbaceous plants was important. The lowlands were then clothed with plants being typical of waterlogged areas. These were occasionally visited by great terrestrial mammals, for instance, *Mastodons* (S. Z. Różycki 1972).

The probable open woodland with small waterbodies was the region of *Amphibia*, *Reptilia* and rodents, and of predatory, hoofed and insectivorous animals. In addition to forest and freshwater faunas a steppe fauna has been found in the upland belt. The latter fauna suggests the occurrence of other climatically controlled plant associations (J. Samsonowicz 1936; J. Stach 1953; M. Młynarski 1955; K. Kowalski 1956; Z. Mossoczy 1959a; R. Gradziński and Z. Wójcik 1966, and others).

WEATHERING AND MORPHOGENETIC PROCESSES

Climatic changes reflected in the gradual change of close forest into open woodland and then into steppe have influenced the type of morphogenetic processes. In the moister phases of the Pliocene both karst and fluvial processes have been revived. Evidence to that are cave and cleft fills which were shown by palaeontological dating to be Middle and Upper Pliocene in age (J. Samsonowicz 1936; and others). Locally, karst shafts enlarged by seasonal torrents, dolines and caves came into being (E. Mycielska 1960, 1965; E. Jońca 1963; M. Pulina 1964; M. Harasimiuk 1975, and others). Changes of both colour and composition of the Pliocene weathered materials are regarded as induced by the climatically controlled changes of the weathering type (Z. Mossoczy 1959b, and others).

On the other hand, the Lower and Upper Pliocene climate, which is believed to have been less moist, did not favour the degradation of areas consisting of hard limestone, for instance, in the eastern foreland of the Świętokrzyskie Mts (D. Kosmowska—Suffczyńska 1966, and others).

THE DEVELOPMENT OF RELIEF

Under warm-temperate and temperate climatic conditions having alternating moister and drier phases, intense denudation took place. The clayey waste was removed by wash to be deposited by seasonal streams in the intramontane Orawa — Nowy Targ Basin, in the northern foreland of the uplands and on the extensive *Cyperaceae* swamps which occupied the central depression. Data for the sub-Carpathian basins are missing.

In the flysch Carpathians four surfaces came into being during the Upper Tertiary. There is a rule that the higher the surface, the harder is the rock on which it is preserved, and the lower the step, the smaller is its extent and softer the bedrock, with few exceptions (L. Starkel 1965). Fragments of flattenings, the levelled mountain ridges and hilly watersheds form the basis for relief reconstruction. The surfaces become younger towards the east.

The highest surface named the Beskidy level (*poziom beskidzki*) (L. Sawicki 1909) survived mostly in the Western Carpathians. The marked height differences (200—500 m) suggest that this was not a planation surface. The Beskidy level became disturbed by later *en bloc* movements (L. Starkel in: M. Klimaszewski 1972).

The second surface named the intramontane level (*poziom śródgórski*) is to be found in the whole Carpathians. It is developed as flattenings in the high foot-hills, as benches on the fringe of the high ranges and as levelled ridges. This surface truncated resistant and moderately resistant rocks. It rises from between 450 and 500 m in the northern foot-hills to 700—1000 m a.s.l. in the Beskidy axis.

The foot-hill level (*poziom pogórski*) is formed of broad levelled ridges at 150—200 m above the valley floors. It rises from between 360 and 420 m a.s.l. in the *Pogórze* (foot-hills) to 600—800 m a.s.l. on the divide. The foot-hill level had the aspect of pediments, overlooked by residual and hard-rock ridges. This level cut across rocks being less resistant than those forming the intramontane level.

The lowest surface or river-side level (*poziom dolinny*) being best developed on soft rocks is of glacis type. It increases in height from between 40 and 50 m in the west to 80—100 m above the valley floors in the east. The step-wise arranged "partial planation surfaces" extend along the valleys into the interior of the mountains (L. Sawicki 1909; J. Smoleński 1937; M. Klimaszewski 1934, 1965; L. Starkel in: M. Klimaszewski 1972). Since the early Pliocene the rivers have flowed from the Carpathians to the developing sub-Carpathian basins (M. Klimaszewski 1934, 1965, and others).

In the flysch Carpathians the age of the surfaces is far from clear (A. Tokarski 1978; W. Zuchiewicz 1984). It seems, however, that the mature relief of both the Tatra Mts and the Beskidy Mts (at 800—1200 m a.s.l.) can be traced back to the Miocene (M. Baumgart—Kotarba 1983; M. Klimaszewski 1965, 1987, and others). In the Tatra Mts and the northern Tatra foreland the "partial planation surfaces" can be correlated by fossil-bearing sediments that were laid down in the subsiding Orawa — Nowy Targ Basin (L. Watycha 1976; K. Birkenmajer 1978). Although no Neogene Tatric gravels have been disclosed there, analyses of both chemical and mineralogical composition of the Neogene clays required from a drilling at Czarny Dunajec (in the Orawa Basin) revealed that they do contain fine Tatra derived materials. According to M. Klimaszewski (1987), the intramontane planation surface is Lower Sarmatian in age. The foot-hill level is Pontian. The riverside level is considered to be Romanian. However, the thrust that has overridden the Lower Sarmatian deposits in the east, where the Beskidy level does not occur, indicates that the three successively younger levels extending along the valleys are post-Sarmatian. Thus, L. Starkel (1969), K. Birkenmajer (1978) and A. Henkiel (1969) date the surfaces in question as Lower Pliocene (Pontian), Upper Pliocene and Lower Quaternary.

In the Tatra Mts, which suffered uplift and very intense degradation, either six (M. Baumgart—Kotarba 1983), or five (M. Klimaszewski 1987) "partial planation surfaces" have developed in between the Badenian and Pleistocene. These are the summit-, montane-, intramontane-, foot-hill- and river-side levels. The High Tatra Mts were higher uplifted than the Western Tatra Mts. The levels are sloping northward, and they are separated by polygenetic valleys. The fourth level passes from the Tatra Mts into the Podhale indicating the common evolution of both units (M. Klimaszewski 1965).

Phases of both uplift and erosion of the margin of the tectonic Orawa — Nowy Targ Basin were associated with the supply of coarse flysch material into the depression (L. Watycha 1976). The Upper Pliocene phase of increased erosion is documented by the Domański Wierch alluvial fan deposits (J. Oszast 1970).

In the erosional-denudational sub-Carpathian basins which developed within the former Carpathian foredeep the soft Miocene deposits were partially removed by the Carpathian, Sudetic and upland rives. The Oświęcim and Racibórz Basins in the west were drained by the Odra River and its tributaries flowing from the Beskid Śląski Mts, the Sudetes Mts and the Silesian Upland. This is indicated by series of either Middle- or Lower Pliocene white quartz gravels that have been laid down by the Nysa Kłodzka River in the Sudetic foreland, and by Carpathian and upland gravels found in the Racibórz Basin (J. Szaflarski 1955; M. Klimaszewski 1958a; K. Klimek in: M. Klimaszewski 1972; S. Dyjor 1983; A. Jahn 1984a, and others). The watershed between waters flowing to the Odra and Vistula Rivers was in the Cracow Gate (K. Klimek in: M. Klimaszewski 1972). Pliocene erosion has partially exposed the upheaved horsts from beneath the thin Miocene cap rock in the southern part of the Silesian-Cracow Upland (M. Klimaszewski 1958a, and others). The successive stages of exhumation associated with epigenesis were controlled by base-level changes in the sub-Carpathian basins (S. Dżułyński et al. 1966, and others).

In the Sandomierz basin the foot-hill level may by traced continuously over the former Carpathian foredeep to both the Sandomierz (Opatów) Upland and the Lublin Upland (M. Klimaszewski 1958a). Selective erosion and denudation led to the formation of a distinct escarpment of the Carpathian Foot-hills. The drainage direction of the denudational Sandomierz Basin has given rise to some controversy. J. Samsonowicz (1934), M. Klimaszewski (1958a) and S. Z. Różycki (1972) postulated that the Proto-Vistula River passed in Pliocene times in a water gap through the Polish Uplands into the central depression in the north. W. Laskowska—Wysoczańska (1975), however, believes that the Vistula was flowing in a south-eastward direction to join the Proto-Dniester River (Annexe D). Subsequent downwarping of the northern part of the Sandomierz Basin in either the Middle Pliocene (W. Laskowska—Wysoczańska 1975), late Pliocene or early Quaternary times (S. Dżułyński et al. 1968, and others) caused sedimentation of the flysch-derived Witów gravel series. Later irregular uplift of the Sandomierz Basin brought about drainage divertion in preglacial *sensu stricto* time (W. Laskowska—Wysoczańska 1971, and others).

During the Pliocene, the relief of the Sudetes Mts was similar to the present one. Along the margin of both the Western and Central Sudetes Mts the broad and levelled foot-hills extended, which correspond to A. Jahn's third relief horizon (1953). This surface also penetrated to the interior of the intramontane depressions. The warm--temperate climatic conditions that prevailed during the Lower or Middle Pliocene favoured the chemical decomposition of feldspar. In the Kłodzko Basin there accumulated quartz and porphyry gravels with sand and kaoline clay derived from the mountains at this period. In Upper Pliocene time, sand, silt and clay containing a rich flora were laid down there. The position of the fossil-bearing clay indicates that the basin floor was towards the end of the Pliocene accordant with the present one (A. Jahn et al. 1984). The Sudetic "white gravels" of the Gozdnica series (S. Dyjor 1966, 1986) also occur in the mountain foreland. Furthermore, quartz fragments, sand and gravels resembling the "white gravels" are recognizable in a kaoline matrix at 50—100 m above the Quaternary valley floors near Jelenia Góra. Probable Pliocene cave fills at Wojcieszów in the Kaczawa Mts were left hanging on the Kaczawa valley-side (M. Pulina 1964). It appears that in contrast to the Eastern Sudetes, the Western Sudetes have suffered uplift towards the end of the Pliocene (A. Jahn 1984; A. Jahn et al. 1984).

In the upland belt Lower Pliocene planation affected areas largely consisting of soft clayey and calcareous rocks (*opoki*, marl). Remnants of this planation surface survived in the denudational basins of the Silesian Upland (J. Lewiński 1914, and others), in the Miechów Upland (M. Tyczyńska 1959; H. Ruszczyńska 1961, and others), the Nida Basin (J. Flis 1954, and others) and in the Sandomierz Upland (J. Samsonowicz 1934; M. Klimaszewski 1958a; D. Kosmowska – Suffczyńska 1966).

TABLE 1. Chronostratigraphy of the Neogene in Southern Poland

Chronostratigraphy of the Neogene in southern Poland (L. Stuchlik 1980)				Orogenic phase and absolute age (million yrs) (K. Birkenmajer 1978)	
QUATERNARY		Eopleistocene		Wallachian phase	- 2
PLIOCENE		Romanian (P_3)		Rhodanian phase	- 3.5
PLIOCENE		Dacian (P_2)			
PLIOCENE		Pontian (P_1)	Upper	Attic phase	- 5
			Lower		
MIO-PLIOCENE	UPPER	Pannonian (MP)	Upper	Moldavian phase	- 7.95
			Lower		- 9.5
MIO-PLIOCENE	UPPER	Sarmatian (M_5)	Upper		- 11.2
			Lower		- 14
MIOCENE	MIDDLE	Badenian	Kosovian		
			Wielician	Styrian phase	- 15.1
			Moravian		- 16 – 17
MIOCENE	LOWER	Karpatian (M_3)			- 20.5
MIOCENE	LOWER	Ottnangian (M_2)			- 21 – 22
MIOCENE	LOWER	Eggenburgian (M_1)		Savian phase	- 23.5 – 24
OLIGO-MIOCENE		Egerian (OM)			30

In the northern part of the Silesian-Cracow Upland the gradual development of the structural scarps resulted in drainage divertions (K. Klimek 1967). In the flat-floored strike valleys the subsequent rivers were flowing north-westwards into the Pliocene sedimentary basin (T. Krzemiński 1965, and others), whereas the southern part of the Silesian Upland was drained south-westward into the Racibórz Basin (J. Szaflarski 1955, and others).

In the Świętokrzyskie Mts further development of the ridge-and-valley topography took place. In the Lublin Upland dissection of the Lower Sarmatian depositional plain (A) in Upper Sarmatian and Lower Pliocene times was accompanied by pedimentation. This produced two planation surfaces (B and C) above which residuals of the oldest relief are rising (A. Jahn 1956; H. Maruszczak in: M. Klimaszewski 1972; M. Harasimiuk 1975). Pedimentation was interrupted by wetter climatic conditions that prevailed at the Lower/Middle Pliocene transition. In the Upper Pliocene increased aridity was associated with the formation of a flattening (D) only within the valleys. This surface corresponds to the river-side level in the Carpathians (M. Harasimiuk 1975).

Throughout the Pliocene the central depression was either an extensive periodically flooded plain (S. Z. Różycki 1972) or a residual lake of marine origin (E. Odrzywolska – Bieńkowa et al. 1979, and others). The lowland area accomodated sedimentation of the clayey Poznań series. These deposits were derived from the waste-covered uplands in the south. In the Mazowsze depression there accumulated the variegated clays, silt and sand attaining a thickness of 160 m (M. D. Baraniecka 1985). In both the Wielkopolska Depression and the Sudetic Foreland the sedimentation of the Poznań series existed well back into the Upper Badenian time, and terminated during the Pontian (A. Stachurska et al. 1973; S. Dyjor 1986). Here the variegated clay is underlain by green clay with glauconite and marine microfossils and by a grey clay being the lowermost member of the Poznań series. The green clay contains gypsum crystals (E. Odrzywolska – Bieńkowa et al. 1979) which bear witness to the rapid dessiccation of the shallow water bodies (S. Z. Różycki 1972).

THE EROSIONAL DISSECTION OF THE LOWER PLIOCENE PLANATION SURFACE

At the late Pliocene and early Quaternary stage the revival of fluvial erosion and subsequent valley incision took place as a result of the Wallachian phase of tectonic movements. Proceeding uplift of the Sudetes Mts, Sudetic Foreland and Polish Uplands was favourable for the dissection of both submontane and lowland plains. In the Polish Lowlands waters were flowing then to the North Sea and to the Arctic Seas. The Odra River, draining the western portion of the Polish Lowlands, the Sudetes Mts, Beskid Śląski Mts and Silesian Upland, flowed westward to the North Sea (B. Krygowski 1952; T. Bartkowski 1955; S. Dyjor 1985a, b, and others).

In north-eastern Poland rivers flowing to the Arctic Seas spread Eopleistocene deposits on the flat depositional plain or slightly truncated surface of the Pliocene clays (M. D. Baraniecka 1985, and others). The first fossil valley system included that of the Proto-Vistula River and its tributaries draining the vicinity of Radom, Łódź and Warsaw. The second system included the Proto-Bug valley extending across the Mazurian Lake District, Lithuania, Latvia and the Ladoga and Onega Lakes to the Arctic Seas (E. Rühle 1955; M. Harasimiuk and A. Henkiel 1981; S. Dyjor 1985a, and others). The above arrangement of the late Pliocene and early Quaternary valleys became disturbed by Pleistocene glaciation of most of the area of Poland.

CONCLUSIONS

A brief review of the geomorphological evolution of Poland's territory revealed that the general type of modification of the landforms inherited from the Tertiary was dependent upon the neotectonic tendencies. In the areas that showed long lasting tendencies towards subsidence the Tertiary features persisted beneath a thick covering of Tertiary marine, lacustrine and terrestrial sediments (the Polish Lowlands, the sub-Carpathian basins). In the less depressed areas there has been a general lovelling of the ancient relief by multiple deposition of allogenic materials upon which new levels developed. The resulting polycyclic surface includes relics of the older, now buried surfaces (e.g., the forelands of the Sudetes Mts and the Świętokrzyskie Mts). On the contrary, in the high standing areas that showed tendencies towards uplift there has been a general lowering of the original Tertiary land surface accompanied by the removal of either regolith mantle or thin cap-rock (e.g., the major Sudetic ranges, the major Silesian-Cracow Upland divide).

The state of both preservation and modification of the Tertiary landscapes is determined by the resistance of the predominant rock complexes. The harder the rocks, the better preserved are the surfaces of the oldest foundation. Most of the successively later surfaces truncated softer rocks (e.g., in the Carpathians). In the present landscape there persisted only fragments of the summit planes, levelled mountain and upland ridges and hilly watersheds. The more or less advanced modifications of the inherited Tertiary land surfaces are due to the changing climates and associated morphogenetic processes of the Pleistocene.

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REFERENCES

Alexandrowicz S. W., 1958, Transgresywne osady miocenu z kopalni Makoszowy i ich pozycja stratygraficzna, *Acta Geologica Polonica* 8, 1, 149—177.

Alexandrowicz S. W., 1963, Stratygrafia osadów mioceńskich w Zagłębiu Górnośląskim, *Prace Instytutu Geologicznego* 39, 5—126.

Areń B., 1957, *Atlas Geologiczny Polski, Zagadnienia stratygraficzno-facjalne. Trzeciorzęd*, 11, Instytut Geologiczny.

Bakker J. P., 1965, Some observations in connection with recent Dutch investigations about granite weathering and slope development in different climates and climate changes, *Zeitschrift fur Geomorphologie*, Supplementband 1: Morphologie des versants.

Baraniecka M. D., 1985, Profil z Ponurzycy na tle osadów plioceńskich i preglacjalnych południowego Mazowsza, in: *Plioceńska i plejstoceńska sieć rzeczna i związane z nią kompleksy osadów gruboklastycznych w Polsce*, PAN, Wrocław., 17—20.

Baraniecka M. D., Sarnacka Z., 1971, Stratygrafia czwartorzędu i paleogeografia dorzecza Widawki, *Biuletyn Instytutu Geologicznego* 254, 157—269.

Bartkowski T., 1955, O budowie głębszego podłoża Koła nad Wartą, *Przegląd Geologiczny* 6, 296—299.

Baumgart—Kotarba M., 1983, *Kształtowanie koryt i teras rzecznych w warunkach zróżnicowanych ruchów tektonicznych (na przykładzie wschodniego Podhala)*, Prace Geograficzne IGiPZ PAN 145.

Biernat S., 1960, Budowa geologiczna kredy opolskiej, *Biuletyn Instytutu Geologicznego* 152, 173—230.

Biernat S., 1968, Problemy tektoniki i morfologii stropu mezozoiku między Bełchatowem a Działoszynem, *Kwartalnik Geologiczny* 12, 2, 296—307.

Biernat S., 1971, Budowa geologiczna i ukształtowanie podłoża kenozoiku w rejonie Bełchatowa, *Biuletyn Instytutu Geologicznego* 254, 37—48.

Birkenmajer K., 1974, Trzeciorzędowe wulkany Graczy na Dolnym Śląsku i ich założenia uskokowe, *Rocznik Polskiego Towarzystwa Geologicznego* 44, 4, 357—575.

Birkenmajer K., 1978, Neogene to Early Pleistocene subsidence close to the Pieniny Klippen Belt, Polish Carpathiana, *Studia Geomorphologica Carpatho-Balcanica* 12, 17—28.

Birkenmajer K., Jeleńska M., Kądziołko—Hofmokl M., Kruczyk J., 1977, Age of deep-seated fracture zones in Lower Silesia (Poland), based on K — Ar and palaeomagnetic dating of Tertiary basalts, *Rocznik Polskiego Towarzystwa Geologicznego* 47, 4, 545—552.

Błaszak M., 1970, Charakterystyka naturalnych surowców dla mas formierskich w utworach krasowych okolic Częstochowy, *Biuletyn Instytutu Geologicznego* 240, 157—243.

Bogacz K., Dżułyński S., Harańczyk C., 1973, Caves filled with clastic dolomite and galena mineralization in disaggregated dolomites, *Rocznik Polskiego Towarzystwa Geologicznego* 43, 1, 59—72.

Bogacz W., Kotarba M., Krach J., 1984, Cechy strukturalne serii węglonośnej południowej części Rybnickiego Okręgu Węglowego, *Rocznik Polskiego Towarzystwa Geologicznego* 54, 3—4, 361—377.

Brykczyński M., 1985, Związki między późnotrzeciorzędową, eoplejstoceńską i współczesną siecią rzeczną Niżu Polskiego, in: *Plioceńska i eoplejstoceńska sieć rzeczna...*, Summaries, PAN, Wrocław, 20—21.

Buła Z., Jura D., 1983, Uwagi o rozwoju molasy rowu przedgórskiego Karpat na Śląsku Cieszyńskim *Przegląd Geologiczny* 12, 659—662.

Chmura K., Lewowicki S., 1962, Kwarcyty trzeciorzędowe okolic Bolesławca na Dolnym Śląsku, *Biuletyn Instytutu Geologicznego* 173, 5—56.

Ciuk E., 1975, Rozwój litologii i sedymentacji utworów trzeciorzędowych w rejonie Tanowa na północny zachód od Szczecina, *Biuletyn Instytutu Geologicznego* 284, 133—162.

Ciuk E., 1977, Nowe stanowiska eocenu górnego i oligocenu dolnego w rejonie Głobic, Szaszorowic i Miechowa (woj. leszczyńskie) w SW części monokliny przedsudeckiej, *Przegląd Geologiczny* 1, 1—5.

Czarnik J., 1972, Paleomorfologia okolic Turka w górnym trzeciorzędzie i plejstocenie, *Studia Geologica Polonica* 40, Wyd. Geol., Warszawa.

Czarnocki J., 1929, O tektonice okolic Łagowa oraz kilka słów w sprawie trzeciorzędu i złóż galeny na tym obszarze, *Posiedzenia Naukowe Państwowego Instytutu Geologicznego* 24, 38—36.

Czarnocki J., 1932, L'Helvétien et les lignites de cet étage dans les environs de Korytnica et de Chomentów, *Posiedzenia Naukowe Państwowego Instytutu Geologicznego* 32, 16—19.

Czudek T., 1971, Geomorfologie vychodni cásti Nízkého Jeseniku, *Rozpravy Československé Akademie Ved, Rada matematickych a přirodnich ved* 81, 7.

Dąbrowski A., 1980, System rowów trzeciorzędowych w obrazie grawimetrycznym, *Przegląd Geologiczny* 3, 169—172.

Deczkowski Z., 1963, Górny trias i jura okolic Wielunia, *Biuletyn Instytutu Geologicznego* 168, *Z badań złóż kruszców.* 4, Wyd. Geol., Warszawa, 87—123.

Deczkowski Z., Gajewska I., 1980, Mezozoiczne i trzeciorzędowe rowy obszaru monokliny przedsudeckiej, *Przegląd Geologiczny* 3, 151—156.

Doktorowicz—Hrebnicka J., 1957, Wzorcowe spektra pyłkowe plioceńskich osadów węglonośnych, *Prace Instytutu Geologicznego* 15, *Z badań flor trzeciorzędowych* 2, 87—165.

Dyjor S., 1966, Młodotrzeciorzędowa sieć rzeczna zachodniej części Dolnego Śląska, in: *Z geologii Ziem Zachodnich*, 2, Wrocław, 287—316.

Dyjor S., 1983, Ewolucja trzeciorzędowych przedgórskich rowów tektonicznych centralnych i wschodnich Sudetów, in: *Współczesne i neotektoniczne ruchy skorupy ziemskiej w Polsce*, 4, Ossolineum, Wrocław.

Dyjor S., 1985a, Plioceński i eoplejstoceński rozwój systemów kopalnych dolin w Polsce na tle ewolucji paleogeograficznej bruzdy środkowoeuropejskiej, in: *Plioceńska i eoplejstoceńska sieć rzeczna...*, Summaries, PAN, Wrocław.

Dyjor S., 1985b, System kopalnych dolin Polski zachodniej i fazy jego rozwoju w pliocenie i eoplejstocenie, in: *Plioceńska i eoplejstoceńska sieć rzeczna...*, Summaries, PAN, Wrocław.
Dyjor S., 1986, Zarys ewolucji tektonicznej w trzeciorzędzie polskiej części bruzdy środkowoeuropejskiej i jej ram południowych, in: *Historia ruchów tektonicznych na ziemiach polskich. Cykl alpejski*, Summaries, Wyd. Uniw. Wrocławskiego.
Dyjor S., in press, Systemy kopalnych dolin Polski zachodniej i fazy ich rozwoju w młodszym neogenie i eoplejstocenie.
Dżułyński S., 1953, Tektonika południowej części Wyżyny Krakowskiej, *Acta Geologica Polonica* 3, 3, 325—440.
Dżułyński S., Henkiel A. Klimek K., Pokorny J., 1966, Rozwój rzeźby dolinnej południowej części Wyżyny Krakowskiej, *Rocznik Polskiego Towarzystwa Geologicznego* 36, 4, 329—343.
Dżułyński S., Krysowska—Iwaszkiewicz M., Oszast J., Starkel L., 1968, O staroczwartorzędowych żwirach w kotlinie Sandomierskiej, *Studia Geomorphologica Carpatho-Balcanica* 2, 63—76.
Embleton C., Demek J., 1984, Hercynian Europe, Introduction, in: *Geormophology of Europe*, The Macmillan Press Ltd., London.
Flis J., 1954, *Kras gipsowy Niecki Nidziańskiej*, Prace geograficzne IG PAN 1.
Gilewska S., 1958, *Rozwój geomorfologiczny wschodniej części Wyżyny Miechowskiej*, Prace Geograficzne IG PAN 13.
Głazek J., Oberc J., Sulimski A., 1971, Miocene vertebrate faunas from Przeworno (Lower Silesia) and their geological setting, *Acta Geologica Polonica* 21, 3, 473—516.
Głazek J., Galewski K., Wysoczański—Minkowicz T., 1977, Nowe dane o krasie kopalnym w Przewornie, Prace Naukowe Uniwersytetu Śląskiego 185, *Kras i Speleologia* 1(10), 81—87.
Gołąb J., 1931, Zarys budowy geologicznej okolic Ostrzeszowa, *Rocznik Polskiego Towarzystwa Geologicznego* 7, 398—401.
Grabowska I., 1983, Palinoflora i plankton paleogeński z NW Polski, *Przegląd Geologiczny* 7, 420—423.
Grabowska I., Piwocki M., 1975, Wiek i geneza iłów toruńskich w okolicy Torunia na podstawie obserwacji palinologicznych i litologicznych, *Biuletyn Instytutu Geologicznego* 284, *Z badań złóż węgli brunatnych w Polsce* 5, 41—70.
Gradziński R., Wójcik Z., 1966, O krasie kopalnym w Polsce, *Prace Muzeum Ziemii* 9, *Prace Geologiczne*, 151—203.
Harasimiuk M., 1975, *Rozwój rzeźby Pagórów Chełmskich w trzeciorzędzie*, Prace Geograficzne IG PAN 115.
Harasimiuk M., Henkiel A., 1981, Kopalne formy dolinne w okolicy Łęcznej i ich znaczenie dla paleogeografii dorzecza Wieprza, *Kwartalnik Geologiczny* 25, 1, 147—161.
Henkiel A., 1969, Rozwój rzeźby dorzecza Strwiąża (Karpaty Wschodnie), *Annales UMCS*, sectio B, 24, 99—148.
Jahn A., 1953, Morfologiczna problematyka zachodnich Sudetów, *Przegląd Geograficzny* 25, 3, 1—9.
Jahn A., 1956, *Wyżyna Lubelska*, Prace Geograficzne IG PAN 7.
Jahn A., 1980, Główne cechy i wiek rzeźby Sudetów, *Czasopismo Geograficzne* 51, 2, 129—154.
Jahn A., 1984, Sudety na przełomie trzeciorzędu i czwartorzędu, in: *Przewodnik Ogólnopolskiego Zjazdu Polskiego Towarzystwa Geograficznego*, Lublin 13—15 IX 1984, część 1.
Jahn A., Łańcucka—Środoniowa M., Sadowska A., 1984, Stanowisko utworów plioceńskich w Kotlinie Kłodzkiej, *Geologia Sudetica* 18, 2, 7—43.
Jawor E., 1983, Utwory miocenu między Krakowem a Dębicą, *Przegląd Geologiczny* 12, 635—641.
Jońca E., 1963, W sprawie wieku studni krasowych w Smerdynie, *Przegląd Geograficzny* 35, 1, 95—99.
Jucha S., 1985a, The Carpathian foredeep in the light of the works done in the Department of Drilling and Oil Engineering of the Academy of Mining and Metallurgy in Cracow, in: *Drilling, Oil and Gas, Zeszyty Naukowe AGH*.

Jucha S., 1985b, New features of structure of the Carpathian foredeep and basement of the Carpathian Mts, *Przegląd Geologiczny* 6, 333—345.

Karnkowski P. H., 1980, Paleotektonika pokrywy platformowej w Wielkopolsce, *Przegląd Geologiczny* 3, 146—151.

Kasiński J. R., 1981, Rekonstrukcja środowiska sedymentacji formacji brunatnowęglowej w niecce żytawskiej na podstawie analizy facjalnej, *Kwartalnik Geologiczny* 25, 4, 796—797.

Kasiński J. R., 1983, Mechanizm sedymentacji cyklicznej osadów trzeciorzędowych w zapadliskach przedpola Sudetów, *Przegląd Geologiczny* 4, 237—243.

Kasiński J. R., 1984, Tektonika synsedymentacyjna jako czynnik warunkujący sedymentację formacji brunatnowęglowej w zapadliskach tektonicznych na obszarze zachodniej Polski, Przegląd Geologiczny 5, 260—268.

Klatkowa H., 1972, *Paleogeografia Wyżyny Łódzkiej i obszarów sąsiednich podczas zlodowacenia warciańskiego*, Acta Geographica Lodziensia 28.

Klimaszewski M., 1934, Z morfogenezy Polskich Karpat Zachodnich, *Wiadomości Geograficzne* 12, 30—44.

Klimaszewski M., 1958a, Rozwój geomorfologiczny terytorium Polski w okresie przedczwartorzędowym, *Przegląd Geograficzny* 30, 1, 3—43.

Klimaszewski M., 1958b, Nowe poglądy na rozwój rzeźby krasowej, *Przegląd Geograficzny* 30, 3, 421—438.

Klimaszewski M., 1965, Views on the geomorphological evolution of the Polish West Carpathians in Tertiary times, in: *Geomorphological Problems of the Carpathians* 1, Vyd. Slov. Akad. Vied, Bratislava, 91—126.

Klimaszewski M., ed., 1972, *Geomorfologia Polski*, tom 1, Polska Południowa, Góry i Wyżyny, PWN, Warszawa.

Klimaszewski M., 1987, The geomorphological evolution of the Tatra Mts of Poland, *Zeitschrift für Geomorphologie* suppl. 65.

Klimek K., 1967, *Rozwój dolin w północnej części Wyżyny Śląsko-Krakowskiej*, Instytut Geografii PAN, Cracow, manuscript.

Kosmowska—Suffczyńska D., 1966, *Rozwój rzeźby w trzeciorzędzie okolic Ostrowca Świętokrzyskiego i Ćmielowa*, Prace Geograficzne IG PAN 54.

Kotlarczyk J., 1979, Węgiel we fliszu karpackim — kilka spostrzeżeń sedymentologicznych, Rocznik Polskiego Towarzystwa Geologicznego, 49, 1—2, 135—154.

Kotlarczyk J., 1986, Tektogeneza mioceńska Karpat, in: *Historia ruchów tektonicznych na ziemiach polskich. Cykl alpejski.* Summaries, Wyd. Uniw. Wrocławskiego.

Kowalski K., 1956, Insectivores, bats and rodents from the early Pleistocene bone breccia of Podlesice near Kroczyce (Poland), *Acta Palaeontologica Polonica* 1, 4, 331—394.

Kowalski W. M., 1979, Rozmieszczenie pstrych utworów na podmioceńskich wychodniach skał karbońskich w południowo-wschodniej części Rybnickiego Okręgu Węglowego, *Annales UMCS* 34, 6, sectio B, 113—122.

Kowalski W. M., 1980—1981, Przejawy sylifikacji w obrębie utworów karbońskich polskich zagłębi węglowych, *Annales UMCS* 35/36, 3, sectio B, 29—35.

Krach W., 1947, Miocen okolic Miechowa, *Biuletyn Państwowego Instytutu Geologicznego* 43.

Krach W., 1962, Zarys stratygrafii miocenu Polski południowej, *Rocznik Polskiego Towarzystwa Geologicznego* 32, 4, 529—556.

Krumbiegel G., 1959, *Die tertiäre Pflanzen- und Tierwelt der Braunkohle des Geiseltales*, A. Ziemsen Verlag, Wittenberg Lutherstadt.

Krygowski B., 1952, Zagadnienie czwartorzędu i podłoża środkowej części Niziny Wielkopolskiej, *Biuletyn Instytutu Geologicznego* 66, 189—212.

Krzemiński T., 1965 *Przełom doliny Warty przez Wyżynę Wieluńską*, Acta Geographica Lodziensia 21.

Książkiewicz M., 1931, Sprawozdanie z badań geologicznych wykonanych w r. 1930 w północnej części arkusza Wadowice, *Posiedzenia Naukowe Państwowego Instytutu Geologicznego* 30, 48—50.

Książkiewicz M., 1972, Karpaty, in: *Budowa geologiczna Polski*, 4, *Tektonika* 3.
Ksiażkiewicz M., Samsonowicz J., Rühle E., 1965, *Zarys Geologii Polski*, Wyd. Geol. Warszawa.
Kuźniar W., 1910, Eocen Tatr i Podhala, *Sprawozdanie Komisji Fizjograficznej PAU* 44, 26—76.
Laskowska—Wysoczańska W., 1971, *Stratygrafia czwartorzędu i paleogeomorfologia Niziny Sandomierskiej i Przedgórza Karpat rejonu rzeszowskiego*, Wyd. Geol., Warszawa.
Laskowska—Wysoczańska W., 1975, *Zagadnienia paleohydrografii Kotliny Sandomierskiej*. Sesja naukowo-sprawozdawcza Pracowni Geologii Czwartorzędu ZNG PAN, Streszczenia referatów, Warszawa, 17—31.
Lencewicz S., 1934, Surfaces d'aplanissement tertiaires dans les Monts Łysogóry, in: *Comptes rendus du Congrès International de Géographie*, Varsovie, 70—71.
Lewiński J., 1914, *Utwory dyluwialne i ukształtowanie powierzchni przedlodowcowej dorzecza Przemszy*, Warszawskie Towarzystwo Naukowe.
Lilpop J., 1957, *Roślinność Polski w epokach minionych*, Wyd. Geol., Warszawa.
Liszkowski J., Stochlak J., 1969, Rozprzestrzenienie i geneza trzeciorzędowych osadów glaukonitowych na obszarze NE obrzeżenia Gór Świętokrzyskich, *Przegląd Geologiczny* 2, 80—83.
Łańcucka—Środoniowa M., 1957, Mioceńska flora z Rypina na Pojezierzu Dobrzyńskim, Prace Instytutu Geologicznego 15, Z badań flor trzeciorzędowych 2, 5—55.
Łańcucka—Środoniowa M., 1966, Tortonian flora from the 'Gdów Bay' in the south of Poland, *Acta Palaeobotanica* 7, 1.
Łyczewska J., 1958, Stratygrafia paleogenu i neogenu Polski północnej, *Kwartalnik Geologiczny* 2, 1, 127—155.
Matl K., Śmigielska T., 1977, Palaeogene marine sediments between Głogów and Sieroszowice (Lower Silesia — Poland), *Rocznik Polskiego Towarzystwa Geologicznego* 47, 1, 11—25.
Michalski A., 1884, Krótki zarys geologiczny południowo-wschodniej Guberni Kieleckiej, Pamiętnik Fizjograficzny 7.
Milewicz J., Grocholski A., 1960, Trzeciorzęd między Bolesławcem a Węglińcem, *Biuletyn Instytutu Geologicznego* 151, 25—69.
Młynarski M., 1955, Żółwie z pliocenu Polski, *Acta Geologica Polonica* 5, 161—214.
Mochnacka K., 1979, Geneza mineralizacji uranowej bloku karkonosko-izerskiego, *Rocznik Polskiego Towarzystwa Geologicznego* 49, 1—2, 195—210.
Mossoczy Z., 1959a, Odkrycie miejsc występowania kości kręgowców kopalnych w okolicach Kłobucka, *Przegląd Geologiczny* 3, 132—134.
Mossoczy Z., 1959b, Zagadnienia wieku jaskiń północnej części Jury Krakowsko—Wieluńskiej, *Speleologia* 1, 4, 201—210.
Mycielska E., 1960, Formy krasowe na północ od Osieka Sandomierskiego, *Przegląd Geograficzny* 32, 4, 575—585.
Mycielska—Dowgiałło E., 1965, Rozwój Geomorfologiczny południowo-wschodniej części Wyżyny Sandomierskiej w górnym miocenie i pliocenie, *Przegląd Geograficzny* 37, 4, 637—649.
Ney R., Burzewski W., Bachleda T., Górecki W., Jakóbczak K., Słupczyński K., 1974, *Zarys paleogeografii i rozwoju litologiczno-facjalnego utworów miocenu zapadliska przedkarpackiego*, Wyd. Geol. Warszawa.
Nowak J., 1927, *Zarys tektoniki Polski*.
Oberc J., Dyjor S., 1969, Uskok sudecki brzeżny, *Biuletyn Instytutu Geologicznego* 236, *Z badań tektonicznych w Polsce* 2, 42—142.
Odrzywolska—Bieńkowa E., Kosmowska—Ceranowicz B., Ciuk E., Giel M. D., Grabowska I., Piwocki M., Pożaryska K., Ważyńska H., Ziembińska—Tworzydło M., 1979, Syntetyczny profil stratygraficzny trzeciorzędu polskiej części północno-zachodniego basenu trzeciorzędowego Europy, *Przegląd Geologiczny* 9, 481—489.
Oszast J., 1970, O wieku stożka Domańskiego Wierchu na podstawie badań palinologicznych, Kwartalnik Geologiczny 14, 4, 843—846.
Oszast J., 1973, The Pliocene profile of Domański Wierch near Czarny Dunajec in the light of palynological investigations (Western Carpathians, Poland), *Acta Palaeobotanica* 14, 1.
Oszast J., Stuchlik L., 1977, Roślinność Podhala w neogenie, *Acta Palaeobotanica*, 18, 1, 46—73.

Oszczypko N., Tomaś A., 1976, Przedtortońska rzeźba przedgórza Karpat między Krakowem a Dębicą oraz jej wpływ na sedymentację miocenu, *Rocznik Polskiego Towarzystwa Geologicznego* 46, 4, 525—548.

Passendorfer E., 1959, Paleogeografia wyspy tatrzańskiej w czasie eocenu, *Biuletyn Instytutu Geologicznego* 149, 259—269.

Pernarowski L., 1963, Morfogeneza północnej krawędzi Wzgórz Niemczańskich, *Acta Universitatis Wratislaviensis* 10, *Studia Geographica* 2, Wrocław.

Piwocki M., 1975, Trzeciorzęd okolic Rawicza i jego węglonośność, *Biuletyn Instytutu Geologicznego* 284, 73—169.

Polichtówna J., 1962, Ostańce Wyżyny Krakowsko—Częstochowskiej, ich geneza i znaczenie w krajobrazie, *Ochrona Przyrody* 28, 255—284.

Połtowicz S., 1974, Wgłębna tektonika brzegu Karpat w okolicy Tarnowa i Pilzna, *Rocznik Polskiego Towarzystwa Geologicznego* 44, 4. 491—514.

Połtowicz S., 1978, Grawitacyjne ześlizgi fliszu brzeżnej strefy Karpat w świetle nowej tektoniki globu ziemskiego, *Rocznik Polskiego Towarzystwa Geologicznego* 48, 3—4, 407—444.

Połtowicz S., Starczewska—Popow A., 1973, Rozwój zapadliska przedkarpackiego między Tarnowem a Przemyślem, *Rocznik Polskiego Towarzystwa Geologicznego* 43, 4, 495—517.

Pożaryska K., Odrzywolska—Bieńkowa E., 1982, Wpływ tektoniki na sedymentację w trzeciorzędzie na Niżu Polskim, *Przegląd Geologiczny* 11, 589—591.

Pożaryska K., Pożaryski W., 1984, Postęp w badaniach geochronologicznych, *Przegląd Geologiczny* 10, 513—517.

Pożaryski W., 1951, *Odwapnione utwory kredowe na północno-wschodnim przedpolu Gór Świętokrzyskich*, Biuletyn Państwowego Instytutu Geologicznego 75.

Paulina M., 1964, *Zjawiska krasowe w Sudetach*, manuscript.

Radłowska C., 1963, *Rzeźba północno-wschodniego obrzeżenia Gór Świętokrzyskich*, Prace Geograficzne IG PAN 38.

Radwańska U., Radwański A., 1984, A new species of inarticulate brachiopods, *Discinisca polonica sp. n.* from the Korytnica Basin (Middle Miocene, Holy Cross Mts, central Poland), *Acta Geologica Polonica* 28, 3—4, 253—269.

Radwański A., 1969, Transgresja dolnego tortonu na południowych stokach Gór Świętokrzyskich (strefa zatok i ich przedpola), *Acta Geologica Polonica* 19, 1, 1—137.

Raniecka—Bobrowska J.,1959, Trzeciorzędowa flora nasienna z Konina, *Biuletyn Instytutu Geologicznego* 13, *Z badań trzeciorzędu w Polsce* 2, 159—252.

Różycki S. Z., 1937, Alb, cenoman i turon w okolicy stacji Złoty Potok (koło Koniecpola), *Sprawozdania Państwowego Instytutu Geologicznego* 9, 1, 19—68.

Różycki S. Z., 1967, Plejstocen Polski Środkowej na tle przeszłości w górnym trzeciorzędzie, 2nd ed. 1972.

Roniewicz P., 1969, Sedymentacja eocenu numulitowego Tatr, *Acta Geologica Polonica* 19, 3, 503—608.

Ruszczyńska H., 1961, Przedplejstoceńskie powierzchnie gradacyjne i stratygrafia osadów czwartorzędowych w dorzeczu Uniejówki, in: *Prace o Plejstocenie Polski Środkowej*, Wyd. Geol., Warszawa, 107—131.

Rutkowski J., 1965, Senon okolicy Miechowa, *Rocznik Polskiego Towarzystwa Geologicznego* 35, 1, 3—53.

Rühle E., 1955, Przegląd wiadomości o podłożu czwartorzędu północno-wschodniej części Niżu Polskiego, *Biuletyn Instytutu Geologicznego* 70, 159—173.

Sadowska A., 1985, Plioceńskie flory Polski południowo-zachodniej, in: *Plioceńska i eoplejstoceńska sieć rzeczna...*, Summaries, PAN, Wrocław.

Samsonowicz J., 1934, *Objaśnienie arkusza Opatów*, 1, PIG, Warszawa.

Samsonowicz J., 1936, Zjawiska krasowe i trzeciorzędowa brekcja kostna w Wężach pod Działoszynem, *Zabytki Przyrody Nieożywionej* 3, 147—155.

Sawicki L., 1909, *Z fizjografii Karpat Zachodnich*, Archiwum Naukowe, Lwów.

Sawicki L., 1925, *Przełom Wisły przez Średniogórze polskie*, Prace Instytutu Geografii UJ 4.

Smoleński J., 1937, W sprawie wieku i genezy krajobrazu Beskidów Zachodnich, *Wiadomości Geograficzne* 15, 59—74.

Stach J., 1953, *Ursus wenzensis*, nowy gatunek małego niedźwiedzia plioceńskiego. Studia nad trzeciorzędową fauną brekcji kostnej w miejscowości Węże koło Działoszyna, część 2, *Acta Geologica Polonica* 3, 1, 103—136.

Stachurska A., Sadowska A., Dyjor S., 1973, The Neogene flora at Sośnica near Wrocław in the light of geological and palynological investigations, *Acta Palaeobotanica* 14, 3, 147—176.

Stahl A., 1933, Das Relief des oberschlesischen Steinkohlengebirges im Lichte der Palaeogeographie, *Jahrbuch der Preussischen Geologischen Landesanstalt* 53, (für 1932), 95—111.

Starkel L., 1965, *Rozwój rzeźby polskiej części Karpat Wschodnich*, Prace Geograficzne IG PAN 50.

Starkel L., 1969, The age of the stages of development of the relief of the Polish Carpathians in the light of the most recent geological investigations, *Studia Geomorphologica Carpatho-Balcanica* 3, 33—44.

Steenis van, C. G. G. J., 1962, The land-bridge theory in botany with particular reference to tropic plants, *Blumea* 11, 2, 235—542.

Stuchlik L., 1980, Chronostratygrafia neogenu Polski południowej (północna część Paratetydy centralnej) na podstawie badań paleobotanicznych, *Przegląd Geologiczny* 8, 443—448.

Szafer W., 1946—1947, Flora plioceńska z Krościenka nad Dunajcem, *Rozprawy PAU*, B, 72.

Szafer W., 1954, *Plioceńska flora okolic Czorsztyna i jej stosunek do plejstocenu*, Wyd. Geol., Warszawa.

Szafer W., 1958, Nowa flora eoceńska w Tatrach. Wiadomość wstępna, *Kwartalnik Geologiczny* 2, 1, 173—174.

Szafer W., 1961, *Mioceńska flora ze starych Gliwic na Śląsku*, Wyd. Geol., Warszawa.

Szaflarski J., 1955, Zarys rozwoju ukształtowania Wyżyny Śląskiej, in: *Górny Śląsk*, *Prace i materiały geograficzne*, 65—116.

Ślączka A., 1977, Rozwój osadów miocenu z otworu wiertniczego Sucha IG — 1, *Kwartalnik Geologiczny* 21, 2, 404—405.

Teisseyre H., 1960, Rozwój budowy geologicznej Sudetów od prekambru po trzeciorzęd, in: *Regionalna geologia Polski*, 3, *Sudety*, 2, 335—358.

Tokarski A., 1978, Orogenesis and morphogenesis of the Outer Carpathians and plate tectonics, *Studia Geomorphologica Carpatho-Balcanica* 12, 29—43.

Traczyk S., 1978, Charakterystyka mineralogiczno-petrograficzna serii pstrych łupków fliszu karpackiego oraz niektóre problemy ich sedymentacji i diagenezy, *Kwartalnik Geologiczny* 22, 2, 323—333.

Tyczyńska M., 1959, Morfologia środkowej części dorzecza Szreniawy, *Dokumentacja Geograficzna* 6.

Uberna T., 1962, Dolomity środkowodewońskie w Winnej koło Łagowa, *Przegląd Geologiczny* 10, 515—519.

Watycha L., 1976, Neogen niecki orawsko-nowotarskiej, *Kwartalnik Geologiczny* 20, 3, 575—585.

Zuchiewicz W., 1984, Ewolucja poglądów na genezę i wiek karpackich powierzchni zrównania, *Przegląd Geologiczny* 8—9, 468—477.

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DEPOSITIONAL MODELS AND ICE-FRONT DYNAMICS IN NORTHWESTERN POLAND: A METHODOLOGICAL APPROACH

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INTRODUCTION

Ice marginal deposits and forms together with the so-called directional elements of tills, streamlined forms and striated rocks are a fundamental source of information about the directions of movement and extent lines of Pleistocene ice sheets. Their informative functions are best in areas of the last glaciation because of freshness of sedimentological and geomorphic record. Hence, they have been most successfully used in establishing the maximum extent of the last Scandinavian ice sheet and its waning phases in the North Polish Plain at an over-regional (Bartkowski 1969; Galon 1968, 1972; Roszko 1968; Różycki 1973) and regional (Bartkowski 1967; Karczewski and Roszko 1972; Kliewe and Kozarski 1979; Kozarski 1962; Rotnicki 1963; Żynda 1967) spatial scale. The growing knowledge on glaciarization processes and ice sheet extension brought also increasing information about ice-front dynamics. It results from descriptions presented in literature on this subject that attention has been mostly focused on two extreme states of ice-front behaviour namely:

(1) the state of high activity and advance recorded by the presence of thrust ridges containing glaciotectonic structures, and

(2) the state of complete passiveness represented by disintegration features, i. e. dead-ice forms and deposits containing gravity deformation structures.

The first state is easily detectable, wheras the other one poses more interpretation problems because the preceding it slow ice-front advance and/or dynamic equilibrium recorded by deposits, except some cases (Kasprzak and Kozarski 1984; Kozarski 1978, 1981; Roszko 1968), has been ignored. Direct causes comprised here deficiences in too general programmes and very few detailed research procedures which did not promote careful identification and advanced interpretation of deposits occurring in marginal zones. As a consequence simplifications in the reconstruction of deglaciation process appeared with an exaggerated preference for the model of vast zonal ice-sheet wastage.

This gave rise to search for new approaches in studies of deposits in marginal zones so that a better, broad and objective basis might be provided for a reconstruction of the last ice sheet behaviour in frontal parts during deglaciation.

FACIES ANALYSIS OF SEDIMENTS

Recent observations and experience from Great Poland Lowland (Kozarski and Kasprzak 1987) indicate that such approaches are attempted in procedures using the facies analysis known from sedimentology (Gradziński et al. 1976). Its variants are successfully applied in glacial geology and geomorphology. Up to now this analysis has been used in few selected test areas (Fig.1) subjected to detailed investigations. It involves (Kasprzak and Kozarski 1984):

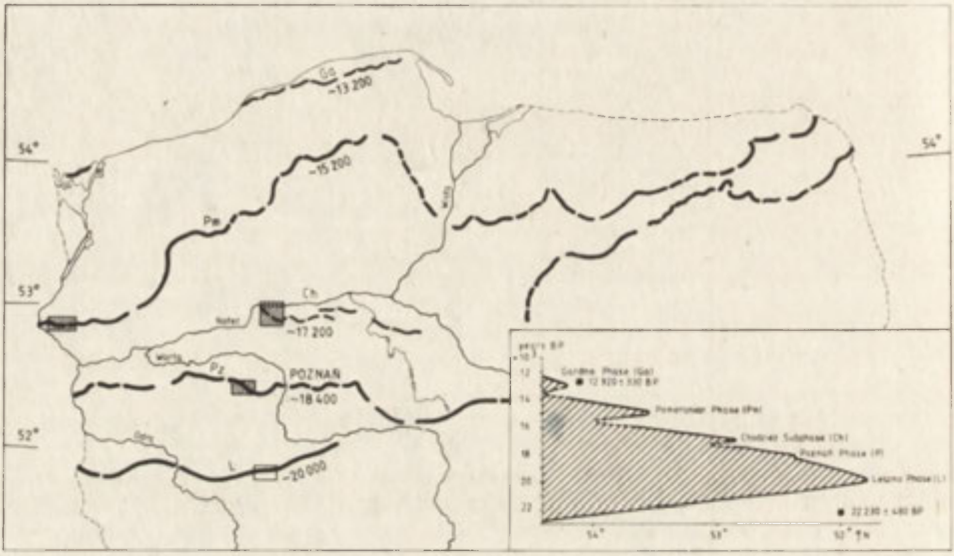

Fig. 1. Location of case study areas and major Vistulian ice sheet positions in northwestern Poland with esimated radiocarbon age BP (after Kozarski, in press).
Hatching shows the case studies mentioned in this paper. Inset: time-distance diagram of the last ice sheet in northwestern Poland (Source: Kozarski, in press)

(1) identification of diagnostic properties of each sediment resulting from its structure and texture,

(2) spatial distribution and thickness of all deposits, and

(3) geological interrelationships between deposits and their geomorphic position.

This variant of analysis is termed (Kasprzak and Kozarski 1984) complex facies analysis which proves helpful in constructing the sheme of local lithostratigraphy associated with a sequence of geomorphic events and palaeogeographic conclusions as well. At full research possibilities, which have not been available as yet, the complex facies analysis should be intended to recognize in detail lithofacies units, i. e. lithosomes (Gradziński et al. 1976) referred to as three-dimensional rock masses displaying constant lithofacies properties.

DIAGNOSTIC DEPOSITS AND STRUCTURES

The construction of depositional models intended to represent the dynamic state of the ice-sheet front must be preceded by ordering and clustering of deposits and structures according to their diagnostic value in the determination of activity or stagnation and disintegration of ice masses. Diagnostic deposits and structures known from extensive literature have been aggregated in this paper into three groups (Table 1) which reflect:

(1) ice-front advance resulting from accumulation (A_c) or alimentation (A_l) excess over ablation (A_b), (2) ice-front dynamic equilibrium (= steady state) when the addition of ice mass (A_c or A_l) is balanced by the ice mass loss due to ablation (A_b) in the marginal zone, and (3) ice-front retreat and disintegration; the former state results from slow recession of an active front ($A_c < A_b$), while the latter one is associated with a change into dead ice and its ablation (A_b) only.

TABLE 1. Deposits and structures diagnostic for ice-front dynamics reconstruction

Ice-front net balance	Diagnostic deposits and structures
$A_c > A_b$ advance	Lodgement till Glaciotectonic macrostructures Glaciotectonic microstructures Mylonitization, deformation till
$A_c = A_b$ steady state	Lodgement till Melt-out till Allochthonous flow till Glaciofluvial deposits supraglacial facies extra-and proglacial facies Glaciotectonic microstructures
$A_c < A_b$ retreat and disintegration A_b	Allochthonous flow till Sandy and stony deposits of rims around closed dead ice hollows flow till facies lacustrine facies Melt-out till Parautochthonous melt-out till and sand Glaciofluvial and glaciolacustrine deposits of in-, intra- and supraglacial facies related to stagnant and dead ice Gravity structures, faults and flexures Diapiric structures (intrusions)

A_c = accumulation (alimentation), A_b = ablation.

Glaciotectonic macro- and microstructures are largely of diagnostic value for the first state. They comprise, for instance, shear planes, flaky-raft structures, compressional faults, various types of folds, including drag-folds which are particularly specific among microsctructures, and mylonitization zones as horizons of primary structure total destruction and mixing of deposits differing in origin and/or facies. Lodgement till which is found on proximal slopes of thrust ridges and push moraines (Kasprzak 1985; Kozarski 1959) remains complementary.

Lodgement till and subglacial melt-out till (Boulton and Paul 1976) in a submarginal position (Kozarski, in press) have primary diagnostic significance for the second state as beds reflecting zones of differentiated debris transport in the warm sole of an active ice sheet or in that subjected to regelation, i. e. the zones of traction and suspension (Boulton and Paul 1976), respectively. Flow tills in ablation end moraines

(Kozarski 1981) which do not need to be manifested in the relief are another diagnostic facies of deposits. They form discontinuous sedimentary units and originate at the surface of the snout to become next accumulated at its base. According to Boulton et al. (1985), two processes may be responsible for the formation of such tills, namely dumping or tectonic stacking in conditions of compressive ice flow. Allochthonous flow tills (Boulton and Paul 1976) in extramarginal position (Kasprzak and Kozarski 1984 ; Kozarski, in press) which are genetically associated with the latter variant of the process play an important diagnostic role, too, since they are also generated by compressive ice flow from the ice sheet sole onto the surface.

Fluvioglacial deposits representing the supraglacial facies formed as fans in the lowermost part of the snout or those representing the extra- and proglacial facies laid down by meltwater immediately before and behind the extent line of the ice-sheet front to form outwash plains are another indicator of the dynamic equilibrium state.

The greatest variety of diagnostic structures and deposits considered jointly allow to reconstruct the state of ice-sheet frontal retreat and/or disintegration. Among them ablation end moraine composed of redeposited flow tills again (Kozarski 1978, 1981) and similar in facies sandy-stony deposits of which rims around hollows are built up (Kasprzak and Kozarski 1984), occur as ice-cored moraine scars.

After melting out of thin dead-ice patches over vast areas, parautochthonous tills (Boulton and Paul 1976) and melt-out sands (Kasprzak and Kozarski 1984) are deposited on flat or gently sloping surfaces. On the other hand, discontinuous sandy-silty or sandy-gravelly lithofacies units which build up kames and eskers or fill up depressions of former ice-dammed lakes in different geomorphic settings are indicative of dissipative or chaotic deposition in different places within the stagnant and dead ice. The effects of dead ice melting have been preserved in those deposits as gravity, fault and flexure structures (McDonald and Shilts 1975), while static pressure of dead-ice blocks produces till wedges (Bramer 1961; Keller 1954; Rotnicki and Wasilowska 1962) or diapirs (Baherjee and McDonald 1975) which have been squeezed up from the substratum into lateral or central parts of eskers.

DEPOSITIONAL MODELS OF SELECTED ICE-SHEET MARGINAL ZONES

The complex facies analysis of deposits and aggregation of deposits and structures according to their diagnostic properties (Table 1) represent a preparatory stage in the construction of a marginal zone depositional model. Likewise the facies model defined by Walker (1980, p. 5), it must fulfil four important functions and act as:

(1) a norm for purposes of comparison,

(2) a framework and guide for future observations,

(3) a predictor in new geological situations, and

(4) a basis for dynamic interpretation of an environment or system which it represents; in our case account is taken of glaciodynamic interpretation.

At the present stage of research, the depositional model is discussed and presented as a local summary sequence (LSS) which contains the basic information about diagnostic deposits and structures, about the primary sedimentary environment, as well as about the dynamic tendency (D_t) of the ice-sheet front which has been established on these grounds. It is easy to note that the above variant of the model fulfils all of the four functions. A similar model but representing a simple methodological and interpretative variant has been recently shown by Kozarski and Kasprzak 1987.

Sets of lithofacies units which have definite diagnostic properties and reflect ice flow in the marginal zone or its stagnation and disintegration serve as a criterion in the evaluation of dynamic state of the ice-sheet front. Among the examples given below (Figs 1 and 2) there are no considerations of a model representing high-energy state of

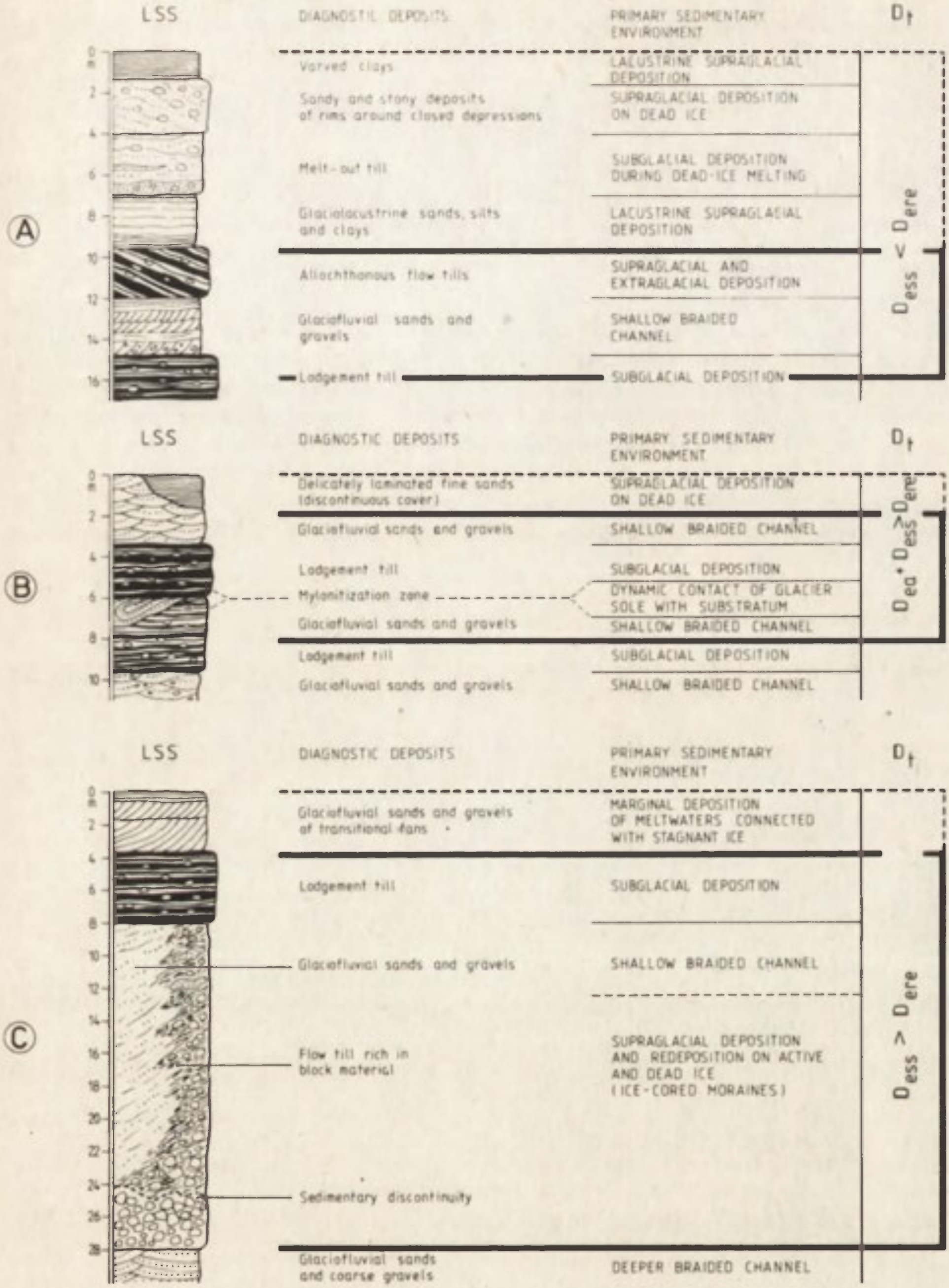

Fig. 2. Depositional models for case study areas (hatching) shown in Fig. 1.
A — Poznań Phase, B — Chodzież Subphase, C — Pomeranian Phase; LSS — local summary sequence, D_t — dynamic tendency

the ice-sheet front, which generates the formation of glaciotectonic macrostructures in thrust ridges and/or push moraines, since such a state is easiest for identification. Attention is focused on the remaining models showing the states of: (1) dynamic equilibrium (= steady state), (2) slow advance and dynamic equilibrium, and (3) retreat and disintegration of the ice-sheet front.

The sedimentological criterion was nearly absent from the earlier attempts to establish the dynamic state of the ice-sheet front (Bartkowski 1967, 1969; Karczewski and Roszko 1972; Roszko 1968). The geomorphic criterion of evaluation prevailed. It was supported by general recognition of the nature of deposits of which the marginal forms were built up and of the existing deformations, especially discontinuous ones. Poor accuracy of research and random distribution of exposures led to subjective assessments. In view of those deficiences, depositional models have been intended to provide objective assessments resulting from meticulously planned and conducted research at test sites (Fig. 1) by means of complex facies analysis, with a choice of uniform criteria, i.e. diagnostic deposits and structures, at disposal to distinguish between sets of lithofacies units, as has been mentioned above.

Under ideal conditions these units should be lithosomes (L) which allow to determine the depositional effectiveness (D_e) resulting from a definite dynamic state of the ice-sheet front and to calculate the cumulative volume of lithosomes, i.e.

$$D_e = L_1 + L_2 + L_3 + \dots L_n.$$

However, in basic research it is difficult to collect material which would allow exact calculation of lithosome volume because of financial problems. Hence, such calculations can be replaced by merely estimation of the depositional effectiveness (D_e) as the mean thickness of a lithofacies unit defined by a limited number of boreholes, as well as large and small pits. This value should by reflected in the local summary sequence (LSS).

Under both ideal and substitute circumstances the dynamic tendency (D_t) typical of a given marginal zone in the case study area may be expressed by a general ratio of depositional effectiveness during the state of dynamic equilibrium (= steady state) of the ice-sheet front (D_{ess}) to that associated with retreat and ice disintegration (D_{ere}), i.e.

$$D_t = \frac{D_{ess}}{D_{ere}},$$

and if account is taken of slow transgression (D_{ea}) which precedes the state of dynamic equilibrium, the ratio as follows:

$$D_t = \frac{D_{ea} + D_{ess}}{D_{ere}}.$$

In case the lithosome volumes are known dynamic tendency can be expressed by numerical data. If merely average values of the thickness of lithofacies units are available the dynamic tendency can be defined as the estimated dominance of one depositional effectiveness over the other (Fig. 2 — A,B,C), i.e.

$$D_{ess} > D_{ere}, \text{ or } D_{ea} + D_{ess} > D_{ere} \text{ and}$$

$$D_{ess} < D_{ere}, \text{ or formally } D_{ess} = D_{ere}.$$

It is just this approach that is applicable to the recently studied portions of marginal zones (Fig. 1) belonging to the Poznań Phase (Kasprzak and Kozarski 1984), the Chodzież Subphase (Kozarski and Kasprzak, 1987), and the earlier studied section of the Pomeranian Phase (Kozarski 1978, 1981). The relationships that are described by depositional models for case study areas of the above mentioned marginal zones (Fig. 1 and 2) are given below.

(1) Clearly distinguishable sets of lithofacies units belonging to the Poznań Phase (Fig. 2 A) are indicative of the dynamic equilibrium (= steady state) as well as ice front retreat and disintegration. Depositional effectiveness throughout dynamic equilibrium (D_{ess}) is smaller than that during retreat and disintegration (D_{ere}). For justified reasons the view (Bartkowski 1967, 1969) on deposits accumulation and mostly shaping of forms in conditions of ice sheet stagnation and disintegration only should be denied.

(2) The depositional model representing the Chodzież Subphase (Fig. 2 B) reveal a well marked dominance of a set of lithofacies units associated with active ice, immediately behind the extent line of this subphase. On the other hand, the amount of sediments deposited in dead-ice conditions remains small. The expression $D_{ea}+D_{ess} > D_{ere}$ for the Chodzież Subphase portion under consideration is complementary to the earlier information (Kozarski, 1959, 1962) about ice-sheet activity, which has been provided by the analysis of glaciotectonic macrostructures.

(3) The depositional model representing the marginal zone portion belonging to the Pomeranian Phase (Fig. 2 C), which lies in the Odra lobe, reveals great dominance of depositional effectiveness of the ice margin at steady state (D_{ess}) over that throughout the ice-sheet retreat (D_{ere}). Flow till lithofacies units of which the ablation end moraine is built up attain the extreme common thickness of 28 metres (Kozarski, 1978, 1981).

Fluvioglacial deposits are also thick. They build up an extensive outwash plain formed when the ice-sheet front occupied the maximum extent line of the Pomeranian Phase. On the other hand lithofacies units which form short and thin transitional fans were associated with stagnant ice only. A great thickness of flow tills very rich in block material gathered in a narrow, scarcely 100 to 200 metres wide, outer strip of the marginal zone also suggest that the state of dynamic equilibrium lasted there for a relatively long time (Kozarski, in press).

CONCLUSIONS

The discussed simple depositional models and procedures by means of which they can be constructed are the result of detailed, painstaking and strongly time-consuming, though still preliminary studies. Nevertheless, it may be assumed that they represent a promising way of determination of the dynamic state of the ice-sheet front throughout the deglaciation process, even in the present stage of research. They are applicable to portions of ice-front zones and their dynamic interpretation, assessment of depositional effectiveness in selected profiles, comparison between case study areas and key sites during lithostratigraphic analyses.

It should be emphasized that the proposed models based on depositional effectiveness are not applicable to areas and sections where lithosomes have been considerably reduced by denudation and erosion. This restriction holds first of all for possible attempts at applying such models for the area of older glaciations where top deposits of convex forms in marginal zones have become strongly reduced.

It is expected that a larger number of studies and their multiplication at test sites located in marginal zones of the last ice-sheet will allow an objective assessment of the state of ice-front dynamics along changeable extent lines and thus, reconstruction of a more realistic palaeogeographic image of the main Vistulian glacial episode.

Depositional models based on the above accepted principles to be improved in the future should also be a component of the verification of theoretical models of the ice-sheet behaviour in marginal zones. Recently Andrews (1982) has pointed out that the future ice-sheet reconstruction will not depend on single representations but multiple ones based on all field and theoretical date. Thus, inductive and deductive research approaches are simultaneously required. The discussed in this paper procedures and models belong to the former of these approaches.

REFERENCES

Andrews J. T., 1982, On the reconstruction of Pleistocene ice sheets, *Quaternary Science Reviews* 1, 1, 1—30.

Baherjee I., McDonald B. C., 1975, Nature of esker sedimentation, in: A. V. Jopling, B. C. McDonald, eds, *Glaciofluvial and glaciolacustrine sedimentation*, Society of Economic Paleontologists and Mineralogists, Special Publication No. 28, 132—154.

Bartkowski T., 1967, *O formach strefy marginalnej na Nizinie Wielkopolskiej*, PTPN, Prace Komisji Geograficzno-Geologicznej 7, 1.

Bartkowski T., 1969, Deglacjacja strefowa deglacjacją normalną na obszarach niżowych (na wybranych przykładach z Polski zachodniej i północnej), *Badania Fizjograficzne nad Polską Zachodnią*, ser. A, 23, 7—30.

Boulton G. S., Paul M. A., 1976, The influence of genetic processes on some geotechnical properties of glacial tills, *Quarterly Journal of Engineering Geology* 9, 159—194.

Boulton G. S., Smith G. D., Jones A. S., Newsome J., 1985, Glacial geology and glaciology of the last mid-latitude ice sheets, *Journal of the Geological Society of London* 142, 447—474.

Bramer H., 1961, Bemerkungen zum Problem der Aufpressungs-Oser, *Geologie* 10, 4/5 480—492.

Galon R., 1968, Przebieg deglacjacji na obszarze Peribalticum, in: *Ostatnie zlodowacenie skandynawskie w Polsce*, PWN Warszawa, 201—216.

Galon R., 1972, Główne etapy tworzenia się rzeźby Niżu Polskiego, in: *Geomorfologia Polski*, 2, PWN Warszawa, 35—110.

Gradziński R., Kostecka A., Radomski A., Unrug R., 1976, *Sedymentologia*, Wyd. Geol., Warszawa.

Karczewski A., Roszko L., 1972, Grundriss des Rückzuges des Inlandeises der letzten Vereisung zwischen der Oder und der Weichsel mit besonderer Berücksichtung des Oderlobus, *Wissenschaftliche Zeitschrift der Ernst-Moritz-Arndt-Universität zu Greifswald*, *Mathematisch--naturwissenschaftliche Reihe* 21, 1, 27—29.

Kasprzak L., 1985, A model of push moraine development in the marginal zone of the Leszno Phase, west central Poland, *Quaternary Studies in Poland* 6, 23—54.

Kasprzak L., Kozarski S., 1984, Analiza facjalna osadów strefy marginalnej fazy poznańskiej ostatniego zlodowacenia w środkowej Wielkopolsce, *Uniw. im. A. Mickiewicza, ser. Geografia* 29.

Keller G., 1954, Drucktexturen in eiszeitlichen Sedimenten, *Eiszeitalter und Gegenwart* 4—5, 158—171.

Kliewe H., Kozarski S., 1979, Zur Verknüpfung von Marginalzonen im Bereich des Oderlobus, *Acta Univ. Nicolai Copernici, Geografia* 14, 46, 21—30.

Kozarski S., 1959, O genezie chodzieskiej moreny czołowej, *Badania Fizjograficzne nad Polską Zachodnią* 5, 45—72.

Kozarski S., 1962, *Recesja ostatniego lądolodu z północnej części Wysoczyzny Gnieźnieńskiej a kształtowanie się pradoliny Noteci-Warty*, PTPN, Prace Komisji Geograficzno-Geologicznej 2, 3.

Kozarski S., 1978, Lithologie und Genese der Endmoränen im Gebiet der skandinavischen Vereisungen, *Schriftenreihe geologische Wissenschaften Berlin*, 9, 179—200.

Kozarski S., 1981, Ablation end moraines in western Pomerania, NW Poland, *Geografiska Annaler* 63A, 3—4, 169—174.

Kozarski S., in press, Osady ablacyjne ostatniego zlodowacenia w Polsce środkowozachodniej: identyfikacja i paleogeograficzne implikacje, *Acta Universitatis Wratislaviensis*.

Kozarski S., Kasprzak L., 1987, Facies analysis and depositional models of Vistulian ice-marginal features in northwestern Poland, in: V. Gardiner, ed., *International Geomorphology* 1986, part II, J. Wiley and Sons Ltd, Manchester 693—710.

McDonald B. C., Shilts W. W., 1975, Interpretation of faults in glaciofluvial sediments, in: A. V., Jopling, B. C. McDonald, eds, *Glaciofluvial and glaciolacustrine sedimentation*, Society of Economic Paleontologists and Mineralogists, Special Publication Nr. 28, 123—131.

Roszko L., 1968, Recesja ostatniego lądolodu z terenu Polski, in: *Ostatnie zlodowacenie skandynawskie w Polsce*, PWN, Warszawa, 65—100.

Rotnicki K., 1963, Zagadnienie zasięgu stadiałów leszczyńskiego i poznańskiego w południowo-wschodniej części Wysoczyzny Gnieźnieńskiej, *Badania Fizjograficzne nad Polską Zachodnią* 11, 133—189.

Rotnicki K., Wasiłowska K., 1962, Oz lubaski, *Badania Fizjograficzne nad Polską Zachodnią* 9, 95—123.

Różycki S. Z., 1973, *Plejstocen Polski środkowej*, 2nd ed. PWN, Warszawa.

Walker R. G., 1980, Facies and facies models, general introduction, in: *Geoscience Canada Reprint Series* 1, *Facies models*, Second Printing, 1—7.

Żynda S., 1967, *Geomorfologia przedpola moreny czołowej stadiału poznańskiego na obszarze Wysoczyzny Lubuskiej*, PTPN, Prace Komisji Geograficzno-Geologicznej 8, 1.

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MAIN PHASES OF EROSION AND ACCUMULATION IN THE PROSNA VALLEY IN THE LAST GLACIAL-INTERGLACIAL CYCLE

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INTRODUCTION

The problem of associating the main phases of erosion and accumulation in river valleys with specific climatic phases of the glacial-interglacial cycle has been known for a long time (Soergel 1921; Penck 1938; Trevisan 1949; Jahn 1956a, b; Schumm 1965; Rotnicki 1974a; Kozarski and Rotnicki 1977, 1978, 1983; Starkel 1983). In the initial period of the investigation of this problem the prevailing view was that expressed by Soergel (1921), among others, that conditions existing in cold periods of the Pleistocene favoured intensive accumulation in river valleys resulting in high valley filling. By the same view, interglacial periods were characterized by deep erosion.

As the time passed, newer and newer studies, kept modifying this view. Jahn (1956a, b) expressed the opinion that the erosion phase appears twice in the glacial-interglacial cycle: first at the turn of an interglacial and a glacial, and then at the transition from the glacial to a temperate period. Schumm (1965) presents a somewhat different view on this problem. According to him, the erosion appears in a late glacial and keeps operating well into early phase of an interglacial, and the tendency towards vertical stabilization of valley floors appears in the optimum of the interglacial. He is of the opinion that in interglacial periods accumulation and aggradation cannot be traced in river valleys. They appear only at the beginning of a next cold period. There is no erosion phase between those of interglacial stabilization and cold aggradation assumed by Jahn (1956a, b).

Later, opinions on this matter underwent further evolution, especially concerning those river valleys which were directly affected by the recession of the last inland ice. In such valleys the phase of deep erosion took place much earlier, viz. when deglaciation started, that is, about 18000 years BP. This was pointed out by Rotnicki (1966, 1974a), Galon (1968), Maruszczak (1968) and Różycki (1972). It should also be added that opinions concerning Holocene tendencies of fluvial processes are not as undivided as those referring to Pleistocene interglacial periods.

AREA OF RESEARCH

The area of the North European Plain is particularly suitable for analysing the connection between the main phases of erosion and accumulation in river valleys and the glacial-interglacial cycle. It follows from the northward slope of the lowland surface.

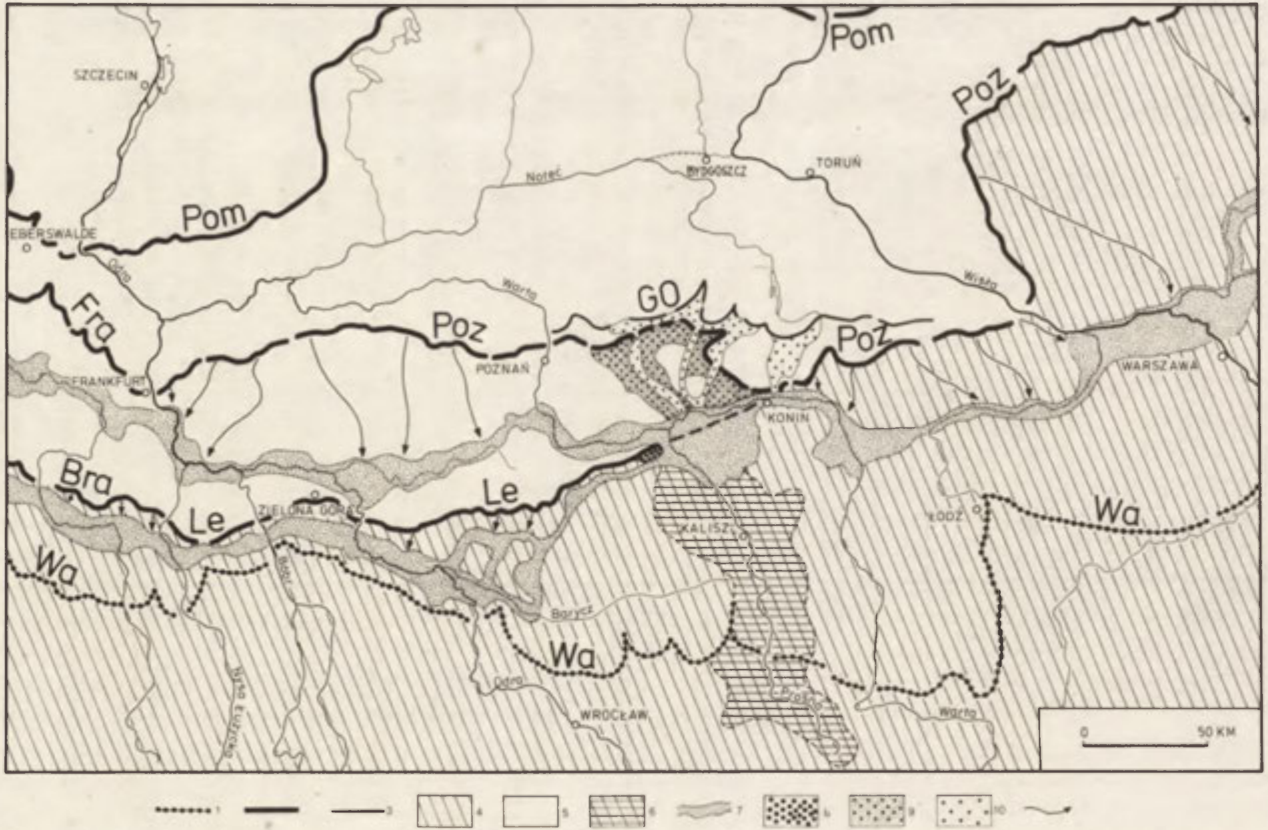

Fig. 1. The Prosna drainage basin against the extent line of the Warta Stage inland ice and the main phases of the Vistulian glaciation with the main draining channels of the ice margins: 1 — extent of the Warta Stage inland ice, 2 — extent of the main phases of the Vistulian inland ice: Le — Leszno Phase, Poz — Poznań Phase, Pom — Pomeraniam Phase, 3 — extent of the last inland ice during the Gniezno Oscillation; 4 — area of Middle Polish Glaciation, 5 — area of the Vistulian Glaciation, 6 — Prosna river drainage basin, 7 — main pradolinas, 8 — outwash plain of the Leszno Phase, connected with the III terrace of the Prosna river, 9 — outwash plain of the Poznań Phase, connected with the IV terrace of the Prosna river, 10 — outwash plain of the Gniezno Oscillation, connected with the VI terrace of the Prosna river, 11 — meltwater flow direction in other places

In the past this slope made it absolutely necessary for fluvio-periglacial waters flowing from south-Polish mountains and uplands to join the melt waters of the inland ice. It resulted in the overlapping of glacial and glaciofluvial sediment series with the periglacial river sediments of the same age, and in the formation of morphologically uniform surfaces of the common water runoff. In this way the morphostratigraphy and lithostratigraphy of the contact of the glacial and extraglacial zones establish certain leading levels separating events preceding a given glaciation from those following it. For this reasons the contact zone of glacial and periglacial areas is particularly useful for the analysis of the appropriate location of erosion and accumulation phases on the curve of climatic changes in the glacial-interglacial cycle.

Out of the rivers of the North Polish Plain the Prosna valley is especially suitable for such studies. Its junction with the Warta valley is situated exactly on the line of the maximum extent of the last glaciation, and during its several phases the Prosna waters were in contact with the melt waters of the inland ice (Fig. 1).

The morphostratigraphic pattern of the Prosna valley terraces and its relation to the outwash levels of the Last Glaciation have been known in a general outline for a long time (Rotnicki 1966, 1974a). Research done in the past five years includes radiocarbon dating of some Pleni-Vistulian deposits series, the determination of the stratigraphic position of deposits forming terrace II of the Prosna in the Grabów Basin, as well as the construction of a general model of erosion and accumulation phases during the last glacial-interglacial cycle.

THE PROSNA TERRACE SYSTEM-DISTRIBUTION, GEOLOGICAL STRUCTURE AND RELATION TO THE FORMS AND DEPOSITS OF THE LAST GLACIATION

In the Prosna valley eight terraces have been distinguished. Not all of them occur along the whole valley of today's Prosna. Their spatial extent and interrelations are shown in Fig. 2.

Terrace I. Only fragments are preserved in the eastern part of the Grabów Basin. It lies at 168—170 m a.s.l. near Wieruszów and at 140—143 m in the northern part of the basin. It is formed of sands and gravels, sometimes agglomerate, of a thickness of 3 to 7 m. They lie on grey-coloured glacial till, usually 7 to 10 m thick. Farther north there are no traces of this terrace.

Terrace II. It is the most extensive area in the Prosna valley, especially in its middle course, in the Grabów Basin. Near Wieruszów it lies at 159—161 m a.s.l. It joins here the high terrace of the Niesób, which fills the Kępno Basin at the southern foreland of the Ostrzeszów Hills (Fig. 3). In the Grabów Basin it slopes toward the north to reach 130—132 m a.s.l. near the village of Sobiesęki situated in the northern fringe of this form. North of the Grabów Basin there are no traces of this terrace in the vicinity of today's Prosna valley, which cuts through the Kalisz Pleistocene Plateau between Kalisz and Chocz to reach the Pyzdry Basin (Fig. 2).

The northward extension of terrace II lies in a valley lowering, dead today, stretching towards the Pyzdry Basin about 15 km east of the present valley. Near Słuszków in the south-eastern part of the Pyzdry Basin, terrace II ends at an altitude of 126 m a.s.l., undercut by lower terrace levels.

Terrace II, whose distribution in the Grabów Basin is presented in Fig. 3, is built of a series of deposits, thick for lowland conditions, amounting to 35—40 m in the middle of the basin. Under them there is glacial till, and here and there glaciotectonically thrusted Tertiary and Quaternary deposits. The situation of the glacial till, high on the fringes of the basin and low in its centre, proves that the Grabów Basin is a form of glacial origin from the Warta stage, transformed later by fluvial processes. In the bottom of terrace deposits in the northern part of the Grabów Basin, on the postglaciotectonic surface

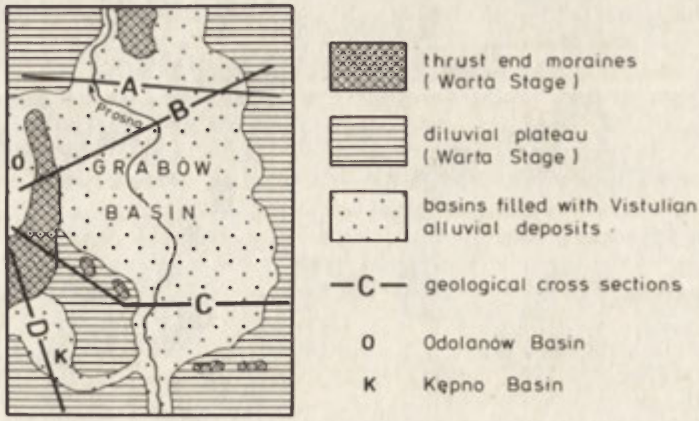

Fig. 3. Location of the geological profiles of the Grabów and Kępno Basins in the drainage basin of the middle Prosna

(Rotnicki 1985), there is a clear valley lowering extending to the north-east. This depression, located 15 km east of the Prosna valley near Kalisz, is a trace of the direction of the Prosna water runoff after the recession of the inland ice of the Central Polish Glaciation (Saalian).

The lithology and stratigraphy of the deposits forming terrace II was studied on the basis of about 150 borings and some dozen exposures (Rotnicki 1986; Rotnicki and Wiśniewski 1986). In the Domasłów and Kępno Basins (Fig. 3), there are peats and gyttja of the Eemian Interglacial in the bottom of deposits of this terrace (Rotnicki and Tobolski 1965a; Rotnicki 1966, 1974b). The terrace series is built there of grey silts and sandy silts about 20 m thick, covered with a sand series some meters in thickness. Under the sands, in the top of the silt series, there is an organic layer of the Denekamp Interstadial dated to 31400 years BP (Fig. 4; Rotnicki and Tobolski 1969; Rotnicki 1974b).

In the recent years a very interesting profile has been discovered at Sobiesęki displaying the structure of terrace II in the northern part of the Grabów Basin. In the bottom of the profile there are coarse-grained sands with admixture of gravel. In other places of the Grabów Basin these sands and gravels rest on glacial till (Fig. 5). They are covered with a bipartite organogenic layer with a thickness of 2.5 m, composed of peats and gyttjas and containing the mollusc fauna. The organogenic layer is divided by a thin band of dark-grey sands. The organogenic series lies at a depth of 30.8 to 33.5 m (Fig. 6). Above it lies a thick fine-grained series built of grey silts, clayey silts and sandy silts intercalated by thin bands of very fine-grained sands. Towards the top of the profile the admixture of sand is more and more frequent. The top of the series lies at a depth of 3.95 m. Thus, it has a thickness of 26 – 27 m. It contains six layers with admixture of organogenic matter. They lie at the following depths: 30.2 – 30.4 m, 29.1 – 29.4 m, 27.5 – 28.0 m, 19.2 – 20.2 m, 7.1 – 7.5 m, and 5.5 – 5.7 m. For the top two organogenic layers 6 datings of various fractions of organic matters were carried out at the C^{14} Laboratory at Gliwice. Two profiles were sampled. The top of the older layer was dated back to from 27780 + 590 years BP to 27960 + 680 years BP (Gd – 757 and Gd – 756), while the younger peat layer was dated back to 26960 + 1050 years BP (Gd – 753) and 26300 + 1140 years BP (Gd – 752). The top part of deposits of terrace II, lying above the the top organogenic layer aged 26,300 years, is formed by medium- and fine-grained sands with a joint thickness of 3.5 – 5 m.

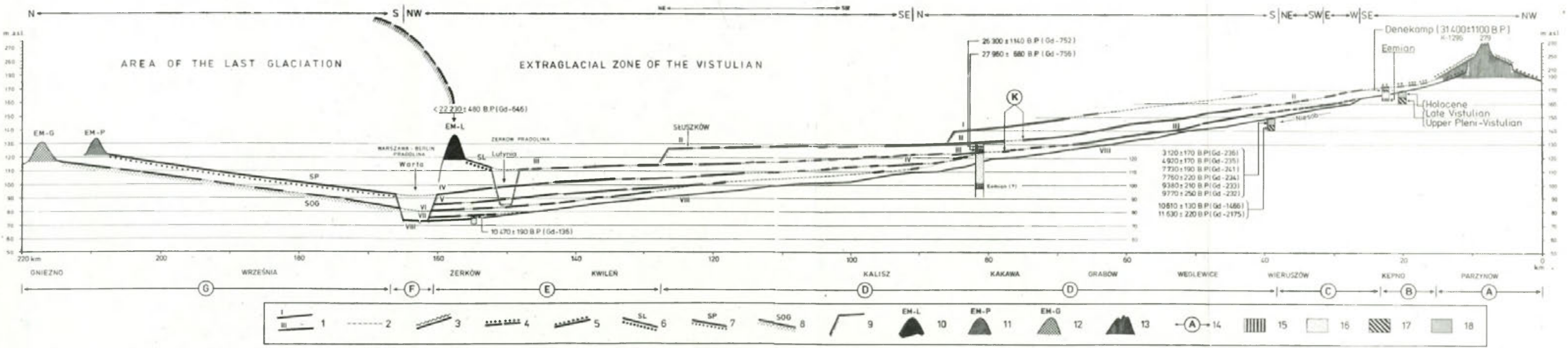

Fig. 2. Longitudinal profiles of the Prosna terraces and their relation to the outwash plains of the Last Glaciation: 1 — terrace shelves, 2 — lack of terraces, 3 — surface of the older periglacial pediment, 4 — older accumulation surface in the Ostrzeszów Hills foreland, 5 — surface of the younger periglacial pediment, 6 — outwash plain of Leszno Phase, 7 — outwash plain of Poznań Phase, 8 — outwash plain of Gniezno Oscillation, 9 — edge undercutting a terrace, 10 — end moraine of Leszno Phase, 11 — end moraine of Poznań Phase, 12 — end moraine of Gniezno Oscillation, 13 — end moraine of Warta Stage (Saalian), 14 — different parts of the Wielkopolska Region: A — Ostrzeszów Hills, B — Kępno Basin, C — Niesob River Valley, D — Prosna River Valley, E — Pyzdry Basin, F — Warszawa—Berlin Pradolina, G — Gniezno—Morainic Plateau, 15 — Eemian peats, 16 — silts and very fine sands of the Vistulian age, 17 — Vistulian peats, 18 — Holocene deposits

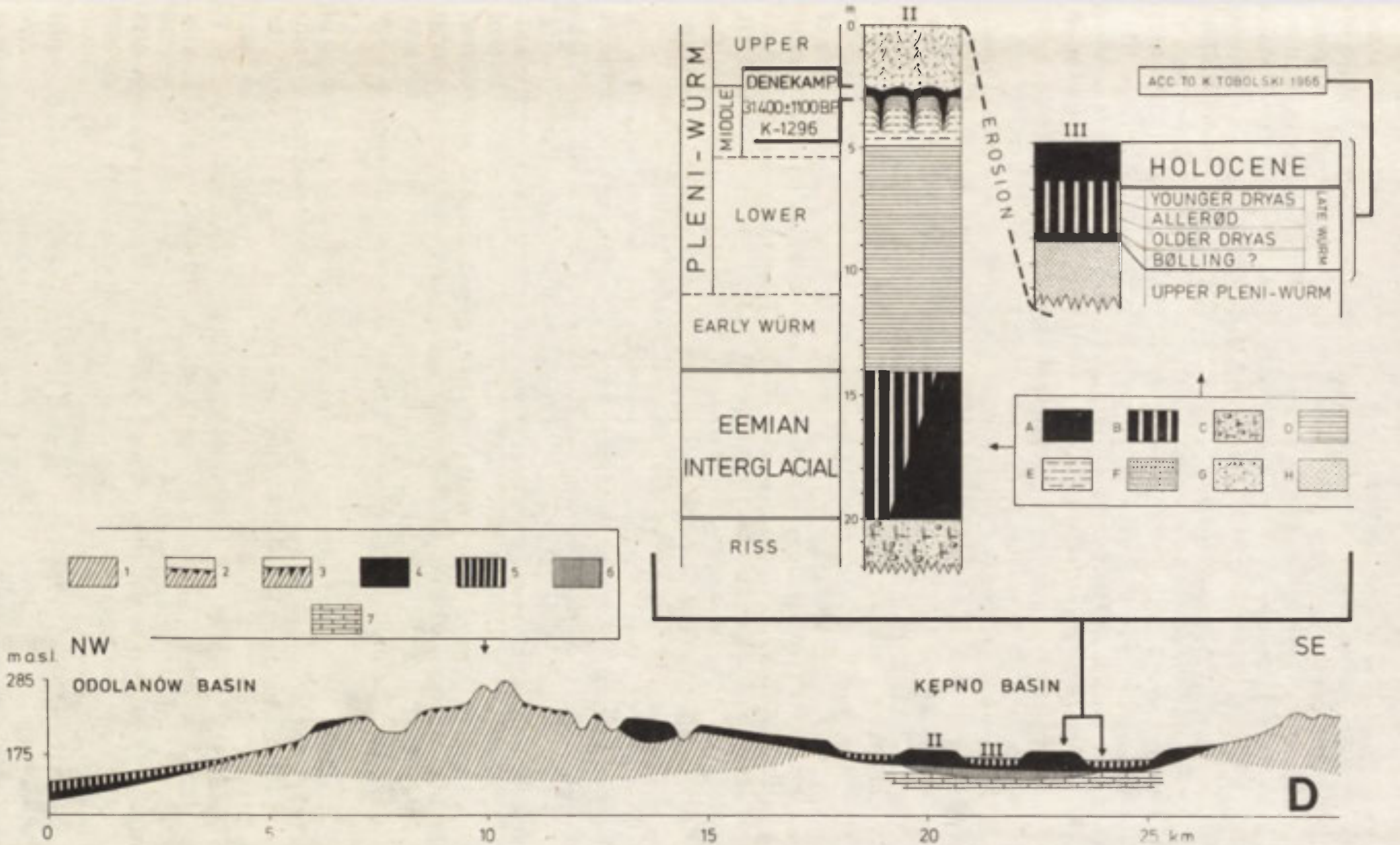

Fig. 4. Stratigraphy of Vistulian deposits in the Kępno Basin (according to Rotnicki 1974b): 1 — glaciotectonically thrusted Neogene and Quaternary deposits, 2 — older periglacial pediment, 3 — younger periglacial pediment, 4 — deposits making up the older surface of periglacial accumulation, 5 — deposits making up the younger surface of periglacial accumulation, 6 — Eem Interglacial organic deposits, 7 — Saalian deposits: A — peats, B — gyttjas and organic silts, C — glacial till, D — grey lacustrine silts, E — sand-rich silts, F — stratified sands, G — various-grained sands with single gravely grains — secondarily disturbed by cryostatic structures, H — medium-grained sands; II — level connected with terrace II of the Prosna, III — level connected with terrace III of the Prosna

Near Węglewice, in the same deposits series, at a depth of 9—10 m from the surface of terrace II, there appears another organogenic layer dated back to 35000—37000 years BP. The similarity of deposits of the Sobiesęki profile to the stratigraphy of deposits of the Kępno and Domasłów Basins, as well as the similarity of the morphological situations of these series, suggest that the two bottom organic layers: peats and gyttjas with the mollusc fauna in the Sobiesęki profile represent the Eemian Interglacial, while the whole silt-sand series with a few organogenic intercalations represents a more or less continuous sedimentation from the beginning of the Lower Vistulian till the Upper Pleni-Vistulian. The palinological analysis of one of the silt layers of the upper part of deposits of terrace II, made once by Tobolski (Rotnicki and Tobolski 1965b), shows that the deposits of this terrace in its upper half were accumulated under a cold climate. The spectrum contains as much as 76% NAP, with the mosses *Drepanocladus exannulatus* and *Scorpioidum scorpioides* dominant. The question of the age of the top deposits of this series lying above an organogenic layer aged 26300 years will be discussed a bit further on.

Terrace III. It is well preserved along the whole course of the middle and lower Prosna; its width, however, is considerably smaller than that of the terrace II. In the Grabów Basin it appears as a strip 3 km wide along both banks of the Prosna. Near Wieruszów it reaches 156 m a.s.l., falling to the north to 130 m near Grabów, 119—123 m near Kalisz, and 110 m in the Pyzdry Basin. In the western part of this basin, in the initial section of the Żerków-Rydzyna Pradolina, terrace III joins the Żerków outwash plain of the Leszno Phase at 110 m a.s.l. (Fig. 2 and 7). The structure of this terrace is diversified. It is an erosion-accumulation type of terrace. Its accumulation cover is formed by fine-grained, finely laminated sands with clear traces of eolization, as at e.g. Jedlec, Macew or Jastrzębniki. The thickness of river deposits of terrace III ranges from 3 to 8 m. The terrace deposits rest on an older substratum, i.e. on the deposits of terrace II in the Grabów Basin or on glacial tills and Tertiary in the gorge section of the Prosna valley between Kalisz and the Pyzdry Basin.

Terraces IV, V and VI. They only appear north of the Grabów Basin (Fig. 2). Terrace IV appears for the first time at the confluence of the Ołobok with the Prosna. It lies there at 120 m a.s.l. It accompanies the lower Prosna valley in the form of a shelf to spread in a fan-like fashion in the area of the Prosna alluvial fan in the Pyzdry Basin, where it joins the outwash plain of the Poznań Phase at the altitude of 92—94 m a.s.l. Terrace V appears even farther north, in Kalisz. It joins the Pyzdry Basin at the level of 85—86 m a.s.l. Terrace VI accompanies the Prosna only along 40 km of its lower course. It joins the outwash terrace of the Gniezno Oscillation in the Warszawa—Berlin Pradolina at 80—81 m a.s.l. (Figs. 2 and 7). The accumulation covers of these terraces, built of sands that are getting coarser on each lower terrace, have a thickness of a few metres.

Terraces III, IV, V, and VI form one system. It follows from the spatial situation of the terraces and their altitudes that terrace III undergoes multiple devergence towards the mouth of the Prosna river in the transition zone from the Grabów Basin to the narrow valley near Kalisz. Near Olobok terrace IV separates from III and then, near Kalisz, terrace V separates from IV; terrace V splits into V and VI near Rokutów. It follows from the relation of the terraces of this system to outwash plains of the Last Glaciation that they are connected with the stay of the last inland ice in the central Wielkopolska, and more precisely with its maximum extent and the first recession phases: Poznań and post-Poznań, and with the Gniezno Oscillation. Terrace III corresponds to this system in the middle course of the river.

Terrace VII and VIII. They form another terrace system. Terrace VII occurs only at the confluence of the Prosna and the Warta, in the fork of the two rivers, because it separates from a bottom terrace. In the top of terrace VII, under dunes, there is a peat layer whose age is 10470+190 years BP (Gd—136), which means that the

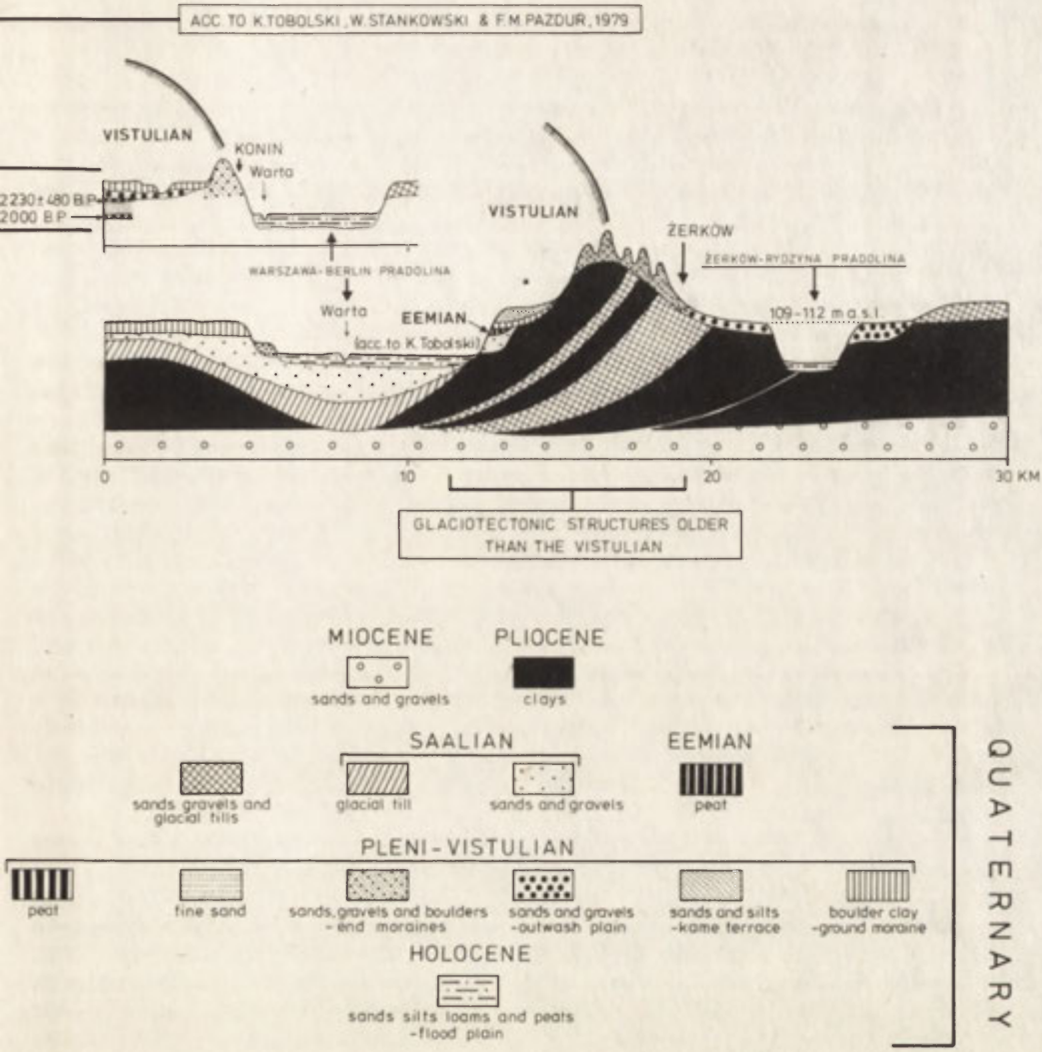

Fig. 7. Maximum extent of the Vistulian Glaciation in the Pyzdry Basin and the zone of contact of fluvioperiglacial waters with glaciofluvial waters of the Last Glaciation (main cross-section according to Rotnicki, 1975)

deposits of this terrace are older (Nowaczyk and Rotnicki 1972; Rotnicki 1972). Since there are deposits from the Alleröd in the bottom of deposits filling the oldest palaeomeanders on the terrace VIII, it follows that the deposits of this terrace are inserted in a dessection coming from at least the Late Vistulian. Thus, terrace VII must have formed in the Upper Pleni-Vistulian, possibly in the Pomeranian Phase. Terrace VIII deposits formed since the Late Vistulian throughout the whole Holocene, as evidenced by palaeomeanders on the surface of the flood terrace differing, though Holocene, in age (Fig. 2).

THE PROBLEM OF THE AGE OF THE EROSION PHASE PRECEDING THE MAXIMUM EXTENT OF THE LAST GLACIATION

In the whole Prosna drainage basin there are distinct traces of heavy deep erosion operating after the formation of terrace II and before that of terrace III; thus, it preceded the Leszno Phase. After the Sobiesęki locality has been discovered, it is possible to determine the stratigraphic position of this erosion phase more precisely. Taking into consideration the youngest radiocarbon date in the top of terrace II deposits (26300 years BP) as well as the fact that this layer is covered with sand deposits 3.5 — 5 m thick, it can be assumed that the aggradation of this terrace was finished about 23000 — 24000 years BP. Considering also the fact that terrace III functioned during the maximum extent of the last inland ice in the Leszno Phase, i.e. about 18000 years BP, it turnes out that the deep erosion which caused the dissection of terrace II operated in a period directly preceding the Leszno Phase. Thus the phase of erosion falls between 23000 — 24000 and 18000 — 20000 years BP.

From the theoretical point of view, the occurrence of an erosion phase at that moment seems to be obvious and understandable. The development of the ice sheet of the Last Glaciation was extremely rapid. Recent results show that in southern Sweden, in Scania, inland ice could have appeared only after 21300 years BP (Berglund and Lagerlund 1981). The development of such a large ice sheet, with a thickness of up to 3.5 km in the centre (Bloom 1971), in such a short time, could not have taken place under a cold continental climate which is assumed for almost the whole Pleni-Vistulian. The development of an ice sheet requires the appearance of the oceanic version of a cool climate (Majdanowski 1958) marked by an increase in precipitation, and hence in humidity. It is hard to suppose that this increase in humidity, certainly caused by a change in the atmospheric circulation in Europe, was merely local and embraced only the area of the centre of the prospective ice sheet. It seems highly probable that this humidification of climate must have felt in adjacent areas, among others on the North European Plain.

Thus far, this phase of erosion conditioned by increased humidity of climate during the growth of the Scandinavian inland ice in the Upper Pleni-Vistulian has not been generally noticed. It is true that some scholars do find traces of one or two phases of erosion in the Vistulian; however, they associate them with different amelioration periods and not with the period of the growth of the last inland ice (Klatkowa 1965; Różycki 1961; Jahn 1956a; Gilewska 1963; Mycielska — Dowgiałło 1967, 1977; Starkel 1982). Earlier, only the phase of erosion in Ostrzeszów Hills (thrust end moraine of the Warta Stage) and the Prosna drainage basin was associated with the growth of the last inland ice (Rotnicki 1966). At that time, though, it was situated in the Paudorf, but it should be kept in mind that the Paudorf was then placed 25000 — 28000 years BP, i.e. in the period immediately preceding the inland ice transgression of the Leszno Phase maximum extent of the Last Glaciation. At that time, the only argument for this stratigraphic position of the erosion phase — apart from the theoretical premises — was that it preceded the formation of terrace III coming from the Leszno Phase.

The radiocarbon dates from the top organogenic layer of the Sobiesęki profile prove the position of intensive deep erosion between 23000 — 24000 and 18000 — 20000 years BP. Only a few authors have taken notice of the humidification of climate at that time (Ivanowa 1969; Geyh and Rodhe 1972; Flohn 1974; Lamb 1974; Andrews and Mahaffy 1976; Andrews, Barry, Bradley, Miller and Williams 1972). Thus far, this humidification phase has not been detected by biostratigraphic methods, either. It is not unlikely that the first biostratigraphic trace of this phase in the Maliniec II layer near Konin, dated to 22300 ± 480 years BP (Stankowski 1979; Tobolski 1979; Pazdur and Walanus 1979). Duphorn (1976) has indicated slight amelioration at the age of this layer on the curve of

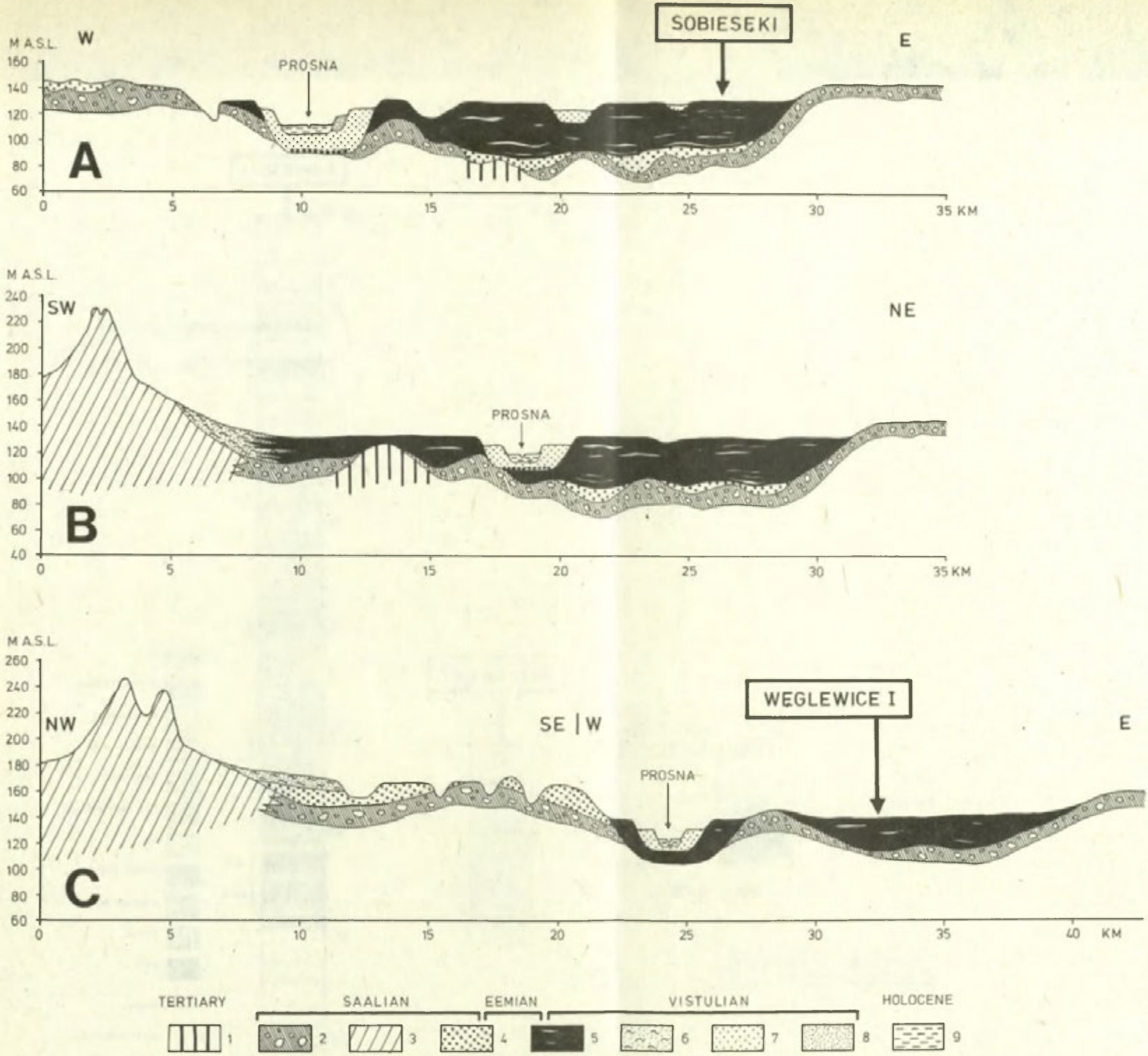

Fig. 5. Deposits filling the Grabów Basin. Tertiary: 1 — Tertiary clays partially glaciotectonically thrusted; Saalian: 2 — glacial till, 3 — glaciotectonically thrusted Tertiary and Quaternary deposits, 4a — glaciofluvial and fluvial sands and gravels; Eemian: 4b — fluvial sands, 5a — peats, gyttjas and silts; Vistulian: 5b — grey-coloured silts and very fine-grained sands with thin peat interbeddings, 6 — periglacial slope deposits, 7 — sands and silts of terrace III (Leszno Phase), 8 — sands of terraces IV—VII (Phases: Poznań, post-Poznań, Gniezno Oscillation, and Pomeranian); Holocene: 9 — sands and muds of flood terrace beginning of accumulation in Late Vistulian, Sobieseki and Węglewice I — location of litho stratigraphic profiles presented in Fig. 6

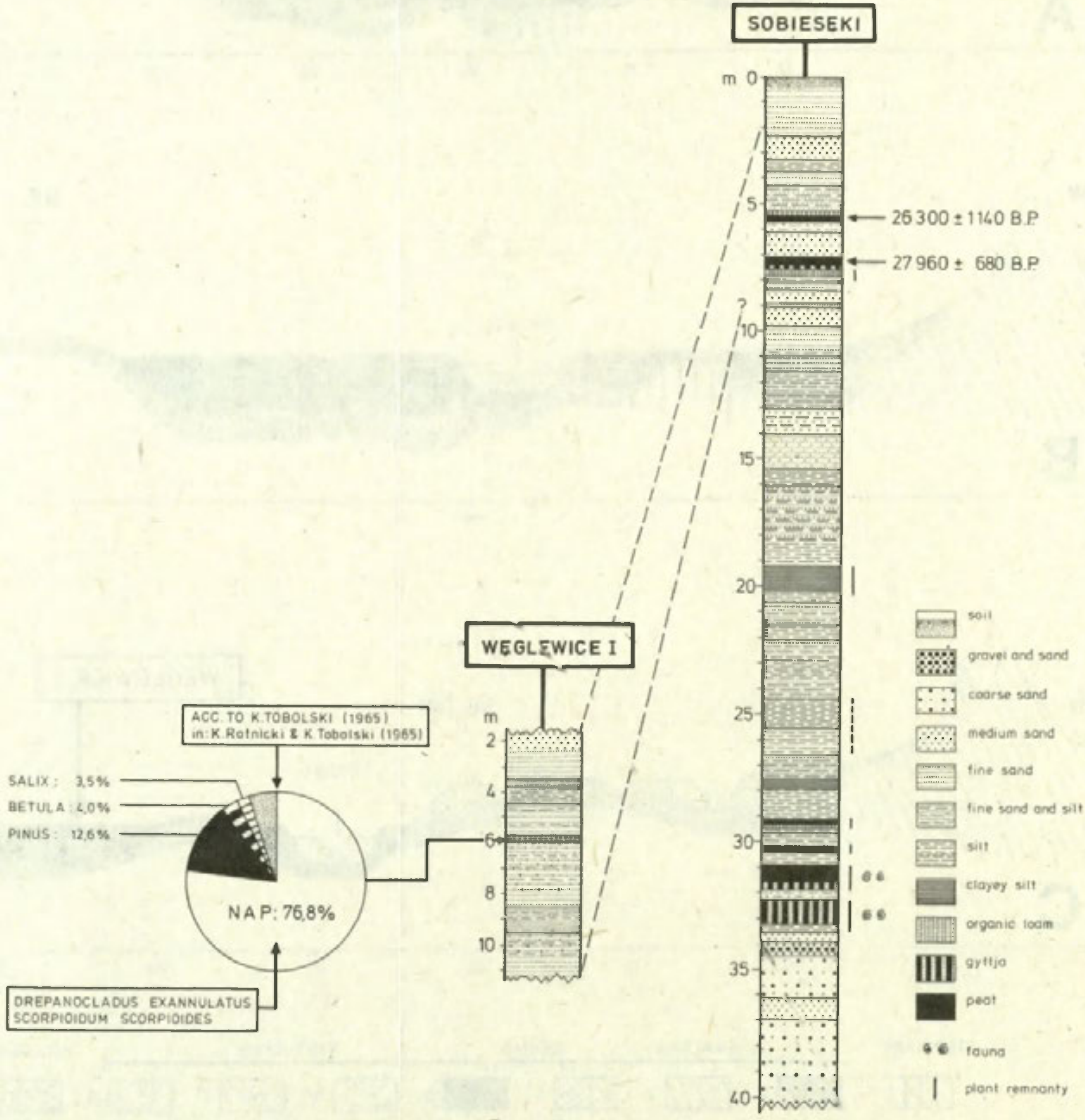

Fig. 6. Stratigraphy of Vistulian deposits in terrace II of the Prosna in the Grabów Basin

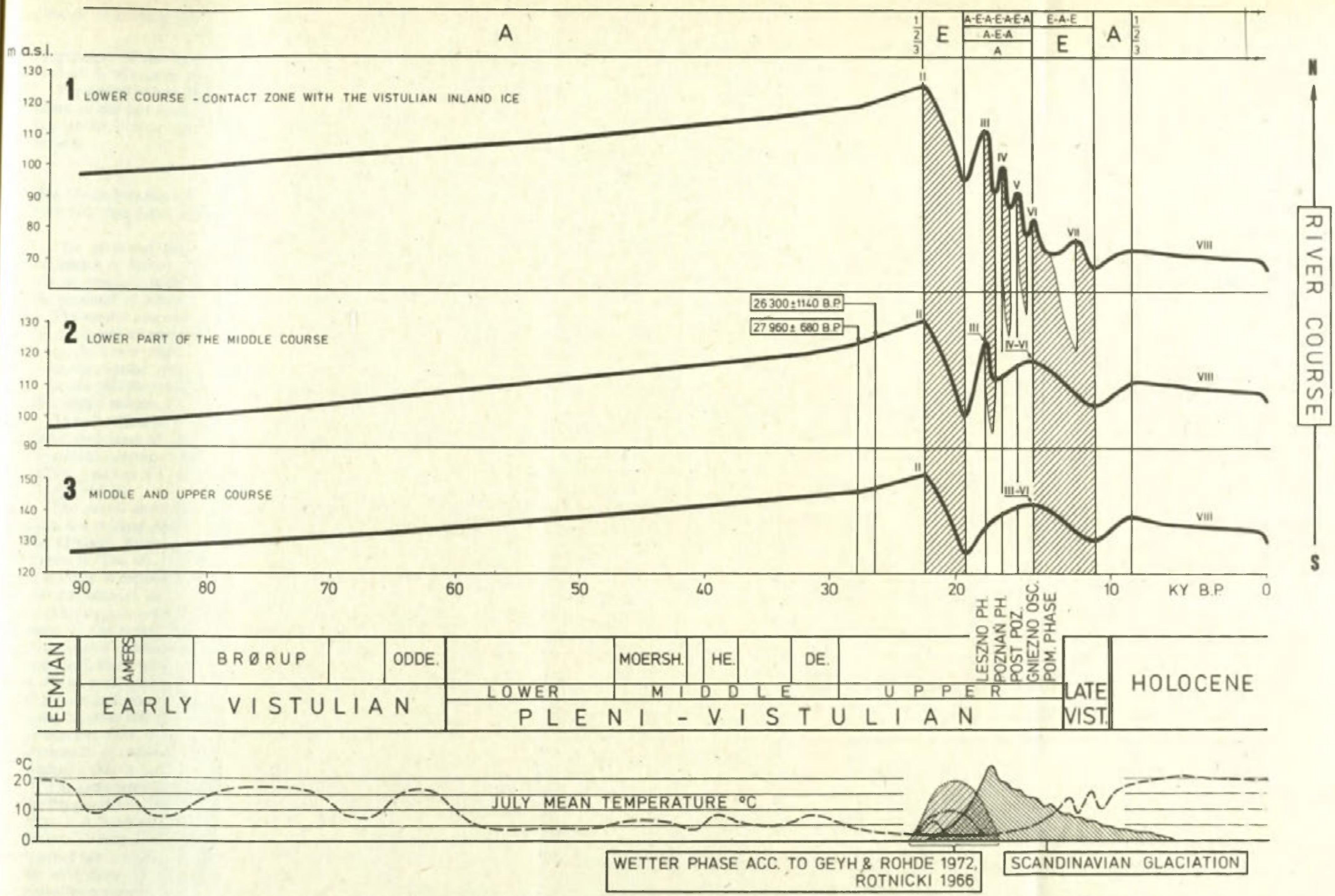

Fig. 8. Spatio-temporal model of the main phases of accumulation and erosion in the Prosna valley during the Vistulian; and Holocene dashed line: mean temperature of July (according to van der Hammen et al. 1967, and Duphorn 1976); dot-dash line: mean temperature of July according to Duphorn (1976); double dot-dash line: mean temperature of July according to the author; dotted line, oblique hachure, dots: period of humidification; solid line, oblique hachure: growth and disappearance of the Scandinavian ice sheet

temperature of the Last Cold Period. A few years later the same amelioration was marked by Kozarski (1980).

The period of humidification and deep erosion preceding immediately the maximum extent of the last Scandinavian inland ice during the Upper Pleni-Vistulian stated at first in the Prosna valley (Rotnicki 1966) is called the Prosna Phase by the present author.

THE MAIN PHASES OF EROSION AND ACCUMULATION IN THE PROSNA VALLEY DURING THE LAST GLACIAL-INTERGLACIAL CYCLE

The presented facts allow the construction of a spatio-temporal model of the tendencies of fluvial processes in the drainage basin of an extraglacial river whose activity was controlled by climatic conditions and accidentally by the direct influence of the presence of inland ice in the zone of the valley mouth.

The model was constructed for three cross-sections of the Prosna valley presenting the processes of accumulation (aggradation) and erosion in three zones:

a) the lower course,

b) the border zone between the lower and middle course, and

c) the middle course.

The upper section of the valley is excluded from considerations for the present.

The situation of points determining the course of the curves of the tendencies of fluvial processes in each of the three mentioned zones is defined in the Cartesian co-ordinate system by the abscissae of age and the ordinates of the altitudes of the bottom and top of a given deposit series forming a definite terrace surface in the Prosna valley (Fig. 8).

The curves show that from the Eemian Interglacial until 23000 – 24000 years BP there is a marked aggradation tendency in the analysed section of the Prosna valley. In the Grabów Basin it is an accumulation of a thicker series of deposits building terrace II. This 40 – metre series lies in an extensive fossil valley lowering, also marked in the way of deposition of the glacial till of the Central Polish Glaciation (Saalian) and the glaciofluvial sands and gravels of the same glaciation that cover it (Fig. 5).

The accumulation of the discussed series starts with the formation of gyttjas and peats. In the Kępno and Domaslów Basins these are Eemian peats (Fig. 4). A similar stratigraphic position is occupied by peats and gyttjas in the Grabów Basin at Sobieskęki. Here, too, they occur in the bottom of deposits forming terrace II. Therefore, it can be presumed that they also formed in the Eemian Interglacial. The accumulation of the silt and silt-sand series lasted till about 23000 – 24000 years BP. It is hard to tell today whether this aggradation tendency was continuous or interrupted by short phases of erosion. One thing is certain: that throughout this long period, lasting tens of thousands of years, the aggradation tendency was decisively dominant. It is this tendency that is responsible for the existence of such extensive surfaces of terrace II.

The long-term aggradation tendency was suddenly interrupted by deep erosion of the Prosna Phase, clear proofs of which occur in the whole of the Prosna drainage basin (Figs. 2, 4, 5, and 8). There is evidence of increasing transportation competence of the Prosna already in the final phase of sedimentation of the series forming terrace II. A proof can be the change of material from silt to sand in the top of this series, above the level dated to 26300 years BP. This phase of erosion, the Prosna Phase, is most probably connected with the growth and transgression of the Scandinavian inland ice, which has been discussed in the previous section of this paper.

In the succeeding period, i.e. during the maximum transgression and then the recession of the last inland ice in the Upper Pleni-Vistulian, the main tendency of fluvial activity was differentiated spatially (Fig. 2 and 8). In the zone of the contact of the

Prosna with glaciofluvial waters in the Pyzdry Basin, several short periods of erosion alternated with those of stabilization and accumulation. The short stops of the ice margin of the Leszno, Poznań and post-Poznań Phases, and the Gniezno Oscillation were the periods of outwash aggradation. The outwash aggradation from the north brough about the aggradation of the floor of the lower Prosna valley in the Pyzdry Basin. It should also be kept in mind that in the extraglacial zone, in which the whole of the Prosna valley was situated, a cold climate still prevailed with permafrost in the substratum — conditions favouring heavy aggradation of the valley floors. Phases of inland ice recession were indicated by the uncovering of the erosion basis lying lower and lower, marked by the incision of the Prosna in the lower section of the valley. At the same time that the system of terrace III to VI developed in the lower part of the valley — as a result of four alternating phases of accumulation and erosion, in middle section the aggradation of the floor at the terrace III level took place. It was conditioned by the climate, which has been pointed out above. This Upper Pleni-Vistulian aggradation of valley bottoms took place throughout the periglacial zone except in those areas in which local factors forced another tendency.

It follows from the above that in one valley, at a relatively short distance from each other, we witnessed two opposing tendencies at the same time: generally, erosion in the lower section of the valley and aggradation in the middle one. There is a transition zone between the zones with the opposing tendencies in which only one short phase of erosion took place separating two local phases of accumulation (terraces III and VI in Fig. 8, diagram 2).

At the decline of the Last Cold Period there again appeared an erosive tendency in the whole of the valley, interrupted only temporarily by a short phase of local aggradation in the very mouth of the Prosna, as a result of which terrace VII was formed. This erosive tendency was stopped basically in the Late Vistulian, at least in the middle course of the valley, since the oldest attested deposits filling the Prosna palaeomeanders occurring at the valley floor in terrace VIII come from the Alleröd.

CONCLUSIONS

The spatio-temporal model of the main tendencies in the fluvial processes in the Prosna valley shows how many factors influence these tendencies and how difficult it is to talk of a simple, universal pattern of the phases of aggradation and erosion at a scale of a larger area, if no such pattern can be applied even to one, relatively small valley.

If, however, we were to reject the influence of local factors differentiating the history of erosion and aggradation phases, we could distinguish two periods in the Prosna valley with diametrically different tendencies:

1. The first period of accumulation and the dominance of the aggradation tendency. It started already in the Last Interglacial and lasted till the rapid growth of the last Scandinavian inland ice.

2. The second period, of the dominance of the erosive tendency. It started with the humidification of the climate followed by the appearance of an extensive ice sheet. This period left its marks both in the lower section of the valley in which fluvial processes were under the direct influence of the ice, as well as in the middle section affected by the climatic condition responsible for the formation of the ice sheet. Depending on the immediate cause, the intensity of this tendency varies. It is readily visible in the successive diagrams of the model (Fig. 8). Anyway, this erosive tendency, which was the strongest during the inland ice recession, became clearly weaker towards the end of the last period. In the Holocene the impact of this tendency is only minimal. Generally, we can speak of a Holocene vertical stabilization of the Prosna valley floor, with the

exception of the latest period conditioned by man's economic activity. There is an increase in the erosive tendency in this period (Fig. 8).

Thus, we can see, the course and stratigraphic position of the main phases of fluvial aggradation and erosion are closest, generally of course, to the picture presented by Schumm (1965), a major exception being the fact that in the Prosna valley the phase of the dominance of the erosive tendency begins much earlier — already during the rapid growth of the last inland ice.

REFERENCES

Andrews J. T., Barry R. G., Bradley R. S., Miller G. H., Williams L. D., 1972, Past and present glaciological responses to climate in eastern Baffin Island, *Quaternary Research* 2, 303—314.

Andrews J. T., Mahaffy M. A. W., 1976, Growth rate of the Laurentide ice sheet and sea level lowering with emphasis on the 115,000 BP sea level low, *Quaternary Research* 6, 167—183.

Berglund B. E., Lagerlund E., 1981, Eemian and Weichselian stratigraphy in South Sweden, *Boreas* 10, 323—362.

Bloom A. L., 1971, Glacial-eustatic and isostatic controls of sea level since last glaciation, in: K. K., Turekian, ed, *The Late Cenozoic glacial ages*, Yale University Press, New Haven and London, 355—380.

Duphorn K., 1976, Kommt eine neue Eiszeit?, *Geologische Rundschau* 65, 3, 845—864.

Flohn H., 1974, Background of a geophysical model of the initiation of the next glaciation, Quaternary Research 4, 385—404.

Galon R., 1968, Ewolucja sieci rzecznej na przedpolu zanikającego lądolodu (Sum.: Evolution of fluvial system in foreland of retreating ice sheet), in: R. Galon, ed, *Ostatnie zlodowacenie skandynawskie w Polsce*, Prace Geograficzne IG PAN 74, Warszawa, 101—120.

Geyh M. A., Rodhe P., 1972, Weichselian chronostratigraphy, C^{14} datings and statistics, 24th International Geological Congress, Montreal, 12, 27—36.

Gilewska S., 1963, *Rzeźba progu środkowotriasowego w okolicy Będzina*, Prace Geograficzne IG PAN 44, Warszawa.

Hammen T. van den, Maarleveld G. C., Vogel J. C., Zagwijn W. H., 1967, Stratigraphy, climatic succession and radiocarbon dating of the last glacial in the Netherlands, *Geologie en Mijnbouw* 46, Maart, 75—95.

Ivanova I. K., 1969, Les loess de la partie Sud-Ouest du territorie Europeen de l'URSS et leur stratigraphie, in: *La Stratigraphie des loess d'Europe*, Paris, 151—159.

Jahn A., 1956a, *Wyżyna Lubelska. Rzeźba i czwartorzęd* (Sum.: Lublin Upland — Relief and Quaternary), PWN, Warszawa.

Jahn A., 1956b, The action of rivers during the Glacial Epoch and the stratigraphic significance of fossil erosion surfaces in Quaternary deposits, *Przegląd Geograficzny* 28, Suppl., 101—104.

Klatkowa H., 1965, Niecki i doliny denudacyjne w okolicach Łodzi, *Acta Geographica Lodziensia* 19, Łódź.

Kozarski S., 1980, An outline of Vistulian stratigraphy and chronology of the Great Poland Lowland, *Quaternary Studies in Poland* 2, 21—35.

Kozarski S., Rotnicki K., 1977, Valley floors and changes of river channel patterns in the North Polish Plain during the Late Würm and Holocene, *Quaestiones Geographicae* 4, 51—93.

Kozarski S., Rotnicki K., 1978, *Problemy późnowürmskiego i holoceńskiego rozwoju den dolinnych na Niżu Polskim*, PTPN, Prace Komisji Geograficzno-Geologicznej 19.

Kozarski S., Rotnicki K., 1983, Changes of river channel patterns and the mechanism of valley-floor construction in the North Polish Plain during the Late Weichsel and Holocene, in: D. J. Briggs, R. S. Waters, eds, *Studies in Quaternary geomorphology*, Geo Books, University Press, Cambridge, 31—48.

Lamb H. H., 1974, Atmospheric circulation during the onset and maximum development of the

Wisconsin/Würm ice age, in: Y. Herman, ed., *Marin geology and oceanography of the Arctic seas*, Springer Verlag, New York, 349—358.

Majdanowski S., 1958, Zagadnienia klimatyczne okresów wydmowych w świetle glacjalnych i postglacjalnych zmian ogólnej cyrkulacji atmosferycznej w Europie środkowej (Res.: Les problémes climatiques des périodes de dunes par rapport aux variation glaciers et postglaciaires de la circulation atmosphérique générale dans l'Europe Centrale), in: R. Galon, ed., *Wydmy śródlądowe Polski*, PWN, Warszawa, 33—51.

Maruszczak H., 1968, Przebieg zjawisk w strefie peryglacjalnej ostatniego zlodowacenia w Polsce (Sum.: The course of the phenomena in the periglacial zone during the Last Glaciation). in: R. Galon. ed., *Ostatnie zlodowacenie skandynawskie w Polsce*, Prace Geograficzne IG PAN 74, 157—200.

Mycielska—Dowgiałło E., 1967, Formy szczelinowe i inwolucyjne w piaskach i żwirach doliny Wisły koło Tarnobrzega (Sum.: Frost fissures and involutions in sands and gravels of the Vistula valley near Tarnobrzeg), *Biuletyn Peryglacjalny* 16, 203—215.

Mycielska—Dowgiałło E., 1977, Channel pattern changes during the Last Glaciation and Holocene, in the Northern Part of Sandomierz Basin and the Middle Part of the Vistula Valley, Poland, in: K. J. Gregory, ed., *River channel changes*, Wiley, Chichester, 75—87.

Nowaczyk B., Rotnicki K., 1972, Wiek wydmy parabolicznej w Łąkach Pyzdrskich w świetle wyników analizy C—14 osadów organogenicznych występujących w spągu osadów eolicznych (Age of the parabolic dune at Łąka Pyzdrska based on radiocarbon analysis, Central part of the Greater Poland), *Sprawozdania PTPN* 2/90, 91—93.

Pazdur M. F., Walanus A., 1979, The Konin-Maliniec site: age assessment by radiocarbon method, in: *Guidebook of Excursion. Symposium on Vistulian Stratigraphy, Poland 1979*, Warszawa, 4.5—4.6.

Penck A., 1938, Das Klima der Eiszeit, *Verhandlungen der III Internazionalen Quartär Konferenz, 1936*, Wien.

Rotnicki K., 1966, *Rzeźba Wzgórz Ostrzeszowskich jako rezultat rozwoju stoku podczas Würmu* (Sum.: The relief of the Ostrzeszów Hills as a result of slope development during the Würm), PTPN, Prace Komisji Geograficzno-Geologicznej 5, 2.

Rotnicki K., 1972, Wiek najniższych poziomów terasowych doliny dolnej Prosny w Kotlinie Pyzdrskiej... (Sum.: Age of lower terraces of the Prosna river in the Pyzdry Basin...), *Sprawozdania PTPN* 2/90, 107—108.

Rotnicki K., 1974a, Pozycja chronologiczna faz erozji w dolinie Prosny po okresie maksymalnego zasięgu zlodowacenia bałtyckiego (The chronological position of erosion phases in the valley of the Prosna after the period of maximum extent of the Last Glaciation), Krajowe Sympozjum „Rozwój den dolinnych...", Wrocław—Poznań 1974, *Streszczenie referatów i komunikatów*, 37—41.

Rotnicki K., 1974b, Slope deveiopment of Riss Glaciation end moraines during the Würm; its morphological and geological consequences, *Quaestiones Geographicae* 1, 109—139.

Rotnicki K., 1975, Stratigraphic evidences of the survival of Riss glaciotectonic structures and forms in the marginal zone of the maximum extent of the last glaciation, *Quaestiones Geographicae* 2, 113—143.

Rotnicki K., 1985, Postglaciotectonic surface as a basis of analysis of fluvial relief development in areas of glacial deposition, *Quaestiones Geographicae*, Special Issue 1, 147—158.

Rotnicki K., 1986, Paleomorfologia północnej części Kotliny Grabowskiej ze schyłku zlodowacenia środkowopolskiego, południowo-wschodnia Wielkopolska (Palaeomorphology of the northern part of the Grabów Basin, the closing phases of the Middle-Polish Glaciation, south-eastern Greater Poland), *Sprawozdania PTPN za 1985 r.*

Rotnicki K., Tobolski K., 1965a, Stanowisko interglacjału eemskiego w Domasłowie pod Kępnem, wiadomość wstępna (Sum.: Site of the Eemian Interglacial at Domasłów near Kępno, preliminary report), *Badania Fizjograficzne nad Polską Zachodnią* 15, 177—186.

Rotnicki K., Tobolski K., 1965b, Pseudomorfozy wieloboków z lodem szczelinowym i stanowisko tundry w peryglacjalnym basenie sedymentacyjnym ostatniego zlodowacenia w Kępnie (Sum.:

Pseudomorphoses on the fissure ice polygons and the locality of tundra in the periglacial sedimentation basin of the Last Glaciation at Kępno), *Badania Fizjograficzne nad Polską Zachodnią* 15, 93–137.

Rotnicki K., Tobolski K., 1969, Interstadiał Paudorf w Kępnie (Sum.: Paudorf Interstadial in Kępno, south-east Greater Poland), *Badania Fizjograficzne nad Polską Zachodnią* 23, Ser. A, 119–127.

Rotnicki K., Wiśniewski J., 1986, Litostratygrafia osadów wypełniających północną część Kotliny Grabowskiej (Lithostratigraphy of Vistulian sedimentary infillings of the northern part of the Grabów Basin, south-eastern part of Greater Poland), *Sprawozdania PTPN za 1985 r.*

Różycki S. Z., 1961, Middle Poland, in: *Guidebook of Excursion "From Baltic to the Tatras"*, Part II, 1, VIth INQUA Congress, Warszawa, 5–116.

Różycki S. Z., 1972, *Plejstocen Polski Środkowej* (The Pleistocene of Middle Poland), 2nd ed., Warszawa.

Schumm S. A., 1965, *Quaternary palaeohydrology, The Quaternary of the United States. A review volume for the VII Congress INQUA*, Princeton, New Jersey, 783–794.

Soergel W., 1921, *Die Ursachen der diluvialen Aufschotterung und Erosion*, Berlin.

Stankowski W., 1979, An outline of geological structure and geomorphology of the Konin surroundings, in: *Guidebook of Excursion. Symposium on Vistulian Stratigraphy, Poland 1979*, Warszawa, 4.1–4.4.

Starkel L., 1977, The palaeogeography of mid- and east Europe during the last cold stage, with west European comparisons, *Philosophical Transactions, Royal Society*, London, B, 280, 351–372.

Starkel L., 1982, ed., *Evolution of the Vistula River Valley during the last 15,000 years*, Part 1, Geographical Studies, Special Issue 1, Wrocław–Warszawa.

Starkel L., 1983, The reflection of the hydrological changes in the fluvial environment of the temperate zone during the last 15,000 years, in: K. J. Gregory, ed., *Background to Palaeohydrology*, Wiley, Chichester, 213–235.

Tobolski K., 1966, Późnoglacjalna i holoceńska historia roślinności na obszarze wydmowym w dolinie środkowej Prosny (Sum.: The Late-Glacial and Holocene history of vegetation in the dune area of the middle Prosna valley), *Prace Komisji Biologicznej PTPN* 23, 1, 1–69.

Tobolski K., 1979, Fossil flora in the Maliniec site, in: *Guidebook of Excursion, Symposium on Vistulian Stratigraphy, Poland 1979*, Warszawa, 4.3–4.5

Trevisan L., 1949, Genese des terrasses fluvialesti en relation avec les cycles climatiques, in: *Comptes Rendus du Congres International de Geographie*, 2, Lisbone 1949, 344–350.

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THE SŁUPIA VALLEY IN THE VICINITY OF SŁUPSK TOWARDS THE CLOSE OF THE VISTULIAN AND IN THE HOLOCENE

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INTRODUCTION

In the investigations of the post-glacial development of river valleys and of the evolution of their channels the rivers flowing from the Pomeranian area directly into the Baltic Sea were generally left out. It does not mean, however, that there are no studies concerning the geological structure and the geomorphology of river valleys in this area. But most of those studies have a general character (cf. Deecke 1911; Bülow 1930; Keilhack 1930; Sylwestrzak 1973, 1978a, b; Piasecki 1976, 1982). The common feature of those studies is that they are but little supported by field investigations. B. Rosa (1964) has made an attempt at a comprehensive solution of the problem of the geological structure of littoral river valleys; making use of geological archivals and of his own observation he has presented his views on the development of littoral river valleys and the problems to be solved. He has identified rather large sections of accumulation in the valley bottoms of littoral rivers (50 km for the Slupia, 13 km for the Łupawa), as well as the effect of the sea level fluctuations on the development of organic covers in the valley bottoms.

In their studies of the Radunia valley Rachocki (1973, 1974, 1981) and Koutaniemi and Rachocki (1981) have noticed the role of late-glacial lakes in the development of river valleys and they have afforded important information concerning the features of fluvial deposits, the rate of fluvial processes and the effect of hydrotechnical interference in the river channel on the course and rate of those processes.

The Pleistocene developmental stage of the catchment area of the Slupia and its valley has been lately the subject of Mojski and Orlowski's (1978) and of Orlowski's (1981, 1983) studies. They have been based on the existing geological material and on a detailed analysis of morphologic conditions.

It seems, however, that the present state of knowledge on the postglacial stage of littoral river valleys development has been correctly evaluated by Kondracki (1978) and Marsz (1984) who have indicated that, up to now, the development of no littoral river valley has been studied.

In 1981–1985 at the Department of Geography of the High School of Pedagogy, Słupsk, the present author supervised work on the following subject: "The evolution and mechanism of transformation of the valley bottoms of the Słupia and the Łupawa". It was part of the interdepartmental problem MR I/25: "Transformation of Poland's geographical environment" coordinated by Professor Leszek Starkel. Substantial help in this research was brought by the High School of Pedagogy in Słupsk, which financed a part of the radiocarbon dating and some other specialistic work.

The study was carried out with reference to the program IGCP No 158 A "Palaeohydrologic changes in the temperate zone during the last 15000 years", subproject A: "Fluvial environment". A guide-book for investigations, prepared and published by Starkel and Thornes (1981) has proved to be extremely useful.

During their work the authors taking part in the investigations published or prepared for print a number of studies (Orłowski 1981; E. Florek 1983, in press; E. Florek, W. Florek, in press; W. Florek 1983, in press a, b, c; Alexandrowicz et al., in press). Further studies are being prepared.

AREA AND RANGE OF INVESTIGATIONS

Several sections of the Słupia and the Łupawa valleys were selected for detailed investigation. Those sections were chosen which could afford the fullest record of the processes that have occurred in the river valleys since the moment of the river outflow formation, up to the present time.

The section of the Słupia River situated near Słupsk was most thoroughly investigated. Here the valley is wide and all the morphologic levels have been distinctly developed and, what is more, they have been preserved up to the present.

A preliminary evaluation of geomorphologic conditions was done with the use of air photos and through the analysis of old maps (E. Florek 1983, in press; E. Florek, W. Florek, in press; W. Florek 1984). Owing to this the principal morphologic levels could be distinguished and considerable advantage was obtained in the analysis of the character and age of a great number of paleomeanders preserved on the flood plain. Geological and geomorphological field investigations have brought information on the structure of glacifluvial and fluvial terrace levels and they have revealed the spatial distribution of the deposit facies of different age and genesis.

Radiocarbon dating of samples collected helped to determine the age of organogenic deposits filling the depressions of different morphologic levels. Most datings have been additionally confirmed by the results of palinologic expertise. In one case (profile Słupsk – South) a full palinologic and diatomaceous analysis was done in a boring situated in a depression of an upper floodplain terrace. Several profiles were examined in respect of changes in the composition of the diatom flora and malacofauna. The data of Poland's archeologic map executed at the moment were also used in order. to determine, at certain points, the interaction of human settlements and environmental change. Finally, several hundred analyses were executed of the granulation of deposits and of the rounding degree of quartz grains of three sand fractions. The petrographic composition of the deposits and the composition of heavy minerals in the fine-grained sand fraction were also examined.

All that work has brought important information on the processes which, from the close of the Vistulian and throughout the Holocene, have led to the formation of the framework of the river system in the Pomeranian northern slope, and then modelled its bottom. This concerns not only natural processes but also those conditioned by man's activities.

THE INVESTIGATED AREA IN THE DEGLACIATION PERIOD

From Lubuń to the confluence of the Gnilna the valley of the Słupia River has developed in a subglacial channel (Sylwestrzak 1978a) of considerable depth (reaching ca 130 m below sea level) cutting through Upper-Campanian deposits; the complexity of the deposits which fill the channel evidences its outlasting (Mojski, Orłowski 1978; Orłowski 1981, see Fig. 1 C).

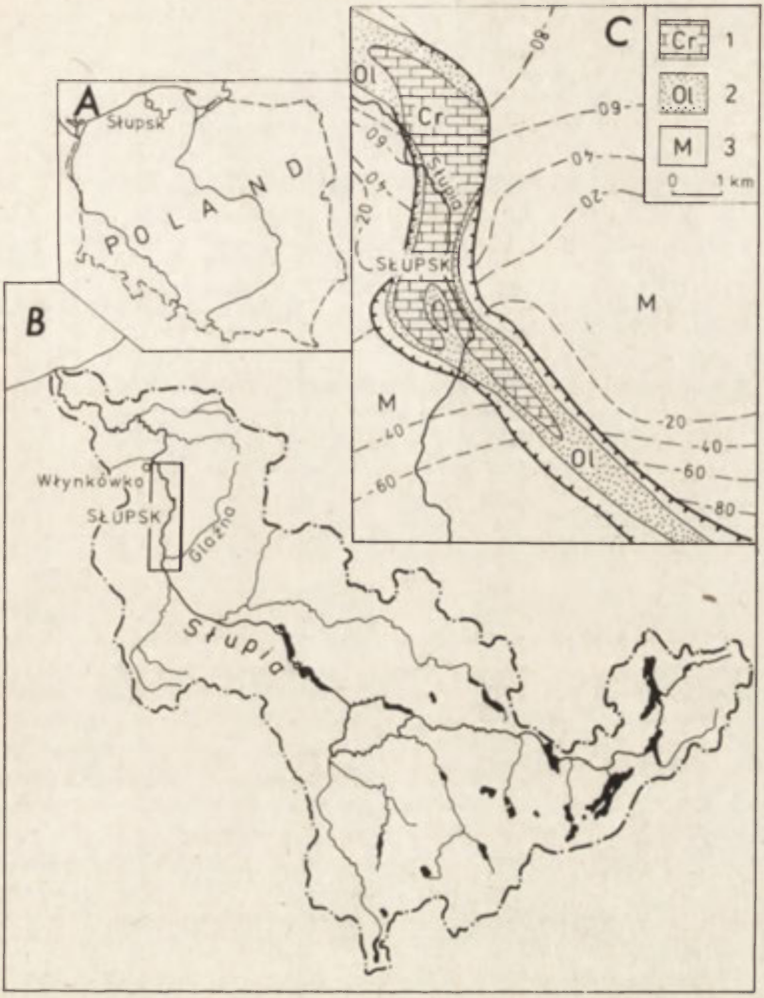

Fig. 1. Location of the study area, A — on Poland's background, B — on the Słupia's river catchment, C — geological sketch of the Sub-Quaternary surface of the Słupsk vicinity (after Mojski and Orłowski, 1978); 1 — Cretaceous sediments, 2 — Oligocene sediments, 3 — Miocene sediments

The highest morphologic level preserved here (45 m a.s.l. in Słupsk) is connected with the draining of water from the thawing continental ice-sheet of the Gardno Phase and with the degradation of blocks remaining from the recession phases previous to the Gardno Phase (Sylwestrzak 1969; Mojski, Orłowski 1978; Orłowski 1981, 1983). In the vicinity of Kwaków, waters flowing from the North joined those flowing from the South and then they flew westwards (Fig. 2).

The water that has formed the lower level of the outwash draining track flew in the same direction (Orłowski 1981, 1983). The remains of the then formed level have survived in Słupsk at 33—35 m a.s.l.

At the close of the Gardno Phase of deglaciation an outflow track was formed in the immediate foreland of the Gardno moraine: it ran westwards through the marginal

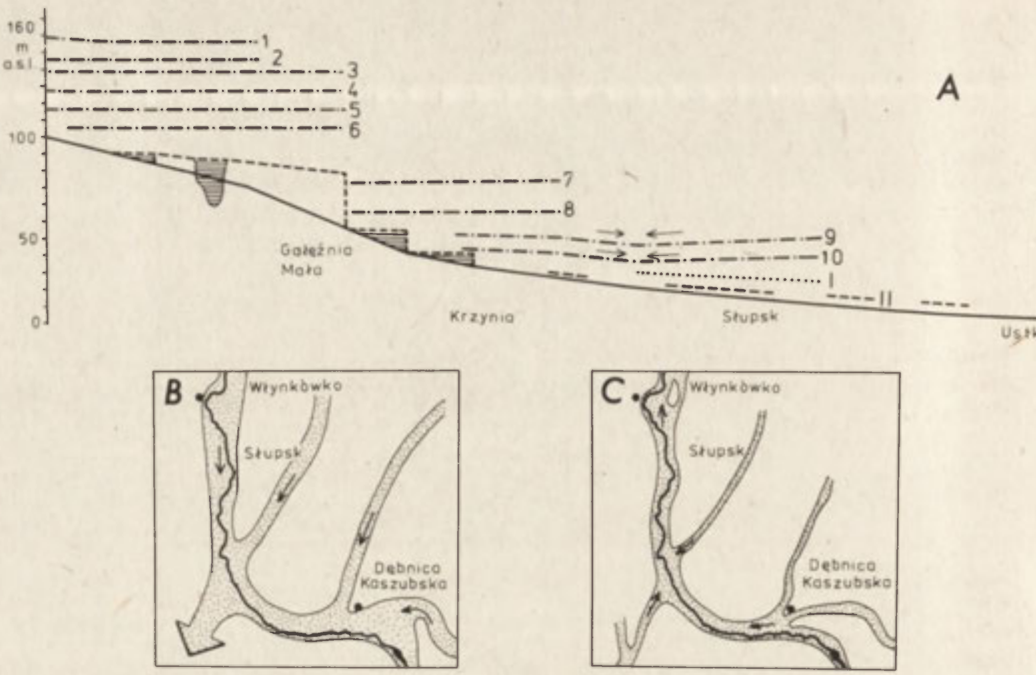

Fig. 2. A — Longitudinal profile of the morphological surfaces within Słupia's valley in section between Żukowskie Lake and the mouth into the Baltic Sea; 1—6 the sandur levels declining to the Miastko vicinity, 7—8 the sandur valley levels connected with the hydrographic junction of the Lętowskie Lake, 9—10 sandur valley levels connected with the hydrographic junction of the Kwakowo region, PP — overflow terrace (I), TN — upper floodplain terrace (II), the floodplain terrace is not marked (extent of the sandur levels after Orłowski, 1981), B — the main directions of outflow in the Gardno Phase, C — the main directions of outlow in the Bölling after breaking of the Gardno Phase marginal forms in Bydlino region

lakes of Wytowno, Modła and Pieńkowo (Giedrojć—Juraha 1949; Sylwestrzak 1969, 1978b; Rosa 1963, 1964, 1968, 1984; Orłowski 1981).

At the same time the morainic forms of the Gardno Phase (?) were broken near Bydlino and the waters flew northwards along the erosion valley of the Słupia. This probably occurred within the level which has been preserved in the southern part of the town, at 22—25 m a.s.l. Its geological structure is very complicated. Glacifluvial rests play here an important part. Investigations carried out by Dawid—Hryniewicz and Hryniewicz (1984) have shown that they are built of fractions over 0φ. Their content exceeds 50 weight per cent, particularly in deeper parts. The granulation histograms are flat, without distinct culmination. Nevertheless the occurrence of bimodal systems can be observed. It is accompanied by poor or medium sorting of the deposits. Vast depressions within this level are filled with organogenic deposits. Their bottom is a 0.5 m thick layer of peat; its bottom has been dated at 10400 ± 160 years B.P. (Gd — 2232). It is overlaid by a 2 m thick layer of calcareous gyttja. The uppermost deposit in the series is a 0.3 m thick layer of decomposed peat with a considerable mineral admixture (Fig. 3).

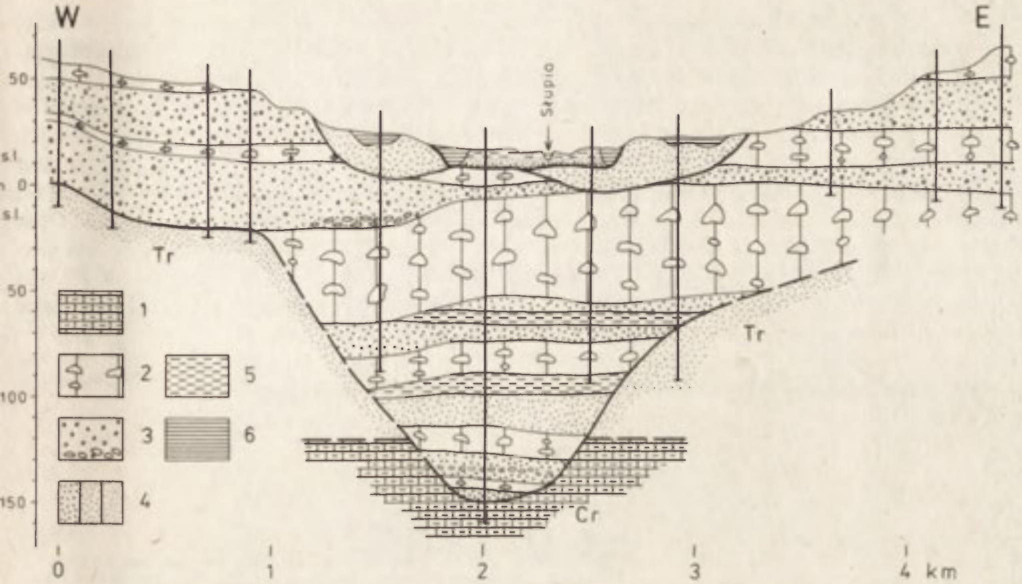

Fig. 3. The geological cross section through the Slupia's valley in Słupsk (after Orłowski, unpublished); 1 — Upper-Cretaceous marlstones, 2 — till, 3 — gravel with pebbles, 4 — sands varying in particle-size, 5 — silts, 6 — organogenic and mineral deposits of the paleomeander infillings

So the filling of the depression is young. It may be supposed that it progressed as the climate became damper, the dead ice or, maybe, the permafrost was thawing due to the development of the thermokarst lake.

Deposits of a different form fill the lower part of a transitory water basin which existed between the northern boundary of the present-day Słupsk and the morainic forms blocking the northward outflow. Deposits of varved clay type were accumulated here and exploited until the 1940s. Now the state of the excavation makes it impossible to find a profile that would be suitable for investigations.

THE SŁUPIA VALLEY DURING THE FORMATION OF RIVER TERRACES

The formation of the ravined section of the Słupia valley, from Włynkówko to Bydlino, probably favoured by the thawing dead ice in the warmer weather of the Bölling, has led to the drainage of the ephemeral Słupsk basin and to the exposure of its bottom. Most probably the water of the Słupia River began to flow into the Modła marginal lake (Sylwestrzak 1969).

The events progressed at great speed. South of the town the valley bottom lowered by several metres (2–5 m) while north of the town there occurred only slight changes. There began the formation of an undoubtedly fluvial terrace which forms, nowadays, an upper floodplain level. An essential part in the structure of that terrace has been played by washed glacifluvial deposits as well as deposits washed out of the grey till horizon (Fig. 3). The layer of fluvial deposits reaches a thickness of ca 4 m and, in places, it is overlaid with a 2 m thick layer of eolian sands. As a rule, the sand in the numerous oblong hollows does not reach 0.5 m. Gyttja and peat overlie it (Fig. 4 and 10).

The bottom of those deposits has been dated at 12300 + 180 years B.P. (Gd – 2116) (post Slupsk-South) and at 11150±210 years B. P. (Gd — 2117) (post Slupsk-Racławicka). Those dates have been fully confirmed in palinologic, diatom and malacologic investigations (W. Florek 1984; Alexandrowicz et al. in press). Near the confluence of the Glażna the lower terrace level is covered with an alluvial fan built by thaw waters, strongly loaded with rubble, from the marginal zone of the Gardno Phase (Fig. 4, and 6). This was probably accompanied by the development of glacifluvial naledis (Kozarski 1975, 1977). Its result is a counter-slope of the surface of mineral deposits, visible in the geological structure of the fan (Fig. 4). The occurrence of permafrost in the cone has also been proved by the analysis of the organogenic filling of the hollows. First of all, the composition of the malacofauna indicates the temporary existence of a thermokarst lake of the alas type (Alexandrowicz et al. in press).

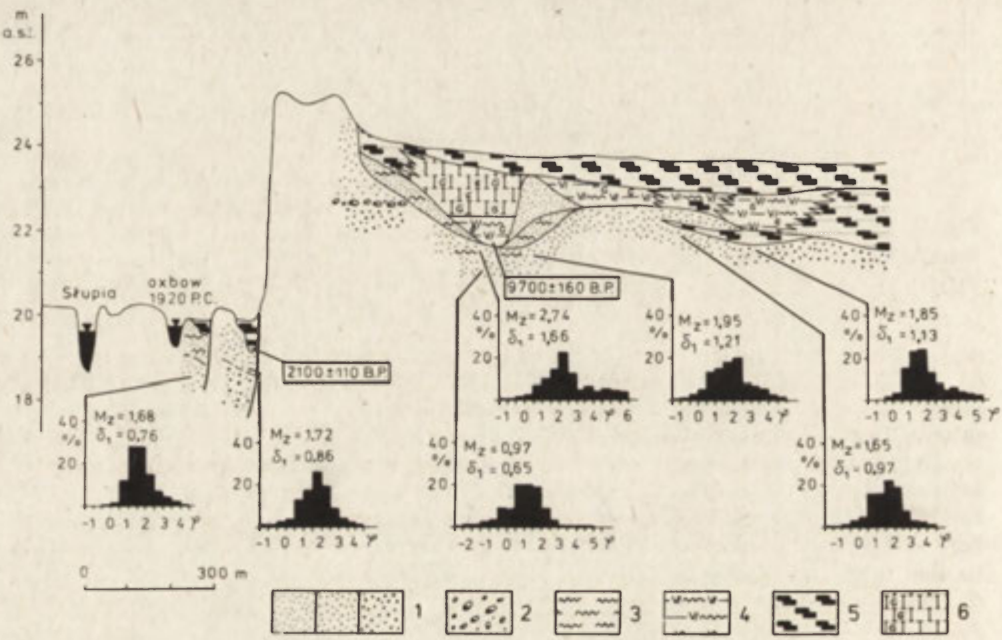

Fig. 4. The geological cross section through the floodplain deposits and the Glazna fan, together with exemplary histograms of particle-size distribution of various sedimentary groups; 1 — sands varying in particle-size, 2 — pebbles, 3 — silts, 4 — gyttja, 5 — lake marl, 6 — malacofaunal remains, 7 — peat

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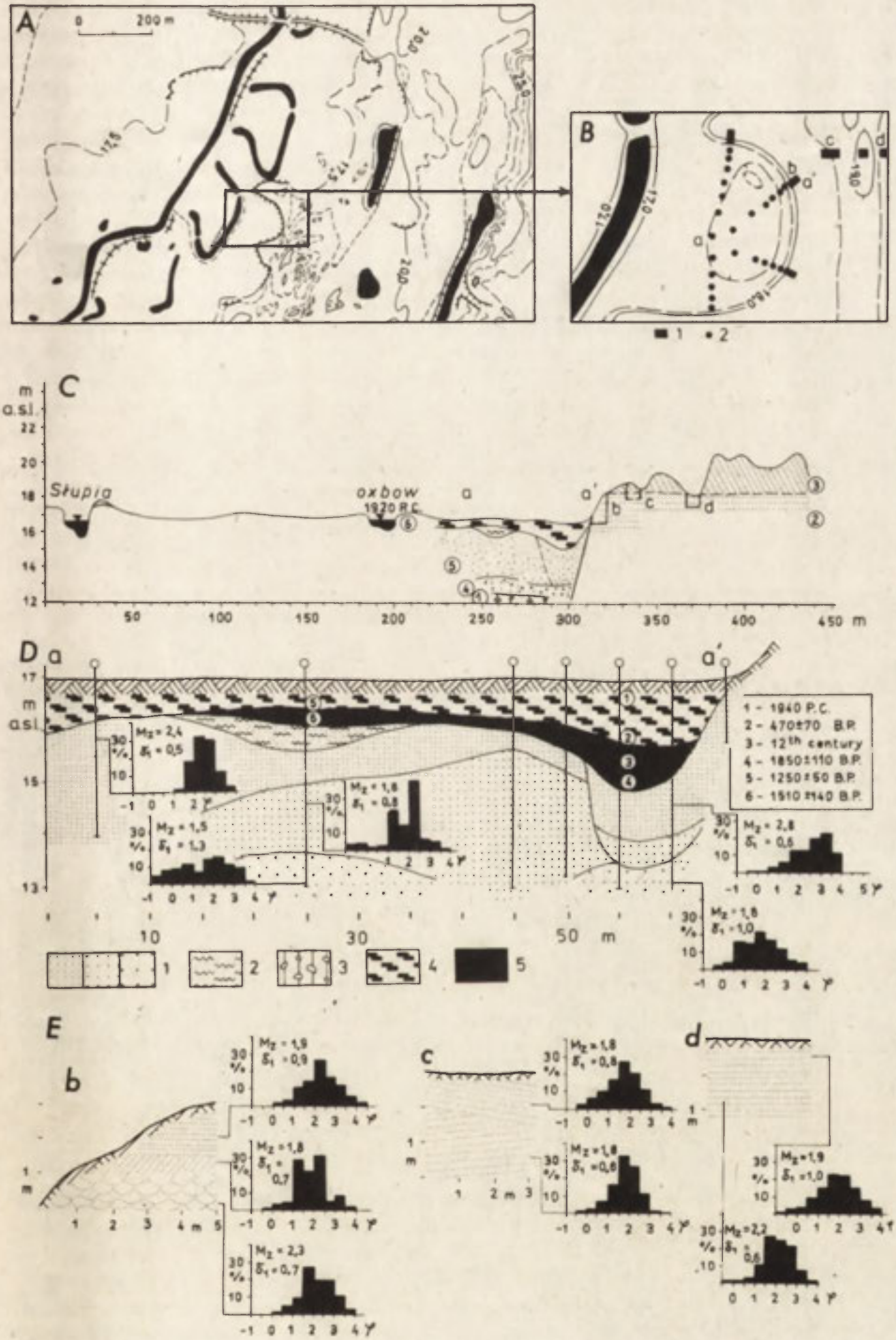
Fig. 5. The cross section through the Slupia's paleomeander south of Słupsk, A — hipsometry of the study area, a—a' — approximate course of the geological section shoved in part C, B — detailed study area: 1 — excavations, 2 — borings, a—a' — course of the geological cross section presented in part D, C —geological cross section through the east part of the floodplain and the upper floodplain terrace: 1 — grey till, 2 — upper floodplain terrace's sands varying in particle-size, 3 — eolian sands, 4 — medium- and coarse — grained sands of the floodplain, 5 — middle- and fine-grained sands of the point bars and paleomeander infilling, 6 — peat, b, c, d, — excavations showed in part E, D — geological cross-section through floodplain sediments (the numbers in circles corresponding to dates given beside): 1 — sands varying in particle-size, 2 — silts, 3 — till, 4 — brown peat, undecomposed, 5 — peat, varying in of decomposition degree, black or brown-grey


A
0 200 m
B
1 2
C
m a.s.l.
Słupia
oxbow 1920 B.C.
a
a'
b
c
d
50 100 150 200 250 300 350 400 450 m
D a
a'
m a.s.l.
1 - 1940 P.C.
2 - 470±70 B.P.
3 - 12th century
4 - 1850±110 B.P.
5 - 1250±50 B.P.
6 - 1510±140 B.P.
M_z = 2,4 δ_1 = 0,5
M_z = 1,6 δ_1 = 0,8
M_z = 1,5 δ_1 = 1,3
M_z = 2,8 δ_1 = 0,6
M_z = 1,6 δ_1 = 1,0
10 30 50 m
E
b
c
d
M_z = 1,9 δ_1 = 0,9
M_z = 1,8 δ_1 = 0,7
M_z = 2,3 δ_1 = 0,7
M_z = 1,8 δ_1 = 0,8
M_z = 1,8 δ_1 = 0,6
M_z = 1,9 δ_1 = 1,0
M_z = 2,2 δ_1 = 0,6

The northward water flow in the horizon of the above-flood terrace did not last long since it was stopped as early as in the Pre-boreal Period (Alexandrowicz et al. in press). In the Boreal Period the Słupia cut into the Late-Vistulian deposits of the lower terrace. Thus the accumulation on that terrace stopped definitely. As a result there is a stratigraphic gap in the deposits filling the hollows, which includes whole periods: the Boreal, the Atlantic and the Sub-boreal. The erosive cut of the Słupia was small – ca 2 m, its reach was determined by the horizon of grey Vistulian till which has been found directly under the deposits which fill the palaeomenders of various age in the lower part of the Słupia (W. Florek 1984, in press b; E. Florek, W. Florek, in press). It has also been stated that, when the flow approximates the mean yearly discharge, the same till horizon is exposed in the current line of the present-day river channel. The presence of that visible lithological barrier has had an essential effect on the limitation of

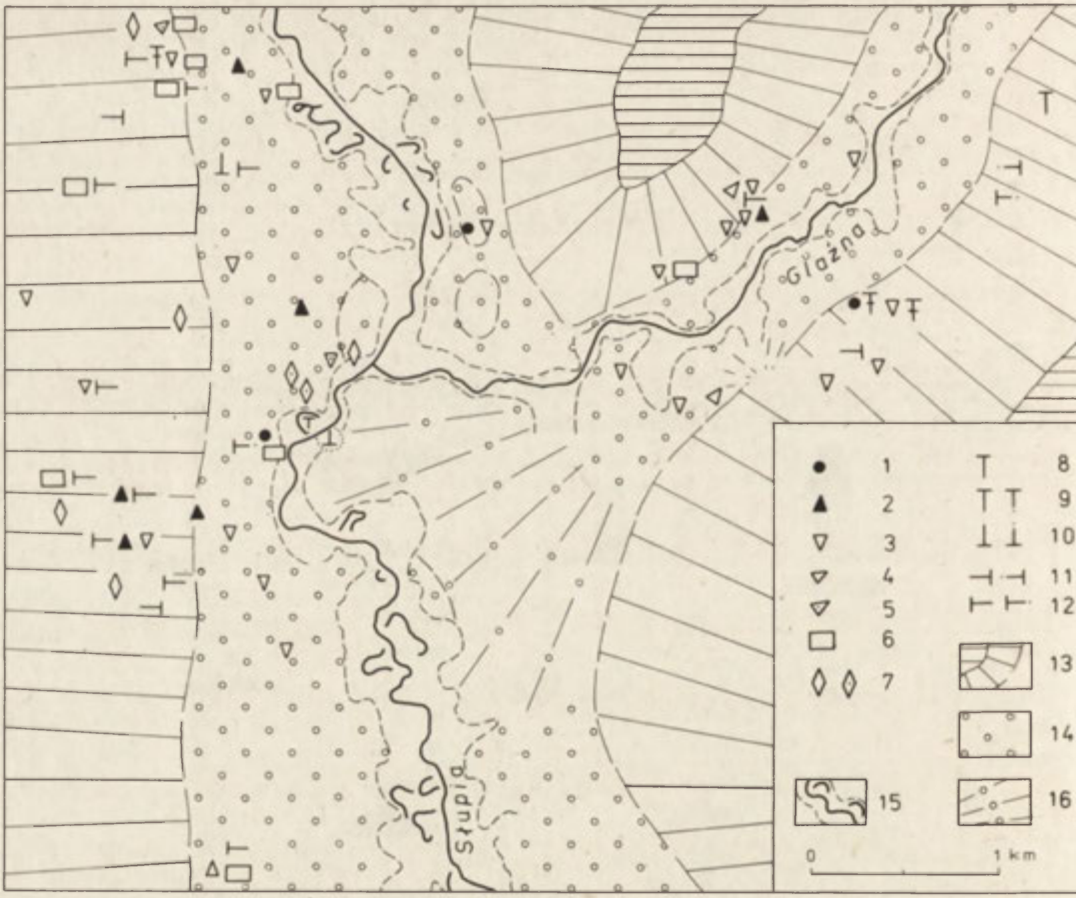


Fig. 6. Distribution of archeological findings in the Glazna confluence area on the bigger geomorphological units background: 1 — funnel beaker culture, 2 — Late Neolithic cultures, 3 — Late Bronze Age and Early Iron Age, 4,5 — Pomeranian culture, 6 — Late la Tene Period, 7 — Middle Ages Period (6/7th — 14th c.), 8 — 6/7th — 8th c., 9 — 8th — 9th c., 10 — 9th — 10th c., 11 — 11th — 12th c., 12 — 12th — 13th c. (all signatures with a dot indicate the settlements), 13 — moraine plateau, its slope and lower edge, 14 — over flood terrace levels, 15 — floodplain and its border, 16 — Glazna alluvial fan

the Słupia deep erosion and it was the principal reason of the vertical stabilization of the river channel in its lower course. It caused an intensive development of lateral migration processes of the river channel; since the moment of formation of the flood plain that migration has become the only radical way of reacting to the changes occurring in the drainage area and in the channels of the Słupia and its affluents. There have not been found, along the Słupsk section of the valley, any palaeomeanders from the earliest existence of the Słupia flood plain, but above the confluence of the Skotawa, in the vinicity of Dębnica Kaszubska, the bottom of the peat filling a palaeomeander has been dated at 8320+170 years B.P. (Gd – 2131). M. Latałowa (1983) has determined the age of this peat, using the palaeobotanic method, as the early Boreal Period.

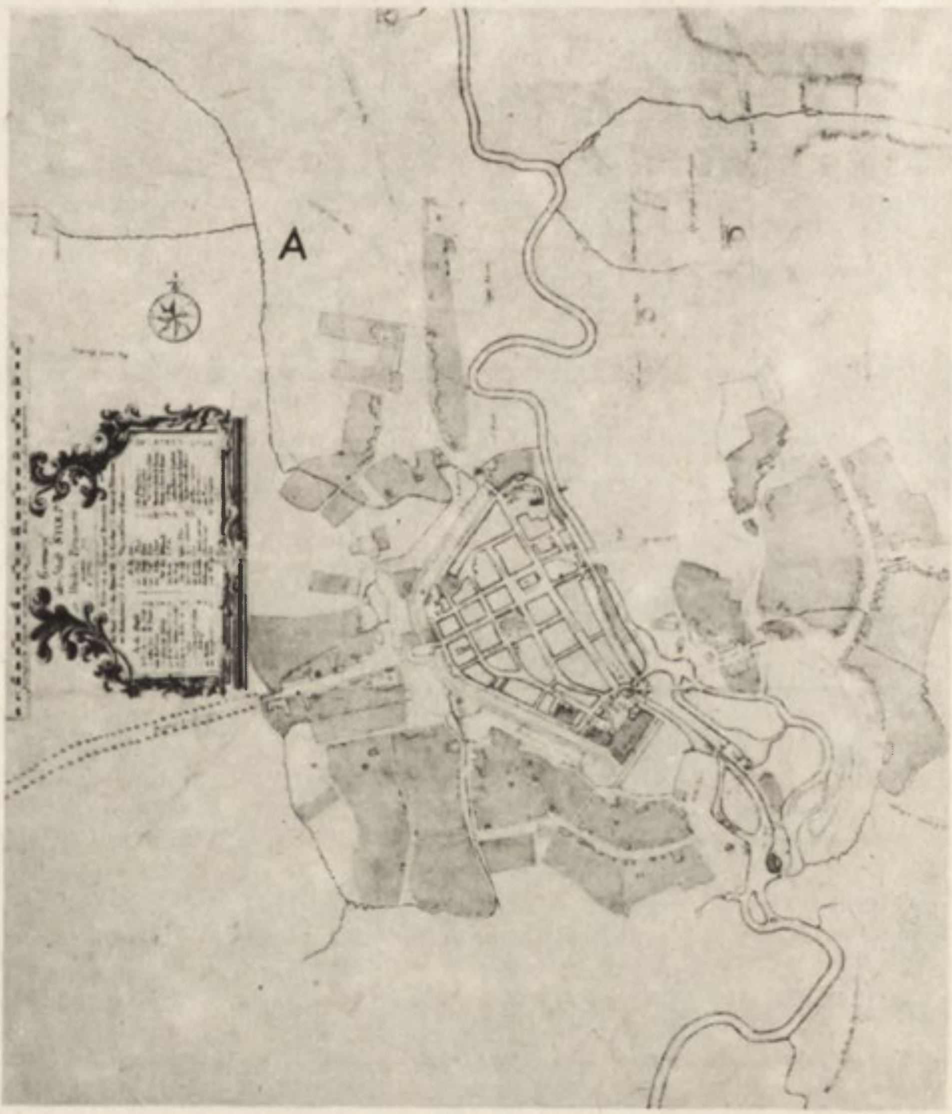
The previously described picture of the floodplain as well as the lithologic limitations and anthropogenic changes have determined that most preserved fossil channels and palaeomeanders are very young (E. Florek 1983, in press), only a few date from older parts of the Subatlantic or Subboreal period (W. Florek 1984; E. Florek, W. Florek in press). This is also confirmed by the dating results of the filling of palaeomeanders south of Słupsk (1850±110 years B.P. – Gd – 864, W. Florek 1983, in press a, b, Fig. 5) and near the Glaźna fan (3020±110 years B.P. – Gd – 2212, W. Florek 1984; E. Florek, W. Florek, in press). All these dates have been confirmed by palinologic expertises carried out by M. Latałowa (1981, 1983).

The dampening of the Słupia valley bottom, from the Glaźna confluence to Włynkówko also occurred in the Subatlantic period; it has been dated, according to the age of the bottom of peat deposits on the floodplain near a palaeomeander south of Słupsk, at 1510+140 years B.P. (Gd – 862) (W. Florek 1983, in press a, b, Fig. 5) and confirmed by the results of a palinological analysis carried out by M. Latałowa (1981).

Fig. 7. View of Słupsk from the east after Eilhardt Lubin's map of Pomerania (1618). Photo Edward Wojtowicz

The dampening has also attained the lower terrace where peat accumulated in depressions left by channels and marginal lakes. The process began in the Subatlantic period (Zachowicz 1983; Alexandrowicz et al. in press). In the preceding period the depressions and the lower terrace were freely penetrated by the Neolithic man, which proves that those areas were dry and easily accessible in the Subboreal (W. Florek 1984, in press c).

The dampening of the Slupia valley was not an isolated fact. At the same time the southern part of the Gardno Lake basin changed from a wet area into a shallow lake and, in its bottom, a younger peat layer was formed (Zachowicz 1977; Rosa 1977; Bogaczewicz—Adamczak et al. 1982), while a great part of the Gardno-Łeba Lowland turned into a marsh (Tobolski 1982). It is also then that the Modła Lake, existing till now, was formed west of the Słupia mouth (Bogaczewicz—Adamczak et al. 1980).



At that time the settlement of population groups belonging to the Lusatian Culture was intensively developing. They settled mainly on slopes of small river valleys (e.g. the Glażna valley — Fig. 6), where hydrogeologic conditions favoured the occurrence of springs, while the valley bottoms, previously formed by extraglacial waters, probably constituted moderately damp meadows, their fertility and size being an optimum for the scale and character of animal-breeding carried out by the Lusatian population (W. Florek in press c). At the same time the valley bottoms of larger rivers (in Pomerania's scale such is, among others the Slupia River) were an area of intensive lateral migration of meander channels and this was accompanied by floods and the progress of paludification in the valley bottom. It is a fact of some importance that the settlements of the Lusatian population in this area had an open character and so those people did not feel it necessary to use the shelter of river meanders and marshes for defence.

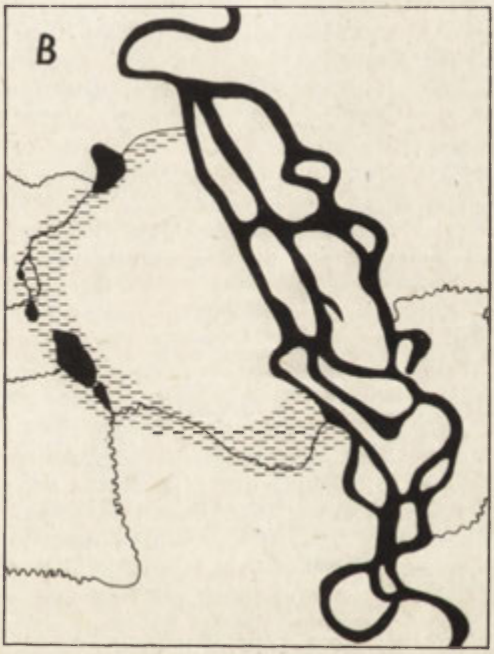


Fig. 8. A — Słupia river in the Słupsk area after Auen's town plan from 1780; B — An attempt at a reconstruction of the course of the Słupia river from a pre-foundation time (13th c.) using the material of Spors (1981) and Gręźlikowski and Szopowski (1983)

In the Middle Ages the settlements were mostly of a closed character, that is why their traces have been found inside the meanders, both in the vicinity of the Glaźna fan (Fig. 6) and in present-day Słupsk. The settling sites were chosen carefully which can be proved by the stability of the selected meanders (in the vicinity of the Glaźna fan an early medieval meander was cut no sooner than in the twentieth century for river-regulation purposes — Fig. 6). The concentration of settlements in the Słupsk region and the creation of a city brought about the development of economic activities and the construction of mills and fulleries. Those works, though rather numerous, did not change conditions in the Slupia channel because their construction did not result in the division of the whole channel. Another factor could have been of some importance:

navigation was carried on along the lower Słupia which brought about the necessity of the constant cleaning of the channel and its banks as the boats were drawn by people or by animals moving along the bank (Szopowski 1962; Czacharowski 1981).

A great number of damming-up devices on smaller streams (mainly mills and weirs for fishing) undoubtedly changed water conditions in the catchment areas but it is impossible, nowadays, to determine the exact scale of that process.

The Słupsk castle and its borough developed at a ford across the Słupia where its channel split into three arms. As an evidence of that fact there are three waving blue lines in the arms of the town (foundation charter in 1310) and the picture of the town on Eilhardt Lubin's map of 1618 gives an idea of the shape of the fourteenth-century channel of the Słupia (Fig. 7). Basing on an eighteenth-century plan of Słupsk one might try to reconstruct the position of the then existing rivers's arms (Fig. 8 B). J. Spors (1981) as well as J. Gręźlikowski and Z. Szopowski (1983) have tried to do it. Early feudal conditions favoured the economic development of Słupsk (Spors 1981). Its growth required new building grounds situated near the castle but safe from floods. To this end higher fragments of the floodplain were raised and some of the numerous arms and paleochannels of the Słupia (mainly East of the present-day channel) were filled-up. This led soon to the formation of a channel consisting, within the tows, of two almost parallel arms (Fig. 7, and 8) of which only one has finally remained. Newer buildings – the renaissance ducal castle, the Dominican and Norbertine cloisters and the so-called New Słupsk, later surrounded with walls, developed on the left river bank touching, on the west, an arched depression filled with not yet dated layers of mineral and organic deposits (among others three layers of peat). The depression was drained by two small streams and in the deepest places there were two small but deep lakes, probably as remains of evorsion kettles which had been preserved owing to dead ice. The streams were periodically dammed up and the water basins thus created fulfilled economic functions and acted as a moat.

Man's activities connected with the economic and spatial development of the town were added to the growth of peat-bogs in the valley bottom, progressing since the beginning of the Subatlantic period; as a result the bottom of the river-bed rose by ca 1 m and the level of the floodplain within the town was lifted by 1–2 m. In some sections of the left-bank valley bottom, south of the town, and in the old city center destroyed during World War II, the area of the old floodplain rose by another 1 m almost reaching the altitudes of depressions in the lower terrace.

The modern, active water control and exploitation has also brought great changes; it began, along the investigated section of the Słupia, at the end of the 19th century by starting hydrogeological observation and the elaboration of the first plans of flood-control and drainage. In the first twenty years of the 20th century many of those plans were realized. A dozen or so power plants were built on the Słupia, in its middle and upper course (the water plant of Młynki-Struga is the oldest one in Poland). In Słupsk a weir damming up water up-stream of the town and the feeding system of a large water-mill were reconstructed, and the river-course was shortened up-stream and down-stream of the town (Fig. 9).

The construction of damming-up steps caused the interception of a great part of coarser rubble so that the particular sections of the Słupia are now autonomous as regards the processes of erosion, transportation and accumulation of coarser fractions; hence a strong connection, observed by J. Dynus-Angiel (1980, 1983) of the features of bottom deposits with the features of local lithologic deposits accompanying the river-channel.

The negative result of drainage is also unquestionable. It means not only drying up valley bottoms, mainly in the lower course of rivers, but also a fast draining into rivers of thaw and precipitation water. However, it is impossible to quote here definite values.

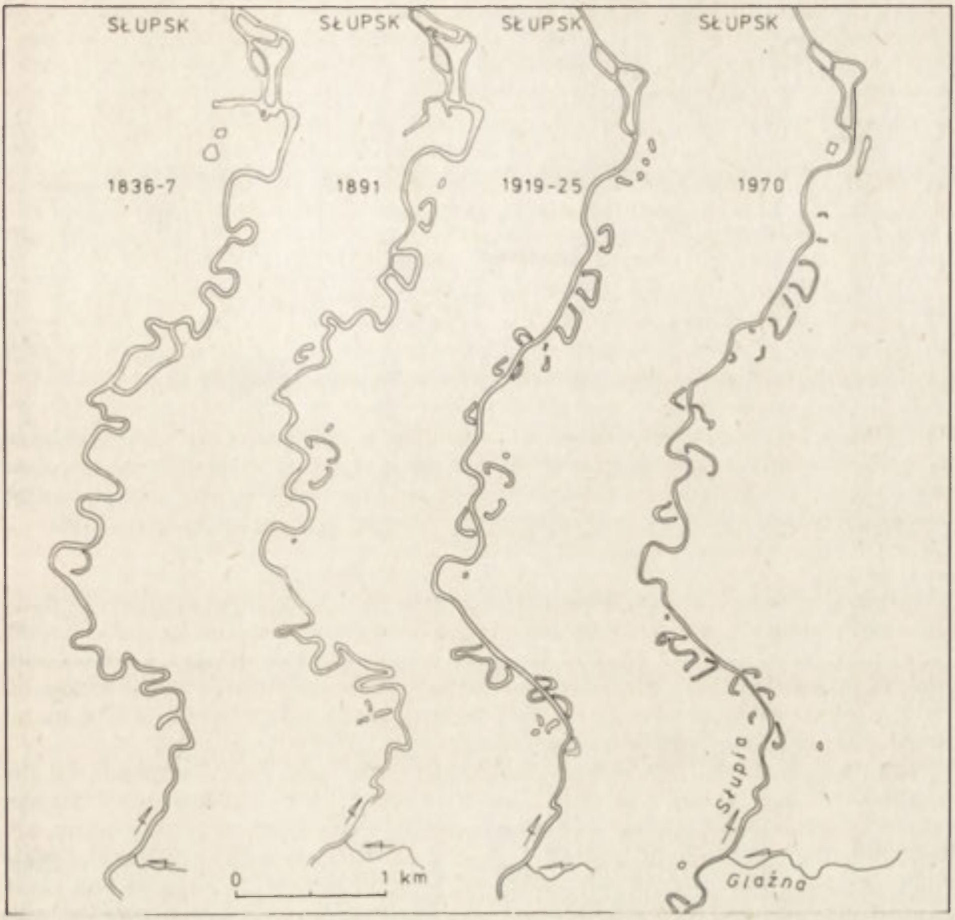


Fig. 9. The Słupia river before and after the regulation between confluence of Glaźna and Słupsk

CHANGEABILITY OF DEPOSITS AND FORMS IN COURSE OF TIME

Investigations carried out have made it possible to determine the regularities controlling the formation of the selected terrace levels (Fig. 10). A common feature of all morphologic levels is the relatively shallow occurrence of the roof of grey Baltic till (according to Mojski and Orlowski's terminology, 1978).

The oldest level (I), called the overflow terrace, is built of sands of various granulation with the prevalence of coarse-grained sands and gravels, most frequently diagonally bedded. The post-lake and post-channel depressions in this terrace are now filled with organogenic deposits: Preboreal peats and calcareous gyttjas with Subatlantic peat in the roof. The principal shape of the mineral surface of the terrace was probably formed between the Oldest Dryas and the Bölling.

The upper floodplain level (II), fully formed by river waters flowing from the Bölling till the Preboreal period inclusive, has a rather complex structure. There occur small rests of coarse-grained glacifluvial deposits and vast surfaces built of sands of different

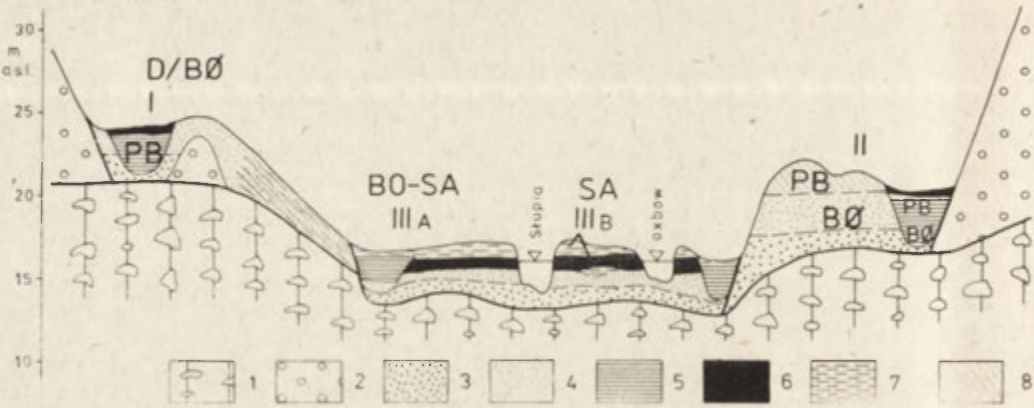


Fig. 10. The scheme of the Słupia's river terraces construction in Słupsk; 1 — till, 2 — glacifluvial gravels and sands, 3 — coarse-grained sand of the channel sub-facies, 4 — fine-grained sands of the bars sub-facies, 5 — organogenic and mineral paleomeander infillings, 6 — Subatlantic peats of the varying terrace levels, 7 — contemporary over-channel sediments, 8 — eolian sands

granulation with the prevalence of medium-grained sand diagonally, tabularly or flatly bedded. The roof of the sandy areas is surmounted by aeolian fine-grained deposits occurring as small dunes of an irregular shape. Post-channel depressions are filled with thick (up to 4 m) layers of peat and calcareous gyttjas which settled from the Bolling till the close of the Preboreal. Here, too, Subatlantic peats occur in the roof. There are no Boreal, Atlantic or Sub-boreal deposits.

The flood plain (III), in its principal shape, was formed from the Boreal till the Subatlantic b. It is built of deposits the variability of which is characteristic of terraces formed by meandering rivers. Coarse-grained deposits of the channel bottom are overlaid by fine-grained mineral deposits of meander backwater or by gyttjas and peats filling the palaeomeanders. On the surface of the flood plain there occurs a peat cover the formation of which began 1500 years ago. The peat deposits are interbedded with sands and silts of crevasse cones.

The youngest element (III B) is formed by an anthropogenic cover of heaped material; it covers the deposits of the floodplain, mainly in the left-bank part of the town, with a 1–2 m thick layer. It consist of debris, brought sand or litter. The regulated river channel (canalized within the town) is accompanied by embankments mostly built of sand and gravel taken from the river bottom while it was cleared and deepened. Embankments were built as early as in the Middle Ages since the dredging of the river channel was necessary for navigation. Even in the early phases of the Middle Ages the construction of embankments was known.

The two highest levels developed when the channel had an irregular shape, and an irregular flow and as the intensity of the water-flow was decreasing.

At the end for the Preboreal and the beginning of the Boreal period the development of vegetation was accompanied by a process of river drainage concentration which resulted in the formation of a meandering channel. Its energy was sufficient to cut into the lower terrace II along 2 metres. Since then the Słupia river channel, along its Słupsk section, has functioned uninterruptedly as a meandering channel. Its rectification and shortening was an artificial operation.

Investigations of the sedimentological features of the fluvial deposits of the Słupia have shown that neither in the distant past (overflow and upper floodplain terrace) nor in more recent times (flood plain) could the river achieve a radical transformation as

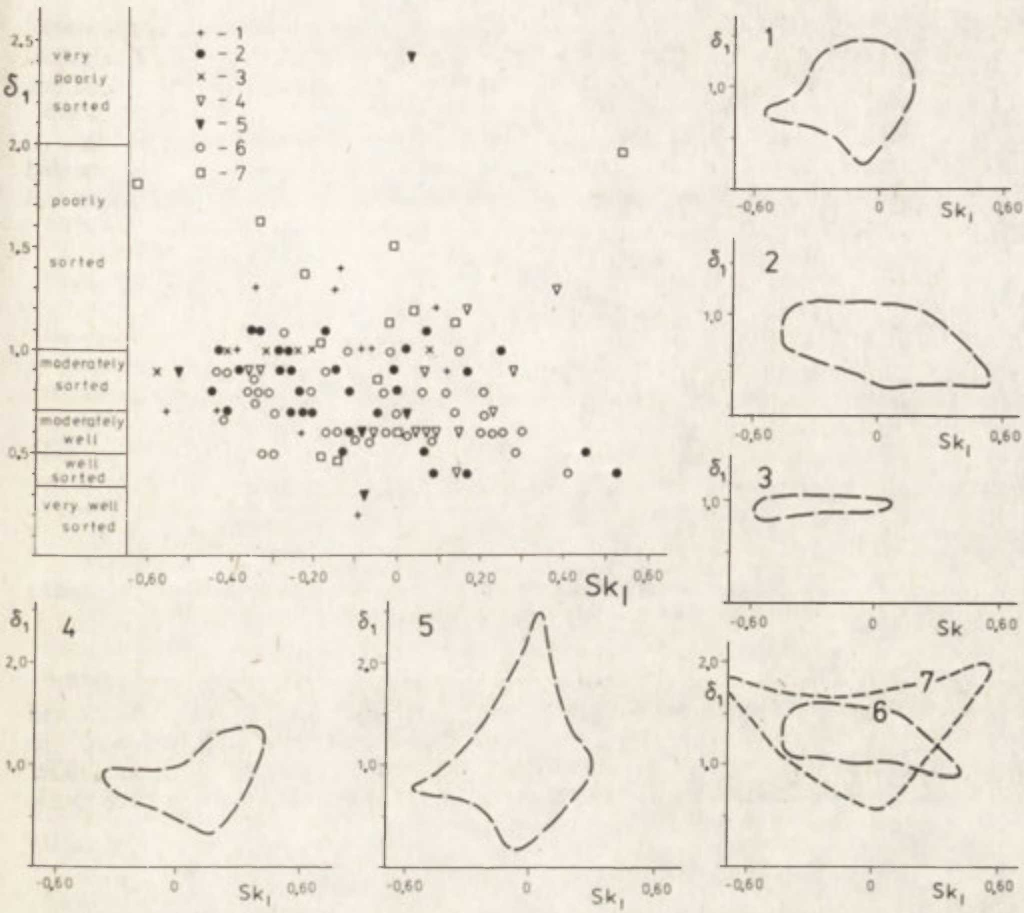


Fig. 11. The relationship between standard deviation (δ_1) and the graphic skewness (SK_I) indicators; sediments of the varying floodplain facies: 1 — channel bed, 2 — point bar, 3 — "cork" of paleomeander, 4 — paleomeander infilling, 5 — flood basin, 6 — fluvial sediments of the upper floodplain terrace, 7 — glacifluvial sediments of the overflow terrace

regards the granulation of deposits (W. Florek in press; E. Florek, W. Florek in press). Nevertheless the sorting activity of this small river has been reflected both in granulation histograms and in the graphic indices of granulation (Fig. 4, 5, 11). It concerns the differences between the deposits of the upper floodplain terrace and those of the floodplain as well as the differences among the particular facial types of the floodplain minerals (W. Florek, in press b).

The quartz grain abrasion also proves that the small distances of transportation and the limited energy of the Słupia could but slightly affect the degree of "fluvialization" of the deposits both of the lower terrace and of the various facies of the floodplain. The differences between the facies are rather the result of the morphoselection of the grains than of their abrasion (E. Florek, W. Florek in press).

More refined technics (surface analysis of quartz grains and of some heavy minerals) have revealed the existence of earlier connections. They are connections with the environment in which the deposits were accumulated and then transported to

Pomerania by the Scandinavian continental ice-sheet. Neither the action of the glacial and the periglacial environment of the close of the Vistulian nor that of the Lateglacial and Holocene fluvial environment could eliminate those features (E. Florek, W. Florek in press).

Finally, it should be noted that since the moment of formation of the overflow (I) terrace in the Słupsk section of the Słupia valley there occurred an uninterrupted tendency towards erosion (Fig. 12). The aggradational growth of the peat cover and

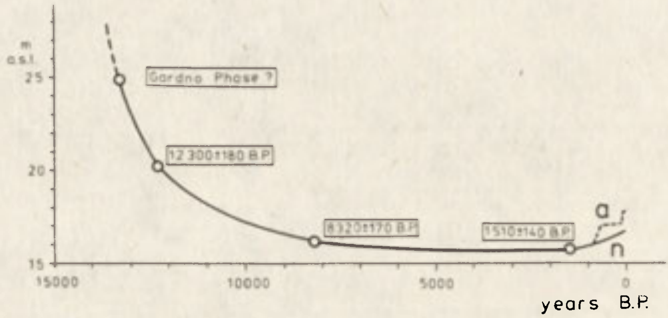


Fig. 12. Changes of the vertical position of the Słupia valley bottom in the Słupsk from the Gardno Phase: n — natural conditioned, a — antropogenic conditioned

crevasse deposits is connected with the Late-Subatlantic dampening of the climate and it concerns the valley section from the confluence of the Glazna to Wlynkowko. No tendency towards the aggradation of the valley can be observed along any other fragments of the lower section of the Slupia. A separate question is the anthropogenic rise of the valley bottom in the urbanized area.

REFERENCES

Alexandrowicz S. W., Florek W., Zaborowska K., Zachowicz J., in press, The Słupia upper floodplain in the vicinity of Słupsk, Pomerania, Poland, *Quaestiones Geographicae* 11.

Bogaczewicz—Adamczak B., Fedorowicz S., Gołębiewski R., Miotk G., 1982, Polodowcowa historia rejonu jeziora Gardno (Sum.: Post-glacial history of the region of Lake Gardno), *Zeszyty Naukowe Wydziału Biologii i Nauk o Ziemi UGd, Geografia* 12, 27—49.

Bogaczewicz—Adamczak B., Fedorowicz S., Miotk G., 1980, Paleogeografia strefy brzegowej południowego Bałtyku w rejonie jeziora Modła (Sum.: The paleogeography of the eastern-Baltic shore zone — Lake Modła region), *Zeszyty Naukowe Wydziału Biologii i Nauk o Ziemi UGd, Geografia* 11, 39—51.

Bülow K. v., 1930, *Erdgeschichte und Landschaftsgestaltung im Kreis Stolp in Pommern*, Stolp.

Czacharowski A., 1981, Miasto w okresie rozdrobnienia feudalnego 1307—1478, in: S. Gierszewski, ed., *Historia Słupska*, Wyd. Pozn., Poznań.

Dawid—Hryniewicz A., Hryniewicz B., 1984, *Morfogeneza doliny Słupi na odcinku od ujścia Glaźny do Słupska*, Zakład Geografii WSP Słupsk, typescript.

Deecke W., 1911, Zur Morphologie und Tektonik Pommern, *Zeitschrift der Deutschen Geologischen Gesellschaft* 63.

Dynus—Angiel J., 1980, *Przestrzenne zróżnicowanie materiału dennego w korytach rzecznych (na przykładzie rzek północnego skłonu Pomorza Zachodniego)*, Ph.D. thesis, Biblioteka Wydziału Geografii i Studiów Regionalnych UW, typescript.

Dynus–Angiel J., 1983, Typy uziarnienia rumowiska dennego rzek północnego skłonu Pomorza Zachodniego (Sum.: Types of grain-size distribution of the northern slope of West Pomerania's rivers bed material), *Przegląd Geafizyczny* 28, 2, 185–193.

Florek E., 1983, Kartograficzny i fotointerpretacyjny obraz zmian biegu koryt Słupi i Łupawy, in: Materiały Zjazdu Geografów Polskich, *Toruń 15–18 września*, Toruń, 89–90.

Florek E., in press, Kartograficzny i fotointerpretacyjny obraz zmian koryta Słupi, *Prace Geograficzne UW.*

Florek E., Florek W., in press, The Słupia river floodplain — its age and development (Pomerania, Poland), *Quaternary Studies in Poland.*

Florek W., 1983, Sedymentologiczne cechy różnowiekowych osadów teraz rzecznych Słupi, in: Materiały Zjazdu Geografów Polskich, *Toruń 15–18 września*, Toruń, 90–91.

Florek W., 1984, *Ewolucja i mechanizm transformacji dolin Słupi i Łupawy* (*sprawozdanie z badań za lata 1981–1984 MR I/25 06.6*), Zakład Geografii WSP Słupsk, typescript.

Florek W., in press a, Wstępne uwagi o wieku osadów budujących dno doliny dolnego odcinka Słupi, *Badania Fizjograficzne nad Polską Zachodnią*, ser. A.

Florek W., in press b, Cechy sedymentologiczne i wiek osadów równiny zalewowej Słupi, *Prace Geograficzne UW.*

Florek W., in press c, Rozwój sieci rzecznej Przymorza na przełomie okresów subborealnego i subatlantyckiego a działalność człowieka (na przykładzie Słupi i Łupawy), in: *Kultura łużycka na Pomorzu*, Słupsk.

Giedrojć–Juraha S., 1949, Moreny czołowe okolic jeziora Gardno, *Czasopismo Geograficzne* 20, 1–4, 239–243.

Grężlikowski J., Szopowski Z., 1983, *675–lecie Słupska — wystawa*, Słupsk.

Keilhack K., 1930, *Geologische Karte der Provinz Pommern und der anschliessenden Teile der Grenzmark, 1:500000*, Berlin.

Kondracki J., 1978, *Geografia fizyczna Polski*, PWN, Warszawa.

Koutaniemi L., Rachocki A., 1981, Palaeohydrology and landscape development in the middle course of the Radunia basin, North Poland, *Fennia* 159, 2, 335–342.

Kozarski S., 1975, Oriented kettle holes in outwash plains, *Quaestiones Geographicae* 2, 99–112.

Kozarski S., 1977, Ukierunkowane wytopiska na sandrach (Sum.: Oriented kettle holes in outwash plains), *Studia Geologica Polonica* 52, 235–248.

Latałowa M., 1981, *Ekspertyza palinologiczna osadów organogenicznych z równiny zalewowej Słupi koło Słupska*, Zakład Geografii WSP Słupsk, typescript.

Latałowa M., 1983, *Wyniki badań palinologicznych osadów z doliny Słupi*, Zakład Geografii WSP Słupsk, typescript.

Marsz A. A., 1984, Główne cechy geomorfologiczne (Sum.: Main geomorphological features), in: B. Augustowski, ed., *Pobrzeże Pomorskie*, Ossolineum, Gdańsk.

Mojski J. E., Orłowski A., 1978, Plejstoceńska forma rynnowa okolic Słupska (Sum.: Pleistocene sub-glacial channel from the vicinities of Słupsk), *Kwartalnik Geologiczny* 22, 1, 171–179.

Orłowski A., 1981, Ukształtowanie powierzchni podczwartorzędowej w dolinie Słupi i obszarach przyległych, *Koszalińskie Studia i Materiały* 2, 191–203.

Orłowski A., 1983, *Litostratygrafia i paleogeomorfologia doliny Słupi*, Ph.D. thesis, Katedra Geografii Fizycznej UGd, typescript.

Piasecki D., 1976, Doliny złożone rzek zachodniego Przymorza (Sum.: Composite valleys of rivers of the Western Baltic littoral), *Czasopismo Geograficzne* 47, 1, 21–32.

Piasecki D., 1982, Ewolucja dolin rzek Przymorza (Sum.: Evolution of river valleys in the Przymorze (maritime region)), *Przegląd Geograficzny* 54, 1–2, 49–67.

Rachocki A., 1973, Wstępne wyniki badań tempa zasypywania zbiornika w Rutkach (Sum.: Preliminary results obtained from investigations of the rate of silting-up of the Rutki storage basin), *Przegląd Geograficzny* 45, 3, 613–618.

Rachocki A., 1974, Przebieg i natężenie współczesnych procesów rzecznych w korycie Raduni (Sum.: Course and intensity of present-day fluvial processes in the Radunia river as example), *Dokumentacja Geograficzna* 4.

Rachocki A., 1981, Forms of sand-gravel streaks in contemporary river channels, *Quaestiones Geographicae* 7, 91—96.

Rosa B., 1963, *O rozwoju morfologicznym wybrzeża Polski w świetle dawnych form brzegowych* (Zsf. Uber die morphologische Entwicklung der Küste Polens im Lichte der Strandforms). Studia Societatis Scientiarum Torunensis, Sec. C, 5.

Rosa B., 1964, O utworach aluwialnych i biogenicznych wyścielających dna dolin rzek nadbałtyckich, ich związku z transgresją morza i znaczeniu dla badań nad neotektoniką obszaru (Sum.: On alluvial and biogenic deposits filling the floors of the rivers of the Polish coastal zone, their connection with the transgression of the sea and their significance in the investigation of the neotectonics of this region), *Zeszyty Naukowe UMK* 10, *Geografia* 3, 85—108.

Rosa B., 1968, Obszar południowobałtycki w okresie ostatniego zlodowacenia i w holocenie (Sum.: The south Baltic area during the Last Glaciation and the Holocene), in: *Ostatnie zlodowacenie skandynawskie w Polsce*, Prace Geograficzne IG PAN 74, 121—155.

Rosa B., 1977, Wyniki badań geologicznych w rejonie jeziora Gardno, *Studia i materiały Oceanologiczne* 19, 309—311.

Rosa B., 1984, Rozwój brzegu i jego odcinki akumulacyjne (Sum.: The development of the shore and its alluvial plain), in.: B. Augustowski, ed., *Pobrzeże Pomorskie*, Ossolineum, Gdańsk.

Spors J., 1981, We wczesnofeudalnym Księstwie Sławieńsko-Słupskim i Gdańsko-Pomorskim w XII—XIV wieku (do roku 1308), in: S. Gierszewski, ed., *Historia Słupska*, Wyd. Pozn., Poznań.

Starkel L., Thornes J. B., eds., 1981, *Palaeohydrology of river basins, British Geomorphological Research Group*, Technical Bulletin 28.

Sylwestrzak J., 1969, Odpływ wód roztopowych na tle recesji lądolodu we wschodniej części Równiny Słupskiej i Wybrzeża Słowińskiego, *Zeszyty Geograficzne WSP Gdańsk* 11, 9—77.

Sylwestrzak J., 1973, *Rozwój sieci dolinnej na tle recesji lądolodu w północno-wschodniej części Pomorza* (Sum.: The development of the valley system against a background of the ice-sheet recession in the north-east part of Pomerania), Uniwersytet Gdański, Gdańsk.

Sylwestrzak J., 1978a, *Rozwój sieci dolinnej na Pomorzu pod koniec plejstocenu* (Sum.: The development of the valley system in Pomorze towards the close of the Pleistocene), Gdańskie Tow. Nauk., Gdańsk.

Sylwestrzak J., 1978b, Zagadnienia morfologii i typizacji dolin północnego skłonu Pomorza (Sum.: Chosen problems of the morphology and classification of the types of valleys on the north slope of Pomerania), *Biuletyn Instytutu Geologicznego* 306, *Z badań czwartorzędu w Polsce* 21, Wyd. Geol., Warszawa, 199—231.

Szopowski Z., 1962, *Małe porty Pomorza Zachodniego w okresie do drugiej wojny światowej*, PWN, Warszawa.

Tobolski K., 1982, Antropogenic changes in vegetation of the Gardno-Łeba Lowland, N. Poland, *Acta Palaeobotanica* 22, 1, 131—139.

Zachowicz J., 1977, Analiza palinologiczna osadów jeziora Gardno, *Studia i Materiały Oceanologiczne* 19, 299—303.

Zachowicz J., 1983, *Sprawozdanie z analiz palinologicznych osadów terasy nadzalewowej Słupi i stożka Glaźny*, Zakład Geografii WSP Słupsk, typescript.

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MORPHOMETRY AND MORPHODYNAMICS OF THE LOWER VISTULA CHANNEL MESOFORMS

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1. INTRODUCTION

Mesoforms are of major importance in the study of channel processes defined by Popov (1977) as an unsteady continuous movement of the river channel bed under the influence of flowing water. Kondratiev et al. (1982: p. 23) account for this in a special way. They have stated explicitly that "the study of mesoforms is a way by means of which the fundamental principles of a typical channel process may be discovered and its logic may be understood". Following earlier suggestions (Kurpianov and Kopaljani 1979), they also state that the channel process is chiefly conditioned by a mechanism for load transport and they establish certain relationships between them. These relationships have also been built up by Schumm (after Shen 1982; p. 13) and by Winkley et al. (1984: p. 84). From these data it can be inferred that a channel pattern (channel process) and an adequate system of mesoforms depend on the quantity and quality of the transported load.

According to Rzhanitsin (1984: p. 130), the mechanism for load transport is a complex process. It occurs continuously in one case or takes place in a series of jerks to become cyclical in another case. Continuous load transport leads to the formation of dunes and sand ripple marks, i.e. microforms, whereas discontinuous transport results in central and lateral bars (islands), i.e. mesoforms.

The term channel mesoforms refers to forms, the size of which corresponds with the channel width and high stability of which is determined by hydraulic geometry of a stream (Kondratiev and Popov 1967; Antropovski 1969). They usually comprise single large sand-gravel waves and fixed lateral bars, the surface of which lies in a zone of average water stages although they are formed at high water stages.

British-American investigators, for example Church and Jones (1982), Cant and Walker (1978), Ferguson and Werritty (1983), Kellerhals et al. (1976), Schumm (1985), use a broader typology of mesoforms. According to Znamenskaya (1976: p. 9), it is partly based on variations of identical forms. N. D. Smith (1978: Table 1) provides a comprehensive description of channel mesoform divisions and nomenclature. He uses as many as 32 terms referring to forms of bar type, which are based on the morphological criterion.

Kondratiev and Popov (1967), as well as Znamenskaya (1976) group mesoforms into longitudinal sand waves, lateral and central bars, depending on channel conditions. A different classification of bars dependent on the degree of form stability (Fig. 11.4) and based on the relationship between form morphology and functions of stream hydraulic resistance and the amount of load (Table 11.2) has been given by Church and Jones (1982). The above divisions and classifications supplemented by other studies, including those by Krigström (1962), Task Force (1966), Collinson (1970), Kellerhals et al. (1976), Barwis (1978), Cant and Walker (1978), Levey (1978), Ferguson and Werritty (1983), Carson (1984) and Schumm (1985), are presented as a scheme in Figure 1. Recurrence of names in a variety of systems in the scheme is due to the complexity of load transport and to the naming of different forms with identical words by different investigators.

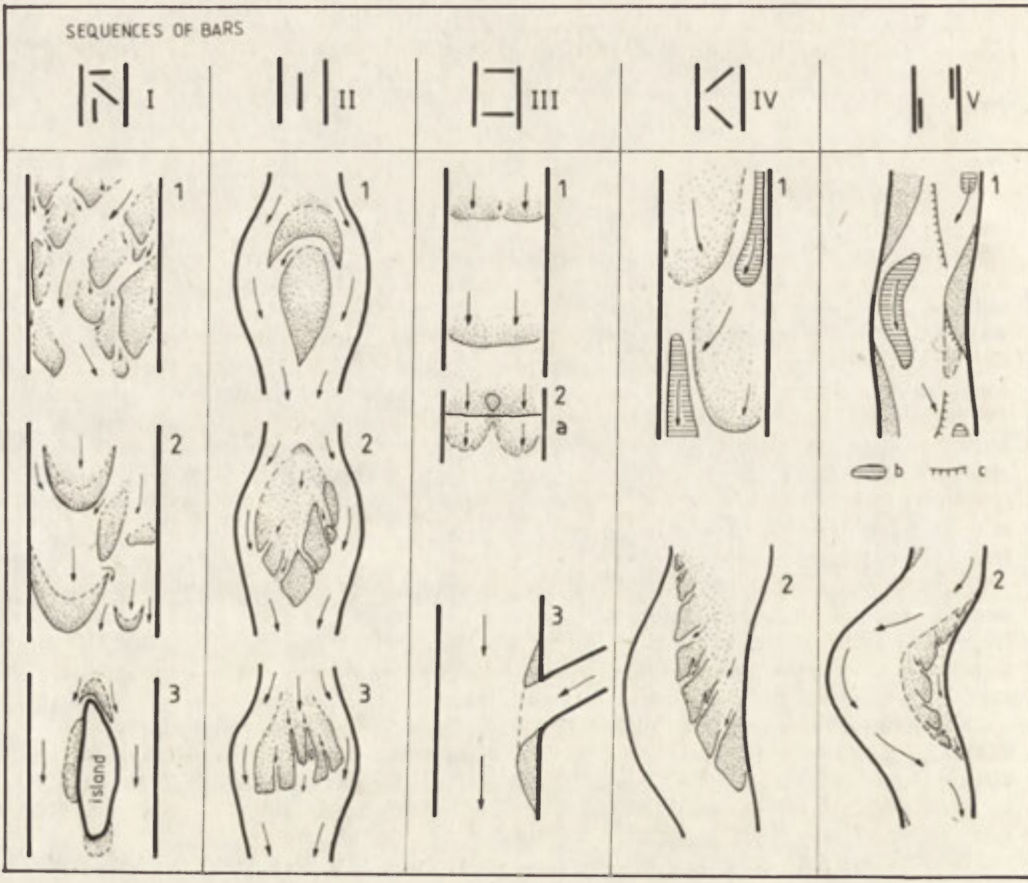

Fig. 1. Schematic system of channel mesoforms in sandy-gravelly-bottomed rivers (see the text for more information).

I. Changing sequence: 1 — braided channel bars; 2 — sandbars: linguloid bars, dunes, longitudinal bars, central bars, mid-channel bars; 3 — near-island bars. II. Midstream longitudinal sequence: 1 — longitudinal bars, crescent-shaped longitudinal bars, 2 — midstream bars, medial or diamond bars, 3 — transverse bars. III. Transverse sequence: 1 — transverse bars, cross-channel bars, 2 — transverse bars with horns (a: developmental stage), 3 — junction bars. IV. Diagonal sequence: 1 — alternate diagonal bars, 2 — diagonal bars. V. Lateral sequence: 1 — lateral bars, side bars, 2 — point-bars, (b) pools, (c) riffles

2. RESEARCH AIM, SCOPE AND METHODS

The objective of this study is to determine the morphometry and morphodynamics of the prevailing mesoforms of the lower Vistula channel, i.e. unsteady movement of load. In view of the fact that these forms are the resultant of load movement intensity at a definite time interval, the Vistula load balance can be estimated.

The study of channel mesoforms is based on the interpretation of large-scale setting-altitude maps, air-photo interpretation, bathymetric maps, cross-sections and longitudinal profiles of the Vistula channel bed. The method of successive comparisons has been used. For the purpose of observing temporary changes in channel bed morphology, multiple geodesic measurements of sandbar dynamics were carried out in two sections subjected to detailed analysis, namely at Nieszawa and Świecie (Fig. 2),

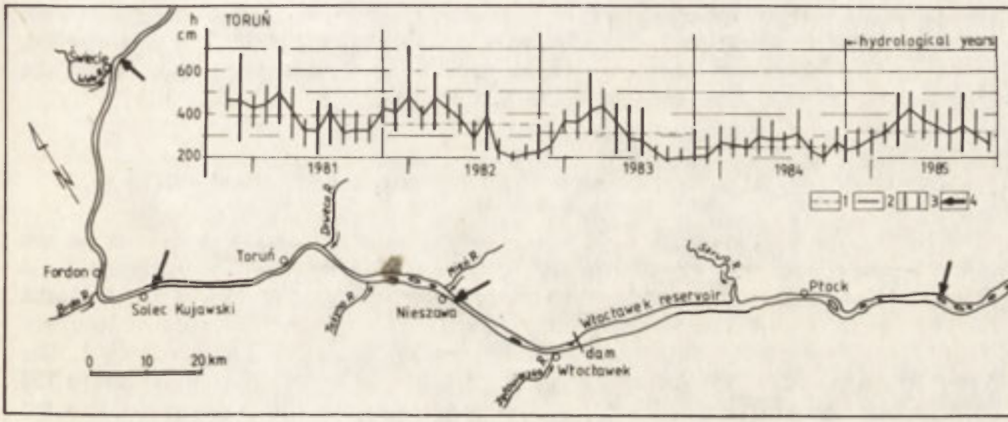

Fig. 2. Sketch-map of the Vistula section under investigation and characteristics water stages for the Toruń water-gauging station: 1 — average annual water stages, 2 — average monthly water stages, 3 — monthly amplitudes of water stage fluctuations, 4 — sections subjected to detailed study (see the text for interpretation)

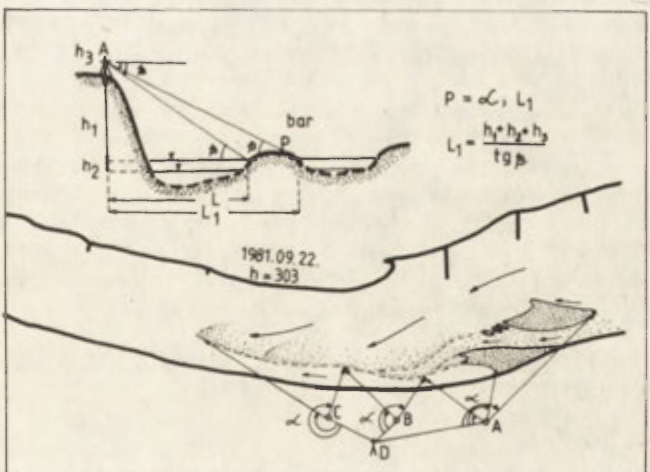

Fig. 3. Sketch-map of the Vistula channel below Nieszawa and a description of the geodesic measurement method (see Chapter 2 for some explanation)

throughout the year. The method of geodesic measurements is presented in a schematic form in Figure 3. It involves geodesic measurements of horizontal and vertical angles in characteristic bar portions at fixed field reference sites (A – C) against fixed observation points (D) (Fig. 3). Rapid performance without the need to use extra measuring equipment (measuring rods) and without other staff is their advantage. On the other hand, their disadvantage is due to spatial constraints resulting from measurements taken at a high vertical angle (high banks) and to time limits depending on water stages at which the bars are either emerging or close to the water table level, i.e. they can be seen. The data thus obtained and the above cartographic material have been supplemented by the hydrological analysis of the Vistula.

The study of the Vistula channel mesoforms was concerned with its lower reaches from the water dam at Włocławek as far as the mouth of the Wda river (Fig. 2). Hence, investigations were carried out along a variably improved reach, i.e. along that displaying a varying degree of channel development. Those detailed investigations were supplemented by the examination of channel processes occurring on the uncontrolled, untamed Vistula before the improvements were initiated (Solec Kujawski) and on the present-day Vistula after passing through Płock (Kępa Polska) (Fig. 2).

3. GENERAL HYDROLOGIC CHARACTERISTICS OF THE LOWER VISTULA

During the year one (spring) or two (spring, summer) peak flows occur on the Vistula with low stages in summer and late autumn. In this century maximum flood wave 916 cm high was recorded at the water-gauging station in Toruń on 31 March 1924. The data were collected from this gauging station because of its central location. The lowest water stage of 121 cm was observed on 20 and 21 December 1961. The annual amplitudes of water stages in the years 1956 – 1985 ranged from 248 cm to 758 cm in 1984 and 1962, respectively. The average annual water stages in Toruń fluctuated between 256 cm and 404 cm in 1984 and 1975, respectively. In the hydrological years 1956 – 1985, the mean was 336 cm at the mean discharge values of about 980 m^3s^{-1}. In view of these mean values recorded over a multi-year period, there existed average hydrologic conditions with extreme values of 718 and 174 cm in 1982 and 1984, respectively, at the annual water stages fluctuating between 248 and 526 cm in 1984 and 1982, respectively, over a period, during which the measurement of bar dynamics was performed, i.e. 1981 – 1985. The average annual water stages fluctuated between 256 to 396 cm in 1984 and 1981, respectively. On the average, they attained 328 cm (Fig. 2) but in general, there was a tendency toward lower water stages from the beginning of that period up to 1984. Since 1970 when the water dam was constructed at Włocławek (Fig. 2), the hydrologic regime of the lower Vistula has changed. Since that year the dynamics of flowing water has decreased in the Vistula section between Płock and Włocławek due to the creation of a water reservoir. Hydrologic activity has increased below the dam where diurnal fluctuations of the water level may approach 3 m (Babiński 1982). The fluctuations are observable along the entire reach under investigation (Machalewski et al. 1974) but they have a marked effect on channel processes as far as the mouth of the Tążyna river (Fig. 2).

4. GENERAL CHARACTERISTICS OF THE LOWER VISTULA CHANNEL DEVELOPMENT

The Vistula channel and the floodplain cut through fluvioglacial and fluvial sandy deposits. During the Holocene it was liable to several conversions from a braided to meandering pattern to become again braided. This was initially due to climatic changes and finally, it was caused by the economic action of man. In the 17th century particularly intense influence of man on the environment, i.e. forest clearance and cultivation of root crops, produced marked changes in the fluviodynamic processes on the Vistula (Falkowski 1967). At that time the Vistula was overloaded with the debris that it carried. Great fluctuations of water level were due to frazil-ice jams. They were accompanied by an impressive local decrease in channel slope and lowering of water table level. The channel contained numerous bars, chiefly central ones, and the intervening lateral limbs. These facts receive confirmation from large-scale maps of the Vistula channel which date back to the 19th century (cf. Babiński 1981, 1982, 1984; Koc 1972; Kopczyński 1982; Tomczak 1971). This untamed character of the Vistula is still predominating in part of the channel after passing the Włocławek reservoir (Fig. 4A). Numerous islands with central bars of high dynamics and in chaotic arrangement (Fig. 4A) rendered navigation difficult. In such a situation as that described above, the construction of controlling structures became necessary.

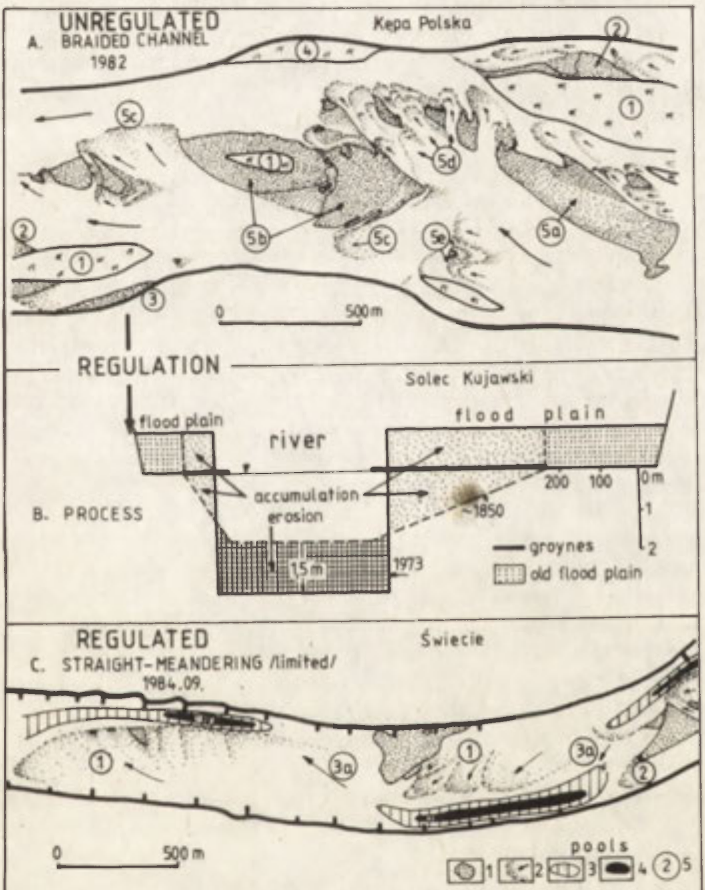

Fig. 4. Development of channel processes on the Vistula due to improvements: A — unregulated Vistula section in the vicinity of Kępa Polska; B — Vistula channel process below Solec Kujawski; C — regulated Vistula section below Świecie, 1 — emerged bar parts, 2 — submerged bar parts, 3 — channel depth on the order of 4—6 m, 4 — channel depth greater than 6 m, 5 — bar labels (see Chapter 5)

The plan of a series of general improvements which was introduced by an engineer, Mr. Sewerin, in 1830 implied that the lower Vistula channel would be slightly meandering and would be 350—375 m wide at the average water stage (Fig. 4B). The improvements were stage-made at varying intensity, especially through the construction of groynes and lateral dams. The improvements were fully made along a section between the mouth of the Tążyna river (Fig. 2) and the Baltic Sea by the end of the last century. On the other hand, a section upstream of the mouth of the Tążyna river was incompletely improved at the turn of the 1940s.

In consequence of the improvements made along the lower reaches, a new morphodynamic channel system results (Babiński 1984). On the one hand, the channel width confined to 350—375 m produced deep erosion leading to permanent deepening of the bed zone by about 1.5 m during over a hundred years but on the other hand, a new floodplain level has formed due to the infilling of zones separated by the groynes. This mechanism is schematically presented in Fig. 4B and in the present author's study of 1984 for a 5-km-long reach in the vicinity of Solec Kujawski (Fig. 2). The narrowing of the channel and its meandering pattern, though not always so, also contributed to changes in the system of channel mesoforms (Fig. 4A and C). As has already been mentioned, chaotically arranged central bars remained predominant mesoforms prior to the improvements (Fig. 4A). Nowadays, diagonal bars together with pools occurring in alternation prevail (Fig. 4C).

5. THE LOWER VISTULA CHANNEL MESOFORMS

Field studies carried out in the years 1978—1985, air photo interpretation and the literature provide the basis of grouping of bars occurring in the lower Vistula channel into the following types which are schematically shown in Figure 1 and are labelled in Fig. 4A and 4C:

(1) with regard to an unimproved braided river section with islands (No 1, Fig. 4A) and to an initially improved intermediate braided-straight river section (Fig. 3):
- (i) lateral, frontal and lower near-island bars (No 2),
- (ii) lateral bars (No 3) and fixed lateral bars (No 4):
  - (a) natural,
  - (b) artificial, separated by groynes,
- (iii) central bars (No 5):
  - (a) longitudinal,
  - (b) crescent-shaped longitudinal bars,
  - (c) linguloid bars,
  - (d) diagonal bars developed from riffles,
  - (e) bars with horns,
- (iv) diagonal bars along an initially improved section,

(2) with regared to an improved straight-meandering (meandering limited by groynes) river section (Fig. 4C):
- (i) diagonal bars terminating in lateral bars (No 1),
- (ii) linguloid bars (No 2),
- (iii) riffles (No 3a) and pools.

6. DYNAMICS OF THE LOWER VISTULA CHANNEL MESOFORMS

The two section of the Vistula channel that have been subjected to morphodynamic analysis are as follows:

(1) an initially improved section at Nieszawa, where there are chiefly lateral, central and linguloid bars (Figs 3 and 5),

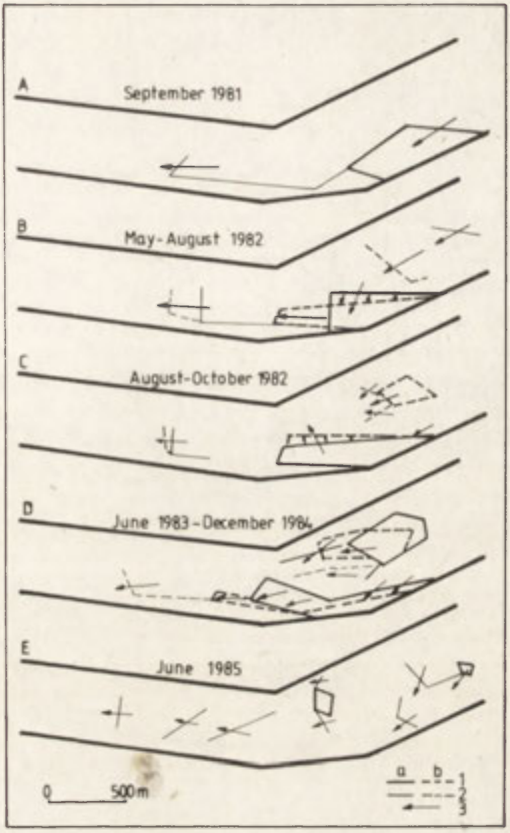

Fig. 5. Scheme of bar dynamics in the Vistula channel section at Nieszawa: 1 — emerged bar parts, 2 — submerged bar parts, a — original setting, b — secondary setting, 3 — main directions of bar shifting

(2) an improved and deepened section at Świecie where diagonal bars and pools occur in alternation (Figs 4C and 6).
The two section are typical and representative of part of the Vistula channel from the mouth of the Zgłowiączka river as far as the Vistula delta area (Fig. 2).

6.1. THE INITIALLY IMPROVED SECTION AT NIESZAWA

The over 2 km long Vistula channel section below Nieszawa is characterized by a changeable channel pattern to which incomplete improvements have been introduced. This variation is due to alterations in the channel width and to associated changes in the hydrodynamic conditions of the river (Figs 3 and 5). The upper part of the section about 600 m wide, which is controlled by groynes to become 500 m wide, changes at the angle of 25° into a 400 m wide straight lower section. Owing to these variations of river hydraulic geometry, local impounding of water and flow velocity loss occur above the point of channel narrowing. This leads to the activation of accumulation of the mineral transported. Thus, in the upstream section there is a concentration of bars, especially central and lateral ones, with numerous linguloid bars (Figs 3 and 5). These forms are the main studied objects.

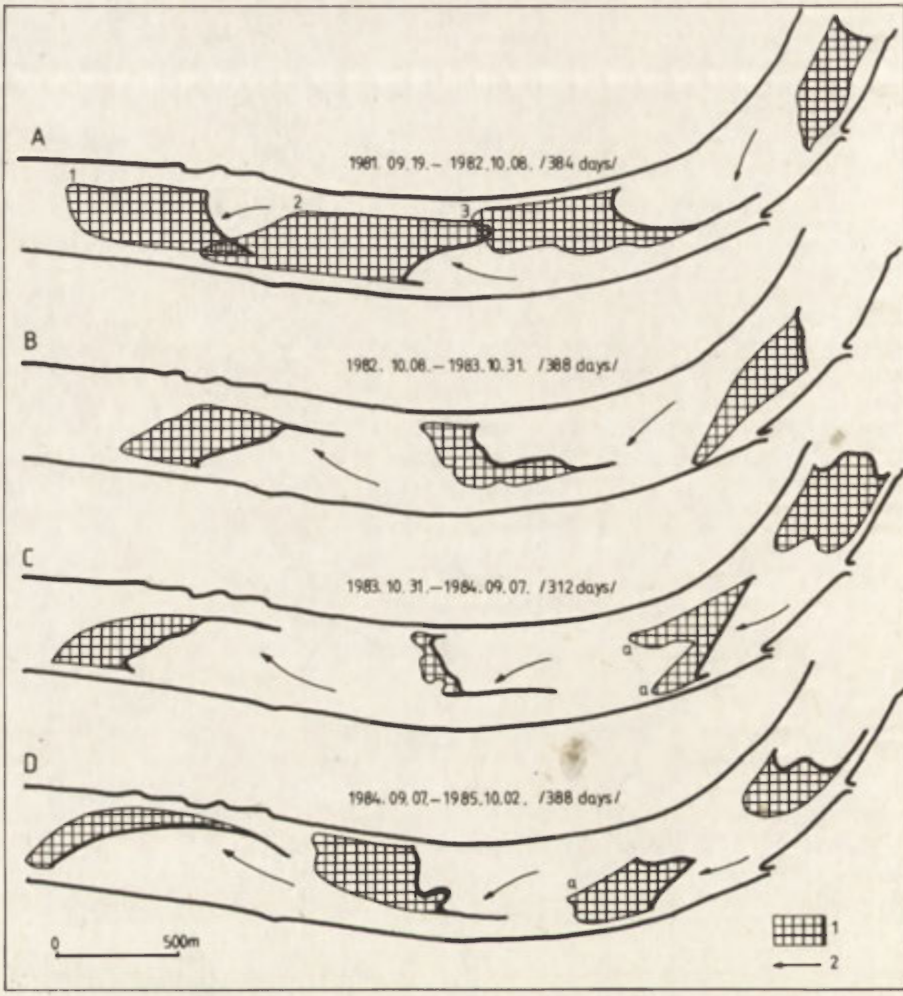

Fig. 6. Diagonal bar edges during particular periods (years) of investigation in the Vistula channel section below Świecie: 1 — superficial part of load transported as bars over a given period, 2 — course of thalweg, a — sites analysed in the text (subchapter 6.2)

The formation of a left-bank lateral bar is associated with hydrodynamic conditions of the river, determined by the setting of a central bar in the upstream widened section of the Vistula at the left bank, as well as with the occurrence of channel bends before the point of narrowing (local impounding of water). Initially, the lateral bar travelled downstream and occupied the major part of the channel width (Fig. 5A). In consequence of the concentration of the river currents adjusted to a smaller width of the channel, the left bank became eroded before the point of narrowing. The effects of erosion involved formation of dynamic linguloid bars and next, rapid downstream movement of the lateral bar face (Fig. 5B). The lower portion of the lateral bar was then separated from the channel bank by a 50—100 m wide limb. This process would have lasted as long as the moment of its disappearance at the point of channel narrowing, which took place after three years (Fig. 5E), if a central bar had not formed in the upstream section. Owing to the presence of this bar, river currents did not occur parallel to the banks of the lower section but struck the emerged part of the lateral bar. This part became rapidly eroded. The material was displaced towards the centre of the channel in the form of sandbars (Fig. 5C and D). It was only when the lateral bar approached the point of channel narrowing and when the upper central bar moved simultaneously downstream that the process ceased to operate in this direction. In consequence, the lateral bar was again liable to lateral erosion until it became completely eroded (Fig. 5E). The eroded material formed a bedform covered by sand waves, i.e. megaripples. Meanwhile a new central bar emerged close to the left bank in the upstream section under investigation (Fig. 5E). It moved towards the left bank in accordance with the hydrodynamic conditions of the river, resulting in a new lateral bar and thus, contributing to the completion of a four-year morphodynamic cycle in that section (Fig. 5A – E).

The four-year studies of the Vistula section below Nieszawa indicate a close relationship of the mesoform dynamics with changes in the hydrologic conditions of the river. However, this relationship holds for a low to average high water stage interval since no typical flood waves occurred then (Fig. 2). It is thus difficult to describe the bed conditions and their changes during periods of floods. This is the reason why the quoted morphometric data about bars refer to that period merely.

Both lateral and central bars attain similar heights and reach average water stages. After the flood, however, they may continue to grow until their heights exceed the average water stages by about 0.2 – 0.3 m. The lateral bars are up to 1100 m long and 230 m wide. Their existence is closely related to the channel morphology, i.e. the concave bank or to hydrotechnological installations, i.e. river groynes.

The central bars with maximum lengths of 650 m and maximum widths of 250 m occur in the channel sections which are more than 450 – 500 m wide, i.e . they are associated with either an unimproved channel or that in which a series of improvements were initiated. Their positions are subject to change at a distance of about 800 – 900 m from each other, with 320 – 420 m wide passages in between, at the average diurnal velocity of 0.41 – 0.60 m (Fig. 7: 0.5 m throughout the 24-hour period or 183 m per year, on the average).

Linguloid bars, in turn, exhibit much higher dynamics. They move at the diurnal velocity of 2.8 to 4.3 m. The maximum diurnal value observable is 7.76 m, which is

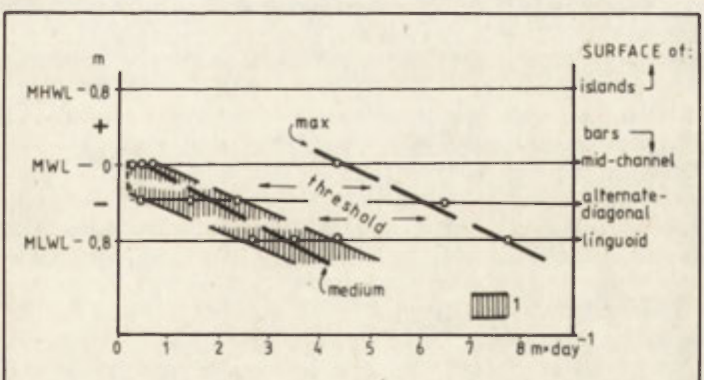

Fig. 7. Mean diurnal rate of bar schifting in the Vistula sections at Nieszawa and Świecie: 1 — zone of average velocities of bar shifting, MHWL — average high water stage, MWL — average water stage, MLWL — average low water stage

indicative of 2833 m per year. However, the annual value is overestimated since it is derived from the average value obtained for a short period of time when the bar movement was highly dynamic. The surface of linguloid bars is about 1.0 m lower than that of lateral and central bars. Their formation and growth is closely related to threshold changes in the hydrodynamic conditions of the river due to changing hydraulic geometry, i.e. channel morphology and hydrologic phenomena. They are built up of material derived from the eroded parts of central, lateral and diagonal bars.

Negative forms, i.e. pools, are found in the immediate vicinity of the above bars. As can be inferred from numerous bathymetric maps, cross-sections and longitudinal profiles of the channel, pools 4—6 m deep in relation to the average water stages occur in the section under investigation. However, they can attain a depth of 8 m in case local obstacles, e.g. groynes, have been created. In view of the above facts and of the heights of particular bars, sandy deposits move as 4.0—4.5 m thick central and lateral bars and 3.0—3.5 m thick linguloid bars. This is of great importance in estimating the bed load balance.

6.2. THE FULLY IMPROVED SECTION AT ŚWIECIE

An over 4 km long channel section below Świecie represents a typically improved channel. It extends from the mouth of the Tążyna river to as far as the Vistula delta area (Fig. 2). The channel is about 420 m wide in this section. The improvements have been made along the width of 375 m. The course of the thalweg is slightly sinuous with numerous straight reaches. A non-typical bend which becomes converted into a straight reach at the angle of 70° is located in the Vistula channel section under investigation (Fig. 4C and 6). Besides the prevailing mesoforms, i.e. diagonal bars, point-bars lie at the convex bank.

A growth cycle of alternate diagonal bars and the rate of shifting of their faces during the period of investigation are presented in Figure 6. In general, these forms are in the shape of elongated tongues, the faces of which attain the shallowest depth (Fig. 4C). Their surface is located at the river level at its middle and middle--low stage average and average-low stage (Fig. 7). Note should be made of the fact that no typical flood waves occurred during the period of investigation (Fig. 2). Hence, their heights may be underestimated. The culminating parts of the bars do not occupy the centre of the channel but lie at either the left bank or the right one in accordance with the alternate location of forms. They are separated from the banks by 30—150 m wide limbs. Only in the case of a bend are right-bank bars attached to the convex bank, resulting temporarily in lateral point-bars.

The downstream movement of the bars under investigation is parellel to the channel banks along straight reaches (Fig. 6). The load transport occurs chiefly over the form alone although some proportions of the deposit may be separated and carried by the lateral currents onto the next lower-lying bar. Hence, more than one bar should be studied in order to compute the magnitude of the load transport. Forms located on river bends diverge from this general scheme of diagonal bar movement. They either become then central bars with a system of linguloid bars with the so called horns in case they lie close to the left bank (Figs 4C point 2 and 6Ca) or form lateral point--bars if they are attached to the right bank. They are, however, transition bars which assume again their original form as they move downstream (Fig. 6). As has been mentioned above, diagonal bars exhibit alternate spacing of forms in the shape of elongated tongues. The distance between their faces range from 700 to 1600 m, most frequently 1200—1400 m (Fig. 7). They are accompanied by pools, the deepening of which may attain 11 m at the low water stage (Fig. 4C). Their length is similar to that of diagonal bars. The pools are 100 to 150 m wide and 5 to 7 m deep. The average height of bars between the upper edge of the culminating part and the flat bottom of a pools is about 4.5–5.0 m. The two types of forms result in a riffle and pool sequence typical of improved straight channels. According to R. Keller (1982), the distance between two successive bars is 5 to 7 times the channel width in the case of such a system of forms. If the values are 1300 and 420 m, respectively, for the section under investigation this ratio becomes 3. A considerable discrepancy between these values is indicative of an unstable system of channel forms which has been produced by the improvements. This instability can be linked to a too straight improvement route.

The dynamics of diagonal bars at average annual stages is shown in Figure 6. The omission is the stage after stage development of forms. It follows from the initial observations that the first period of investigation was characterized by increased dynamics of forms (Fig. 6A). Bars 1–3 shifted over a distance of 620–900 m (713 m on the average) during 384 days. This implies that the mean diurnal velocity of bar shifting is 1.86 m (Fig. 7). The maximum velocity of shifting of the diagonal bars has been found by investigating diagonal bar 1 which moved by as much as 390 m from 6 March to 5 May 1982, i.e. during 60 days, the average diurnal value being 6.5 m (Fig. 7).

In the next years the rate of shifting became slower (Fig. 6). The bars moved at a velocity of 0.42 to 1.6 m throughout the 24-hour period. An uneven movement rate and changeable magnitude of load transport are largely due to changing hydrologic conditions on the Vistula. There was considerably higher water dynamics over the first period of investigation. Two flood waves 718 and 592 cm high were recorded at the Toruń gauging station on 8 January and 10 March 1982, respectively. There were hardly any high stages in the next three years (590 cm recorded on 15 March 1983) but there was a concurrent tendency toward uniform water stages (Fig. 2). Thus, the rate of bar shifting bears a close relationship to the Vistula hydrologic conditions.

6.3. THE RATE OF SHIFTING OF THE VISTULA CHANNEL MESOFORMS

Rzhanitsin (1984) suggests that river transport in the form of mesoform shifting is not uniform but occurs cyclically in a series of jerks. According to the Soviet research workers, this discontinuous load transport is linked to the so called channel-forming river discharges (e.g. Chalov and Byelyi 1975, 1984; Vlasov and Chalov 1981). This concept refers to discharge values (water stages) at which the most intensive channel modification and thus, accelerated load transport take place. The above workers give 1 to 3 such ranges of channel-forming discharge values for the Soviet rivers and point out that flood discharge is not always most effective in each case. Low stages may also have a marked effect on load transport. Some British-American workers consider this fact similarly by identifying thresholds in the fluvial environment (e.g. Schumm 1973 and 1977; Bull 1979; Coates and Vitek 1980; Howard 1980; D. G. Smith 1980; Winkley et al. 1984; Ritter 1985).

As field studies have shown, the transport of load in the form of bars on the Vistula occurs in two modes; it is uniform and cyclical. The continuous transport of load in the form of central, diagonal and linguloid bars proceeds at the velocities of 0.4–0.6, 0.4–2.4 and 2.8–4.3 m, respectively, throughout the 24-hour period (Fig. 7, No 1). This uniform shifting of forms is disturbed by thresholds of bar growth. The load transport becomes then temporarily accelerated to the velocities of 4.33, 6.5 and 7.76 m throughout the 24-hour period for particular bars, respectively (Fig. 7). Unfortunately, the non-occurrence of floods makes it difficult to provide full characteristics of this transport (Fig. 7).

The load transport in a series of jerks is due to a relationship between bar surface height and water table level, affected by the river hydrological regime (discharge). The moment the water table either enters or leaves the surface of bars, islands or a floodplain and is capable of transporting the load is a threshold for the development of channel processes. In the case of the section at Nieszawa (subchapter 6.1) thresholds for accelerated load transport occur at discharge values characteristic of the heights of the surface of central-lateral and linguloid bars and a floodplain and in the case of the Świecie section (subchapter 6.2), they are identified at discharge values characteristic of the heights of the surface of diagonal bars and a floodplain and sporadically, of linguloid bars (Figs 7 and 8). From the above it follows that a section of a braided or partially improved river is characterized by three channel-forming discharges, i.e. thresholds, for channel processes, whereas an improved section exhibits merely two thresholds (Fig. 8).

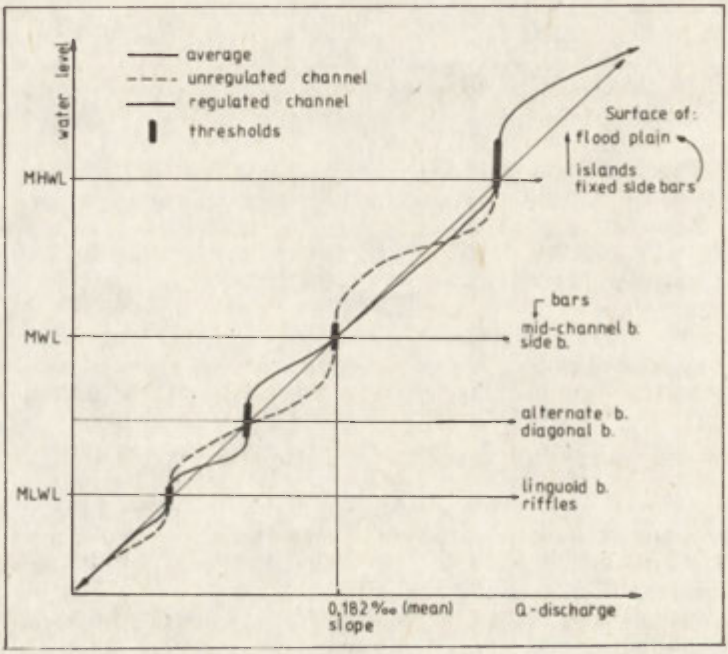

Fig. 8. Scheme of threshold channel process development on the lower Vistula (see Fig. 7 for key to symbols MHWL, MWL and MLWL)

A section below the Włocławek water dam is a specific example of threshold channel adjustments (Fig. 2). In consequence of diurnal water level fluctuations approaching 3 m, there was a rise in the frequencies of thresholds from several throughout the year to a few during a day and thus, increased load transport occurred. This led to erosion of all bars below the dam. They form only at a distance of over 16 km below the reservoir. This evidence points to the importance of channel-forming discharges and threshold load transport in the channel process.

7. LOAD TRANSPORT BALANCE IN THE LOWER VISTULA CHANNEL

The computed values based on morphometric data from channel mesoforms and average velocities of shifting (subchapter 6.2) show that the Vistula carries about 940 thousand cubic metres of sandy material in the form of bars along the improved section below Świecie throughout the average hydrological year, i.e. 1981 – 1982, the first period out of those under investigation (Fig. 6A). In the next three years the magnitude of load transport became reduced annually to 420—460 thousand cubic metres as a result of lower discharge values and the non-occurrence of a flood wave. From the mean figures, it can be inferred that the Vistula river carries 420 thousand to 1.0 million cu m of material along the improved reaches in the dry year and in the average year, respectively. The non-occurrence of flood waves and the absence of a more humid spell did not allow full characteristics of this problem to be provided. This is also the reason why no computations were made for the section below Nieszawa. It has been merely assumed that it resembles the unimproved section above the Włocławek reservoir and so it must be characterized by similar-magnitude load transport. According to Skibiński (1985) who quotes Sliwinski's computations based on the analysis of cross-sections about 1.4 million cu metres of sediments were deposited annually in the Włocławek reservoir in the average years 1971—1978, whereas the rate of sedimentation was 4.0 million cubic metres per year at a humid spell.

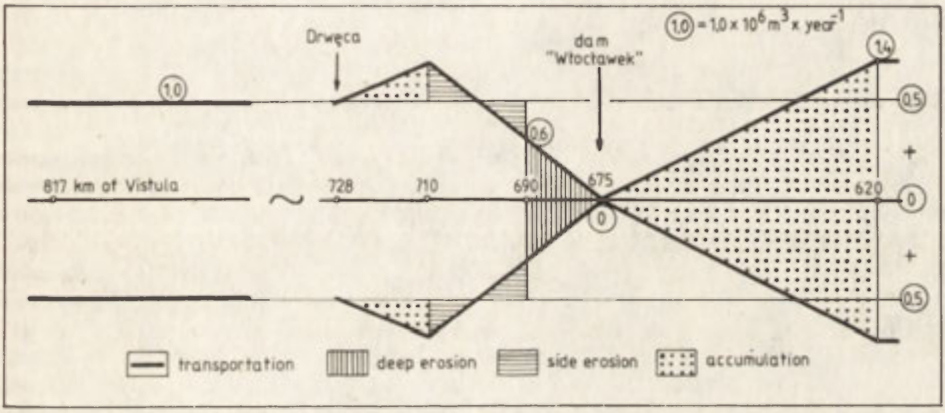

Fig. 9. Load transport balance for the Vistula from Kępa Polska to Świecie in the average year

It has been established that the magnitude of load transport is 1.4 and 1.0 million cu m of material along the unimproved and improved sections, respectively, throughout the average hydrological year. These data provide the basis of construction of the load balance scheme (Fig. 9) for the entire section under investigation (Fig. 2), with reference to an erosional-accumulation process operating below the Włocławek water dam as well.

8. FINAL REMARKS

The results of multi-year studies of channel processes operating on the lower Vistula indicate that the section under investigation exhibits changeable growth and dynamics of mesoforms and thus, varies in respect of load transport characteristics. This is due to many economic causes, i.e. man's involvement in these events. The initial cause involved stage-made different-intensity improvements. In the recent hundred years they led to the magnification of differences between the upper reaches which retained their original features of a braided river and the lower reaches where the channel became narrowed and deepened by about 1.5 m, a new floodplain level formed and a new sequence of diagonal bars and pools was produced. The aim of the postwar improvements was to get rid of those differences. The preliminary improvements made then were, however, stopped because a series of water dams were, planned on the lower Vistula. The Włocławek water dam constructed in 1962–1968 not only failed to improve the waterway but even contributed to its deterioration. It was intended to work in a system of nine planned reservoirs. In consequence, it inhibited the transport of about 1.4 million cu m of sediment load throughout the year and intensified the process of deep erosion below the dam. On the average, about 0.6 million cu m of sand-gravel deposit was eroded annually over the 17-year period and was next accumulated in a section lying at a distance of 16 km downstream of the dam (Fig. 9).

As can be inferred from brief characteristics of the load transport balance (Fig. 9), the creation of the reservoir led to greater variation of the erosional-accumulation process operating in that section. The situation may be radically changed in case a series of dams are constructed on the lower Vistula or intensive improvements are made in a braided section and dilapidated old hydrotechnic installations become replaced.

REFERENCES

Antropovski V. Ye., 1969, Gidromorfologicheskiye zavisimosti i ikh dalnyeishiye razvitye, *Trudy GGI* 169, 34–86.

Babiński Z., 1981, Wstępna charakterystyka morfologiczna dna koryta Wisły dolnej na odcinku Włocławek-Chełmno w świetle map i fotointerpretacji (Sum.: Preliminary morphological characteristics of the Lower Vistula bed on the section Włocławek-Chełmno in the light of maps and photointerpretation), *Gospodarka Wodna* 4–5, 105–109.

Babiński Z., 1982, Procesy korytowe Wisły poniżej zapory we Włocławku (Sum.: The influence of the water dam in Włocławek on fluvial processes of the Vistula River), *Dokumentacja Geograficzna* 1–2.

Babiński Z., 1984., The effects of human activity on changes in the Lower Vistula channel, *Geographia Polonica* 50, 271–282.

Barwis J. H. 1978, Sedimentology of some South Carolina tidalcreek point bars, and a comparison with their fluvial counterparts, in: A. D. Miall, ed., *Fluvial sedimentology*, Calgary, Alberta, 129–161.

Bull W. B. 1979, Threshold of critical power in streams, *Geological Society of America Bulletin* 90, 453–464.

Cant D. J. Walker R. G., 1978, Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada, *Sedimentology* 25, 625–648.

Carson M. A. 1984, Observations on the meandering-braided river transition, the Canterbury Plains, New Zealand, *New Zealand Geographer* 40, 1, 12–17.

Church M., Jones D., 1982, Channel bars in gravel-bed rivers, in: R. D. Hey, J. C. Bathurst, C. R. Thorne, eds, *Gravel-bed rivers*, John Wiley and Sons Ltd, 291–338.

Coates D. R., Vitek J. D. 1980, Perspectives on geomorphic thresholds, in: D. R. Coates, J. D. Vitek, eds, *Thresholds in geomorphology*, George Allen and Unwin, London, 3–23.

Collinson J. D., 1970, Bedforms of the Tana river, Norway, *Geografiska Annaler* 52A, 31–56.

Chalov R. S., Byelyi B. V., 1975, Rayonirovaniye territorii Sibirii po kharakteru rusloformiruyushchikh dyeyatelnosti riek, *Meteorologiya i Gidrologiya* 12, 76–88.

Chalov R. S., Byelyi B. V., 1984, Regionalnye osobyennosti rusloformiruyushchikh razkhodov vody na ryekakh Sredney Azii, *Izvestia AN SSSR* 5, 107–116.

Falkowski E., 1967, Ewolucja holoceńskiej Wisły na odcinku Zawichost-Solec i inżyniersko-geologiczna prognoza jej dalszego rozwoju (Sum.: Evolution of the Holocene Vistula from Zawichost to Solec with an engineering-geological prediction of further development), *Z badań geologiczno-inżynierskich w Polsce* 4, *Biuletyn Instytutu Geologicznego* 198, 57–150.

Ferguson R. I., Werritty A., 1983, Bed development and channel changes in the gravelly river Feshie, Scotland, *Sedimentology* 6, 181–193.

Howard A. D., 1980, Thresholds in river regimes, in: D. R. Coates, J. D. Vitek, eds. *Thresholds in geomorphology*, George Allen and Unwin, London, 227—258.
Keller R., 1982 Die Wasserbilanz in Deutschland, *Zentralblatt fur Bakteriologie, Parasitenkunde, Infektionskrankenheiten und Hygiene*, Abt. I, Orig. B. 172, 122—133.
Kellerhals R., Church M., Bray D. I., 1976, Classification and analysis of river processes, Proceedings of the American Society of Civil Engineers, *Journal of the Hydraulics Division* 102, No HY 7, 813—829.
Koc L., 1972, Zmiany koryta Wisły w XIX i XX wieku między Płockiem a Toruniem (Sum.: 19th and 20th centuries' changes in Vistula channel between Płock and Toruń), *Przegląd Geograficzny* 44, 4, 703—719.
Kondratiyev N. Ye., Popov I. V., 1967, Metodicheskiye predposylki k postanovke setevikh nabludenii za ruslovym protsessom, *Trudy GGI* 144, 118—149.
Kondratiyev N. Ye., Popov I. V., Snishchenko B. F., 1982, *Osnovy gidromorfologicheskoi teorii ruslogo protsessa*, Gidrometeoizdat, Leningrad.
Kopczyński S., 1982, *Stosunki wodne Basenu Grudziądzkiego i jego otoczenie* (Sum.: Water relationships in the Grudziądz Basin and its surroundings), Studia Societatis Scientiarum Torunensis, Sec. C, 9, 4.
Krigström A., 1962, Geomorphological studies of sandur plains and their braided rivers in Iceland, *Geografiska Annaler* 3—4, 328—346.
Kurpianov V. V., Kopaliani Z. D., 1979, Resume of research into river channel changes conducted in the USSR, *Bulletin des Sciences Hydrologiques* 24, 345—349.
Levey R. A., 1978, Bedforms distribution and internal stratification of coarse-grained point bars, Upper Congaree River S. C., in: A. D. Miall, ed., *Fluvial sedimentology*, Calgary, Alberta, 105—129.
Machalewski W., Miłkowski M., Rozwadowski J., 1974, Wpływ stopnia wodnego we Włocławku na warunki żeglugowe Wisły dolnej (Sum.: Impact of the Włocławek water step on the navigation conditions on the Lower Vistula), *Gospodarka Wodna* 3, 114—117.
Popov I. V., 1977, *Zagadki rechnogo rusla*, Gidrometeoizdat, Leningrad.
Ritter D. F., 1985, Regional landform thresholds, *NASA Conference Publication, Global Mega-Geomorphology*, Arizona, 23—26.
Rzhanitsin N. A., 1984, Osnovi teorii rusloformiruyushchikh protsessov, in: *Dinamika i termika riek i vodokhranilishch*, Nauka, Moskva, 124—139.
Schumm S. A., 1973, Geomorphic thresholds and complex response of drainage systems, in: M. Morisava, ed., *Fluvial geomorphology*, State University New York, Binghampton, N. Y., 253—284.
Schumm S. A., 1977, *The fluvial system*, John Wiley and Sons Ltd, New York.
Schumm S. A., 1985, Patterns of alluvial rivers, *Annales Earth Rev. Planet. Sci.* 13, 5—27.
Shen H. W., 1982, The wandering rivers, Water International Publications, IWRA by Elsev. Sequoia S. A., Lausanne 7, 10—17.
Skibiński J., 1985, Charakterystyka morfologiczna Wisły na odcinku Modlin-Włocławek oraz zmiany powstałe w ukształtowaniu dna w wyniku spiętrzenia wody w Zbiorniku Włocławskim (Morphological characteristics of Vistula channel from Modlin to Włocławek and bottom changes brought about by water damming in Włocławek reservoir), in: W. Majewski, ed., *Powódź zatorowa na Wiśle w rejonie Zbiornika Włocławek w zimie 1982 r.*, Wydawnictwa Geologiczne, Warszawa, 35—49.
Smith D. G., 1980, River ice processes: Thresholds and geomorphologic effects in northern and mountain rivers, in: D. R. Coates, J. D. Vitek, eds, *Thresholds in geomorphology*, George Allen and Unwin, London, 323—343.
Smith N. D., 1978, Some comments to terminology for bars in shallow rivers, in: A. D. Miall, ed., *Fluvial sedimentology*, Calgary, Alberta, 85—89.
Task Force on bed forms in alluvial channels of the committee on sedimentation, 1966, Nomenclature for bed forms in alluvial channels, Proceedings of the American Society of Civil Engineers, *Journal of the Hydraulics Division* 92, No. HY 3, 51—64.

Tomczak A., 1971, *Kępa Bazarowa na Wiśle w Toruniu w świetle badań geomorfologicznych oraz archiwalnych materiałów kartograficznych* (Res.: L'ilot du Bazar sur la Vistule á Toruń a la base des recherches geomorphologiques et des materiaux cartographiques anciens), Studia Societatis Scientiarum Torunensis, Sec. C, 7, 6.

Winkley B. R., Schumm S. A., Mahmood K., Lamb M. S., Linder W. M., 1984, *New developments in the protection of irrigation, drainage and flood control studies on rivers*, International Commission on Irrigation and Drainage, Symposium, Fort Collins, 69—111.

Vlasov B. N., Chalov R. S., 1981, Rusloformiruyushchiye raskhody rieki Volgi, *Izvestia AN SSSR* 6, 75—81.

Znamenskaya N. S., 1976, *Donniye nanosy i rusloviye protsessy*, Gidrometeoizdat, Leningrad.

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THE ROLE OF PRECIPITATION WATERS IN DECALCIFICATION OF LOESSES IN WESTERN ROZTOCZE (THE CATCHMENT AREA OF THE UPPER SANNA RIVER)*

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INTRODUCTION

Considerable areas of south Poland are covered by loesses (Jahn 1950, 1956; Maruszczak 1972b, 1980; Jersak 1973). The content of carbonates in the mineral composition of these deposits varies from a few to over ten percent.

The purpose of this paper was to determine the role of precipitation waters in decalcification of carbonate loesses on the example of the upper Sanna catchment area. Leaching of loess-like deposits is not discussed in this paper.

METHODS

In the studies the following procedures were applied: a) determination of calcium carbonate content in loess at various levels and in the underlying calcium rock; b) laboratory examinations of the solubility of calcium carbonate in distilled water and magnesium contained in carbonate loesses in samples with disturbed structures and in monoliths; c) field studies of leaching carbonates out of loesses by precipitation waters; d) determination of the chemical composition of spring waters occurring in the area studied and in the Sanna river water.

Methods of quantitative analysis as well as conductometric and potentiometric measurements were also used.

PHYSIOGRAPHY OF THE UPPER SANNA CATCHMENT AREA

The studies cover the catchment area of the upper Sanna river situated in the south-east part of Poland, in the contact zone of Western Roztocze and the Lublin Upland (Fig. 1).

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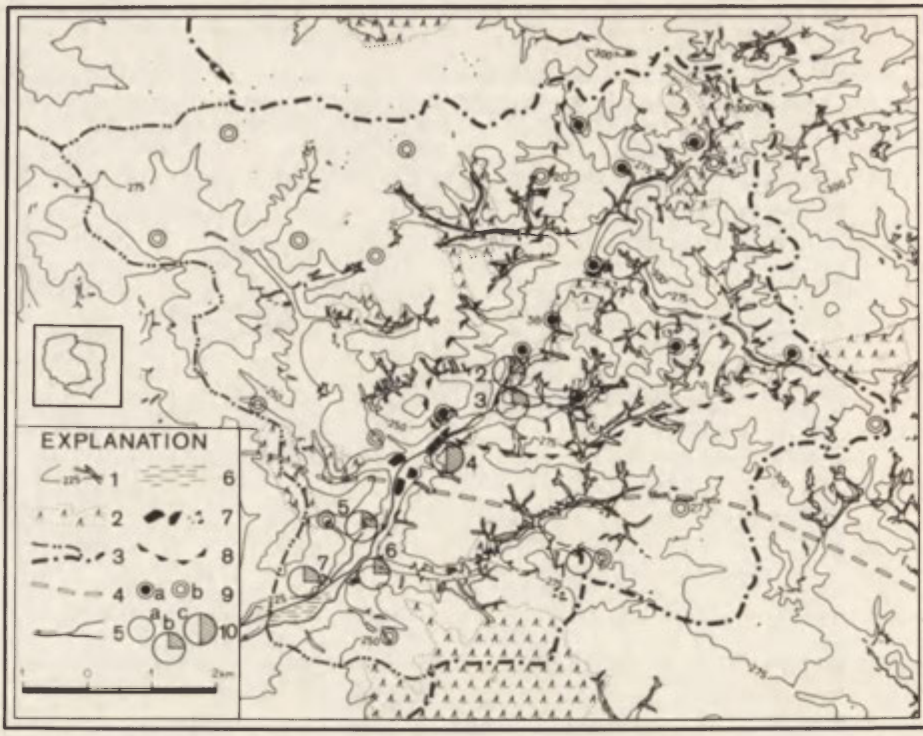

Fig. 1. The catchment area of the upper Sanna river: 1 — contour lines and ravines, 2 — woods and coppices with shrubs, 3 — watersheds of II and III order, 4 — contact zone of the water-bearing cretaceous horizon (form N) and Tertiary-Miocene horizon (from S), 5 — water streams in the Sanna river valley, 6 — wet land areas, 7 — fish-nurseries and other depressions filled with water (*stoczki*), 8 — southern borderline of deep carbonate loesses in the catchment area, 9 — more important drillings made by the author; a — in carbonate loesses, b — in decalcified loesses (and in other deposits), 10 — springs in output categories (in $dm^3 \cdot s^{-1}$): a — below 1,0, b — 1.0—50.0, c — over 100.0

The catchment area up to the water-level recorder closing the region studied is 76.5 km^2, and its dry part above the spring No 2 (Fig. 1) is 29 km^2. The Sanna river is 51.3 km long and flows directly into the Vistula river as its right tributary. The length of the Sanna within the catchment area is about 6.0 km.

The bedrocks of significant importance in hydrological processes in the catchment area are upper Cretaceous (Campanian) deposits, which are overlaid by Tertiary limestones of various thickness. South of the borderline of underwater occurrence in Neogene limestones (Fig. 1), the thickness of these rocks does not exceed 10—20 m. North of this line the deposit of limestones is distinctly thinner, and it disappears completely in many places.

The Cretaceous-Tertiary rock is overlaid by Pleistocene deposits: silt loesses, sandy-silty, clay-silty and loam rocks. Because of a considerable distribution of loesses and their thickness (up to 9.0 m), being the objective of this paper, these rocks should be given a greater attention.

J. Malinowski (1964) documented in Roztocze five loess horizons, the most widespread of which are horizon II and IV. In the report of H. Maruszczak (1972b,

1980) horizon II and IV correspond to the terms "older loess" and "younger loess" respectively. According to this author younger loesses of the last glaciation are the most common in Poland's south-east area, which overlie older loesses. These deposits are characterized by vertical lithological variations.

On the basis of textural features, H. Maruszczak (1972 a) distinguished unstratified and cryptostratified loesses of aeolian and aeolian-deluvial facies, as well as stratified loesses of deluvial, solifluctional and colluvial facies, classified with regard to the sedimentation environment as subaerial and subaqueous loesses. In the catchment area of the upper Sanna river there occur both unstratified and stratified loesses and, according to morphological classification — top, slope and valley ones.

Field and laboratory studies made it possible to determine the range of carbonate loesses (Fig. 1) covering the catchment area in about 32% (24 km^2).

In water infiltration into loesses the shape of their grains and diameters play an important role. According to H. Maruszczak (1972 b) the average grain size of younger loess ranges from 0.020—0.035 mm, but the upper profile layers have slightly coarser grains. Another important feature is general porosity which was estimated at 25—51% by J. Malinowski (1964). From the studies of J. Tokarski (1951), and J. Tokarski and K. Oleksynowa (1951) it appears that quartz grains 0.1—0.01 mm in diameter of silt fraction are always sharp-edged, which favours the existence of capillary systems in aeolian loesses. In S. Zawadzki's opinion (1972) the total porosity in carbonate systems is constituted in 18% by micropores ($\varnothing<2\cdot10^{-7}$ m), in 64% by mezopores ($\varnothing 2.10^{-7}-3\cdot10^{-5}$ m) and in 18% by macropores ($\varnothing>3\cdot10^{-5}$ m).

In decalcification of loesses the basic role is played by infiltration of free water which can percolate gravitationally into water-bearing horizons almost only through macropores. However, both the number and diagrams of these channels determine low values of the filtration and permeability coefficients. Taking this into consideration, Z. Pazdro (1983) determined loesses as rocks of medium parameters: the filtration coefficient ranging from 10^{-4} to 10^{-5} $m\cdot s^{-1}$, and permeability coefficient from 10.0—1.0 darcys. J. Malinowski (1974) empirically determined the values of filtration coefficients for loesses of Western Roztocze, ranging from 4.7×10^{-6} to 5.0×10^{-7} $m\cdot s^{-1}$ for aeolian loesses and from 1.15×10^{-5} to $1.1\cdot10^{-6} m\cdot s^{-1}$ for solifluction loess. From the hydrological studies carried out in this region (Janiec 1984) it appears that these values should be considered as indices. It can also be concluded from A. I. Budagovski's studies (fide: Nazarov 1981) that the differentiation of the filtration coefficient is big even in a small area (1 m^2), which, expressed e.g. by the variation standard, is 0.50. In the process of chemical changes of precipitation waters percolating gravitationally through the loess cover, alterations occurring in carbonate loesses are most significant, but they are less significant in decalcified loesses. The presence of specific macroelements in water depends on the chemical composition of the dissolved rocks, whereas the kind of minerals and their susceptibility to leaching limits the mineralization horizon and their molar ratios (ions) in water, at least in the period of its infiltration through the aeration zone.

Silica (SiO_2) predominates in the chemical composition of loesses, the content of which is about 70% by weight in unweathered deposits. Aluminium, iron, calcium and magnesium oxides occur in considerably lesser amounts. These compounds form minerals in loesses, which include quartz low-soluble in water, aluminium oxides, silts of feldspars slightly faster or readily soluble in aggressive infiltration waters, calcium and magnesium carbonates, iron oxides and clay minerals (hydromicas).

From the determinations of carbonates in the loess profiles of the upper Sanna catchment area it appears that: a) there occur carbonate and decalcified loesses (totally carbonate-free deposits have not been recorded), b) carbonate loesses are considerably decalcified, most frequently down to the depth of 3 m, c) the amount of carbonates most frequently found in unweathered loesses is 5.0—6.0% of $CaCO_3$ (Fig. 2A) and is usually

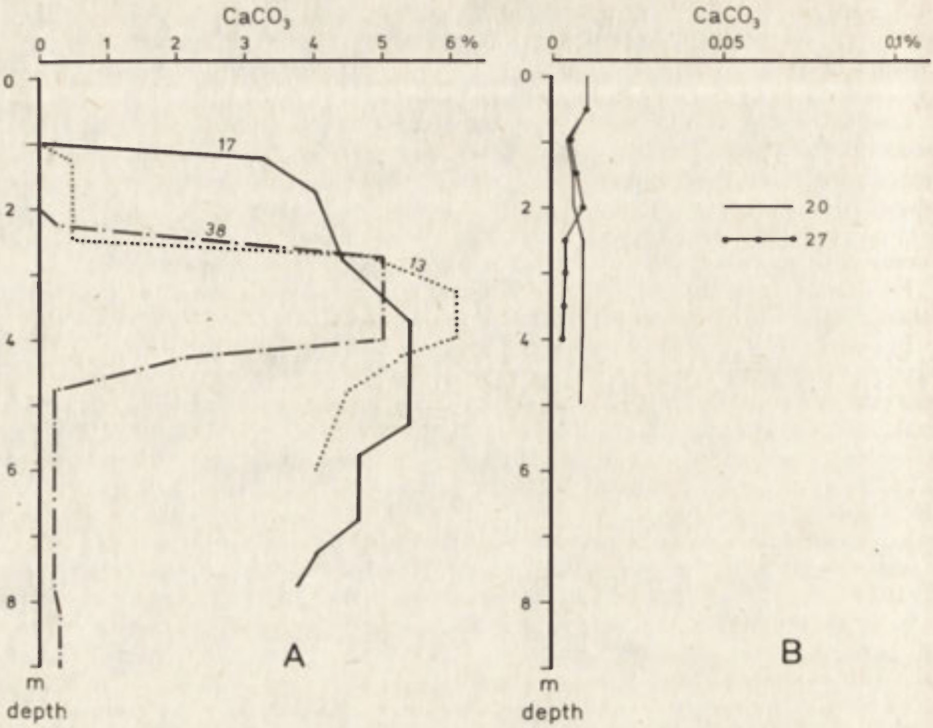

Fig. 2. Content of calcium carbonate ($CaCO_3$) in selected profiles of carbonate loesses (A) and decalcified loesses (B); 13, 17, 20, 27, 38 — numbers of profiles localized in Fig. 1.

found at a depth of 3.0—6.0 m, d) in the NW and S part of the area studied loess-like deposits occur which are considerably decalcified, containing below 0.1% of $CaCO_3$ (Fig. 2B), but leaching of these rocks by precipitation waters is not discussed in this paper.

DECALCIFICATION OF LOESSES

Studies of decalcification of loesses are interesting both from theoretical and practical point of view. They concern the mechanism of this phenomenon, the amount of displaced carbonate matter, the pedological (degradation of soils) and geomorphological (shape changes of inclines) consequences, changes of the quality of waters infiltrating through the aeration zone, as well as the possibility of predicting changes of the geographical environment of loess areas.

Dissociation of carbonates contained in loesses is connected with aggression of infiltration waters, which depends on the total content of CO_2 and H_2CO_3. The investigation of the chemical equilibrium of dissolved carbonates can be facilitated by assuming the existence of ideal water, i. e. such a system in which only H_2O — CO_2 — $CaCO_3$ occur. Water pH is then an index of carbonate equilibrium. Negligence of the dissociation of other substances in water not belonging to this system does not eliminate the expectation that they increase the value of the solution ionic strength. Therefore, the activity coefficients of ions, their share in the solution ionic strength should be calculated from Debye—Hückle's law, and corrections in calculations of different constants, particularly of the solubility ratios should be done (Barnes 1964; Hem 1985; Langmuir 1968; Muxart and Birot 1977). It is known that $CaCO_3$ solubility is also effected by other salts contained in the solution, particulary those with their common ion, which decrease the solubility of calcium carbonate.

Inaccuracies in investigations of decalcification processes of loesses, resulting from the complications presented are not dealt with in this paper.

Taking advantage of the ideal system in further considerations of loess decalcification, the relation $H_2O-CO_2-H_2CO_3-pH$ should be clarified. Carbonic acid (H_2CO_3) may occur in water only at a considerable dilution, and its quantity depends on the amount of carbon dioxide dissolved in water. In aqueous medium H_2CO_3 dissociates in two stages, and the dissociation constants are: $K_1 = 3.2313 \cdot 10^{-7}$ $mol \cdot dm^{-3}$ and $K_2 = 3.2360 \cdot 10^{-11}$ $mol \cdot dm^{-3}$. K_1 and K_2 values for 10°C were determined as arithmetic means of those obtained from various sources (fide: Muxart and Birot 1977). The konwledge of K_1 and K_2 makes it possible to determine the potential content of aggressive CO_2 ($CO_2 + H_2CO_3$) in initial water, knowing the values of HCO_3 and H^+ which can be easily measured. When equilibrium is established it is also possible to determine CO_2 content by solving the equations of I and II degree of carbonic acid dissociation, as well as the equation of $CaCO_3$ solubility ratio whose value for 10°C is $S_{CaCO_3} = 3.9367 \cdot 10^{-9}$ ($mol \cdot dm^{-3}$).

LABORATORY STUDIES OF LOESSES WITH DISTURBED STRUCTURES

The initial programme was designed to determine temporal changes of the reaction of distilled water which was in contact with carbonate loesses and limestone *opoka* of upper Cretaceous (Campanian). In the upper Sanna catchment area chiefly limestones constitute the bedrock of loesses. At the same time the water-bearing cretaceous horizon is alimented by the precipitation waters infiltrating through the loess cover. The results of one of the laboratory experiments are illustrated in Fig. 3.

From them information was obtained on changes of water reaction (pH) as carbonate equilibrium index during the first 90 minutes of water contact with the rocks mentioned above in conditions of open system at 18°C and continuous inflow of

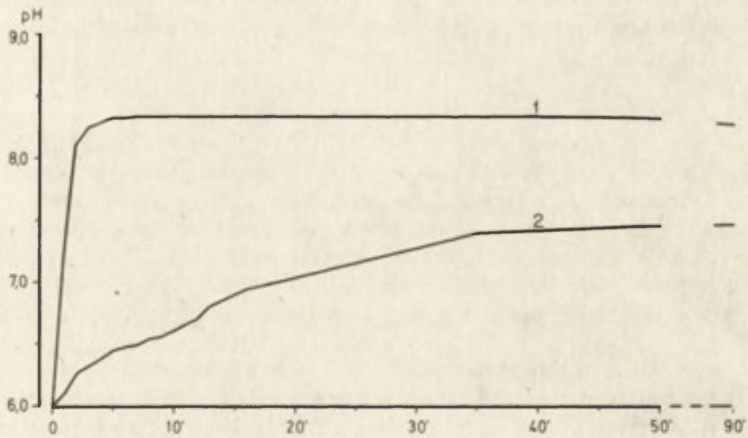

Fig. 3. Changes of water reaction (pH) dissolving loess (1) and cretaceous rock (2) in the time (T) of initial 90 minutes of water contact with rocks

atmospheric CO_2. In this experiment various development of the phenomenon studied was observed in both cases; pH changed from 6.0 to 8.32 in a water sample with loesses after 7 minutes, whereas in another sample with limestone water reaction increased only by 0.48 pH in the same time.

In another experiment 49 carbonate loess samples of 50 g each were weighted and flooded with 100 ml of redistilled water. The specific conductivity of initial water (at 10°C) was 2.0 $\mu S \cdot cm^{-1}$ and the pH = 5.84. The studies were carried out at a constant temperature (10°C); water reaction was measured potentiometrically, and specific conductivity conductometrically. On the frist day all samples were examined twice and then the examination was repeated after 1 day, 2, 3, 5, 7, 9, 13 and 15 days. The studies were finished when in two successive measurements no change (in plus) was found in water conductivity, which coincided with determination of pH. It should be mentioned that in the experiment arbitralily was assumed: a) use of redistilled water, b) disintegration of loess samples, c) mixing of samples after every measurement.

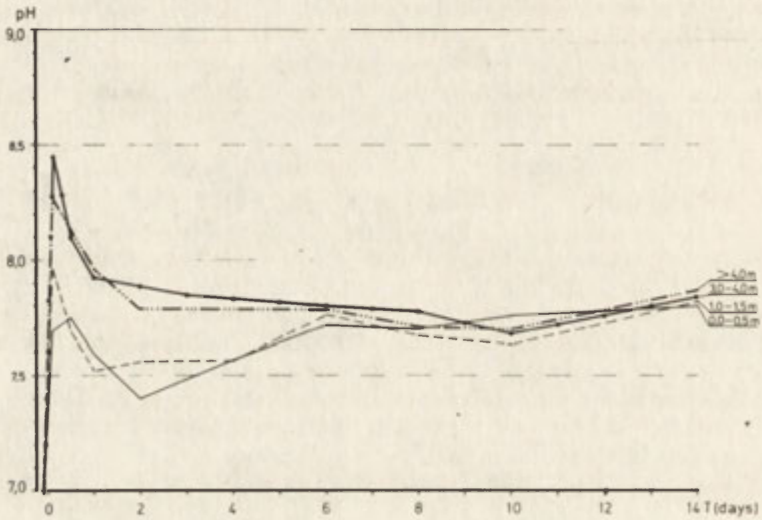

Fig. 4. Changes of water reaction (pH) in the time T accompanying dissolution of carbonate loesses with disturbed textures; mean values from samples of selected depths

The results of water reaction changes are presented in Fig. 4. It is to be added that the studies also comprised the roof (almost decalcified) parts of these profiles. For illustration the results from five depth intervales were chosen. From the curve courses it appears that the reaction of the waters being in contact with loesses of different horizons was alkalic during the experiment. In the first 6 days duality in the course of pH curves can be seen — a little lower values in the upper parts of the profiles. After this period the values of all samples were stabilized on the level of pH 7.6—7.8 (buffer). Attention is drawn by the fact that in all water samples the evolution of the features lasted for about two weeks.

In small intervals of variance the specific conductivity and mineralization of waters are almost directly proportional. However, the relationship of these values with the level of water pH is indirectly included in the dissociation schemes, because the number of dissociated $CaCO_3$ molecules essentially depends on the content of H_2CO_3 ($H_2CO_3 + CO_2$) in water. In the case of calcium carbonate dissociation the increase by

1 mval·dm^{-3} of calcium and hydrogen carbonates is accompanied by increased water conductivity (at 10°C) by 39 and 32 μS·cm^{-1} respectively.

From the experiment carried out it appears that increased water conductivity was connected with a higher percentage of carbonates content in the rock, hence the upper layers of profiles showed a relatively low electric conductivity. At full saturation of waters with carbonates under laboratory conditions maximum conductivity exceeded 160–200 μS·cm^{-1}, and the mean value (at 10°C) was 145 μS·cm^{-1}. The following reaction rate was observed: after 5–6 h the average saturation of waters was 28%, 61% after 24 h, 79% after 2 days and 90% after 5 days.

When pH and water conductivity were measured, $Ca^{2+}+Mg^{2+}$ were determined in all samples.

Among 49 waters with samples of carbonate loesses, examined for the content of calcium and magnesium the highest (over 70%) was found in the interval 1.25–1.75 mval·dm^{-3}. The mean arithmetic content of $Ca^{2+}+Mg^{2+}$ in the whole study series was 1.47 mval·dm^{-3}. This value converted into weight units of calcium carbonate was 74 mg·dm^{-3} of $CaCO_3$. This is illustrated in Fig 5. It shows the determination results of total hardness converted into weight units of Ca^{2+}. It should be stressed that to illustrate the solubility of carbonates in water the same profiles were chosen for which exemplary $CaCO_3$ content in rocks was shown (Fig. 2). The distribution of total hardness (as Ca^{2+}) in various parts of profiles of carbonate loesses is presented by curves (profiles) No 13, 17 and 38, whereas Nos 20 and 27 refer, as examples, to

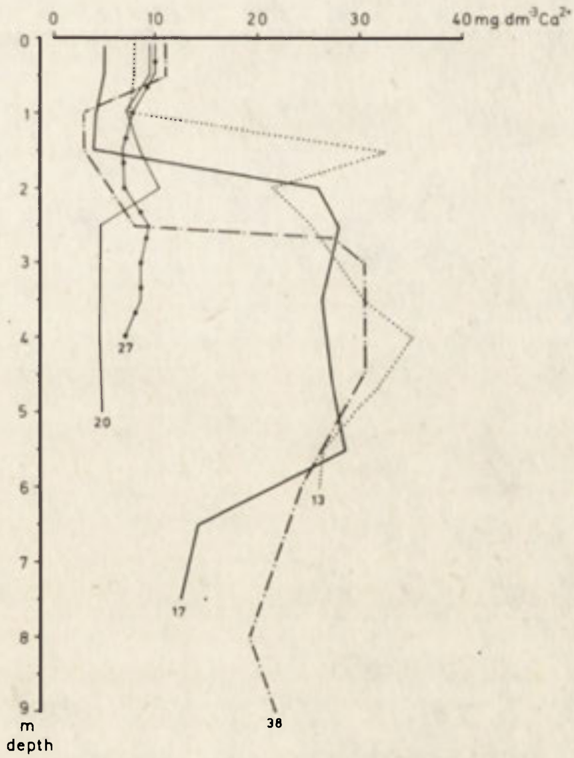

Fig. 5. Total hardness (presented arbitrarily in gravimetric units of Ca^{2+}) in waters dissolving loesses with disturbed textures; Nos 13, 17, 38 – carbonate loesses, 20 and 27 – decalcified loess-like deposits

loess-like deposits considerably decalcified. Attention is drawn by the fact that a leached horizon with a small amount of carbonates occurs in carbonate loesses at a depth of 1.0 – 1.5 m. Below this depth or a little lower the curves show a rapid increase of calcium content. It is characteristic that in the depth interval 3.0 – 6.0 m there occurs the greatest possibility of $CaCO_3$ leaching by infiltrating waters. It is also worth noting that this depth interval shows the highest content of carbonates in loesses.

EXPERIMENTS ON LOESSES WITH NATURAL STRUCTURES

After obtaining initial information concerning solubility of carbonates contained in loesses, another programme of laboratory experiments was completed by using fresh loess monoliths prepared in the field in 50 and 100 cm polyvinyl chloride pipes. Through these profiles a stream of redistilled water was passed at a constant hydrostatic pressure of about 5 hPa. The basic parameters of initial water were: t = 18°C, pH = 6.0 and specific conductivity of water $K = 2.0\ \mu S \cdot cm^{-1}$.

The aim of using fresh loess monoliths was to utilize natural partial pressure of carbon dioxide in soil air, which, on the basis of the author's studies and other sources, should be considered as considerable (Buckman and Brady 1971; Deines et al. 1974; Nazarov 1981; Uggla 1981).

The first determinations of water were done after 3, 7 or 23 h from the beginning of the experiment. Determinations of successive water samples were done after 27, 31, 49, 57, 72, 78, 84 hours and after 140 h on closing the series. In each water, pH, specific conductivity (10°C), total hardness and calcium content were examined. The mean values of these characteristics are shown in Fig. 6.

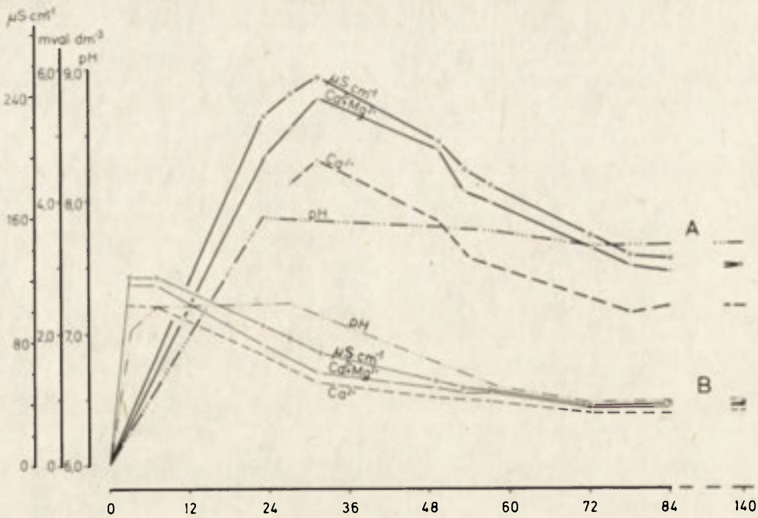

Fig. 6. Qualitative changes of waters infiltrating through monoliths of upper, decalcified parts of loess profiles (unweathered — B) and carbonate loesses (A)

Studies of water reaction were carried out in two groups: in profiles-monoliths taken from roof parts (considerably decalcified) of the profiles and from their underlying carbonate loesses. In Fig. 6, in group B of diagrams, the course of pH curve is presented in the time as the mean values from the studies of all individual samples of water infiltrating through the roof parts of the profiles, and in group A of water infiltrating through unweathered loesses.

From the studies (on 12 profiles) the following conclusions can be drawn: 1. During the first day (24 h) of a stream water flowing through the profiles the readings were close to those recorded in the studies with disintegrated samples. Later a distinct decrease in pH values was observed in decalcified – eluvial loess horizons. 2. From the calculations of H_2CO_3 as a ratio of HCO_3 and H^+ to K_1 it appears that CO_2 partial pressure in loess monoliths was about 30 – 35 hPa, and the equivalent water pH_t was lower (5.0–5.5) than that of the water before infiltration (pH = 6.0).

Another important problem is probable occurrence of the so called exchangeable sorption in loesses. In the cases studied complications with pH may also result from potential possibility of cation H^+ entry into the sorption complex of loesses. In P. Schachtschabel's opinion (Scheffer and Schachtschabel 1976) the entry of cations into the sorption complex is a selective reaction which largely depends on the kind of the colloid. From the description of the loesses of the area studied it appears that the main colloidal component in loesses is illite and kaolinite. According to this author illite demonstrates a great selective sorption in relation to H^+. It follows that a change of water solution pH, besides dissociation of carbonate, may to some extent result from saturation of the solid phase with hydrogen cations.

The specific conductivity of waters infiltrating through the loess monoliths was distinctly differentiated. This resulted from the chemico-mineral composition of the rocks studied, as well as from the time interval in which the profile was continuously leached. The mentioned relationship occurring in dual profiles of carbonate loesses is illustrated in Fig. 6. Attention is called by the fact that the highest responses of the coductivimeter occurred in the first phase of experiment. In the case of loesses of the lower, unweathered part of profiles the specific conductivity of water (10°C) was from 200 to 260 $\mu S \cdot cm^{-1}$ (Fig. 6A) and was two times higher in comparison to the results obtained for the roof parts of profiles (Fig. 6B). It should be stressed that these properties are close to those which were recorded in studies with disintegrated samples.

Laboratory studies on leaching of carbonates in freshly prepared loess monoliths should be recognized as close to those carried out in the field. Deviation from exact simulation of natural infiltration and decalcification consisted almost exclusively in the use of redistilled water. Reconnaissance tests by using precipitation and thaw waters indicate that the method applied is appropriate, and small differences in the results of studies can be caused both by the kind of water used and the individual character of the monolith studied.

The studies comprised determinations of total and carbonate hardness, the content of Ca^{2+} ions; magnesium content was calculated from the differences of milliequivalents $(Ca^{2+}+Mg^{2+})$ – Ca^{2+}. The determinations were performed synchronically with measurements of pH and specific conductivity. The curves of total hardness are similar in their shape to those of specific conductivity of water (Fig. 6). This results from distinct dependence of conductance on water mineralization. However, total mineralization of waters in loesses is largely determined by dissociation of carbonates, which goes together with infiltration processes. During the experiments a decrease in the content of calcium and magnesium was found in subsequent water portions passing through the profiles-monoliths.

The conclusions concering evolution of the chemical composition of waters, drawn from the applied infiltration should be considered reliable. This is connected with the probability of maintaining natural partial pressures of CO_2 in freshly prepared rock monoliths through which water was flowing. These pressures are relatively high, which might account for the intensity of biochemical changes associated with evolution of considerable amounts of CO_2 in the roof layers of loess covers. Partial pressures, in relation to atmospheric, which increased manifold in the rocks, caused in accordance with Henry's law an increase of P_{CO_2} in infiltration waters, and a rapid decrease of pH must have been effected by these changes. The consequence of increased aggressiveness of acid waters on carbonates was their high hardness, particulary recorded in the initial period of studies (over 5.5 $mval \cdot dm^{-3}$). Attention is attracted by the fact that waters infiltrating through loesses exceeded spring waters of the upper Sanna catchment area with respect to the total hardness.

The coefficient Mg/Ca can be a genetic index of waters, which can help to estimate the environment in which water has reached a state of saturation with carbonates or has been close to it. The present studies of the content of both ions in waters dissolving carbonate loesses in monoliths have shown that this coefficient is 0.227. In waters of springs which drain reservoirs of the water-bearing cretaceous horizon underlying carbonate loesses in the Sanna river catchment area, the value of Mg/Ca coefficient is 0.203.

FIELD STUDIES OF THE DECALCIFICATION OF LOESSES

Precipitation water supplying underground reservoirs can infiltrate: a) directly after touching the ground surface, b) after a period of stagnation on the surface, most frequently in small land depressions (grooves, furrows, pits, ditches and other microforms), c) after surface or hypodermic translocation of its molecules from places located higher than the depressions. In each of these cases mentioned initial water infiltrating through rock pores has different physical and chemical features, above all a different aggressiveness on carbonates.

In the upper Sanna catchment area no systematic studies of the ionic composition and physical features of precipitation waters were carried out. However, attention should be given the results obtained from measurements made sporadically in 1982-1983. They concerned both rain water (4 measurements) and that coming from snow (fresh and stale — 8 measurements). In the light of these studies the reaction of precipitation waters was 5.4 to 6.0, total hardness ranged from 2.0—4.0 $mg \cdot dm^{-3}$, HCO_3 — 4.0 to 6.0 mg and Cl^- from 2.0—3.0 $mg \cdot dm^{-3}$, and specific conductivity of rain waters was determined from 14.0—25.0 $\mu S \cdot cm^{-1}$.

In waters obtained from snow the parameters studied were respectively: 5.0—6.3 (pH), 2.0—3.0 $mg \cdot dm^{-3}$ (Ca+Mg), 3.0—6.0 $mg \cdot dm^{-3}$ (HCO_3), 1.0—6.0 $mg \cdot dm^{-3}$ (Cl^-) and 10—29 $\mu S \cdot cm^{-1}$ (specific conductivity at 10°C). From the fragmentary data it appears that pollution of precipitation waters in the upper Sanna catchment area is much lower than, e.g., in the Western Carpathian Mts (Pawlik—Dobrowolski 1983).

Precipitation waters, on touching the topographic surface of the loess area can directly infiltrate into the ground only in small amounts. Water permeation is determined by low values of the filtration coefficients. Therefore, we should consider the features of those waters which, after a period of stagnation or local surface flow, begin to percolate gravitationally through loess pores and channels. Such waters were examined in two periods of spring surface flow (1982—1983). Water samples from the surface flow were taken from 8 localities situated in the catchment area studied. In waters stagnating on the surface the reaction was observed to range from 6.77—7.08. In determinations of other features the following variations were found (in $mval \cdot dm^{-3}$): total hardness — 0.62—0.90, Ca^{2+} — 0.44—0.80, Mg^{2+} — 0.10—0.18, HCO_3^- — 0.15—0.70, Cl^- — 0.08—0.11, SO_4^{2-} — 0.05—0.12. Waters from surface flow had slightly higher values: pH — 7.68—8.00, total hardness (in $mval \cdot dm^{-3}$) — 0.62—1.52, Ca^{2+} — 0.58—1.35. Mg^{2+} — 0.04—0.25, HCO_3^- — 0.55—0.90, Cl^- — 0.13—0.33 and SO_4^{2-} — 0.30—0.37. The specific conductivity of stagnating waters was 57—94 $\mu S \cdot cm^{-1}$, whereas that of waters flowing on the land surface — 55.122 $\mu S \cdot cm^{-1}$.

From the observations it appears that the primary features of precipitation waters had radically changed before the infiltration process started. As regards other changes, water reaction altered significantly in the direction of alkalinity. In such a case the term of initial water can hardly be referred to the time interval in which decalcification of loesses starts in vertical horizons. It is undoubtedly that infiltration waters recover a weakly acid reaction on passing through the soil profile; however, they create additional difficulties in observations of water saturation with carbonates under conditions of an open system, being itself complicated.

WATERS IN CARBONATE LOESSES

Decalcification of carbonate loesses by infiltrating waters was studied in three natural profiles (No 13, 17 and 38) and in several loess walls. Water for studies was taken at vertical 0.5 m distances to the depth of 2.5 m. The water infiltrating into loess was drained away into vessels by means of troughs installed horizontally. Because of technical difficulties only pH and total hardness were examined in the water filtrates.

The experiments confirmed quantitative and qualitative changes in waters, which were recorded in studies of loess monoliths in the first experimental phase.

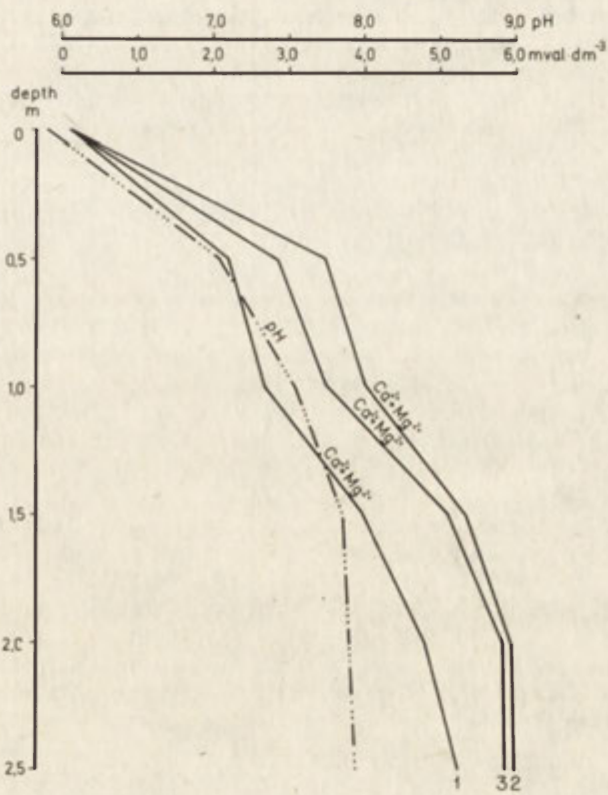

Fig. 7. Changes of total hardness ($Ca^{2+}+Mg^{2+}$) and reaction (pH) in waters infiltrating through profiles of carbonate loesses; 1 and 2 — extremal values found, 3 — mean value of $Ca^{2+}+Mg^{2+}$ in waters

From the studies of natural profiles it can be assumed that every dm^3 of water percolating through carbonate loess leaches from this rock about 5.0–6.0 mval of calcium magnesium. In terms of standard units of calcium carbonate this gives 250 to 300 mg of $CaCO_3$. In this process the reaction of infiltrating water changes to weakly alkaline (pH 7.6–7.8) and is an index of dynamic equilibrium of water saturation with carbonates (Fig. 7). Water possessing such features is not aggressive any more, i.e. it does not dissolve $CaCO_3$ during its gravitational flow into the saturation zone. Thus the field experiment has proved that loess cover plays a protective role for a carbonate deposit to which infiltration water is flowing.

SPRINGS AND THE SANNA RIVER

After passing the saturation zone, infiltration waters undergo infiltration processes according to the hydraulic gradient. The Sanna river valley plays a draining role. This is manifested by the occurrence of spring complexes of considerable output in the valley (Fig. 1).

Studies on changes of the chemical composition of the waters of all valley springs of the upper Sanna river were carried out from July 1981 till June 1983, which were systematic from the autumn 1982. From the analysis of water pH variations in the springs and the Sanna river it appeared that over the study period it was more alkalic in the river water than in that of springs. Periodical pH decrease of river water always coincided with surface flow. It was also observed that decreased pH in spring waters was recorded with a small delay. This confirms the assumption that in natural waters occurring in carbonate rocks any reaction decrease is caused by the inflow of fresh infiltration waters of a relatively high CO_2 partial pressure.

An analysis of a temporal pH change in water flowing out onto the surface due to snow thawing and defrosting of the ground (resumption of infiltration) indicates that after a period shorter than one month, fresh precipitation waters appear.

Parallel to pH changes, a temporal variation of the level of carbonates in water was recorded. The temporal decrease of both the features studied in spring waters during gravitational inflow should be interpreted by increased water infiltration through the cover of decalcified loesses. As it appears from unpublished studies such waters are characterized by weakly acid reaction and a small amount of carbonates.

In the annual flow of the Sanna river the underground flow is of decisive importance. On the basis of the material possessed by the author it can be assumed for many years that the modulus of underground flow in the upper Sanna catchment area amounts 4.2 $dm^3 \cdot s^{-1} \cdot km^2$, and the value of the flow in the profile closing the area studied is 320 $dm^3 \cdot s^{-1}$ (with underground flow).

To determine the value of the material in the form of ions carried outside the system (catchment area), it is necessary to estimate mineralization of the total water in the river. In the studies carried out in the Sanna catchment area mineralization of the river water in the periods of underground flow was taken into consideration. Conductometric measurements were made 6–8 times in a month. Apart from that, chemical analyses of waters from springs and the Sanna river were performed at the site were the water level recorder was located, once or twice a month in order to determine the activity coefficients of the water studied. The mean conductivity was calculated by two methods: as medial value from weighted means of the specific conductivity of spring waters and as arithmetic mean of river water conductivity from two hydrological years (1982 and 1983). It is interesting to note that these values do not differ. The mean value of specific conductivity (365 $\mu S \cdot cm^{-1}$) was converted into specific resistance ($t = 10°C$).

For determination of water mineralization in the river water for geomorphological purposes, Doroszewski's formula modified by the author (Janiec 1982) was used:

$$M\,(\mathrm{mg \cdot dm^{-3}}) = \frac{1\,009\,000}{\varrho\, 10^{\circ}\, \gamma \pm} - C_{Ti}$$

in which M is the total mineralization of water reduced by the value C_{Ti}, $\varrho\ 10^{\circ}$ is the water resistance at 10°C, $\gamma\pm$ is the mean activity coefficient, and C_{Ti} is the total content initial carbon. The value $\gamma\pm$ is determined by Debye—Hückel equation for waters of known ionic strength, whereas value C_{Ti} is mainly CO_2 dissolved in water infiltrating through soil into the bedrock and inherent in HCO_3^- ion present in waters. Carbon dioxide is here mainly of organic origin and should not increase denudation which occurs in the lithosphere. The value C_{Ti} was determined from molar ratios of $Ca(HCO_3)_2$ dissociating in water. The numerical value of this parameter is 84 mg $\cdot$ dm^{-3} for mean conductivity of the Sanna water. For two years studied (1982—1983) the mean weighted value of mineralization was:

$$M = 419.0\ \mathrm{mg \cdot dm^{-3}} - 84\ \mathrm{mg \cdot dm^{-3}} = 335\ \mathrm{mg \cdot dm^{-3}}$$

The ratio of underground flow volume and mineralization constitutes the value of ionic flow which amounts:

$$10\,139\,450\ \mathrm{m^3 \cdot year^{-1}} \cdot 0.000335\ \mathrm{ton \cdot m^{-3}} = 3\,396.7\ \mathrm{ton \cdot year^{-1}}$$

The index of chemical denudation, which is 44.4 ton $\cdot$ km^{-2} $\cdot$ year^{-1} is expressed by the quotient of ionic flow and the catchment area surface.

On the basis of the distribution of carbonate loesses in the upper Sanna catchment area (about 24 km^2) and the mineralization level of waters infiltrating through these deposits, it can be assumed that about 30% of chemical denudation comes from decalcification of unweathered loess (carbonate) deposits. The value of ionic flow calculated for this part of the catchment area is over 1 000 tons $\cdot$ year^{-1}.

CONCLUSIONS

From the studies on decalcification of carbonate loesses it appears that precipitation waters gravitationally percolating through unweathered loess covers are always saturated with carbonates to an extent at least equal that of waters flowing out from springs. Such waters do not cause leaching of carbonate rocks underlying loesses.

The quantitative and qualitative determination of chemical denudation in loesses as a results of studies of decalcification processes allows us to predict changes and progressing degradation of the geographic environment of loess areas. To prevent further decalcification of carbonate loesses, which directly leads to podzolization of soils formed on these rocks, liming should be used. Undertaking such a treatment, indices of underground flow should be taken into consideration. In the case of the upper Sanna catchment area this index was determined at over 130 mm.

REFERENCES

Barnes I., 1964, *Field measurement of alkalinity and pH*, Geological Survey Water-Supply Paper 1535-H, Washington.

Buckman H. O., Brady N. C., 1971, *Gleba i jej właściwości* (Polish translation; The nature and properties of soils), PWRiL, Warszawa.

Deines P., Langmuir D., Harmon R. S., 1974, Stable carbon isotope ratios and the existence of a gas phase in the evolution of carbonate ground waters, *Geochemica et Cosmochimica Acta* 38, 1147—1165.

Hem J. D., 1985, *Study and interpretation of the chemical characteristics of natural water*, Geological survey Water-Supply Paper 2254, Washington.

Jahn A., 1950, Less, jego pochodzenie i związek z klimatem epoki lodowej (Sum.: Loess, its origin and connection with the climate of the glacial epoch), *Acta Geologica Polonica* 1, 3, 257—302.

Jahn A., 1956, *Wyżyna Lubelska. Rzeźba i czwartorzęd* (Sum.: Geomorphology and Quaternary history of Lublin Plateau), Prace Geograficzne IG PAN 7.

Janiec B., 1982, Badania denudacji chemicznej metodą konduktometryczną (Studies over a chemical denudation by a conductometric method), *Annales UMCS, Sec. B*, 37 119—140.

Janiec B., 1984, *Wody podziemne w strefie południowo-zachodniej krawędzi Wyżyny Lubelskiej*, Wyd. Geol., Warszawa.

Jersak J., 1973, *Litologia i stratygrafia lessu wyżyn południowej Polski* (Sum.: Lithology and stratigraphy of the loess on the southern Polish uplands), Acta Geographica Lodziensia 32.

Langmuir D., 1968, Stability of calcite based on aqueous solubility measurements, *Geochimica et Cosmochimica Acta* 32, 835—844.

Malinowski J., 1964, *Budowa geologiczna i właściwości geotechniczne lessów Roztocza i Kotliny Zamojskiej między Szczebrzeszynem i Turobinem* (Geological structure and geotechnical properties of loesses in Roztocze and in the Zamość Basin, between Szczebrzeszyn and Turobin), Prace Instytutu Geologicznego 41.

Malinowski J., 1974, *Hydrogeologia Roztocza Zachodniego*, Prace Hydrogeologiczne IG, seria specjalna 6.

Maruszczak H., 1972a, Zagadnienie klasyfikacji lessów w Polsce, in: *Przewodnik sympozjum krajowego. Litologia i stratygrafia lessów w Polsce*, Wyd. Geol., Warszawa, 17—23.

Maruszczak H., 1972b, Podstawowe cechy genetyczne i stratygraficzne lessów Polski południowo—wschodniej, in: *Przewodnik sympozjum krajowego. Litologia i stratygrafia lessów w Polsce*, Wyd. Geol., Warszawa 89—135.

Maruszczak H., 1980, Stratygrafia i chronologia lessów w Polsce (Stratigraphy and chronology of the loesses in Poland), in: *Przewodnik seminarium terenowego. Stratygrafia i chronologia lessów oraz utworów glacjalnych dolnego i środkowego Plejstocenu w Polsce SE*, 43—54.

Muxart T., Birot P., 1977, *L'alteration meteorique des roches*, Paris.

Nazarov T. V., 1981, *Gidrologicheskaya rol pochvy*, Leningrad.

Pawlik—Dobrowolski J., 1983, *Zmiany składu chemicznego wód powierzchniowych pod wpływem zanieczyszczeń obszarowych w Karpatach Zachodnich*, Falenty.

Pazdro Z., 1983, *Hydrogeologia ogólna*, Wyd. Geol., Warszawa.

Scheffer F., Schachtschabel P., 1976, *Lehrbuch der Bodenkunde*, Stuttgart.

Tokarski J., 1951, Chemiczno-mikroskopowa analiza lessu z okolic Mogiły pod Krakowem, *Sprawozdania z Czynności PAU* 52, 3, p. 200.

Tokarski J., Oleksynowa K., 1951, A special analysis of loess from Mogiła near Kraków, *Comptes Rendus Mensuels des Séances de la Classe des Sciences Mathematiques et Naturalles*, Academie Polonaise des Sciences et des Lettres 5—10, p. 25.

Uggla H., 1981, *Gleboznawstwo rolnicze*, PWN, Warszawa.

Zawadzki S., 1972, Charakterystyka fizycznych właściwości wybranych utworów lessowych Wyżyny Lubelskiej, in: *Przewodnik sympozjum krajowego. Litologia i stratygrafia lessów w Polsce*, Wyd. Geol., Warszawa, 64—65.

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DYNAMICS OF SHALLOW GROUNDWATER TEMPERATURE IN THE VISTULA ABANDONED VALLEY, THE PLOCK BASIN

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1. INTRODUCTION

The temperature of shallow groundwater is markedly affected by air temperature, the height of water table and its changes throughout the year, thermal conductivity and ground moisture in the zone of aeration, slope exposure and terrain cover, i.e. land use patterns and built-up areas. Thus, the problem is complex but little attention has been devoted to it in the hydrological study.

The textbooks of hydrogeology contain the basic information concerning factors which affect groundwater temperature and the depth of diurnal, seasonal and annual temperature fluctaution zones (Marchacz 1960; Pazdro 1964). These problems also dominate the latest textbook of groundwater hydrology (Kriz 1983). The problems of groundwater temperature dynamics fall outside their scope.

Few detailed studies on the thermal properties of shallow groundwater are concerned with selected aspects of the problem. An example concerns a few characteristic wells where Mikulski (1963) pointed out the dependence of groundwater temperature on the thickness and geological structure of the zone of aeration. Skibniewska (1963, 1964), Nawrocka and Sadowska (1964) dealt with groundwater temperature fluctuations at different depths in relation to air temperature in Poland's territory. Dynowski (1968) has established that groundwater temperature depends on thermal conductivity of the water-bearing and overlying rocks. Relationships between air temperature and water temperature in two wells differing in geological structure were considered by Gutry-Korycka (1969). She suggests that shallow groundwater temperature dynamics is largely dependent on air temperature, thermal conductivity of the ground in the zone of aeration, annual fluctuations of depths to water and groundwater character, i.e. chemical composition.

Publications pertinent to the influence of slope exposure and land use patterns, i.e. terrain cover, on groundwater temperature have been lacking so far in the Polish scientific literature. One of the objectives of the present study is to assess the significance of these factators.

2. RESEARCH AIM AND STUDY AREA

The studies were concerned with the identification and description of shallow groundwater temperature dynamics down to the depth of 3 m in sandy-gravelly deposits of the Vistula valley, especially

(1) the identification of annual varitations and amplitudes of groundwater temperature with reference to air temperature and the thickness of the zone of aeration,

(2) the observation of the influence of local factors, especially slope exposure, ground moisture and land use patterns (terrain cover) on groundwater temperature.

The studies were carried out in the western portion of the Płock Basin, in an area adjacent to the lower part of the Włocławek water dam (Fig. 1). Its main morphologic characteristics are flat, frequently marshy terrace levels along the Vistula, which are locally diversified by dunes and lakes. Glaciofluvial sediments, i.e. up to 40—50 m thick sands with admixed gravel, dominate the profile of Quaternary deposits. Fluvio-lacustrine and lake sediments comprising fine and silt-sized sands occur locally in close vicinity to terrain depressions. The bottoms of the depressions are filled with organogenic sediments, i.e. peats and alluvial material. Fine dune sands are of eolian origin.

The thickness of the zone of aeration reaches 2—3 m at a maximum but it increases to higher values in dune—covered areas. The ground with moderate and high permeability prevails in the arid layer. It accounts for 80 and 98 percent of the surface area and depth of 3 m, respectively (Glazik 1978). Peats have expanded considerably; they occupy 16 and 0.3 percent of the surface area and the 3m depth, respectively. The ground I (tills, clays) which is basically impermeable is sporadically found in the zone of aeration (1 percent). Owing to this, groundwater table is free and there exist suitable conditions for infiltration in the study area.

3. RESEARCH METHODS

For the purpose of establishing relationships between groundwater temperature and air temperature in connection with the thickness of the zone of aeration, data from 10 control wells located at various sites of the Vistula abandoned valley (Fig. 1) have been used. Most wells are situated at higher upper floodplain level III and within the extent lines of a low-lying area adjacent to the lower part of the reservoir. Two wells are found at higher terrace level IV. Over the period of observation, i.e. in the year 1971—1972, water level in wells was measured every day, while the measurement of water temperature was performed twice a week. Data concerning air temperature were provided by a meteorological station at Wistka Królewska which is not more than 7 km distant from the farthest well.

The average depths to water vary and fluctuate between 0.5 and 3.0 m in the wells. All the wells have similar geological structure. Fine sands prevail to the depth of 0.5—1.0 m and are underlain by medium-grained, scarce coarse sands and gravels (Glazik 1978). Ground permeability, as a rule, increases with depth.

The location of the wells in flat land within farms protects severly microclimatic conditions and thus, groundwater temperature against the influence of differences in terrain cover and slope exposure. It can also be inferred that variations of thermal conductivity and ground moisture are not great because of similar geological structure and the location of farms in higher-lying areas, far from wet regions. This has allowed tracing of the relationships between groundwater temperature and air temperature in connection with the depth to water.

In order to establish the effect of local factors, especially slope exposure, ground moisture and land use patterns, on groundwater temperature, an area lying close to the station of the Institute of Geography and Spatial Organization of the Polish Academy of Sciences at Dobiegniewo has been chosen for the study area (Fig. 1). Thirteen piezometers have been installed in a few hydrogeological profiles perpendicular to the reservoir. They are located on dune slopes with a different exposure and in areas with a

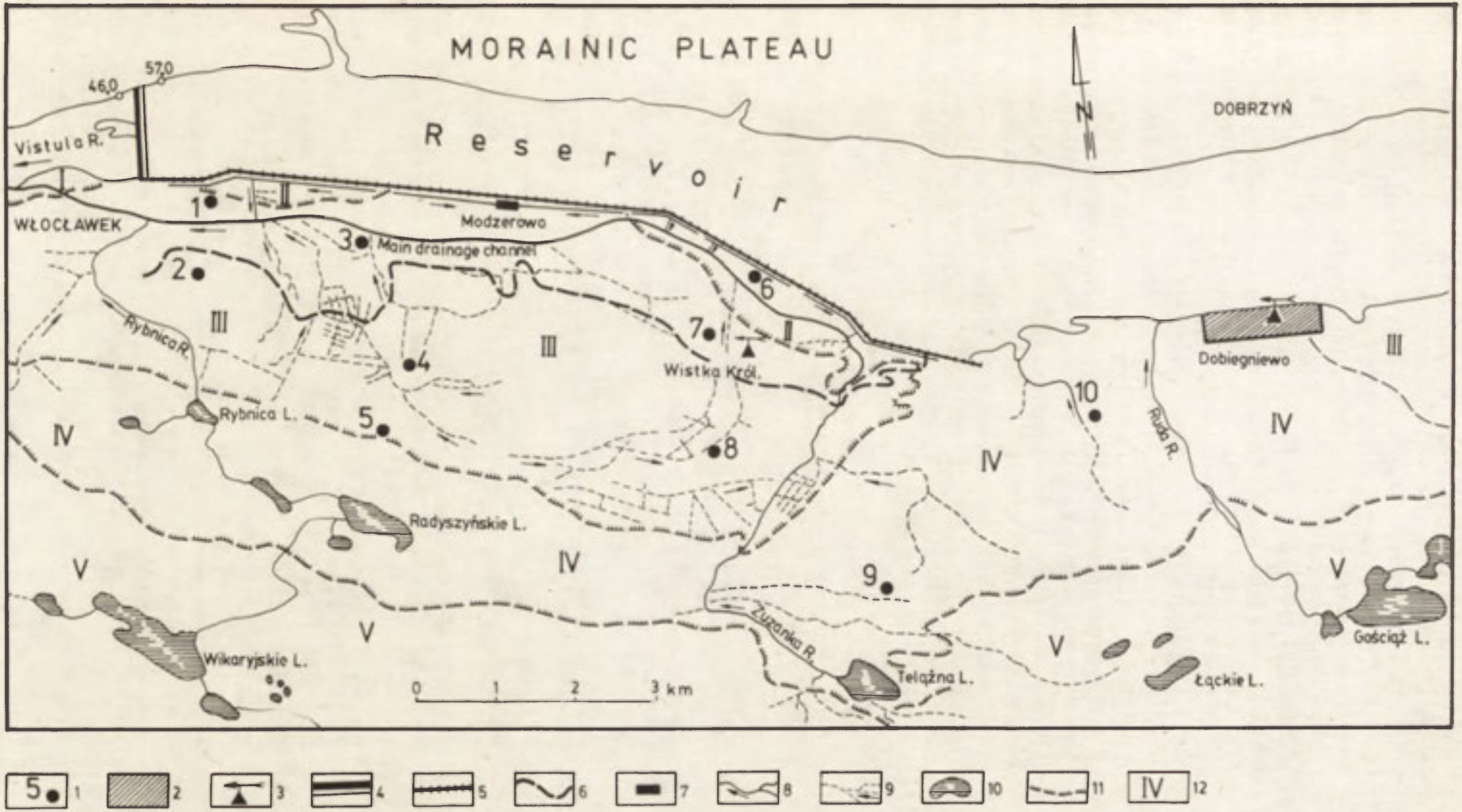

Fig. 1. Study area: 1 — observation wells, 2 — area subjected to detailed study, 3 — meteorological stations, 4 — frontal dam, 5 — lateral dam, 6 — boundary of low-lying area, 7 — pumping station, 8 — permanent streams, 9 — periodical streams, 10 — lakes, 11 — terrace scarps, 12 — terrace number

different land use pattern and differing moisture, i.e. a forest, meadow, pasture and barren land. Detailed location of piezometers is presented in the follwing pages.

Piezometers are 10 cm diameter polyvinyl chloride tubes. Metal taps are attached to their ends. Orifice and terrain datum was levelled around the piezometers. Over the period of observation, 1983—1984, the measurement of water level and temperature was performed by the use of piezometers once a week. Water temperature was measured with a thermometer many times until the mercury column became stable. A better solution would be provided by fixing thermometers in the piezometric tubes and taking them out merely in order to read off the figures or using a thermistor thermometer. The thickness of a water layer in the piezometers was at least 0.2 m at low levels.

It was difficult to evaluate the influence of local factors on graundwater temperature dynamics because of differing depths to water in the piezometers. They ranged from 0.2 to 1.4 m. An attempt was made to eliminate the influence of the thickness of the zone of aeration on groundwater temperature by analysing data from the piezometers with identical depth to water and with identical geological structure. Data concerning air temperature were acquired from the meteorological station of the Institute of Geography and Spatial Organization of the Polish Academy of Sciences at Dobiegniewo, located in the centre of the detailed study area (Fig. 1). Two-year observation cycles are too short to provide a full description of groundwater temperature dynamics. Nevertheless, they have allowed a lot of underlying characteristics relevant to the problem to be identified.

4. EFFECT OF AIR TEMPERATURE AND AERATION ZONE THICKNESS ON GROUNDWATER TEMPERATURE

The data from observation wells characterized by similar geological structure, similar moisture of the zone of aeration and similar setting have provided the basis for studying fluctuations and annual amplitudes of water temperature at differing thickness values for the zone of aeration. Relationships have been established between groundwater temperature and depths of groundwater occurrence during particular months throughout the year.

4.1. ANNUAL VARIATIONS OF WATER TEMPERATURE IN WELLS

Annual variations of groundwater temperature are largely dependent on air temperatures (Fig. 2). Over the period of observation, the highest water temperatures were recorded in all wells in August, whereas they became lowest in winter from January till March. Over the spring-summer period from April till August water temperature was lower than air temperature. During the autumn-winter period from September till March, the opposite held. Identical air and water temperatures were recorded at the turn of March and of August.

The effect of air temperature on groundwater temperature became lessened rapidly with increasing depth. It was extremely marked in the deepest well (2.5 m). This can be inferred from more uniform water temperatures as the thickness of the zone of aeration increased (Fig. 2). The greatest variations of water temperatures in the wells with differing depths occurred at spells of extreme air temperatures, i.e. winter-summer. During the summer half of the year water temperature dropped off with depth, whereas it increased in the winter half of the year. From April to May and from September to October, water temperatures in the wells became uniform, irrespective of the well depth.

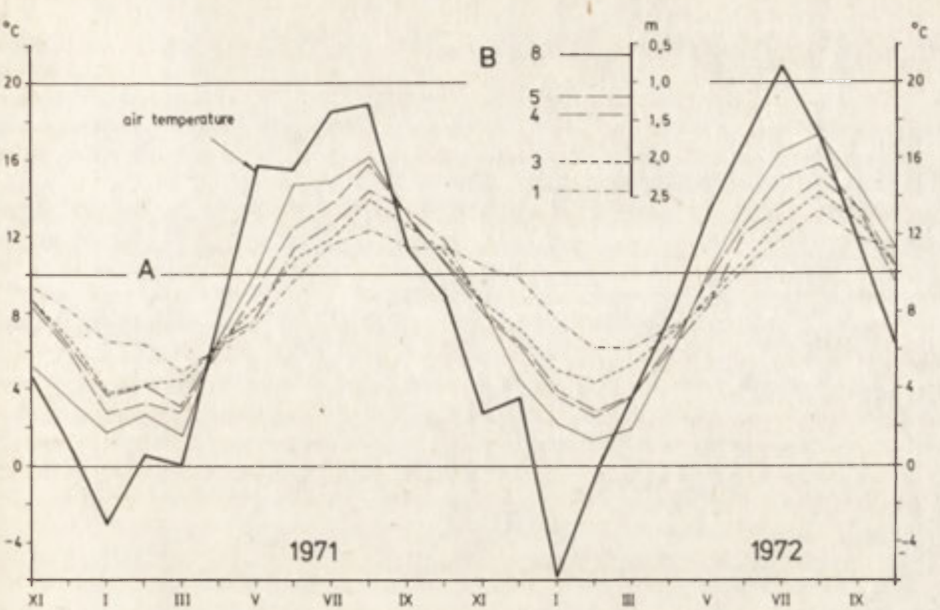

Fig. 2. Mean monthly water temperatures in wells with differing depths against air temperatures, the Wistka Królewska station: A — mean monthly air and groundwater temperatures, B — mean depths to water, 1971—1972 (see Fig. 1 for well number)

4.2. RANGE OF GROUNDWATER TEMPERTATURE FLUCTUATIONS AND AMPLITUDES

As the depth increases, fluctuations of groundwater temperature are subject to rapid change (Fig. 3). During the period of observation water temperatures fluctuated between 0.1 and 17.7°C and 4.0 and 13.6°C in the wells where the zone of aeration was up to 1 m

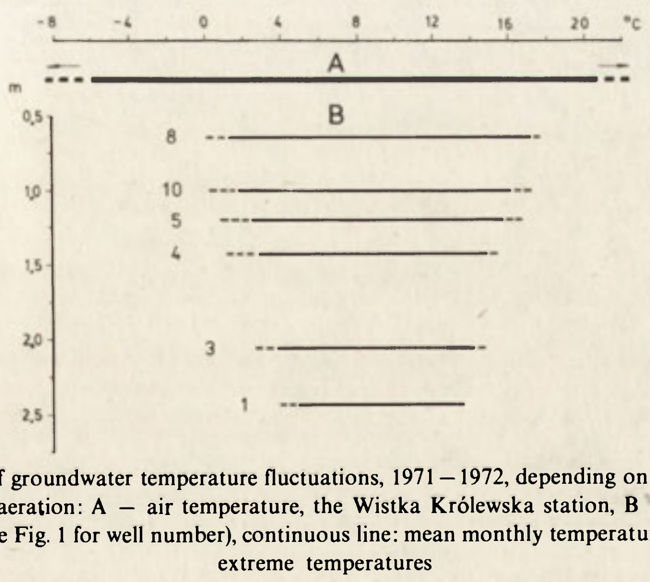

Fig. 3. Range of groundwater temperature fluctuations, 1971—1972, depending on mean thickness of the zone of aeration: A — air temperature, the Wistka Królewska station, B — groundwater temperature (see Fig. 1 for well number), continuous line: mean monthly temperatures, broken line: extreme temperatures

thick and in a well 2.5 m deep, respectively. In the winter there was a thin ice cover in shallow wells up to 1 m deep.

The above data are indicative of strong influence of air temperature fluctuations on groundwater temperature in a 2.5 m thick ground layer. As the depth increases, water temperature fluctuations become reduced, resulting in a fall in maximum temperature and a rise in minimum temperature.

There is a close relationship between the annual water temperature and well depth but it is not based on a simple linear equation (Fig. 4). The shallower the depths are, the more marked rises of the annual water temperature amplitudes occur. In 1972 the annual water temperature amplitudes were higher at a given depth than those in 1971. This is illustrated in Figure 4 in the form of curves displaced to the right along the axis of amplitudes.

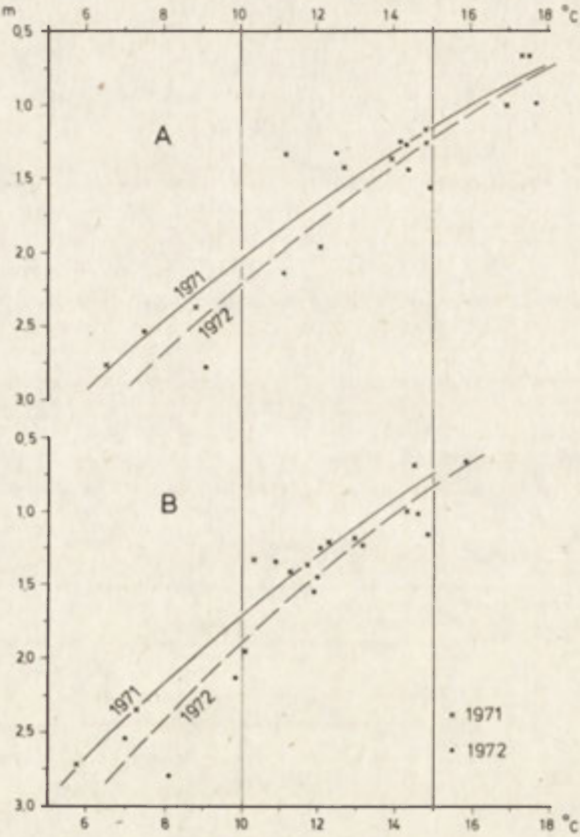

Fig. 4. Dependence of annual groundwater temperature amplitude or mean annual thickness of the zone of aeration A — annual amplitudes of extreme water temperatures measured twice a week, B — annual amplitudes of extreme mean monthly water temperatures

In 1971 and 1972 the mean annual air temperatures recorded at the station at Wistka Królewska were nearly identical (8.3 and 8.0°C). The annual amplitudes of mean monthly air temperature differed to a larger extent. They were 22.0 and 26.7°C in 1971 and 1972, respectively. This difference became reflected in higher water temperature amplitudes in 1972. Thus, it can be inferred that annual amplitudes of shallow groundwater temperatures depend on the annual amplitudes of air temperature. It is not possible to consider this relationship in more detail because of a short period of observations.

The annual amplitudes of groundwater temperature decrease rapidly with depth (Fig. 4). They were twice lower at the depth of 2.5 m than at that of 1.0 m.

4.3. RELATIONSHIPS BETWEEN THE MEAN MONTHLY GROUNDWATER TEMPERATURE AND THE THICKNESS OF THE ZONE OF AERATION

The data from the ten wells represent the basis for plotting the relationships between the mean monthly temperature and the depth to water for each month separately. A

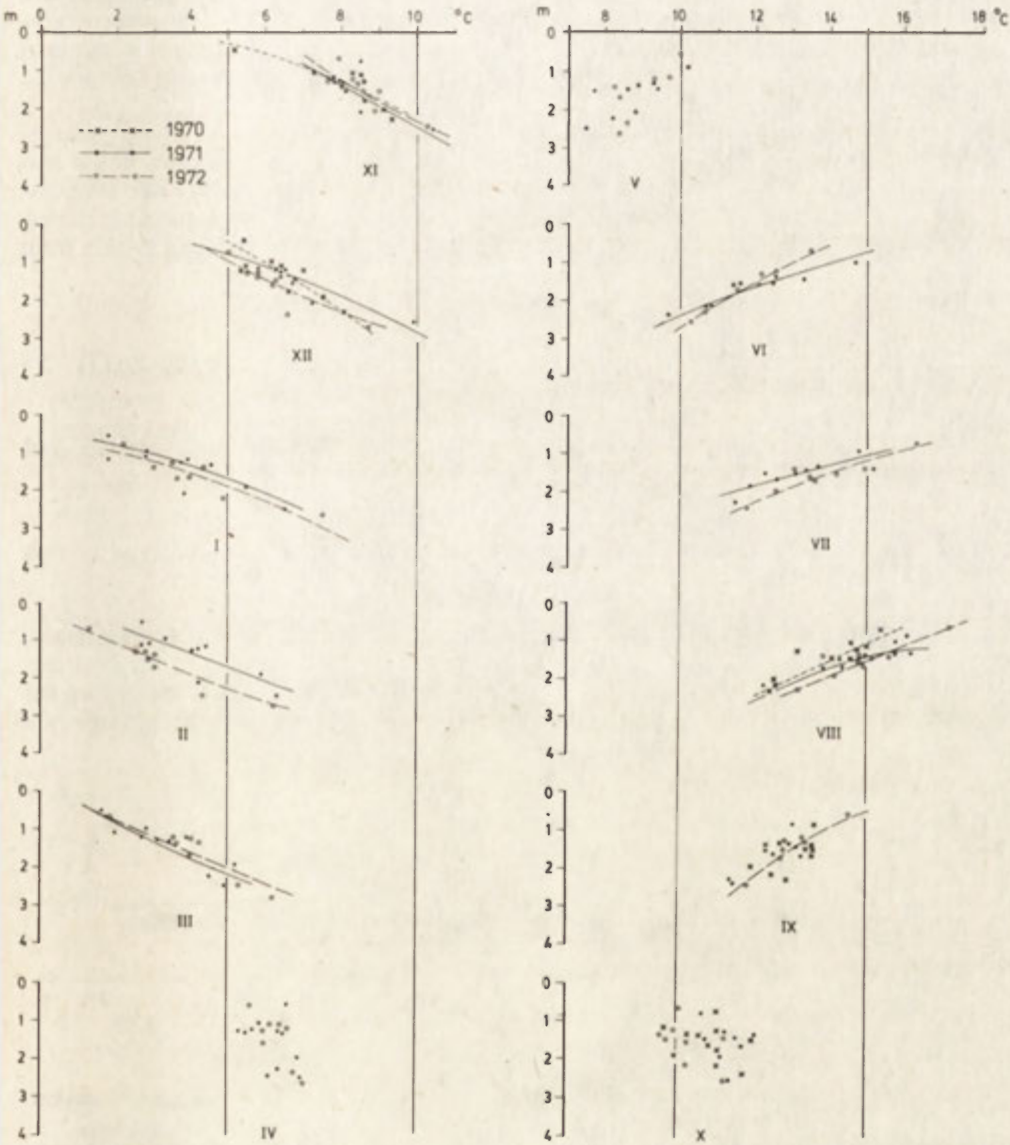

Fig. 5. Dependence of mean monthly water temperatures in wells on mean monthly depths to water throught the year, August 1970—October 1972

dynamic illustration of annual variations in shallow groundwater temperature has been thus provided. It is dependent on air temperatures and the thickness of the zone of aeration (Fig. 5).

During the winter half of the year groundwater temperature increased with depth due to slower rates of cooling of the deep ground layers. From November the water temperature decreased gradually at the same depths to attain its minimum over a period between January and March. In April the water temperature became uniform in the wells due to rapid warming of the surface ground layers, the temperature of which approached gradually that of the deep layers.

The reverse thermal properties occurred during the summer half of the year. Water temperature decreased with depth due to slower warming of the deep ground layers. From May the water temperature tended permanently to be higher at the same depths to attain its maximum in August. In October uniform water temperatures were again recorded in the wells due to rapid cooling of the ground layers immediately beneath the surface.

The above relationships are applicable to shallow groundwater whose zone of aeration is composed of merely sandy deposits. Owing to a small thickness of the arid layer, air temperature fluctuations become rapidly reflected in groundwater temperature. The relationships of water temperature with the depth if its occurrence during a given month showed minor variations during particular years, resulting largely from differences in air temperature.

5. EFFECT OF SLOPE EXPOSURE, GROUND MOISTURE AND LAND-USE PATTERNS ON GROUNDWATER TEMPERATURE

The effect of local factors on groundwater temperature is exemplified by two hydrogeologic sections located near the station of Dobiegniewo.

5.1 EFFECT OF LOCAL FACTORS ON GROUNDWATER TEMPERATURE IN A HYDROGEOLOGIC SECTION EAST OF THE DOBIEGNIEWO STATION

Four piezometers were located in the section under investigation (Fig. 6). Piezometer 1 was 15 m distant from the reservoir. It was situated in close vicinity to the northern fringes of a forest on river alluvium (poplar, ash tree). Piezometer 2 was located on the southern fringe of the forest on river alluvium at the base of the north-facing slope of a

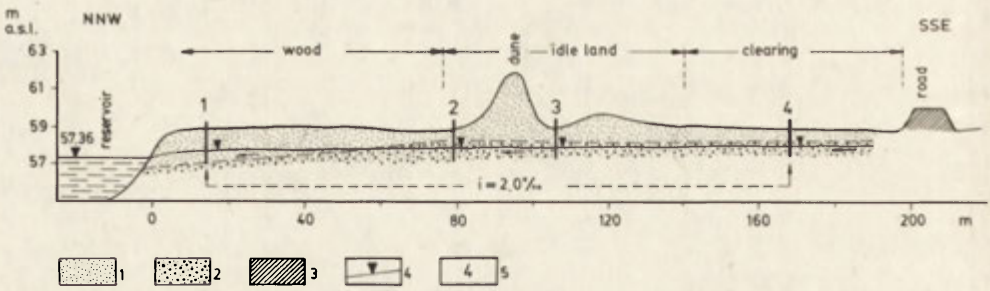

Fig. 6. Hydrogeological section along the line of piezometers 1–4, east of the Dobiegniewo station: 1 – fine and medium-grained sands, 2 – coarse sands containing gravel and pebbles, 3 – embankment, 4 – groundwater and reservoir table levels on 27 October 1982, 5 – piezometer number

small dune. The dune is, as a rule, without vegetation (unconsolidated sand). Juniper and dwarf pine are of local occurrence. Piezometer 3 was located at the base of the south-facing slope of the dune. There is xerothermal turf containing mosses and lichens around the piezometer. It is typical of warm and arid sites. Piezometer 4 was located in the clearing inhabited by grass and young birches.

All the piezometers were located in an area with similar geological structure. Medium-grained sands locally covered with fine dune sands spread over the surface. Beneath is a continuous layer of water-soaked coarse sands containing gravel and pebbles.

The characteristics of depths to water and extreme water temperatures in piezometers 1—4 are presented in Table 1.

TABELE 1. Depts to water and extreme water temperatures in piezometers 1—4, 1983—1984

Piezometer number	Depth to water in cm		Water temperature in °C		
	mean	amplitude	max.	min.	amplitude
1	113	37	14.1	3.9	10.2
2	90	52	16.5	4.0	12.5
3	91	51	18.7	3.0	15.7
4	77	56	16.5	2.5	14.0

Marked variations of water temperature in the piezometers are largely due to differences in warming and cooling rates between the ground layers immediately beneath the surface, depending on slope exposure and land use patterns (terrain cover). Most uniform water temperatures were recorded in the forest (piezometer 1). As opposed to bare soil, temperatures tended to be lower in the summer and higher in the winter. A relatively low water temperature maximum in the forest can be linked to

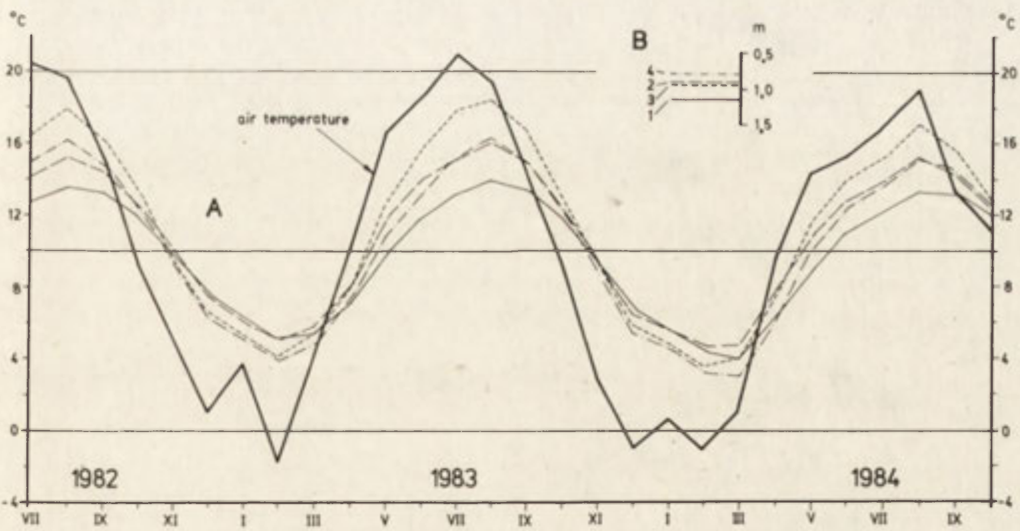

Fig. 7. Mean monthly water temperatures in piezometers 1—4, July 1982—October 1984, against mean monthly air temperatures, the Dobiegniewo station: A — mean monthly air and groundwater temperatures, B — mean depths to water, 1983—1984 (see Fig. 6 for piezometer number)

slower warming of the ground surface due to the presence of obstacles to sun-ray penetration in the form of tree crowns, the shrub layer and litter. A somewhat greater thickness and higher moisture of the zone of aeration, as well as microclimatic influence of the reservoir may be of some significance.

The maximum water temperature at the base of the south-facing dune slope (piezometer 3) was 2.2°C higher than that on the north-facing slope (piezometer 2) at the same thickness values for the arid layer and at the same amplitudes of water levels and as much as 4.6°C higher than that in the forest. Intense warming of the south-facing slope in the summer and considerable loss of heat in the winter (exposed slope) condition a high water temperature amplitude. The water temperature maximum in the clearing was identical with that on the north-facing slope on the forest fringe, while its minimum decreased to lowest values.

The greatest variations of water temperature in the piezometers were recorded in July and August (Fig. 7). This implies that the effect of slope exposure and terrain cover on the microclimatic conditions and thus, groundwater temperature was extremely marked in the summer. During the winter a snow cover serves as a heat insulator. The occurrence of a snow cover, as well as hampering of the growth of a vegetation cover, freezing of surface water, etc., contribute considerably to a reduction in microclimatic differences due to differing slope exposure and changeable land use patterns. In consequence, the goundwater temperatures were more uniform during the winter than during the summer.

5.2. EFFECT OF LOCAL FACTORS ON GROUNDWATER TEMPERATURE IN A HYDROGEOLOGICAL SECTION IN THE VICINITY OF THE DOBIEGNIEWO STATION

Six piezometers were installed in the section under investigation (Fig. 8). Piezometer 5 was located on a gently northward-inclined moist meadow above the water. Piezometers 6 and 7 were situated at the foot of the west-facing slope of a small dune within a grass-covered pasture. Piezometer 8 was located on a moist meadow stretching at the foot af a gently inclined slope with a southern exposure. Piezometers 9 and 10 were located in a typical dense forest on river alluvium.

Medium-grained sands covered locally with fine dune sands are largely found on the surface along the section line. The centre of the section is occupied by a terrain depression (intermittent marshy ground) drained through a ditch to the Włocławek reservoir after the damming of the Vistula water. The depression is filled with

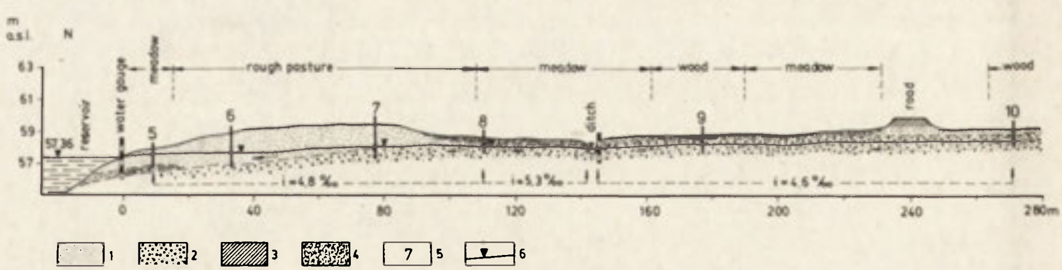

Fig. 8. Hydrogeological section along the line of piezometers 5—10, in the vicinity of the Dobiegniewo station: 1 — fine and medium-grained sands, 2 — coarse sands containing gravel and pebbles, 3 — embankment, 4 — fine and medium-grained sands containing organic particles (muck), 5 — piezometer number, 6 — groundwater, drainage ditch and reservoir table level on 27 October 1982

medium-graiend sands containing large proportions of organic particles. The substratum of the surface deposits is a continuous layer of water-saturated coarse sands containing gravels and pebbles.

A survey of the section has shown that there is an undergound watershed on a meadow in the vicinity of piezometer 8. It does not fit the terrain morphology. The watershed is the result of groundwater drainage through a ditch, which modifies a general north direction of groundwater runoff.

The influence of terrain morphology on groundwater temperature can be clearly seen in piezometers 5 and 8 (Table 2). They are characterized by similar geological

TABLE 2. Depths to water and extreme water temperatures in piezometers 5—10, 1983—1984

Piezometer number	Depth to water in cm		Water temperature in °C		
	mean	amplitude	max.	min.	amplitude
5	36	48	15.0	2.3	12.7
6	138	43	15.8	3.9	11.9
7	135	41	15.4	3.8	11.6
8	36	51	16.5	2.0	14.5
9	45	45	15.1	2.5	12.6
10	53	67	14.5	1.9	12.6

structure, identical depths to water and water table amplitudes and similar ground moisture due to the location of piezometers on moist meadows. During the period of investigation, water temperatures in piezometer 5 tended to be lower during the summer and higher during the winter. This was due to a different exposure of slopes or different conditions of warming and cooling of the ground surface. Piezometer 5 is located on the "cool" slope with a northern exposure, while piezometer 8 is located on the "warm" slope with a southern exposure. More uniform water temperatures in piezometer 5 may be additionally produced by the microclimatic influence of the reservoir. The north-facing slope directed towards the reservoir is subject to undoubtedly more intensive thermal influence of the reservoir than the opposite slope more distant from the reservoir.

A comparison of water temperatures in piezometer 5 with those in piezometers 6 and 7 is quite interesting. The thickness of the zone of aeration was 1 m greater in the latter piezometers located at arid sites than in piezometer 5 located on a moist meadow. During the summer water temperature tended to be higher in piezometers 6 and 7 rather than in piezometer 5 in spite of greater depths. Heat storage conditioned by ground moisture is of major importance. The stored heat increases with increasing moisture. Thus, the moist ground becomes warmed up at a slower rate than the arid ground. In consequence, groundwater temperature is higher in arid regions during the summer when the thickness of the zone of aeration and that of a terrain cover are identical. Note should be made of the fact that relatively high water temperatures and the lowest temperature amplitudes were reached in piezometers 6 and 7. This was largely due to greater depths to water.

The maximum water temperature was lower in a forest on river alluvium (piezometer 10) than in other piezometers. As has already been mentioned, tree crowns, the shrub layer and litter, as well as higher ground moisture serve as heat insulators. During the winter the water temperature was low. The minimum temperature of 1.9°C was reached during a spell of spring meltwater release in March 1984. The accumulation of meltwater in a small depression around the piezometer can account for this. In the case of a small thickness of the zone of aeration and high ground permeability, the infiltrating meltwater affects directly the groundwater temperature. During the winter and summer somewhat higher water temperatures were recorded in piezometer 9 located in a young forest on river alluvium.

5.3. IMPORTANCE OF THE THICKNESS OF THE ZONE OF AERATION AND LOCAL FACTORS IN SHALLOW GROUNDWATER TEMPERATURE DYNAMICS

For the purpose of studying the intensity of the effect that a varying thickness of the zone of aeration and local factors have on groundwater temperature throught the year,

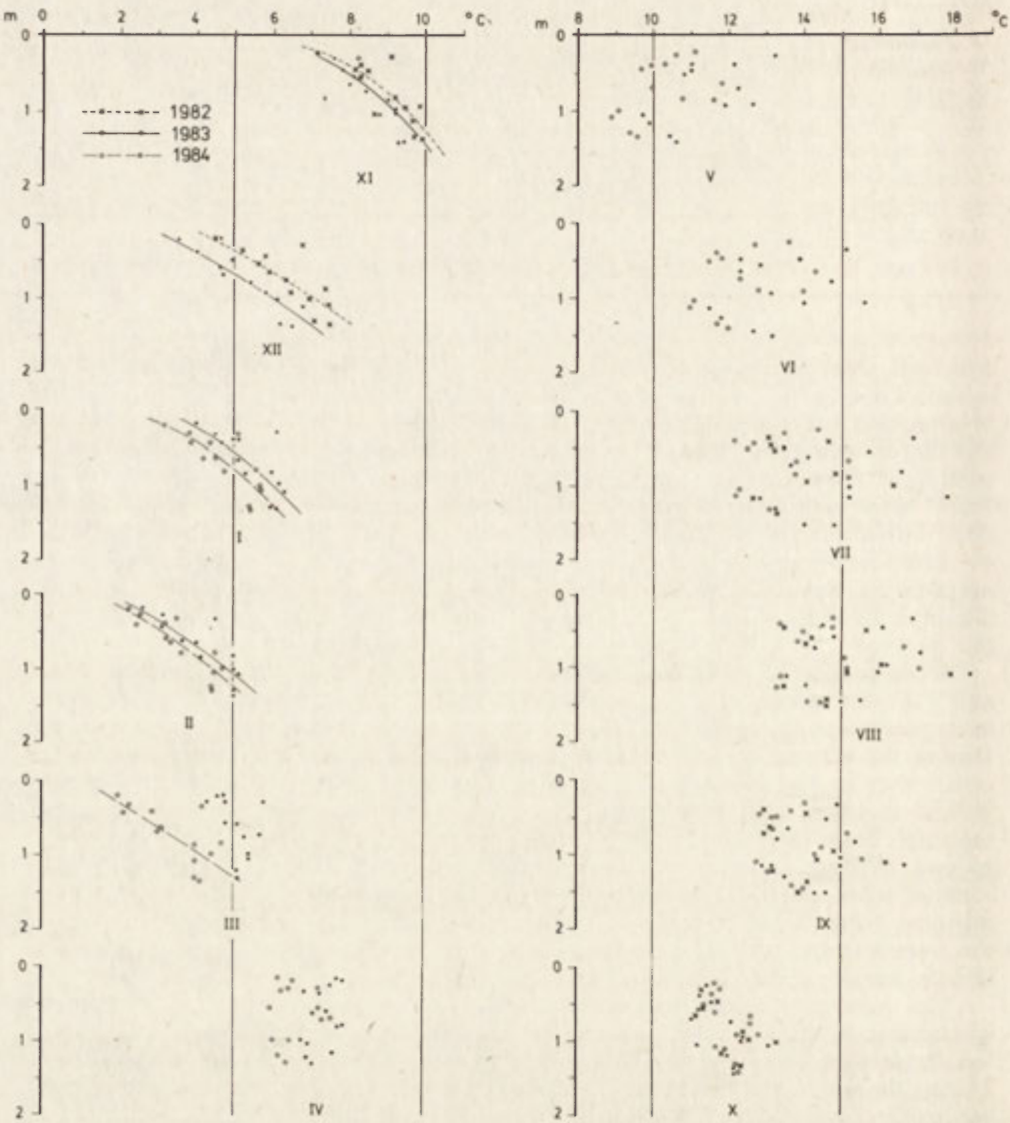

Fig. 9. Dependence of mean monthly water temperatures in piezometers on mean monthly depth to water throughout the year, July 1982 — October 1984

relationships between the mean monthly temperatures and depths to water have been plotted on the basis of data from 13 piezometers (Fig. 9). The relationships which hold for observation wells located in identical field settings have been presented before (Fig. 5).

From the comparison between Figures 9 and 5 for different periods of observation, it follows that relationships between the thickness of the zone of aeration and groundwater temperature were distinct in both the observation wells and piezometers during the winter. This is indicative of lower intensity of local factors, especially land use patterns (terrain cover), slope exposure and moisture of the zone of aeration, which affect the shallow groundwater temperature during the winter. Differences in the rate of cooling between the ground layers immediately beneath the surface are, to a great extent, reduced by a snow cover, hampered growth of the vegetation cover, frozen surface water, etc.

During the summer months water temperature in the observation wells located in similar field settings was also affected by the thickness of the arid layer (Fig. 5). However, no relationship can be established between the thickness of the zone of aeration and water temperature in the piezometers located in different field setting (Fig. 9). A large scatter of points indicates that local factors have a marked effect on the shallow groundwater temperature during the summer whereas depths to water are of less significance. As has been pointed out before, great variations of groundwater temperature during the summer are conditioned by slope exposure, ground moisture and land use patterns (terrain cover).

6. CONCLUSIONS

Annual variations of shallow groundwater temperatures reflect the meso-and microclimatic conditions. Water temperature in wells with similar geological structure, location and moisture of the zone of aeration is largely dependent upon air temperature and depth to water. The influence of air temperature on groundwater temperature becomes rapidly lessened with increasing depth. The annual water temperature amplitude and the thickness of the zone of aeration depends on the annual air temperature amplitude and the thickness of the zone of aeration. During the summer months water temperatures fall with depth, whereas they increase during the winter months. During the spring and autumn uniform water temperatures are experienced in wells with differing depths.

Microclimatic conditions affected by a different exposure of slopes, differing ground moisture and land use patterns (terrain cover) have a marked effect on shallow groundwater temperatures. It has been established that local factors affect groundwater temperature during the summer whereas the thickness of the zone of aeration is of secondary significance. Great variations of water temperature during the summer are due to differences in cooling rates between the ground layers immediately beneath the surface. The highest water temperatures were reached on exposed slopes with a southern aspect at the same thickness of the zone of aeration, whereas the lowest temperatures were recorded in the forest. Water temperature tended to be lower on a moist meadow than in arid regions. As the stored heat increases with moisture, the moist ground becomes warmed up at a slower rate than the arid region. During the winter months water temperature is largely dependent on the thickness of the zone of aeration and the influence of local factors is not marked. Microclimatic discrepancies resulting from a different exposure of slopes and terrain cover are, to a large extent, reduced by a snow cover, hampered growth of a vegetation cover, etc. Therefore, groundwater temperature becomes more uniform during the winter than during the summer.

REFERENCES

Dynowski J., 1968, *Z badań nad termiką wody gruntowej regionu krakowskiego* (Sum.: An analysis of ground water thermic conditions in environment of Cracow), Zeszyty Naukowe UJ, Prace Geograficzne 21.

Glazik R., 1978, Wpływ zbiornika wodnego na Wiśle we Włocławku na zmiany stosunków wodnych w dolinie (Sum.: The effect of a reservoir on the Vistula near Włocławek in the changes of water conditions in the valley), *Dokumentacja Geograficzna* 2—3.

Gutry—Korycka M., 1969, Z zagadnień termiki płytkich wód gruntowych (Sum.: From the problems of thermal conditions in the shallow ground water), *Wiadomości Służby Hydrologicznej i Meteorologicznej* 5(17), 1—2.

Križ H., 1983, *Hydrologie podzemnich vod*, Academia, Praha.

Marchacz W., 1960, *Hydrogeologia* (Hydrogeology), Warszawa.

Mikulski Z., 1963, *Zarys hydrografii Polski* (An outline of hydrography of Poland), Warszawa.

Nawrocka E., Sadowska E., 1964, *Temperatury wód gruntowych w Polsce w pięcioleciu 1959—1963* (Temperature of ground water in Poland during 1959—1963), Materiały PIHM 128.

Pazdro Z., 1964, *Hydrogeologia ogólna* (General hydrogeology), Warszawa.

Skibniewska H., 1963, Wpływ termiki wód gruntowych na ich własności fizyczne i chemiczne (Effects of thermal regime of ground water on their physical and chemical characteristics), *Gazeta Obserwatora PIHM* 4.

Skibniewska H., 1964, *Wyniki obserwacji temperatury wód gruntowych w Polsce* (Results of ground water measurements in Poland), Prace PIHM 84.

MODÉLISATION DU TOPOCLIMAT COMPTE TENU DE L'ANALYSE DU BILAN THERMIQUE DE LA SURFACE ACTIVE DE LA TERRE

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1. INTRODUCTION

La méthode de délimitation des topoclimats (zones homogènes du point de vue topoclimatique), élaborée au Laboratoire de Climatologie de l'Institut de Géographie et d'Aménagement du Territoire de l'Académie Polonaise des Sciences (PAN), consiste en l'analyse de la différenciation spatiale de la structure du bilan thermique de la surface active. La grandeur des flux particuliers de chaleur était définie, jusqu'à présent, à l'aide de valeurs relatives (J. Paszyński 1980). Pour valeurs relatives de la composante donnée du bilan thermique on adoptait les écarts, positifs ou négatifs, des valeurs qui apparaissent par temps de faible nébulosité, avec une vitesse du vent au-dessous de 2 ms^{-1}, avec une humidité modérée du sol sur les terrains plats à horizon découvert et couverts d'un tapis herbeux d'environ 10 cm.

A l'aide de cette méthode, quelque peu modifiée, on a établi une série de cartes topoclimatiques pour différentes régions de la Pologne (J. Grzybowski 1980, 1983a, 1983b, 1986 et autres). La méthode appliquée permet de présenter la différenciation spatiale des topoclimats dans des conditions de temps du type radiatif. Il est vrai que dans les légendes des cartes s'étaient trouvées des remarques sur la variabilité des topoclimats dans différents états du milieu et dans différents types de temps mais ce n'étaient que des informations générales, descriptives. Le problème de la présentation de la différenciation spatiale des topoclimats par différents types de temps restait toujours ouvert. De plus, les qualifications relatives s'étaient avérées insuffisantes.

La variabilité locale de la structure du bilan thermique, tant dans le temps que dans l'espace, est si grande qu'il était utile d'entreprendre une tentative de modélisation de cette structure dans les conditions physico-géographiques changeantes. Dans d'autres centres de recherche, qui s'occupent de l'analyse du bilan thermique de la surface active, sont entrepris, depuis peu, des essais de modélisation des grandeurs des composants particuliers de ce bilan grâce à l'application de la télédétection (W. H. Terjung, P. A. O'Rourke 1984; B. Itier 1980: B. Itier, Ch. Riou 1982 et autres). Cela a augmenté avant tout l'étendue de l'information sur la différenciation spatiale de la température de la surface active qui est un élément important permettant de calculer les flux particuliers de chaleur. Bien que B. Seguin (1984) écrive que les photos de satellite permettront l'étude des changements saisonniers et à court termę, il souligne cependant que l'on n'est pas encore parvenu à une continuité de mesures respectives à l'aide de satellites stationnaires. C'est pourquoi, à notre avis, les modèles mathématiques de simulation pour la détermination des flux de chaleur, et particulièrement pour l'établissement des cartes topoclimatiques, sont toujours nécessaires.

2. BUT DE L'ETUDE

L'auteur a entrepris un essai d'établissement d'une formule simple permettant d'évaluer la grandeur approximative des flux particuliers de chaleur à partir de l'analyse des traits de la surface active qui, de manière décisive, influencent la grandeur de ces flux. Le but de ce travail était donc de préciser quels sont les traits qui influencent de manière décisive la grandeur des flux particuliers de chaleur, de présenter ces traits sous une forme permettant leur application dans la formule et, ensuite, d'élaborer les formules eux-mêmes. L'élaboration de ce type de formules créerait, selon les visées de l'auteur, une base pour l'analyse spatio-temporelle des variabilités de la structure du bilan thermique en valeurs absolues, dans différents types du milieu géographique, dans différentes phases de la période végétative et par différents types de temps.

3. MATÉRIAUX DE BASE

Pour l'analyse de l'influence des propriétés physiques de la surface active sur la grandeur des flux particuliers de chaleur, nous avons utilisé aussi bien les résultats de certaines recherches menées au Laboratoire de Climatologie de l'Institut de Géographie de l'Académie Polonaise des Sciences à Varsovie que les résultats des recherches publiés et non publiés, mis à notre disposition par leurs auteurs pour leur utilisation dans la présente étude. Ces résultats proviennent des types variés du milieu géographique de la zone tempérée: champs cultivés sur plateaux (friches, champs de betteraves, de froment, de maïs, de pommes de terre) pendant différents stades phénologiques et par différents types de temps, ainsi que des prairies, sèches et humides, et des terrains boisés (bois de pins et de bouleaux). En plus nous avons utilisé les résultats des mesures des flux de chaleur provenant des régions désertiques et des surfaces découvertes des réservoirs d'eau. Une analyse plus détaillée des matériaux de base sera publiée séparément.

Dans la plupart des résultats des mesures, utilisés dans notre étude, le flux de chaleur latente (E) n'était pas calculé directement mais comme le reste du bilan $E = Rn-H-S$ (où Rn = rayonnement net, H = flux turbulent de chaleur sensible, S = flux de chaleur dans le sol). En conséquence nous avons choisi, pour la suite de notre analyse ceux des flux qui étaient calculés directement c'est-á-dire Rn, H et S.

Des matériaux dont il était question plus haut on avait obtenu, pour chacun des flux signalés, pres de 4.000 données. Elles représentent différentes périodes d'observation: depuis une journée jusqu'à plus de dix jours; elles informent sur le cours journalier, ou étalé sur 24 heures, des valeurs momentanées des flux de chaleur mesurés par intervalles allant d'une heure à 15 minutes. Elles représentent aussi bien le temps ensoleillé que nuageux avec des précipitations.

4. CHOIX DES ÉLÉMENTS DU MODÈLE

Il est difficile de déterminer les élements du milieu géographique qui influencent, de manière décisive, la grandeur des flux particuliers de chaleur du fait du nombre important de facteurs qui exercent, directement ou indirectement, une influence sur la structure du bilan thermique. Au résultat des essais antérieurs pour definir ces éléments (J. Grzybowski 1984a, 1984b) et de l'analyse de la littérature (B. Itier, N. Katerji 1983; D. H. Miller 1982) l'auteur a admis que la plus grande influence sur la grandeur du flux turbulent de la chaleur sensible (H) est exercée, en plus du rayonnement solaire, par l'humidité du sol et le stade phénologique des plantes.

Plus l'humidité du sol diminue, plus la quantité d'énergie du rayonnement utilisée pour chauffer les couches d'air au niveau du sol augmente. Le stade du développement des plantes, et aussi la phase des travaux des champs, décident de la capacité des plantes à l'évapotranspiration. La part des surfaces vertes dans la surface analysée, et aussi la hauteur et la densité du tapis végétal constituent, elles aussi, les traits du stade du développement des plantes qui influent sur la grandeur des flux H et E. De plus, la grandeur du flux H peut diminuer à la suite des précipitations. Au cours des travaux précédents (J. Grzybowski 1984a, 1984b, 1986) on avait analysé aussi l'influence de la vitesse du vent. Cependant, au résultat d'un plus grand nombre d'informations réunies, il s'est avéré qu'avec une vitesse du vent inférieure à 5 ms^{-1}, ce facteur pouvait être négligé.

A partir de l'analyse des résultats des études du flux de chaleur, dans le sol, on a admis que la variabilité journalière de ce flux est le plus influencée, en plus du rayonnement solaire, par la conductibilité thermique du sol, par l'humidité du sol et aussi par la hauteur et la densité du tapis végétal. Ce ne sont évidemment pas les seuls traits de la surface active qui influent sur la grandeur du flux de chaleur dans le sol. Ils sont toutefois les plus importants dans le cas d'un temps du type radiatif. Lorsqu'augmente l'importance de l'advection, le réchauffement du sol (ou le retardement de son refroidissement) peut avoir lieu aussi dans la soirée. Le flux Rn joue alors un rôle moins important tandis que grandit l'importance de la température actuelle de l'air. Les précipitations causent, le plus souvent, une diminution supplémentaire du flux S.

Comme on a constaté au cours d'analyses antérieures, l'influence de la conductibilité thermique des sols sur la grandeur du flux S se révèle avant tout au moment où change l'humidité du sol. Les grandeurs de la conductibilité thermique des sables et des argiles, donc des terres qui dominent sur la Plaine Polonaise, diffèrent presque insensiblement. Cette différenciation augmente seulement à mesure que grandit l'humidité de ces terres. Les sols tourbeux dont les propriétés thermiques et d'humidité diffèrent sensiblement de celles des sols minéraux constituent la seule exception. Dans l'analyse ultérieure, la conductibilité thermique des sols était traitée comme une fonction de leur humidité.

5. MODÈLE DE SIMULATION DES FLUX H ET S ET SA VÉRIFICATION

L'influence des traits particuliers de la surface active sur la grandeur des flux H et S avait été étudiée à travers l'analyse de la fonction $H = f(Rn)$ et $S = f(Rn)$, ce qui a été décrit séparément (J. Grzybowski, 1984a, 1986). En cet endroit nous rappellerons seulement que pour l'analyse de la fonction $H = f(Rn)$ et $S = f(Rn)$ on avait choisi les résultats des recherches au cours desquelles ne changeait qu'un seul des traits du milieu jugé décisif pour la grandeur du flux donné. Les autres traits ne subissaient aucun changement. De cette manière, en analysant p. ex. les résultats des recherches sur la structure du bilan thermique au-dessus des champs à humidité du sol, densité des herbes et conditions météorologiques semblables, on obtenait l'image de la fonction $S = f(Rn)$ dans les conditions où la hauteur des plantes changeait.

En marquant l'humidité du sol comme a_*, la hauteur des plantes comme h_*, leur densité comme g_*, et en admettant l'invariabilité des conditions météorologiques nous obtenons:

$$S = f(h_*, g_*, a_*, Rn)$$

En divisant $\frac{S}{Rn}$ nous obtenons $h = \operatorname{tg}\alpha$ qui est le coefficient dépendant de la hauteur h_*.

De manière semblable on a analysé l'influence des autres traits de la surface active obtenant, successivement, les coefficients pour les variables h, g et a dans des conditions où trois de ces variables restent constantes et la quatrième subit des changements. On avait analysé aussi la fonction $H = f(Rn)$ dans des conditions de changements successifs de l'humidité du sol, de la hauteur et de la densité du tapis végétal et aussi de la part des surfaces vertes dans l'unité spatiale analysée. On a obtenu de cette manière les coefficients propres aux différents états du milieu environnant.

Les flux H et S constituent une certaine partie du flux Rn. Nous pouvons écrire cela sous la forme

$$H = xRn \text{ et } S = yRn$$

où x et y indiquent les ensembles de ces traits de la surface active qui causent la diminution de la grandeur du flux Rn jusqu'a la grandeur des flux H ou S.

Renouant avec les régularités constatées nous pouvons écrire les équations

$$H_1 = a_1 h_1 l \ p \ Rn$$
$$S_1 = a_2 h_2 g \ p \ Rn$$

dans lesquelles H_1 et S_1 indiquent la grandeur des flux de chaleur calculés à l'aide de la méthode de simulation, a — coefficients dépendant de l'humidité de la surface active (a_1 pour le flux H_1 et a_2 pour le flux S_1), h — coefficients dépendant de la hauteur de la végétation (h_1 pour le flux H_1 et h_2 pour le flux S_1), l — coefficient dépendant de la part des surfaces vertes dans la surface étudiée, g — coefficient dépendant de la densité des plantes, p — coefficient dépendant de la somme des précipitations (mm) le même pour les flux H et S).

Les deux équations ont été vérifiées sur 300 résultats des mesures des flux Rn, H et S, qui n'avaient pas été utilisés au cours de l'élaboration des équations. Ces mesures constituent donc un matériau indépendant. Les fonctions calculées $H_1 = f(H)$ et $S_1 = f(S)$, où H et S indiquent les flux déterminés à l'aide des mesures des gradients de la température et de la vitesse du vent, et H_1 et S_1 — ce sont les mêmes flux calculés à l'aide de la méthode de simulation. Pour ces fonctions on avait calculé l'équation de la droite:

$$H_1 = 0{,}92 \ H + 6{,}21 \qquad r = 0{,}93$$
$$S_1 = 1{,}02 \ S - 1{,}20 \qquad r = 0{,}94$$

Les deux coefficients de la corrélation sont essentiels du point de vue statistique. L'erreur relative des calculs du flux turbulent de chaleur sensible (H_1) pour les heures du jour est de 20%, par contre pour le flux de chaleur dans le sol (S_1) est de 15%. Ce sont les valeurs moyennes de l'erreur. Pour le flux S_1 il est le plus petit (env. 5—10%) dans les heures autour de midi, pour le flux H_1 par contre pour les heures avant midi par jours à température ne dépassant pas 20°C.

La méthode élaborée peut être appliquée pour les zones plates, situées entre $\varphi = 45°$N et $\varphi = 55°$N constituées par des terres sablonneuses et argileuses. Cette méthode peut être appliquée pour les heures du jour, dans des conditions instables, pendant les mois de la période végétative. Elle permet d'obtenir les valeurs moyennes des flux H et S pour environ 15 minutes, parce que cette période des mesures est caractéristique pour la majorité des matériaux de base. Dans le cas du flux S la formule ne prend pas en compte les changements qui résultent du décalage de phase par rapport au flux Rn dans les heures de l'après-midi.

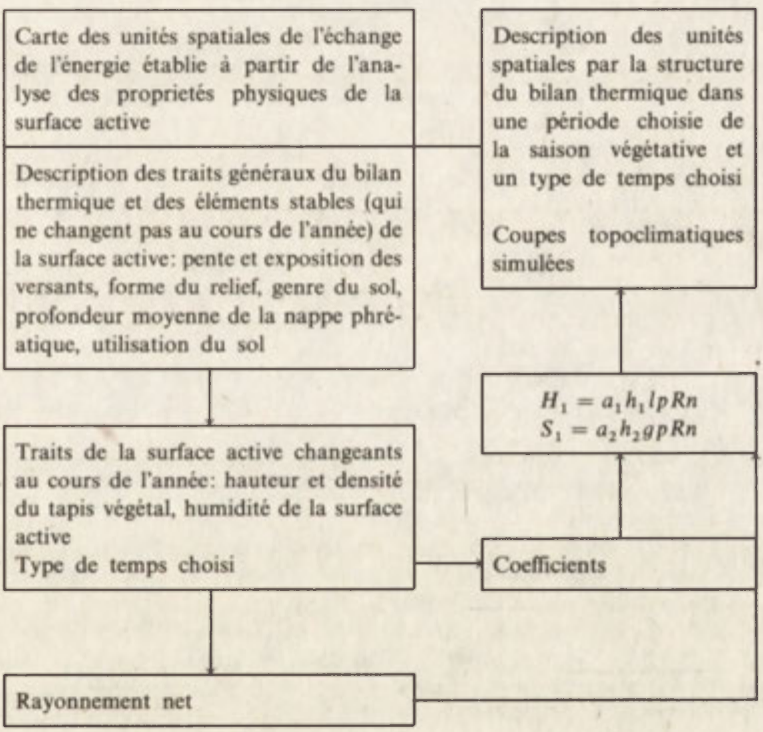
On peut admettre pour le rayonnement net une valeur quelconque de même que les valeurs des autres traits de la surface active qui changent au cours de la période végétative. La grandeur de ce flux peut être calculée aussi directement, sur le terrain, tout comme la grandeur des éléments particuliers du milieu naturel pris en compte dans les formules. Les coefficients particuliers, et leurs valeurs, dans différents états du milieu géographique avaient été élaborés sous forme de tableaux (J. Grzybowski 1985).

6. CONCLUSIONS

La méthode de simulation pour calculer les valeurs approximatives des flux de chaleur, présentée dans cette étude, n'a pas pour but de remplacer les mesures prises sur le terrain avec un appareillage approprié. Elle peut toutefois, malgré toutes ses limitations et ses simplifications, nous rendre conscients de l'étendue des variation des coefficients particuliers du bilan thermique dans différents types du milieu géographique et dans des types choisis de temps, avec la grandeur quelconque du rayonnement net (Rn) et des traits particuliers du milieu. Cela a une énorme importance dans les pronostics des changements topoclimatiques résultant p.ex. de l'afforestation ou de la deforestation, des améliorations foncières etc.

Le schéma des étapes de la procédure pour définir la valeur approximative des flux particuliers de chaleur est représenté sur la Figure 1. Il en découle que la carte de

Fig. 1. Schéma d'établissement de la carte de l'échange de l'énergie entre l'atmosphère et le sol et de calcul des valeurs des flux de chaleur particuliers.



l'échange de l'énergie entre l'atmosphère et le sol peut contenir des informations générales sur la structure du bilan thermique par un temps du type radiatif et en pleine saison de végétation. Il est possible d'appliquer les définitions relatives ou calculées selon la méthode de simulation proposée. La légende de la carte devrait contenir aussi la description de ces traits de la surface active qui subissent des changements au cours de l'année, c'est-à-dire de l'humidité du sol, de la hauteur et de la densité du tapis végétal etc. Cela rendra possible l'évaluation de la grandeur des flux particuliers de chaleur dans différents états du milieu. Les valeurs des flux H et S (et en conséquence aussi du flux E en tant que reste du bilan: $E = Rn - H - S$), dans des conditions choisies du développement de la végétation, peuvent constituer un supplément de la carte et l'élargissement de son contenu. Elles peuvent aussi être utilisées sur des coupes à travers les types choisis du milieu géographique, dans un type de temps admis et dans un mois donné. Elles peuvent aussi être élaborées sous forme de graphiques du déroulement de leurs valeurs momentanées, à un moment donné de la journée, au cours de la période végétative, au-dessus des différents types de la surface active.

L'établissement des cartes des topoclimats, à la lumière de la méthode proposée, consisterait en une documentation détaillée de ceux des éléments de la surface active qui, de manière décisive, influencent la grandeur des flux particuliers de chaleur. Cette documentation devrait contenir l'analyse de la variation des traits biotiques et abiotiques de la surface active dans les mois successifs de la période végétative. L'attribution à ces éléments de coefficients appropriés et leur substitution dans les formules proposées peut augmenter l'étendue de l'information sur le climat local au cours de n'importe quel mois de la période végétative.

BIBLIOGRAPHIE

Grzybowski J., 1980, *Mapa topoklimatyczna Kotliny Biebrzańskiej 1:100 000* (Carte topoclimatique de Kotlina Biebrzańska 1:100000), manuscrit, Laboratoire de Climatologie de l'Institut de Géographie et d'Aménagement du Territoire de l'Académie Polonaise des Sciences, Varsovie.

Grzybowski J., 1983a, Wyróżnianie i klasyfikacja jednostek wymiany energii na powierzchni czynnej na przykładzie Kotliny Biebrzańskiej (Delimitation and classification of energy exchange units on the active surface on the example of the Biebrza Basin), *Przegląd Geograficzny* 55, 2, 341—358.

Grzybowski J., 1983b, An attempt at the delimitation and classification of energy exchange spatial units on the example of the Beskid Makowski mountain range, dans: X Międzynarodowa Konferencja Meteorologii Karpackiej, Kraków 16—20 IX 1981, *Zeszyty Naukowe UJ* 680, Prace Geograficzne 57, 207—211.

Grzybowski J., 1984a, Essai d'établissement d'un modèle de l'influence des éléments du milieu géographique sur le bilan d'énergie de la surface active de la terre, *Geographia Polonica* 50, 225—232.

Grzybowski J., 1984b, Simplified method of calculation of the components of the earth's surface heat balance, Symposium on Applied Climatology 21st August 1984, *Zürcher Geographische Schriften* 14, 83—86.

Grzybowski J., 1985, *Wyróżnianie i klasyfikacja topoklimatów na podstawie właściwości fizycznych powierzchni czynnej* (Délimitation et classification des topoclimats à partir des propriétés physiques de la surface active), manuscrit, Laboratoire de Climatologie de l'Institut de Géographie et d'Aménagement du Territoire de l'Académie Polonaise des Sciences, Varsovie.

Grzybowski J., 1986, Mapa wymiany energii między atmosferą a podłożem jako podstawa analizy funkcjonowania środowiska geograficznego (Map of energy exchange between atmosphere and the underlying ground as a basis for analysis of the functioning of the environment), *Przegląd Geograficzny* 58, 1—2, 131—142.

Itier B., Riou Ch., 1982, Une nouvelle méthode de détermination de l'évapotranspiration réelle par thermographie infrarouge, *Journal des Recherches Atmosphériques* 16, 2, 113—125.

Itier B., Katerji N., 1983, Une expression simplifiée de la résistance aérodynamique d'un couvert végétal, *Acta Oecologica* 4, 3, 215—227.

Paszyński J., 1980, Metody sporządzania map topoklimatycznych (Méthodes d'élaboration des cartes topoclimatiques), *Dokumentacja Geograficzna* 3, 13—28.

Seguin B., 1984, Bioclimatological aspects of crop production, dans: A. Berg ed., *Application of remote sensing to agricultural production forecasting*, Rotterdam.

Terjung W. H., O'Rourke P. A., 1984, Summer energy budget simulation for a latitudinal transect along the East Coast of the Americas, *Journal of Climatology* 4, 2, 111—122.

GEOGRAPHIA POLONICA 53, 1987

TYPES OF BIOCLIMATE IN POLAND

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INTRODUCTION

Atmospheric environment continuously affects man's organism, yet with changing intensity, depending on time and space. This is caused by the action of various groups of climatic impulses which bring about more or less desirable functional, metabolic, and morphological changes in man's organism, especially where the impulses display high intensity, exceeding the ability of a living organism to maintain psychosomatic balance.

The bioclimatology of man knows many stimuli which affect human organism through: skin, respiratory system, organs of smell, audition and sight, and throught the nervous system. All of them can be classified into three basic groups (Kozłowska – Szczęsna, Krawczyk, Błażejczyk and Kuczmarski 1985):

– the group of physical impulses,
– the group of chemical impulses,
– the group of biological impulses.

Taking as the basis the intensity of external stimuli (chiefly physical) which affect man's organism, a map of types of bioclimate in Poland has been compiled (Fig. 1), showing which impulses, or groups of them, play the main role in shaping bioclimatic conditions in different parts of Poland. The criterion by which the types of bioclimate have been distinguished was the intensity of impulses, determined on the basis of bioclimatological scales and comparison with the norms used for evaluation of bioclimatic conditions. The values of individual bioclimatic elements were analyzed for different ranges, with their respective degrees of impulse intensity (Kozłowska – Szczęsna et. al. 1985; Kozłowska – Szczęsna 1986 a, b). The response of human organism to atmospheric stimuli depends in the first place on the intensity of those stimuli; too weak impulses may result in loss of adaptability or excess vulnerability, moderate ones have an invigorating, hardening and therapeutic effect, while too strong an impulse may be detrimental and may strain the organism (Flemming 1979; Kozłowska – Szczęsna 1984a, Kozłowska – Szczęsna and Grzędziński 1984). The action of individual impulses may be amplified or weakened, among other factors by their joint occurrence (synergism); in that case, an otherwise irrelevant and weak impulse can amplify the effect of action of other impulses.

Four bioclimatic types have been distinguished in Poland: strong stimulating bioclimate (type I), moderate stimulating bioclimate (type II), mild stimulating bioclimate (type III), and weak stimulating bioclimate (type IV). Two subtypes have been marked within each of these four types: bioclimate of forested terrain, with spare features (subtype A), and bioclimate of urbanized terrain, with strain features (subtype B). The initial material for this paper was taken from data collected by nearly 100 state-run meteorological stations and posts between 1961 and 1970. The

Fig. 1. The types of Poland's bioclimate: I — strong stimulating bioclimate, II — moderate stimulating, III — mild stimulating, IV — weak stimulating; subtypes: A — bioclimate of forested areas with spare features, B — bioclimate of urbanized areas with strain features.

decade was then marked by hot summer in 1963, cool summers in 1962 and 1965, frosty winters 1962/63 and 1969/70, wet summers in 1966 and 1970, and dry summers in 1964 and 1969; in all, the period had embraced a range of extreme weather conditions and, therefore, be accepted as representative for evaluation of Poland's bioclimate.

It is noteworthy that non – Polish bibliography gives few examples of cartographic analyses made from the point of view of man's bioclimatology. One such example is a map of West Germany (Becker and Wagner 1972).

DISTINGUISHING THE TYPES OF BIOCLIMATE IN POLAND – AN ATTEMPT

Most of the Polish territory is under the weak stimulating bioclimate (type IV) marked by mean values of bioclimatic elements and indices. It embraces the expanse of the Middle-Polish Lowlands and the Northern sub-Carpathians, both of which are characterized by hardly diversified terrain relief that naturally offers no hindrance to free penetration by oceanic air masses, with westerly circulation, and of continental currents, with easterly circulation. Free circulation of the air tones down stimuli. The climatic conditions of that zone are typical of Poland and the most widespread. The zone is inhabited by most of the country's population which is adapted to just these conditions. A change of location within this type of bioclimate requires only minor adaptation and re-adaptation upon return home. There is a slight difference between climatic impulses in the western and eastern, and in the northern and southern limits of this typological unit. River valleys, too, are marked by slightly different bioclimatic conditions, with occasional, passing domination of mild or even moderate stimuli, depending on the kind of ground and on surface cover.

The mild stimulating bioclimate (type III) occurs both in positive and negative landforms, covering the South-Baltic and East-Baltic Lake Districts, the Silesian-Cracovian uplands, the Central Małopolska uplands, part of the Eastern Małopolska upland and part of the Sudetes and Carpathian uplands. Occasional moderate impulses tend to occur in river valleys, and they depend on the kind of ground, depth of the valley and surface cover. The bigger the share of fenland in the investigated area, the worse the bioclimatic conditions, owing to increased air humidity and possibility of sultriness on hot summer days. The mild stimulating bioclimate of surface elevations (250—500 m a. s. l.) tends to be modified by the solar or windy exposition of the slopes and diversification of surface cover and relief, which allows to dose climatic impulses.

The moderate stimulating bioclimate (type II) embraces seaside lowlands of the South-Baltic Lake District, submontane basins and the parts of elevations up to 750 m a.s.l. The moderate impulses are a result above all of weaker solar radiation and lesser wind speed than on the Baltic coast or in higher parts of the mountains. There are periods of occasional strong stimuli as well, depending on landform, solar or windy exposition, surface cover, and influence of sea breeze in the north or foehn in the south of Poland. Passing strong impulses in narrow river valleys and basins may be caused by considerable contrasts between daytime and nighttime temperatures and air humidity, as well as by residual cool air, frequent fogs, sultriness, and shortage of solar energy owing to the horizon shading by surrounding hills. The moderate character of the impulses can turn into strong also in the mountains due to high levels of solar radiation and considerable wind speed.

The strong stimulating bioclimate (type I) embraces the Baltic coast strip and higher part of the mountains (over 750 m a.s.l.). The values of the meteorological elements studied stay within the ranges perceived as strong impulses of radiation, thermal-humidity, or mechanical. Compared to the interior, the coastal areas are characterized by domination of factors strongly stimulating man's organism. A peculiar feature of this type of terrain is the presence of a favourable sea aerosol in the air. An important factor is the sea- and land-breeze. The former brings with it a sudden chill after a period of heat, which may cause perturbations in heat and water balances of human organism. The central part of the Baltic coast is marked by strongest climatic stimuli, with slightly weaker factors in the western part of the coast — on the Pomeranian Bay (*Zatoka Pomorska*), and in the east — on the Bay of Gdańsk.

Large, inland reservoires of water show similar bioclimatic effects as the sea, yet on a lesser scale. Areas surrounding both man-made and natural reservoires tend to have moderate to strong stimuli, depending on the size of the reservoire and development of its nearest vicinity. Both the big inland water reservoires and the Baltic sea produce a warming-up effect in the autumn and a cooling effect in the spring.

Bioclimatic conditions in the Tatra and Karkonosze mountains are similar to those in medium—high mountains; there is considerable intensity of stimuli, brought about by high intensity of solar radiation, strong winds, and decrease of temperature and molecular pressure of oxygen with increasing altitude. From the bioclimatic point of view, an important factor are mountain foehns; they raise the temperature and reduce humidity of the air, and cause sudden changes in air pressure which may in turn have an adverse impact on man's psychological and physical feeling.

The map of bioclimatic types (Fig. 1) shows two subtypes (A and B) in each of the four major types. The numerical characteristic of the types are shown in Table 1.

TABLE 1. Characteristics of the types of bioclimate in Poland

Mean yearly values, 1961 – 1970	type I strong stimulating	type II moderate stimulating	type III mild stimulating	type IV weak stimulating
Sunshine duration (in hrs.)	1350 – 1616	1435 – 1580	1300 – 1550	1400 – 1650
Relative sunshine duration (%)	30 – 35	32 – 35	39 – 34	31 – 36
Cloudiness in 2nd term (%)	61 – 78	66 – 73	63 – 76	65 – 75
Number of days with cloudiness ≤ 50%	84 – 145	115 – 140	80 – 120	95 – 140
Number of days with 100% cloudiness	100 – 160	120 – 180	115 – 200	130 – 160
Temperature in 2nd term (°C)	0.2 – 9.9	7.5 – 11.0	8.5 – 11.3	9.0 – 11.5
Extreme temperature amplitudes (°C)	52 – 63	55 – 68	57 – 67	60 – 67
Number of days with max. temperature ≥ 25°C	0 – 19	6 – 30	19 – 37	30 – 40
Number of days with max. temperature ≥ 30°C	0 – 3	1 – 3	2 – 6	3 – 6
Number of days with min. temperature ≤ – 10°C	6 – 60	15 – 50	20 – 40	20 – 38
Number of days with max. temperature ≤ – 10°C	1 – 20	1 – 7	2 – 8	2 – 7
Relative humidity in 2nd term (%)	66 – 85	67 – 75	65 – 70	65 – 72
Number of days with sultriness: e ≥ 18.8 hPa	0 – 15	5 – 16	7 – 26	15 – 20
Wind speed in 2nd term ($m \cdot s^{-1}$)	4.0 – 5.5	2.0 – 4.0	2.0 – 4.5	2.5 – 4.6
Number of days with winds ≥ 8 $m \cdot s^{-1}$	40 – 110	20 – 80	40 – 60	20 – 60
Number of days with rainfalls ≥ 0.1 mm	145 – 250	165 – 210	130 – 180	150 – 180
Number of days with snow cover	60 – 215	60 – 130	80 – 105	60 – 100
Number of days with fog	55 – 215	30 – 95	20 – 70	25 – 80
Number of days with thunderstrom	17 – 35	18 – 30	15 – 30	10 – 25

Wooded areas are dominated by spare features of the bioclimate (subtype A), which are an effect of the toning action of vegetation on radiation, thermal-humidity, and mechanical stimuli. The main bioclimatic significance of the forest is that it improves the hygiene of the air, absorbing dust and gaseous pollutants, while at the same time muffling noise and enriching the air with aromatic substances (the biological impulses). The positive psychological impact of greenery on man is not insignificant, either.

Urbanization and industrialization have and adverse impact on the radiation, thermal-humidity, mechanical and acoustic stimuli, which is why these types of areas have been marked as a separate subtype (B). Urbanized and industrialized areas tend to display wide diurnal amplitude of air temperature, considerable variability of wind speed (depending on street pattern), variable amount of solar radiation at street surface (depending on the height and tighteness of land development), and considerable intensity of noise. There is also high physical and chemical air pollution (the chemical impulses), All of these strain human organism.

CONCLUSION

Summing up, one can say that bioclimatic conditions are not the same all over Poland. The weak stimulating bioclimate, is the most widespread type, and is typical of the Polish lowlands. The second most widespread is the mild stimulating bioclimate of the northern and southern uplands and larger river valleys. Despite considerable instability of weather, a characteristic feature of Poland's climate, lowland terrains enjoy greater stability of either good or bad weather as compared with the seabound or mountaineous regions. A change of location within these bioclimatic types requires either a relatively short period of adaptation and re-adaptation upon return, or none at all. The period good for climatotherapy there lasts from April through October; the best time being end of spring and beginning of summer (May, June) as well as beginning of autumn (September). The aforementioned terrains of weak and mild bioclimatic conditions can be useful for rest and resort treatment of especially people advanced in years, convalescents, and generally all those who feel ill under stronger stimuli (Kozłowska—Szczęsna 1984 a and b; Kozłowska—Szczęsna and Grzędziński 1983, 1984). Beneficial from that point of view are localities situated on plains, far from any vast fenlands (to avoid possibility of sultriness on warm, summer days), and characterized by clean air and presence of forest or other wooded area (to protect from excessive solar radiation and strong winds).

The moderate stimulating bioclimate should be seen as the intermediate type between the weak and mild lowland bioclimates and the strong stimulating bioclimate of seabound or mountaineous terrains. Moderate stimuli have an invigorating, hardening and even therapeutic effect on man's organism. This type of bioclimate occurs along a narrow strip in the northern part of Poland, covering the seaside lowlands except the sea coast, and the Sudetes and Carpathian uplands — in the south. Their bioclimatic conditions are recommended for rest and resort treatment of childern and are a good location for sports training camps.

The strong stimulating bioclimate is observed at the Baltic coast and in the mountains. The central part of the coast is characterized by particularly strong impulses which tend to decrease both east- and westwards. The westerly coast, on the Pomeranian Bay, and the easterly coast on the Bay of Gdańsk, have bioclimatic conditions for treatment of children and elderly people with well-functioning thermoregulating systems and considerable adaptability. Favourable climatotherapeutic period at the seaside is from June through September, but even then day-to-day changes of weather are relatively frequent and take place on half of all the days in a month.

The mountains are marked by high variability of bioclimatic conditions, from moderate on the slopes to strong down at valley bottoms and at the peaks, which may be a counterindication for geriatric treatment. Diversified terrain relief modifies intensity of stimuli, thus enabling to dose them. However, the latter is possible only up to the upper limit of the forests, as long as it is possible to take advantage of wooded terrain. The strongest stimuli can be encountered at peaks of mountains, due to increased solar radiation and greater wind speed. The danger of sunstrokes is very high, especially where snow covers the peaks. Clean mountain air and reduced amount of allergens have a positive effect in diseases of the respiratory system and in allergies — except in blooming time, i.e. May and June. There are conditions for climatotherapy in the uplands and the mountains almost throught the year (except beginning of winter) and they are particulary favourable in September and October. Radical day-to-day changes of weather occur chiefly in winter and early spring.

Both in the mountains and at the seaside, variability of stimuli is high not only in terms of space, but also as a result of considerable variability of weather. There is, therefore, high, short-time variability of stimuli. It should be remembered that in a strong stimulating climate one undergoes a period of adaptation whose duration depends on one's individual sensitivity and condition of health. Those who will feel well in a strong stimulating type of bioclimate will be young people and adults with well-functioning thermoregulating systems and smooth adaptability, as well as those who are insensitive to rapid changes of weather.

Forests and woodland have an important effect on bioclimatic conditions, which depends on the size of the forest, species composition of the trees, density of the stand, the forest's vertical structure, type of soil, and relief. Wooden terrains tone down bioclimatic conditions, and this effect manifests in: cleanliness of the air and the latter's being enriched with aromatic substances, reduced solar radiation, slowed wind speed, diminished durinal range of temperature and air humidity, absorption of gaseous pollutants, interception of descending dust, interception of precipitations by crowns of trees, muffling of noise, longer remaining snow cover and, psychological influence.

Urbanization and industrialization also have an important impact on bioclimatic conditions, yet in a negative sense. The action of individual impulses changes, among other factors because they tend to occur together (synergism). The main feature of city bioclimate is poor quality of the air. A sign of poor quality of the air is the "pollution dome" hovering above. The layer of polluted air over cities may be as much as several kilometres thick, depending on atmospheric conditions. Pollution reduces transparency of the atmosphere and weakens the effect of solar radiation, including the biologically active ultraviolet radiation, and brings about more frequent fogs. Gas-saturated fog ("smog") is a health hazard for man. Polluted air contains increased amount of pathogenic microorganisms.

REFERENCES

Becker F., Wagner M., 1972, Das Bioklima in der Bundesrepublik Deutschland, *Geographisches Taschenbuch 1970/1972*, Wiesbaden.

Flemming G., 1979, *Klima – Umwelt – Mensch*, VEB Gustav Fischer Verlag, Jena.

Kozłowska – Szczęsna T., 1984a, Bioklimat polskich uzdrowisk jako podstawowa cecha ich warunków środowiskowych (Bioclimate of Polish resorts as the essential feature of their environmental conditions), *Dokumentacja Geograficzna* 1 – 2, 13 – 88.

Kozłowska – Szczęsna T., 1984b, Les conditions bioclimatiques en tant que basé d evaluation du milieu géographique des stations de cure polonaises, *Geographia Polonica* 49, 129 – 138.

Kozłowska – Szczęsna T., ed., 1986 a, Wyniki badań bioklimatu Polski, cz. 1 (Results of bioclimatic research of Poland, part 1), *Dokumentacja Geograficzna* 3.

Kozłowska – Szczęsna T., 1986b, Klimat Polski a zdrowie człowieka (Poland's climate and human health), *Problemy Naukowe Współczesności*.

Kozłowska – Szczęsna T., Grzędziński E., 1983, Rola uzdrowisk polskich w leczeniu geriatrycznym (The role of Polish health resorts in geriatric treatment), *Problemy Uzdrowiskowe* 1 – 2 (183/184), 41 – 54.

Kozłowska – Szczęsna T., Grzędziński E., 1984, The role of Polish health resorts in geriatric treatment, *Archives of Gerontology and Geriatrics*.

Kozłowska – Szczęsna T., ed., 1985, *Metody badań bioklimatu człowieka* (The methods of man's bioclimate research), Problemy Uzdrowiskowe 1 – 2 (207/208).

GEOGRAPHIA POLONICA 53, 1987

A MODEL FOR BIOCLIMATIC EVALUATION AND TYPOLOGY OF HEALTH RESORTS AND RECREATION AREAS. CONCEPT OF A METHOD

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Applied climatology has perceived a clear need of objective evaluation and typology of climates for specific purposes. Bioclimatology, too, is in need of ways to objectively evaluate geographical environment in health resorts for their usefulness for recreation and climatotherapy.

The paper presents an attempt to employ simple mathematical and physical models for evaluation of bioclimatic conditions of a given area and typology of selected facilities. Data collected from 1961 till 1970 from 19 Polish health resorts, located in different bioclimatic conditions, were used to check functioning of the model (Fig. 1).

The descriptive method and quality classification are the most frequently used techniques in evaluation of bioclimatic conditions. Their chief deficiency, however, is that they are too subjective. Technique of terrain evaluation which would involve a model has been rarely employed so far, the first such attempts being one by Warszyńska

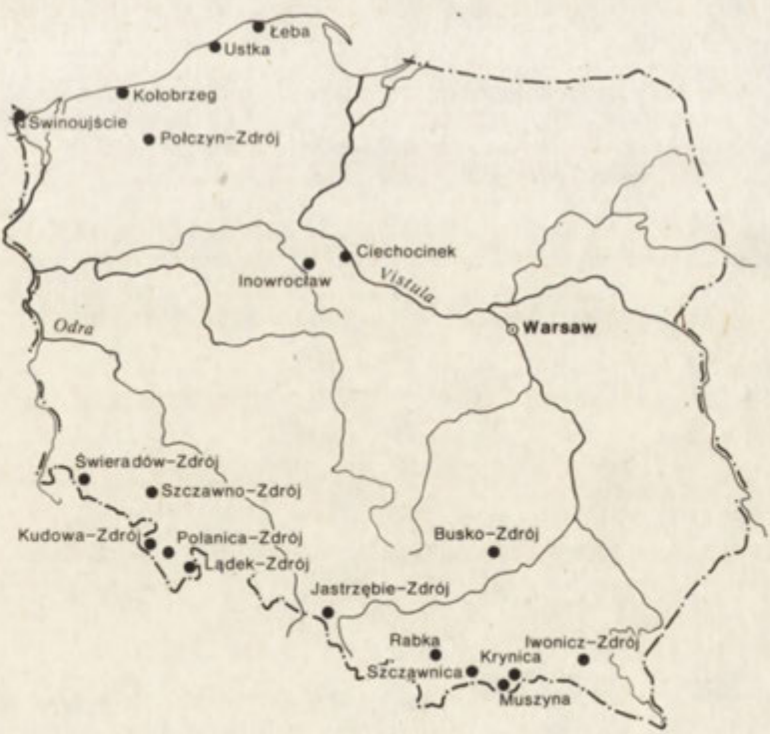


Fig. 1. Location of Polish health resorts included in the study

(1973) — from the point of view of tourism, and Błażejczyk (1980 a, b, 1983) — for bioclimatology.

Comprehensive bioclimatic evaluation of health resorts and recreation areas should take into account not only general climatic conditions, but also other components of geographical environment, which perceivably modify the local climate. They include in the first place, relief and land organization.

GENERAL MODEL FOR EVALUATION

A mathematical model of the exponential function has been taken as the basis for a general model of evaluation of bioclimatic conditions:

$$y = x^z$$

where: x describes the quantity characteristics of a given element of geographical environment, while z — its quality characteristics. The values of y were assumed to be alternating within the interval from 0 to 1. A 0 value means complete absence of favourable features in an environment, 1 — an ideal state with nothing to hamper climatotherapeutic treatment and recreation. To fit this assumption, the values of x must also remain within the range of variable from 0 to 1, whereas z may take any value from 0 to $+\infty$; $z<1$ will increase the values of x, while $z>1$ will diminish it. With $z = 0$ (best possible quality characteristics), y will equal 1, whatever the value of x.

The quantity and quality characteristics of individual elements of geographical environment were evaluated with the use of the model employed in physics to determine intensity of electric current, and based on the Ohm law. It is defined by Kostrowicki (1970) as the model of system efficiency (S):

$$S = \frac{P}{O},$$

where: potential (P) are environmental characteristics suitable for climatotherapy and recreation, and resistance (O) are such conditions which hamper or even make it impossible to stay outside the buildings, and which impair qualities of the environment. The model requires such choice of numerical values, so as, with the quantity characteristics — $0 \leqslant S \leqslant 1$. As for the quality characteristics, the conditions where $P/O > 1$ have been regarded as favourable. Thus to meet the requirements posed by the exponential function model, the calculations must assume $z = 1/S$.

In this way, the final model of bioclimatic usefulness of individual elements of geographical environment for climatotherapy and recreation will be:

$$y = \left(\frac{Px}{Ox}\right)^{\left(\frac{Oz}{Pz}\right)}$$

where: Px — potential of quantity characteristics, Ox — resistance of quantity characteristics, Pz — potential of quality characteristics and, Oz — resistance of quality characteristics.

SPECIFIC MODELS OF EVALUATION

Evaluation of climatic conditions proposed here employs a typology of weather developed by Błażejczyk (1979), from the point of view of man's recreation needs.

The quantity characteristics in the model were frequencies of occurrence of various classes of weather in each successive month. For potential (Px), the model assumed weathers favourable for climatotherapy and recreation (I_1, I_2, I_3, I_4), and for resistance (Ox) — the weathers definitely unsuitable for staying out of facilities.

Quality characteristics of climatic conditions were determined by parametres of weather day-to-day variability (Błażejczyk, 1980 a), namely: frequency of changes of low (i_m) and high (i_d) intensity, stability of low intensity (s_m) and high intensity changes (s_d), and probability of three successive days of the most favourable weathers (P_I) as well as unfavourable ones (P_{III}) for climatotherapy and recreation.

The complete model for evaluation of a climate's usefulness for climatotherapy and recreation (Kl) is as follows:

$$\mathrm{Kl} = \left(\frac{I_1+0.75I_2+0.5I_3+0.2I_4}{1+III}\right)^{\left[\left(\frac{1+i_d}{1+i_m}\right)\left(\frac{1+s_d}{1+s_m}\right)\left(\frac{1+P_{III}}{1+P_I}\right)\right]}$$

The values of Kl index obtained were grouped into intervals of different usefulness for recreation and climatotherapy:

Interval	Kl value	Evaluation of climatic conditions
A	$\leqslant 0.200$	unfavourable
B	$0.201-0.400$	little favourable
C	$0.401-0.600$	average
D	$0.601-0.800$	favourable
E	$\geqslant 0.801$	very favourable

Favourable and very favourable climatic conditions exist in Polish health resorts from May through October, while in wintertime weather conditions are little favourable for climatotherapy or recreation.

Obviously, also other characteristics of climatic conditions can be employed instead of the weather typology and related parameters as used in this paper. In that event, evaluation scale would need to be altered, too.

The basis for evaluation of terrain relief from the point of view of the needs of recreation and climatotherapy were maps of relative direct solar radiation in a given area on a month-to-month basis.

The quantity characteristic of the index of bioclimatic evaluation of relief was the weighted average value of relative direct radiation (expressed in fractions of maximum direct radiation) for the whole investigated area.

Treated as quality characteristic was the share of terrains displaying relative direct radiation lower than that observed on terrains where direct radiation exceeds the total radiation at horizontal surface in a given month.

The formula for bioclimatic evaluation of relief for climatotherapy and recreation (Rt) is as follows:

$$\mathrm{Rt} = 0.005 \sum_{i=1}^{20} (Nw_i \cdot s_i)^{\left(\frac{1+Nw_{<0}}{1+Nw_{>0}}\right)}$$

where: Nw_i — centre of the i-th interval of relative direct radiation, s_i — the share in the total area of terrains within the i-th interval of relative direct radiation, $Nw_{<0}$, $Nw_{>0}$ — the share of terrains on which relative direct radiation is lower or higher than the total at horizontal surface in a given month.

The present study employs the Stružka method (1956) for the calculation of relative direct radiation. In Struzka's tables, the values of relative direct radiation oscillate, in the different months, from 0% in December – at northerly exposed slopes, to 200% in June – on southerly exposed slopes (compared to 1/12th of the yearly value of direct radiation). Hence the 0.005 coefficient in the formula. For other methods of determining slope radiation, the coefficient and evaluation scale must be modified accordingly. The values of the Rt index obtained may be grouped into intervales corresponding to the different bioclimatic evaluations of relief from the point of view of recreation and climatotherapy:

Interval	Rt value	Relief evaluation
A	$\leqslant 0.250$	unfavourable
B	0.251–0.500	little favourable
C	0.501–0.750	average
D	0.751–0.850	favourable
E	$\geqslant 0.851$	very favourable

Usefulness of relief for climatotherapy and recreation is clearly dependent on the time of the year due to considerable variability of solar radiation throughout the year.

Highest relief evaluation indices (Rt) are found on areas of diversified relief, i.e. in the mountains, and the lowest on flat areas with monotonous relief.

Land organization was evaluated for climatotherapy and recreation with the use of land use maps and air pollution data.

As the potential of quality characteristics for the index of bioclimatic evaluation of land organization, the share in a given area of zones occupied by forest, parks or orchards was assumed (*Zw*), and as resistance – the proportion of tightly built-up urban areas (*Zm*).

The potential of quality characteristics was expressed in *Am* coefficient indicating the impact of sea aerosol in formation of bioclimatic conditions. The resistance of quality characteristics was air pollution coefficient *Op*.

The formula for bioclimatic evaluation of land development from the point of view of the needs of recreation and climatotherapy (Zt) is as follows:

$$\mathrm{Zt} = \left(\frac{Zw}{1+Zm}\right)^{\left(\frac{Op}{Am}\right)}$$

Here, air pollution coefficient is dependent on total yearly amount of dustfall and changes in dustfall level throughout the year. The liminal value assumed was the dustfall of 100 $\mathrm{mg \cdot km^{-2} \cdot year^{-1}}$. For localities with dustfall greater than that, the *Op* coefficient was calculated as follows:

$$Op = \frac{\text{dustfall}}{100} \text{x weight, depending on month}$$

In the different months, the weight factor is: I – 0.79, II – 0.82, III – 1.0, IV – 1.10, V – 1.18, VI – 1.19, VII – 1.17, VIII – 1.15, IX – 1.03, X – 1.07, XI – 0.89, XII – 0.74. If the calculated value of *Op* was less than 1, then *Op* was assumed to be *Op* = 1. For localities with dustfall lower than the 100 $\mathrm{mg \cdot km^{-2} \cdot year^{-1}}$, *Op* – in the period from May through October – was assumed to be equal to the weight factor ascribed to each month, and for the period from November through February, *Op* = 1.

The values of the *Op* coefficient can also be calculated on the basis of data showing concentration of toxic chemicals in the air. Unfortunately, no such data were available for this research.

The *Am* coefficient applies only for seaside localities and those where sea aerosol is spread in an artificial way. For all others, $Am = 1$. The coefficient has been calculated as follows:

$$Am = \frac{1}{1 - Sa}$$

where *Sa* is the share of zones with increased level of sea aerosol in the air within total area of locality.

The values of the Zt index can be grouped into intervals of different bioclimatic usefulness of land organization for climatotherapy and recreation.

Interval	Zt value	Evaluation of land organization
A	$\leqslant 0.200$	unfavourable
B	0.201 – 0.300	little favourable
C	0.301 – 0.400	average
D	0.401 – 0.500	favourable
E	$\geqslant 0.501$	very favourable

The values of the Zt coefficient for Polish health resorts display little variability throughout the year, yet considerable diversification in space. Many of the Polish health resorts have very favourable land organization, but there is also a significant group of ones with land organization unsuitable for climatic treatment and recreation.

MODEL FOR COMPREHENSIVE BIOCLIMATIC EVALUATION

It is a common knowledge that man is not affected by separated elements of his geographical environment, but by a combination of them all. Usefulness of geographical environment for recreation and climatotherapy is modified, in varying degrees, by land development and relief. Formula of comprehensive bioclimatic evaluation of a locality or area will be as follows:

$$\mathrm{Bk} = \frac{\mathrm{Kl} + \mathrm{Zt} + 0.5\,\mathrm{Rt}}{3}$$

The values of the Bk index have been grouped into intervals of different usefulness of bioclimatic conditions for climatotherapy and recreation:

Interval	Bk values	Bioclimatic conditions evaluation
A	$\leqslant 0.175$	unfavourable
B	0.176 – 0.317	little favourable
C	0.318 – 0.459	average
D	0.460 – 0.574	favourable
E	$\geqslant 0.575$	very favourable

Highest Bk values among selected Polish health resorts have been found to occur in the mountain resorts of Polanica, Świeradów and Krynica (Fig. 2). Lowest Bk are in health resorts located in central and northern Poland: Busko, Inowrocław and Kołobrzeg.

The different resorts have sometimes considerable differences in yearly variability of bioclimatic evaluation index; from unfavourable or little favourable conditions during the wintertime months, to favourable and very favourable in summer (Fig. 3). Generally, the most suitable time for climatotherapy and recreation in Poland is the period from April through October, and in the mountain resorts also during the winter months.

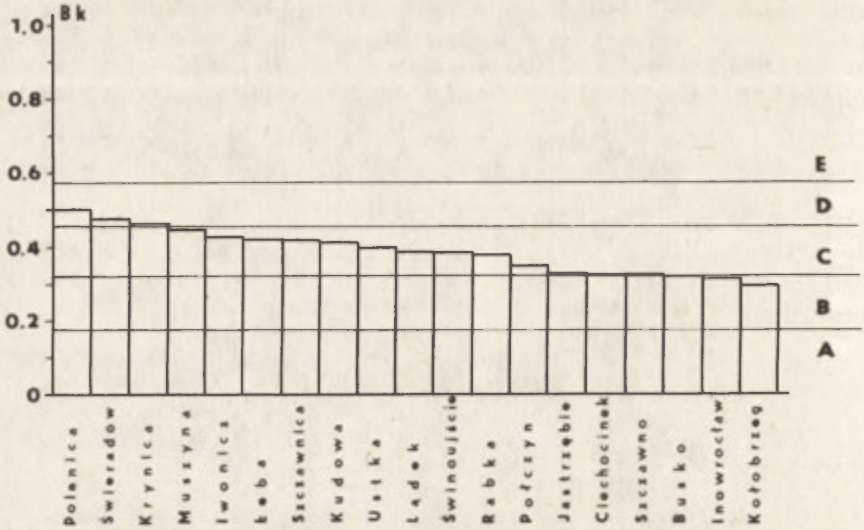


Fig. 2. Mean yearly values of the index of comprehensive bioclimatic evaluation (Bk) of selected health resorts in Poland: A — unfavourable, B — little favourable, C — average, D — favourable, E — very favourable

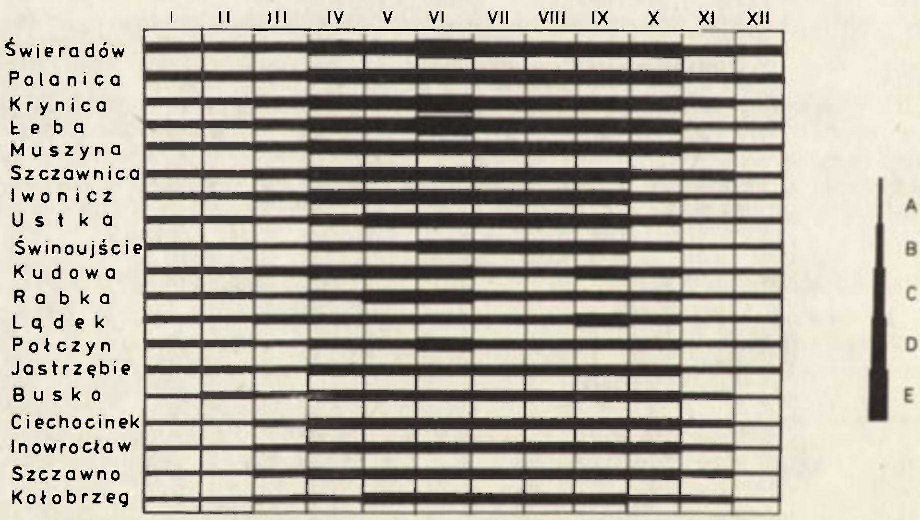


Fig. 3. Yearly values of indices of comprehensive bioclimatic evaluation Bk (interval values) in selected Polish health resorts: A — ⩽0.175, B — 0.176—0,317, C — 0.318—0.459, D — 0.460—0.574, E — ⩾0.575

MODEL FOR BIOCLIMATIC TYPOLOGY OF HEALTH RESORTS AND RECREATION AREAS

Health resorts displaying related characteristics have been grouped together with the use of numerical taxonomy. The characteristics were indices of evaluation of climate, relief and land organization for each of the four seasons of the year. They were the quantity characteristics. Their values were established separately for each of the resorts under study.

The degree of similarity between the resorts can be determined with the use of one of the taxonomical measures of distance, namely mean differences measure (Sokal and Sneath 1963).

$$d_{k,l} = \frac{\sum\limits_{k,l=1}^{19} \sum\limits_{i=1}^{p} |x_{ik} - x_{il}|}{p},$$

where $d_{k\,l}$ — taxonomical distance between resorts k and l (absolute value), x_{il} — value *of* i-th characteristic at resort k, x_{il} — value of i-th characteristic at resort l, p — total number of characteristics (here: 12).

The values of the different characteristics do not need to be normalized, as they are dimensionless indices placed within the interval 0—1.

The next stage in the process of establishing the bioclimatic types of health resorts was the choice of grouping procedure. Here, one of the best such procedures was used, the Sokal and Michener method of joint pairs, which enabled to draw a number of divisions of the resorts into groups.

The criterion of goodness applied to determine which of the division was best was mean distance inside each group:

$$d_j = \frac{1}{n_j} \sum_{k,l=1}^{n_j} d_{j,kl},$$

where: d_j — mean distance in group j, $d_{j\,kl}$ — distance between resort k and l in group j, n_j — number of resorts in group j, k,l — resorts classed into group j.

The results have been rendered graphically with the use of Czekanowski's diagram method. Instead of descriptive denomination of the degree of similarity between the resorts, the diagram here shows the intervals of the assumed measure of distance (Fig. 4).

The above grouping procedures allowed to divide the resorts into seven groups in which they are similar among one another for their assumed characteristics. Each of the groups (types) of resorts is characterized by specific features of geographical environment:

— type 1 (Szczawno, Połczyn, Inowrocław, Ciechocinek, Busko) — land organization unfavourable for climatotherapy and recreation;

— type 2 (Kudowa, Lądek, Szczawnica, Rabka) — at least average or favourable climatic conditions year long;

— type 3 (Jastrzębie) — land organization dangerous to man's health;

— type 4 (Polanica, Muszyna, Krynica, Iwonicz) — average or favourable climatic conditions, relief and land organization year long;

— type 5 (Łeba, Ustka, Świnoujście) — considerable seasonal oscillation of climatic conditions; relief little favourable for climatotherapy and recreation;

— type 6 (Świeradów) — very favourable conditions of land organization;

— type 7 (Kołobrzeg) — little favourable climatic conditions, relief and land organization.

CONCLUSIONS

Bioclimatic evaluation and typology of Polish health resorts, involving the above models, has allowed an objective assessment of the different localities and in consequence has offered a chance for more rational planning of therapeutical and recreational seasons. Objective evaluation of environmental conditions paves the way also for correct planning of development of the different localities.

At a different point in time, in the future, the same model and parameters of evaluation can be used to re-evaluate the areas covered by this study and thus find out about the trends of modern changes in geographical environment.

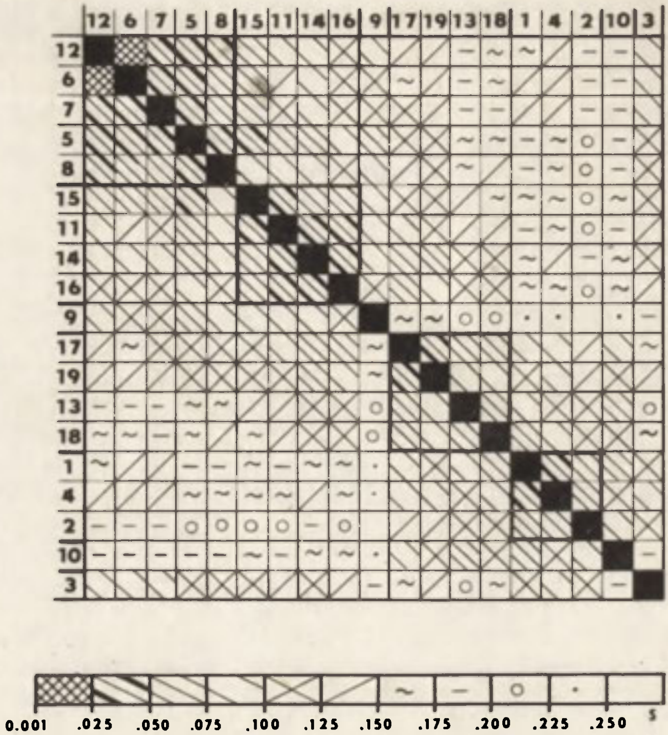


Fig. 4. Diagram of similarity indices for health resorts in Poland included in the study: 1 — Łeba, 2 — Ustka, 3 — Kołobrzeg, 4 — Świnoujście, 5 — Połczyn, 6 — Inowrocław, 7 — Ciechocinek, 8 — Busko, 9 — Jastrzębie, 10 — Świeradów, 11 — Kudowa, 12 — Szczawno, 13 — Polanica, 14 — Lądek, 15 — Rabka, 16 — Szczawnica, 17 — Muszyna, 18 — Krynica, 19 — Iwonicz

The above model for bioclimatic evaluation and typology of health resorts and recreational areas is applicable to any geographical conditions if their typical parameters — characterizing local environmental properties — are taken into the calculations.

REFERENCES

Błażejczyk K., 1979, Typologia pogody na potrzeby klimatoterapii (Weather typology for the needs of climatotherapy), *Dokumentacja Geograficzna* 2, 12—24.

Błażejczyk K., 1980a, Próba oceny klimatu uzdrowiska metodą modelową (An attempt at evaluation of climate of a health resort by the method of modelling), *Przegląd Geograficzny* 52, 1, 115—124.

Błażejczyk K., 1980b, Metoda modelowa oceny bioklimatu uzdrowisk (The model method of evaluation of health resorts' climate), *Dokumentacja Geograficzna* 3, 66—71.

Błażejczyk K., 1983, Bioklimatyczna ocena i typologia uzdrowisk Polski (Bioclimatic evaluation and typology of Polish health resorts), *Dokumentacja Geograficzna* 3.

Kostrowicki A. S., 1970, Z problematyki badawczej systemu człowiek-środowisko (Problematics in the investigation of the relation: man-environment), *Przegląd Geograficzny* 42, 1, 3—16.

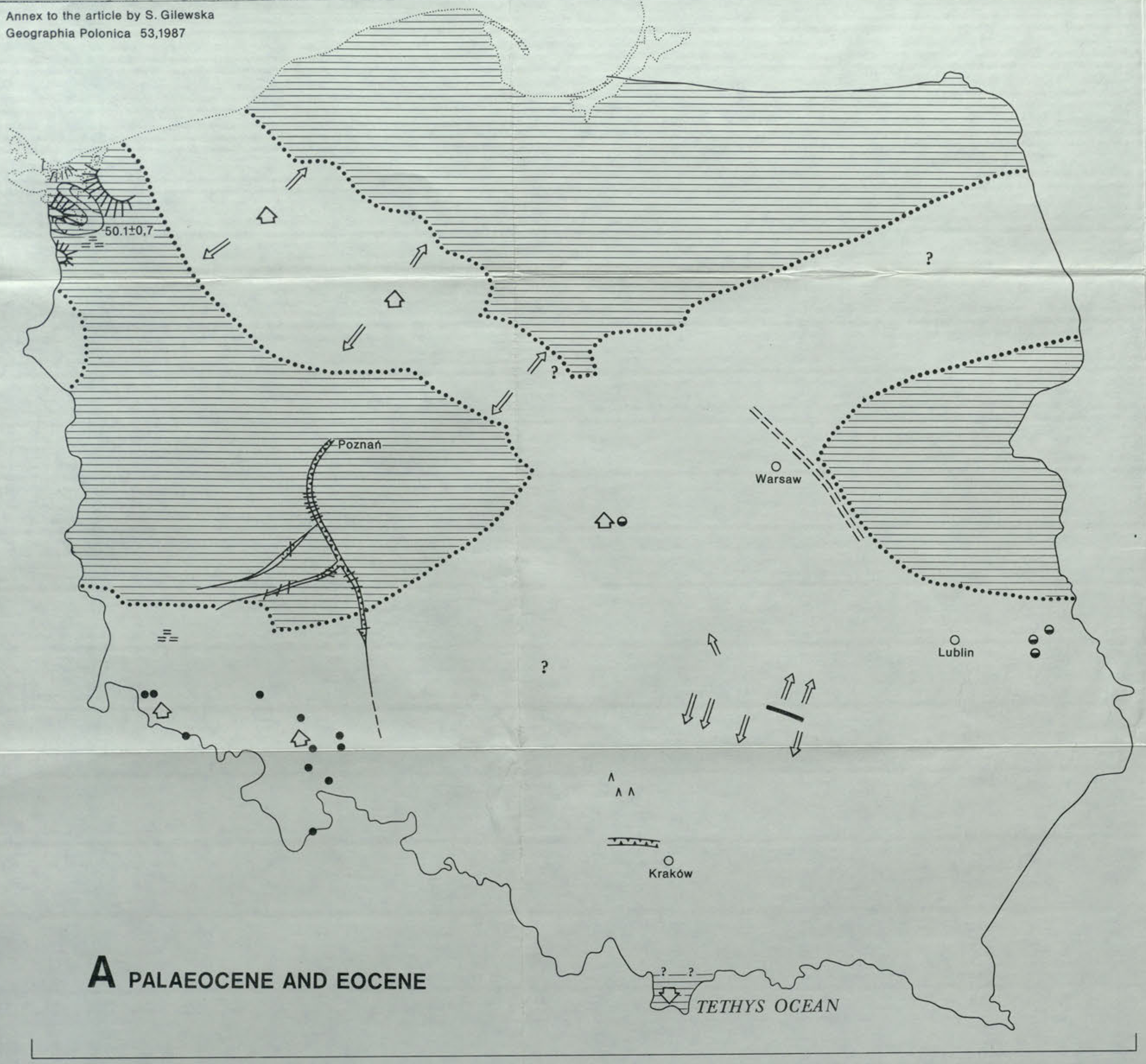
Sokal R. R., Sneath P. H. A., 1963, *Principles of numerical taxonomy*, W. H. Freeman and Co., San Francisco.

Stružka V., 1956, *Meteorologicke pristroje a mereni v prirode*, Statni Pedagog. Nakład., Praha.

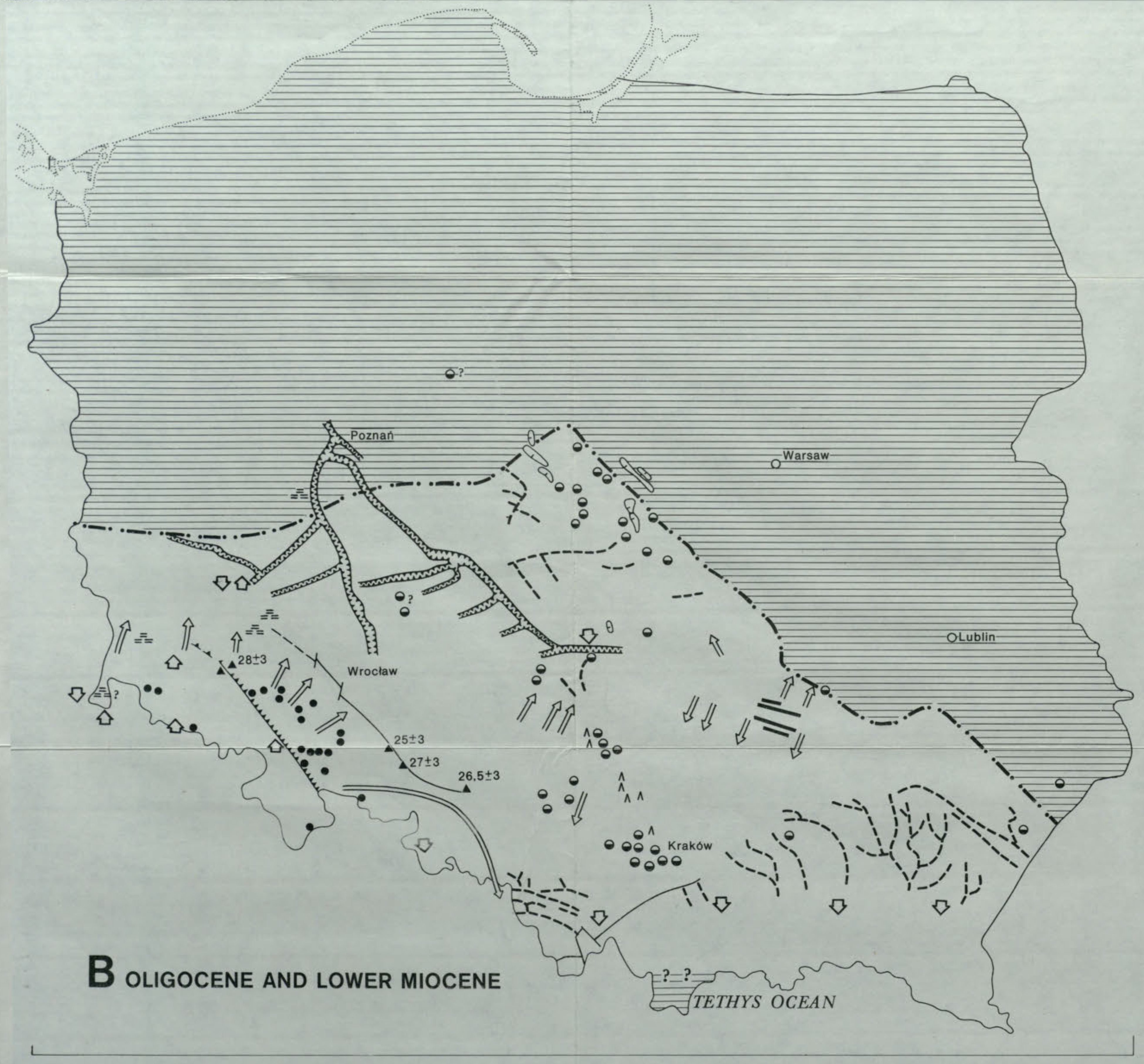
Warszyńska J., 1973, *Ocena zasobów środowiska naturalnego dla potrzeb turystyki* (Evaluation of natural environment resources for tourist needs), Zeszyty Naukowe UJ, Prace Geograficzne 36.

Annex to the article by S. Gilewska
Geographia Polonica 53,1987

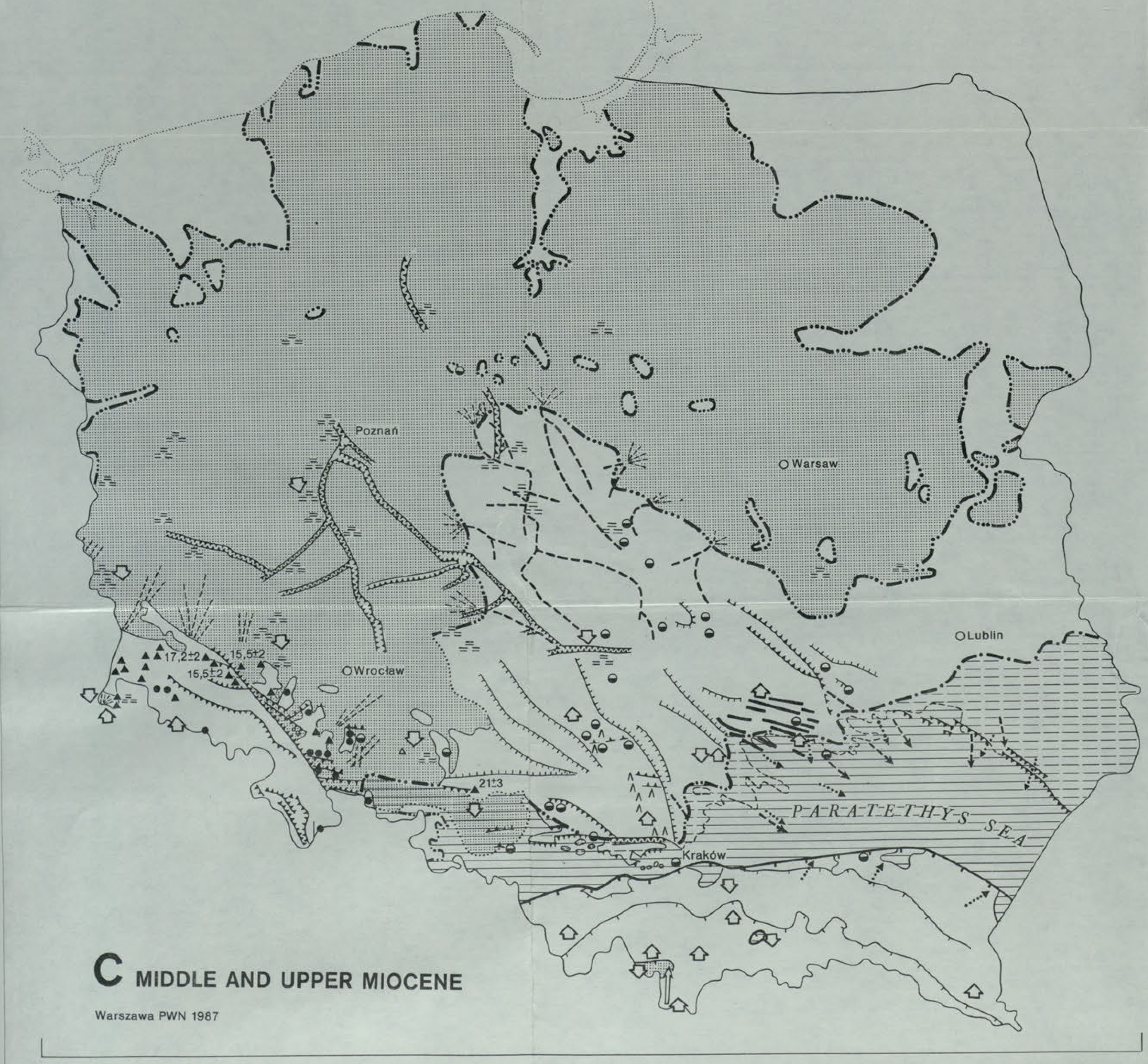
The geomorphic evolution of Poland's territory in Tertiary times



A PALAEOCENE AND EOCENE

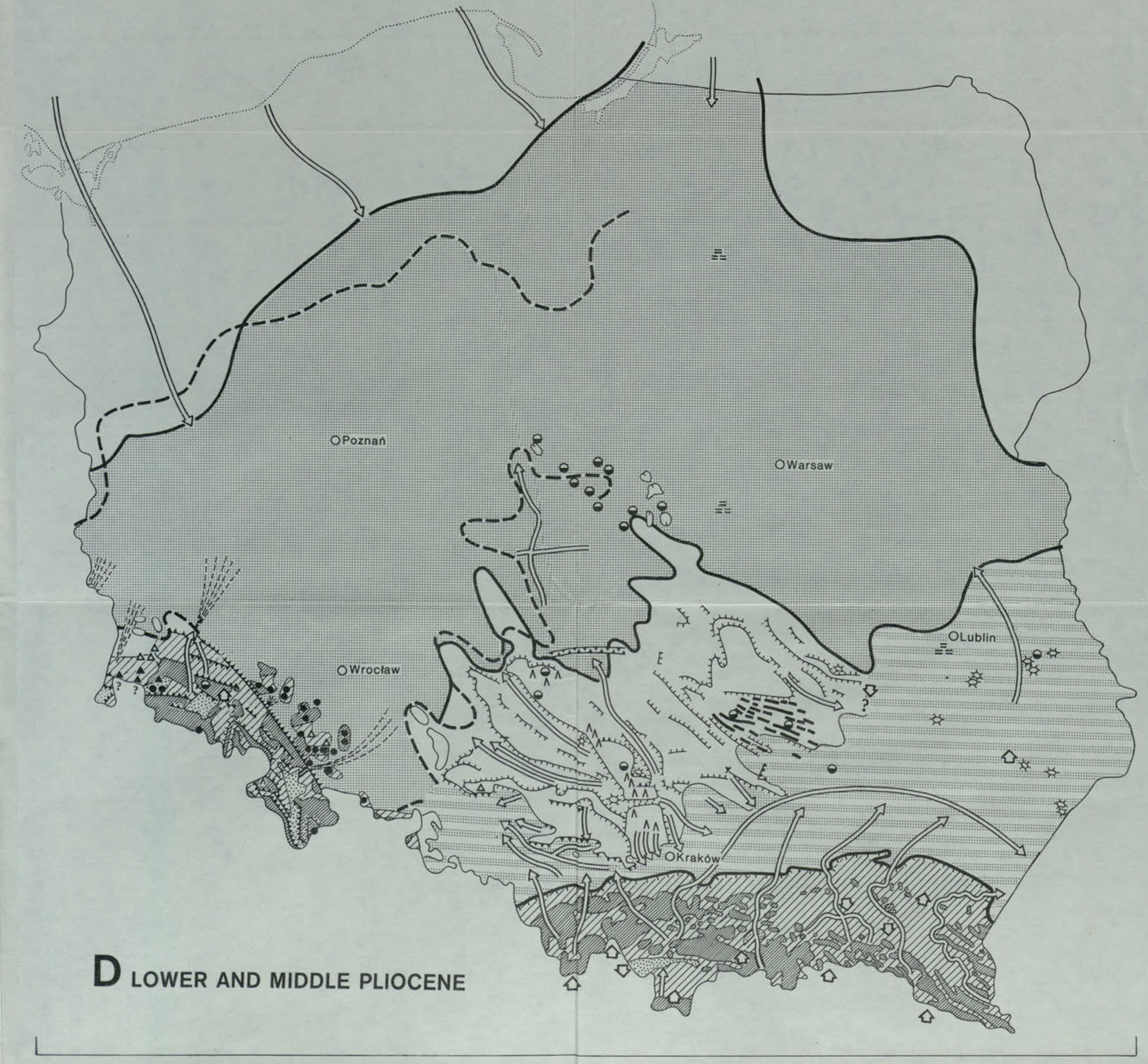


B OLIGOCENE AND LOWER MIOCENE



C MIDDLE AND UPPER MIOCENE

Warszawa PWN 1987



D LOWER AND MIDDLE PLIOCENE

-denuded land-mass
-sea
-depositional plain
-erosional-denudational plain
-extent of the late Eocene marine transgression
-extent of the Upper Oligocene marine transgression

extent of the Paratethys transgression:
a) during the Badenian
b) during the Sarmatian
-present-day extent of the Miocene marine deposits
-extent of the Sarmatian inland sedimentary basin
-present-day extent of the Miocene inland sediments

limit either of periodically flooded areas or of a residual lake during the Pliocene:
a) according to S.Z. Różycki (1972)
b) according to S. Dyjor (1987)
-tectonic scarps and major fault zones
-grabens
-post-orogenic collapse basins
-extensive closed depressions of tectonic and probable karstic origin
-scarp following the front of both flysch and Miocene nappes
-scarp following the front of flysch nappes
-active volcanoes
-inactive volcanoes
-areas showing uplift tendencies
-areas showing tendencies towards subsidence
-cuestas and other denudational escarpments
-outliers and uponliers
-Carpathian Foot-hills scarp of both tetonic and denudational origin
-mountains
-foot-hills
-intramontane basins
-hard-rock ridges
-inselbergs
-karst residuals (mogotes)
-karst depressions true to scale
-other karst depressions and caves
-elongated depressions of an unknown origin
-fluvial valleys
-important water gaps
-alluvial fans
-drainage direction

supplies of deltaic materials:
a) during the Upper Badenian
b) during the Lower Sarmatian
-bogs giving rise to brown coal seams (selected)
15,5±2 -absolute age (million years)

0 50 100 km

CONTENTS OF VOLUMES

GEOGRAPHIA POLONICA

Cena zł 400,—