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# A high-resolution record of Late-Glacial and Early-Holocene climatic and environmental change in the Czech Republic

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**Abstract**

Recent discovery of thick buried lake sediments in the Třeboň Basin, South Bohemia, presents an exceptional opportunity to study the Late-Glacial history of eastern Central Europe. High-resolution investigation of pollen, plant macrofossils, algal remains, and lithology for the Late-Glacial and Early-Holocene sediments of former Lake Švarcenberk yielded a well-founded palaeoclimatic and palaeovegetational data that can be compared with the results from other parts of a west–east European transect, taking into account the oceanic/continental gradient and its influence for palaeoenvironmental conditions. The results demonstrate that the effect of North Atlantic oceanic changes during the last glacial–interglacial transition extended to the investigated area. Nevertheless, significant differences in timing, intensity, and character of vegetational response to these climatic changes have been found between the area under study and the western part of Central Europe. These differences can be ascribed to increased seasonality, specific regional mesoclimatic and soil conditions, and possible local glacial refugia for pine. The results of sediment chemical analyses indicate a close correspondence between climatic, vegetational, and soil development in lake catchments. 

## 1. Introduction

Organic deposits suitable for palaeoecological research usually began to form about 13 ka BP in western and northwestern Europe, classified into different Late-Glacial phases by pollen analysis. In non-glaciated, continental regions of Central Europe, this subdivision is usually not possible, either because minerogenic sediments contain no pollen or because sediments simply did not accumulate at this time. Particularly, lacustrine sequences are very rare in these regions. The profile under study in the Czech Republic is a unique example with an extensive and well-stratified Late-Glacial record. High sediment-accumulation rates permit the detection of brief Late-Glacial climatic oscillations, so that comparisons can be made with numerous results from western and northwestern Europe, where the basic biostratigraphic and climatostratigraphic concepts have been developed (Iversen, 1954; Mangerud et al., 1974; Watts, 1979). As postulated by Ruddiman and McIntyre (1981) and later recognised in terrestrial records within the areas adjacent to the North Atlantic (e.g. Lowe et al., 1994; Walker, 1995), the rapid climatic changes during the last glacial–interglacial transition can be ascribed to large-scale shifts in the position of the oceanic Polar Front, which have amphi-Atlantic or even global effects (e.g. Peteet, 1995).

The goal of the present study is to test the validity of some of these concepts as applied to the eastern part of Central Europe. The great distance from the North Atlantic and from major ice sheets, as well as the high degree of regional environmental diversity, are the most important factors that could cause certain differences in climate and provoke distinct biotic responses to climatic changes in the area under study. It is well known that the response of populations (e.g. plant populations) to climatic change is likely to be greatest near the margin of their distributional limits (Watts, 1979). Rapid climatic changes during the last glacial–interglacial transition usually affected only local populations and did not permit long-distance migrations (Ammann, 1989). This caused a high degree of inter-regional biological diversity, depending on local availability of species. For example, in the intermontane basins of the Western Carpathians, some 250 km east of the study area, local

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populations of *Pinus cembra, P. sylvestris, Larix decidua*, and *Picea abies* responded to the onset of Late-Glacial climatic amelioration by expansion to higher altitudes from their locally present glacial refugia (Ralska-Jasiewiczova, 1980; Jankovská, 1984; Rybníček and Rybníčková, 1994). This pattern is completely different from any other recognised in Europe north of the Alps, which probably had no local glacial refugia.

In spite of the problems with absolute dating of Late-Glacial events due to the occurrence of $^{14}C$ plateaus, the biostratigraphic subdivision of NW European Late-Glacial is today well fixed on an absolute time scale (Ammann and Lotter, 1989; Wohlfarth, 1996, Hoek, 1997). In this paper, detailed correlation of recognised biostratigraphic boundaries with events described from other parts of Europe is not possible on an absolute chronological basis due to the lack of sufficent AMS radiocarbon dates of terrestrial plant remains. Biostratigraphical terminology is therefore followed in this paper. Local pollen assemblage zones (PAZ) are correlated with regional climatostratigraphical units according to Mangerud et al. (1974) and Ammann and Lotter (1989). This correlation is only roughly confirmed by available $^{14}C$ dates.

## 2. The study area and site

The study site is situated in South Bohemia in the flat landscape of the Třeboň Basin (which has an area of about 700 $km^2$ and maximum relief undulation of 20–40 m). Sandy and clayey Cretaceous sediments with locally superimposed Tertiary sediments constitute the principal geological substratum. Depressions are filled with Quaternary alluvial silt and gravel, aeolian sands, and particularly peat bogs. The content of clay in soils generally increases with depth, and soil aeration is reduced accordingly. The soil nutrient content is generally poor. Calcium carbonate deficiency is common, potassium is sufficient only in deep soil horizons, nitrogen content is low, and concentrations of phosphate are moderate (Husák and Hejný, 1978). Most soils are leached and show a tendency towards podzolization. The soil reaction is mostly highly acidic (pH up to 3.3). Various types of podzols and sandy or peaty gleysols prevail.

The present climate is suboceanic and is determined by prevailing westerly air masses, already significantly reduced in moisture by passage across central Europe. The region is somewhat sheltered by the Šumava-Bavarian Forest highlands. Macroclimatic conditions are strongly modified by presence of extensive wetlands: Frequent occurrence of fog is typical for the region. Annual mean precipitation is 622 mm (January being the driest month), annual mean temperature is 7.4°C (see Fig. 1b).

The potential vegetation of the Třeboň Basin would generally be silver fir-oak woodlands (*Abieti-Quercetum*), in waterlogged areas bird cherry-pendulate oak and -alder woodlands (*Quercus robur-Padus avium* and *Alnus glutinosa-Padus avium* communities), alder carrs (*Carici elongatae-Alnetum*), and reed swamps and tall-sedge communities (*Phragmito-Magnocaricetea*) (Neuhäuslová, 1998). Vast transitional peat bogs are still dominated by forests of Bog-pine (*Pinus rotundata*). Acidophillous pine (*Pinus sylvestris*) forests with oak (*Quercus robur*), birch (*Betula pubescens*), and spruce (*Picea abies*) prevail on dry soils of river terraces and aeolian sands.

The Třeboň Basin, originally an inaccessible swampy area, remained largely a wilderness until the 13th century. During Late Medieval time, it developed into a cultural landscape of fish culture and forest plantations, and fishponds still constitute a characteristic element of the landscape.

The first palynological investigation of the area was carried out by Rudolph (1917). While his primary focus was the investigation of plant macrofossils in several peat bogs, he supplemented his results by analysis of some types of arboreal pollen. According to his results, the basal age of some investigated deposits was later established to be of "Kiefern-Zeit" (Rudolph and Firbas, 1922). The early postglacial age of most peat deposits in Třeboň basin was later confirmed by Klečka (1926, 1928) and Štěpánová (1930). Small pollen counts and exclusive focus on arboreal pollen were the disadvantages of these early palynological investigations. In the early 1960s, Jankovská started her palaeoecological investigations of Třeboň Basin, using modern approaches. Her work (Jankovská, 1980) forms the basis of our knowledge of the area.

The former lake Švarcenberk is situated 4 km south of the city of Veselí nad Lužnicí (49°9′N, 14°42′E) at 412 m a.s.l. Limnic sediments are overlain by peat, which formed after the natural infilling of the lake at approx. 5500 BP according to $^{14}C$ dating. Nowadays, the site is heavily influenced by intensive management. Between 1698 and 1701 a dammed fishpond was constructed directly on the site, and its waters almost completely flooded over the peat and the underlying lake sediments. The only presumed remnants of the original vegetation cover are the small patches of tall sedge and *Sphagnum* communities (*Eriophorion gracilis* and *Rhynchosporion albae*) in western part of the locality.

For local hydrology the presence of several strong artesian springs is characteristic. The underground water ascends along a deep tectonic fault and is rich in iron oxides. The former lake was presumably supplied almost exclusively by this artesian water. The activity of underground water sources was apparently independent of general climatic fluctuations in the past. As a result, lake level remained almost constant over the millennia

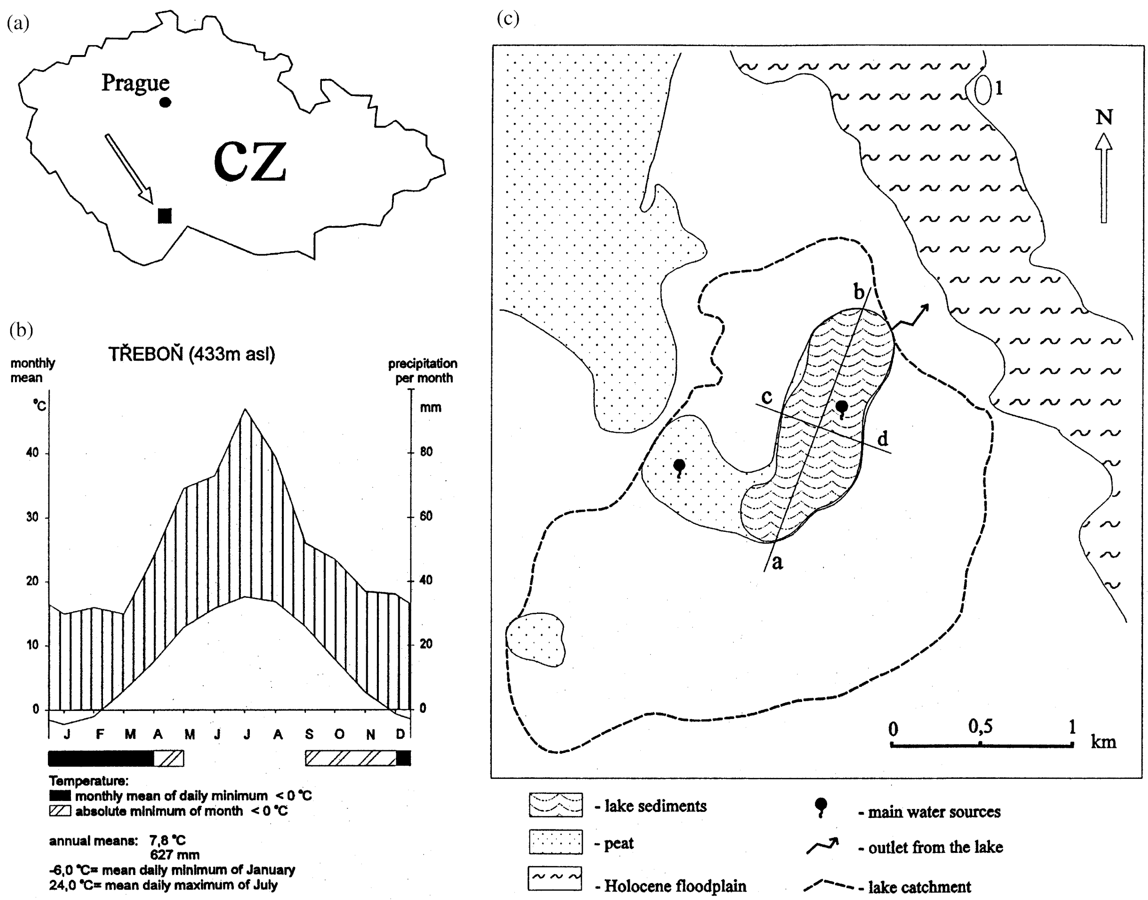


Fig. 1. The study site. (a) Location of the study area within the Czech Republic. (b) Climatic diagram for Třeboň (ca. 15 km south from the site) derived from 50 yr of observations. (c) Quaternary geology and topography of the site. 1—"Vlkovský přesyp" aeolian sand dune, a,b,c,d—two cross-sections used for stratigraphic investigation. Their crossing point is in the centre of the basin, where the "main profile" is situated.

and cannot be used as a climatic indicator. The lake drained into nearby Lužnice River. For evaluating the regional pollen rain on the site, the vicinity to the river floodplain is important.

Numerous aeolian deposits are situated along the river floodplain. One of the biggest and most prominent sand dunes, "Vlkovský přesyp", is situated 1200 m from the former lake basin. Its relative height over surrounding terrain (6 m), its size (60 × 80 m), and the unvegetated character make it a prominent structure on the landscape. According to its morphology, "Vlkovský přesyp" sand dune was formed by easterly winds. The material constituting it is derived directly from the nearby river floodplain (carried by wind for tens or hundreds of meters) according to mineralogical analyses (Chábera, 1982).

The existence of the former Lake Švarcenberk was noted for the first time by Vlasta Jankovská in the late 1970s. In her study, which focused on the vegetational development of Třeboň Basin (Jankovská, 1980), she presents a pollen diagram and macrofossil analysis obtained from an open pit. Her profile comprised about 1.5 m of lake sediments, and she correctly assumed that she dealt with the littoral facies of a larger lake. Unfortunately, no stratigraphic data were obtained. The present study fully confirms Jankovská's original assumption.

## 3. Methods

### 3.1. Field methods, sediment description, and subsampling

During the pilot study, the extension and stratigraphy of the former lake basin was studied by coring in a 100m × 100m grid (sampling distances were reduced along the shores). For subaquatic coring, a boat was used. Two right-angle transects across the basin were chosen as reference sections (shown in Fig. 1c and 2). They have their crossing point in the centre of the basin, where the "main profile" is situated. The distances in metres and the geographical position (N, S, E, W) in relation to this zero point is given by the core labels. The cores were levelled according to fishpond water level. The coring was performed with Russian-type corer

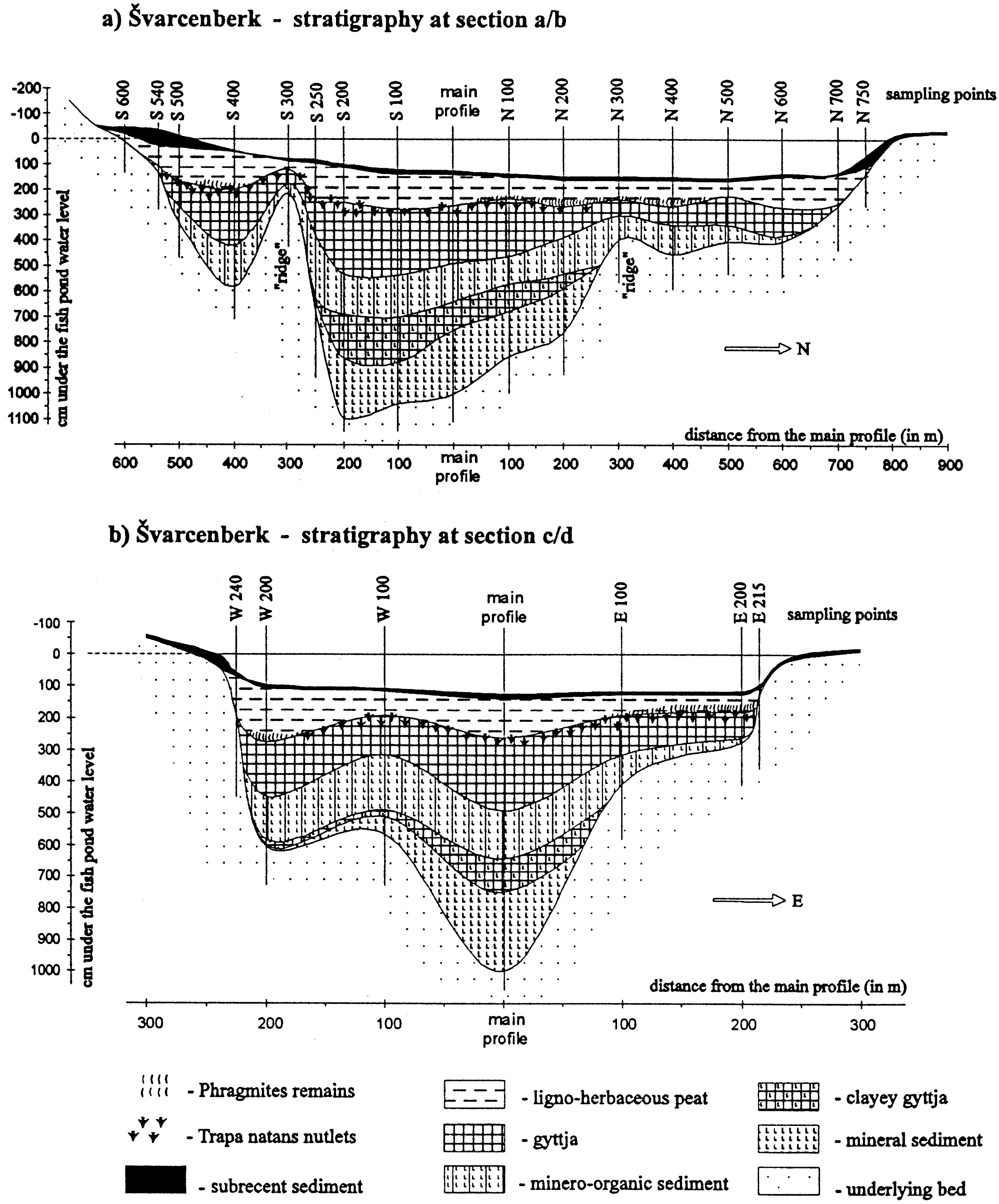


Fig. 2. Two selected orthogonal stratigraphic cross-sections (their position described on Fig. 1) through the Švarcenberk lake basin.

(Jowsey, 1966) 5 cm in diameter. The sampling point S500 was opened by hand-made pit. Correlation of the individual cores across the basin was achieved by visual stratigraphic description and further confirmed by pollen analysis in some cases.

The core in the centre of former lake was selected as the standard profile. The central core will most probably show a continuous record without hiatuses, and it is more likely to give an "average" picture of the events in the basin and its catchment (without the background of local "noise", which is assumed to be greatest along the shores). This "main profile" actually consists of seven separate parallel cores taken close together in order to obtain enough material for all kind of analyses. The coring was performed using the Russian-type corer 5 cm in diameter. This type of device permits complete recovery of the section penetrated. All cores comprising the "main profile" were correlated according to their visual lithostratigraphy. This correlation only confirmed the satisfactory accuracy of parallel sampling.

Sediment description follows the system of Troels-Smith (1955) as modified by Aaby and Berglund (1986). The colours were determined according to standard (Munsell) soil colour charts. The subsampling intervals depended on the required temporal resolution and the sample volume needed for each analysis. Closer subsampling was undertaken in order to get enough material for additional analyses if needed (as advised by Moore et al., 1991). The stratigraphy of "Vlkovský přesyp" sand dune was studied in an open ditch.

### 3.2. Macrofossil analysis

After subsampling for other analyses, the remaining material was used for macrofossil analysis. Contiguous samples 10 cm long were cut, and the volume of each was determined (approximately 250 ml in all cases). Macrofossils were extracted by heating each sample for 5 min in a 5% potassium hydroxide (KOH) solution and sieved with running water. Sieves with mesh sizes of 200, 300 and 700 μm were used. The residues were examined under a dissecting stereomicroscope. The absolute number of each kind of macrofossil was recalculated to a standard volume of 500 cm$^3$ fresh sediment. For determination of the seeds/fruits, a reference collection and the atlas of macrofossils (Kac et al., 1965) were used.

### 3.3. Pollen analysis

The samples used for pollen and other microfossil analyses were prepared by a modified acetolysis method. As the Late-Glacial part of the core had a more or less mineral character, the samples were pre-treated with concentrated (35%) cold hydrofluoric acid (HF) for 24 h (Faegri and Iversen, 1989; Moore et al., 1991). The extracted microfossils were lightly stained by 0.3% safranin and mounted in a liquid glycerol-water (1:1) mixture. In each sample at least 1500 pollen grains were counted, except in the lowermost samples poor in pollen, where only 500–700 grains were identified. For pollen identification, the following keys were used in addition to a reference collection: Faegri and Iversen (1989); Moore et al., 1991; Punt (1976–1996). Pollen nomenclature follows ALPADABA (*Alpine Palynological Data-Base*, housed at the Geobotanical Institute, Bern).

An attempt was made to subdivide *Alnus viridis* and *Betula nana* in the pollen diagram. *Alnus viridis* pollen was found to be relatively easily separable. More difficulties were connected with *Betula nana* pollen identification. As the difference in pollen morphology between *Betula nana* and arboreal birch species is largely quantitative (Birks, 1968; Usinger, 1975), exact separation was impossible in all cases. When uncertainties occurred, problematic pollen was ranked as "*Betula* indet." This category represents typically about 30% of total *Betula* pollen in the Late-Glacial sequence. It is included in the pollen sum but is not shown in pollen diagram. Algae and other microfossils were identified with the help of the publications of Van Geel et al. (1981, 1983, 1989), Jankovská (1983), and Jankovská and Komárek (1982).

Absolute pollen concentrations were estimated by the volumetric method (Davis, 1965). As the volumetric method has been proved by Walker et al. (1994) to give results with a relatively high standard deviation, pollen-concentration determinations were not used for pollen-influx calculations. The pollen-concentration curve has been included only in the main pollen diagram (Fig. 3) to give a rough idea about absolute pollen content of the sediment.

The lowermost Late-Glacial and particularly Pleniglacial sediments contained reworked Tertiary pollen. An attempt at reliable separation of these pollen grains was made in order to avoid possible misinterpretations of Late-Glacial pollen spectra and to obtain a tool for assessment of erosion rates within the lake basin and its catchment. In this approach it is assumed that reworked pollen grains embedded in Late-Glacial and Holocene sediments originated primarily from eroded Tertiary subsoil sediments in the lake catchment, including lake shores. Some of the Tertiary taxa are easily recognised according to morphology (e.g. *Engelhardtia, Carya, Liquidambar*), but some are difficult or even impossible to separate in a morphological basis only. Fortunately, the state of preservation of the Tertiary pollen was relatively poor, giving pollen grains a "ghostly" appearance. It has been assumed that different degree of fossilisation resulted in different chemical composition and physical properties of the exine. Considering this assumption, a simple method was developed to confirm the Tertiary origin of some pollen grains. Pollen preparations, already stained by 0.3% safranin, were bleached in 40% ethanol for 1 min. Then they were centrifuged and transferred to glycerol-water mounting medium again. As a result of this procedure, all pollen grains were bleached to different degrees, but the Tertiary ones became almost completely transparent and hence easily separable from the Late-Glacial ones.

### 3.4. Standard pollen diagram

The selection of types included in the pollen sum is always an important stage in the interpretation of palynological results. Excluded from the sum, therefore, are types potentially produced by the local aquatic and marsh vegetation of the lake. In case of the Late-Glacial

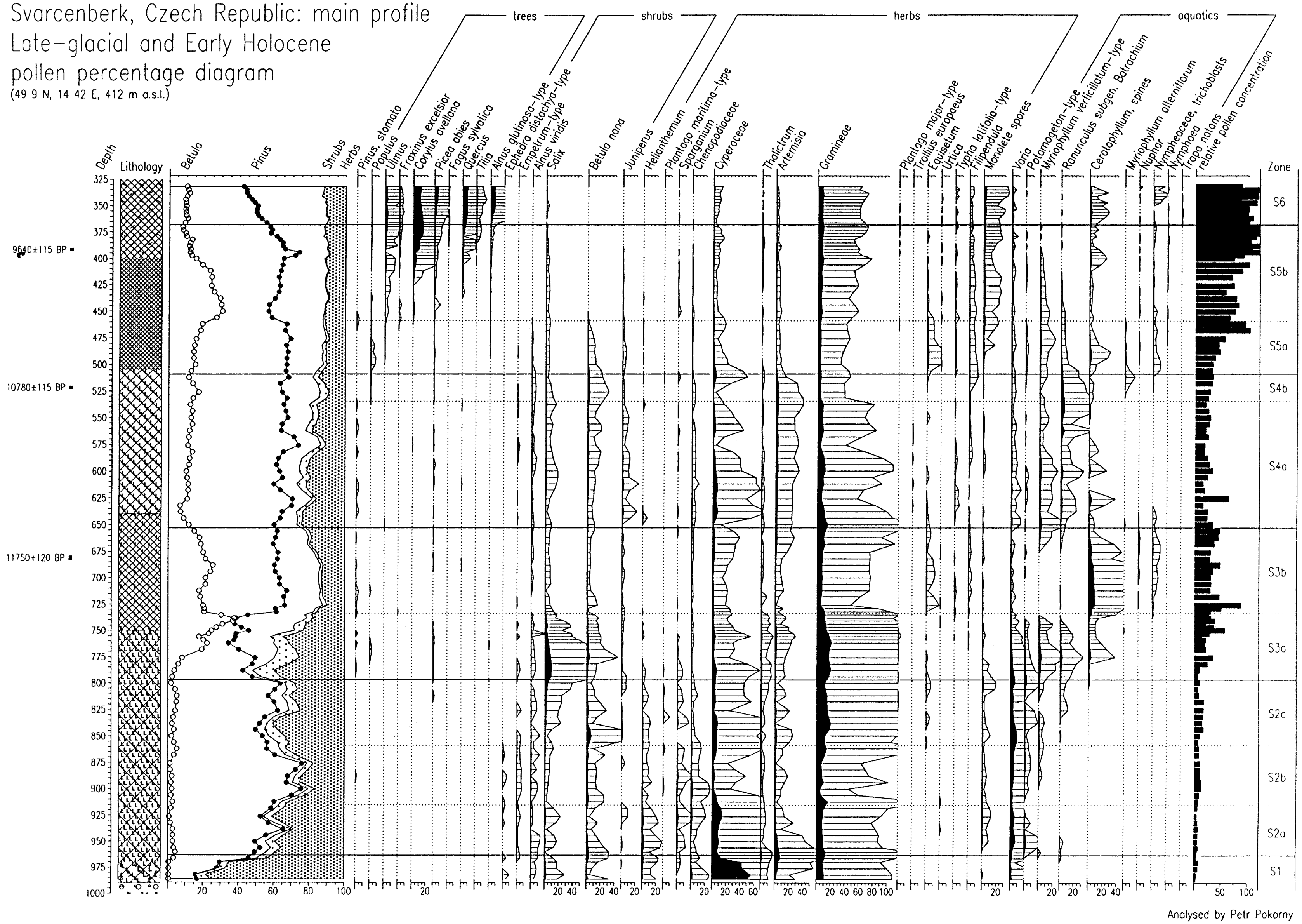


Fig. 3. Late-Glacial and Early-Holocene pollen diagram of the "main profile". Only selected types are included.

however, it is difficult to evaluate exactly the role of certain wetland taxa (e.g. Cyperaceae, *Equisetum, Salix*) in the regional vegetation cover, as we do not have exact modern analogues to the Late-Glacial communities (Watts, 1979; Moore, 1979). Percentage values were therefore calculated on the basis of the AP + NAP pollen sum, excluding only submerged and floating-leaf aquatics but including monolette and *Equisetum* spores (in many plant communities, these taxa usually have an ecological role equivalent to that of higher plants). Concealed, corroded, degraded, and well preserved but indeterminable pollen grains were put together and labelled "varia" in the pollen diagram. Printing of the diagrams was made with the TILIA computer program, written by E. C. Grimm (Springfield, Illinois).

The pollen diagram was zoned visually, on the basis of both presence and abundance of taxa. A more formalised approach to delimit the local pollen assemblage zones was also applied on the basis of three different constrained classification procedures implemented in the computer program ZONE (Lotter and Juggins, 1991). Consistency among results from these three different zonation procedures provided the basis for confirmation and further specification of visually delimited local PAZ (Figs. 4–6).

### 3.5. Sediment chemistry and $^{14}C$ analyses

Numerous palaeolimnological studies have shown how sediment chemical properties may be interpreted in terms of processes acting within the lake and the surrounding catchment (e.g. Na, K, and Mg content in sediments directly reflect the intensity of weathering and erosion). These processes are often directly or indirectly related to climatic parameters, and their understanding may enable further climatic reconstructions (Engström and Wright, 1984; Dearing, 1991).

Total carbon and nitrogen content was determined by combustion at 950°C in pure oxygen, with subsequent conductivity detection of C and N oxides (in Heraeus CHN-Rapid Analyser). Carbonate content was measured by sodium hydroxide titration to neutral pH after dissolution of 0.5 g sample in 0.5 M hydrochloride acid and boiling for 20 min (after Hammarlund and Buchardt, 1996). Total organic carbon content was calculated from the difference between total carbon and carbonate carbon. The elements Ca, Mg, K, and Fe were analysed by atomic emission spectrometry in the Analytical Laboratory of the Institute of Botany, Academy of Sciences of the Czech Republic, with a Unicam 9200X AAS instrument.

Radiocarbon dates used in the present study are all AMS dates determined from bulk sediment samples and individual plant macrofossils (for sample description see Table 1). The disadvantage for the Late-Glacial sequence under the study is the absence of sufficient terrestrial plant macrofossils for dating purposes. The dates from gyttjas, clayey lake sediments, and aquatic plant macroremains are known to give ages that usually exceed those obtained from terrestrial macrofossils (Törnqvist et al., 1992). This effect is often ascribed to the hardwater error. In the present study a hardwater error is expected to be relatively small, as the sediments contain negligible amounts of carbonates. Radiocarbon analyses were carried out by the Radiocarbon Dating Laboratory, Department of Quaternary Geology, Lund, Sweden. The samples have been pretreated with HCl and NaOH. Age calculations are based on a $^{14}C$ half-life of 5568 yr. For the purposes of simplicity and comparability, dates are expressed in uncalibrated $^{14}C$ years before present (BP) unless otherwise stated.

### 3.6. Multivariate data analysis

In an attempt to study the similarities among individual pollen spectra, a detrended correspondence analysis (DCA) was made for a complete Late-Glacial and Early-Holocene data set. Pollen percentages of individual taxa were used as input data for the CANOCO 3.10 (ter Braak, 1990) program. DCA results, plotted as ordination diagrams on DCA axes 1 and 2, show the overall trends in the data (Fig. 7). Similar samples are close together and dissimilar samples are far from each other.

To study the relationship between the composition of individual pollen spectra and sediment chemical composition, a canonical correspondence analysis (CCA) (ter Braak, 1986) was made for the combined data set, including pollen percentages and chemical data. Program CANOCO 3.10 was again used. Plot of the centroid of sediment-chemistry variables in the intersection of CCA axes 1 and 2 was constructed (Fig. 8) in order to summarise the results graphically.

## 4. Results

### 4.1. Lake basin, its origin and stratigraphy

The extent of lake deposits within the basin was mapped in detail from approximately 120 hand borings. The altitude of individual stratigraphic transitions was obtained by relative levelling. The former lake was found to have a maximum surface of 0.51 km$^2$, and the ratio of the surface to drainage basin to be about 1:8 (Fig. 1a). Two lithological cross-sections (Fig. 2) show the morphometry and the infilling of the depression. Correlation of the individual cores across the basin was achieved by visual stratigraphy, which reflects well the environmental conditions during the time of sedimentation (e.g. Younger Dryas sediments are characterised over the entire basin by the presence of three distinct

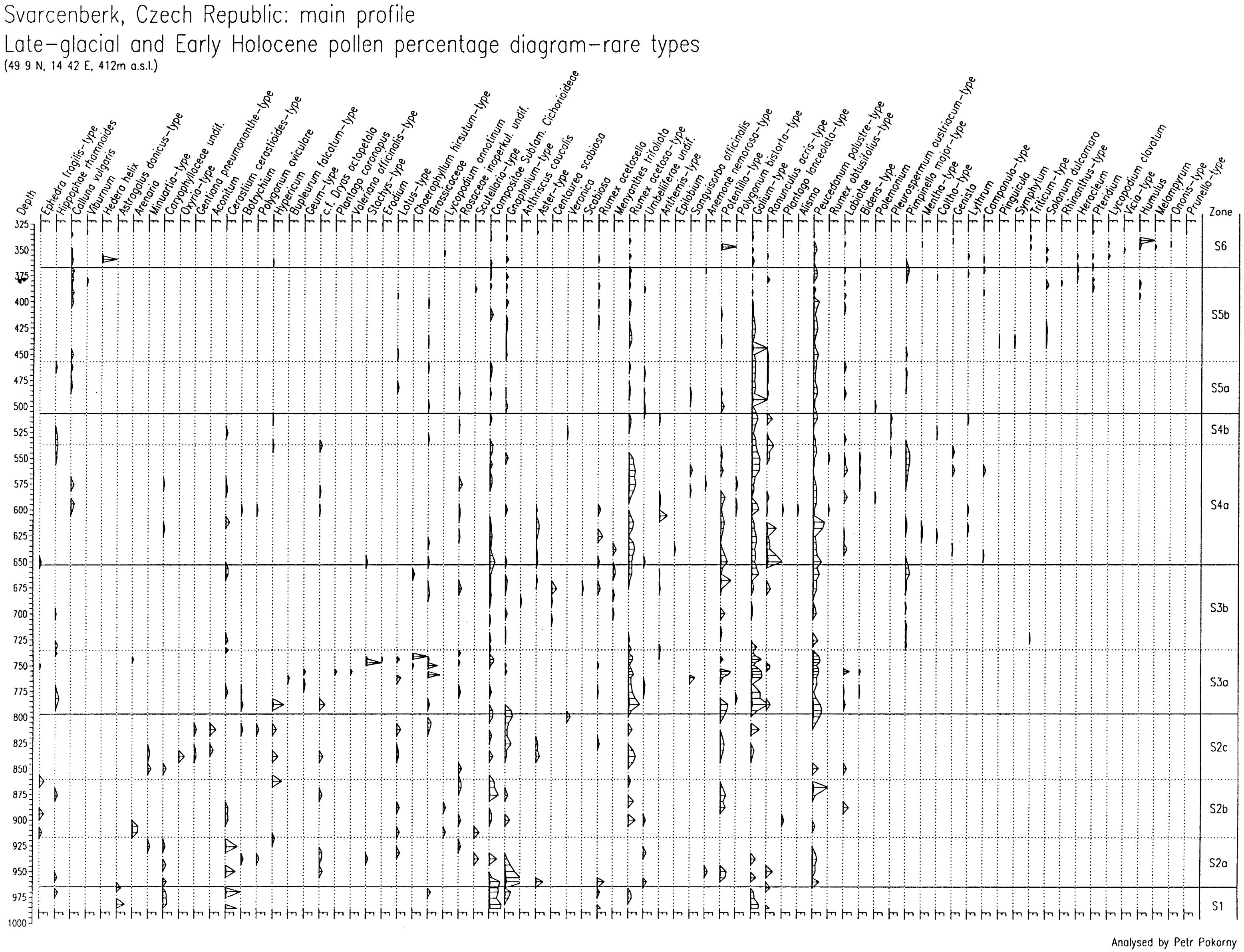


Fig. 4. Late-Glacial and Early-Holocene pollen diagram of the "main profile". Rare pollen types ordered according to results of weighted averaging.

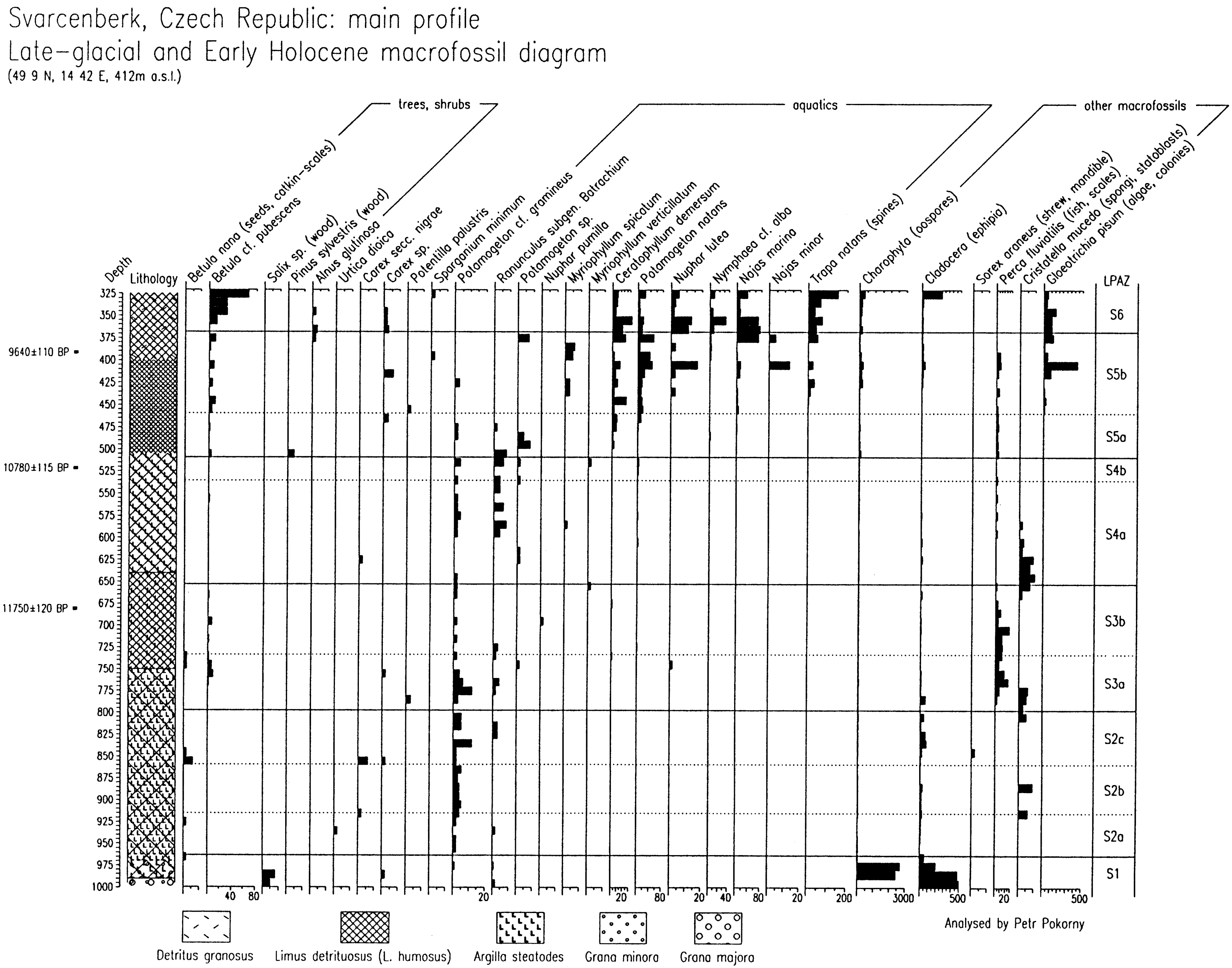


Fig. 5. Late-Glacial and Early-Holocene macrofossil diagram of the "main profile". Absolute numbers of the finds were recalculated to standard volume of 500 cm$^3$ of fresh sediment. If not stated differently, all finds represents seeds.

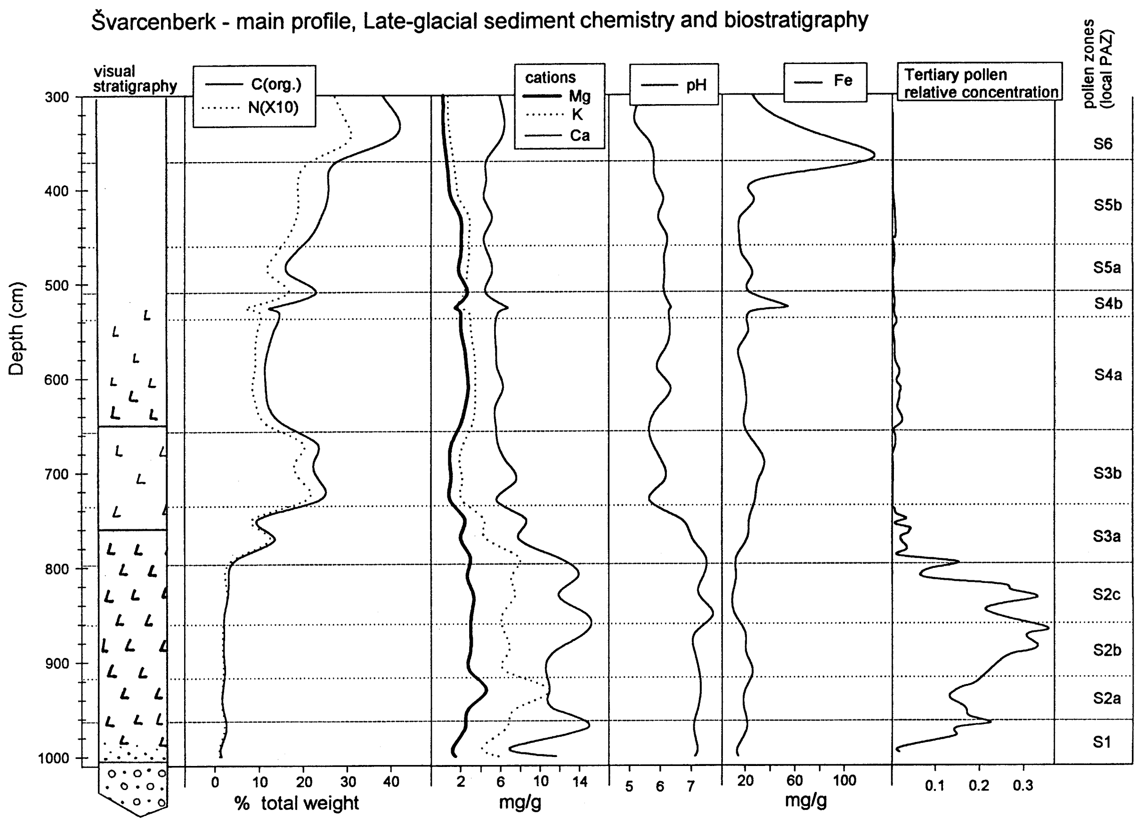


Fig. 6. Sediment composition of the "main profile" correlated with local pollen zonation. Simplified stratigraphic column shows different proportion of sand (dots and circles) and silt (angles) components in organic sediments.

Table 1
AMS radiocarbon dates from Švarcenberk littoral (S500) and central ("main profile") cores and "Vlkovský přesyp" sand dune (LuA-4645)

| Lab. no. | Core label/depth | Method | Type of material | Measured $^{14}C$ age |
|---|---|---|---|---|
| LuA-4297 | S500: 200 cm | AMS | *Trapa natans* nut | 6340 ± 110 BP |
| LuA-4589 | "Main p.": 324–327 cm | AMS | *Trapa natans* nut | 6350 ± 100 BP |
| LuA-4590 | "Main p.": 390–393 cm | AMS | Woody stem fragment | 9640 ± 115 BP |
| LuA-4591 | "Main p.": 520–523 cm | AMS | Bulk gyttja sample | 10,780 ± 115 BP |
| LuA-4738 | "Main p.": 680–683 cm | AMS | Alkali-soluble fraction from gyttja | 11,750 ± 120 BP |
| LuA-4645 | Surface of a fossil soil | AMS | *Pinus* charcoal fragments | 11,260 ± 120 BP |

yellowish layers of aeolian material, each about 1 cm thick). The borders distinguished between several litostratigraphic units can therefore be assumed to be roughly time-parallel. In two littoral sampling points, S500 and JC-7B (the latter made and studied in SW part of the basin by Jankovská, 1980), pollen analysis confirmed this assumption. Only the limno-thelmatic contact, marking the time of final infilling of the lake, was expected to be metachronous over the basin. Radiocarbon dating in the littoral part (sampling point S500, with a date of 6340 ± 110 BP directly at the limno-thelmatic contact) and the central part of the lake (the "main profile", with an almost identical date of 6350 ± 100 BP at a level significantly under the limno-thelmatic contact) confirmed this expectation.

The striking features of the basin morphometry are its kidney-shaped form, surprising depth and declivity (the presence of unusually steep slopes), and relatively great age of its infilling. Unfortunately, no radiocarbon date exists from the basal sediments, but their age is estimated around 16,000 BP from the pollen-analytical results. On the basis of these findings, the origin of such structure can be best explained as the remnant of a huge Pleniglacial ground-ice lens—an open-system pingo.

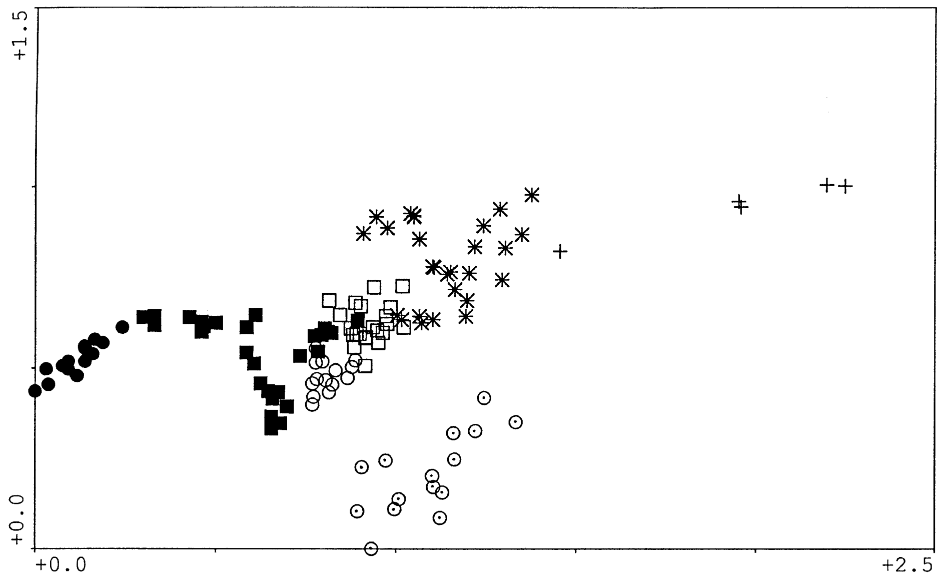


Fig. 7. DCA ordination results of percentage pollen data from the "main profile". The use of the symbols follows pollen zonation (LPAZ): Full circles—zone S6, full squares—zone S5, open squares—zone S4, open dotted circles—subzone S3a, open circles—subzone S3b, stars—zone S2, crosses—zone S1.

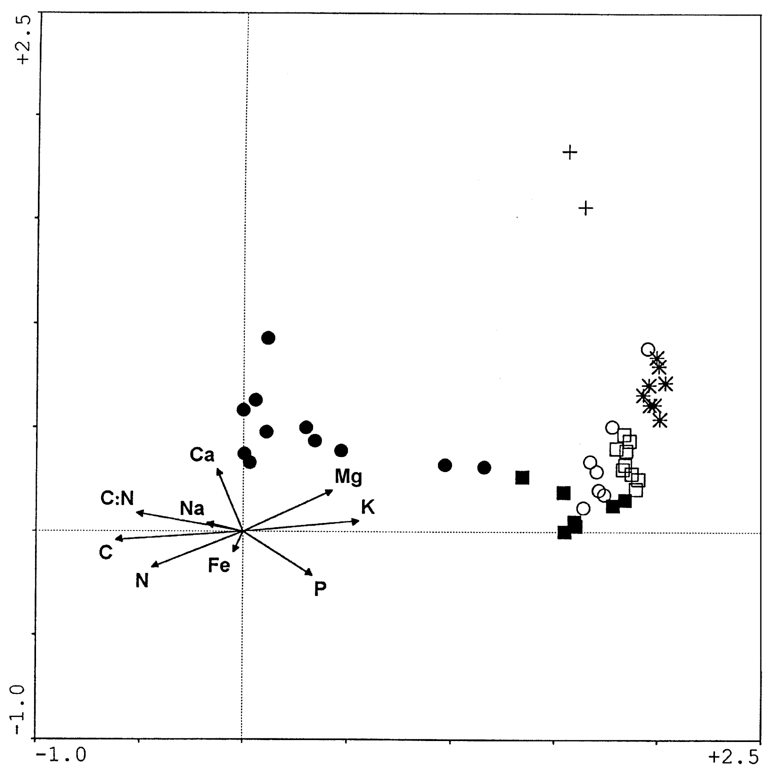


Fig. 8. CCA ordination results. Percentage pollen data together with sediment chemistry data (as environmental variables) have been used. The use of the symbols follows pollen zonation (LPAZ): Full circles—zone S6, full squares—zone S5, open squares—zone S4, open circles—zone S3, stars—zone S2, crosses—zone S1.

The absence of a distinct rampart around the former lake is not surprising, as the original geomorphology of the site was completely disturbed by human action, especially during the construction of the fishpond in the 17th century. Moreover, some pingos do not dome very high over the terrain, although they are relatively large in diameter (Washburn, 1980). These structures do not form a distinct rampart after they collapse. What is more surprising in case of former lake Švarcenberk is the unusually big size of its depression. Considering this fact and the observation of the "ridges" (Fig. 2) dividing the basin into three main parts, the origin of the lake can be best viewed as the remnant of some kind of a compound pingo structure.

In the littoral parts of the former lake basin, only a thin layer of Late-Glacial sediments is present, completely lacking deposits older than Younger Dryas. This can be explained by intensive reworking of the shores during Late-Glacial rather being the result of lower lake levels during this period. As the lake was fed almost exclusively by artesian water, water-level remained constant over the entire period of its existence, and water-level reconstructions cannot be used as a climatic indicator, unfortunately. After the final infilling of the lake (dated to approx. 5500 BP), oligotrophic peat started to accumulate.

### 4.2. Main biostratigraphic events and geomorphic processes

Pollen stratigraphy of the "main profile" has been subdivided into six local pollen assemblage zones (PAZ) and eleven subzones, which are used as a framework for the discussion of the results. Because of terminological problems (e.g. Ammann and Lotter, 1989; Walker, 1995) I have decided to subdivide the diagram in this way rather than into Firbas pollen zones (Firbas, 1949), as is traditionally done in Central Europe. The absence of analogous results over a wide region discourages the use of regional pollen zonation. The local PAZ are later compared (Fig. 9) with European climatostratigraphical units according to Mangerud et al. (1974) and Ammann and Lotter (1989) and with the $\delta^{18}$O curve of the Greenland ice core GISP2 (Stuiver et al., 1995).

#### 4.2.1. Zone S1

The lowest sediments of the lake-basin sequence consist of fine silt with coarser sand. Absolute pollen concentrations in the sediment are low. The zone is characterised by high NAP values, suggesting open herbaceous vegetation. Grasses, Cyperaceae, Chenopodiaceae, *Betula nana, Alnus viridis, Salix* (most likely some dwarf willow species), *Thalictrum*, and *Artemisia* were important components of the vegetation. Macroscopic stem fragments of *Salix* sp. were found in the sediment. Sporadic pollen finds of *Ephedra* (both *Ephedra distachya* and *E. fragilis* types) are difficult to interpret, as this type of pollen can be dispersed over long distances (Huntley and Birks, 1983; Lang, 1994). *Pinus* values are below 30%, and *Betula* values do not exceed 5%; both can be ascribed to long-distance transport. *Helianthemum* is indicative of bare,

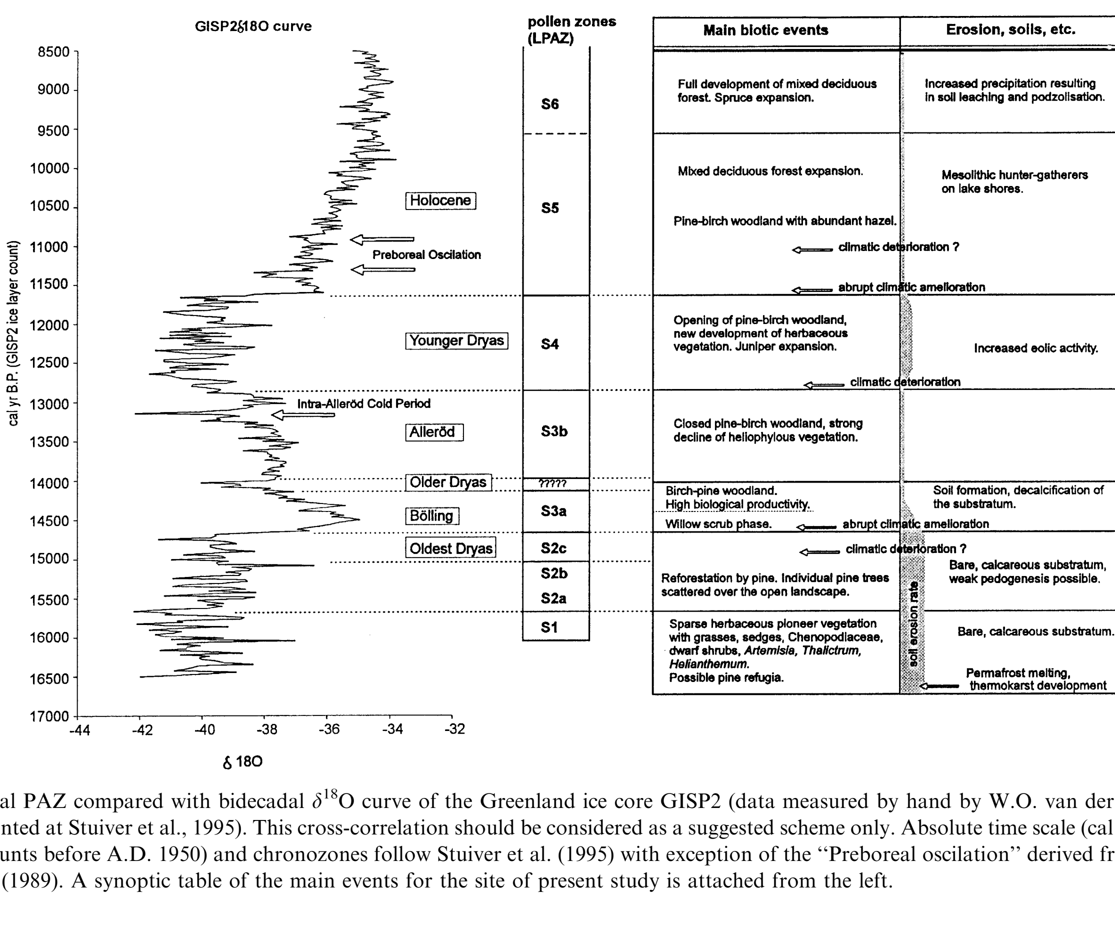


Fig. 9. Local PAZ compared with bidecadal $\delta^{18}$O curve of the Greenland ice core GISP2 (data measured by hand by W.O. van der Knaap from graph presented at Stuiver et al., 1995). This cross-correlation should be considered as a suggested scheme only. Absolute time scale (cal yr BP; yearly ice-layer counts before A.D. 1950) and chronozones follow Stuiver et al. (1995) with exception of the "Preboreal oscilation" derived from Ammann and Lotter (1989). A synoptic table of the main events for the site of present study is attached from the left.

calcareous substratum (Hoek, 1997). The abundance of Cyperaceae pollen in relation to *Artemisia* pollen indicates prevailing moist conditions for zone S1. This could be the result of the presence of permafrost and its progressive decay from the surface. Pollen of submerged water plants is absent, and the seeds of *Ranunculus* subgen. *Batrachium* and *Potamogeton* cf. *gramineus* are rare, while Charophyta oospores (cf. *Chara strigosa*, a pioneer species with subarctic distribution) are exceptionally abundant. *Ranunculus* subgen. *Batrachium* is a climatic indicator for maximum July temperatures of at least 10°C, while *Hippopha rhamnoides* presence (found as pollen) suggests at least 11°C (Huizer and Izarin, 1997). Due to the absence of plant macrofossils suitable for $^{14}C$ dating and the mineral character of the sediment, absolute dating of this zone unfortunately was not possible. The age is only roughly estimated to be around 16,000 BP.

Very low organic content of the sediments suggests low productivity in the lake as well as in its catchment. The lower Mg, Ca, and K content, if compared to the S2 zone, is most likely the result of high proportion of coarse sediment (and high accumulation rates), rather than lower allogenic sediment input to the lake basin.

All the results point to severe climatic conditions, the absence of well-developed soils, and an open, treeless vegetation of steppe and tundra during S1. A significant difference in the character of the vegetation between this and the higher zones was confirmed also by DCA results (Fig. 7), for the samples of the S1 zone are plotted in the ordination diagram far from the main population of samples, indicating significantly different composition of their pollen spectra.

#### 4.2.2. Zone S2

This zone starts with a rise in *Pinus* pollen percentages up to 50%, together with a decrease in *Cyperaceae* and a prominent *Artemisia* peak. Pollen of submerged water plants occur for the first time. *Betula* pollen curve is continuous during the entire zone, but is too low (under 5%) to suggest the local occurrence of tree birch. As in the previous zone, grasses, Cyperaceae, Chenopodiaceae, *Alnus viridis,* dwarf willow*, Thalictrum, Artemisia,* and *Betula nana* were important components of the vegetation. Seeds and catkin-scales of dwarf birch (*Betula nana*) were found in the sediment. *Empetrum* pollen percentages around 1% suggest local presence but not the development of extensive *Empetrum*-dominated heaths (according to quantitative criteria given by Huntley and Birks, 1983). Relatively high *Helianthemum* percentages (up to 3%) point to the presence of calcareous substratum. This accords well with the results of sediment chemical analyses, showing high Ca content. This finding is in sharp contrast with present-day conditions in the area under study, for most soils are leached, highly acidic, and almost completely lacking calcium carbonate. The absolute pollen concentrations in the sediment are still low during this zone.

The silty FeS-coloured sediments in the centre of the basin suggest anoxic conditions, as iron-sulphide deposition usually occurs under prolonged or permanent stratification of the lake (Engström and Wright, 1984). Sediment organic content is only around 3% and reflects low productivity in the lake and its catchment. High content of Na and K and of reworked Tertiary pollen suggest high erosion rates and the absence of stable soils in the catchment.

#### 4.2.3. Subzone S2a

The rise in *Pinus* percentages up to values around 60% suggests the onset of pine expansion in the area as the result of some climatic amelioration. Individual pine trees must have been scattered more or less sporadically in the landscape of steppe or tundra-like character. The rise in *Artemisia* shortly before this time can be also considered as the first signal of warming. The find of *Urtica dioica* seed (together with *Urtica* pollen) is particularly interesting from the point of environmental reconstruction, for this nitrophillous plant, requiring mean July temperatures at least 8°C (Bos, 1998), is distributed in eutrophic habitats today. This is in sharp contrast to the general picture of the S2 subzone, which points to prevailing oligotrophic and distrophic conditions in the lake and its catchment (see also the low sedimentary N content). There must have existed some favourable, nutrient-rich microhabitats (bird-manured patches, for example), where *Urtica dioica* could have prospered.

#### 4.2.4. Subzone S2b

The signs of progressive climatic amelioration are traceable from the pollen record during the onset of S2b subzone as *Pinus* values rise up to 75%. The local presence of pine is confirmed also by the find of *Pinus* stomata. Pine became a successful competitor for light with the heliophyllous vegetation. Overshading occurred in moist habitats, as seen from the decline of *Salix* and Cyperaceae pollen percentages, but also in drier habitats, as *Artemisia* and Gramineae declined during *Pinus* maxima as well. If we compare the pollen concentration curve (attached to the pollen diagram in Fig. 3) for S2b subzone with the others, particularly S3 and S4, zones, it become obvious, that the rise in *Pinus* is not due to the percentage effect caused by generally low regional pollen production (Moore et al., 1991). The dating of this subzone is unfortunately uncertain, due to the absence of appropriate material for $^{14}C$ analysis again. It is only roughly estimated to range between 15,000 and 14,000 BP.

#### 4.2.5. Subzone S2c

The decline of pine and new expansion of heliophyllous herbs in this subzone can be correlated with the Oldest Dryas chronozone (*sensu* Stuiver et al., 1995). Unfortunately, exact $^{14}C$ dating of this event is again missing. In the upper half of S2c subzone, a new pine expansion anticipates the transition to zone S3. An interesting macrofossil find, a shrew (*Sorex* cf. *araneus*, det. by I. Horáček) mandible, was made within this subzone. The mandible comes from an adult specimen, but its size is unusually small. Its phenotypic appearance significantly differs from contemporary populations. Shrews were in general rare constituents of late Pleistocene faunal assemblages. This find is particularly interesting from the palaeoecological point of view, for shrews are characteristic of sparsely wooded landscapes (I. Horáček, pers. comm.), consistent with the palaeobotanical results that suggest sparse tree cover during that time.

#### 4.2.6. Zone S3

Marked vegetation changes are characteristic for this zone. Reforestation of the landscape by birch and pine resulted in decline in heliophyllous herbs. Absolute pollen concentrations in the sediment were about three times higher than in the underlying zone. Soil development under forested conditions led to the decrease in erosion rates (see progressive decline in sedimentary Mg and K and sharp decline in reworked Tertiary pollen). Decalcification of the substratum continued up to the maximum extent (see the decline in Ca down to the values comparable with those in the Holocene). Increased organic production resulted in gyttja sedimentation. Abundant perch (*Perca fluviatilis*) scales were found in the sediment. Perch is an anambitious fish genus that can survive even in subarctic lakes. The fry are produced in large quantities and feed on planktonic or benthic organisms. Adults feed mostly on their own young and reach high population densities. Such an interesting cannibalic food chain has been described from several contemporary Siberian lakes (Holčík, 1977; Karasev, 1987).

#### 4.2.7. Subzone S3a

This subzone starts with a rise in *Salix* pollen percentages, followed by *Betula* increase. *Salix* probably formed the shrub belt in front of the *Betula* forest-line (Gaillard, 1985; Hoek, 1997), so high values of *Salix* may be expected in advance of a *Betula* expansion. In the second half of S3a subzone, tree birch became dominant in the pollen spectra in place of NAP and *Pinus*. *Betula* cf. *pubescens* seeds and catkin-scales were found in the sediment. This assemblage reflects the development of open boreal birch woodland with dispersed pine trees. The lower limit of mean July temperature needed for tree birch colonisation is usually taken as 10°C, but 12°C is the optimum for the development of *Betula pubescens* woodland (Birks, 1993). An open character of the forest can be inferred from the pollen diagram, for most of the heliophyllous herbs typical for steppe and tundra communities are still abundant. In the aquatic environment, abrupt climatic amelioration caused the expansion of submerged macrophytes, including *Ceratophyllum demersum*, which appears for the first time in this subzone. The S3a subzone is correlated with the Bölling chronozone.

#### 4.2.8. Subzone S3b

The basal spectrum of this subzone coincides with a strong decrease in Betula from 40% to about 20% and a new increase in *Pinus* up to values about 60%. The forest cover became more closed, as reflected by a decrease in all open-communities indicators. *Helianthemum* and *Plantago maritima*-type disappeared from the spectra completely. *Filipendula, Typha latifolia, Nymphaea*, and *Nuphar* appeared in and around the lake, pointing to minimum July temperatures, at least 12°C (Huizer and Izarin, 1997). *Ceratophyllum demersum* became the dominant submerged aquatic. Gradual *Betula* decrease is recorded during the upper half of this subzone. S3b is correlated with the second half of the Late-Glacial interstadial—the Alleröd chronozone. This correlation has been confirmed by the radiocarbon date 11 750 $\pm$ 120 BP.

#### 4.2.9. Zone S4

The values of *Betula* decrease to about 10%, whereas *Pinus* percentages are generally the same as in the preceding zone. *Alnus viridis, Salix, Betula nana,* Chenopodiaceae, and *Artemisia* values increase again, suggesting a new climatic deterioration. A similar increase, but even more prominent, is recorded in the *Juniperus* curve. Juniper percentages reach a maximum in this zone. Vegetation change reflecting climatic deterioration is recorded also in the lake: *Ceratophyllum* spines values decrease, and Nympheaceae trichoblasts and *Nuphar* pollen are completely lacking in favour of a large amount of *Myriophyllum verticillatum* and *Ranunculus* subgen. *Batrachium* (both recorded as macrofossils). The presence of *Typha latifolia* pollen in the entire zone can be considered the proxy for minimum July temperatures of at least 12°C (Iversen, 1954; Ammann, 1989), i.e. at the same range as those inferred for the preceding zone. This suggests that climatic deterioration was the result of an increase in continentality rather than a decrease in summer temperatures. Absolute pollen concentrations are lowered relatively to S3 zone, and the sedimentation character changes to more minerogenic again (with lower organic carbon content). Slight increases in erosion indicators (Mg, K, reworked Tertiary pollen) are observed as well.

Although clear evidence of climatic deterioration (manifested most likely as the increase in continentality) is found in this zone, this climatic oscillation did not result in complete deforestation. The pine woodland only became somewhat more open. Zone S4 is correlated with the Younger Dryas chronozone, and this correlation is confirmed by the $^{14}$C date 10 780 ± 115 BP obtained from the position slightly below the upper zone limit. It is obvious from DCA ordination diagram (Fig. 7) that climatic deterioration led to vegetational reversion, as can be concluded from the shifted position of the samples along the main ordination axis (*x*-axis in the diagram).

The stratigraphical investigation of "Vlkovský přesyp" sand dune has revealed a fossil soil buried under more than 5 m of aeolian sands. A distinctive layer of pine charcoal fragments buried by aeolian sands is dated 11,260 ± 120 BP and implies that the formation of "Vlkovský přesyp" sand dune dates to very beginning of theYounger Dryas period.

#### *4.2.10. Subzone S4a*

This subzone is quite uniform, with only slight fluctuations in *Pinus* values and some more or less synchronous fluctuations in Cyperaceae and Gramineae that are difficult to interpret.

#### *4.2.11. Subzone S4b*

Slightly higher values of *Betula nana, Artemisia*, and *Ranunculus* subgen. *Batrachium* mark the lower limit of the subzone. When these herbs subsequently decrease, the first possible evidence of climatic improvement is recorded by the increase in *Filipendula* and onset of *Populus* rational limit.

#### *4.2.12. Zone S5*

The transition from S4 to S5 zone is characterised by a decrease in all NAP taxa, especially *Alnus viridis, Betula nana*, Chenopodiaceae, and *Artemisia* among the most important. At the same time, *Betula, Populus, Filipendula*, and *Equisetum* percentages rise, while those of *Salix* decrease. Most of the thermophyllous trees have their empirical limits (*sensu* Faegri and Iversen, 1989) at the beginning of this zone, and the development of mixed deciduous forest is observed during its second half. *Picea abies* already occurs in low percentages much earlier (during the entire Late-Glacial), but these early finds can be ascribed to long-distance transport of easily dispersed pollen rather than local occurrence. Organic production generally increases, and the erosion rate is low. Organic sediment (gyttja), rich in macrofossils, accumulated again in the basin. This zone is correlated with the beginning of the Holocene, the Preboreal chronozone.

#### *4.2.13. Subzone S5a*

The peak of *Populus* and *Equisetum* and relatively high *Filipendula* values are characteristic for this zone, pointing to the development of wet meadows. *Populus tremula* may have been favoured as a pioneer tree in areas that were left open during the preceding period (Ammann et al., 1994). *Alnus viridis* and *Betula nana*, the most characteristic tundra elements surviving from the preceding zone, were gradually outcompeted by developing forests, and their values fall to zero. As in the preceding period, pine was still dominant in the regional forest cover, but during the transition to S5b subzone birch started expanding in place of *Pinus*. The short decrease in organic production at 480 cm cannot be attributed to any climatic oscillation without exact time control.

#### *4.2.14. Subzone S5b*

Boreal forest dominated by birch was slowly replaced by mixed deciduous forest in this subzone. This change is reflected by the gradual decrease in the *Betula* curve to 10%. On numerous sites with unfavourable sandy soils, *Pinus* stands still occurred, being favoured in competition with arriving thermophyllous trees. Pine forest persistence could also be attributed to prevailing continental character of the climate. Numerous macrofossil finds of *Najas marina, Najas minor*, and *Trapa natans* are dated to about 9800 BP (calculated by linear extrapolation from the two adjacent dates), permitting climatic reconstruction for this period. *Najas marina* suggests a mean July temperature above 15°C (Lotter, 1988), and *Trapa natans* even more. According to Gams (1926) and Jorga et al. (1982), water chestnut requires mean July water temperature not below 20°C and in May, when the flowers develop, at least 12°C. The rapid change to warmer climatic conditions is also evidenced by appearance of macroscopic colonies of the thermophyllous blue-green alga *Gloeotrichia pisum* (Van Geel et al., 1989).

Possible palynological evidence for the short Preboreal cold climatic oscillation was found, but it is weak: a short *Pinus* peak accompanied by the fall in *Corylus, Ulmus*, and *Quercus* percentages is dated slightly before 9640 ± 115 BP. Another explanation may be the opening of the vegetation by fire, but this hypothesis was not confirmed by higher presence of charcoal particles in the corresponding layer.

#### *4.2.15. Zone S6*

Full expansion of mixed deciduous forest started in this zone. The pollen curve of *Pinus*, hitherto very high, started decreasing as the result of competition with deciduous trees even on poor substrates, where soils had developed since the beginning of the Holocene. Alder carr developed on infilling margins of the lake, as reflected in the strong rise of *Alnus glutinosa* pollen curve and the occurrence of alder wood and seeds in the littoral sediments (correlated with the main profile by pollen analysis and $^{14}C$ dating). The development of highly productive alder stands around the lake probably contributed to increased organic deposition (increased organic carbon content).

At the beginning of the zone, a prominent Fe peak dated to about 8600 BP may be explained as the reflection of intensive leaching caused by a more humid climate in conjunction with the build-up of raw humus on the soil surface (Engström and Wright, 1984; Starkel, 1990). In the pollen diagram, the same period is characterised by *Picea abies* expansion. The lower boundary of zone S6 can be correlated with the beginning of the Boreal period.

## 5. Discussion

### 5.1. Initial warming

The origin of the former lake Švarcenberk can be best explained as the remnant of a huge Pleniglacial ground-ice lens, an open system pingo. A similar thermokarst origin has been suggested for several semicircular depressions in The Netherlands, Belgium, France, Germany, and Poland (Washburn, 1980; de Gans, 1988; Hoek, 1997). The sandy geological substratum, the presence of several strong artesian springs on the site, and its location close to the river are factors known to be favourable for pingo formation (de Gans, 1988; Pissart, 1988). The occurrence of such thermokarst phenomena has certain climatic significance and may be used for mean annual air temperature reconstruction, suggesting this to be $-1°C$ or lower (Mackay, 1988). The presence of a pingo remnant with a depth of almost 12 m indicates the minimum permafrost thickness. By ca. 16 ka BP, advanced warming is recorded in the eastern Alps, for at least some glaciers had receded more than two-thirds of their original LGM length (Lundqvist and Saarnisto, 1995). In the Swiss Alps, a very fast decay of glacier ice must have occurred between about 18 and 15 ka BP (Ammann et al., 1994). At about that time, North American and Scandinavian ice sheets were in full retreat (Lundqvist and Saarnisto, 1995; Tyráček, 1995). Thermokarst lakes are usually the first indicators of climatic amelioration in the periglacial zone (Lundqvist and Saarnisto, 1995). The basal age, estimated to be around 16,000 BP for the bottommost sediments of the Švarcenberk central core, represents the minimum possible age of the lake that originated by permafrost thawing. This suggests the change from high-arctic to somewhat warmer conditions. The vegetation cover during this time can be reconstructed from the pollen spectra. Treeless vegetation of steppe and tundra character prevailed in the area.

*Pinus* percentages ranging between 60% and 75% are characteristic for S2 local PAZ. Relatively straightforward evidence of climatic amelioration is present, especially in the S2b subzone, when *Pinus* percentages reach their maximum. Pine by then was a successful competitor for light against the heliophyllous vegetation, for indicators of certain open communities declined. Reconstructed vegetation cover during the S2b subzone can be characterised as a mixture of shrub-heath and steppe patches with scattered pine trees. From this point of view, the macrofossil find of a shrew (*Sorex* cf. *araneus*) mandible in the lake sediments is particularly interesting, for shrews are characteristic of sparsely wooded landscapes (I. Horáček, pers. comm.).

High *Pinus* percentages recorded at Švarcenberk during the pre-Bölling period notably exceeded those found elsewhere in Europe at that time and suggest the local occurrence of *Pinus* stands. Huntley and Birks (1983) conclude that pollen values $>50\%$ indicate local dominance of pine. Ammann et al. (1994) suggested the Oldest Dryas *Pinus* percentages around 20% originated from long-distance transport. The increase of *Pinus* from 20% to 65% is defined as the rational limit by Ammann and Lotter (1989) and Lotter et al. (1992). Surface pollen samples of the woodland-steppe ecotone in Inner Mongolia suggest *Pinus* pollen percentages exceeding 70% indicate dense local pine woodland (Liu et al., 1999). In the ecotone itself and in the edge of the steppe zone these values decrease to 30% or lower. According to Poser (1948) the poleward limit of forest in Europe approximates the $10°C$ isotherm for July (inferred minimum July temperatures for S2 zone are $11°C$). In the Alpine region, and also elsewhere in western Central Europe, pine expands about 13,000 BP or even later in the Alleröd chronozone (e.g. Watts, 1979; Hoek, 1997; Bos, 1998). This pine expansion is not accompanied by other indicators of climatic warming and therefore is considered to be caused by lagged immigration (Gaillard, 1984, 1985). The early pine expansion can be ascribed to the rapid response of *Pinus* populations expanding from locally present glacial refugia shortly after climatic amelioration. Favourable mesoclimatic conditions in the marshy area of Třeboň basin (high local humidity) might have played a significant role in early reforestation as well. In continental parts of Central Europe, *Pinus sylvestris* and possibly some other demanding species might have persisted locally through the entire glacial maximum. Pine is known to tolerate and even reproduce under extremely severe climatic conditions (usually in dwarf forms), either dry, windy, or cold. Unfortunately, it is usually not possible to discriminate between a pollen peak caused by a population invading from another region and one caused by local expansion of the species previously present but climatically little favoured

(Watts, 1979). This is true mainly in the case of pine pollen transported from a long-distance.

Unfortunately, exact dating of the period of first pine expansion is missing, although it certainly antedates the Bölling chronozone and ranges approximately to the period between 15,000 and 14,000 BP. This result is consistent with similar evidence from southwestern European pollen records (de Beaulieu and Reille, 1984; Jalut et al., 1992; de Beaulieu et al., 1994) and the findings from northwestern Norway (Vorren et al., 1988; Alm and Birks, 1991), where early signs of climatic amelioration occur around 15 ka BP. In northern Germany (Meiendorf Interval) and The Netherlands (Epe Interval), there is also evidence for some short-lived warmer oscillation around 15 ka BP (Menke, 1968; Kolstrup, 1980). At about that time, weak traces of initial pedogenesis in the eastern part of Central Europe indicate a short break in loess deposition during the period of slightly warmer and wetter climate (Tyráček, 1995).

The terminal phase of S2 zone, reflected in pollen diagram from Švarcenberk as the period of new *Pinus* decline and a new expansion of open-communities indicators, can be interpreted as the episode of certain climatic deterioration and can be correlated with the Oldest Dryas chronozone (sensu Ammann et al., 1994; Stuiver et al., 1995). In the Alps as well as in Scandinavia and North America, the Oldest Dryas is known to be a period of temporary glacier readvance (Lundqvist and Saarnisto, 1995).

High values of sedimentary Mg, K, and Ca during the entire pre-Bölling period may be explained as derived from eroding, unstable soils. Low nutrient status of the lake and its catchment together with low productivity (N and organic C values are very low during that time) were primarily caused by low energy input into the ecosystem. The local PAZ boundaries recognised in the pre-Bölling record from lake Švarcenberk are compared (in Fig. 9) with $\delta^{18}O$ curve of the Greenland ice core GISP2 (Stuiver et al., 1995). A certain degree of correspondence exist between these two.

### 5.2. Late-Glacial interstadial

An abrupt warming is recorded in the areas adjacent to the North Atlantic around 13 ka BP, during the Oldest Dryas–Bölling transition (Lowe et al., 1994). Reforestation by birch and later by pine is recorded over most of NW and Central Europe during that time. Closing of the immigrating forest canopy is usually anticipated by the period of *Juniperus* expansion. Juniper was an important pioneer shrub after a period of prevailing herbaceous vegetation, and its phase is prominent especially in the Alpine region and in Britain. Farther to the east, this shrub appears to be much less significant in the Late-Glacial vegetation (Huntley and Birks, 1983). In the area under study, juniper plays only a minor role, for no *Juniperus* peak (or only an ambiguous one) occurs during the Oldest Dryas–Bölling transition. Instead, a prominent *Salix* peak characterises this transitional phase. Local edaphic conditions (prevailing waterlogged soils) may have played a role, decreasing the importance of juniper and favouring moisture-demanding pioneer willow shrubs. The species represented by the pollen are not known due to the lack of *Salix* macrofossils. Gaillard (1985) suggested that *Salix caprea* may have grown as a pioneer species before expansion of birch woodland. The *Betula nana* peak accompanying that of *Salix* is most likely also the result of climatic improvement, reflecting shrub-heath development prior to the closing of forest canopy. This resembles the *Betula nana* phase during the very beginning of the Late-Glacial Insterstadial in the Swiss Alps (Lotter et al., 1992; Ammann et al., 1994). The response of locally present species to climatic amelioration precedes the immigration of species and shows no lag-phase.

In the second half of the S3a subzone, tree birch became dominant in the pollen spectra in favour of NAP and *Pinus*. *Betula*-dominated forest developed, outcompeting most of the heliophyllous plant communities. *Betula* and *Pinus* pollen percentages are almost the same, indicating the diminished role of pine in fully developed Interstadial forests. Shortly after, a marked transition is recorded in the pollen diagram: *Betula* percentages suddenly decrease again in favour of *Pinus*, pointing to the change of proportion between these two in the regional forest cover. Heliophyllous herbs generally decrease to their Late-Glacial minima during this *Pinus* phase, indicating that this vegetation change represents a progressive closing of the forest canopy. *Pinus* expansion may indicate increasingly severe conditions, particularly in winter, i.e. an increase in continentality (Walker, 1995). This event can be correlated with the Bölling–Alleröd transition, which appears to be abrupt in the pollen record. No distinct transitional phase attributable to Older Dryas is present. If this oscillation occurred in the area under study, it had only a small impact on the vegetation. A short-lived climatic deterioration for the Older Dryas is generally recognised in central and western Europe (e.g. Walker, 1995), although in some sites its recognition is out of temporal resolution of the analyses or out of the climatic threshold of plant communities involved. It is well known, that the response of a plant population to climatic change is likely to be greatest near the margin of its tolerance (Watts, 1979). For example, in Switzerland the Older Dryas (often correlated with the "Aegelsee oscillation" in this region) is not detectable at low-lying sites but is more apparent at higher altitudes (above ca 600 m a.s.l.), where the climatic limits of indicator species have been crossed (Lotter et al., 1992; Ammann et al., 1993). During the Late-Glacial Interstadial, marked climatic gradients developed over Europe, with temperature differences as much as 6–7°C within a few hundred kilometres (Lowe et al., 1994). Under these conditions, the response of vegetation to climatic changes must have been very diverse in relation to geographical position of the site.

The Late-Glacial Interstadial appears to be a period with significantly increased organic production, as reflected in the sharp transition from minerogenic to organic sedimentation (with high organic carbon and nitrogen content in the sediment). The declining values of Mg and K in the Interstadial sediments are the result of the formation of clay minerals in the soil horizons, as soils developed progressively in the lake catchment. During episodes of relatively stable soils, deep weathering of mature soil profiles should diminish the base content of mineral material prior to its erosive removal and sedimentation in lake basins (Engström and Wright, 1984). The same process of soil development is recorded also in lowland loess plateaus of the Czech Republic. Loess formation, which is characteristic of late Pleniglacial, generally terminates during the Bölling phase, and initial pedogenesis takes place during that time (Ložek and Cílek, 1995). The decalcification of soil horizons, together with expanding forest, were responsible for ultimate decrease in *Helianthemum* percentages.

### 5.3. Younger Dryas

The Younger Dryas as a biozone is widely recognised over most of Europe. Concerning the duration and amplitude, this climatic oscillation is the most important during the whole Late-Glacial period (Lotter et al., 1992). YD climatic deterioration, dated roughly between 11 and 10 ka BP, is correlated with a readvance of polar waters into the North Atlantic. Although the consequences of this event are registered more strongly at the sites near the ocean fringes of northern Europe, it is apparent today that regional changes in climatic regime may have been just as great in southern and eastern regions of Europe as in the northern part of this continent (Lowe and Watson, 1993; de Beaulieu et al., 1994; Khotinsky and Klimanov, 1997). Although problems of absolute dating accompany the recognition of Younger Dryas event (the "$^{14}$C plateau"; Ammann and Lotter, 1989), it has been described from many sites in the world and today is believed to be a global event (Peteet, 1995). For the territory of the Czech Republic, almost no reliable between-site comparison has been possible for the Younger Dryas. Only at Vracov, southern Moravia (Rybníčková and Rybníček, 1972), is it marked by a small *Juniperus* and *Salix* rise after 10,765 BP, but its identification is difficult.

At the present study site, clear evidence of climatic deterioration is ascribed to the Younger Dryas chronozone. The values of *Betula* decrease, whereas those of *Alnus viridis, Salix, Betula nana,* Chenopodiaceae, and *Artemisia* increase. Proxy evidence suggests that climatic deterioration was an increase in continentality rather than a decrease in summer temperatures (see also Ammann, 1989), for reconstructed minimum July temperatures are at least 12°C. The same values are reconstructed for western Poland (Walker, 1995). The sedimentation character changes back to more minerogenic, and erosion intensity rises. The increase in erosion indicators is only indistinct, suggesting that soil development was not interrupted completely during this time. The Younger Dryas can be subdivided in the Švarcenberk standard profile into two phases, indicating climatic amelioration (increase in humidity?) some time before the onset of Holocene warming. The subdivision of Younger Dryas into an older phase with colder and more arid climate and the younger phase with warmer and wetter climate has also been reported from some other sites in Europe (from Norway and Poland; Goslar et al., 1993; Birks et al., 1994), while in the Alps and in most of Western Europe the younger phase has been suggested to be somewhat drier (Walker, 1995).

The formation of extensive aeolian deposits in the region under study is dated to the beginning of Younger Dryas chronozone: Stratigraphic investigation of one of the most prominent sand dunes ("Vlkovský přesyp", situated near Švarcenberk lake basin) has revealed a fossil soil buried under aeolian sands with a distinctive layer of pine charcoal fragments on its surface. This situation resembles conditions in The Netherlands, where the "Usselo soil layer" formed during the Alleröd period of lower aeolian activity. "Usselo-layer" has been dated from surface charcoal fragments to between 11,400 and 10,300 BP, with average date around 11,000 BP (Hoek, 1997). These results resemble those from "Vlkovský přesyp", where a radiocarbon date $11{,}260 \pm 120$ BP has been obtained from very similar stratigraphical situation. The formation of soils during the Alleröd period required stable climatic conditions with less aeolian activity and relatively dense vegetation cover. On the other hand, the formation of aeolian sand dunes requires severe climatic conditions and sparse vegetation cover. Acceleration of aeolian activity during the Younger Dryas accords well with the results of pollen analysis, which point to a certain opening of the forest relatively to the preceding Alleröd period. The morphology of sand dunes points to prevailing easterly winds in time of their formation.

### 5.4. Early-Holocene

There is abundant evidence throughout Europe for a rapid rise in temperature at around 10,000 BP, although precise dating of this event is difficult because of another "radiocarbon plateau" at about that time. Over many areas of central and northwestern Europe, Younger Dryas open communities were replaced within less than 500 yr by *Betula/Pinus/Corylus* woodland (Walker, 1995). The preservation of *Pinus*-dominated forest during the entire Early-Holocene and the relatively late development of deciduous forest was connected with the persistence of a continental climate during that time and generally low nutrient status together with the sandy character of soils. Pine-dominated forests persisted in the area until the Boreal increase in humidity, although deciduous forests started to develop in favourable locations somewhat earlier. The rapid temperature rise during the Preboreal is indicated in the lake environment by the early occurrence of *Najas marina, Najas minor*, and *Trapa natans* macrofossils. *Najas marina* suggests a mean July temperature not below 15°C (Lotter, 1988), and *Trapa natans* even more. This proxy evidence suggests that the present-day values were reached as early as about 9800 BP.

In a number of proxy records from mainland Europe, there are indications of a cold climatic oscillation during the first millennium of the Holocene: the "Preboreal oscillation" of the Swiss Plateau (Lotter et al., 1992) or the "Youngest Dryas" of northern Germany (Behre, 1978). In western Norway, a readvance of the Jostedalsbreen ice cap has been dated to ca. 9100 BP (Nesje et al., 1991), while an abrupt fall in snow accumulation (associated with a fall in North Atlantic sea surface temperatures) is recorded in the GISP2 Greenland ice core some 400 yr after the end of the Younger Dryas (Alley et al., 1993). Equivalents of the European Preboreal oscillation are supposed to be found also in North America (Lowe et al., 1994). Possible palynological evidence for Preboreal climatic oscillation has been found also at Švarcenberk, but it is relatively weak: a short *Pinus* peak accompanied by the fall in *Corylus, Ulmus*, and *Quercus* percentages and dated slightly before $9640 \pm 115$ BP can be the result of some short cooling episode.

The Early-Holocene sediment record from Švarcenberk lake comprises a prominent Fe peak dated to around 8600 BP. It may be best explained as the reflection of intensive leaching caused by sudden climatic humification (Engström and Wright, 1984; Starkel, 1991). In the pollen diagram, the same period is characterised by *Picea abies* expansion. There is probably some connection between these two phenomena, as spruce grows preferably on waterlogged soils and is able to produce highly acidic, raw humus, promoting intensive leaching. The gradual development of nutrient-poor acid soils was an important factor in the Holocene vegetation development, as emphasised by Iversen (1964). The building-up of raw humus on the soil surface and resulting reducing conditions may have released Fe from the soil, and it travelled to the lake in solution or bound in organic complexes. A similar peak in Fe has been described from lowland areas of the Czech Republic, where Early-Holocene debris is cemented by limonite and goethite (Ložek and Cílek, 1995). Also in Poland, the beginning of the Holocene is characterised by inwash of dissolved iron into the lakes, and this is interpreted as the first stage of intensive soil leaching (Pawlikowski et al., 1982). In southern Sweden, the Fe content of several Early-Holocene lake sediments is very high. Digerfeld (1972, 1975) attributed this peak to early leaching from Late-Glacial soils in the catchment and subsequent transport by groundwater to the lake.

## 6. Conclusions

The results of this study permit the following conclusions:

1. The Late-Glacial biostratigraphy of the investigated site can be subdivided into different local pollen assemblage zones (PAZ). These biostratigraphic units can be correlated with general chronostratigraphic and climatostratigraphic subdivision of the last glacial–interglacial transition as established in Europe.
2. Three distinct cold events interrupted the Late-Glacial climatic amelioration: The first occurred before 13,000 and is correlated with the Oldest Dryas. Another regressive phase left only a weak signal and subdivides the Late-Glacial Interstadial into distinct *Betula* and *Pinus* phases. It is correlated with the Older Dryas. The third regression, corresponding to the Younger Dryas chronozone, is the most prominent one. It resulted in the reduction of the regional forest cover and a new expansion of open herbaceous communities. Aeolian activity accelerated during this time. The continental character of the climate continued for about 500 yr after the onset of Holocene warming. It is likely that the observed Late-Glacial and Early-Holocene climatic oscillations can be attributed to the same processes that acted in the western parts of Europe, i.e. the large-scale shifts in the position of the North Atlantic Polar Front (Ruddiman and McIntyre's model).
3. The response of regional vegetation to the Late-Glacial climatic changes was different from much of NW and western part of Central Europe. The main difference is the early development of primeval pine forest during climatic amelioration antedating the onset of the Late-Glacial Interstadial. Pine dominance is then characteristic for the entire Late-Glacial and Early-Holocene in the investigated area, while juniper is only of minor importance. The role of *Juniperus* as a pioneer shrub was therefore adopted by willow species.

4. There is strong correspondence between vegetational development, soil development, and the intensity of geomorphic processes acting within lake and its surroundings. The synchroneity of these processes has been primarily provoked by external climatic forcing.

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