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Timing of the final disappearance of permafrost in the central European Lowland, as reconstructed from the evolution of lakes in N Poland

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Accumulation of sediments in the lake basins of the Starogard Lakeland, Northern Poland, an area which was entirely ice-covered during the last glaciation, started at different times, beginning during the Late Glacial. Sedimentation continued till the beginning of the Holocene (Preboreal). The principal factor causing the asynchronous start of the lake development was the variation in melting processes of buried dead-ice blocks. The preservation of dead-ice masses in some depressions until the Preboreal leads to the conclusion that the ultimate disappearance of permafrost in the study area occurred only at the beginning of the Holocene.

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INTRODUCTION

During the Late Weichselian transgression of the ice sheet towards the line of the Last Glacial Maximum (LGM), the climate was harsh. It is generally accepted that almost the entire area that nowadays is Polish territory outside the LGM then was covered by continuous permafrost, with average annual temperatures from –7°C in the west to –9°C in the east (Kasprzak, 2003; Mojski, 2005). The periglacial conditions significantly transformed the relief of these areas, which – despite the time passed since, and also despite the growing influence of Man – still show numerous characteristic periglacial features (Starkel, 2005).

Much discussion goes on, however, about the possible presence of permafrost in the area of Poland that was covered by ice during the last glaciation. The discussion concerns both the possibility of permafrost being preserved under the ice sheet during ice advance (anaglacial phase) and the possible development of "fresh" permafrost in areas undergoing deglaciation (cataglacial phase). Important for this discussion was a model presented by Kozarski (1995), which takes into account the advance of permafrost to the area just in front of the retreating ice sheet, where simultaneously the older, more deeply seated, permafrost disappeared. This model, with some adaptations, was also considered applicable to the deglaciated areas of Northern Germany (Liedtke, 1993; Böse, 1995; Vaikmäe *et al*., 1995).

The results of recent studies from several areas covered by an ice sheet during the last glaciation, especially the Laurentian area in North America, suggest that permafrost under the ice sheet did not disappear completely. It is also likely that permafrost was limited to a small area within the range of subglacial water circulation (French and Harry 1990; Clayton *et al*., 2001; Lacelle *et al*., 2004). Similar considerations on the possibility of partial preservation of permafrost under the ice sheet during its advance refer to the deglaciated area in Poland. In this context, permafrost under the ice sheet may have played an important role in the development of the subglacial drainage system and the formation of subglacial channels (Piotrowski *et al*., 2009).

Consequently, a significant relationship exists between the successive stages of the ice sheet's retreat and the developmental stages of periglacial conditions. Periglacial transformation of the relief is least evident in the areas which were the last ones that became exposed again after retreat of the ice sheet (Liedtke, 1993; Kozarski, 1995). It is commonly accepted that continuous permafrost existed in most parts of the deglaciated area of Poland, and that permafrost was discontinuous only north of the maximum extent of the ice sheet of the Pomeranian Phase (Kozarski, 1995). The main controversy in the specialist literature concerns the time of the complete disappearance of permafrost in the central European Lowland. This took place during the Bølling-Allerød interval according to Liedtke (1993), Böse (1995), Kozarski (1995) and Marks (1996), but others suggest the Preboreal (Gołębiewski, 1981; Florek, 1991; Błaszkiewicz, 2005).

The present author studied the evolution of lake basins and river valleys in Northern Poland (Błaszkiewicz, 1998, 2005, 2010) and found unambiguous evidence for the presence of permafrost in the area until the beginning of the Holocene or the end of the Preboreal. This evidence is presented here in the form of dated melt-out of buried dead-ice blocks and of the timing of formation of glacial lakes on the present-day Polish territory.

STUDY AREA

The study area is located in Northern Poland and includes the northern part of the Tuchola Forest and the Starogard (Kociewie) Lakeland. This area was entirely covered by the last Weichselian ice sheet (Fig. 1). During its retreat, a distinct marginal zone associated with the Pomeranian Phase developed here. It is characterized by the proximal zones of outwash plains, and by the sedimentary contact between the ice and the upper proximal level of the outwash plain in the Stara Kiszewa region (Błaszkiewicz, 1998, 2003; Fig. 2). The current state of knowledge regarding the character of the Pomeranian Phase in the study area does not allow a precise reconstruction, but the absence of both marginal forms with glacitectonic structures and individual till layers suggests that it represents here only a stagnation of the ice sheet in the hinterland during the overall ice retreat.

The lack of interstadial organic deposits dating from before the Pomeranian Phase implies that the Pomeranian Phase can be dated only indirectly. Based mainly on dating of the later Gardno Phase (14.5–14.3 ka ^{14}C BP following Rotnicki and Borówka, 1994, 1995; Rotnicki, 2001), it is deduced to range from 16.2 ka ^{14}C BP (Kozarski, 1995) to 15.2 ka ^{14}C BP (Marks, 2002). This age range may also apply to the formation of the depressions which would become lake basins during the Late Glacial and early Holocene.

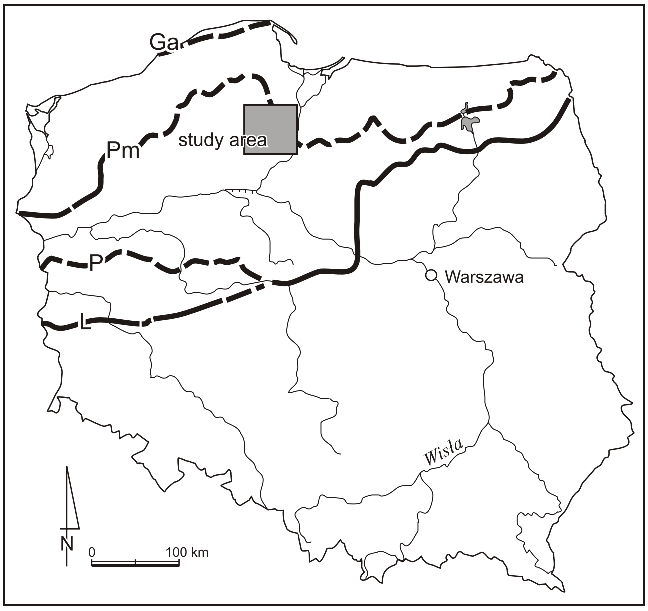

Fig. 1. Map of Poland with the location of the study area and the maximum

Extent of the Late Vistulian ice sheet (after Kozarski, 1995): L – Leszno Phase, P – Poznań Phase, Pm – Pomeranian Phase, Ga – Gardno Phase

The geomorphology of the study area in the forefield of the Pomeranian Phase is dominated by extensive outwash plains built of sand/gravel deposits up to 30 m thick. The most important surface features are the numerous subglacial channels associated with the glaciation and the dune fields formed during the main dune-forming phase in the Younger Dryas (Błaszkiewicz *et al*., 2006). The hinterland of the Pomeranian Phase is dominated by an undulating moraine plateau with numerous subglacial channels which are often incorporated in river-valley systems. The plateau areas commonly show a characteristic kame-melt relief, which indicates the patchy nature of the ice sheet disappearance during the retreat following the maximum extent of the Pomeranian Phase (Błaszkiewicz, 1998).

Permafrost remained present in the study area when the ice sheet retreated. This is evident from, among other features, thermal contraction cracks, oriented kettle-holes on the outwash plains and frost-related transformations of soils (Błaszkiewicz, 2005). The presence of lake basins is another line of evidence, and their long-term preservation, thanks to the presence of buried blocks of ice, and eventual disappearance are indications of the duration of the permafrost and its ultimate disappearance.

PRESERVATION OF DEPRESSIONS AS EVIDENCE FOR THE PRESENCE OF BURIED DEAD-ICE

A distinct time gap exists between the creation of the young glacial basins (which developed during glaciation) and the appearance of the lakes themselves (e.g., Więckowski, 1966; Stasiak, 1971; Gołębiewski, 1976; Niewiarowski, 1986, 1988, 1989, 1995, 2003; Ralska-Jasiewiczowa and Starkel, 1988; Florek, 1991; Błaszkiewicz and Krzymińska, 1992; Nowaczyk, 1994; Marks, 1996; Błaszkiewicz, 1998, 2002, 2005). Similar time gaps have been documented for other areas of the Weichselian Glaciation, for instance in Northern Germany (e.g., Gripp, 1964; Nitz, 1984; Pachur and Röper, 1987; Wünnemann, 1993; Böse, 1995; Nitz *et al*., 1995; Schlaak, 1997; Endtmann, 1998; Kaiser, 2001; Brande, 2002; Homann *et al*., 2002; Schlaak and Schoknecht, 2002), Belarussia and the Baltic states (Ekman, 1992; Zernickaya and Pavlovska, 1994; Almquist-Jakobson, 1995; Davydowa and Servant-Vildary, 1996; Zernitskaya *et al*., 2000; Matvieyev, 2002; Wohlfarth *et al*., 2002). They are also well-documented for deglaciated areas of the United States and Canada (Florin and Wright, 1969; Porter and Carson, 1971; Driscoll, 1980; Last and Vance, 2002; Schwalb and Dean, 2002; Eyles *et al*., 2003).

The geomorphological study of the lake basins has provided a wealth of evidence for the hypothesis that the fundamental cause of this time gap was a long-term preservation of the lake basins by dead-ice blocks, which, at least initially, took place under periglacial conditions. In the case of certain depressions in Northern Poland, the time gap is more than a thousand years (Błaszkiewicz, 2005). The origin of ice-preserving basins of mainly channel origin is controversial. Basically, it is assumed that it was glacial ice from the collapse of the channel roofs once the subglacial channels had been formed (e.g., Nowaczyk, 1994; Kehew and Kozłowski, 2007). A meltwater origin of the ice should, however, also be considered: water which formed subglacial channels was probably undercooled and under high hydrostatic pressure. Hence, it may have frozen rapidly immediately after the cessation of its movement.

These processes were beyond doubt influenced by permafrost preserved under the ice sheet from the time of its advance. Another argument for a meltwater origin of the buried ice is the limited amount of moraine material in the mineral substratum of the subglacial channels. The melting of glacial ice should have left large quantities of such material. The ice filling the depressions after their development should thus be considered as buried ice, rather than as dead-ice. The latter term should be applied exclusively to ice with a glacial genesis, i.e. ice which lost contact with the active or stagnant part of an ice sheet (Ben and Evans, 1998; Schomacker, 2008). Regardless of the origin of the ice preserved in the depressions, abundant data showing its presence were collected during research in young glacial areas. The most important are:

– the existence of various types of depressions and, above all, of deep subglacial channels within the younger outwash and a marginal lake surface within both large river valleys and spillways (Woldstedt, 1921, 1926; Bartkowski, 1953; Galon, 1982; Niewiarowski, 1986;

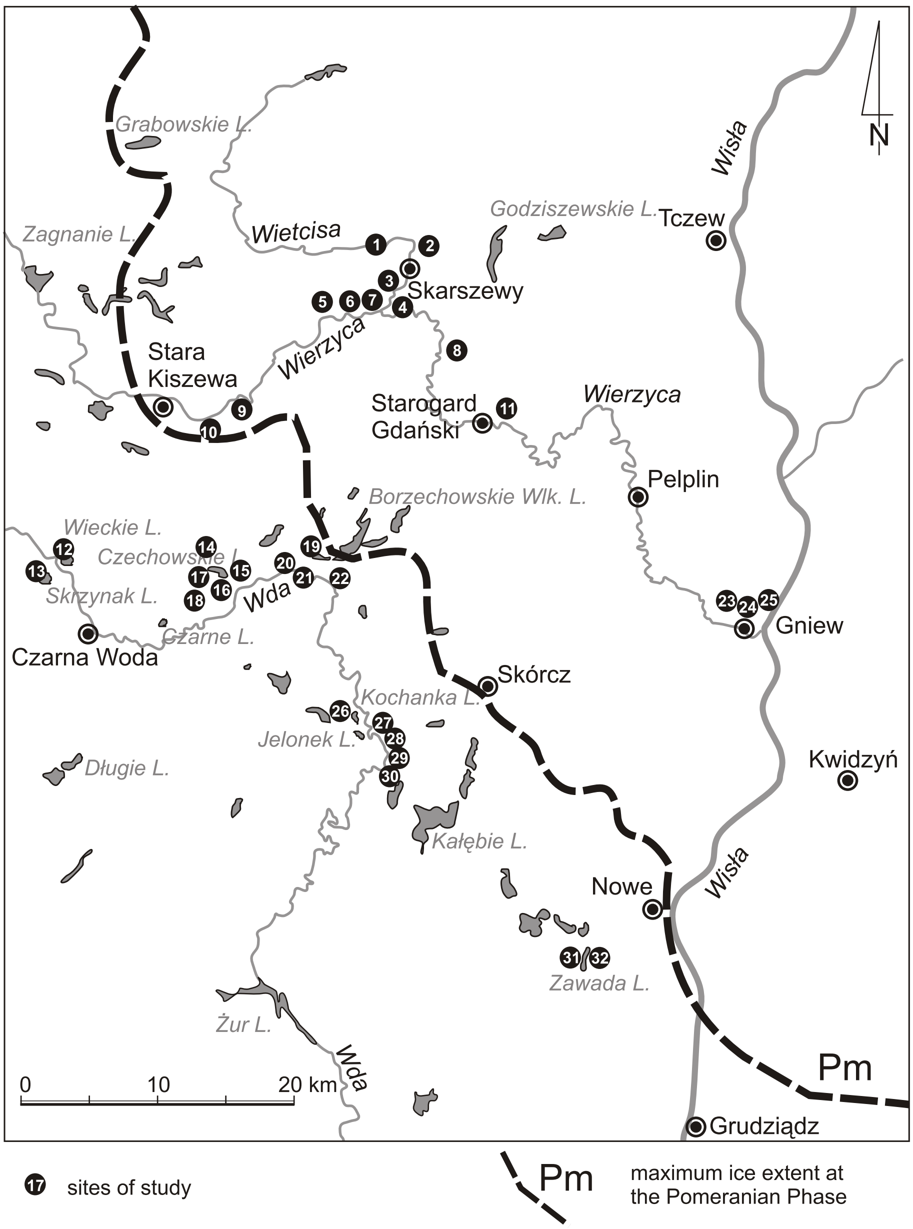

Fig. 2. Location of the sites in the Starogard Lakeland where the beginning of the lake development was determined by means of palynological and radiocarbon dates

Nowaczyk, 1994; Błaszkiewicz, 2005; Jorgensen and Sandersen 2006);

– the existence of deformation structures of which the formation was a consequence of melting of the ice bottom and which occur within the mineral substratum of the lake basins (primary mineral cover on dead-ice, consisting of fluvioglacial and glaciolimnic sediments as well as morainic material from the subsequent melting of dead-ice) but also in the lacustrine bottom sediments (examples of such structures were found for instance in the depression near Stara Kiszewa, where the bottom part of some of pre-Allerød lake sediments was cut by a network of antithetic faults; Fig. 3); similar structures were also documented from glaciolimnic sediments of the Gniew marginal lake (Błaszkiewicz, 2005; Błaszkiewicz and Gruszka, 2005);
– the presence of the so-called basal peat or bottom peat which underlies the lake sediments. The basal peat is thin, with an average thickness of 5–10 cm, and is located at the bottom of lake sediments, directly on the mineral substratum of the lake basin. The accumulation of this peat on top of blocks of dead-ice buried in the bottom of lake basins is indicated by its sedimentary genesis, with a steeply inclined surface of over 10° (Strahl and Keding, 1996; Helbig, 1999; Kaiser, 2001), as well as at locations concordant with the morphology of the mineral substratum. An important argument is the synchronicity of the peat accumulation, regardless of its current altitude at the bottoms of the basins (Gross, 1937; Woldstedt, 1952; Wieckowski, 1966; Florek *et al*., 1999; Kaiser, 2001; Homann *et al*., 2002; Błaszkiewicz, 2005). Additional evidence are comprises the geomorphological positions of the basal peats from both Allerød and Preboreal in the channels of river valleys in relation to the fluvial structures of the erosive sections of water gaps (Błaszkiewicz, 2002, 2005; Lorenz and Schult, 2004).

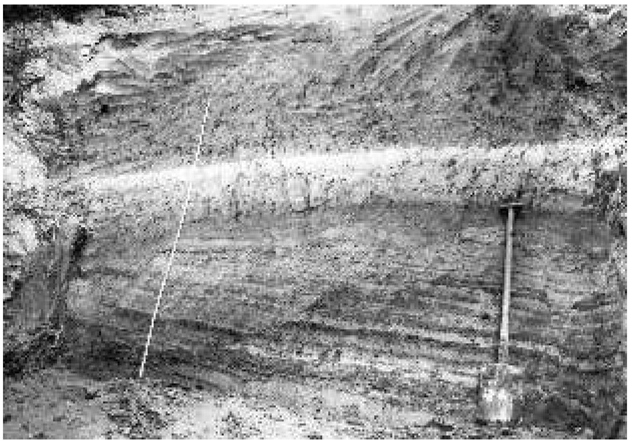

Fig. 3. Geological structure of pre-Allerød lake level in the end depression near Stara Kiszewa. In the lower part, a system of complementary faults (antithetic deformations), turning into flexures in the upper part, is visible. The deformations disappear within the light layer of the lacustrine chalk. The silt and sandy laminae above the lacustrine chalk are not deformed (photo by Cz. Kuchta)

A layer with plant remains found in the bottom part of the lacustrine sediments of many lakes in Minnesota (Florin and Wright, 1969) can be considered as an equivalent of the basal peat. The main difference is the composition of the vegetation in the two organic layers. In the case of the central European Lowland, this was mostly moss growing under tundra conditions, while it was a relatively high forest dominated by spruce trees in North America (Wright and Stefanova, 2004).

MELTING OF BURIED DEAD-ICE BLOCKS: THE START OF LAKE DEVELOPMENT

During previous research in this area, the present author determined the beginning of the development of lakes in 30 lake basins (Fig. 2 and Table 1), which occur both in the immediate hinterland of the ice sheet when it had reached its maximum extent during the Pomeranian Phase (the Wierzyca catchment area) and in the area just in front of the ice (the Wda catchment area). The vast majority of the basins appear to represent subglacial channels created by the erosion of subglacial streams. Some of these subglacial channels must, however, taking into account that they contain low-lying morainic levels with a subglacial relief that has developed locally (drumlin-like forms), have been eroded also by the ice (apart from the erosion by subglacial streams). Field observations also indicate the existence of older, pre-Weichselian structures in some of the larger depressions (Błaszkiewicz, 1998, 1999, 2005).

The basins also differ from one another with respect to their relationships with the present-day hydrographical network. In addition to the depressions which became incorporated in river valleys (Wierzyca, Wda, Wietcisa), endorheic basins appear also to exist. Six of the basins still contain lakes (the Skrzynka, Borzechowskie Małe, Czechowskie, Zawada, Jelonek and Kochanka lakes), while the other lakes have already disappeared. Complex melting of buried ice must have occurred. The melt-out processes varied in intensity from the moment of ice sheet retreat, and lasted in some depressions until the end of the Preboreal. In some of the basins, they occurred in successive stages and on successive morphological levels, which eventually resulted in the development of the present-day local river network and caused hiatuses in the lacustrine sedimentation. The principal mechanism for the melting of the ice that was buried in the depressions was obviously downwasting in the sense of Schomacker (2008). Buried ice melting by backwasting was common in the case of positive forms of ice-cored moraines.

Considering the timing of the lacustrine sedimentation and its Late Glacial development, three major morphogenetic groups of lake basins can be distinguished (Błaszkiewicz, 2005; Fig. 4):

– basins where the beginning of lacustrine sedimentation started before the Allerød;
– basins where lakes started to form during the Bølling-Allerød interval;
– basins where the beginning of lacustrine sedimentation started at the beginning of the Holocene, i.e. the Preboreal.

Table 1

Sites in the Starogard Lakeland where the beginning of the lake development was determined, by means of palynological and radiocarbon dates

Site no.	Profil	Geomorphic sitting	Material dated	Depth [m]	Lab. no	BP ^{14}C	Age cal.^{14}C years BP$^{(*}$	Palynological analysis	References
1	Wolny Dwór	peat plains	peat layer	14	Gd-15144	12 380 ±270	14615 ±520	AL	Błaszkiewicz (2005)
2	Wilcze Góry	peat plains	peat layer	10.50	–	–	–	AL	Błaszkiewicz (2005)
3	Skarszewy 1	peat plains	peat layer	7.20	Gd-15158	12 500 ±220	14763 ±470	BO/AL	Błaszkiewicz (2005)
4	Skarszewy 2	lake terrace	peat layer	2.5	–	–	–	AL	Błaszkiewicz (2005)
5	Malary	peat plains	peat layer	6.0	Gd-5937	9660 ±100	10993 ±165	PB	Błaszkiewicz (1998)
6	Więckowy 1	peat plains	peat layer	17.7	Gd-11156	9680 ±110	11006 ±176	PB	Błaszkiewicz (2005)
7	Więckowy 1	peat plains	peat layer	17.6	Gd-15140	9460 ±220	10752 ±309	PB	Błaszkiewicz (2005)
8	Linowiec	peat plains	peat layer	6.2	–	–	–	PB	Błaszkiewicz (1998)
9	Boże Pole Szlacheckie	lake terrace	peat layer	6.6	Gd-6311	13 010 ±220	15851 ±543	PAL	Błaszkiewicz (1998)
10	Palubinek	peat plains	peat layer	11.6	Gd-4620	12 600 ±240	14933 ±486	BO	Błaszkiewicz (1998)
11	Kochanka	peat plains	peat layer	10.7	–	–	–	PB	Błaszkiewicz (1998)
12	Wieck	fan–delta	gyttja	14.5	–	–	–	AL	Błaszkiewicz (2005)
13	Skrzynka Lake	bottom lake	gyttja	16.6	–	–	–	BO/AL	Błaszkiewicz (2005)
14	Piece	peat plains	gyttja	12.0	–	–	–	BO/AL	Błaszkiewicz (2005)
15	Iwiczno	peat plains	peat layer	13.1	Gd-13052	10 700 ±350	12454 ±475	AL	Błaszkiewicz (2005)
16	Czechowskie Lake	bottom Lake	gyttja	23.5	–	–	–	AL	Błaszkiewicz (2005)
17	Czechowskie Peatlands	peat plains	peat layer	6.2	Gd-10932	12 500 ±170	14771 ±413	BO/AL	Błaszkiewicz (2005)
18	Czarne	peat plains	peat layer	–	–	–	–	AL	Błaszkiewicz (2005)
19	Borzechowskie Małe Lake	bottom lake	peat layer	18.8	–	–	–	AL	Błaszkiewicz (2005)
20	Borzechowo 2	peat plains	peat layer	13.2	Gd-12387	9720 ±150	11054 ±221	PB	Błaszkiewicz (2005)
21	Borzechowo 3	peat plains	peat layer	9.65	Gd-12391	9888 ±130	11415 ±200	PB	Błaszkiewicz (2005)
22	Biała Góra	peat plains	peat layer	11.1	Gd-12393	9860 ±130	11383 ±205	PB	Błaszkiewicz (2005)
23	Gniew 1	peat plains	peat layer	16.1	Gd-14047	12 250 ±350	14490 ±590	AL	Błaszkiewicz (2005)
24	Gniew 2	peat plains	peat layer	16.0	Gd-11470	11 450 ±120	13972 ±561	AL	Błaszkiewicz (2005)
25	Gniew 3	peat plains	peat layer	15.95	Gd-13045	11 850 ±390	13348 ±166	AL	Błaszkiewicz (2005)
26	Kochanka Lake	bottom lake	peat layer	22.7	Poz-29716 (AMS)	11 730 ±60	13611 ±136	–	this study
27	Smolniki	lake terrace	peat layer	5.74	Poz-26683 (AMS)	12 590 ±70	14941 ±302	BO/AL	this study
28	Szlaga 10	peat plains	peat layer	14.45	Poz-26780 (AMS)	9380 ±50	10612 ±61	PB	this study
29	Szlaga 11	peat plains	peat layer	12.54	Poz-26742 (AMS)	9330 ±50	10544 ±78	PB	this study
30	Skorzenno	peat plains	peat layer	17.90	–	–	–	BO/AL	this study
31	Zawada Lake	bottom lake	peat layer	18.50	–	–	–	PB	Błaszkiewicz (2005)
32	Zawada Peatlands	peat plains	peat layer	10.0	Poz-3628 (AMS)	9760 ±60	11183 ±46	PB	Błaszkiewicz (2005)

* – calibration in program *CalPal*; PAL – pre-Allerød; BO – Bølling; AL – Allerød; PB – Preboreal

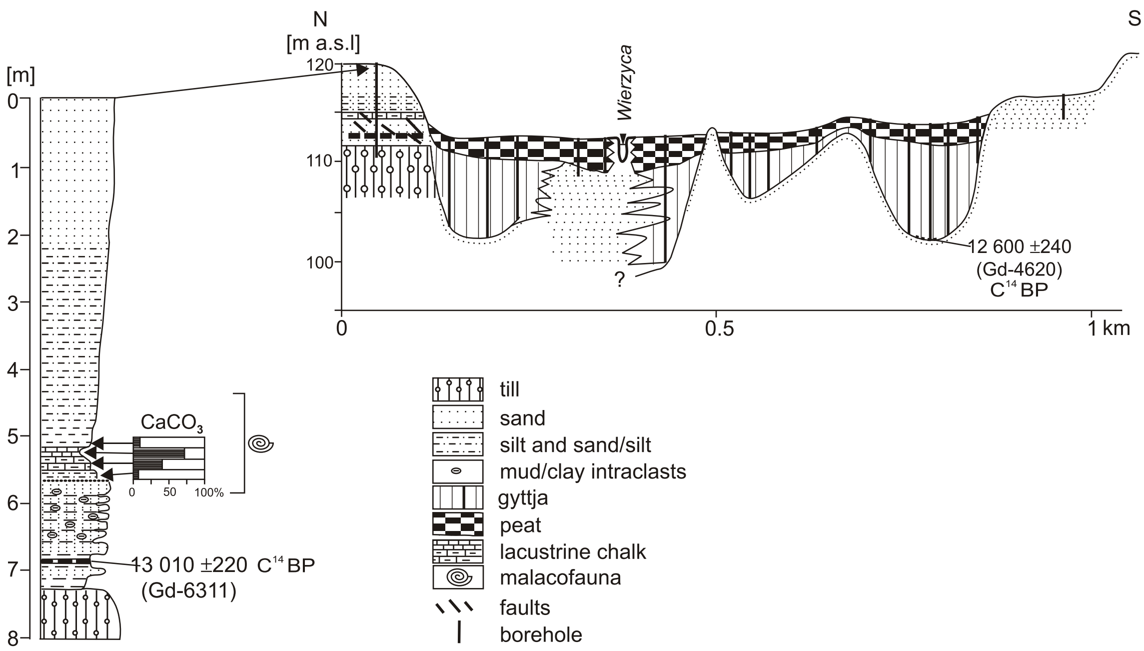

Fig. 4. Cross-section through the bottom of a depression near Stara Kiszewa

THE PRE-ALLERØD LAKES

The term “pre-Allerød” is not very precise and includes the possibility of accumulation of lake sediments in both Bølling and pre-Bølling (Meiendorf after Menke, 1968). It applies to most situations in which more precise palaeobotanic dating is difficult, for example due to the limited thickness of biogenic sediments within the mineral deposits, without a clear stratigraphic continuity with the biogenic sediments from the younger, i.e. Allerød time-span.

Lacustrine sediments of pre-Allerød lakes are not commonly reported from the central parts of lake basins, but from high-lying levels on the slopes of the depressions, usually from a few to several metres above the current lake-water surface or the modern peat plains (Nitz, 1984; Chrobok *et al*., 1986; Niewiarowski, 1986, 1988, 1989; Błaszkiewicz and Krzymińska, 1992; Nowaczyk, 1994; Nitz *et al*., 1995; Błaszkiewicz, 1998).

Undoubtedly the development of pre-Allerød lakes took place in a periglacial climate, when dead-ice blocks were present, as indicated by the deformation structures in the deposits, and with small melt-out forms at the surface of morphological levels. Sediments of pre-Allerød lakes are rarely found in the deepest parts of modern lakes. Hence, it can be inferred that they were relatively shallow, insignificant, and spread over the lower parts. Their occurrence was as a rule confined to the slopes of the depressions. The pre-Allerød lakes had a short lifetime because they were quickly filled or drained due to the development of the river network and the integration of the individual basins into valley systems.

The most representative site with these deposits occurs in the depression near Stara Kiszewa, in which a fragmented pre-Allerød lacustrine level is present at an altitude of 3–8 m above the floodplain of the Wierzyca River and a few metres below the moraine plateau. The most characteristic elements in the morphology of this level are small melt-out depressions. The level includes a 7 m sand/mud/carbonate lacustrine succession on top of a sandy ablation till (Fig. 4). The lowermost unit of this succession consists of alternating massive muds and massive sands with a large amount of mud/clay intraclasts. At a depth of 6.6 m, between layers of clayey mud, a discontinuous peat layer is present. Its radiocarbon age is 13 010 ±220 years BP (Gd-6311), and it contains a Younger Dryas flora (with, among other taxa, *Salix*, *Betula nana*, *Dryas octopetala*, *Selaginella selaginoides*). Directly above these deposits, a 0.5 m thick carbonate layer occurs containing a malacofauna with cold-loving species of Holarctic range (Błaszkiewicz and Krzymińska, 1992); it is covered by clayey/muddy/sandy deposits that form a succession 5 m thick.

All deposits in the bottom part of the exposure, down to the carbonate layer, are cut by a network of normal faults with vertical displacements up to 20 cm (Fig. 3). Along the fault lines, continuous deformations appear at the bottom of the carbonate layer in the form of small folds, while the overlying layer is not disturbed. The geological structure of the level records the full life cycle of a pre-Allerød lake, from the shallow backwaters of a wetland character, through a gradual deepening of the lake basin as a result of melting of the ice in the substratum, to the shallowing of the lake as a result of rapid supply of clastic material.

The above pre-Allerød lake was drained in the middle Bølling towards the nearby channel. This is indicated by radiocarbon (12 600 ±240 years BP – Gd-4620) and palynological datings of the peaty mud located at the base of the lacustrine sediments in the centre of the marginal depression (biogenic-alluvial floodplain) that already had formed in the younger-generation lake. Lack of pre-Allerød sediments suggests that the functioning of these oldest lakes was limited to small spaces between the slopes of the depressions, and the blocks of dead-ice filling most of these depressions.

The analogy with the oldest generation of the lakes in the Biesenthal Basin near Berlin (Nitz *et al.*, 1995) is large. Moreover, such a situation indicates a time gap in lacustrine sedimentation between the accumulation in pre-Allerød lakes and the successive younger generation of lakes which developed at the lower morphological levels of the depressions. In the case of the marginal depression near Stara Kiszewa, it lasted only until the middle Bølling.

THE BØLLING-ALLERØD LAKES

Despite documented cases of pre-Allerød developments of lakes, most sediment infilling of lakes in basins of the young glacial areas started during the Bølling–Allerød interval (e.g., Więckowski, 1966; Stasiak, 1971; Niewiarowski, 1989; Ralska-Jasiewiczowa and Starkel, 1988; Nowaczyk, 1994; Nitz *et al*., 1995; Żurek, 1995, 1996; Davydova and Servant-Vildary, 1996; Jakuszko, 1999; Kaiser, 2001). Subsequently, when the largest Late Glacial global warming occurred, the main phase of melting of buried dead-ice blocks took place. In some depressions, the formation of these lakes was preceded by a short peat-forming phase. After melting of the buried dead-ice blocks, these peats occupied the lowest positions of the lake bottoms of the modern lake basins.

Directly above the basal peat layer, deep lacustrine sedimentation took place during the Late Glacial. This lacustrine sedimentation was continuous, which indicates rapid melting of the dead-ice, resulting in a supply of water to and total exhumation of these basins during the Allerød (Fig. 5). A particularly good indicator of the accelerated pace of melting of the dead-ice blocks and the origin of these lake basins already in the Allerød is the undisturbed character of the deposits overlying the Allerød gyttja. These deposits show well-developed continuous lamination in the lower parts of cores from Jelonek Lake and Czechowskie Lake in the Tuchola forests (Błaszkiewicz, 2005). Geomorphological evidence indicates that the water level of the lakes was commonly slightly higher immediately after the development of the lakes than it is nowadays (Błaszkiewicz, 2005).

THE PREBOREAL LAKES

Sedimentation started in four of the subglacial channels under study during the Preboreal. In all cases, the development of early Holocene lakes was preceded by a peat-forming phase (Fig. 5). The Preboreal basal peats are much thicker than the Allerød peat (average 10–40 cm). The peats are situated directly on the mineral substratum of the channels. The absence of older lacustrine or peat deposits at the base of the Preboreal basal peat indicates that the basins of these lakes never acted as sedimentary basins during the Late Glacial.

The warming of the Bølling-Allerød interval resulted in a gradual lowering of the top of the buried ice (Fig. 5). Drainage could, however, restrict the intensity of this process. A simultaneous increase in the thickness of the mineral cover by morainic material falling from the material on top of the melted ice should also be considered. Accumulation of the basal peats on the mineral cover started as a rule during the beginning of the Preboreal, whereas deep-lake accumulation took place at the end. This is indicated by the fast melting of buried dead-ice blocks, leading to the fomation of the lake basins at the end of the Preboreal. The youngest glacial lakes in the young glacial area could then thus develop, and their further development depended on the climate and the hydrological conditions (Błaszkiewicz, 2005; Słowiński, 2010).

CAUSES OF THE TEMPORAL VARIATION OF THE MELT-OUT PROCESSES

As mentioned above, the various lake ages were determined by different durations of the preservation of the blocks of dead-ice and different developments of the melt-out processes. The thickness and the type of the mineral cover had beyond doubt also an important influence (Galon, 1982; Niewiarowski, 1989; Nowaczyk, 1994; Böse, 1995), and this holds also for the presence of peats above the blocks of buried ice (Błaszkiewicz, 2005).

The field data clearly indicate that, in addition to the mineral cover, a particularly important factor influencing the preservation of dead-ice blocks until the Preboreal was the continuous drainage of the depressions (Błaszkiewicz, 2002, 2005). This condition was met in particular by the basins located in the vicinity of a stream that caused downward erosion during the Late Glacial. On the other hand, in places where the morphological conditions favoured a longer persistence of stagnant water, the thermal effect of the water on the underlying dead-ice led to melting and the formation of lake basins as early as the beginning of the Late Glacial.

Accelerated melting, which is a process with positive feedback, took place after the first season during which the lake had become sufficiently deep to avoid freezing of its water down to the bottom (Homann *et al.*, 2002; Błaszkiewicz, 2005). A good example of such conditions is the channel of the Borzechowskie lakes where it meets the valley of the Wda: the differences in age of the lakes depend on the relation of individual parts of the channel to the draining surface (Błaszkiewicz, 2005). The section of the subglacial channel in which lacustrine sedimentation, preceded by a phase of fen development, occurred only in the Preboreal, was drained during the Late Glacial by the quickly downwards eroding Wda River (Figs. 6 and 7). Both the local morphological conditions, especially the slope of the part of the channel that was concordant with the axis of the valley, and the permanent lowering of the base level of the Wda Valley, favoured rapid drainage. The buried ice in the bottom of the channel was therefore not exposed to the higher temperature of the water and could escape melting longer. Only with the rapid climatic amelioration during the second half of the Preboreal and the stabilisation of the erosive and accumulative processes at the bottom of the Wda Valley, did this entire section of the channel became a lake (with running water).

The palaeogeographical development along the Wda channel was, obviously, more complex. The morphological conditions in other parts of this channel, where the basin of Małe

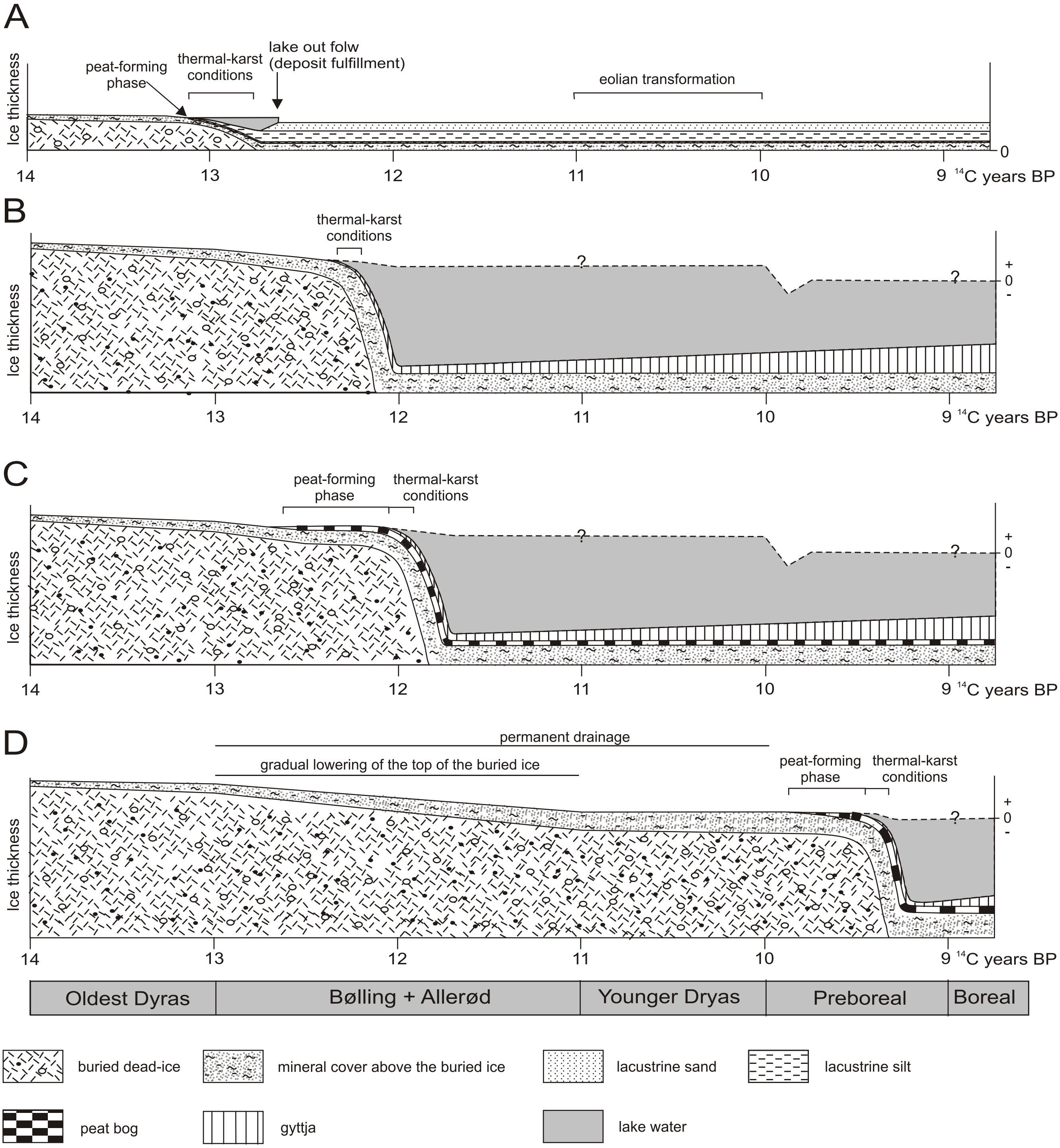

Fig. 5. Melt-out development of the buried dead-ice and related lake development in the Late Glacial and early Holocene

A – pre-Allerød lakes; **B** – lakes of the Bølling-Allerød interval; **C** – lakes of the Bølling-Allerød interval with basal peat bog; **D** – early Holocene lakes; 0 – level of contemporary lakes or peat bog plains

Borzechowskie Lake developed, did not, on the other hand, enable drainage of the meltwater that was supplied as a result of the global warming during the Bølling-Allerød interval. The thermal influence of the water on the underlying ice thus led there to the rapid formation of a deep lake. Already during the Allerød, after a brief peat-forming phase, calcareous gyttja started to develop in the deeper part of the lake. Complete melting of the buried ice and full development of the lake basin then took place. It must be emphasized in this context that the melting of buried dead-ice blocks under thermal-karst conditions was very fast, regardless of the time-span during which the basins were preserved.

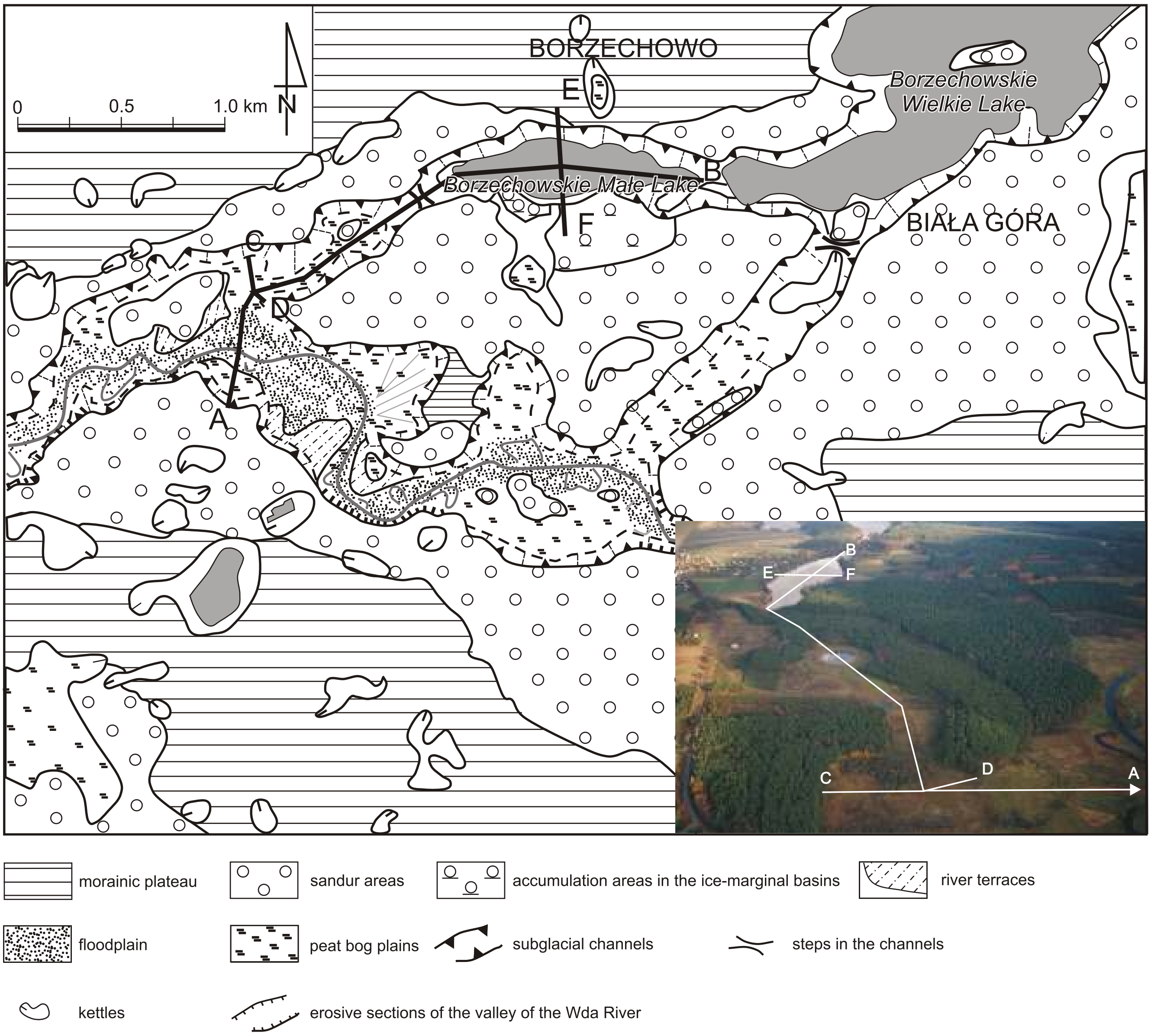

Fig. 6. Geomorphological map of the contact zone of the sublacial channel of the Borzechowskie lakes with the valley of the Wda River

BURIED ICE AND PERMAFROST – DISCUSSION

Although the main melting phase of the dead-ice blocks and the formation of the lakes took place during the Bølling-Allerød interval, basins with buried ice survived in eastern Pomerania until the Preboreal (Błaszkiewicz, 2003, 2005). The relationship between the duration of the preservation of the basins and the presence of permafrost is often emphasized (e.g., Nitz, 1984; Böse, 1995; Kozarski, 1995). Long preservation of dead-ice bodies after the disappearance of permafrost is difficult to accept (Marks, 1996), although the marginal zones in Iceland show that it is possible: the area in front of, among other glaciers, the Tungnaarjökull and Skeidararjökull, where no permafrost is present and the average annual temperature is about 4°C, still contain blocks of glacial dead-ice in ice-moraine ramparts, kame forms and outwash levels from the end of the Little Ice Age (Wiśniewski *et al.*, 1997; Worsley, 1997; Andrzejewski, 2002). Obviously, gradual melting of buried ice takes place, but the intensity of this process is determined primarily by the local morphological conditions affecting mass movements, and by the thickness of the mineral cover and the presence of vegetation (Krüger and Kjaer, 2000; Kjaer and Krüger, 2001; Schomacker, 2008).

Under permafrost conditions, ice blocks can be fully preserved if the mineral cover is thicker than the depth of the active layer (Boulton, 1967; Grześ, 1986, 1987; Araźny and Grześ, 2000). Ice that was buried this way becomes part of the permafrost and its further development is related to the permafrost evolution. In turn, buried ice affects the degradation of permafrost significantly; this was observed for permafrost with a large amount of ground ice in northwestern Canada (Murton, 1996, 2001; Dyke and Brooks, 2000), and for pingo forms in Central Asia (Babiński, 1982).

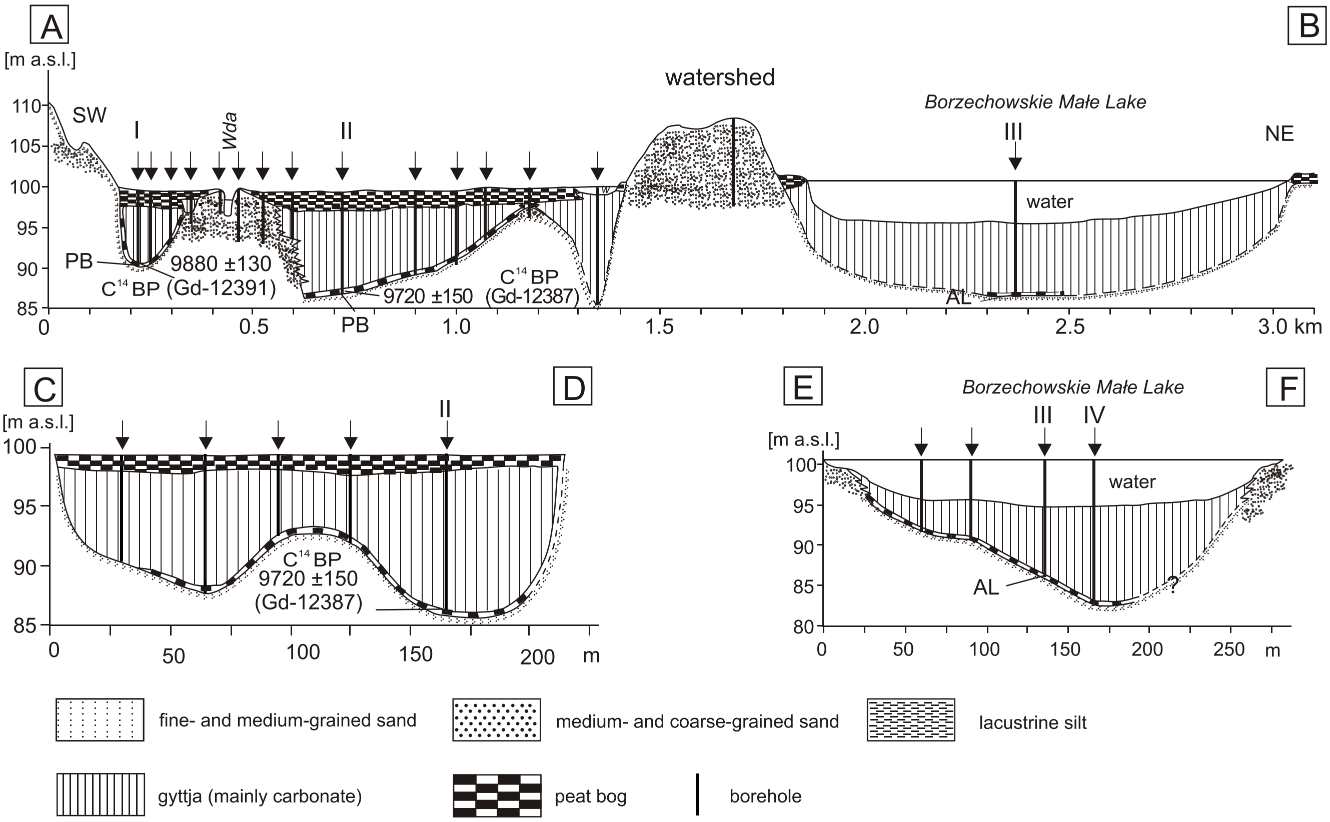

Fig. 7. Geological cross-sections across the bottom of the sublacial channel of Lake Małe Borzechowskie – Wda River

AL – Allerød; PB – Preboreal; for location of the cross-sections see Figure 6

Once reservoirs with water that does not freeze to the bottom in winter develop over buried blocks of dead-ice, the thermal effect of the water on the dead-ice underneath leads to rapid melting and the formation of a lake basin. This happened at the beginning of the Late Glacial (Błaszkiewicz, 2003). Consequently, where blocks of dead-ice were present, taliks spread down from the surface, significantly accelerating the degradation of the permafrost.

The buried ice had a chance to be preserved longer, however, even until the Preboreal, in places where the relief favored continuous drainage. This condition was especially fulfilled by the basins that were located in the vicinity of a stream, and that were subject to ongoing downward erosion during the Late Glacial (Błaszkiewicz, 2005). The beginning of the phase change from ice to water requires large amounts of energy; hence, dead-ice caused a significant delay for such a change in thermal regime, at least in its immediate vicinity. Large blocks of ice must have been powerful "reservoirs" of cold, and they must thus have influenced the persistence of low temperatures (below or close to 0°C) in the mineral environment. This created suitable conditions for the preservation of relic permafrost throughout the Bølling-Allerød interval.

The cooling of the Younger Dryas favored further preservation of dead-ice blocks. The dominance of mineral sediments at the bottom of the then lakes indicates intensive slope processes that could develop thanks to the sparse vegetation of the parkland tundra (Madeyska, 1998). Most probably discontinuous permafrost reappeared in the Younger Dryas, and at places with surviving relict ice partial aggradation even took place. This conclusion is supported by the climate models, which suggest that permafrost was then probably present (Renssen *et al.*, 2000). The rapid warming in the early Holocene must, obviously, quickly have made disappear the surviving blocks of ice and the surrounding permafrost. Where buried dead-ice blocks still were present, the youngest generation of lakes appeared at the end of the Preboreal.

In this context, the results of temperature measurements in deep cores in the Suwałki region (including borehole Udryń IG 8) are interesting. At a depth of about 550 m, thermal inversion was recorded, interpreted as an echo of permafrost (Szewczyk, 2002, 2005). In 2010, at a depth of 357 m, a frozen soil layer was found in the nearby Szypliszki borehole (Szewczyk and Nawrocki, 2011). All these boreholes are located in the area covered by the ice sheet of the last glaciation. Such a deep range of primeval permafrost suggests that, when this area was covered by the ice sheet, not all permafrost which had formed during the cataglacial phase of the Late Weichselian disappeared, but it is possible that relics of permafrost from earlier Weichselian phases had been preserved.

Permafrost degradation under an ice sheet is likely to be limited to the upper layer of the permafrost, in which circulation of subglacial water can take place. The permafrost forming from the surface downwards can become connected with deeper relic permafrost during ice sheet retreat. This supports the possibility of long-term preservation of certain basins by buried blocks of ice until the Preboreal.

Palaeozoological and palaeobotanical studies form the best source of our knowledge about the palaeogeography and the environment of the Late Glacial and early Holocene of Poland (Starkel, 1977; Kozarski, 1991; Ralska-Jasiewiczowa, 1991). Plant and animal habitats adapted quickly to environmental changes, mainly evoked by climate change. The response of most of the abiotic components to these changes took place, however, with some delay. The biotic environment of the end of the Allerød, of course excluding the Younger Dryas, was almost the same as that of the Holocene. In the area of the Tuchola Forest, birch/pine communities with an admixture of aspen dominated (Miotk-Szpiganowicz, 1992). The reaction of the "glaciation underground" to the Late Glacial climate change was, however, far slower and depended largely on local factors and on the temperature distribution in the lower subsoil.

CONCLUSIONS

It has been found on the basis of the lowermost lacustrine sediments that the start of lacustrine sedimentation was asynchronous in the study area, which had been covered by an ice sheet during the last glaciation. Most lakes appeared in the Bølling-Allerød interval, but older lakes had already formed before the Allerød, whereas other lakes came into being only at the end of the Preboreal.

The main reason for the difference in the formation of the lakes was due to the diversity of melt-out processes of the buried dead-ice blocks. The melting intensity differed after the time of ice sheet retreat and lasted in some depressions until the end of the Preboreal. Particularly important for the preservation of blocks of dead-ice was the constant or interrupted character of the drainage of the depressions, especially for the depressions in the vicinity of a water course that was subject to downward erosion during the Late Glacial.

On the other hand, local morphological conditions favored longer preservation of stagnant water, and the thermal effect of water above the dead-ice led to rapid melting and to the establishing of a lake basin at the beginning of the Late Glacial. The melt-out of buried dead-ice blocks, under thermokarst conditions, was very fast regardless of how long that dead-ice had survived. The long-term preservation of the lake basins as a consequence of the presence of buried dead-ice blocks not only indicates the presence of Late Glacial permafrost in the areas covered by the last ice sheet, but also provides evidence for the timing of its ultimate disappearance. The probable age of the final degradation of permafrost in the study area is the end of the Preboreal.

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