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Climate reconstruction from stable-isotope composition of the Mazovian Interglacial (Holsteinian) lake sediments in eastern Poland

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ABSTRACT:

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Sediments of palaeolakes located in eastern Poland representing the Mazovian Interglacial (Holsteinian) and the initial part of the succeeding Middle-Polish glaciation (Saalian), are unique in Europe. These sediments are very thick (up to 55 m) and homogeneous, composed of lake marl and calcareous gyttja. They are thoroughly interpreted in terms of geological and palaeobiological studies (palynology, macrofossils, diatoms and malacofauna) and represent continuous deposition. Analysis of stable oxygen and carbon isotopes in these deposits and of the shells of the malacofauna enables interpretation of the changes in the palaeoclimate and sedimentary environment that occurred in lake basins during this part of the Pleistocene. Maximum $\delta^{18}O$ (-3,6‰) and minimum values of $\delta^{13}C$ (-6,4‰) correspond to the first part of the interglacial optimum, while minimum $\delta^{18}O$ values (-10,1‰) and maximum $\delta^{13}C$ values (10,0‰) correspond to the coldest period, directly preceding the following glaciation. Climatic changes are well documented by isotopic curves correlated with the results of pollen analysis. The isotopic curves indicate relatively cool climatic conditions at the climatic optimum of the Mazovian Interglacial. This may have been caused by increased atmospheric precipitation that led to deepening of the lakes, and/or by influx of ground waters enriched in light isotopes. In the upper part of the successions, corresponding to the initial stage of the following glaciation, the concentration of ^{18}O and ^{13}C increases, which was probably associated with the lake shallowing and with enrichment in heavy isotopes through evaporation under conditions of a cool steppe climate and/or with the redeposition of "warm" interglacial deposits from shore areas exposed as the result of lake shallowing. The isotopic curves clearly illustrate changes in the relative influence of maritime and continental air circulation during the Mazovian Interglacial. The studied lakes were oligo- or mesotrophic with dominant influence of continental air circulation before the interglacial climatic optimum. Maritime air circulation prevailed during the interglacial climatic optimum and the lakes became eutrophic. Cyclic climatic cooling during the post-interglacial period is recorded in oxygen isotopic curves trending towards their lower values. The carbon isotopic curves reach higher values, which is related to climatic cooling and shallowing of the lakes, caused by their infilling with deposits.

Key words: Paleoclimate, Mazovian Interglacial (Holstenian), Isotope.

INTRODUCTION

Climatic fluctuations that occurred during the Pleistocene, ranging from interglacials to glaciations, are indicated by variations in the oxygen stable-isotope ratio ($^{18}O/^{16}O$) in carbonates, resulting from temperature-dependent fractionating processes between carbonate and water.

The operation of these processes has been clearly demonstrated by numerous studies of borehole cores of deep-sea sediments (SHACKLETON & OPDYKE 1973, IMBRIE & *al.* 1973, PISIAS & *al.* 1984, MARTINSON & *al.* 1987, SHACKLETON & *al.* 1992). Deep-sea sediments allow the interpretation of a complete core that is usually without depositional gaps. However, pollen diagrams, which jointly with isotopic analysis, are used to determine Pleistocene warm periods on the continents, are not readily applicable to marine deposits. Isotopic studies of lake sediments do not always provide reliable results. This is partly caused by frequent depositional changes in lake basins, resulting from their limited size and depth, hydrogeological instability (irregular influx of ground and surface waters), or by increased evaporation (EICHER & SIEGENTHALER 1976, EICHER & *al.* 1981, FRITZ & POPLAWSKI 1974, EICHER & *al.* 1991, RÓŻAŃSKI & *al.* 1993).

Numerous depositional gaps in terrestrial deposits and in the supply of allochthonous calcite to lake basins seriously hinders the interpretation of isotopic curves (BEAULIEU & *al.* 1994, NITYCHORUK & *al.* 1999).

A more thorough interpretation of climatic changes was obtained from studies of vein calcite and cave travertines in Nevada (WINOGRAD & *al.* 1992), northern England and Canada (GASCOYNE 1992) and Norway (LAURITZEN 1995). The isotopic record obtained is not quite so clear as in deep-sea sediments, as a result of supply of genetically various waters. However, use of Th/U dating (WINOGRAD & *al.* 1992), makes the oxygen isotope curves to represent the same periods, expressed in particular isotope stages (EMILIANI 1955). However, correlation of pollen spectra and isotope stages is not entirely unambiguous. The same isotope stages in terrestrial and in marine sediments can be represented by slightly different isotope curves.

The numerous studies made so far on deep-sea cores have resulted in numerous isotope curves for the last few million years (SHACKLETON & OPDYKE 1976, SHACKLETON 1989). Studies of ice cores obtained from Antarctica and Greenland (JOHNSEN & *al.* 1972, LORIUS & *al.* 1985, DANSGAARD 1987, DANSGAARD & *al.* 1993, JOHNSEN & *al.* 1995, GROOTES & STUIVER 1997) contributed to more precise determination of climatic changes, especially for the last 130 ka. During this period there were two warming phases, namely the Holocene and the Eemian Interglacial, separated by the Vistulian Glaciation. They were all assigned to oxygen isotope stages 1 – 5e. It is much more difficult to correlate older interglacials, determined on the basis of pollen analysis, with a particular isotope stages. For example, the Mazovian Interglacial (Holsteinian), well known from studies of palaeolake sediments, is correlated with oxygen isotope stage 11c (SARNTHEIN & *al.* 1986, REILLE & BEAULIEU 1995, LINDNER & MARCINIAK 1998b), 9c (ZAGWIJN 1992) or 7e (LINKE & *al.* 1985) in deep-sea sediments.

Problems with stratigraphical correlation of deep-sea and terrestrial deposits does not diminish a significance in interpretation of oxygen and carbon stable isotopes for reconstruction of climate and basin sedimentation. STUIVER (1970), LANG (1970) and FRITZ & *al.* (1975) showed that isotope analysis can be successfully used for freshwater sediments. Combining palaeobotanical results with isotope data enables more detailed palaeoclimatic conclusions for the continents to be drawn. Such attempt have brought more satisfactory results recently, resulting in very good characterisation of sediments from the Late Glacial to the Holocene (FRITZ & POPLAWSKI 1974, EICHER & SIEGENTHALER 1976, EICHER 1979, EICHER & *al.* 1981, SIEGENTHALER & EICHER 1986, RÓŻAŃSKI 1987, MCKENZIE & EBERLI 1987, HOLLANDER & *al.* 1988, EICHER & *al.* 1991, FRENZEL 1991, LOTTER & *al.* 1992, RÓŻAŃSKI & *al.* 1992, KUC & *al.* 1993, VON GRAFENSTEIN & *al.* 1994, 1996, KUC & *al.* 1998, RÓŻAŃSKI & *al.* 1998).

The results of studies of the older Pleistocene deposits were also published for the Vistulian (BÖTTGER & *al.* 1994, JĘDRYSEK & *al.* 1995), Eemian (LITT & *al.* 1996, DRESCHER-SCHNEIDER & PAPESCH 1998), Mazovian (Holsteinian) (MÜLLER & HÖFLE 1994, NITYCHORUK & BIŃKA 1998, NITYCHORUK & *al.* 1995, 1998, 1999), Ferdynandovian (Cromerian) (KSZYSZKOWSKI & *al.* 1996) and recently also for the Augustovian (Bavelian?) (NITYCHORUK & *al. in press*).

GEOLOGICAL AND PALAEOBOTANICAL INTERPRETATION OF PALAEOLAKE SEDIMENTS OF THE MAZOVIAN INTERGLACIAL (HOLSTEINIAN)

The deposits of palaeolakes discovered in central-eastern Poland are well documented (Text-fig. 1). The best described localities, out of about thirty, are at Biała Podlaska (KRUPIŃSKI 1984-85, 1988a, b; KRUPIŃSKI & *al*. 1986, 1988; MARCINIAK & LINDNER 1995), at Komarno (LINDNER & al. 1988, KRUPIŃSKI & LINDNER 1991), Ossówka (LINDNER & al. 1990, NITYCHORUK 1994a), Hrud (LINDNER & *al*. 1991, NITYCHORUK 1994a), Borsuki (NITYCHORUK & BIŃKA 1994), Woskrzenice (BIŃKA 1994, BIŃKA & NITYCHORUK 1995), Kaliłów (BIŃKA 1994, BIŃKA & NITYCHORUK 1996), Małaszewicze (NITYCHORUK 1994a), Pawłów Nowy (NITYCHORUK 1994a, KRUPIŃSKI 1996b), Wilczyn (BIŃKA 1994, BIŃKA & *al*. 1996, 1997) Grabanów (KRUPIŃSKI 1995, LINDNER & WYRWICKI 1996), Lipnica (NITYCHORUK & BIŃKA 1995, BIŃKA & *al*. 1997), Zakrz

(ALBRYCHT & *al*. 1997), Lachówka Mała and Pokinianka (NITYCHORUK 1999), at Ortel Królewski and Rossosz (ALBRYCHT & *al*. 1995).

The numerous locations of lake sediments form a palaeolake district in southern Podlasie (NITYCHORUK 1994a; ALBRYCHT & *al*. 1997; PAVLOVSKAYA & NITYCHORUK *in press*). Pollen analysis (LINDNER & *al*. 1991; KRUPIŃSKI & NITYCHORUK 1991; NITYCHORUK 1994a; BIŃKA 1994; KRUPIŃSKI 1995; BIŃKA & NITYCHORUK 1995, 1996; BIŃKA & NITYCHORUK 1998, 1999; NITYCHORUK & BIŃKA 1994; NITYCHORUK & *al*. 1997), diatoms (LINDNER & *al*. 1990, 1991; MARCINIAK & LINDNER 1995; MARCINIAK 1998) and shells of snails from lake sediments – *Viviparus diluvianus* (KUNTH) and *Lithoglyphus pyramidatus* MILDF. (LINDNER & *al*. 1991; NITYCHORUK 1994a, b, 1995; ALBRYCHT & *al*. 1995; KRUPIŃSKI & SKOMPSKI 1995), enabled the palaeolake district to be assigned to the Mazovian Interglacial (Holsteinian). It probably continues southward (DYAKOWSKA 1952, 1956; BREM 1953; SOBOLEWSKA 1956; ŚRODOŃ 1969; KARASZEWSKI 1972; KARASZEWSKI & RÜHLE 1976; JANCZYK-KOPIKOWA 1981; WOJTANOWICZ 1983; HARASIMIUK & *al*. 1988; LINDNER & *al*. 1991; WINTER 1991) and northward, mantled with glacial deposits of the Middle-Polish Glaciations (Saalian) (GOŁĄBOWA 1957, BORÓWKO-DŁUŻAKOWA 1973, BAŁUK 1991, BAŁUK & MAMAKOWA 1991, BAŁUK & *al*. 1991, ALBRYCHT & *al*. 1995, MAMAKOWA 1998).

Palynological studies at Ossówka (KRUPIŃSKI 1995) indicated that the entire Mazovian Interglacial (Holsteinian) is recorded in these calcareous lake sediments. A dozen or so metres of the overlying calcareous deposits represent a cool early-glacial period that probably belong to the first Middle-Polish glaciations, i.e. the Liviecian in Poland (according to LINDNER 1988,

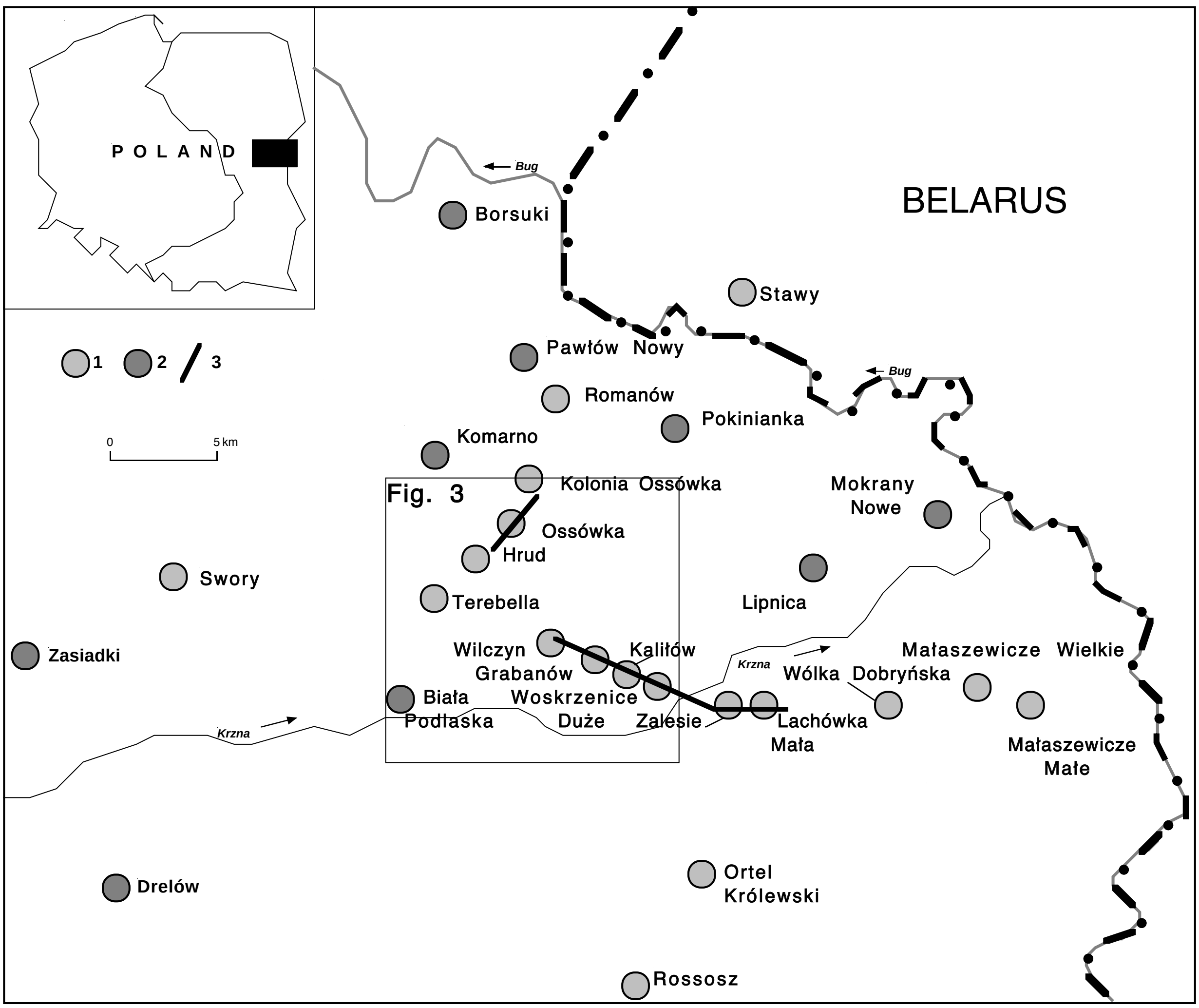

Fig. 1. Mazovian Interglacial localities in eastern Poland; 1 – calcareous deposits; 2 – bituminous shales and peats; 3 – geological cross-sections (cf. Text-figs 2, 3)

LINDNER & MARCINIAK 1998a, b), or to an early stage of the Saalian in Germany (according to MÜLLER & HÖFLE 1994). This cooling phase will be described herein as an early stage of the Middle-Polish glaciations.

The locations of the Mazovian Interglacial deposits in the Podlasie region can be divided into three groups based on the origin of the lake basins and the sedimentary environment (NITYCHORUK 1994a).

The first type consists of trough palaeolakes up to a dozen kilometres long filled with calcareous deposits – calcareous gyttja and lake marl (Text-figs 1-4), locally over 30 m thick. According to NITYCHORUK (1994a) these basins are arranged SW-NE (Ossówka – Hrud and Rossosz – Ortel Królewski) and W-E (Wilczyn – Grabanów – Kaliłów – Woskrzenice – Lachówka Małaszewice Małe), corresponding to tectonic structures in the Palaeozoic rocks, noted by ŻELICHOWSKI (1972, 1974) and POŻARYSKI (1974, 1986). Palynological studies suggest that sedimentation continued during the Interglacial in these lakes without major interruptions and terminated at the beginning of the following glaciation.

The second type consists of complex palaeolakes located on the morainic plateau (Biała Podlaska, Komarno, Mokrany Nowe, Lipnica and Pokinianka; Text-fig. 1). The sediments are: bituminous shales, peats, shaly peats and silts (NITYCHORUK 1994a) up to a dozen metres thick. These lakes are small, with diameters of a few hundred metres, which resulted in unstable depositional conditions and poor preservation of sediments.

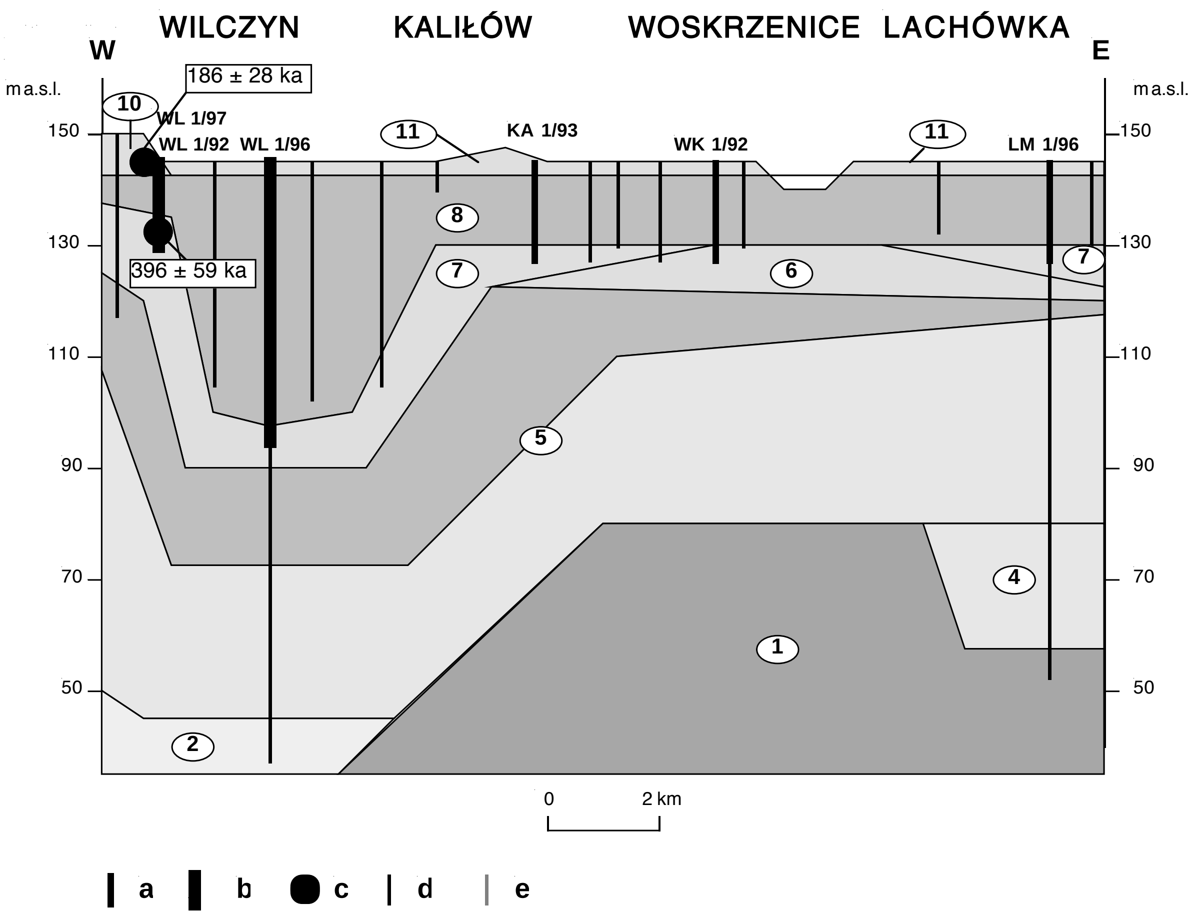

Fig. 2. Geological cross-sections of lake sediments of the Mazovian Interglacial at Wilczyn, Kaliłów, Woskrzenice Duże and Lachówka Mała (after NITYCHORUK *& al.* 1999, modified); for location of the cross-sections see Text-figs 1 and 4;
Cretaceous: 1 – chalk; Tertiary: 2 – glauconite sands; Quaternary: Nidanian Glaciation (Menapian): 3 – till; Ferdynandovian Interglacial (Cromerian): 4 – sands and muds with fragments of mollusc shells and plant debris; South-Polish Glaciation (Elsterian): 5 – sands, muds and sills; 6 – till; 7 – vario-grained sands; Mazovian Interglacial (Holsteinian) and the initial part of the Middle-Polish Glaciation: 8 – calcareous gyttja and lake marl; 9 – muds and silty muds; Middle-Polish Glaciation (Saalian): 10 – till; Middle-Polish Glaciation (Saalian) – Holocene: 11 – sands, muds, peats and peat alluvia; a – palynological analysis; b – oxygen and carbon isotope analysis; c – TL dating; d – other boreholes; e – boreholes outside the cross-section

The third type includes the palaeolakes of complex origin located in areas with tectonic structures in the Palaeozoic rocks. The best examples are at Pawłów Nowy and Romanów (Text-Fig. 1), where a 25 m thick complex of both calcareous and sapropelic deposits was identified (calcareous gyttja and bituminous shales). These lakes are relatively large (diameters of 1-2 km) and have diversified shorelines. Calcareous deposits, interlayered with bituminous shales, were found in the central part of the basin and in small bays.

The three above listed types of lakes occurring in the Mazovian Interglacial (NITYCHORUK 1994a) were supplemented by a fourth type, which was interpreted from boreholes in the vicinity of Janów Podlaski (LINDNER & MARCINIAK 1998a).

Palaeolake deposits are covered by a thin layer (2-3 m) of glacial, fluvioglacial, slope and valley-fill deposits of different ages. They overlie tills, fluvioglacial sands and limnoglacial silts and clays of the Sanian Glaciation (Elsterian) (WOJTANOWICZ 1983, LINDNER 1988, NITYCHORUK 1994a, MARKS & *al.* 1995).

The stratigraphy of deposits older than the Mazovian Interglacial is presented on geological cross sections (Text-figs 2, 3). They show that the lake series was preceded by two glacial episodes of the South-Polish glaciations (Elsterian). Although there are fluvial deposits of older interglacials, no corresponding palaeolake sediments have been found so far (LINDNER 1988, NITYCHORUK 1994a). The Quaternary deposits are underlain by Tertiary glauconitic sands and Cretaceous chalks (Text-figs 2, 3).

The palaeolake sediments directly underlie the interglacial series and the bottom parts of these lake sediments were dated by the thermoluminescence method from 431.7 to 132 ka BP (PRÓSZYŃSKI & *al.* 1989, NITYCHORUK 1994a, KRUPIŃSKI 1995, BIŃKA & *al.* 1997). The thermoluminescence dating of the deposits that cover the interglacial lake series (NITYCHORUK 1994a) are: basal sediments from the beginning of the Middle-Polish glaciations (Saalian) from 233 to 208 ka BP; glacial deposits and their residues from 224 – 154 ka BP.

The exact duration of the Mazovian Interglacial cannot be determined on the basis of the listed data. The data could indicate oxygen isotope stages 11, 9, or 7. Equally contradictory results come from ESR and U/Th dating (SARNTHEIN & *al.* 1986).

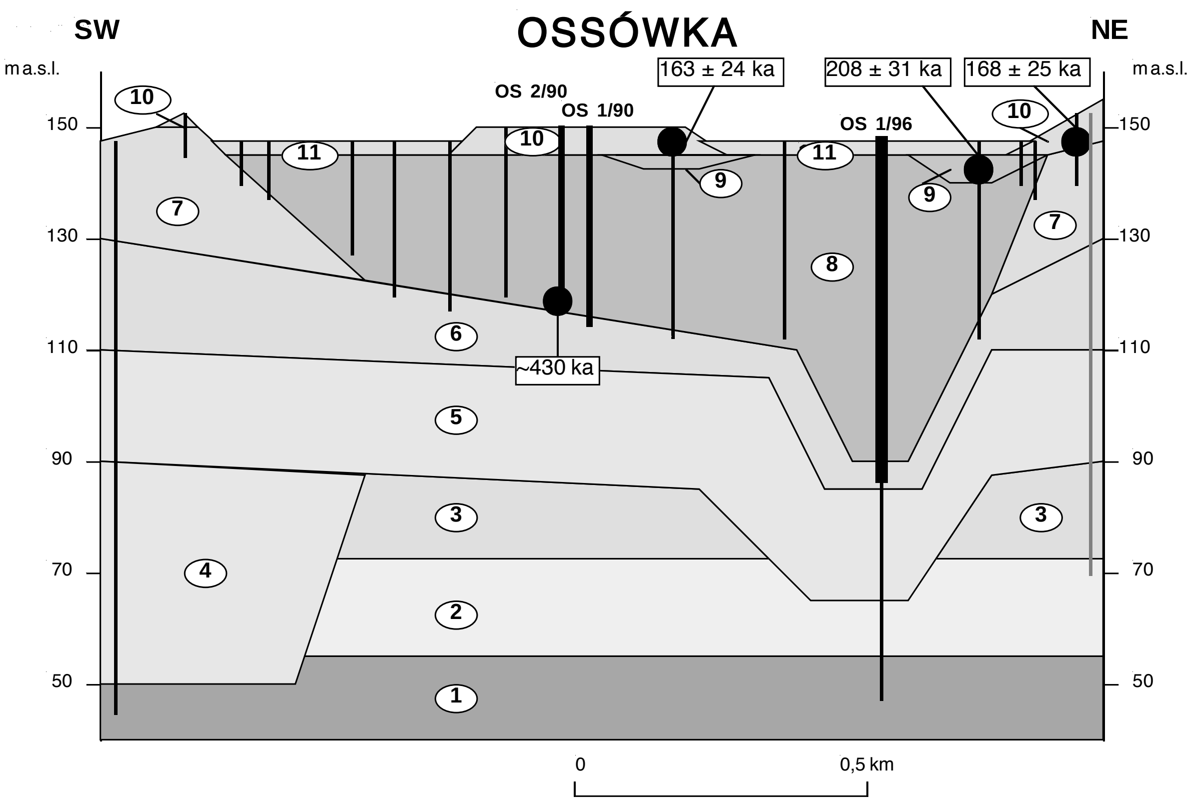

Fig. 3. Geological cross-sections of lake sediments of the Mazovian Interglacial at Ossówka (after NITYCHORUK & *al.* 1999, changed); for location of the cross-sections see Text-figs 1 and 4, for explanation see Text-fig. 2

Results of the latest boreholes

In 1993 – 1996 the author assisted in drilling boreholes in eastern Poland. These were, firstly, several hundred boreholes to a depth of 30 m drilled by the Geological Enterprise in Warsaw to document the lake marl (with the author determining the location of suitable sites). Eight deep, fully cored, cartographic boreholes were drilled for the Detailed Geological Map of Poland in 1:50 000 sheets Biała Podlaska and Rokitno (NITYCHORUK 1999). In the Biała Podlaska Sheet, boreholes were drilled into calcareous lake deposits at Ossówka and Wilczyn, as these had the greatest potential of producing a complete succession. These boreholes produced cores with unusually thick calcareous lake deposits (55 m at Ossówka and 35 m at Wilczyn Text-figs 2, 3, 5). The cores from Ossówka contained laminated deposits at depths 54.5 – 53.3 m, 51.5 – 27.3 m and 26.0 – 23.0 m (Text-fig. 8).

The two most representative successions of calcareous deposits are those of the Ossówka OS 1/96 and Wilczyn WL 1/96 boreholes (Text-figs 4, 5).

OS 1/96

0.0-0.2 m	sandy soil, grey
0.2-1.8 m	sandy alluvia, rusty-white, HCl^{+}
1.8-2.5 m	clay silt, steel-grey, HCl^{+}
2.5-4.2 m	calcareous gyttja, green-grey
4.2-10.3 m	calcareous gyttja, olive, with mollusc shells, at 5 m a 2 cm thick interlayer of fine-grained sand
10.3-10.5 m	calcareous gyttja, grey
10.5-11.1 m	calcareous gyttja, bright olive
11.1-13.0 m	calcareous gyttja, grey, compact
13.0-17.0 m	calcareous gyttja, olive with mollusc shells
17.0-20.5 m	calcareous gyttja grey, homogeneous, compact with numerous fish scales
20.5-21.2 m	calcareous gyttja olive with abundant mollusc shells
21.2-23.0 m	calcareous gyttja, dark grey
23.0-26.0 m	calcareous gyttja with annual lamination, olive
26.0-27.3 m	calcareous gyttja olive-grey
27.3-29.5 m	calcareous gyttja with annual lamination, grey-olive
29.5-33.5 m	calcareous gyttja with annual lamination, olive-grey
33.5-38.7 m	lake marl with annual lamination, grey, compact
38.7-40.5 m	calcareous gyttja with annual lamination, black
40.5-48.4 m	lake marl with annual lamination, dark grey and black (at 40-45 m well preserved hornbeam leaves)
48.4-51.5 m	gyttja with annual lamination, black, slightly inclined in the bottom
51.5-53.3 m	calcareous silt, black, slightly interrupted, without laminations
53.3-54.5 m	calcareous silt, grey-olive, very thinly laminated
54.5-56.5 m	clayey silt, grey, lamination 1 cm thick, interbedded with silty sand, bright grey
56.5-61.9 m	clay with boulders, very sandy, grey, with gravel, HCl^{+}

WL 1/96

0.0-0.4 m	sandy humus soil
0.4-2.0 m	clayey-sandy alluvium, white-rusty, HCl^{+}
2.0-6.2 m	calcareous gyttja, grey
6.2-6.6 m	calcareous gyttja, black with very abundant mollusc shells
6.6-10.0 m	calcareous gyttja, grey-black
10.9-19.3 m	calcareous gyttja, grey
19.3-32.0 m	lake marl, grey-brown with traces of lamination and streaking
32.0-34.2 m	lake marl, grey-brown with thin silty-sandy laminae
34.2-37.0 m	calcareous gyttja, grey-olive with distinct millimetre-thick lamination
37.0-37.3 m	sand with silt and sandy silt with organic parts
37.3-37.4 m	gravel of Scandinavian rocks

Palynological analyses proved that deposition of the laminated series took place during the Mazovian Interglacial (Holsteinian) and the following part of the Middle-Polish Glaciation (Saalian). Counting of annual laminae enabled the duration of the interglacial to be estimated at about 20 ka, which is a few thousand years longer than was indicated by MEYER (1974), MÜLLER (1974) and KRUPIŃSKI (1995), and is similar to the vein calcite age determined by the Th/U method (WINOGRAD & *al.* 1992). The post-interglacial part of the core represents the predominant boreal pine-birch forests and of the 5 cool episodes with subarctic climate, and the cold climatic episode just before the ice sheet advance. Pollen analysis of the 55 m core enabled the most complete interpretation of the Mazovian Interglacial (Holsteinian) in this part of Europe. This is similar to the interpretation of the previously studied 34.0 m OS 1/90 and 29.0 m OS 2/90 cores (Text-figs 3-5) from the same palaeolake basin at Ossówka. Pollen analysis (BIŃKA 1996) of the Wilczyn WL 1/96 core proves that the lower 30 m of the sediments represent the Mazovian Interglacial (incomplete, with a reduced basal sequence and a stratigraphical gap at the end of the climatic optimum), and the upper 5.0 m represent the cool early-glacial period. The calcareous sediments at Wilczyn, consisting of lake marls and calcareous gyttja, were deposited extremely quickly with an average rate of about 2 m per 1000 years (i.e. 2 mm a year).

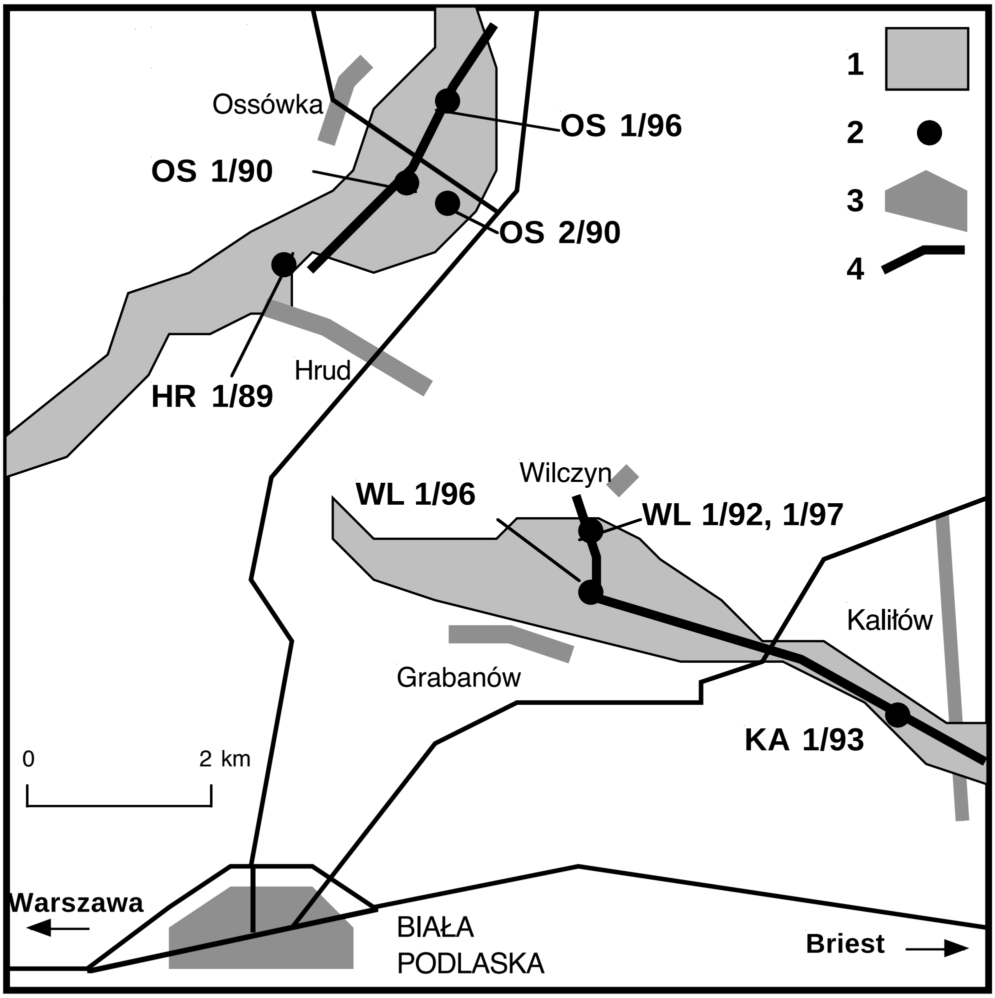

Fig. 4. Location of lake deposits of the Mazovian Interglacial at Ossówka and Wilczyn; 1 – calcareous deposits; 2 – boreholes; 3 – settlements; 4 – geological cross-sections (cf. Text-figs 2, 3)

STUDY METHODS

Oxygen and carbon isotope analyses and basic geochemical studies were initiated in 1996/1997 by the author in the research programme at the University of Göttingen, at the Institute of Geology with Professor J. SCHNEIDER, and at the Institute of Geochemistry with Professor J. HOEFS, and sponsored by the Alexander von Humboldt Foundation.

The Mazovian Interglacial sediments are characterised by a relatively long sedimentation period (about 50 000 years), as well as by homogeneity of interglacial and post-interglacial deposits consisting mainly of highly calcareous lake sediments. This suggests a complete depositional record, enabling valuable results to be obtained.

Isotope analyses were initially conducted for calcareous samples spaced every 0.5 m, and were later condensed, completed and repeated at the most interesting levels. In the case of laminated deposits each sample was collected from 5 annual layer set and isotope analyses were additionally carried out on single annual layers, with the dark and light laminae being analysed separately. Isotope analyses were also carried out on different species of molluscs. Cores of calcareous sediments from the palaeolakes at Ossówka and Wilczyn (Text-figs 4, 5) were studied and compared to each other in order to obtain more complete results.

At Ossówka, the analyses were carried out on samples from the entire 55 m long OS 1/96 core (147 analyses of sediments, 90 analyses of annual light and dark laminae, and 20 analyses of molluscs), from the 15.0 to 25.0 m interval of the OS 1/90 core (44 analyses of sediments), and also on molluscs from the HR 1/89 core – Text-fig. 4 (15 analyses). At Wilczyn the isotope analyses were carried out on 70 sediment samples from the entire 35 m WL 1/96 core, and on the interval 1.5-6.5 m interval of the WL 1/97 core (29 analyses of sediment and 64 analyses of molluscs). In addition, reinterpretation of isotope analyses for the 3.5 to 9.0 m interval (23 analyses of sediment) of the core Kaliłów KA 1/93 – Text-figs 4-5, was undertaken by Professor COLEMAN at the Postgraduate Research Institute for Sedimentology at Reading University (NITYCHORUK & *al*. 1995, COLEMAN & *al. in press*).

Altogether 502 oxygen isotope determinations and 502 carbon isotope determinations were made, together with about 100 repeat determinations on the same samples in order to check the reproducibility of the results obtained.

Oxygen and carbon isotopes analyses were conducted using the classical MCCREA method (1950). The CO_2 obtained by sample reaction with 99% orthophosphorous acid was analysed using a Finnigan MAT 251 gas spectrometer at the Institute of Geochemistry, University of Göttingen. The concentration of ^{13}C and ^{18}O isotopes in the analysed samples are presented as $^{18}O/^{16}O$ and $^{13}C/^{12}C$ isotope ratios versus the PDB standard. The latter is a Cretaceous belemnite from the Pee Dee formation and is the laboratory working standard used in low-temperature carbonate studies. The analytical error is ± 0,1‰ for $\delta^{13}C$ and ± 0,2‰ for $\delta^{18}O$.

Concentration of carbonates was determined in all the studied cores, concentration of carbon and sulphur were determined for the OS 1/90, OS 1/96 and WL 1/96 cores. For the OS 1/96 and WL 1/96 cores, without pollen diagrams, palynological interpretations were undertaken by

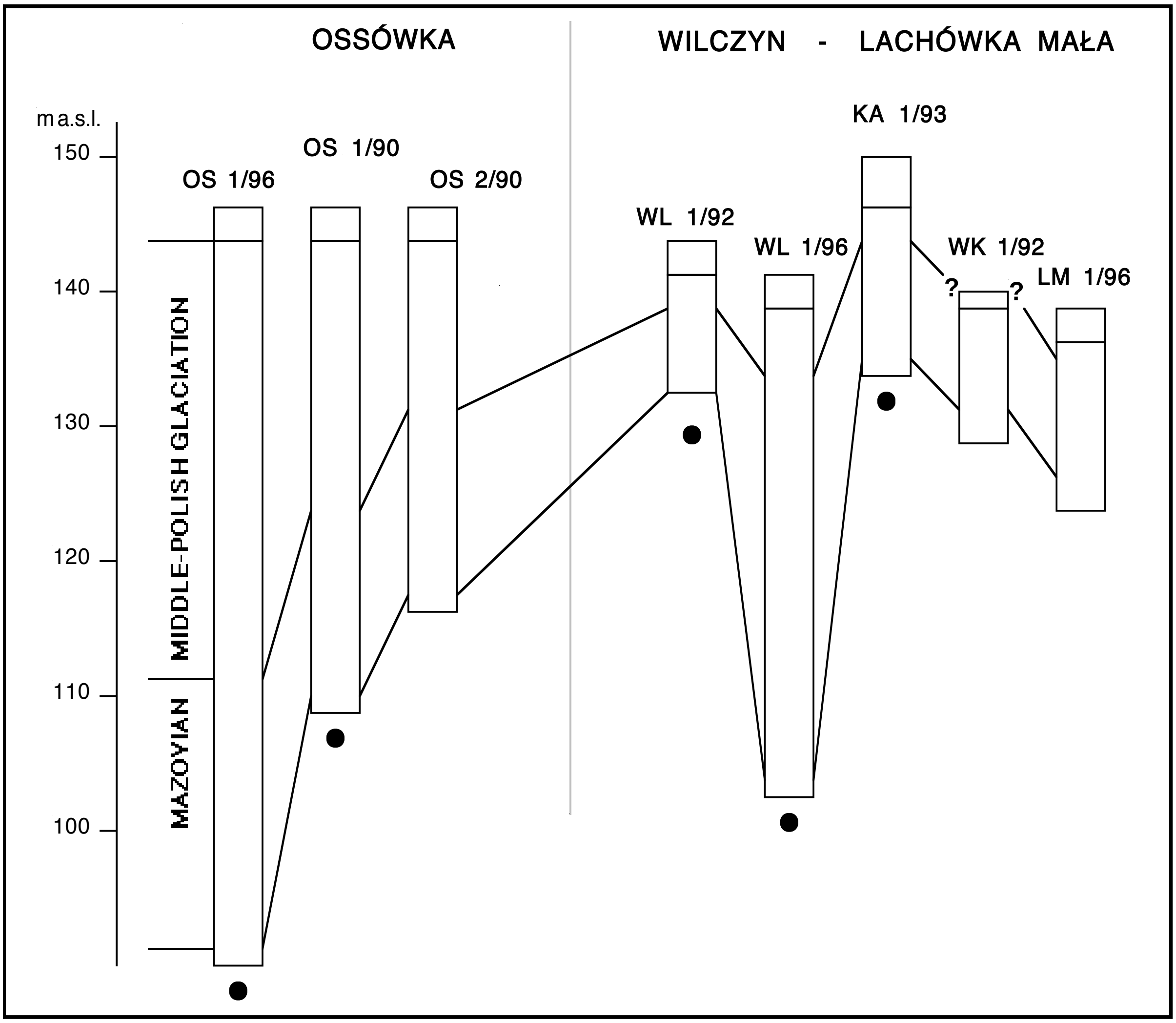

Fig. 5. Correlation of boreholes through lake deposits at Ossówka and Wilczyn; black points indicate boreholes with oxygen and carbon isotope analyses (after NITYCHORUK *& al.* 1999, modified); correlation based on palynological analyses

Dr K. BIŃKA. This enabled exact correlations of the isotope analyses with pollen diagrams from other cores (BIŃKA 1994, 1996, KRUPIŃSKI 1995, BIŃKA & NITYCHORUK 1996, BIŃKA & *al.* 1997).

CLIMATIC INTERPRETATION OF POLLEN RECORD IN THE MAZOVIAN INTERGLACIAL AND EARLY GLACIAL OF THE MIDDLE POLISH GLACIATION

In order to allow a better understanding and interpretation of the isotopic data, changes in vegetation communities during the Mazovian Interglacial and the initial part of the succeeding glaciation were reconstructed on the basis of pollen analysis.

The isotopic composition was determined for the carbonate sediments and for the molluscs of the two palaeolakes Ossówka and Wilczyn (Text-figs 4, 5), for which numerous palynological studies were carried out. There is a more complete interpretation based on pollen analysis for the basin at Ossówka, which comprises the following locations: Ossówka (LINDNER & *al.* 1990; KRUPIŃSKI 1995, 1996a) and Hrud (LINDNER & *al.* 1991). The basin at Wilczyn comprises more successions examined in terms of pollen analysis: Wilczyn (BIŃKA 1994; BIŃKA & *al.* 1996, 1997), Kaliłów (BIŃKA 1994, BIŃKA & NITYCHORUK 1995), Woskrzenice Małe, (BIŃKA *in* NITYCHORUK 1999, ALBRYCHT & *al.* 1997) and Grabanów (KRUPIŃSKI 1995). This list indicates that both palaeobasins were studied by two palynologists: Dr hab. K. M. KRUPIŃSKI and Dr K. BIŃKA who both concluded that the lakes at Ossówka and Wilczyn existed during the Mazovian Interglacial.

However, there are several differences between the two authors in the description of particular pollen periods and in the interpretation of the interglacial climatic optimum. The interpretations of pollen analyses for the lake sediments at Ossówka (boreholes OS 1/90 and OS 2/90) – according to K. M. KRUPIŃSKI (1995) and at Wilczyn (Borehole KA 1/93) – according to K. BIŃKA (BIŃKA 1994, BIŃKA & NITYCHORUK 1995), will be discussed separately.

Ossówka

Early interglacial

The beginning of the Mazovian Interglacial (Text-figs 6, 8) was dominated by boreal climate. There were long, humid and frosty winters and warm summers. High humidity of soil and air pertained. It is represented by pollen periods: **I** – protocratic part of the interglacial, with zones A, B (average July temperature ~13°C), and **II** – boreal and followed by boreal-temperate forest, with zone C and D (average July temperature ~17°C, – average January temperature ~ –4°C).

Interglacial optimum

The interglacial climatic optimum (pollen period **III,** zones E to H – Text-figs 6, 8), represents the predominance of mixed forests and a prevailing warm humid climate with humid winters and relatively dry summers in zone E (mean July temperature ~20°C). A slight cooling phase is noted in zone F, coinciding with the clear influence of continental climate with dry winters and distinctly lower humidity, associated with decreased precipitation. Zone G (average July temperature ~ 20°C, average January temperature ~ –1°C) was characterised by temperate, mild and humid climate with a long vegetation period, and also by mild and short winters. Precipitation was high, especially during the vegetation season. A similar climate continued in zone H, but with a slight decrease in average summer and winter temperatures.

Late interglacial

The late interglacial is represented by pollen period **IV** with boreal forests. The bipartite palynological zone J is characterised by a boreal climate with subarctic features (average July temperature ~ 16°C, average January temperature ~ –3°C).

Early glaciation

The climate changed into subarctic in pollen period **V**. This was the first of a series of cooling phases, marked by shrubby-tundra communities with very abundant juniper (Text-figs 6, 8). This period consists of three pollen zones: K, L and M (average July temperature ~ 13°C). In zone K drying was accompanied by a fall in temperature, especially during the longer winters. In zone L, aridity and the thermal contrast between seasons were slightly reduced, which resulted in a longer vegetation season. A distinct climatic amelioration took place in pollen zone M, when humidity and insolation increased. A short phase of pine forest during pollen period **VI**, zone N, when a return to boreal climate took place, was interrupted by another cooling phase – pollen period **VII**, associated with the disappearance of forests and with the development of shrubby-tundra fauna in a subarctic climate.

During pollen periods **VIII** (zone R) and **X** (zone V) temperate pine forests were the dominant element of

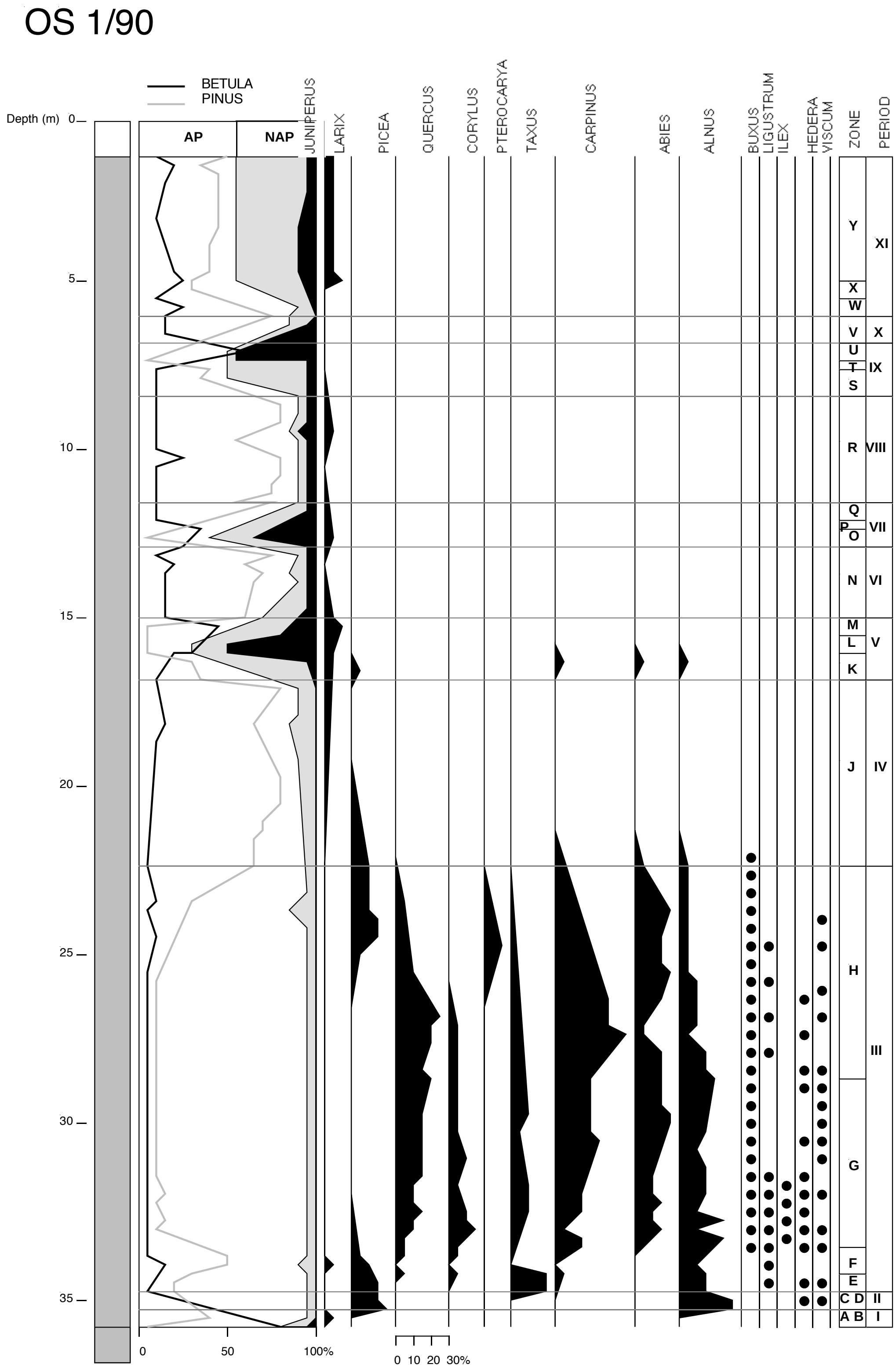

Fig. 6. Palynological diagram of the Mazovian Interglacial (Holsteinian) deposits at Ossówka OS 1/90 after KRUPIŃSKI (1995), greatly simplified; for lithological explanations see Text-fig. 8

the flora (average July temperature ~ 13°C). Considerable aridity during summers and winters, and a high temperature amplitude between these seasons clearly indicated the prevalence of continental climate. Cooling phases separated phases of forest development: pollen period **IX** (zones S, T, U) and pollen period **XI** (zones W, X and Y) (average July temperature ~ 11°C) were other pollen periods with subarctic climate. In zone S a shallowing of the lake occurred, accompanied by an abundance of rushes and shallows. Zone T was characterised by dry continental air masses and only in zone U did warming and an increase in humidity occur. Similar climatic changes can be observed in zones W, X and Y. The forest communities retreated, accompanied by lake shallowing and a decrease in $CaCO_3$. Aeolian processes developed in an arid and cool climate.

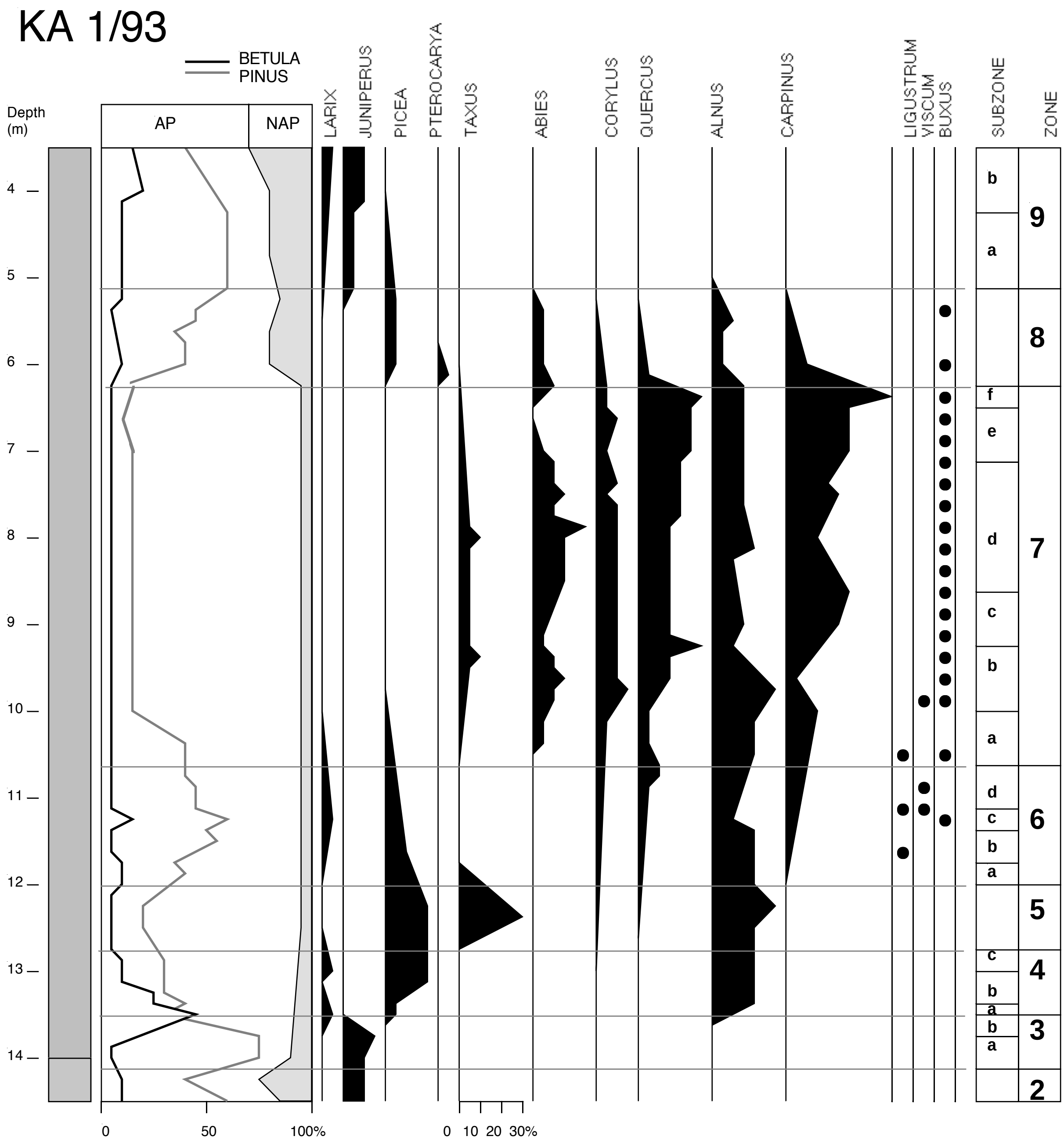

Fig. 7. Palynological diagram of the Mazovian Interglacial (Holsteinian) deposits at Kaliłów KA 1/93 after BIŃKA & NITYCHORUK (1996), greatly simplified; for lithological explanations see Text-fig. 8

Wilczyn

Late glacial

Deposition at Wilczyn starts in the pre-Interglacial period (Text-fig. 7), designated as pollen zone **1** (Cyperaceae-Gramineae). A cold climate predominated at that time, with a moss-grown tundra at higher elevations and with domination of sedge in lower, permanently wet locations (average July temperature ~ 7°C). Pollen zone **2** (*Juniperus-Hippophae*) (average July temperature ~ 10°C) was characterised by subarctic air masses with possible minor influences of maritime air masses. Diversified plant cover appeared at this time with species occurring in the lake water column and at a shore zone. The plants indicate the oligo- or mesotrophic character of the lake.

Early interglacial

The interglacial succession (pollen zone **3** – *Pinus-Betula*) started with the predominance of birch forest (average July temperature ~ 13°C) and was followed by pine forest with larch and juniper (average July temperature ~ 16°C). Larch and pine also occurred in subzones *a* and *c* of pollen zone **4** (*Picea-Alnus*), when climatic conditions with continental features prevailed. In subzone *b*, a culmination of ash preferring maritime climate occurred, which can be linked to the oceanisation of the climate caused by the reconstruction of the Gulf Stream in the northern Atlantic, and possibly also a small transgression of the Holsteinian sea (BIŃKA & *al.* 1997). In pollen zone **5** (*Taxus*), the abundance of yew indicates oceanic air masses, which favoured the development of trees with higher humidity requirements – spruce, alder and ash. At that time there was a high water level in the lake, but small annual fluctuations occurred. The next pollen zone **6** (*Pinus-Larix*) is divided into four subzones. In subzones *a* (*Picea-Alnus*) and *b* (*Carpinus-Quercus*) there was continental influence.

In subzone *b* there was a temporary slight increase in temperature, which was accompanied by drying out of the habitats of spruce and alder, reflected in their declining curves (Text-fig. 7) and in the rising curves of hornbeam and oak. In subzone *c* (*Pinus-Betula*) a distinct climatic cooling was recorded (*see* MÜLLER 1974). Subzone *d* (*Carpinus*) indicates a gradual climatic improvement.

Interglacial climatic optimum

Pollen zone **7** (*Carpinus-Abies*) indicates optimal climatic conditions in the Mazovian Interglacial. Fluctuations in the climate are shown by the mutual relationships between fir, hornbeam and oak. Fir is a stenothermal tree with distinctly higher climatic requirements than hornbeam and oak. Thus, in the zones with fir, the climate was warmer and more humid, with distinct maritime features. The zone shows cyclic humidity changes, which favoured fir during periods with a more humid climate, and oak and hornbeam during periods with lower precipitation.

Late interglacial and early glacial

In the pollen zones after the interglacial climatic optimum, i.e. **8** (*Pinus-Picea-Pterocarya*) and **9** (*Pinus-Juniperus*), the area covered by stenothermal elements decreased and communities characteristic of continental conditions expanded. The elimination of stenothermal species continued as the climate transformed into an even more continental one, and one characteristic of boreal areas. The plant communities constituted sparse forests or forest-steppes. Shallowing of the lake basin occurred, associated with an increased trophicity. Pollen zone **10** (*Artemisia*-Chenopodiaceae) contained steppe plant communities.

Summarising the above comments on plants during the Mazovian Interglacial and the resulting climatic changes, it is clear that periods of continental and maritime influence alternated at that time in eastern Poland. The continental climate in this part of Europe was characterised by warm and dry summers, and by frosty and long winters, which resulted in a long period of lake ice cover and a short vegetation season. The maritime climate had mild temperatures during the summer and relatively high temperatures during the winter, and was mild and humid. A long vegetation season and the lack of frequent frosts enabled the development of stenothermal plants. The maritime influence resulted in increased humidity, with higher precipitation in both summer and winter, which in turn led to higher ground lake water levels.

Eastern Poland, with its exceptionally well preserved palaeolake sediments from the Mazovian Interglacial, is additionally located at the boundary of the influence of continental and maritime climates, and is characterised by variation in isotope values. These factors are very important for the present study.

OXYGEN ISOTOPES IN LAKE SEDIMENTS

The oxygen isotope composition in authigenic carbonates generally directly results from the isotopic composition of the aqueous solution from which the carbonates precipitate. This composition is influenced by two major processes. The first is evaporation, intensification of which causes lake shallowing and enrichment in heavy isotopes (EPSTEIN & MAYEDA 1953, CRAIG 1961, MCKENZIE & EBERLI 1987). The second process is an influx of fresh water into the lake, which causes its deepening and relative depletion in the heavy isotope ^{18}O (MCKENZIE & EBERLI 1987, TALBOT 1990), which is manifested in minimum values on the isotopic curves.

Additional very important factors influencing the oxygen isotope ratios are: distance from the sea, altitude and atmospheric circulation (DANSGAARD 1964, RÓŻAŃSKI & *al.* 1993).

The shallowest lakes show seasonal variation of ^{18}O, with the highest values during the summer months. The largest concentration of ^{18}O is observed between August and October, which is associated with intensive evaporation (WACHNIEW & RÓŻAŃSKI 1998). Small seasonal isotopic changes do not occur in large lakes, which mostly react to long-term climatic changes (FRITZ & POPLAWSKI 1974). Lakes located in the north are covered by ice during the winter, which limits evaporation and in turn decreases the concentration of ^{18}O. The lowest ^{18}O contents are observed during spring thaws when there are large influxes of water enriched in the light isotope ^{16}O. A winter without ice cover increases evaporation especially in smaller lakes, which leads to a strong enrichment in the heavy isotope (FRITZ & POPLAWSKI 1974).

Stagnating lakes have higher ^{18}O content, while lakes with prominent surface and/or sub-surface influxes and outlets have the lowest one (FRITZ & POPLAWSKI 1974).

CARBON ISOTOPES IN LAKE SEDIMENTS

Carbon isotope composition in authigenic lake carbonates results from the isotopic composition of hydrocarbon (HCO^{3-}), which occurs as dissolved inorganic carbon (DIC) in lake waters (FRITZ & POPLAWSKI 1974).

The ^{13}C content in sediment is mainly influenced by exchange between CO_2 in the water and in the atmosphere, by the volume of incoming groundwaters and the influx of dissolved carbonates, by plankton photosynthesis and by CO_2 production during the decay of organic matter (CRAIG 1953, IRWIN & *al.* 1977, RÓŻAŃSKI & *al.* 1998).

In fact, balance in the water-atmosphere system is achieved very rarely. $\delta^{13}C$ of the precipitated carbonate could theoretically rise then to about 2 – 4 ‰ (FRITZ & POPLAWSKI 1974). Intensive mixing of water and ventilation of a shallow basin could contribute to achieving the equilibrium state.

Incoming groundwaters have a higher content of ^{12}C, because of their contact with soil CO_2, especially during summer, when root respiration is increased (KUC & *al.* 1998). Ground and river water discharges from areas with intensive vegetation also have higher contents of ^{12}C, which is associated with intensive oxidation of isotopically light organic matter Carbonates formed by methane oxidation are strongly depleted in ^{13}C compared to carbonates formed in isotopic equilibrium with CO_2 and produced from the oxidation of organic matter or from photosynthesis (JĘDRYSEK 1994, JĘDRYSEK & *al.* 1994, 1995).

An increase in ^{13}C concentration may result from intensive production of organic matter in lakes with a long retention period, or from the limitation of methanogenesis by the influx of oxygenated waters (JĘDRYSEK 1994, JĘDRYSEK & *al.* 1994, 1995).

Lake surface waters usually have the highest content of ^{13}C, while deeper waters have lower values, as they are affected by biological activity (BOTTINGA 1969, WACHNIEW & RÓŻAŃSKI 1998). Decay of organic matter and plant remains in aerobic conditions produces CO_2 with $\delta^{13}C$ values under –20 ‰ (FRITZ & POPLAWSKI 1974).

Changes in carbon isotopes in lakes are dependent on the extent of photosynthesis by plants, which increases with rising temperature. Aquatic plants prefer the lighter carbon isotope ^{12}C, which results in an enrichment of the heavy isotope ^{13}C during the precipitation of carbonates (PARK & EPSTEIN 1960, HOLLANDER & *al.* 1988). Enrichment of ^{13}C in the system can be further increased by intensive evaporation, by forcing the dissolved CO_2 out of the system and by increase of water hardness during deposition of the carbonates (STUIVER 1970). In fact, each lake is characterised by its own specific concentration of ^{13}C, depending on variations in the DIC content originating from the decay of organic matter (WACHNIEW & RÓŻAŃSKI 1998). This concentration results to a certain degree from the biological activity of water plants in the lake, which is influenced by climate. Therefore, studies of carbon isotopes together with oxygen isotopes will provide palaeoenvironmental information.

RESULTS AND DISCUSSION

Isotope concentration for palaeolake carbonate sediments in the Ossówka basin

Borehole OS 1/96 – $\delta^{18}O$

In the final phase of the South Polish Glaciation (Elsterian), the $\delta^{18}O$ values reached –5.0 and –6.0‰ (Text-fig. 8). These relatively high values refer to laminated clayey muds, 1 cm thick, alternating with muddy sands, between 54.5 and 56.5 m. The pollen spectrum is typically glacial. Therefore, the higher $\delta^{18}O$ values

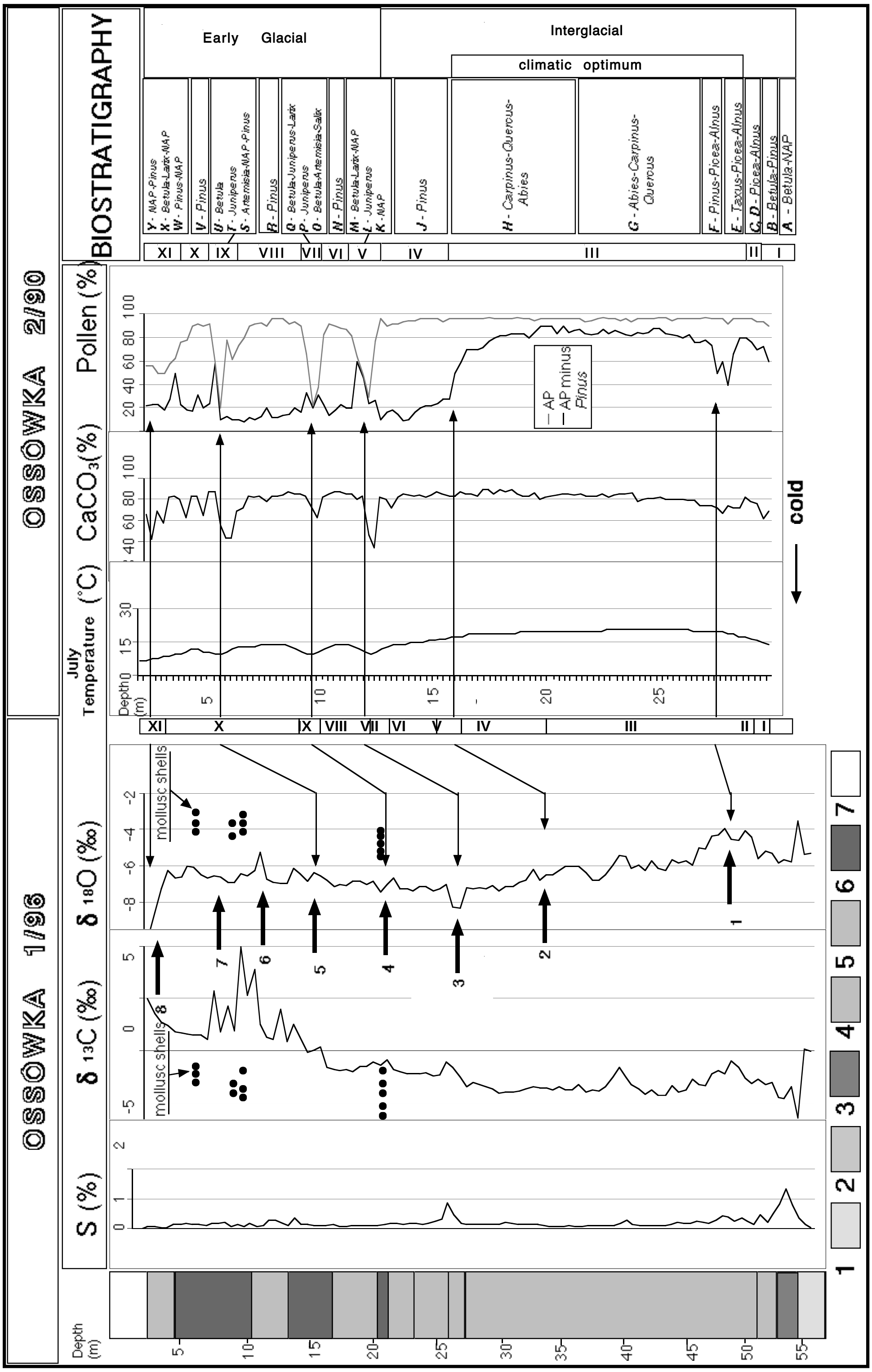

Fig. 8. Biostratigraphical correlation of the cores of boreholes OS 1/90 and OS 1/96 at Ossówka. Arrows indicate cooling phases; correlation based on palynological analyses and $CaCO_3$ content (after NITYCHORUK & *al.* (1999) modified); I, II, III... - palynological periods (after KRUPIŃSKI 1995); A, B, C... – palynological zones; AP MINUS *Pinus* – tree pollen pine excluded; AP – tree pollen; average temperature of the warmest month based on palynological analyses of KRUPIŃSKI (1995); 1 – late glacial clays and silts; 2 – sands and sandy muds with admixture of organic matter; 3 – very thin-laminated calcareous lake deposits; 4 – homogenus calcareous lake deposits; 5 – laminated calcareous lake deposits; 6 – concentrations of mollusc shells; 7 – covering deposits

resulted from the contribution of carbonates from a melting ice sheet. Lamination of the deposits indicates the presence of a water reservoir at that time.

The boundary between the glaciation and the interglacial is very clearly reflected in the oxygen isotope curve and is expressed by an increase in $\delta^{18}O$ values to –3.6‰. This change also relates to the presence of very thinly (0.5 mm thickness) layered carbonate muds between 54.5 and 53.3 m. This part of the succession is represented by pollen period **I** (zones A and B), with average temperatures of 13°C for the warmest month. Since this part of the succession was formed in the initial period of the interglacial, $\delta^{18}O$ reached the highest values, values that were not subsequently attained during the interglacial climatic optimum. This can be ascribed to the influx of allochthonous calcite into the basin, even though $\delta^{13}C$ values are rather low in the same part of the succession (Text-fig. 8). Increased precipitation from maritime air circulation can also have result in an increase in the heavy oxygen isotope content. This problem will be discussed more thoroughly in the chapter dealing with isotopic determination in non-glacial laminae.

A sharp deflection of the $\delta^{18}O$ curve toward higher values is followed by a decrease in $\delta^{18}O$ values to –6.2‰, accompanied by disappearance of lamination in the sediments.

Pollen period **II**, during which the development of boreal and boreal-temperate forest (in its younger part) occurred (Text-fig. 6), coincided with an increase in $\delta^{18}O$ values (Text-fig. 8) up to –3.9‰. Fluctuations of the isotopic curve reach 1.5‰, which could have been caused by the circulation of maritime air.

At the beginning of the interglacial climatic optimum (zone E, according to KRUPIŃSKI 1995) an increase in $\delta^{18}O$ values to about –4.0‰ took place. Temperate climate predominated, with cool winters and relatively dry summers. A slight cooling in zone F is reflected in changes of $\delta^{18}O$ values from –3.6‰ to –4.6‰ (Text-fig. 8) and coincided with the obvious influence of continental climate.

Pollen zones G and H were characterised by temperate humid climate with a long vegetation season and mild winters. Maritime air masses predominated, favouring the development of stenothermal plants (BIŃKA 1994, KRUPIŃSKI 1995). Nevertheless, the isotopic curve turns gradually towards lower values, averaging –6.0‰, and does not indicate warming. The most probable explanation of this paradox is to infer that a distinct deepening of the lake basin took place at the same time (*see* BIŃKA 1994), which led to a decreased concentration of heavy oxygen isotopes (EPSTEIN & MAYEDA 1953, CRAIG 1961, MCKENZIE & EBERLI 1987). In pollen zones G and H, which are rather homogeneous with respect to the development of the plant cover (Text-fig. 6), a feature that is characteristic of stable climatic conditions, the $\delta^{18}O$ values vary over about 1.5‰ (Text-fig. 8), oscillating between –5.3‰ and –6.8‰. The first distinct increase in the isotopic values (at 43.5 m) coincides with the first minimum of fir pollen, associated with a short-term increase in thickening of the laminae in the sediment. A second and much more distinct increase in isotopic values begins at 40.0 m and coincides with a second minimum of fir pollen. A climatic interpretation of this situation is presented in the chapter on changes in the plant cover at Wilczyn (BIŃKA 1994). As suggested there, fir tolerated warmer and more humid climate with a distinct maritime influence. Therefore, a retrieval of fir coincided with drainage around the lake, resulting from the increased influence of continentality. Initially it caused slight shallowing of the lake, accompanied by an increase in the concentration of the heavy isotope ^{18}O. The influence of continental climate resulted later precipitation enriched in lighter oxygen isotopes. This resulted in an increase, followed by a decrease, in $\delta^{18}O$ values. During the first $\delta^{18}O$ increase, a short-term climatic change, clearly visible in pollen spectrum (Text-figs 6, 20), coincides with a slight increase toward higher values of the oxygen isotopic curve, which sufficiently explains the discussed process. However, during the second $\delta^{18}O$ increase, corresponding to the second minimum pollen fir concentration, the crisis observed at Ossówka was too serious i.e. long-lasting (2m of sediments, which represent about 2000 – 3000 years) and distinct (increase in $\delta^{18}O$ values more than 1‰) to make such an interpretation satisfactory. In most palaeolakes in eastern Poland, as well as in Central Europe (MÜLLER 1974), this crisis caused a major break or termination of lacustrine sedimentation.

To explain the above situation, thin sections of laminated sediments from selected locations were prepared in the Institute of Geology at the University of Göttingen. In deposits at about 40.0 m, the content of angular quartz grains of volcanic origin was several times greater than in the remaining part of the succession (Plate 1). The presence of volcanic ash during the Mazovian Interglacial optimum was also found in the Kaliłów KA 1/93 core. After dissolution of the calcium carbonate and organic components, microscopic analysis of all of the quartz grains in the sample indicated that angular grains of volcanic origin constituted 20 – 30% of the total, with their lowest percentage in the lower parts of the succession.

It may therefore be suggested that in pollen period **III**, zone H, subzones 2-5 (according to KRUPIŃSKI 1995

– Text-fig. 6) and zone **7** (*Carpinus-Abies*), subzones e-f (according to BIŃKA & NITYCHORUK 1996 – Text-fig. 7), a climatic crisis occurred which could have been related to a catastrophic volcanic eruption. The scale of this possible catastrophe is indicated by the distance of thousands of kilometres between the region in question and the nearest volcanoes active at that time. The presence of tephra in the Placlaux (Holsteinian) lake sediments (BEAULIEU & REILLE 1995, REILLE & BEAULIEU 1995) of the Placlaux crater, may indicate south-central France as the possible source of the volcanic ash.

The presence of volcanic tephra has been detected in Holocene and pre- Holocene lake sediments (MERKT *& al.* 1993, KENNET & THUNELL 1977), and in ice cores (ZIELINSKI *& al.* 1996, 1997). The influence of volcanic eruptions on climatic changes has been studied for a long time (LAMB 1970, BRAY 1974, RAMPINO *& al.* 1979, 1988; KENNETT & THUNELL 1977; SCHMINCKE *& al.* 1991; RAMPINO & SELF 1993). In addition, a series of works is being published at present that summarises interdisciplinary studies of such volcanoes as: Agung (1963), El Chichon (1982) and Pinatubo (1991) and their influence on the climate (ANGELL 1996, GRAF *& al.* 1996, HANSEN *& al.* 1996, KONDRATYEV 1996, UCHINO 1996). The results of these investigations suggest that very large volcanic eruptions deliver ash and gas pollution up to the highest parts of the atmosphere. Volcanic pollution causes a temperature rise in the stratosphere (ANGELL 1996, LABITZKE & LOON 1996), while below and at the Earth's surface, a decrease in average temperature of about 0.2–0.5°C occurs (JONES & KELLY 1996, KONDRATYEV 1996). These processes occur because part of the solar radiation is absorbed by particles of volcanic origin in the upper parts of the atmosphere, resulting in cooling in the lower parts depleted in this radiation. Large volcanic eruptions known from historical time such as Tambora (1815) and Krakatau (1883) had a great influence on the weather during the succeeding years. The year 1816 passed into history as the year without summer (SCHÖNWIESE 1995, PFISTER 1999). The direct climatic effects of large volcanic eruptions are observed during the following 1 – 3 years (UCHINO 1996, PITARI *& al.* 1996, HANSEN *& al.* 1996).

Many palaeoclimatic analyses indicate, however, that the effects of volcanic eruptions are also observed over a longer time scale. Studies of borehole cores, with respect to the presence of volcanic ash in the last 20 million years revealed (KENNETT & THUNELL 1977) that a marked increase in volcanic activity occurred 3 – 5 million years ago, which could have supported the growth of ice bodies in the arctic areas. At the beginning of the Quaternary, about 2 million years ago, the volume of volcanic ash increased even more, coinciding with the onset of cold periods over the entire planet. Intensification of volcanic activity observed during the last thousand years coincided with a cool period called the Little Ice Age (LAMB 1970). A factor of great importance from a climatic point of view is the presence of sulphate groups – SO_4 in the atmosphere, which fall to the ground with precipitation.

In the OS 1/96 core, the content of sulphur in the sediment rises concomitantly with the first occurrence of volcanic ash (Text-fig. 8); this might have been caused either by precipitation containing sulphur, or by oxidation of FeS or H_2S as a result of lake shallowing (BAAS-BECKING & KAPLAN 1956).

Based on a decrease in $\delta^{18}O$ values of about 1‰ at that time, a temperature decrease of about 1-2°C might have taken place. This could have been associated with the volcanic crisis and/or with continentalisation, reflected in the decrease of fir in the pollen spectra. Climatic changes caused a fall in water level in the Ossówka lake, which survived the crisis without sedimentary hiatuses only because of its exceptional size. In smaller lakes such as Wilczyn, Biała Podlaska or Komarno, a major depositional break occurred.

In the final stage of the interglacial, i.e. pollen period **IV**, a drop in $\delta^{18}O$ values below –7.0‰ took place, which corresponds well with the climatic cooling shown by pollen analyses (Text-figs 6, 8).

During the early glaciation, pollen period **V**, the climate changed into a subarctic one. A distinct cooling phase, marked by shrubby-tundra communities is clearly reflected by a decrease in $\delta^{18}O$ values from –7.2 to –8.6‰ (Text-fig. 8). However, the pollen spectrum corresponding to this cooling phase is slightly delayed in relation to changes in the isotopic curve, because of the slower reaction of the plant cover to climatic changes. A repeated increase in $\delta^{18}O$ values to –7.0‰ was associated with the transition from the cool pollen period **V** to the warmer pine period **VI** (Text-figs 6, 8). The next cooling phase, pollen period **VII**, was on a smaller scale. It is expressed by a 0.3‰ decrease in $\delta^{18}O$ values. The third, fourth and fifth cooling phases, although distinct in the pollen spectrum (Text-Fg. 6), are not clearly reflected in the isotopic curves (Text-fig. 8). Continued global cooling, associated with the development of glaciation, resulted in an increase in $\delta^{18}O$ values to an about –6.5‰, and even as high as –5.6‰. Such a trend is unquestionably associated with lake shallowing caused by gradual infilling the lake basin with sediments. In a closed basin an increased concentration of heavy oxygen isotopes took place in a gradually decreasing water body (EPSTEIN & MAYEDA 1953, CRAIG 1961, MCKENZIE & EBERLI 1987), suggesting an interpretation opposite to cooling.

Borehole OS 1/96 – $\delta^{13}C$

In the final phase of the glaciation preceding the Mazovian Interglacial the carbon isotope curve reaches about 0‰ (Text-fig. 8). Higher $\delta^{13}C$ values are similar to those in the Cretaceous rocks, pieces of which are common in glacial deposits and come from glacitectonically eroded deposits about 15 km to the north. The only moderately low $\delta^{18}O$ values (about –5.0‰) do not reflect the severe climatic conditions prevailing at that time. This effect may have been caused by the presence of carbonates from the Cretaceous rocks, for which the $\delta^{13}C$ values range from 0.1 to 0.2‰.

The beginning of the interglacial coincides with a marked deflection of the $\delta^{13}C$ curve to –6.5‰, which is the lowest value noted at Ossówka (Text-fig. 8). Such a deflection could have been associated with the occurrence of water plants and their decay under aerobic conditions, resulting in the production of CO_2 with more negative $\delta^{13}C$ values (FRITZ & POPLAWSKI 1974). The end of sedimentation of very fine laminated deposits was accompanied by an increase in $\delta^{13}C$ values to about –3.3‰.

Gradual warming coincided with a continued rise in $\delta^{13}C$ values to –3.3‰, and to –2.5‰ in the upper part. The curve $\delta^{13}C$ is rugged which, as in the case of the $\delta^{18}O$ curve, may have been caused by the temporary influence of maritime air circulation. Climatic changes caused by marine impact, were expressed by higher plant production and by water level fluctuations in the lake. The beginning of the interglacial climatic optimum (according to KRUPIŃSKI 1995 – zone E – Text-figs 6, 8), was accompanied by a further increase in $\delta^{13}C$ values to about –2.0‰. A deflection of the $\delta^{13}C$ curve to –0.6‰ during a cooling phase in pollen zone F reflects the influence of continental air, which is documented by changes in the plant cover (KRUPIŃSKI 1995). The higher $\delta^{13}C$ values could have been caused by lake shallowing and its oxidation (JĘDRYSEK 1994, JĘDRYSEK & *al.* 1994, 1995).

Pollen zones G and H indicate a temperate and humid climate with a long vegetation season and mild winters. This was the period in the Mazovian Interglacial with predominantly optimal climatic conditions. The $\delta^{13}C$ curve shows a gradual negative trend to –4.0‰ associated with the deepening of the lake (Text-fig. 8).

In pollen zones G and H, with homogeneous development of the plant cover reflecting stable climatic conditions, the $\delta^{13}C$ values change from –3.0 to –4.0‰. A sharp positive deflection in the $\delta^{13}C$ curve begins at 40.0 m and coincides with the second minimum of fir pollen (Text-figs 6, 8). The $\delta^{13}C$ values reach a maximum of –1.5‰, which resembles the values that obtained during the earlier cooling phase and lake shallowing in pollen zone F. As already described in the discussion of the $\delta^{18}O$ curve, an increase in the volcanic ash content of the sediment occurs at this level in the OS 1/96 core. Volcanic eruptions accompanied by continental climate resulted in lake shallowing, which is confirmed by the increase of ^{13}C content. In the final part of the interglacial the $\delta^{13}C$ values (Text-fig. 8) range from –3.0 to –4.0‰, showing a similar range to that during the interglacial optimum. The isotope curve is rather smooth, reflecting stable climatic conditions and a return to the hydrological conditions that preceded the crisis.

During the early glaciation (pollen period **V**) the climate changed into a subarctic one. Distinct cooling, marked by shrubby-tundra communities, was accompanied by continentalisation and lake shallowing, and is clearly reflected by the increase in $\delta^{13}C$ values from –3.3 to –0.9‰ (Text-fig. 8). A repeated drop in $\delta^{13}C$ values to –2.0‰ is associated with the transition from pollen period **V** to the warmer pine period **VI**. The following cooling phase (pollen period **VII**) is recorded by a rise in $\delta^{13}C$ values from –2.0 to 0.0‰.

Lake shallowings and the prevalence of stable hydrological conditions are well reflected by the isotope curve of the OS 1/96 core below 17 m (Text-fig. 8). Above this depth the $\delta^{13}C$ curve shows extreme values from –2.5‰ to 10.0(!)‰, which is the highest one. There is also a change in the type of the sediment into that of a shallow lake, with grey, homogeneous, compact gyttja with numerous fish scales being replaced upwards at 17 m by olive gyttja with mollusc shells (Text-fig. 8).

The increasing cooling associated with glaciation was accompanied by cyclical drying in a continental climate with steppe features, when the lake basin was filled with sediments. In a closed shallow basin the concentration of heavy carbon isotopes increased due to a gradually decreasing volume of water.

Isotopic determination of mollusc shells

Isotope concentration in mollusc shells reflect the content of $\delta^{18}O$ in a sediment. Variation of $\delta^{18}O$ in individual species (FRITZ & POPLAWSKI 1974) are caused by different environmental preferences e.g. deep or shallow water. Palaeotemperatures based on the results of isotope studies of mollusc shells cannot therefore be easily determined. These isotope ratios seems to be more useful in palaeoenvironmental or palaeohydrologic interpretations (FRITZ & POPLAWSKI 1974).

The content of ^{13}C in mollusc shells is controlled by the Dissolved Inorganic Carbon (DIC) (FRITZ & POPLAWSKI 1974).

In the OS 1/96 core, the isotopic concentrations were determined for mollusc shells at 21.3 m, 10.0 m, 8.9 m, and 6.4–6.5 m (Text-fig. 8). There are differences between the values for the sediments and for the mollusc shells contained in them and also differences between individual mollusc species. The $\delta^{18}O$ record for the sediments is lower than for the shells, while the $\delta^{13}C$ record is higher for the sediments and lower for the shells. A similar relationship is observed in the WL 1/97 core (Text-fig. 16), where the difference in $\delta^{18}O$ values is about 2‰, compared with about 7‰ in the case of $\delta^{13}C$ values. Moreover, in both of the above-mentioned cores and in the HR 1/89 core, the records for individual mollusc species from the same zone differ from one another (Text-fig. 9). As indicated by other studies (FRITZ & POPLAWSKI 1974), the environments of deep and shallow water species exert a great influence on the isotope record. This is illustrated by isotope values measured at different water depths in Lake Gościąż at different seasons of the year (WACHNIEW & RÓŻAŃSKI 1998). The thermal stratification has a major influence on isotopes during summer and autumn when their concentration is higher in the epilimnion and lower in the hypolimnion (WACHNIEW & RÓŻAŃSKI 1998). Therefore, shells of littoral molluscs will reflect a higher isotope record (typical of warm water) than the fauna which lives in a deeper zone. Large differences between the isotopic records measured for the same mollusc species collect-

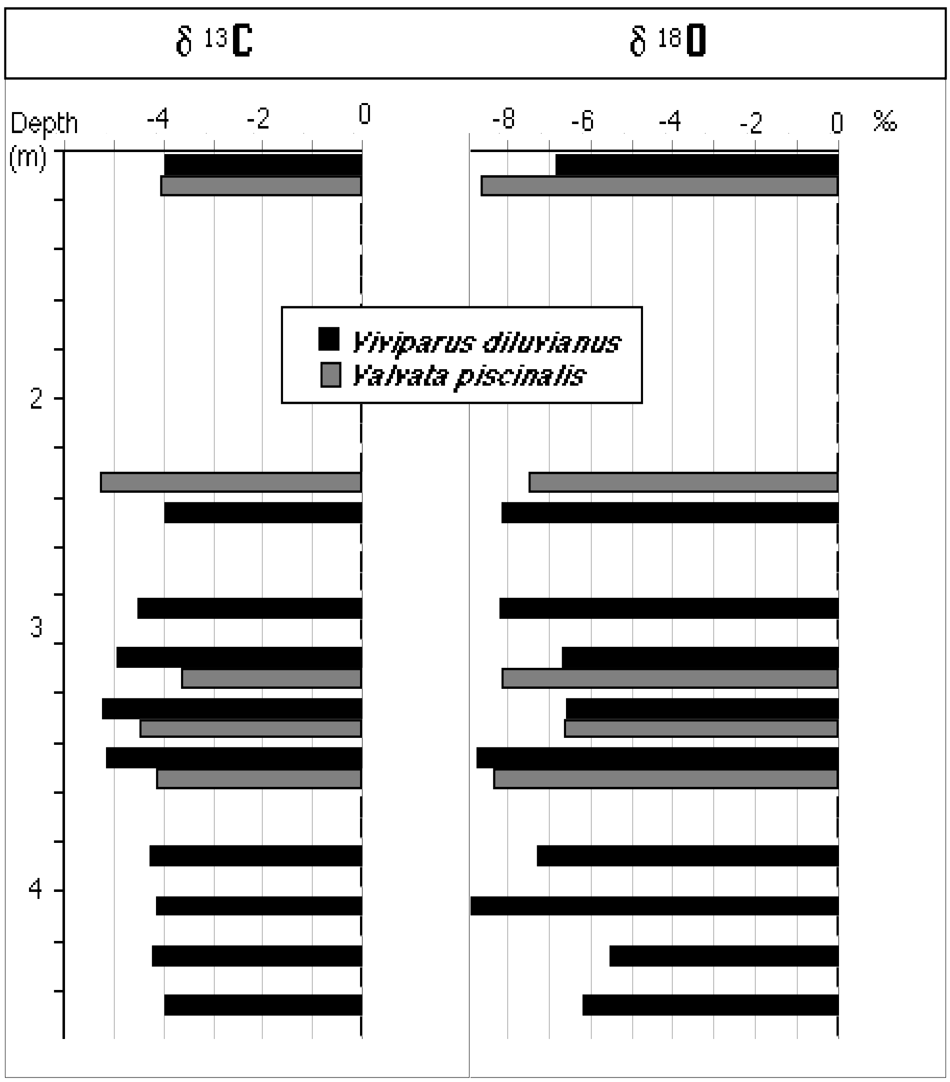

Fig. 9. Oxygen and carbon isotopes for shells of *Valvata piscinalis* and *Viviparus diluvianus* in the Hrud HR 1/89 borehole

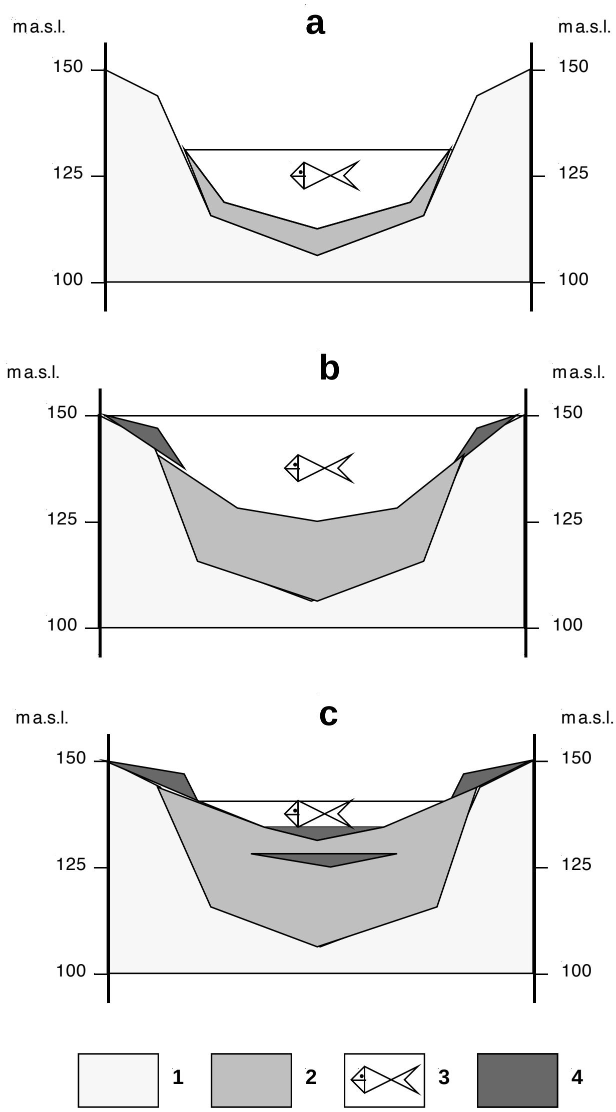

Fig. 10. Calcareous deposition during the Mazovian Interglacial (Holsteinian) and in the begining of the Middle-Polish Glaciation (after NITYCHORUK *& al.* 1999); a – beginning of the Mazovian Interglacial; b – interglacial climatic optimum; c – early Middle-Polish Glaciation; 1 – non-lake deposits; 2 – calcareous deposits; 3 – water; 4 – redeposition

ed from the same depositional horizon, may result from redeposition (*see* NITYCHORUK & *al.* 1999, NITYCHORUK & *al. in press*). The mollusc shells from the shore zone are subjected to continuous wave and current transport. In addition, water level changes enable the mixing of shells of different ages and climatic periods. Shells of snails, which can contain air in their whorls are especially susceptible to transport. Such transport over great distances may lead to confusing interpretations. For example, on the present shores of Gdańsk Bay there are abundant shells of the freshwater snail *Viviparus*, transported by flood-waters of the Vistula (ALEXANDROWICZ 1985). At Ortel Królewski (ALBRYCHT & *al.* 1995), shells of *Viviparus diluvianus* were deposited by the transport of westerly winds over the lake surface and form a layer 5 m thick.

Considerable differences in the isotopic record for the above snails species collected in the early-glacial deposits of the central zone of lake Ossówka (Text-figs 8, 10), result from an admixture of shells from interglacial littoral deposits, exposed and transported into the central part during lake shallowing (NITYCHORUK & *al.* 1999).

Stable isotopes in laminated lake sediments

General remarks

Occasionally, depositional processes occurring annually in water reservoirs lead to the formation of alternating light and dark laminae. Such sedimentary structures, called varves by DE GEER (1919), in the case of proglacial basins, represent a one-year period and therefore find application in precise dating of the sediments. According to KELTS & HSÜ (1978), such lamination forms when:

a. the lake basin is deep and stratified, with rich production in the epilimnion and organic matter precipitates through a thermally stratified water column,
b. toxic, anaerobic conditions prevail on the lake bottom, eliminating the fauna that homogenises the sediment,
c. there are no near-bottom currents and no intensive transport of detritic material occurs, which could interrupt seasonal deposition,
d. there is limited production of gases from the decay of organic matter, bubbles of which could escape from the sediment and destroy its structure.

The most important are biochemical and organodetritic annual laminae in lake sediments.

Biochemical laminae occur in eutrophic lakes and under anaerobic conditions (STURM & LOTTER 1995). The characteristic succession during a year results from chemical and biological processes (LOTTER 1989). Pale-coloured spring-summer laminae are predominantly composed of biogenic calcite accompanied by diatoms with concentric shapes. The dark colour of the August and September laminae results from the deposition of calcite, elongated diatoms and Chrysophyceae cysts, together with organic matter.

Organodetritic laminas occur solely in oligotrophic overflow lakes with an influx of detritic material (STURM 1979). The summer layers are dark and consist of particles from the spring snow melting and summer floods, when much organic matter is transported into the lake. A pale layer contains mainly clay minerals and is devoid of organic matter.

The above results show that the colour of a lamina cannot always be attributed to a particular season. The occurrence of different types of laminae in the succession is possible, and depends on lake trophy.

Facial variation of annual laminae also corresponds to climatic fluctuations, which was demonstrated for the Late Glacial and early Holocene laminated deposits in the Lake Merfelder Maar in Germany (BRAUER & *al.* 1999). Detailed observations enabled four facies of laminae, characteristic of various climatic conditions, to be distinguished.

Various methods are used to determine, which non-glacial lake laminae formed in certain seasons of the year. There are palynological (MÜLLER 1974), geochemical and micropalaeontological determinations (KELTS & HSÜ 1978; STURM 1979; RZEPECKI 1985; SAARNISTO 1985; LOTTER 1989; KELTS & TALBOT 1990; NUHFER & *al.* 1993; GOSLAR 1993a, b, 1998a, b, c; STURM & LOTTER 1995; BRAUER & *al.* 1997, 1999).

Microscopic analysis of annual laminae in Lake Gościąż based on the presence of cysts of Chrysophyceae, diatoms of the sub-class Centricae, quartz grains, organic matter, quartz crystals, vivianite, pyrite aggregates and pine pollen (GOSLAR 1993b, 1998c) suggested that the maximum concentration of

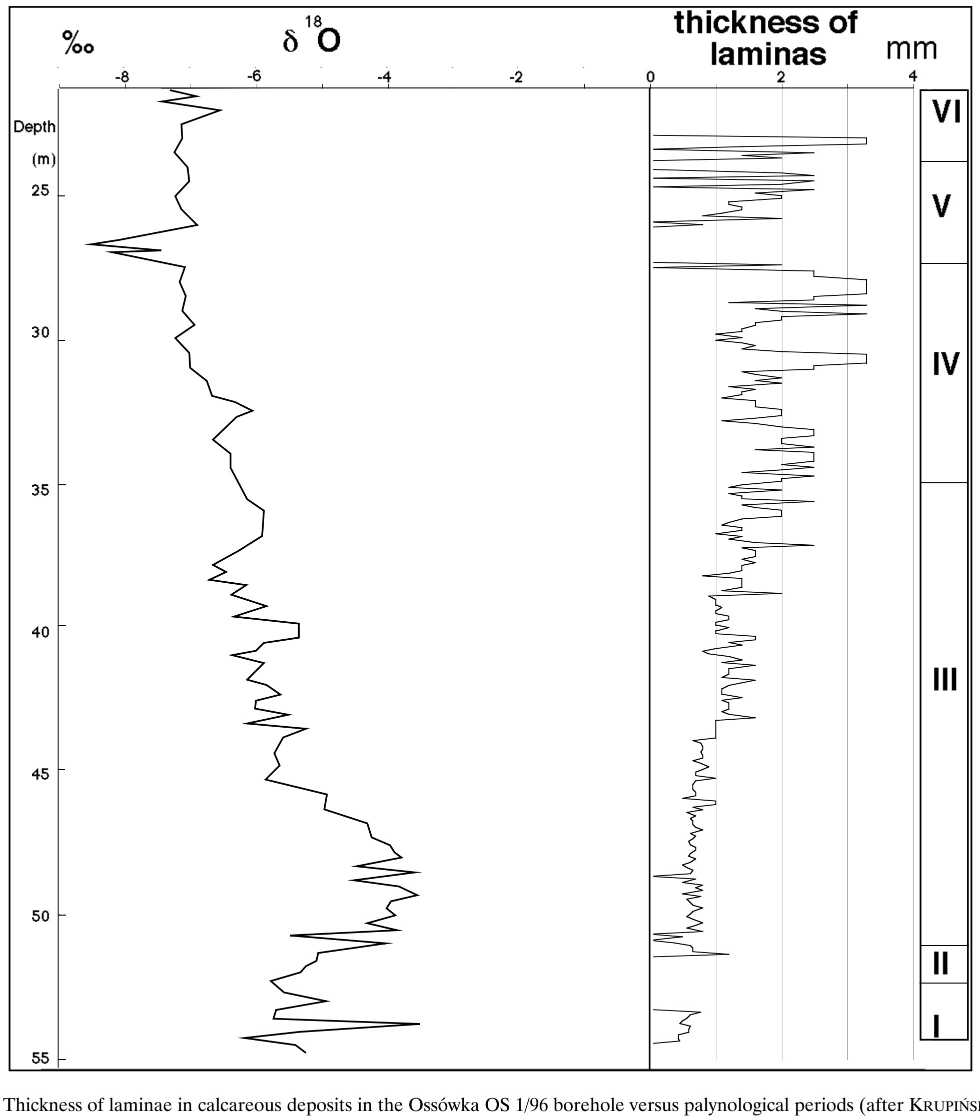

Fig. 11. Thickness of laminae in calcareous deposits in the Ossówka OS 1/96 borehole versus palynological periods (after KRUPIŃSKI 1995) and oxygen isotopes

cysts and *Stephanodiscus* represented spring deposition, large calcite crystals formed during summer, and higher contents of organic matter represented autumn-winter.

Oxygen and carbon isotopes in annual lacustrine laminae – a new approach

The proposed by the author separate isotope determinations for dark and light laminae were expected to discover which laminae, dark or light, corresponded to certain seasons, and thus which trophic conditions characterised the different seasons. The proposed isotopic method enables, regardless of global climatic trends, the determination, on an annual basis, which layers have higher or lower values of $\delta^{18}O$ and $\delta^{13}C$. This method does not require the presence of pollen and diatoms in the sediment. Determination of a season from the isotopic record for dark and light laminae is based on the isotopic content for sediment and water of present-day lakes, determined for particular seasons of the year and at various depths (GAT 1970, 1995; FRITZ & POPLAWSKI 1974; EICHER & SIEGENTHALER 1976; WACHNIEW & RÓŻAŃSKI 1998).

The $\delta^{18}O$ and $\delta^{13}C$ record of modern carbonates in Lake Gościąż during the spring is lower than during the summer-autumn period (WACHNIEW & RÓŻAŃSKI 1998). This difference results from temperature variation between seasons influencing evaporation. The difference is as high as 3‰ for $\delta^{18}O$ and 1‰ for $\delta^{13}C$.

The $\delta^{13}C_{DIC}$ value is also influenced by lake depth. In summer-autumn the thermal stratification has a major influence on $\delta^{13}C_{DIC}$, and is higher in the epilimnion and lower in the hypolimnion (WACHNIEW & RÓŻAŃSKI 1998).

Oxygen and carbon isotopes were determined for almost the entire OS 1/96 core, representing the Mazovian Interglacial (excluding a non-laminated part at 53.3–51.4 m), and the part of the core which represents the deposits of the late interglacial (from the depth 35 m to 27.3 m). Moreover, the isotopes were measured for a small fragment of the early-glacial deposits, at depth 26.0 to 23.0 m (Text-figs 11, 12). At 54.5–53.3 m and 52.2–38.7 m the laminae are very well or moderately well preserved, while in the remaining part of the succession the preservation is poor or very poor (Plate 2). A general upward increase in the thickness of the laminae is observed throughout the laminated part of the succession (Text-fig. 11, Plate 2).

However, variation in the thickness of the laminae may result from climatic fluctuations (BRAUER & *al.* 1999). Therefore, thickness variation of the laminae in the OS 1/96 core will be discussed in the context of changes in plant cover shown by a pollen diagram of the Os 1/90 core (Text-fig. 6) (KRUPIŃSKI 1995), and also changes in the $\delta^{18}O$ record (Text-fig. 8).

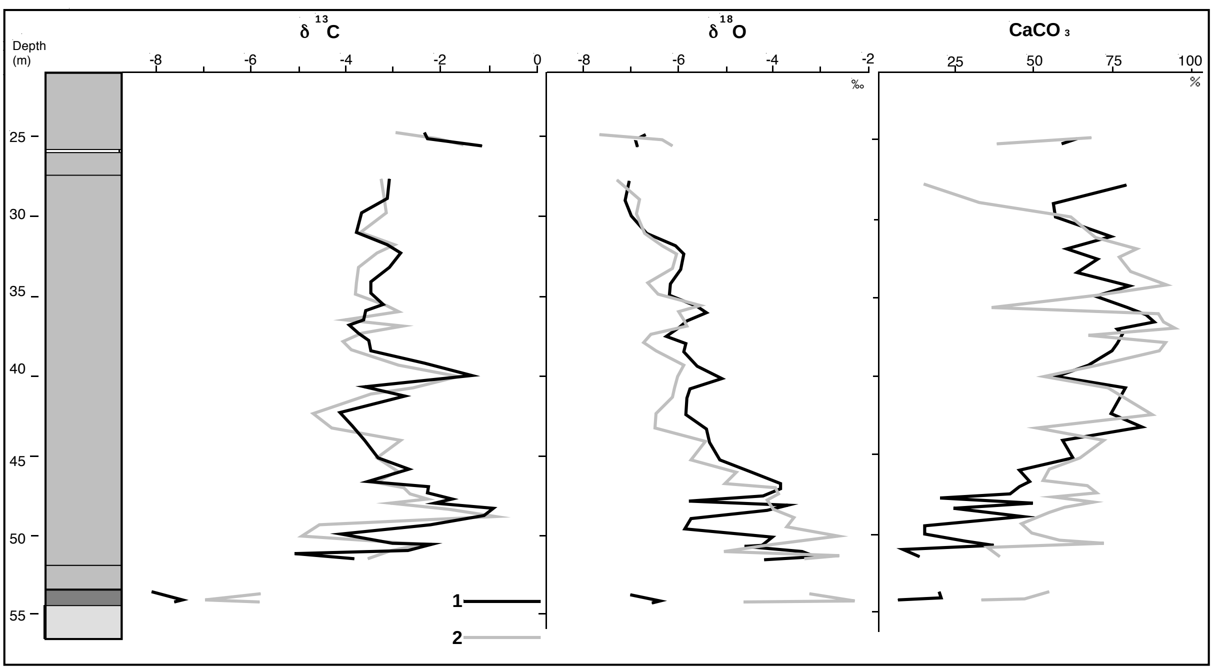

Fig. 12. Oxygen and carbon isotopes, and carbonate content for dark (1) and light laminae (2) in the Ossówka OS 1/96 borehole; for lithological explanations see Text-fig. 8

Thickness of non-glacial annual laminae in the OS 1/96 core

Relationship to climatic changes

Clayey muds and muddy sands of the last phase of the South-Polish Glaciation (Elsterian) form 1 cm thick laminae at 54.5-56.5 m, but these have not been yet analysed. Lamination of the deposits indicates the presence of a temporary water basin at that time.

The boundary between the glaciation and the interglacial is very clearly visible on the oxygen isotope curve (Text-fig. 8). It is expressed by a rapid rise in $\delta^{18}O$ values to –3.6‰. In the pollen diagram this part of the succession represents pollen period **I**, reflecting gradual warming with distinct continental features. Laminated deposits preserved at 54.5-53.3 m (Text-fig. 11) are very thinly layered (0.5 mm). A decrease in the heavy isotope, ^{18}O content coincides with the disappearance of lamination and is associated with the transition to pollen period **II**. The climate became more humid, accompanied by the reappearance of laminated deposits at 51.5 m (Text-fig. 11). Variation in the oxygen isotope contents of 1.5‰ at that time may be explained by alternating continental and maritime influence, which seems to be confirmed by a change in the thickness of the laminae from 0.4 mm to 0.8 mm, and by the absence of lamination at 50.9 and 50.7 m respectively. The average thickness of the laminae in this section is 0.7 mm (Plate 2). Another short break in lamination at 48.7 m coincides with the cooling and distinct continentalisation marked by pollen zone F of pollen period **III**. Less favourable climatic conditions are reflected in a decrease in thickness of the laminae from 0.7 to 0.5 mm.

Pollen zones G and H are characterised by a temperate and humid climate with a long vegetation period and mild winters. The thickness of the laminae increased markedly, on average from 0.7 to 1.2 mm (Text-fig. 11). This feature is particularly relevant to the discussion by palynologists regarding the lower boundary of the climatic optimum of the Mazovian Interglacial. According to KRUPIŃSKI (1988a, 1995), this boundary runs between pollen periods **II** and **III**, that is between zones D and E (Text-fig. 6). According to BIŃKA (BIŃKA 1994, BIŃKA & NITYCHORUK 1995, 1996), on the other hand, the boundary lies higher (Text-fig. 7), at the beginning of pollen zone **7** (*Carpinus-Abies*).

Thicker laminae during the prevailing maritime climate at the interglacial climatic optimum could have been associated only to a limited extent with redeposition because of the consistent plant cover and overall stable hydrological conditions. The thickening of the laminae probably resulted from warming. This intensified photosynthesis and the withdrawal of CO_2 from the lake water, thereby contributing to increased precipitation of carbonates. Intensive bio-production, stimulated by favourable thermal conditions, also intensified organic sedimentation and the formation of thicker laminae (Plate 2). Changing content of $CaCO_3$ in the sediment, from about 60% to 80%, and the presence of abundant diatoms, confirm the operation of these processes.

The increasing thickness of the laminae at about 44.0 m may support the opinion of BIŃKA (1994) regarding the position of the lower boundary of the climatic optimum of the Mazovian interglacial. During the interglacial climatic optimum, which was stable with respect to the development of plant cover, fluctuations in the content of oxygen and carbon isotopes indicate minor climatic changes.

The first distinct deflection in the $\delta^{18}O$ curve (at 43.5 m) coincides with the first minimum of fir pollen, accompanied by a short-term thickening of laminae from 1.1 to 1.6 mm (Text-fig. 11). The second deflection in the curve is much more marked. It occurs at 40.0 m and coincides with the second minimum of fir pollen and with an increase in thickness of the laminae from about 1.2 mm to 1.6 mm (Plate 2), followed by a thinning of the laminae to 1.0 mm over about a 2 m long length of core. A climatic explanation of this was presented in the chapter on the interpretation of isotopes in the OS 1/96 succession. A retrieval of fir was caused by slight continentalisation of the climate, accompanied by lake shallowing. The latter was linked to an increase in trophicity due to the presence of the same amount of nutrients in a smaller volume of water. As a result of the increased nutrient concentration, algae flourished, and the intensive charge of CO_2 from water intensified the precipitation of carbonates and the sedimentation of organic matter. Lake shallowing can also be associated with a higher influx of terrigenic matter. Thinner laminae could have been caused, on the other hand, by slightly lower production of carbonates and reduced photosynthesis by plankton.

The second crisis at Ossówka was probably caused by the previously described volcanic catastrophe, which resulted in less favourable climatic conditions. In the post-crisis sediments there was no lamination, and the diatom content, previously the major component, drastically decreased.

In the final phase of the interglacial a drop of $\delta^{18}O$ values to below –7.0‰ occurred, which corresponds well to the cooling shown by the tree pollen (KRUPIŃSKI 1995). Thickening of the laminae to an average of 2.5 mm, i.e. 1.2 mm more than at the interglacial optimum, is not significantly reflected in the reduction in $CaCO_3$

content of the sediment from about 70 to 60% (Text-figs 11, 12) and indicates only a slight increase in transport of clastic material into the basin. High carbonate production and a high sedimentation ratio of organic matter under colder climatic conditions can be explained by gradual shallowing of the lake, accompanied by an increase in its trophicity. It led to concentration of nutrients in a smaller volume of water. Continuing lake shallowing resulted in a rise of the content of *Pediastrum* (KRUPIŃSKI 1995).

During the early glaciation the climate became subarctic in pollen period **V**. A distinct cooling phase, marked by scrubby-tundra communities, is very well reflected by the lower $\delta^{18}O$ values and the absence of laminae (27.3–26.0 m). A gradual climatic amelioration resulted in deeper lake water and was associated with the deposition of laminated deposits (26.0–23.0 m) (Text-figs 11, 12). Initially severe climatic conditions did not favour rapid deposition and the laminae were less than 1 mm thick. The transition from cool pollen zones (K and L) to the warmer one (M) was marked by thicker laminae (average 2 mm).

Isotopes in non-glacial annual laminae in the OS 1/96 core – $\delta^{18}O$

$\delta^{18}O$ values for the first laminated part of the core in the 54.5-53.3 m interval (Text-fig. 11) varies considerably between the light and dark layers, a feature that is not repeated in the other part of the core. $\delta^{18}O$ values for the dark laminae reach –6.5 – –7.0‰, while the values for the light laminae range between –4.5‰ and –2.0‰. In the most extreme case, the difference reaches 4.0‰ (Text-fig. 12). The rather low $\delta^{18}O$ values for the dark laminae correspond approximately to the prevailing climatic conditions of long frosty winters and warm summers.

High soil moisture and air humidity were dominant features at that time (KRUPIŃSKI 1995). High $\delta^{18}O$ values for the light laminae may have been caused by transport of allochthonous calcite into the lake. Such a process may also be suggested by a minor admixture of sand occurring cyclically in lake marls in the core from Woskrzenice (BIŃKA 1994). According to BIŃKA (1994) these do not simply indicate a short distance from the shore, but may result from storm precipitation causing transport of coarser mineral material into a deeper part of the basin. Allochthonous calcite could have been transported to the lake from an area with incomplete plant cover. In particular, a considerable quantity of carbonate of marine origin was present in every deposit in the lake basin. As described above, these carbonates were derived from glacial rafts of Cretaceous rocks a dozen or so kilometres to the NW (ALEXANDROWICZ & ŚLUSARCZYK 1963, ALEXANDROWICZ & RADWAN 1992).

As the allochthonous calcite could have been transported by storm precipitation, this process seems to have not taken place in winter, when the lake was covered with ice, but rather during a warmer season i.e. late spring and summer. More $CaCO_3$ in the light laminae (30 – 50%) than in the dark ones (about 20%), could have resulted from a contribution of allochthonous calcite, but may also indicate deposition during a warmer season when photosynthesis was more intensive (Text-fig. 12). Calcium carbonate can also precipitate in lake water through physico-chemical processes. Calcium carbonate is less soluble in water at lower temperature. Therefore, cooler hypolimnion waters unsaturated with calcium carbonate, warm up after reaching the lake surface during the spring circulation and become over-saturated (KELTS & HSÜ 1978).

The spring layer will therefore be lighter coloured and richer in $CaCO_3$. The pattern of lower isotopic ratios of oxygen for the dark laminae and higher for the light laminae in this part of the core is the opposite of that obtaining in present-day lakes (GAT 1970, 1995, EICHER & SIEGENTHALER 1976, WACHNIEW & RÓŻAŃSKI 1998).

After a break (between 53.3 and 51.5 m), caused by the absence of lamination in the sediments (Text-figs 11, 12), the concentrations of isotopes in the annual laminae are less differentiated (about 2‰), but in most samples (to 46.7 m) the light laminae still reach higher $\delta^{18}O$ values than the dark ones. The average content of $CaCO_3$ in the light laminae is 60%, and 30% in the dark laminae (Text-fig. 12). The only two levels where there are higher $\delta^{18}O$ values in a dark lamina than in a light one occur at 50.9 and 48.0 m. In both cases, these levels coincide with higher $CaCO_3$ content in the sediment, which probably resulted from more intensive bio-production in the lake under the warmer climatic conditions. In the first case, this warming may have resulted from the circulation of maritime air postulated by BIŃKA (BIŃKA 1994, BIŃKA & NITYCHORUK 1995, 1996, BIŃKA & *al.* 1997) in subzone *b* of the *Picea-Alnus* pollen zone – (Text-fig. 7). The second case may be linked with a warming phase following a cooling phase in pollen period **III**, zone E (KRUPIŃSKI 1995 – Text-fig. 6) and, according to BIŃKA (BIŃKA 1994, BIŃKA & NITYCHORUK 1995, 1996), in subzone *d* of the *Pinus-Larix* pollen zone (Text-fig. 7). Warm and humid climate predominated at that time, with humid winters and relatively dry summers. Dry summers could have caused higher $\delta^{18}O$ values in the light laminae, as enrichment in the heavy oxygen isotope took place, resulting from intensive evaporation.

At 46.7 m, $\delta^{18}O$ values for both annual laminae were about –4.0‰ (Text-fig. 12). In the upper part of the section, the $\delta^{18}O$ values subsequently decrease to about –6.0‰. This tendency coincides with distinct warming indicated in pollen diagrams (BIŃKA 1994, KRUPIŃSKI 1995), which is assumed to indicate the beginning of the Mazovian Interglacial climatic optimum (according to BIŃKA 1994 – Text-fig. 7). In contrast to the situation in the lower part of the succession, above 46.7 m light laminae reach lower and dark laminae higher $\delta^{18}O$ values, as in present-day lakes (GAT 1970, 1995, EICHER & SIEGENTHALER 1976, WACHNIEW & RÓŻAŃSKI 1998). In addition, the $CaCO_3$ content of the sediment rises to an average of 70-80%, with more calcium carbonate present in the light laminae (Text-fig. 12).

The above relationship continues during the entire interglacial optimum (Text-fig. 12). Only in its final part was there a rise in $\delta^{18}O$ values for the light laminae that was twice as high as for the dark laminae, and was accompanied by a rapid decrease of calcium carbonate content in the light laminae. Such a situation was due to a temporary climatic deterioration, reflected by changes in $\delta^{18}O$ values in the OS 1/90 core (Text-fig. 13), accompanied by a higher content of birch pollen (*see* the chapter on isotopes relating to the OS 1/90 core).

Isotopic curves for the OS 1/96 and 1/90 cores show (Text-figs 8, 13) that cooling in the pollen period of the final interglacial phase did not take place immediately. This hypothesis is confirmed by the fact that the $\delta^{18}O$ record for the annual laminae retains the distribution of values characterising the interglacial climatic optimum (Text-fig. 12). It was not until 2000 – 3000 years after the interglacial optimum that there was a gradual increase of $\delta^{18}O$ values, with higher values in the light laminae than in the dark laminae, coinciding with a gradual cooling and a more continental climate. A rapid decrease in the $CaCO_3$ content in the light laminae (from 70% to 20%) also took place.

During the early glaciation the climate changed into a subarctic one (pollen period **V**). In zone K ground drainage occurred simultaneously with a fall in tempera-

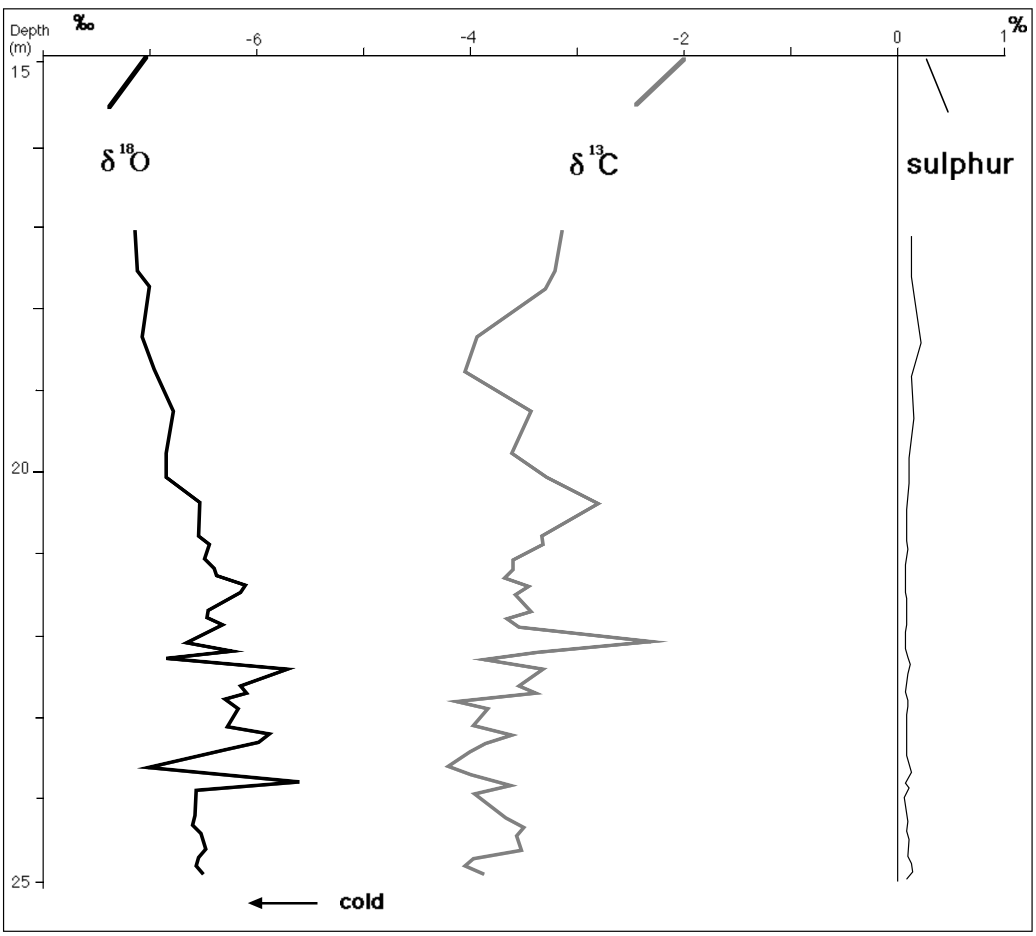

Fig. 13. Oxygen and carbon isotopes curves and sulphur in the Ossówka OS 1/90 borehole succession from the end of the Mazovian Interglacial optimum and the early glaciation

ture, especially in winter. The absence of lamination in the sediments was caused by lake shallowing. In zone L, drainage and the thermal contrast between the seasons became slightly smaller, which resulted in a longer vegetation period (KRUPIŃSKI 1995). The $\delta^{18}O$ values for the light laminae were high, which was caused by dryness in summer and by intensive evaporation. A distinct warming phase occurred in zone M, when humidity and intensity of solar radiation increased. The $\delta^{18}O$ values decreased for the light laminas and increase for the dark ones.

Summarising, the oxygen isotope ratios in the lower part of the succession, when a cooler climate with continental features prevailed, with dry summers and cool and long winters, are higher for the light laminae (spring-summer) and lower for the dark ones (autumn-winter). The maritime climate during the interglacial climatic optimum, with humid summers and mild winters, was characterised by the opposite relationship. The light laminae have lower $\delta^{18}O$ values than the dark laminae, which is similar to the pattern in present-day lakes (GAT 1970, 1995, EICHER & SIEGENTHALER 1976, WACHNIEW & RÓŻAŃSKI 1998).

The isotopic ratios for some samples from the lowermost and uppermost parts of the interglacial succession, with measured values opposite to the general trend, record the short-term influences of maritime climate at the beginning of the interglacial periods, with the prevalance of a continental climate, and the short-term influences of continental climate at the end of these periods dominated by the maritime climate.

Isotopes in non-glacial annual laminae in the OS 1/96 core – $\delta^{13}C$

In the laminae from the lake sediments at Ossówka (OS 1/96) the carbon isotopes change in a similar way to the oxygen isotopes. Maritime and continental climate is reflected in the isotopic curves for the light and dark laminae. At 54.5–53.3 m the $\delta^{13}C$ values are the lowest in the entire succession, which is the result of a relatively cool climate and low bio-production in the lake (Text-fig. 12). In the light laminae the range of $\delta^{13}C$ values is –6.0 to –7.0‰, while in the dark laminae it is –7.5 to –8.0‰.

Such low values of the carbon isotope ratios may indicate a slight contribution of older carbonates in the sediment, since the $\delta^{13}C$ values for rocks of the Cretaceous complex range from 1.8 to 2.3‰. Higher $\delta^{13}C$ values in the light laminae may be due to a longer period when the lake was covered by ice. Long-term ice cover on the lake could have resulted in a shorter vegetation season. This then meant that the light layer formed in summer, when photosynthesis increased as the temperature rose. Water plants prefer the lighter carbon isotope ^{12}C, resulting in enrichment in the heavier isotope ^{13}C in the precipitating carbonates (PARK & EPSTEIN 1960, HOLLANDER & *al.* 1988).

At the beginning of the interglacial climatic optimum (in the understanding of KRUPIŃSKI 1995 – Text-fig. 6) $\delta^{13}C$ values continuously increased to –1.0‰ (Text-fig. 12). Variations in the $\delta^{13}C$ values in the light laminae confirms the dependence between isotopic ratios and climatic changes towards continentalisation and oceanisation, which was presented in the discussion on oxygen isotopes. A general rising trend of $\delta^{13}C$ values in this part of the succession reflected lake shallowing.

The interglacial climatic optimum was associated with warming and with lake deepening. The isotope curve drops gradually to lower values, finally even below –4.0‰ (Text-fig. 12). The dark laminae have higher $\delta^{13}C$ values then the light ones, firstly due to the influx of groundwaters containing more light isotopes because of their contact with CO_2 in the soil, especially during summer when root respiration is more intensive (KUC & *al.* 1998). The second reason is that ground and river waters coming from areas with dense vegetation have lower $\delta^{13}C$ values, associated with intensive oxidation of organic matter rich in light isotopes (JĘDRYSEK 1994, JĘDRYSEK & *al.* 1994, 1995).

Higher $\delta^{13}C$ values in the dark laminae (autumn-winter) than in the light laminae during the interglacial climatic optimum were common and coincide with what is observed in modern lakes (WACHNIEW & RÓŻAŃSKI 1998), but the reverse also occurred. Exceptions to this rule, i.e. higher $\delta^{13}C$ values in a light lamina occur at about 43.5, 40.5, 37.0 and 36.0 m. The first two episodes can be identified in the pollen diagrams (BIŃKA 1994, KRUPIŃSKI 1995) as a decrease in the content of fir pollen which, as already discussed above, was associated with lower precipitation, drainage and slight continentalisation. These changes resulted in a reduction of about 20% in the calcium carbonate content of the light laminae, which can be explained by a lower biological activity in the lake. An enrichment of the light laminae in ^{13}C could have been caused in turn by intensive evaporation under dry climatic conditions (STUIVER 1970).

The other two events correspond to an increased content of birch pollen, which can be interpreted as resulting from the influence of the circulation of continental air at the end of the interglacial climatic optimum. Cooling, drainage, a shorter vegetation season and drop in organic production in the lake, contributed to the increased content of the heavy carbon isotope.

The next cooling episode, observed in the final phase of the interglacial (about 30 m), is marked by increasing $\delta^{13}C$ values in the light laminae (Text-fig. 12) and is linked with lake shallowing resulting from infilling of the basin with sediments and a change in the relationship between the water volume in the epilimnion and hypolimnion – E/H (*see* RÓŻAŃSKI & *al.* 1998).

Summarising, both carbon and oxygen isotopes measured for single laminae in the annual cycle are a good indicator of climatic changes at particular phases of lake development. The results of analysis of the alternating curves obtained from light and dark laminae enabled the reconstruction of temperature-humidity changes corresponding to the periods of prevailing influence of continental or maritime air circulation.

OS 1/90

Isotopic values were determined for selected parts of the 36.3 m long core of the OS 1/90 borehole (Text-fig. 13), for which a pollen diagram had already been published (KRUPIŃSKI 1995). The studies focused on the last part of the Mazovian Interglacial climatic optimum and the initial phase of the following glaciation (core interval 15.0 – 25.0 m). This part of the core corresponds to zones H-5 to M (according to KRUPIŃSKI 1995 – Text-fig. 6).

OS 1/90 – $\delta^{18}O$

During the last phase of the interglacial optimum, when a mild climate with maritime influence prevailed (Text-fig. 13), the δ^{18}O values stay at –6.5‰. At 23.8 m the δ^{18}O curve rapidly rises about 1‰, which may be identified with temporary warming. At 23.6 m the δ^{18}O values decrease about 1.4‰, which reflects a cooling phase. The warming phase interpreted from the isotopic curve is indicated in the pollen diagram by an increased content of birch pollen (*Betula*) from about 5% at 23.6 m to 30% at 23.5 m. A few percent increase in the content of *Pediastrum* at 23.6 m, suggests that cooling resulted in a short-term, slight shallowing of the lake, which increased its trophicity. However, this event is not recorded in as an increase in δ^{18}O values what may suggest a marked temperature drop. A very similar situation is also noted in the higher part of the succession (Text-fig. 13). Initially, following a cooling phase, δ^{18}O values decreased first to –6.0‰ and then to –6.3‰. At 22.5 m they increased to –5.8‰, and finally decreased at 22.4 m to –6.9‰. At 22.5 m the curve of birch pollen indicates a rise of 10%. This was also accompanied by a slight increase of *Pediastrum*. The interglacial climatic optimum terminated after a second culmination of birch pollen (KRUPIŃSKI 1995). Gradual decrease of δ^{18}O toward negative values may indicate a very gradual transition from a warm into a post-optimum climate.

Initially, after a second cooling determined on the basis of fluctuations of the oxygen curve and the increased content of birch pollen, δ^{18}O values increased to –6.2‰, followed by their continued decrease to below –7.0‰. A distinct cooling phase associated with pollen period **V**, and clearly recorded by the isotopic curve of the OS 1/96 core, has not been studied in this core.

OS 1/90 – $\delta^{13}C$

During the Interglacial climatic optimum, δ^{13}C values in the OS 1/90 core oscillate around –4.0 – –3.5‰ (Text-fig. 13). Until the second cooling phase, marked by the increased content of birch pollen, both the oxygen and carbon isotopic curves have approximately similar shapes, which may indicate the hydrology of a closed lake basin. Only after the second cooling phase do the curves no longer correlate with each other and the δ^{13}C values rise to –2.3‰. This temporary increase may have been caused by lake shallowing, associated with the second birch cooling phase. The next decrease in δ^{13}C values indicates a stable sedimentary environment. The next increase in δ^{13}C values to –2.8‰ coincides with the extinction of stenothermal trees, after which a cooling phase recorded by pollen period **V**, is represented by the isotopic curve reaching –4.0‰. This probably indicates the lake shallowing reflected in a rise of *Pediastrum*. The cooling which began the glaciation was not studied in this core, however δ^{13}C values increased to –2.0‰ immediately after this period.

Stable isotopes in fossil carbonate lake deposits in the Wilczyn basin

WL 1/96 – $\delta^{18}O$

The basin at Wilczyn (Borehole WL 1/96; Text-fig. 14) started to develop during pollen zone **3** (*Pinus-Betula*), with already relatively high average July temperatures in the range 14-17°C. The deposition began (37.0-37.3 m) with sands with muds and sandy muds with organic matter.

No isotope analyses of these sediments were carried out, as the results would have been inconclusive. Grey-olive gyttja with distinct millimetre-thick laminae, which overlies sands with mud, has been subjected to isotope analysis. A high δ^{18}O value at the bottom (–7.0‰ – Text-fig. 14) is caused by lack of cool pre-interglacial deposits, and corresponds well to the prevailing mild temperatures (BIŃKA 1994). A gradual warming coincided with higher δ^{18}O values. The following samples reached –6.0‰ and, after a small drop to

–6.4‰ and a subsequent rise to –5.5‰, another drop to –6.4‰ occurred (Text-fig. 14). The studied samples belong to the pollen zone **4** (*Picea-Alnus*), which consists of three subzones. In the subzones *a* and *c* continental climate prevailed, while subzone b was dominated by maritime climate. A drop in $\delta^{18}O$ values can be correlated with the subzones *a* and *c*, while an increase in values is connected with subzone *b*. Higher $\delta^{18}O$ values may be correlated with a maritime climate and with precipitation containing more of the heavy isotope ^{18}O (*see* RÓŻAŃSKI & *al*. 1993). A small transgression of the Holsteinian sea, which possibly occurred in subzone *b* of pollen zone **5** (*Taxus*) (BIŃKA & *al*. 1997), may have caused another increase of $\delta^{18}O$ values to –5.5‰.

A drop of in $\delta^{18}O$ values to about –6.0‰ was followed by another increase to –5.5‰. This event can be confidently correlated with pollen zone **6** (*Pinus-Larix*) and more precisely with subzone *b* (*Carpinus-Quercus*). Despite a continental climate at this time, a warming took place, accompanied by draining of the habitats occupied by spruce and alder. This resulted in a decrease in spruce and alder pollen, and an increase in hornbeam and oak pollen (Text-fig. 7). Dryness of the climate could have caused a small lowering of the lake water level, and this in turn could have contributed to a higher concentration of heavy isotopes (higher $\delta^{18}O$ values). In subzone *c* (*Pinus-Betula*) a distinct cooling phase is recorded (MÜLLER 1974, BIŃKA 1994). Interpretation of the isotopic data suggest another change, reflected (at about 26.5 m) by a drop in $\delta^{18}O$ values from about –6.0 to –6.7‰.

Pollen zone **7** (*Carpinus-Abies*) was dominated by a mild climate with clear maritime influence. The $\delta^{18}O$ values reflect neither a warming nor oceanisation, as they are more negative that in the lower part of the succession (Text-fig. 14). The highest value recorded was –6.0‰, and the average in the pollen subzones a (*Carpinus-Corylus*), *b* (*Abies*), *c* (*Quercus-Carpinus*) and *d* (*Abies*) was about –6.5‰. In subzone *e* (*Quercus-Carpinus*) it was –7.4‰, and at the fir pollen minimum it was –7.5‰, after which it rose to –6.5‰ in subzone *f* (*Abies*). A warmer and more humid climate in the zones with increased fir pollen does not satisfactorily explain such a big rise (1‰) in $\delta^{18}O$ values in subzone *f* and subsequent fall in subzone *e*. Changes of the oxygen isotopic curve may have been

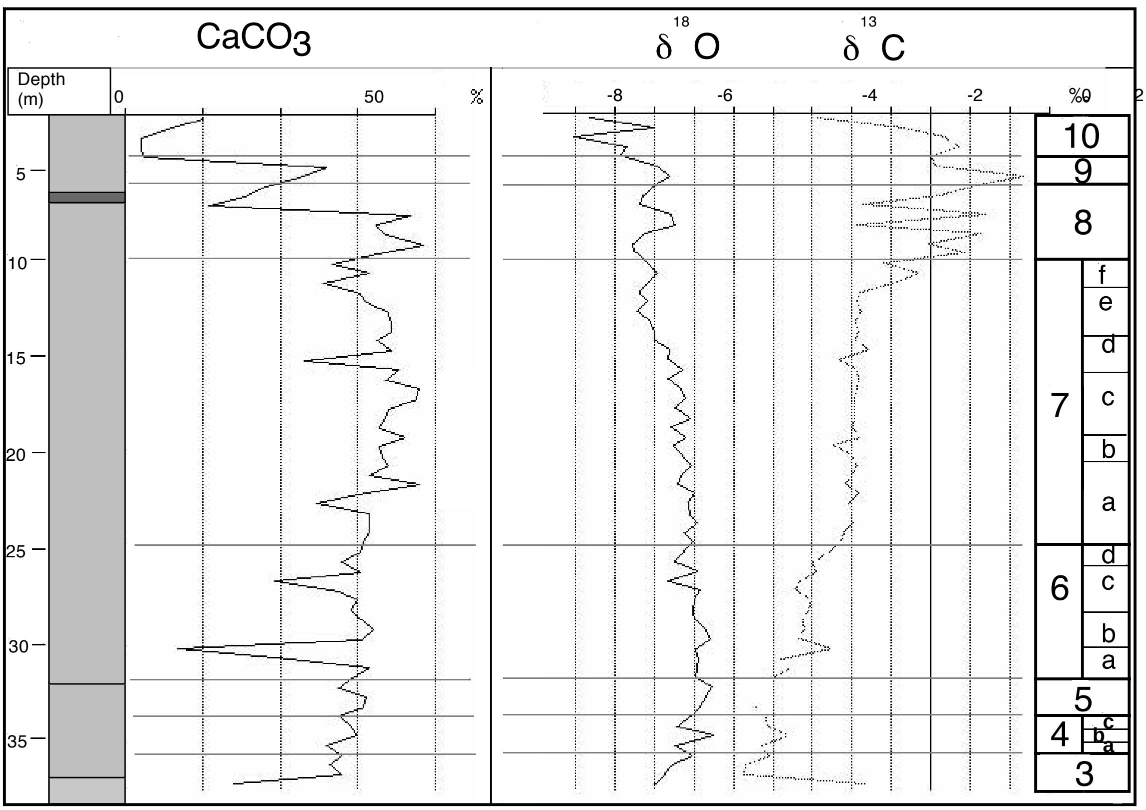

Fig. 14. Oxygen and carbon isotopes and calcium carbonate content in the Wilczyn WL 1/96 borehole versus the palynological zones (3, 4) and subzones (a, b, c) after BIŃKA & NITYCHORUK (1995, 1996), BIŃKA *& al.* (1997); for lithological explanations see Text-fig. 8

associated with the volcanic activity described in the discussion of the OS 1/96 core, and/or with the continentalisation that led to cooling and to lake shallowing. The latter is found in the increased content of organic matter, marked by a change in the sediment colour (at 10.0 m) from grey to black. The basin at Wilczyn, which was largely filled with deposits and was therefore shallower than the lake at Ossówka, hardly survived the volcanic crisis, which is expressed by a depositional break. In pollen zone **8** (*Pinus-Picea-Pterocarya*), which followed the climatic optimum, a decrease in $\delta^{18}O$ values from –6.3‰ to –7.4‰, and a reduction in the carbonate content from 90% to 25% (Text-fig. 14), clearly reflect further cooling. A distinct drop of the water level in the lake is documented at 6.2-6.6 m by a black gyttja with numerous bivalve and snail shells. In pollen zone **9** (*Pinus-Juniperus*), a further cooling took place which, however cannot be interpreted from the $\delta^{18}O$ curve, as the latter records a rise in values from –7.4‰ to –6.6‰. A further distinct shallowing of the lake was associated, despite a cooling, with a higher trophicity, reflecting an increase in *Pediastrum*.

The pollen zone **10** (*Artemisia*-Chenopodiaceae) contains steppe plant communities that indicate a further cooling, associated with a decrease in $\delta^{18}O$ values to –9.0‰.

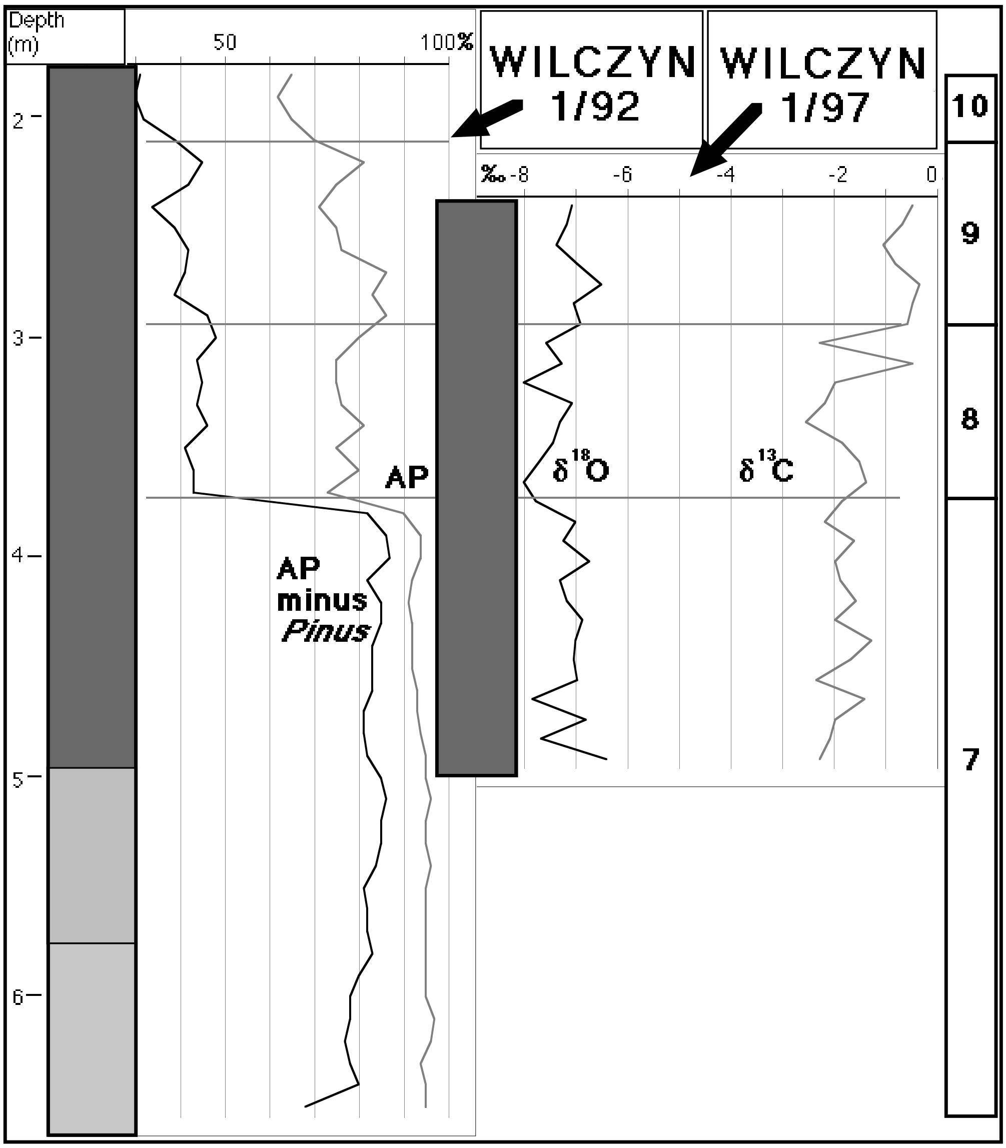

Fig. 15. Oxygen and carbon isotopes for the lake deposits in the Wilczyn WL 1//97 borehole; contents of tree pollen AP minus *Pinus* and all tree pollen (AP) versus the biostratigraphic zones after BIŃKA *& al.* (1996, 1997); for lithological explanations see Text-fig. 8

WL 1/96 – δ [13]C

The δ[13]C values reach –1.6‰ at the beginning of development of the basin at Wilczyn (Text-fig. 14). Gradual warming is associated with a decrease in δ[13]C values to –4.7‰. The following increase, up to –3.6‰, might have been caused by maritime influence. After a small decrease in δ [13]C values to –4.4‰, a rise to –2.5‰ took place, followed by another fall to –3.4‰. The δ[13]C values remained below –3.0‰ until a cooling phase in subzone *c* (*Pinus-Betula*). Pollen analysis indicates a warming after the cooling phase, i.e. the interglacial climatic optimum, accompanied by an increase in δ [13]C values from about –3.0 to –2.0‰.

At about 10.0 m, there is a fall in δ[13]C values, accompanied by a reduction in carbonate content from 90% to 25% (Text-fig. 14). These changes resulted from the volcanic crisis, which influenced deposition in the lake. The δ[13]C curve shows an increase in values at this level from –1.8 to –0.4‰, but higher up it reaches the maximum value of 2.3‰. This is unquestionably caused by a considerable drop of water level in the lake, which is documented by a change in the type of deposits and the presence of mollusc shells.

Despite cooling, a further distinct shallowing of the lake is accompanied, by a higher trophicity, indicated by an increased content of *Pediastrum*.

Boreholes WL 1/92 and 1/97

Both of the boreholes at Wilczyn were drilled close to one another (Text-fig. 4). Borehole WL 1/92 was used to construct a pollen diagram (BIŃKA 1994, BIŃKA & *al.* 1996, 1997). Borehole WL 1/97 supplied a core of lake sediments and mollusc shells for the isotopic studies. In this chapter results of these studies are discussed.

The pollen diagram spans the core interval from 1.8 to 6.5 m, while the isotopic analysis was conducted for the interval from 2.3 to 4.9 m (Text-fig. 15).

WL 1/97 – δ[18]O

Large fluctuations of the oxygen isotopic curve in the lower part of the WL 1/97 core between 6.4 and 7.9 m (Text-fig. 15) were probably associated with water level changes in the palaeolake, which coincide with changes in the contents of *Pediastrum* in the pollen diagram (BIŃKA & *al.* 1996, 1997). δ[18]O values in the upper part of the succession oscillate around –7.0‰ until the second minimum of fir pollen occurs. This was followed by the volcanic crisis discussed above, which is marked in lake sediments by a decrease in δ[18]O values to –8.0‰ and by a drastic decrease in tree pollen in the pollen diagram (BIŃKA & *al.* 1996, 1997). The end of the volcanic crisis was associated with warming and with an increase in δ[18]O values to the previous level of –7.0‰. The retrieval of stenothermal trees, characteristic of a cooling phase, does not correlate with a decrease in the oxygen isotope values, since the latter actually increase. This fact can be probably be explained by lake shallowing, which is also reflected in the increased content of *Pediastrum*. The shallowing resulted in a higher content of the heavy isotope [18]O.

WL 1/97 – δ[13]C

During the interglacial climatic optimum δ[13]C values oscillated around –2.0‰ (Text-fig. 15), what was probably caused by variations in the palaeolake water level, associated with varying humidity of the air. The volcanic crisis, at the end of the climatic optimum, did not cause significant changes in δ[13]C values, which merely increased from –2.1 to –1.4‰. On the other hand, the rapid decrease in content of stenothermal tree pollen and the absence of pollen subzone *f* of the *Carpinus-Abies* zone, could have resulted from a sedimentary hiatus caused by lake shallowing.

The fact that the studied core represents a littoral zone, which is highly susceptible to water level changes in the lake, also confirms the adopted interpretation. Due to a sedimentary break, the volcanic crisis is not expressed by rapid changes in the isotopic curves, as is common at Kaliłów.

The small variability of the isotopic curves from the littoral sediments, together with the probable existence of depositional gaps, makes the isotopic record incomplete, and therefore of little use for a detailed paleoclimatic and paleoenvironmental interpretation (*see* KUC & *al.* 1998).

Isotopes in mollusc shells

Isotopic studies of sediments and mollusc shells from the littoral zone of the lake at Wilczyn yielded information regarding water level changes. For this purpose, two oxygen isotopic curves were compared: one for the snail *Valvata piscinalis* and the other for the deposits (Text-fig. 16). The comparison revealed that some parts of the two curves were similar and coincided with stable conditions of lake development. Lack of correlation of the curves corresponds to a climatic crisis which led to lake shallowing and was accompanied by erosion and redeposition by waves, as was described in the chapter on the OS 1/96 core (Text-fig. 10).

The results of isotopic analysis of mollusc shells from the WL 1/97 core suggest the following conclusion.

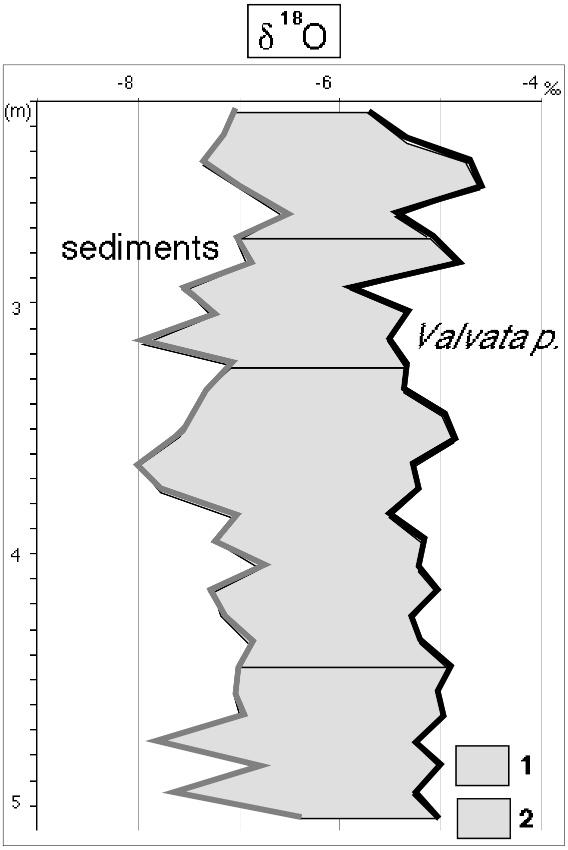

Fig. 16. Correlation of oxygen isotopes for calcareous sediments and shells of the snail *Valvata piscinalis* in theWilczyn WL 1/97 borehole; 1 – positive correlation; 2 – lack of correlation or negative correlation

Isotopic values from shells of *Valvata piscinalis* only were selected for comparison with isotopic values from sediments for three reasons. Firstly, this snail was more common in the core than other species, providing a good chance of obtaining a continuous isotopic curve. Secondly, the isotopic values obtained for *Valvata piscinalis* corresponded better to the results obtained from the lake sediments. The third factor was the opportunity of selecting better preserved, complete shells of *Valvata piscinalis*.

The influence of bad preservation of the shells on the results of isotopic analysis requires further explanation. For the isotope preparation a complete snail shell was used, which was carefully cleaned, washed and ground. It was quite easy to find complete shells of *Valvata piscinalis,* whereas complete shells of *Viviparus diluvianus* occurred only rarely. To check a possible effect of incomplete shells of *Viviparus diluvianus* on the results several determinations of successive growth (ontogenetic) segments of a single shell were carried out (Text-fig. 17). Three specimens were selected for the measurements, two of which had the growth segments arranged in a similar way. The results obtained suggest that the growth segments visible on a shell of *Viviparus diluvianus* were formed during a single year, and therefore represent the period of growth from spring to autumn. The identical arrangement of growth segments on two shells indicates that the snails lived at the same time. Considerable differences of isotopic values in a single shell may influence the measurements if a damaged and incomplete specimen is examined. The use of complete shells with a similar number of growth segments, i.e. corresponding to the same age of the snail, enables comparable isotopic results to be obtained. Single annual growth segments of a shell, if studied at the beginning, i.e. the part formed during spring, and at the end, corresponding to autumn growth (Text-fig. 17), may give information on isotopic values for the vegetation season in a single year.

KA 1/93 – $\delta^{18}O$

The isotopic analysis of the Kaliłów core (Text-figs 2, 4, 18) was carried out on the 3.5–9.0 m interval (NITYCHORUK & *al.* 1995). On the pollen diagram (BIŃKA & NITYCHORUK 1996), this interval comprises pollen zone **7** (*Carpinus-Abies*), from subzone *c* (*Quercus-Carpinus*) through *d* (*Abies*) and *e* (*Quercus-Carpinus*) to *f* (*Abies*) and the following zones **8** (*Pinus-Picea-Pterocarya*) and **9** (*Pinus-Juniperus*) (Text-figs 7, 18). The $\delta^{18}O$ values for the interglacial climatic optimum reach –7.0‰ in the second part of subzone *c*. In subzone *d* (*Abies*), $\delta^{18}O$ values increase to –6.8‰. An increased content of fir pollen is characteristic of a warmer and a more humid climate with distinct maritime features, which is reflected by the isotopic curve. Subzone *e* (*Quercus-Carpinus*) was characterised by lower precipitation and by slightly worse climatic conditions, which are confirmed by a low $\delta^{18}O$ value of –7.3‰. In the following subzone *f*, with fir pollen, a distinct decrease in $\delta^{18}O$ values to –8.0‰ occurred, recording a climatic crisis that is clearly visible in the isotopic curves (Text-fig. 18) and which was probably caused by the increased volcanic activity discussed in the chapter on the OS 1/96 core. As in the WL 1/96 and 1/97 cores, there was a break in deposition at that time, caused by the lake shallowing.

In pollen zone **8** (*Pinus-Picea-Pterocarya*), which followed the climatic optimum, a gradual increase in $\delta^{18}O$ values to a maximum of –7.2‰ is observed (Text-fig. 18). Later, a gradual decrease in $\delta^{18}O$ values to –7.8‰ was accompanied by worsening climatic conditions in pollen zone **9** (*Pinus-Juniperus* zone). This resulted in a more restricted area covered by stenothermal elements and in an expansion of the communities characteristic of a continental climate (BIŃKA & NITYCHORUK 1996). The continued climatic deterioration led to a distinct shallowing of the lake, recorded by the increased content of *Pediastrum.*

KA 1/93 – δ^{13}C

In subzones *c* and *d* of the *Carpinus-Abies* pollen zone **7** (Text-fig. 18) the δ^{13}C curve oscillates between –2.0 and –1.0‰. In subzone *e* there was a rise in δ^{13}C values to –0.2‰, which probably corresponded to a shallowing of the lake, with lower precipitation in the subzones with oak and hornbeam. A marked deflection of the δ^{13}C curve toward negative values in subzone *f*, accompanied by a drop in inorganic C in the sediment from 4% to 7% (Text-fig. 18) and by an increase in C_{org} (NITYCHORUK & *al.* 1995), was probably caused by the volcanic crisis. In the pollen zones after the climatic optimum a distinct cooling occurred, accompanied by shallowing of the lake and higher trophicity. The carbon isotopic curve had its maximum value of 0.8‰ at that time.

CLIMATIC CHANGES IN THE MAZOVIAN INTERGLACIAL BASED ON VARIATION OF OXYGEN AND CARBON STABLE ISOTOPES

The isotopic curves from the lake sediments of the Ossówka OS 1/96 core represent the oxygen and carbon isotopic ratios in the Mazovian Interglacial and during the long (dozens of thousands of years) initial phase of the Middle-Polish Glaciation. At Wilczyn, the isotopic curves represent the Mazovian Interglacial and a short post-interglacial period (Text-fig. 19). There is a distinct similarity in the curves from both lakes, despite some differences resulting from the varying character of the basins. At Ossówka, the lake basin was deeper and therefore more stable, as is indicated by the deposition of lake marls with annual lamination and by similar

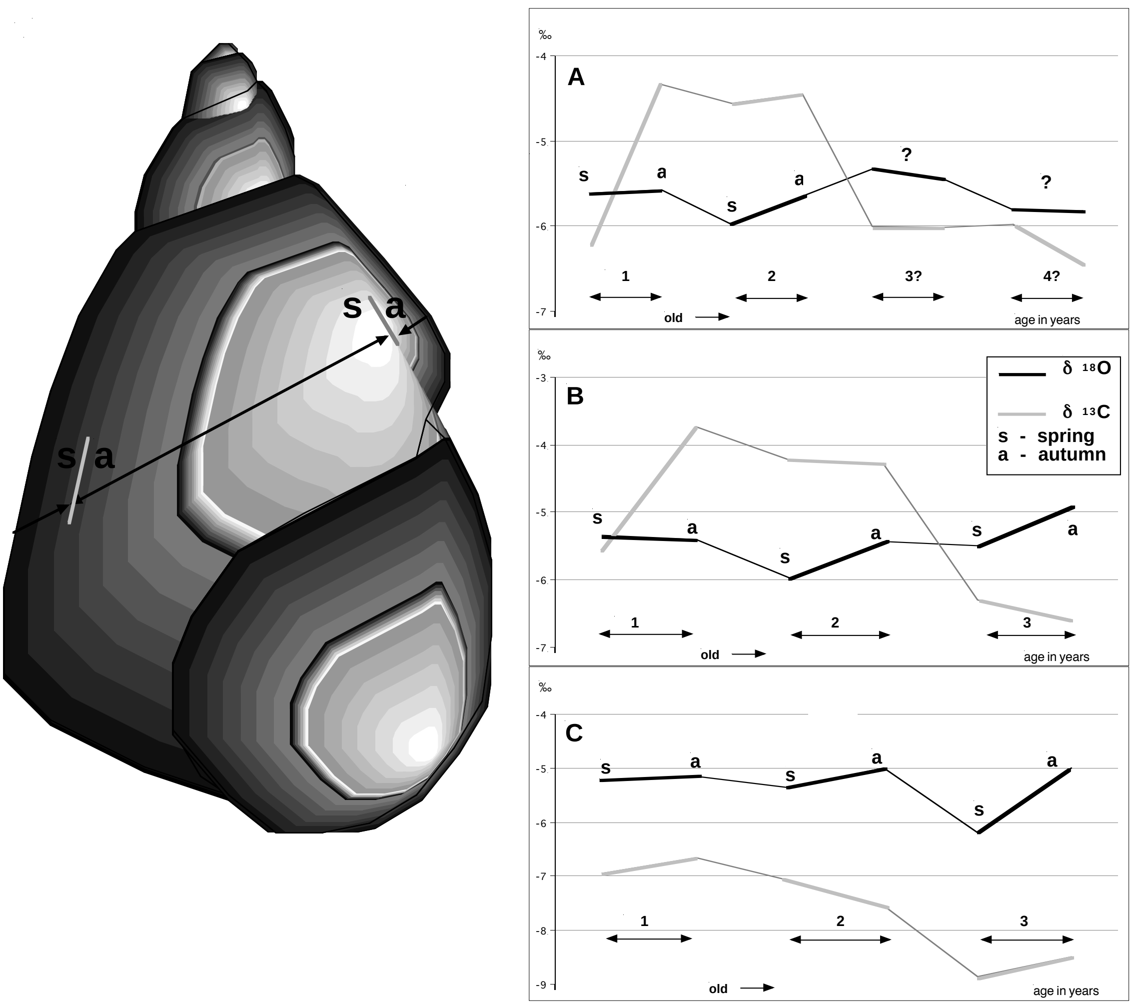

Fig. 17. Oxygen and carbon isotope for three shells (A – C) of *Viviparus diluvianus*; thickened fragments of curves correspond to annual shell growth: spring (s), and towards autumn (a); the almost identical curves for A and B suggest the same age with identical annual growth; specimen C with different annual growth also has different isotope ratios at the beginning and the end of the annual growth segment, and therefore lived at different time; question marks refer to uncertain choice of the correct growth lines in the opercular zone of the shell A

shapes of the $\delta^{18}O$ and $\delta^{13}C$ curves (Text-fig. 8). Good correlation of the $\delta^{18}O$ and $\delta^{13}C$ curves is a typical feature of closed calcareous lakes, while a lack of correlation indicates open lakes (EICHER & SIEGENTHALER 1976, FRITZ & *al.* 1975, TALBOT 1990). By extrapolation, good correlation of isotopic curves indicates a closed palaeolacustrine system. At Wilczyn, there was an overflow lake running from west to east (Text-figs 1, 4) over a distance of 30 km (NITYCHORUK 1994a). Differences between $\delta^{18}O$ and $\delta^{13}C$ values at Wilczyn were greater than at Ossówka at the beginning of the interglacial and at its optimum (about 2‰), which probably resulted from different lake types and the prevailing depositional conditions.

Rich pollen material and palaeoclimatic interpretation (BIŃKA 1994, KRUPIŃSKI 1995, NITYCHORUK & *al.* 1995, BIŃKA & NITYCHORUK 1995, 1996, BIŃKA & *al.* 1996, 1997, NITYCHORUK & *al.* 1999) enable correlation of the isotopic curves with certain climatic phenomena. Based on isotopic analysis of sediments and mollusc shells in relation to the climatic fluctuations inferred from the pollen spectra and the varying thickness of the annual laminae, 11 local isotopic horizons were distinguished for the Mazovian Interglacial in the OS 1/96 core (Text-fig. 20). Isotopic horizons and subhorizons were distinguished on the basis of variations in the oxygen and carbon isotope ratios, expressed as $\delta^{18}O$ and $\delta^{13}C$ values and illustrated by the corresponding isotopic curves. Distinct climatic phenomena were distinguished as separate isotopic horizons based on trends in the isotopic curves in various borehole successions and their correlation with the results of paleobotanical studies.

In addition, four periods of increased maritime influence were distinguished prior to the interglacial cli-

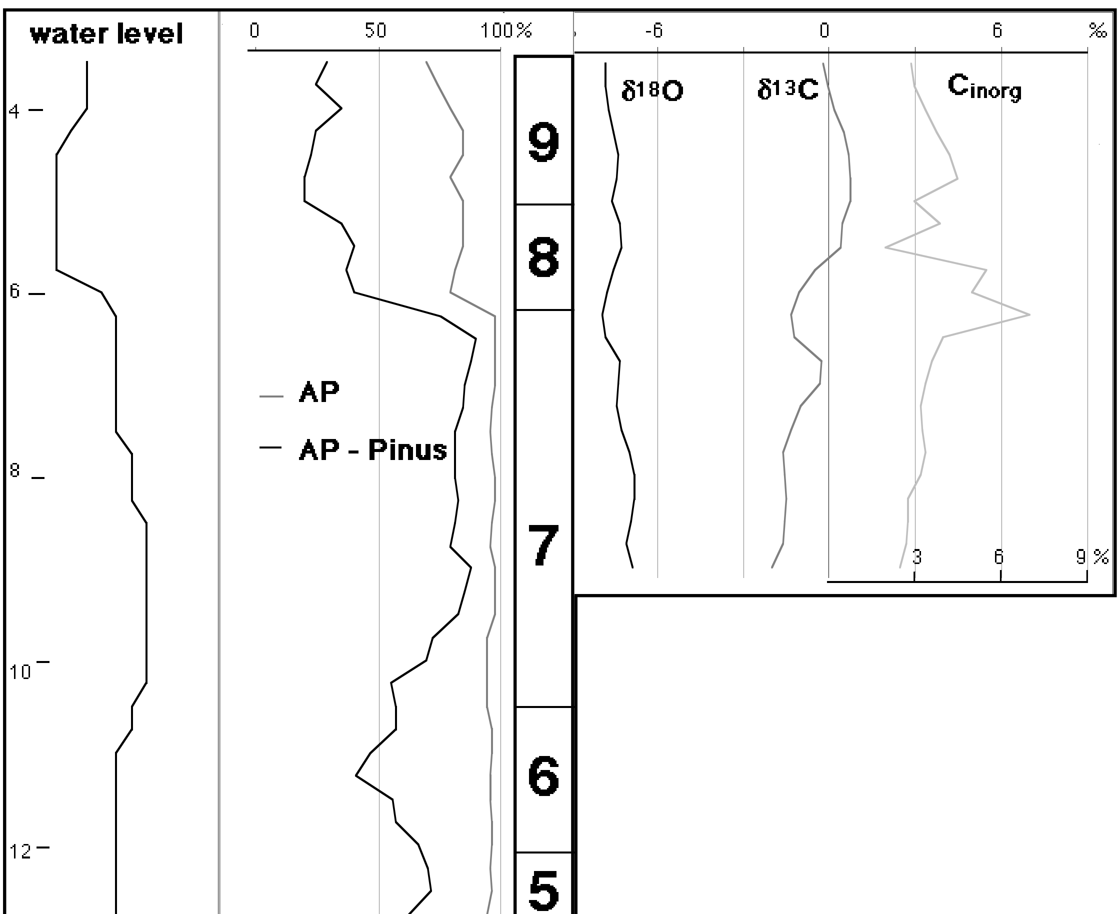

Fig. 18. Oxygen and carbon isotopes for calcareous lake deposits of the Kaliłów KA 1/93 borehole versus variation in inorganic carbon content (C_{inorg}); AP – tree pollen; AP minus *Pinus* – tree pollen withaut *Pinus* pollens; fluctuations of water level in the lake based on palynological studies (after NITYCHORUK & *al.* 1995)

matic optimum. These might have resulted from changing sea currents and/or from transgression of the Holsteinian sea. Moreover, the fluctuations of the water level of Ossówka Lake were estimated (Text-fig. 20).

Isotope horizon 1 – termination of the South-Polish Glaciation (depth 56.5-54.5 m)

The lowermost samples on which isotopic analyses were carried out were collected from the muddy-sandy deposits that accumulated directly after the retreat of the ice sheet of the South-Polish Glaciation (Text-fig. 20). Deposition of the 2 m thick series continued for at least 200 years in a rather dry and cool period with sub-arctic tundra. During summer the water level was higher, while the winters were dry. Poor plant cover led to increased washing of mineral material from the soil of the surrounding land into the lake.

The $\delta^{18}O$ curve runs smoothly between –5.0 and –6.0‰, as does the $\delta^{13}C$ curve between –0.5‰ and 0.0‰. Relatively high $\delta^{18}O$ and $\delta^{13}C$ values in a cool, sub-arctic climate correspond to the presence of marine carbonates from glacial deposits around the lake.

Isotope horizon 2 – beginning of the Mazovian Interglacial (54.5–47.0 m)

Isotope subhorizon 2 a (54.5-53.3 m)

The rapid change from glaciolacustrine to thinly laminated lake deposits (0.5 mm thick – Text-fig. 11) in this interval corresponds to a marked deflection in both the oxygen and carbon isotope curves. The $\delta^{18}O$ values increase from –6.2‰ to –3.5‰ and the $\delta^{13}C$ values decrease from 0.0‰ to –6.4‰ (Text-fig. 20). The time of deposition of this 0.7 m core length can be estimated as about 1500 years. The lake changed from a shallow post-glacial basin into a deeper and more stable one.

Changes in the water level at the beginning of the Mazovian Interglacial, based on examination of the diatoms (MARCINIAK 1998) differed according to the lake type. As at Ossówka, a high water level was also observed at Krępiec (comparable sized lake), while a low water level occurred in a small lake at Biała Podlaska.

The initial development of the basin at Ossówka was probably due to precipitation of waters rich in heavy oxygen isotopes from maritime air masses (*see* RÓŻAŃSKI & *al.* 1993).

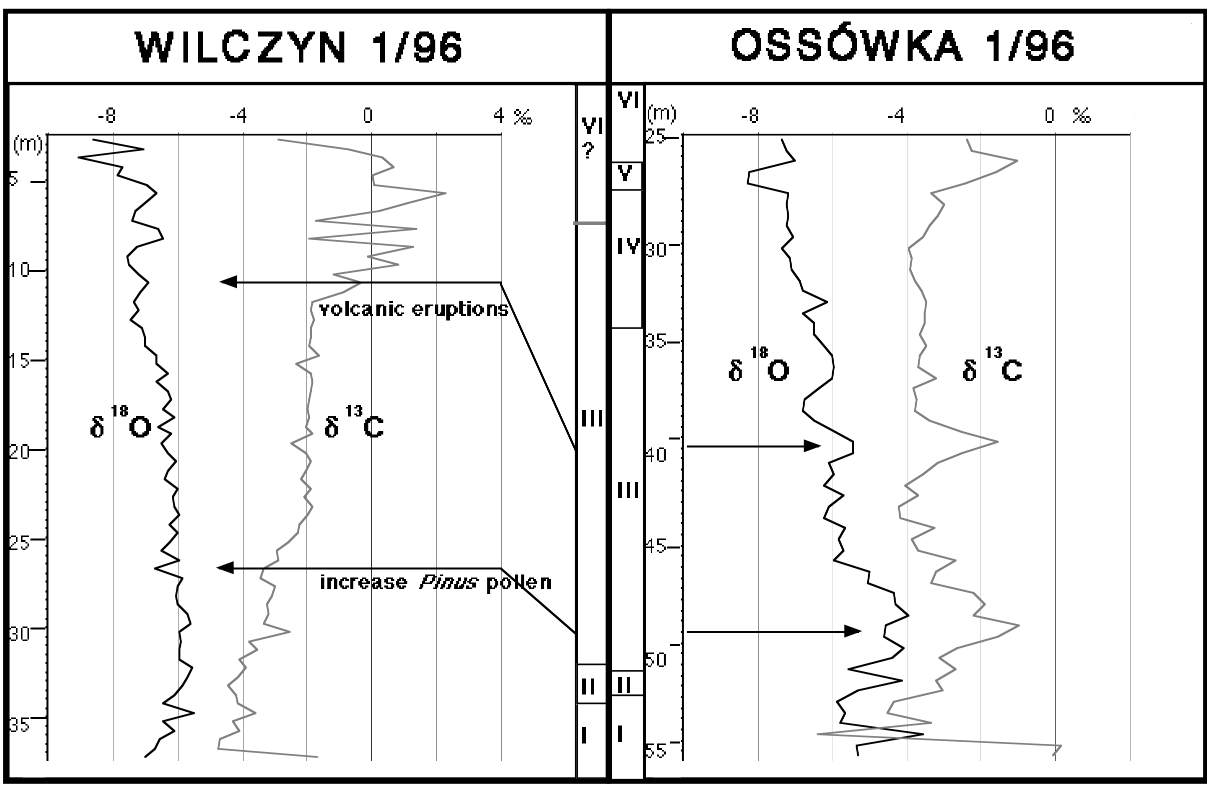

Fig. 19. Isotopes for the Wilczyn WL 1/96 and Ossówka OS 1/96 boreholes versus the palynological periods after KRUPIŃSKI (1995); arrows point to corresponding levels in the succession; in the lower part of the successions, a cooling phase is expressed by an increased content of pine pollen; in the upper part a cooling phase is expressed by a period of higher volcanic activity

Maritime air circulation was probably associated with warm sea currents, dependent on deglaciation in the North Atlantic. A simultaneous transgression of the Holsteinian sea could have occurred (HINSCH 1993, BIŃKA & *al.* 1997).

An improvement of hydrological conditions in the initial phase of the interglacial was not accompanied by a distinct warming (pollen subzone A/B – Text-fig. 6). A cool climate still predominated, expressed by the development of birch-pine forest. The $\delta^{13}C$ curve trended toward lower values (Text-fig. 20), probably indicating a climatic warming or the influx of ground or river water, together with low photosynthesis (EICHER & *al.* 1981, JĘDRYSEK & *al.* 1995).

A similar rapid increase in $\delta^{18}O$ values is observed at the boundary of the late Vistulian and the Holocene in the sections of Lake Gościąż (KUC & *al.* 1993, 1998, GOSLAR & *al.* 1995) and in the ice cores from Greenland (DANSGAARD & *al.* 1993). It was accompanied by a climate change (RALSKA-JASIEWICZOWA & GEEL 1993), change of depth of Lake Gościąż (SZEROCZYNSKA 1993, 1998a, b) and, as at Ossówka, by a reduction in thickness of the laminae (Text-fig. 11) that was also recorded in the Late-Glacial – Holocene sections from the lakes Attersee (SCHNEIDER & *al.* 1987, 1990), Meerfelder Maar (BRAUER & *al.* 1997) and Gościąż (GOSLAR 1998a, 1998b).

Isotope subhorizon 2 b (53.3-51.5 m)

Birch-pine forest was the dominating element, accompanied by continentalisation (pollen zone B – Text-fig. 6). It resulted in lower $\delta^{18}O$ values of about –5.8‰ (Text-fig. 20), probably associated with cooling. The absence of lamination in the sediments was probably due to lake shallowing (Text-fig. 20). This was also recorded by MARCINIAK (1998) in palaeolake deposits from the Mazovian Interglacial at Adamówka, Krępiec and Biała Podlaska. Shallowing also occurred in Lake Gościąż at the beginning of the Holocene (PAZDUR & *al.* 1995, STARKEL & *al.* 1998).

An increased content of sulphur to 1.45% is a very characteristic feature in deposits from this period (Text-fig. 8). Each of the following continental, dry and cool phases associated with lake shallowing is marked by increased contents of sulphur. This sulphur may be attributed to oxidation of FeS and H_2S (BASS-BECKING & KAPLAN 1956) as a result of lake shallowing.

In the Great Salt Lake in Utah (MCKENZIE & EBERLI 1987), there is a distinct relationship between a decrease in $\delta^{18}O$ values and carbonate content in the sediment and lake shallowing, i.e. between factors indicating a cooling phase.

Isotope subhorizon 2 c (51.5-50.0 m)

A change in sedimentation to calcareous gyttja (at 51.5 m) was followed by distinct annual lamination (Text-figs 11, 20). The July temperature rose from 14°C to 17°C, the organic production in the lake increased and was accompanied by a gradual increase in $\delta^{18}O$ and $\delta^{13}C$ values.

At 51.0 m, a rapid decrease in $\delta^{18}O$ values to –5.5‰ took place, coinciding with more sulphur in the sediments (Text-figs 8, 20), which may be correlated with shallowing and cooling. The next increase in $\delta^{18}O$ values coincided with an increased $CaCO_3$ content, resulting from intensive bioproduction in the lake associated with a rise in temperature. The latter was probably associated with the inflow of the maritime air masses postulated by BIŃKA (BIŃKA 1994, BIŃKA & NITYCHORUK 1995, 1996) in subzone *b* of the *Picea-Alnus* pollen zone, (Text-fig. 7). It coincided with an increase in ash and *Hedera* which, according to BIŃKA (BIŃKA & NITYCHORUK 1996), indicates a more humid climate. Another continentalisation of the climate is reflected in a decrease in $\delta^{18}O$ values of 1.5‰ (Text-fig. 20). An increased sulphur content in the sediments suggests yet another lake shallowing (Text-figs 8, 20).

A very similar sequence of climatic-hydrologic phenomena was noted in the palaeolake at Krępiec, where in the *Picea-Alnus* pollen zone (according to JANCZYK-KOPIKOWA 1981), shallowing occurred, followed by deepening and another shallowing MARCINIAK (1998).

The successive period in the climatic development during the Mazovian Interglacial is characterised by distinct climatic oceanisation, indicated by yew and *Hedera* (pollen zone E) reaching their maximum for the entire interglacial (Text-fig. 20). At Grazin (ERD 1969) transgressive deposits with marine diatoms occur in the horizon with yew. Similarly, a marine transgression was observed at Billbrook (HINSCH 1993, LINKE & HALLIK 1993), as well as in a small area in Lithuania and Latvia (KONDRATIENE & GUDELIS 1983). Maritime climate in eastern Poland is expressed by stable conditions of larger water basins (BIŃKA & NITYCHORUK 1996), supported by humid winters and summers (KRUPIŃSKI 1995).

Small lake basins, like the one at Biała Podlaska, despite a favourable climate, were subjected to shallowing (MARCINIAK 1998). This could have been caused by the surface supply of these basins receiving less water due to the development of plant cover resulting in higher water retention in the surrounding area.

The isotopic curves indicate stable climatic conditions at that time and reach high values (Text-fig. 20). Air masses connected with oceanisation were able to reach far to the east and therefore $\delta^{18}O$ values were the high-

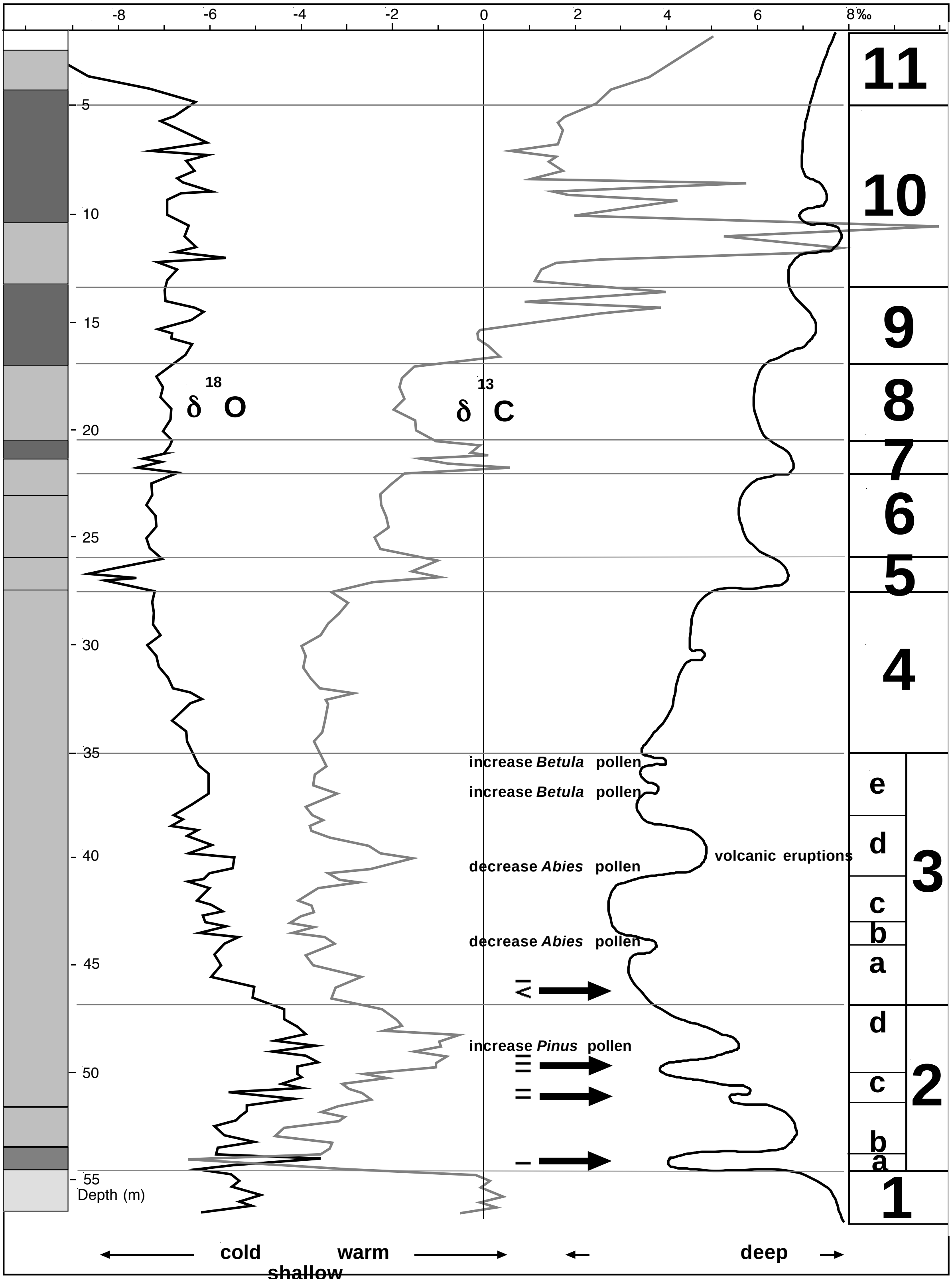

Fig. 20. Isotope horizons (1,2...) and sub-horizons (a,b...) of the Mazovian Interglacial and the initial part of the Middle Polish Glaciation, versus probable changes of water level in the palaeolake and isotopes for calcareous sediments in the Ossówka OS 1/96 borehole; arrows indicate increased marine influence; for lithological explanations see Text-fig. 8

est for the entire interglacial (*see* RÓŻAŃSKI & *al.* 1993). According to the pollen studies, the yew period was not the warmest during the Mazovian Interglacial (MÜLLER 1974, BIŃKA & NITYCHORUK 1996). The lake at that time was oligo- or mesotrophic (BIŃKA 1994, MARCINIAK & LINDNER 1995).

Isotope subhorizon 2d (50.0-47.0 m)

A considerable influence of continental climate (pollen zone F) was followed by a maritime climate. Dry winters caused a general drop of water level, which was also enhanced by the increasing summer temperatures.

This period was characterised by fluctuations of the isotopic curves (Text-fig. 20) over about 1‰, although values remained high. $\delta^{18}O$ values reached a maximum value of –0.5‰ at that time, similar to the values reached during the transition between the South-Polish Glaciation and the Mazovian Interglacial, when a dry and cool climate prevailed. An increased sulphur content in the sediment indicates a drop of water level in the lake (Text-figs 8, 20).

Continentalisation increased and terminated in a distinct cooling phase (no. 1 on Text-fig. 8). This cooling phase is clearly visible on numerous pollen diagrams from the Mazovian Interglacial (Holsteinian) in Europe (MÜLLER 1974). In eastern Poland such cold conditions were expressed by an increased content of pine and birch pollen and by a drastic drop in the pollen of the other trees, especially the stenothermal genera (BIŃKA 1994, BIŃKA & NITYCHORUK 1995, 1996 – Text-fig. 7; KRUPIŃSKI 1995 – Text-fig. 6). A few small lakes in this lake district dried out completely (NITYCHORUK 1994a) and clastic sediments occurred in other lakes, indicating shallowing and redeposition. The palaeolakes at Ossówka and Wilczyn survived this period well. Short-term breaks in lamination (Text-figs 11, 12) and a lowered $CaCO_3$ content in the gyttjas (from 70 to 60%) at Ossówka, are the only signs of shallowing. Decreased isotopic values, which reached over 1‰ at Ossówka and around 0.5‰ at Wilczyn (Text-figs 8, 14), are another indication of a cooling phase.

Isotope horizon 3 – climatic optimum of the Mazovian Interglacial (47.0-35.0 m)

Isotope subhorizon 3a (47.0-44.0 m)

The warmest phase during the entire interglacial followed the above-described climatic crisis (zones G and H according to KRUPIŃSKI 1995 – Text-fig. 6). According to BIŃKA (1994 – Text-fig. 7) this part of the succession can be considered as the beginning of the interglacial climatic optimum. Warming and oceanisation of the climate coincided with a number of important features in the sediments:

— decreased oxygen isotope values, caused by lake deepening (Text-fig. 20);
— an increased content of $CaCO_3$ (on average 70 – 80%), as an effect of higher bioproduction in the lake (Text-fig. 8);
— a thickening of laminae from 0.8 to 1.2 mm due to intensive precipitation of calcium carbonate and increased deposition of organic matter (Text-fig. 11);
— higher $\delta^{18}O$ values in the dark laminae than in the light laminae, accompanied by oceanisation, a pattern opposite to that prevailing since the beginning of the interglacial (Text-fig. 12).

In the pollen analysis, stenothermal trees such as hornbeam, oak, fir and hazel dominate, and *Viscum* and *Buxus* are also indicators of a warmer climate (BIŃKA 1994 – Text-fig. 7; KRUPIŃSKI 1995 – Text-fig. 6). A development of fir, in particular, is a good indicator of warming and associated oceanisation of climate. Transgressive deposits of the Holsteinian sea, dated to the beginning of the interglacial climatic optimum, have been recorded in numerous sections in Germany (LINKE & HALLIK 1993, HINSCH 1993). A humid, warm and mild climate favoured a long vegetation season (KRUPIŃSKI 1995). Large lakes reached the greatest depths at that time (Text-fig. 20). A small lake at Biała Podlaska became shallower (MARCINIAK 1998), probably due to infilling with sediments. In the rather uniform pollen zones G and H there were stable climatic conditions, and $\delta^{18}O$ values ranged over about 1.5‰, reaching from –5.2 to –6.8‰.

Isotope subhorizon 3b (44.0-43.0 m)

The first distinct change in the oxygen isotopic curve (at 43.5 m) coincided with the first minimum of fir pollen, accompanied by a short-term increase of the thickness of the laminae in the sediments (Text-figs 8, 11, 12). A retrieval of fir coincided with a lower water level caused by continentalisation (BIŃKA 1994). This fall in water level led to a slight shallowing of the lake, accompanied by a rise in ^{18}O. A continental climate resulted subsequently in precipitation with lower ^{18}O values and/or in a slight cooling. All this resulted in an increase, followed by a decrease, in $\delta^{18}O$ values (Text-fig. 20).

Isotope subhorizon 3c (43.0-41.0 m)

Another increase in fir pollen results from a warming, although the $\delta^{18}O$ values slightly decreased. The $\delta^{13}C$ values also decreased and the water level in the lake could therefore have been higher (Text-fig. 20).

Isotope subhorizon 3d (41.0-38.0 m)

The second marked deflection in the oxygen and carbon isotopic curves during the interglacial climatic optimum coincided with a minimum of fir pollen and is much more distinct than the previous one. It starts at about 40 m (Text-fig. 20). The crisis observed at Ossówka, Wilczyn and Kaliłów was relatively long (about 2000 – 3000 years) and severe ($\delta^{18}O$ values decreased by more than 1‰; Text-figs 8, 14, 18). $\delta^{13}C$ values increased by 2‰. Additionally, an increased content of sulphur in sediments may indicate a considerable shallowing of the lake (Text-figs 8, 20). As shown above, the data from pollen zone H, subzones 2-5 (according to KRUPIŃSKI 1995 – Text-fig. 6) and pollen zone **7** (*Carpinus-Abies*), subzones *e-f* (according to BIŃKA & NITYCHORUK 1996 – Text-fig. 7) suggest a climatic crisis caused by volcanic activity. The influence of the latter on climate in this part of Europe was significant, and is indicated by depositional hiatuses in lake sediments in Poland (BIŃKA & NITYCHORUK 1995, 1996), Germany (MÜLLER 1974) and Belarus (PAVLOVSKAYA & NITYCHORUK *in press*). Stratigraphical gaps in the Mazovian Interglacial lake sediments at several localities are confirmed (NITYCHORUK 1994a) by pollen examination of the succession at Ossówka (KRUPIŃSKI 1995), which probably represents the deposition throughout the Mazovian Interglacial. According to KRUPIŃSKI (1995), the completeness of the palynological diagram for this locality enabled the palynological zonation of the Mazovian Interglacial to be established.

Isotope subhorizon 3e (38.0-35.0 m)

In the late Mazovian Interglacial climatic optimum the influence of continental climate was expressed by fluctuations of the isotopic ratios, which are well recorded in the OS 1/90 core (Text-fig. 13) and coincided with a rise in birch pollen.

A few percent increase of *Pediastrum* at this time suggests a short-term slight shallowing of the lake, accompanied by higher trophicity caused by cooling (Text-fig. 20). The average temperature could have been 1-2°C, contributing to a gradual retrieval of the stenothermal elements of the flora.

Isotope horizon 4 – the late Mazovian Interglacial (depth 35.0-27.3 m).

A gradual decrease in $\delta^{18}O$ values and an increase in $\delta^{13}C$ values resulted from a gradual transition of the interglacial climate (pollen period **III** – Text-fig. 6) into a cooler post-interglacial one (pollen period **IV**). It was accompanied by another cooling phase (no. 2 on Text-fig. 8), which coincided with a decrease in stenothermal tree pollen (KRUPIŃSKI 1995) and with the predominance of pine (up to 90% in pollen zone J – Text-fig. 8). A boreal climate with continental features was characterised by a slightly lower July temperature falling from 20 to 17°C and finally to 14°C. Analysis of the oxygen isotopic ratios for the final part of the interglacial indicates a rapid cooling (Text-figs 8, 13).

The isotopic values in the annual laminae maintained the distribution pattern typical of the interglacial climatic optimum (Text-fig. 12), confirming this hypothesis. Higher oxygen isotopic ratios in the light laminae than in the dark laminae coincided with gradual cooling, drying and continentalisation 2000 – 3000 years after the interglacial climatic optimum. Continued shallowing of the lake is reflected by an increased thickness of annual lamination in the sediments (Text-fig. 11). A rapid decrease in $CaCO_3$ content from 70 to 20% in the light laminae also occurred (Text-fig. 12).

Isotope horizon 5 – first cooling phase at the beginning of the Middle-Polish Glaciation (27.3-26.0 m)

A distinct cooling phase at this horizon marks the beginning of the Middle-Polish Glaciation, at 27.5 m in the upper part of the Ossówka OS 1/96 core (Text-fig. 6). The climate was not uniform. The initial part of the glaciation was marked by 5 separate cooling phases (3 to 7 on Text-fig. 8), during which the climate changed into a subarctic one. All of these cooling phases show similar pollen spectra. Pollen analysis of the OS 1/90 core (pollen periods **V, VII, IX, XI**, there are no deposits representing the fifth cooling phase) showed that an increase in birch, juniper, larch and grass pollen was accompanied by a decrease in the previously dominant pine pollen (KRUPIŃSKI 1995 – Text-fig. 6).

The occurrence of juniper pollen, in particular, indicates a less dense forest and a subarctic climate. The post-interglacial cooling phases, pollen periods **V** (zones: K, L, M) and **VII** (zones: O, P, Q) KRUPIŃSKI (1995) on the Ossówka pollen diagrams, were characterised by varying humidity and thermal conditions. The most severe continental climates, with long and temperately humid winters and with dry and short-lasting vegetation seasons, were represented by pollen zones K and O. During the third subarctic cooling phase $\delta^{18}O$ values (Text-fig. 8) decreased to –8.5‰ (Text-figs 8, 20) and subsequently increased to the values from before the cooling phase, although cool conditions still prevailed in zones L and P, and a distinct warming phase occurred in zones M and Q.

A decrease in $\delta^{18}O$ values was accompanied by a decrease in the content of $CaCO_3$ in the sediment from 60 – 80% to 40 – 50%, but, as in case of the isotopic values, the calcium carbonate content already rose in zones L and P. A decrease in $\delta^{18}O$ values and in the carbonate content in pollen zones K and O can be interpreted by reduced bioproduction in the lake and by transport of clastic material at the less dense plant cover, and the probable presence of permafrost in these zones (KRUPIŃSKI 1995). If a seasonal frost did exist at that time, it could have considerably influenced groundwater movement and the transport to the lake of the carbonates, from which the lake marls and calcareous gyttja were derived.

The restricted DIC supply to the lake caused an increase in $\delta^{13}C$ values. The limited influx of groundwater contributed to the water level lowering in the lake (Text-fig. 20).

The change in colour of the gyttja from grey into dark-grey was caused by the increased content of organic matter and sulphur in the sediments (Text-fig. 8), probably associated with lake shallowing.

Isotope horizon 6 – warming phase (26.0-22.0 m)

The return of a boreal climate caused an increase in $\delta^{18}O$ values to the previous level i.e. –7.2‰. The content of calcium carbonate and of tree pollen also increased, which confirms a warming phase. A deepening of the lake occurred (Text-fig. 20).

Isotope horizon 7 – second cooling phase (22.0-20.5 m)

The second cooling phase (no. 4 on Text-fig. 8) was associated with a marked negative deflection in the $\delta^{18}O$ curve. A decrease in $\delta^{18}O$ values to –7.5‰ was accompanied by lower evaporation under cool climatic conditions. In the post-interglacial period (Text-fig. 20) the $\delta^{13}C$ curve shows a reverse trend to that of the $\delta^{18}O$ curve. This pattern, which is the opposite of that obtaining during the interglacial, was probably associated with concentration of organic matter due to considerable shallowing of the lake and a long retention period, which favoured an increase in $\delta^{13}C$ values (JĘDRYSEK 1994, JĘDRYSEK & *al.* 1994, 1995).

Isotope horizon 8 – warming phase (20.5-17.0 m)

A return to a continental climate with marked temperature differences between the warmest and coldest months of the warmest and the coldest months and a distinct aridity took place (KRUPIŃSKI 1995), and was probably accompanied by a slight deepening of the lake (Text-fig. 20).

Isotope horizon 9 – third cooling phase (17.0-13.0 m)

Drying could have led to a considerable shallowing of the lake (KRUPIŃSKI 1995).

Isotopic horizon 10 – a non-subdivided warming phase between the fourth and fifth cooling phases (13.0-5.0 m)

The lake still existed after the third cooling phase (no. 5 on Text-fig. 8), but only as a relict one. This is shown by the complete absence of correlation between the oxygen and carbon isotopic curves and by the extreme values attained. $\delta^{18}O$ values increased and reached –5.6‰, which is similar to the value for the interglacial climatic optimum (Text-fig. 20). The $\delta^{13}C$ values reached as high as 10.0‰. Such high $\delta^{13}C$ values could have been due to considerable lake shallowing as a result of intensive evaporation (*see* GONFIANTINI 1986, HOEFS 1996) in a very dry climate with steppe features. The areas occupied by the lakes were therefore reduced, which in turn contributed to the emergence of littoral sediments along the shore. These were deposited during the interglacial climatic optimum and contain mollusc shells. The latter were derived from interglacial deposits and were redeposited in a cool post-interglacial lake (Text-fig. 10).

Isotope horizon 11 – sixth cooling phase (5.0-2.5 m)

The last cooling phase was the most severe and was associated with ice sheet advance. This is shown by the disappearance of calcareous gyttja, changes in the plant cover and by $\delta^{18}O$ values reaching –10.0‰ (Text-fig. 20).

FINAL REMARKS

Interpretation of the oxygen and carbon isotopic curves from the cores of the palaeolake deposits in eastern Poland was carried out in order to reconstruct the climate during the Mazovian Interglacial (Holsteinian) and the initial part of the following Middle-Polish Glaciation (Saalian). The results of this interpretation correspond to the palaeobotanical ones. The principal difference is that the $\delta^{18}O$ values during the interglacial climatic optimum inferred from pollen analysis, reached lower values than during the beginning of the interglacial. The isotopic values also do not reflect the existing climate in the cool period at the beginning of the glaciation, because they are high there (Text-figs 8, 20). As already discussed, these differences result mainly from variations in the water level in the lake and from the transport of allochthonous calcite. The carbon isotopic curve, in particular, may reflect changes of the ratio between the water layer in the epilimnion versus that in the hypolimnion E/H. It results from oscillation of lake water level in the case of short-term climatic changes, and depends mainly on infilling of the lake basin with sediments, which is associated with higher $\delta^{13}C$ values in the upper part of the succession (*see* RÓŻAŃSKI & *al.* 1998).

Each cooling phase is associated with lower $\delta^{18}O$ values, less carbonate in the sediments and lake shallowing (*see* MCKENZIE & EBERLI 1987). In the lakes of eastern Poland this cooling has to be identified with continentalisation, when summers were warmer and winters more severe. The continentalisation resulted in general drying, which had a direct influence on the shallowing of the larger lakes and was associated with precipitation with lower isotopic values (*see* RÓŻAŃSKI & *al.* 1993).

The isotopic curves obtained for the various lakes have different shapes, which results from the different prevailing hydrological regimes and from varying hydrological conditions, lake size and depth. A large number of factors affecting sedimentation in the lakes, which influence the isotopic values, make the comparison of isotopic curves for lakes and oceans even more difficult or even impossible. Moreover, the author does not attempt to identify which deep-sea oxygen isotope stage corresponds to the Mazovian Interglacial (Holsteinian) in the sections of lake deposits in eastern Poland.

The opportunity of making such a correlation does not lie directly in comparing identically shaped isotopic curves, but rather in recognising the existence and intensity of the observed climatic changes. If the cooling that resulted from volcanic activity during the late Mazovian Interglacial climatic optimum occurred over a large area, its influence on other isotopic curves, including those of deep-sea sediments, should be examined.

However, since deep-sea isotopic curves are relatively general, with the data representing relatively broad intervals, the detection of short-term climatic episodes, such as the volcanic crisis discussed in this paper, is difficult.

Correlation of the Mazovian Interglacial to a deep-sea oxygen isotope stage was not, however, the main aim of this study. It was much more important to draw attention to the new opportunities that result from the interpretation of stable oxygen and carbon isotopes in the lake sediments. In numerous cases, the isotopes were found to be more precise tools for recording palaeoenvironmental changes than the palaeobotanical and palaeozoological changes, since the latter exhibit a certain inertia caused by the slow reaction of plants and animals to climatic changes.

Isotopic studies of single non-glacial lamina in complete cores of lake deposits is a very promising technique. Interpretation of isotope values separately for the dark and light laminae, and their representation by separate curves, appears to be a sensitive indicator of oceanisation and continentalisation.

Isotopic determinations of annual growth segments of large water snails can provide similar data regarding continentalisation and oceanisation.

The volcanic activity documented by deflections in isotopic curves and by analysis of thin sections enables a new look at the problem of rapid climatic changes. A cooling phase caused by volcanic activity in the Mazovian Interglacial resulted in the disappearance of most of the lakes in eastern Poland. Cooling and continentalisation, accompanied by probable pollution of lake water by acid rain containing sulphur of volcanic origin, could have resulted in the extinction of snails such as *Viviparus diluvianus* KUNTH and *Lithoglyphus piramidatus* MILDF.

CONCLUSIONS

Climatic conclusions

The interpretation of the isotopic curves obtained from the lake sediments were based on a simple rule i.e. a trend toward more negative values indicates a cooling phase, and a trend towards more toward more positive values indicates a warming phase

The material presented in this study indicates that palaeoclimatic changes can be identified by means of isotopic curves from lake sediments, even where these curves are complex. The quality of such interpretations can be significantly improved by correlation of the isotopic data with a pollen diagram, particularly if the latter has been prepared from the same core. The contents of carbonates and of some trace elements also seems to very significant.

In general, the isotopic curves for palaeolake sediments from the Mazovian Interglacial in eastern Poland clearly document the climatic changes shown in pollen diagrams. There are, however, two levels in the succession where the isotopic data provide apparently inconsistent indications. During the Mazovian Interglacial climatic optimum, the isotopic curves appear to indicate relatively cool climatic conditions. This contradictory effect was probably caused by increased precipitation leading to lake deepening by raised level of lake waters and/or by influx of groundwater enriched in light isotopes.

In the upper part of the successions, corresponding to the initial phase of the Middle-Polish Glaciation, the isotopic values reach unexpectedly high levels. This was probably associated with lake shallowing and with enrichment in heavy isotopes through evaporation, under conditions of a cool steppe climate, and/or with the redeposition of re-emerged "warm" interglacial deposits from the near-shore zones.

The isotopic curves clearly illustrate maritime and continental air influence during the Mazovian Interglacial. Before the interglacial climatic optimum, the lakes were oligo or mesotrophic, with dominating influence of continental air circulation.

In non-glacial laminae from the lower part of the succession, when a cooler climate with continental features, dry summers and humid winters prevailed, the $\delta^{18}O$ values are higher for the light laminae (spring-summer) than for the dark laminae (autumn-winter). The isotopic data from the laminated sediments formed under the maritime climatic conditions of the interglacial optimum, with humid summers and mild, dry winters, exhibit the reverse relationship, i.e. the light laminae have lower $\delta^{18}O$ values than the laminae, as in modern lakes.

The influence of maritime climate dominated in the interglacial climatic optimum, and the lakes became more eutrophic. The lakes were largest and deepest at that time.

The extinction of molluscs is connected with the interglacial climatic optimum. Two particularly important species, *Viviparus diluvianus* KUNTH and *Lithoglyphus pyramidatus* MILDF., are characteristic of the Mazovian Interglacial (known also from *Paludinal* Bank). *Viviparus diluvianus* occurs only in the warmest part of the Mazovian Interglacial; its extinction was probably caused by a rapid change of climate (NITYCHORUK 1994a, 1995, NITYCHORUK & *al.* 1995), which could have resulted from volcanic activity and the related acid rainfall.

The volcanic activity had a crucial influence on lacustrine sedimentation and climate during the Mazovian Interglacial. The cooling caused by pollution of the atmosphere by volcanic ash in the second half of the interglacial climatic optimum resulted in lake shallowing, sedimentary hiatuses or the disappearances of lakes. The isotopes appeared to be a sensitive indicator of the climatic change caused by volcanic activity.

Cyclic climatic cooling phases in the post-interglacial period are recorded by marked negative deflections in the $\delta^{18}O$ curves and by even more marked positive deflections in the $\delta^{13}C$ curves, features that are associated with lake shallowing. In the initial phase of the following glaciation, a continental climate predominated, and simultaneously the lake became smaller, leading to an increased content of organic matter and heavy isotopes.

Methodological conclusions

The isotope curves from lake sediments indicate an earlier and more spontaneous response to climatic fluctuations and changes in the sedimentary environment than do the pollen diagrams. The interpretation of the isotopic curves commonly provided better information than the pollen diagrams, especially with respect to climatic changes, or extreme phenomena such as volcanic activity.

Therefore, isotopic studies constitute an important tool for paleoenvironmental and paleoclimatic interpretation.

The influence of maritime air circulation is associated with precipitation characterised by higher oxygen isotopic values. Precipitation connected with continental circulation is characterised by lower oxygen isotopic values (RÓŻAŃSKI & *al.* 1993). In addition to the isotopically heavy rainfall, a milder climate connected with an increased maritime influence, should be reflected in a trend of the isotopic curve toward higher values. However, a maritime climate is also reflected in higher humidity and lower evaporation. Together with more intensive precipitation, these factors result in deepening of lakes, especially the closed ones. Considerable lake shallowing may result in a lower content of heavy isotopes and in a shift of oxygen isotopic curves toward negative values, which was a common feature during the Mazovian Interglacial climatic optimum.

The isotopic concentrations in lacustrine mollusc shells play an important role in studies of palaeolake districts. However, good results can be obtained only for the *in situ* fauna. The isotopic concentrations determined for large snails need to provide a mean representative sample. Otherwise, the value represents a given annual season and cannot be compared with the average for a sedimentary period of a few years duration. Different mollusc species need to be interpreted separately (*see* FRITZ & POPLAWSKI 1974).

The small variability in isotopic ratios in deposits from the littoral zone and the existence of common sedimentary hiatuses result in incomplete isotopic data and these are therefore of little use for paleoclimatic and paleoenvironmental interpretation.

The $\delta^{13}C$ curve presents various fluctuations. These coincide with lake shallowing. Generally, $\delta^{13}C$ values increase up-section. This is connected with deposition and gradual shallowing of the lake (*see* RÓŻAŃSKI & *al.* 1998).

The oxygen and carbon isotopic ratios for single laminae are good indicators of changes in the climatic conditions. The isotopic curves from the light and dark laminae indicate a continental air circulation, or the temperature-humidity relationship in the palaeoenvironment.

Annual growth segments of shells reflect a growth period beginning in the spring and ending in the autumn. They provide information on isotopic ratios in the vegetation season during a year.

The oxygen and carbon isotopic curves are different (Text-fig. 20). Four types of correlation have been distinguished:

1. Positive correlation is typical of closed lake basins (EICHER & SIEGENTHALER 1976, FRITZ & *al.* 1975, TALBOT 1990). It occurs under generally stable hydrologic and climatic conditions, without relationship to the prevailing cold or warm climate. This is the most common type in the OS 1/96 core.
2. Negative correlation indicates open lakes (EICHER & SIEGENTHALER 1976, FRITZ & *al.* 1975, TALBOT 1990).
   a) $\delta^{18}O$ trends towards higher values, while $\delta^{13}C$ trends toward lower ones. Such a case occurs during initial warming of climate. In the OS 1/96 core (Text-fig. 8) this is seen at the transition from the glaciation to the interglacial. A similar negative correlation occurs in the Holocene deposits of Lake Gościąż at the Younger Dryas/Preboreal boundary (KUC & *al.* 1993, 1998), and at the Allerod/Younger Dryas boundary (KUC & *al.* 1998).
   b) $\delta^{18}O$ is characterised by lower values and $\delta^{13}C$ by higher values. This is the result of considerable cooling connected with lake shallowing. In the OS 1/96 core a cooling phase occurred in the initial phase of the Middle-Polish Glaciation.
3. The curves are chaotic and cannot be correlated. The isotopic values are excessive with respect to climatic conditions. Such a situation appears in a shallow basin with very unstable hydrologic conditions, considerable changes in water level and, in extreme cases, a fall.

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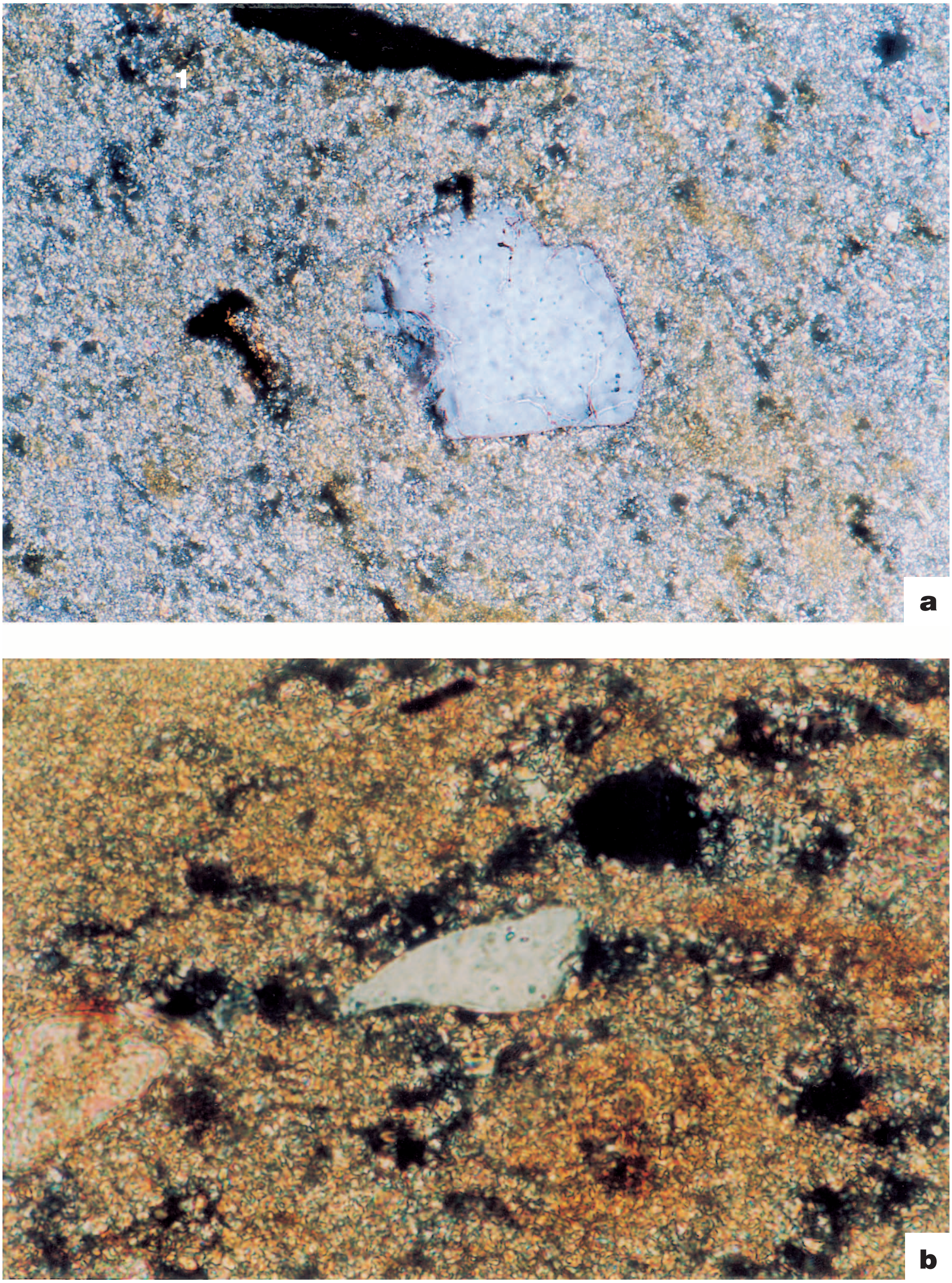
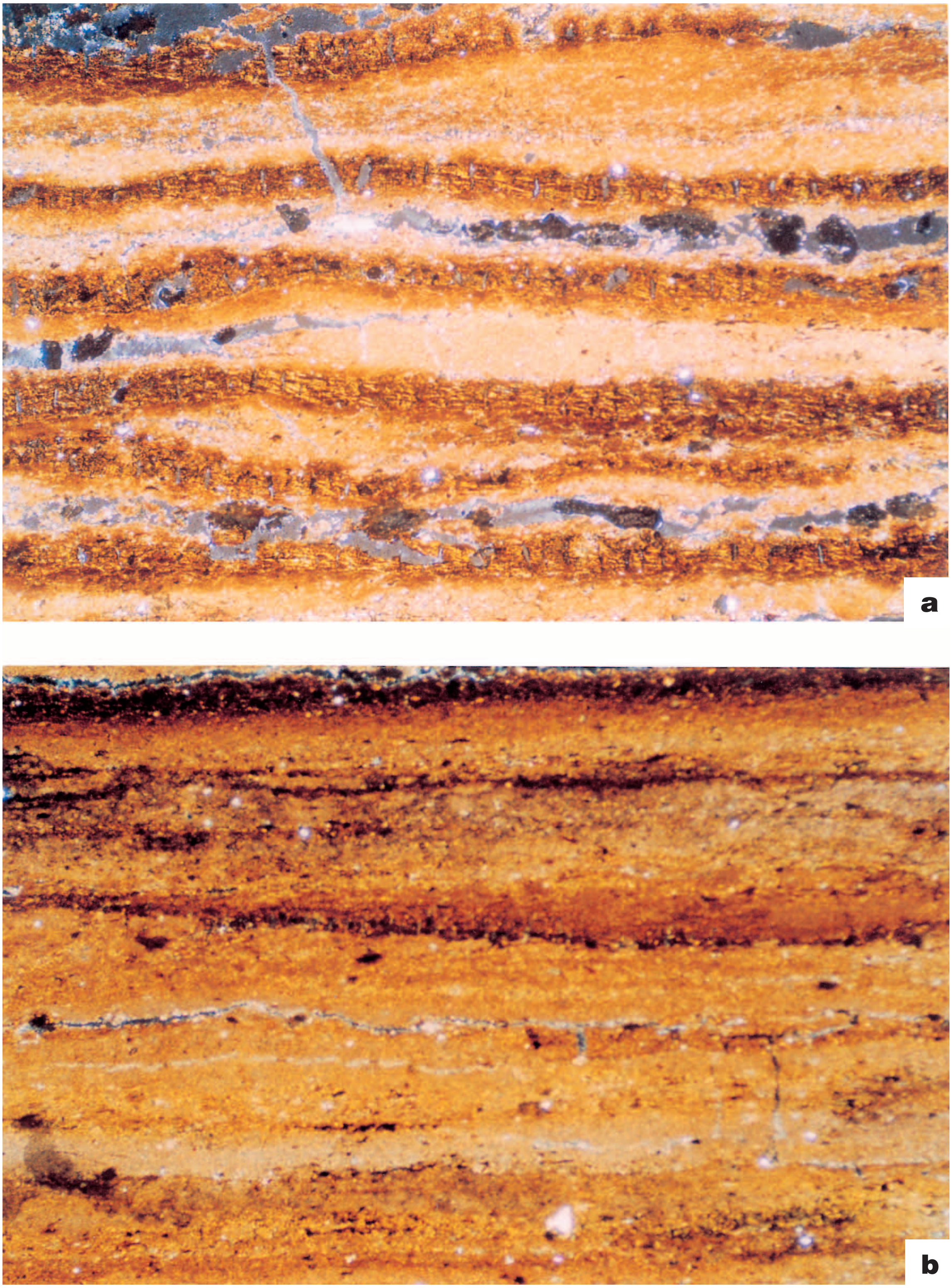


Angular quartz grains of volcanic origin in lake deposits of the Mazovian Interglacial in the Ossówka OS 1/96 borehole, a – at 38.2 m, b – at 400 m; × 560



Laminated lake deposits of the Mazovian Interglacial in the Ossówka OS 1/96 borehole, a – 54.3 m; b – 45.5 m; × 50