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Stable oxygen and carbon isotopes in Late Glacial–Holocene freshwater carbonates from Belarus and their palaeoclimatic implications

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Abstract

Four sets of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves were derived from Late Glacial to mid-Holocene spring and lacustrine carbonates from Belarus. Stratigraphic division was carried out by palynological studies and ^{14}C -dating. It was revealed that variations of isotopic compositions and the dynamics of terrestrial vegetation were affected to some extent by common climatic forcing: birch expanded during relatively warm and/or arid phases (higher $\delta^{18}\text{O}$ and/or $\delta^{13}\text{C}$), fir spread during relatively cold and/or humid phases (lower $\delta^{18}\text{O}$ and/or $\delta^{13}\text{C}$), and grasses were abundant during relatively cold (lower $\delta^{18}\text{O}$) or arid phases (higher $\delta^{13}\text{C}$).

A comparison of four pairs of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves allowed the recognition of several palaeoclimatic events at a regional scale. Two main cold events interrupted by a warmer interval occurred in the Younger Dryas. A short-term episode of active permafrost decay at the end of the Younger Dryas and in the very beginning of the Preboreal caused an abrupt rise of lake levels followed by their rapid fall. A shift toward a lower precipitation–evaporation balance and lower lake levels took place roughly in the middle of the Boreal. A trend of decreasing temperatures throughout most of the Boreal culminated in a short-lived very cold phase at ~ 8100 ^{14}C year BP or ~ 9050 cal. ^{14}C year BP, according to the proposed age model. In the Atlantic 1 warmer phases were coincident with the phases of the lower precipitation–evaporation balance. An episode of significant short-term warming occurred in the last third of the Atlantic 1.

Correlation of the $\delta^{18}\text{O}$ curve derived from authigenic calcites precipitated between the beginning of the Preboreal 2 and the end of Atlantic 1 in the Ptich river valley with that from the GISP2 Greenland ice core allowed good matching of the trends and spikes (i.e. of the air temperature changes) under assumption that the ages of the Ptich pollen-based time model are 600–1000 years too old.

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1. Introduction

Sediments from carbonate lakes and calcareous spring deposits can produce continuous isotopic records of environmental change that respond rapidly to climate forcing. Oxygen and carbon isotopic data can be combined with palaeobotanic and lithological

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studies to develop a synthetic palaeoenvironmental picture. Marked changes in oxygen isotope ratios obtained from carbonate rich sediments may be considered as time-parallel markers at regional and possibly supra-regional scales and can serve as a basis for the development of regional “event stratigraphies” (Lowe et al., 2001).

Oxygen and carbon isotopic curves spanning different intervals of the Late Glacial and Holocene were obtained and interpreted in terms of palaeotemperatures (oxygen isotopes) and hydrological changes (oxygen and carbon isotopes) in some European countries and regions with modern temperate climate: in Great Britain (Andrews et al., 1994; Jones and Marshall, 2002); in Germany (Pachur and Roper, 1984); in the Alps (Eicher and Siegenthaler, 1976, 1984; Eicher et al., 1981; Schwander et al., 2000); in Sweden (Morner and Wallin, 1977; Hammarlund, 1994); in Poland (Rozanski, 1987; Kuc et al., 1993); and in the Baltic region (Punning et al., 1984; Martmaa et al., 1981). Most of these investigations were carried out in the area where modern climate is strongly influenced by Atlantic air masses. The climatic influence of the Atlantic essentially decreases to the east of 17–20 meridians, as was inferred from the distribution pattern of oxygen isotope ratios in recent meteoric precipitation in Europe (Rozanski et al., 1993), and from other evidence. Therefore, eastward of this longitude band, climatic changes in the past might have been driven by another mechanism than that in Western Europe. Belarus is located between 23°11' E and 32°47' E, and may be a good area to test whether climatic shifts to the east of the 20th meridian, synchronous or quasi-synchronous to those in Western Europe, can be recognized in the Late Glacial–Holocene. Thus, isotopic data on Belarusian carbonate sediments may contribute to the reconstruction of a spatial–temporal pattern of climatic changes in Europe.

Isotopic investigations may be of special interest for Belarusian palaeogeographers because palaeoclimatic reconstructions available for the region are based exclusively on the analysis of pollen spectra. Using the so-called “informative statistical method”, Elovicheva and Bogdel (1987), Elovicheva (1993), and Velichko et al. (1997) have reconstructed incredible, perfectly parallel precipitation and temperature changes during the Late Glacial–Holocene. This approach was criticized in the literature (Guiot et al.,

1993), and the results are to be verified by other proxy lines (including isotopic ones). Lake level changes in Belarus were also reconstructed mainly by means of terrestrial and aquatic pollen analysis and to a lesser extent by lithological studies (Zernitskaya et al., 1998), and the conclusions obtained need further support from other methods.

Until 2000, the only pair of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves from Belarus was described in detail (Punning et al., 1988). It was derived from marls of the biggest Belarusian lake Naroch. The stratification of the sequence was based on two radiocarbon dates from basal layers but mainly on the intuitive ageing of isotopic trends and spikes. No pollen stratigraphy was available.

In our paper we present the results of an investigation of four freshwater calcareous deposits from Belarus. Some information (an attempt to correlate oxygen curves) has been published earlier (Makhnach et al., 2000). No comparison of the isotopic trends with the dynamics of vegetation and lithological compositions was made. The aims of the current study are threefold: firstly to provide additional sets of isotopic data on Late Glacial–Holocene carbonates from a poorly studied region located at the transition from maritime to continental climatic conditions; secondly to compare isotopic, lithological and palynological information and to interpret $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records in terms of palaeoenvironmental changes; thirdly to test whether oxygen and carbon isotopes reflect regional climatic and hydrological changes as well as supra-regional events, documented in other European palaeoenvironmental archives.

2. Study area

This paper presents palynological, lithological and isotope data, which were obtained on a spring deposit Ptich, Central Belarus, and on three lakes (Teklits, Okono and Bezymiannoe) from the Belarusian Poozerje (the Lake Land), Northern Belarus. All of the sites are located within the areas where groundwaters are close to saturation with respect to calcite (Fig. 1).

The Ptich calcareous tufa deposit (53°48'N, 27°25'E, ~ 200 m asl) is located on the banks of the river Ptich within the Minsk Upland, 28 km from the river source and 10 km to the southwest of Minsk.

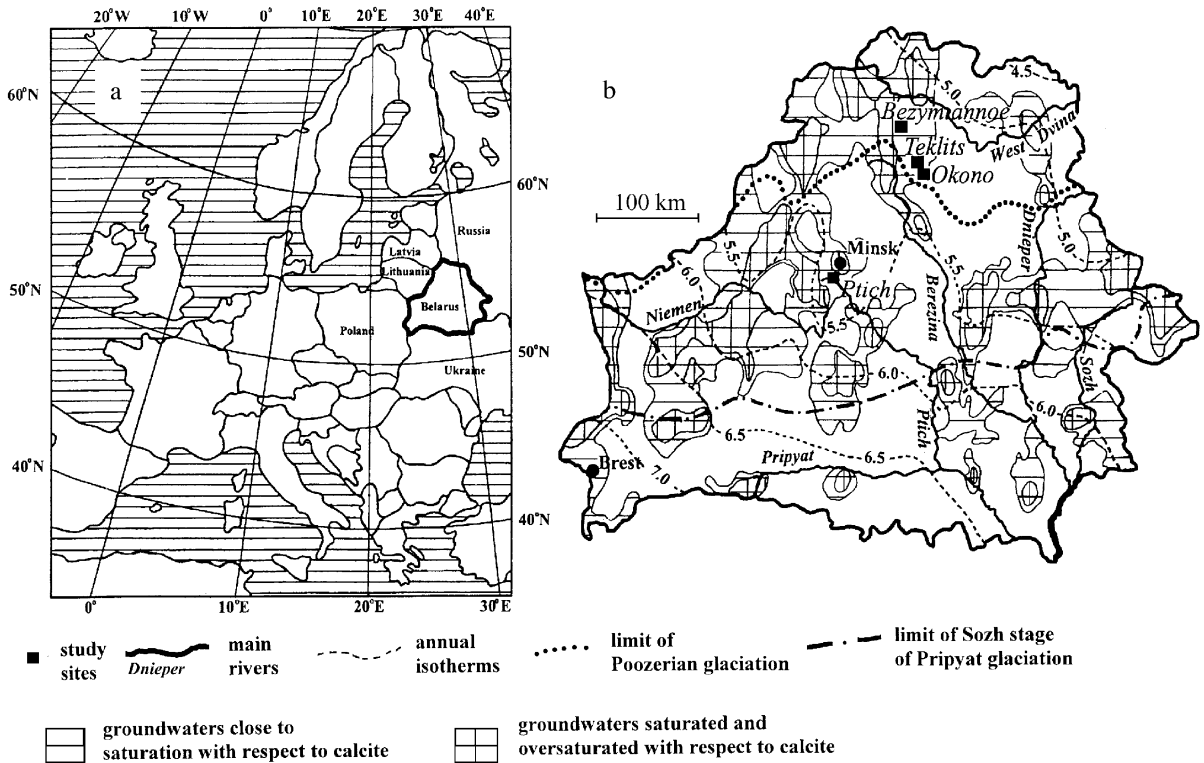


Fig. 1. Belarus in the context of Europe (a), and the location of the study sites (b). Annual isotherms are according to *Entsyklapedyja pryrody Belarusi* (1984); limits of the maximum advances of the Poozerian ice sheet (marine isotope stages 2–5d), and of the Sozh stage (marine isotope stage 6) of the Pripyat glaciation are according to *Geologija Belarusi* (2001); areas of groundwaters characterized by the indices of saturation (I_s) with respect to calcite ranging from -0.5 to 0 (waters close to saturation) and from 0 to $+0.5$ (saturated and oversaturated waters) are as in *Makhnach and Gulis* (1997). Note that in the country, I_s values range from -6.0 to $+0.5$. In Belarus, the ^{18}O content of meteoric precipitation was measured within the framework of worldwide monitoring (*Rozanski et al., 1993*) in Minsk and Brest, which are indicated by bold circles.

The Minsk Upland is a terminal moraine ridge formed during the Sozh stage of the Pripyat glaciation. Tills that are up to 200 m thick and glaciofluvial Quaternary deposits are underlain by Middle Devonian and Upper Cretaceous carbonate rocks. Deep glacial meltwater channels formed during the advance of the Sozh ice sheet and are a characteristic feature of the terrain. Upstream, the river Ptich occupies one of these channels. Near the villages of Priluchki, Priluki, and Letskovshchina, the 5–10 m wide meandering river flows into a deeply incised valley. The valley slopes ($\sim 10^\circ$) rise to 220–240 m from the valley floor (200–201 m asl). In the vicinity of the site, narrow and wide regions of the valley alternate (up to 0.5–1.0 km). The deposit is located in one of these wide segments, has a roughly round shape and a diameter

of ~ 200 m. The maximum thickness of carbonate sediments is about 5 m.

Recent geological, lithological, carpological, and malacological studies (*Kalicki et al., 1997*) enabled the reconstruction of the history of the upper Ptich valley in the Late Quarter. The spring genesis of carbonate sediments was proved by poor lake and abundant terrestrial malacofauna, and by a deficiency of aquatic plant macrofossils. According to *Kalicki et al. (1997)*, during the period of carbonate precipitation the floor of the Ptich valley was an inundated surface with numerous springs. The channel containing the Ptich River was either weakly expressed or did not exist at all. Carbonate precipitation started in the Allerod and ceased probably in the earlier Atlantic when some climatic or local geomorphological event

led to an incision of the river and to the forming of a tufa terrace. Better drainage of the valley floor caused the decline of tufa development. Calcareous tufa is overlain by 0.8–1.5 m thick slope deposits. A buried soil horizon overlain by slope deposits was found at the top of calcareous tufa approximately 30 m from our sampling site. The date of 3310 ± 110 ^{14}C year BP or 3830–3270 cal. ^{14}C year BP (GD-10083) obtained from this buried soil (Kalicki et al., 1997) indicates the minimal age of an episode of active slope erosion in the valley.

The Ptich calcareous deposit is one of the thickest spring deposits in Belarus, and some other thick tufas were found in close proximity to the site in the upper Niemen and upper Ptich catchment areas. We chose to investigate the Ptich site with the hope that in the future we would be able to quantitatively evaluate the similarity of several isotope records derived from a defined area. The profile was sampled in July 1998.

Lake Teklits ($54^{\circ}56'\text{N}$, $28^{\circ}30'\text{E}$, 166.3 m asl, Lepel district, Vitebsk oblast) is located in the Serguch river basin on the Upper Berezina glaciofluvial plain, which was formed during the retreat of the Poozerian glacier. The average thickness of Quaternary deposits within the plain is 100–110 m (Matveev et al., 1988). The area of the lake is 2.02 km², the length is 2.0 km, the maximum width is 1.3 km, the maximum depth is 3.2 m, the catchment area is 32.9 km², and the maximum thickness of bottom deposits is 5.5 m. Woods and shrubs occupy 33% of the catchment area. Teklits is an overflowing eutrophic lake with elements of dystrophy. Two streams enter the lake (one of them from Lake Lukonets, which is located 500 m to the north of Teklits at 166.4 asl). Lake Lukonets does not receive water from upstream lakes. Lakes Teklits and Lukonets are fed by underground waters and surface runoff. The river Serguch, which belongs to the Dnieper catchment area, flows out of Lake Teklits and enters lake Olshitsa (160 asl), 12 km to the SW from Teklits.

The slopes of the lake depression are gentle, 3–8 m high on the north, west and south sides and are moderately steep and 10–12 m high to the east. Northern and western shores are occupied by a 100–200 m wide peat bog. Ancient shorelines, which can be recognized in the lake vicinity, indicate that in the past, the lake level might have been approximately

1.5 m higher, and the area roughly two times wider. A fall in the lake level might have been caused by incision of the river Serguch. Total dissolved solids in the lake water is ~ 230 ppm. Currently, sedimentation in the lake is dominated by black organic gyttja in the profundal zone, and by brown sandy weakly carbonate gyttja in the sublittoral zone. Modern rhizoconcretions composed of pure calcite form in near-shore soils, indicating that soil waters are saturated with respect to calcite. The core was taken in February 1997 from the ice surface in the sublittoral zone, 200 m from the northwestern margin of the lake in a water depth of 2.2 m.

Lake Okono ($54^{\circ}50'\text{N}$, $28^{\circ}32'\text{E}$, 154.3 m asl, Lepel district, Vitebsk oblast) is located in the basin of the river Esa (West Dvina basin) on the Upper Berezina glaciofluvial plain, 20 km to the south of the lake Teklits. Lake Okono is 4.13 km² in area, is 2.9 km long, is 1.8 km at its maximum width, and has a maximum depth of 2.2 m. The catchment area is 46.8 km². The maximum thickness of bottom deposits is 4.9 m. The lake is eutrophic with a relatively long water residence time. Okono does not receive discharge from upstream lakes. One stream enters the lake, and a small river Okonitsa flows out of the lake and enters Lake Bereshcha (152.1 m asl), 2 km to the south of Lake Okono. The slopes of the depression are low and gentle on the north, 2–5 m high on the south and southeast, and 6–10 m high on the east and west. The shores are 0.3–0.5 m high, and a 700-m-wide peat bog borders the northern, western and southern margins of the lake. Geomorphological features of the lake vicinity indicate that the lake level could have been 0.7–1.0 m higher in the past, and that during arid climatic phases the lake might experience closing. Total dissolved solids in the lake water is ~ 170 –185 ppm. Recent bottom sediments are represented by black organic gyttja almost barren of a clastic fraction and carbonates in the central part, and by brown carbonate gyttja with an admixture of sand, silt and authigenic carbonates in the sublittoral zone. The core was taken in February 1997 from the ice surface in the sublittoral zone, 100 m from the northeastern shore in a water depth of 1.7 m.

The neighbouring lakes Teklits and Okono lie in the catchment areas of two different major rivers (the Dnieper and the West Dvina). Therefore, we expected that common features of the isotopic curves derived

from the sediments of Lakes Teklits and Okono would reflect regional climatic changes rather than local hydrological events.

The almost completely overgrown dystrophic lake Bezymiannoe or Lozoviki (55°16'N, 28°07'E, 173.7 m asl, Glubokoe district, Vitebsk oblast) is a remnant water body located within northwestern terrestrialized margins of the adjacent long and narrow Lake Sviadovo. These lakes are part of the Ginkovo–Bezymiannoe–Sviadovo–Dolgoe chain of lakes, which occupies a deeply incised depression crossing the terminal moraine ridge of the Sventiany Upland from NW to SE. The Sventiany Upland was formed during the maximum advance of the Poozerje glacial sheet. Lake Bezymiannoe lies in the local watershed between the major lakes Ginkovo (159.2 m asl) and Sviadovo (172.3 m asl). The lake receives discharge from surface runoff and underground seepage and is connected with lake Sviadovo by a periodic canalized stream. All four lakes of the chain are surrounded by steep slopes, which are higher than 30 m. That is why relatively small fluctuations in the precipitation–evaporation balance in the past could have caused major changes in water depths, which might be reflected in the isotopic compositions of lake carbonates. In particular, recent lithological and palynological studies of four cores obtained from the deepest (up to 53 m deep) Belarusian lake Dolgoe, geomorphological investigations of the lake vicinity, as well as bathymetric mapping, revealed significant changes in Lake Dolgoe levels in the Late Glacial–Holocene (Zernitskaya et al., 2001). At present, the area of Lake Bezymiannoe is only 0.0006 km², the length is 0.035 km, the maximum width is 0.02 km, the maximum depth is 70 cm, water clearness is 30 cm, and the catchment area is less than 1.0 km². Bezymiannoe is the thickest deposit of lake sediments and bog accumulations that have been cored in the northern Belarus so far. This circumstance, as well as the availability of data on level changes in neighbouring Lake Dolgoe, led us to select the Bezymiannoe site for our investigation. The core was taken in July 1998 from the terrestrialized margin 10 m to the south of the water window.

The modern climate of Belarus is characterized by mean annual precipitation ranging from 533 to 706 mm, mean annual temperature of 4.4–7.4 °C (see Fig. 1), mean January temperature of –8.4 to –4.1

°C, and mean July temperature of 17.0–19.7 °C (Entsyklapedyja pryrody Belarusi, 1984; Klimat Belarusi, 1996). January isotherms (which are not depicted in Fig. 1) run in the NNW–SSE direction. The country is located within the sharpest west–east gradient of mean annual and winter temperatures in Europe (in the transition zone between maritime and continental climatic conditions), and Belarusian lakes may have experienced extreme climatic transitions during the Late Glacial and Holocene that would be recorded in the isotopic composition of authigenic calcites.

3. Field and analytical methods

Cores from lakes were obtained with a 10 cm diameter, 1 m long Russian corer. Calcareous spring tufa Ptich was sampled in an open trench down to a vertical depth of 210 cm, and with a 5 cm diameter, 0.8 m long Russian peat sampler from the deeper part of the deposit. One core was obtained from each site. Sediments were sampled at 2–10 cm intervals.

Pollen analyses were completed using standard preparation and analytical procedures (Berglund and Ralska-Jasiewiczowa, 1986), and 500 pollen grains were counted for a sample.

Percentages of organic matter (OM) were determined by loss of weight on ignition (LOI) at 500 °C for 1 h. Chemical macro-elements (expressed in the forms of oxides) in the ash residue were measured using an X-ray-fluorescence method. In this paper we consider the CaO contents only. Assuming that all of the Ca is in CaCO₃ and taking into account atomic masses of calcium, oxygen and carbon, we calculated CaCO₃ (total carbonate) abundances in the ash residue by multiplying CaO content by 1.78. The amount of inorganic, non-carbonate material (non-carbonate detrital fraction) was calculated as the difference between 100% and the sum of percentages of organic matter and total carbonate.

Semi-quantitative estimates of the mineral content of powdered bulk samples were determined by standard X-ray diffraction techniques. An aliquot of the powdered sample was scanned from 15° to 50° at 2°/min using Ni-filtered Cu K radiation at 30 kV, 20 mA.

Sediment samples for isotopic analysis were freeze-dried and then sieved through a 125 µm mesh

to prevent coarse shell fragments contaminating the carbonate content. Samples of carbonate gyttja and carbonate peat were roasted under vacuum at 380 °C for 1 h. Samples of lake marls and spring tufas underwent no thermic pretreatment. Stable isotope analyses were done by analyzing carbon dioxide produced during the reaction of carbonate with 100% H₃PO₄ at 25 °C. Cryogenically trapped and purified carbon dioxide was analyzed on a MI 1201 B mass-spectrometer. Results are expressed as permil deviations from the PDB carbonate standard and have an analytical precision of ± 0.1 ‰ for carbon and ± 0.2 ‰ for oxygen.

Radiocarbon determinations on organic matter and carbonate fractions were performed using benzene synthesis and liquid scintillation counting according to the methods adopted in the Institute of Geological Sciences, Minsk (Mikhailov et al., 2003). Organic fractions for dating were obtained using the “acid–alkali–acid” (AAA) method of pretreatment. The carbonate fraction was extracted from calcareous tufa by ultrasound disintegration, sifting, washing in distilled water and baking of the sample at 450 °C for 2 h. Radiocarbon dates were calibrated using the calibration program Cal-20 (Van der Plicht, 1993).

Isotope, lithological and palynological datasets were compared by means of correlation and regression (including multiple linear regression) analyses using STATISTICA software.

4. Methods of time scale construction

Determining the age of freshwater carbonate sediments is fraught with difficulties because radiocarbon dates on aquatic organic material or carbonates are not absolutely reliable due to the hard water effect. Therefore, in some cases, an indirect chronology based on pollen zonation appears to be the most reliable. Principles of chronostratigraphic division of Belarusian Late Glacial–Holocene profiles by palynological studies are described in detail in *Geologija Belarusi* (2001, pp. 372–386), and may be summarized as follows.

The chronology of main regional vegetation changes in the Late Glacial–Holocene in Belarus was developed as a result of the analysis of pollen

diagrams derived from 30 radiocarbon-dated profiles. Sixty-five age determinations were available.

Regional and sub-regional pollen assemblage zones (RPAZ and SRPAZ) were attributed to stratigraphic units (periods and sub-periods) of the Late Glacial–Holocene. A revised Scandinavian periodization (Mangerud et al., 1974) is traditionally used in Belarus. It includes the following periods: the Oldest Dryas (Oldest D), Bolling (BØ), Older Dryas (OD), Allerod (AL), Younger Dryas (YD), Preboreal (PB), Boreal (BO), Atlantic (AT), Subboreal (SB), and Subatlantic (SA).

Pollen spectra of the Allerod, 11,800–10,900 ¹⁴C year BP or ~ 13,750–12,800 cal. ¹⁴C year BP (the oldest time interval of the Late Glacial–Holocene that is comprehensively studied by Belarusian palynologists), are dominated by *Pinus*, *Ephedra*, *Juniperus* and *Betula* s. *Fruticosa* are frequently present. The Allerod–Younger Dryas boundary can be recognized by an abrupt decrease of *Pinus* and rise of *Picea* and non-arboreal pollen (NAP). In Late Glacial spectra, reworked pollen of *Quercus*, *Tilia*, *Ulmus*, *Carpinus* can be present (not more than 1–2%). The Preboreal 1 (10200–9800 ¹⁴C year BP or ~ 11950–11000 cal. ¹⁴C year BP) is characterized by RPAZ *Pinus*–*Betula*–NAP or *Pinus*–*Picea*–NAP. A wide peak of *Betula* (sometimes up to 70%), the beginning of the continuous presence of *Ulmus*, and almost entirely absent *Picea*, is peculiar for the Preboreal 2 (9800–9000 ¹⁴C year BP or ~ 11000–10000 cal. ¹⁴C year BP). The rational limit of *Corylus* marks the beginning of the Boreal (9000–7800 ¹⁴C year BP or ~ 10000–8500 cal. ¹⁴C year BP). The Boreal can be divided into three sub-periods in the sequences with very detailed sampling (at 2–3 cm intervals), but usually it is divided into Boreal 1 + 2 (9000–8400 ¹⁴C year BP or ~ 10,000–9400 cal. ¹⁴C year BP, RPAZ *Pinus*–*Corylus*–*Ulmus* or *Pinus*–*Betula*–*Corylus*–*Ulmus*) and Boreal 3 (8400–7800 ¹⁴C year BP or ~ 9400–8500 cal. ¹⁴C year BP, RPAZ *Pinus*–*Corylus*–*Quercetum mixtum*). Among the criteria for recognition of the Boreal 2–Boreal 3 boundary are notable rises in *Corylus* and sometimes in *Picea*, *Betula*, and *Salix*. The beginning of the Atlantic is marked by and abrupt rise in *Quercetum mixtum* and *Alnus*. The boundary between the Atlantic 1 (7800–6600 ¹⁴C year BP or ~ 8500–7450 cal. ¹⁴C year BP) and Atlantic 2 (6600–6000 ¹⁴C year BP or ~ 7450–

6850 cal. ^{14}C year BP) can be established by the rise in *Picea* and in some cases by the rational limit or by the beginning of the continuous presence of *Fraxinus*. RPAZ Q.m.–*Alnus–Corylus* is characteristic of Atlantic 3 spectra. At the beginning of the Subboreal, *Picea* increases abruptly. The Subboreal includes three sub-periods: Subboreal 1 (5000–4200 ^{14}C year BP or \sim 5750–4700 cal. ^{14}C year BP, RPAZ *Picea* or *Picea–Quercus*), Subboreal 2 (4200–3200 ^{14}C year BP or \sim 4700–3400 cal. ^{14}C year BP, RPAZ *Pinus–Picea–Quercus–Carpinus*) and Subboreal 3 (3200–2700 ^{14}C year BP or \sim 3400–2800 cal. ^{14}C year BP, RPAZ *Picea–Pinus–Betula*). The Subatlantic is divided into Subatlantic 1 (2700–2000 ^{14}C year BP or \sim 2800–1950 cal. ^{14}C year BP, RPAZ *Pinus–Picea–Quercus*), Subatlantic 2 (2000–1000 ^{14}C year BP or \sim 1950–900 cal. ^{14}C year BP, RPAZ *Pinus–Betula–Quercus–Carpinus*, sometimes with abundant *Picea* and decreased *Betula* in the northern Belarus), and Subatlantic 3 (from 1000 ^{14}C year BP or \sim 900 cal. ^{14}C year BP to present day). In the latter sub-period, human influence on vegetation becomes clearly visible (abundant NAP, *Betula*). In general, it should be borne in mind that during the entire Late Glacial–Holocene, *Picea* was relatively abundant in northern Belarus, less abundant in the central and eastern parts, and least abundant in the south and west.

During the course of time scales construction, two radiocarbon dates from Ptich and 11 dates from Bezymiannoe were taken into account. Special comments will be made in each case of discrepancy between the radiocarbon ages and the pollen-based chronostratigraphy.

We understand that the use of Scandinavian stratigraphic terms may lead to confusion in correlation with West-European sequences. Therefore, it should be emphasized that our chronological scales are mainly of regional use and do not imply accurate synchronicity between the periods (Allerod, Younger Dryas, etc.) in Belarus and elsewhere in Europe. They reflect the current state of chronostratigraphic studies in Belarus, which is certainly not perfect. The fact that many of the 65 ^{14}C dates mentioned above were obtained on aquatic organic matter accumulated in lakes with relatively hard water suggests that the regional chronology may be too old.

5. Results

5.1. Ptich

Eight LPAZ are distinguished in the profile and attributed to certain stratigraphic units according to the criteria described above. A specific feature of the pollen spectra of this profile located in a deeply incised valley is a deficiency of *Picea* and the broad-leaved species (Fig. 2). Brown carbonate peat, overlying yellow silty loam poor in pollen accumulates in the Allerod, possibly in the final Allerod. Grey clayey carbonate tufa accumulates from the beginning of the Younger Dryas. The YD unit is recognized only by abundant *Betula* and *Cyperaceae* herbs. In the beginning of the Preboreal 2, clayey calcareous tufa gives way to white calcareous tufa. In the upper half of the BO1+2 unit, the maximum amount of *Pinus* and deficiency of *Betula* are noted. The BO2–BO3 boundary is not clearly seen. It is distinguished according to the ^{14}C date of 8200 ± 120 ^{14}C year BP or 9440–8720 cal. ^{14}C year BP, which is obtained on an organic fraction from peaty tufa (i.e. on terrestrial organic material), and the age may not appear to be too old due to the hard water effect. In the uppermost part of the BO3 unit a notable increase in NAP takes place. Above the *Betula–Pinus* LPAZ (BO3), carbonate tufa poor in pollen occurs. Pollen concentration is insufficient for the diagram construction, however pollen grains of the broad-leaved trees are relatively more abundant than in the underlying sediments, indicating the transition to the Atlantic. The ^{14}C date of 7890 ± 160 ^{14}C year BP or 9200–8360 cal. ^{14}C year BP (135–140 cm) is obtained on a carbonate fraction, and the age is probably too old due to the hard water effect. Sediment thickness suggests that tufa accumulation might continue roughly till the end of the Atlantic 1. Carbonate tufa is overlain by slope deposits (dark grey silty loam) accumulated in the Subboreal 3 and Subatlantic after a break in sedimentation.

OM in Ptich is represented mainly by thin peat layers intercalating carbonate tufa, i.e. by terrestrial OM. OM content varies from 1% to 48%, but does not exceed 7% over most of the profile (Fig. 3). In the middle of the BO1+2 unit (two samples) and at the top of the first third of the AT1 unit (three samples) OM content is close to 0%. The maximum content was noted in carbonate peat in the AL unit. Several

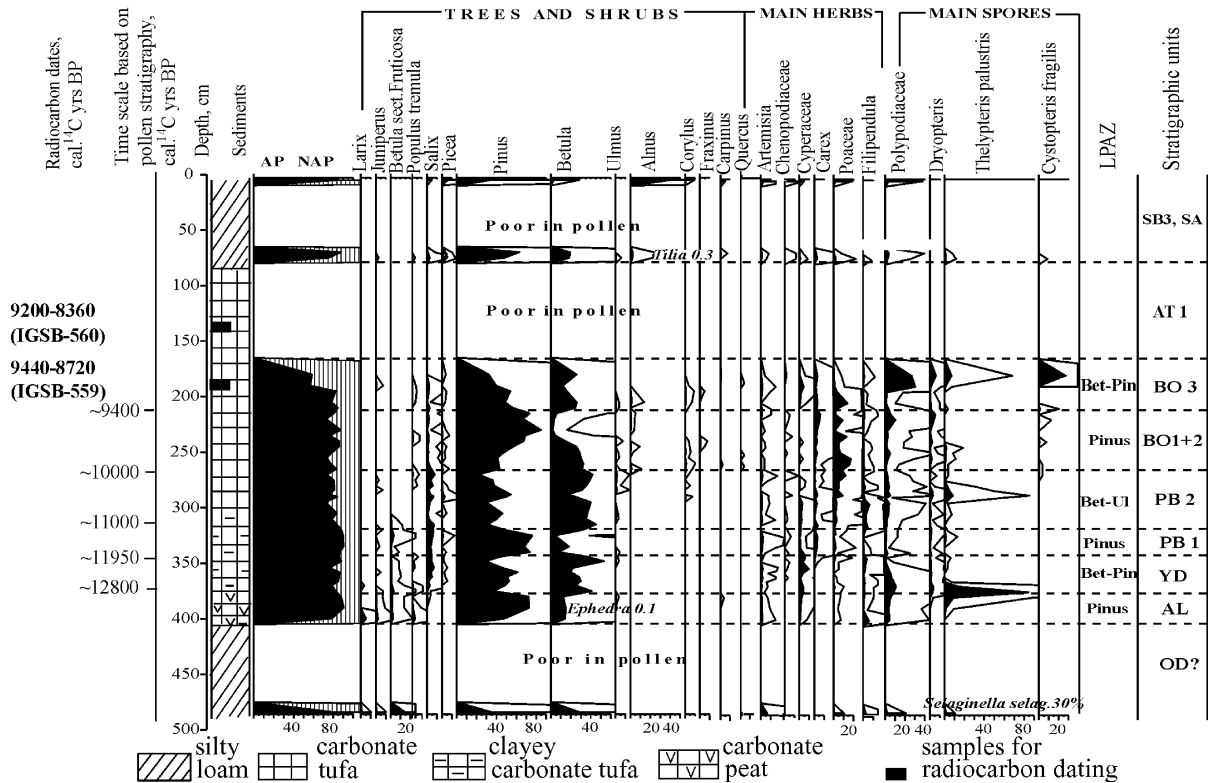


Fig. 2. Simplified pollen diagram derived from the calcareous tufa Ptich. The date of 7890 ± 160 ^{14}C year BP (9200–8360 cal. ^{14}C year BP) was obtained on a carbonate fraction, and the age may be too old due to the hard water effect. The date of 8200 ± 120 ^{14}C year BP (9440–8720 cal. ^{14}C year BP) was obtained on an organic fraction from a layer of peaty calcareous tufa.

narrow and wide spikes of higher OM content can be seen around PB1–PB2, PB2–BO1, BO2–BO3 boundaries and in the upper third of the AT1 unit.

The amount of the non-carbonate detrital fraction is a maximum at the AL–YD boundary. At the PB1–PB2 boundary this fraction decreases abruptly, and from the end of PB2 until the middle of AT1, tufa containing less than 7% of detrital material accumulates. In the lower half of the AT1 unit the amount of the non-carbonate detrital fraction is less than 1%. In the purest carbonate tufa (the upper third of PB2—the middle of AT1), the content of CaCO_3 in the ash residue ranges between 93% and 99.5%.

The values of $\delta^{18}\text{O}$ in the two lowermost samples (-6.8% and -5.8%) mainly reflect the isotopic composition of a mixture of detrital and authigenic carbonates. Within carbonate tufa and carbonate peat,

$\delta^{18}\text{O}$ varies from -13.5% to -6.9% , and the median is -11.3% . In the AL unit, a positive peak is observed, and at the top of the AL unit the values show a minor negative peak. In the upper half of the YD unit, a negative peak is noted. In the BO1+2 unit and through the lower half of the BO3 unit there is a monotonic trend of decreasing values with the minimum for the sequence (-13.5%). This trend is punctuated by a minor positive spike in the lower half of the BO3 unit. A positive spike is observed near the BO–AT boundary, and a narrow peak with the maximum value of -6.9% is noted in the upper third of the AT1 unit.

Statistical analysis does not reveal a significant correlation between $\delta^{18}\text{O}$ and percentages of individual pollen taxa. Nevertheless, $\delta^{18}\text{O}$ shows a weak covariance with the *Betula* + *Pinus* sum ($r = +0.35$).

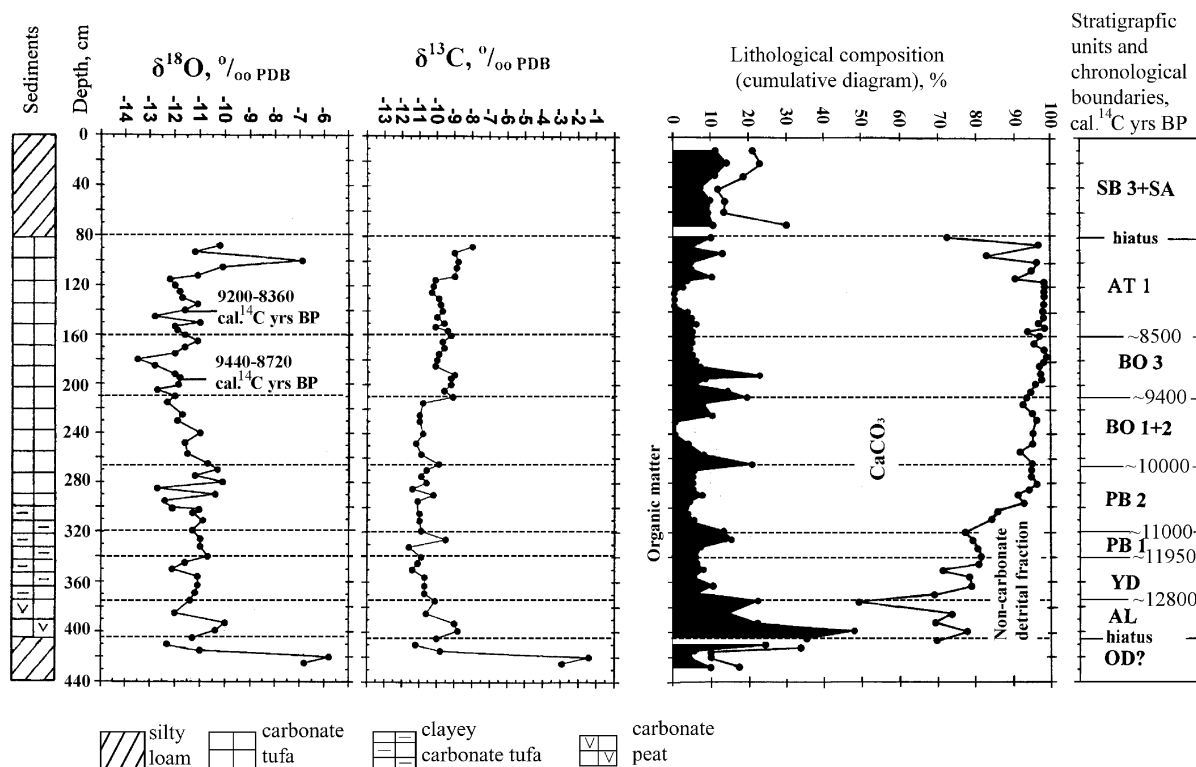


Fig. 3. Stable isotopes in bulk carbonates and the lithological composition of the calcareous tufa Ptich.

Maximum $\delta^{13}\text{C}$ values (-1.4‰ and -2.9‰) are observed in the two lowermost samples and reflect the carbon isotopic composition of a mixture of detrital and authigenic carbonates. Within carbonate peat and carbonate tufa, $\delta^{13}\text{C}$ varies from -11.6‰ to -8.0‰ , and the median value is -10.1‰ . A positive spike in the AL unit, further monotonic decline toward the base of the PB1 unit to the sequence minimum $\delta^{13}\text{C}$, a positive peak below the PB1–PB2 boundary, a shift toward more positive values at the BO2–BO3 boundary, a minor shift toward more negative values in the middle of the BO3 unit, and the last shift to more positive values in the upper third of the AT1 unit are the main features of the carbon isotope curve.

The synchronicity of positive peaks and spikes of $\delta^{13}\text{C}$ and of increased OM contents is of interest. The values of $\delta^{13}\text{C}$ negatively correlate with the *Betula* + *Pinus* sum ($r = -0.45$). Covariance of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is low ($r = +0.33$) and statistically insignificant.

The relationship between the isotopic compositions and the main arboreal taxa, spreading at the expense of non-arboreal species, can be described by the multiple linear regression equation: *Betula* + *Pinus*, $\% = 61 - 0.47\delta^{13}\text{C} + 0.30\delta^{18}\text{O}$ ($R = 0.54$). Note the negative coefficient for $\delta^{13}\text{C}$ and the positive coefficient for $\delta^{18}\text{O}$.

5.2. Teklits

Ten LPAZ or stratigraphic units are distinguished in the profile. Accumulation of chemogenic lake sediments (grey lake marls) begins in the Younger Dryas (Fig. 4). At the base of the AT1 marls give way to grey silty marls, and in the upper third of the unit a layer of yellowish grey carbonate sand occurs. At the upper boundary of the AT1 unit, the deposition of marls with a less abundant clastic fraction resumes. In the SB2 + 3 unit, *Pediastrum boryanum* alga (indicative of high trophic status or low lake level) is absent.

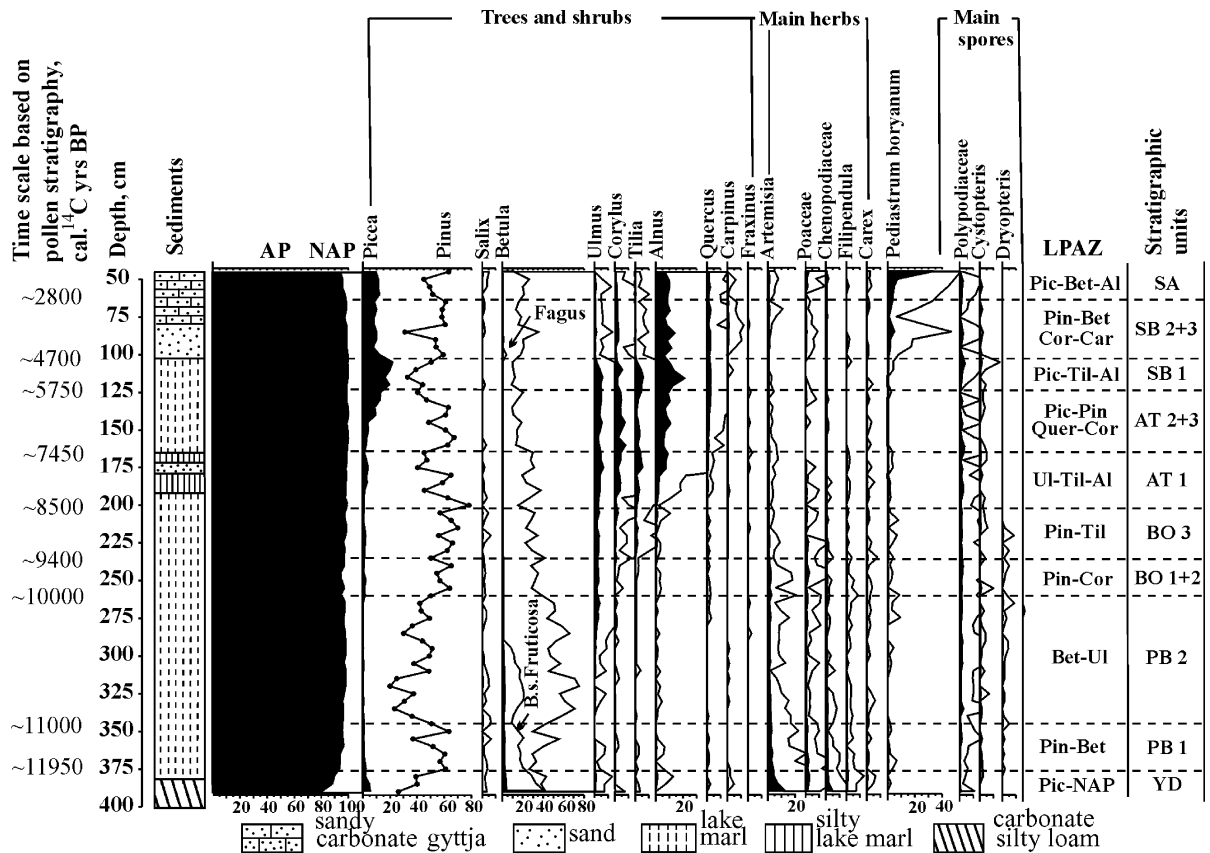


Fig. 4. Simplified pollen diagram derived from lacustrine deposits, Lake Teklits.

At the base of the SB2 + 3 unit *P. boryanum* algae rises abruptly. At that time, marls give way to sand. In the middle of the SB2 + 3 unit the type of sedimentation changes again, and dark brown sandy carbonate gytja starts to accumulate.

OM percentages vary from 15% to 68% (Fig. 5). A wide spike of increased OM (31–36%) is noted in the PB1 unit. Relatively low contents are observed in the AT1 unit (in the sand layer). In the AT2 + 3 and SB1 units OM contents (27–40%) are higher than in older sediments (PB2–AT1). In a sand layer within the SB2 + 3 unit OM percentages are as low as 3%. OM contents in the uppermost samples of the SB2 + 3 unit and in the SA unit (52–69%) are maximal for the sequence.

Through most of the profile, the content of the non-carbonate detrital fraction ranges between 10% and 17%. Increased values are seen at the base of the core

(the YD unit), at the BO2–BO3 boundary, in the upper half of the AT1 unit and in the SB2 + 3 unit (maximum content). The content of CaCO₃ in the ash residue ranges between 71% and 85% through most of the core.

$\delta^{18}\text{O}$ varies from -17.7‰ to -9.5‰ , and the median value is -13.4‰ . The neighbouring samples frequently show values differing by several permil. The minimum is noted in the lowermost sample within the YD unit. Below the YD–PB1 boundary there is a shift toward more ¹⁸O-enriched carbonates. The next strong shift toward more positive values occurs at the top of the PB1 unit. In the lower third of the PB2 unit a positive spike, and roughly in the middle of the PB2 unit, a negative spike, of $\delta^{18}\text{O}$ can be seen. Sediments of the Boreal period exhibit a threefold structure: a negative spike in the lower half of the BO1 + 2 unit, a positive spike around the BO2–

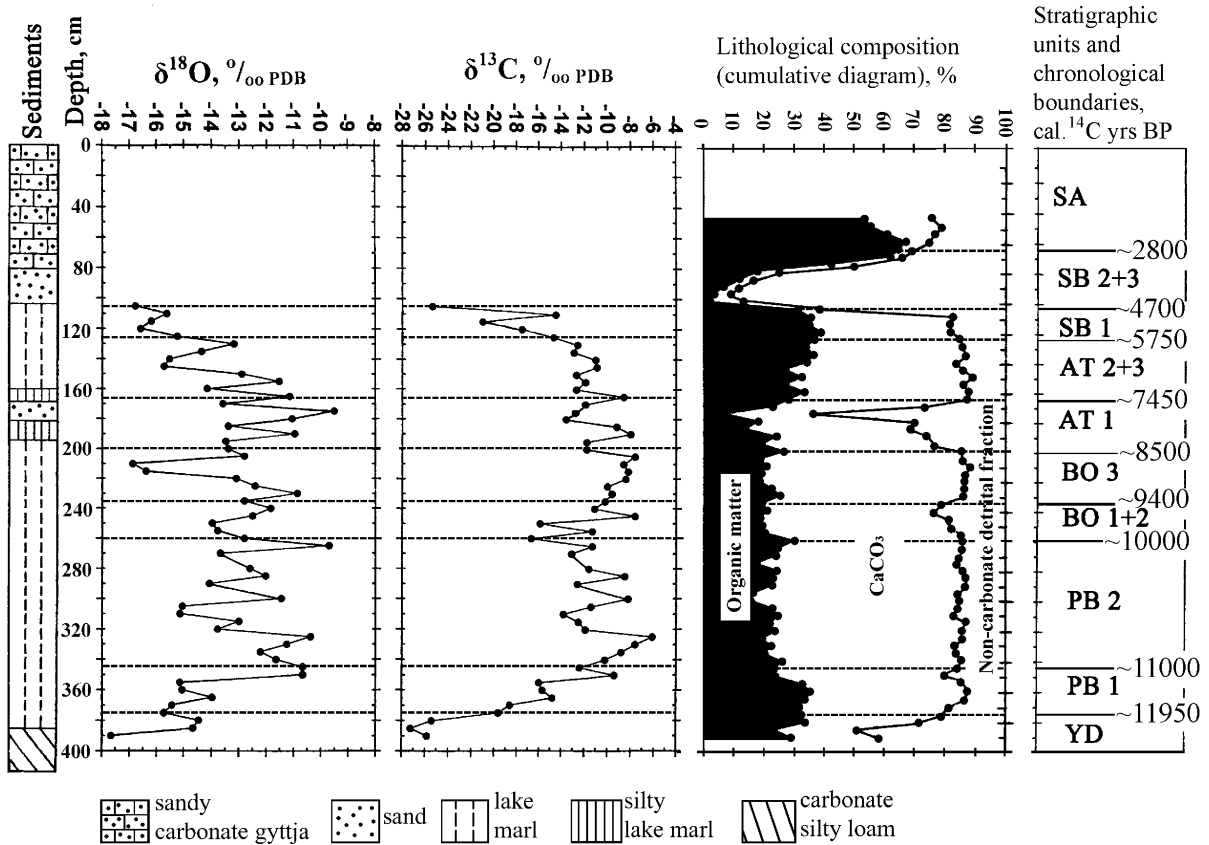


Fig. 5. Stable isotopes in bulk carbonates and lithological composition of lacustrine deposits, Lake Teklits.

BO3 boundary, and a strong negative spike above the middle of the BO3 unit. The $\delta^{18}\text{O}$ values culminate in the value of -9.5‰ near the top of the AT1 unit. Upward, there is a general trend of decreasing values.

Statistical analysis shows that $\delta^{18}\text{O}$ correlates positively with *Betula* ($r=+0.29$, $n=57$) and negatively with *Picea* ($r=-0.42$). In the sediments deposited before the expansion of the broad-leaved trees, which are the main competitors of *Betula* and the conifers (280–390 cm, YD, PB1, the lower third of PB2), a close positive correlation with *Betula* ($r=+0.80$, $n=22$) and a negative correlation with *Picea* ($r=-0.84$) are both visible.

The $\delta^{13}\text{C}$ values fall into an extremely wide range of 21.1‰ (from -27.2‰ to -6.1‰) with a median value of -11.9‰ . At the base of the core there is a trend of increasing values (from -27.2‰ to -6.1‰ , i.e. from the minimum at the top of the YD

unit to the maximum at the top of the lower third of the PB2 unit). The next remarkable feature of the $\delta^{13}\text{C}$ curve is a shift from low values (about -17‰) to more positive values roughly in the middle of the BO1+2 unit (-7.6‰ at 245 cm). The BO3 unit exhibits a monotonic trend of increasing values. Above the top of the AT1 unit, a trend of decreasing values (up to the value of -25.5‰ in the uppermost sample) is noted.

In Teklits, sediments relatively rich in OM show lower $\delta^{13}\text{C}$ values: the correlation coefficient r between $\delta^{13}\text{C}$ and OM is -0.60 . If the values $< -20\text{‰}$ are ignored, $\delta^{13}\text{C}$ correlates positively with NAP ($r=+0.51$, $n=43$) and negatively with *Picea* ($r=-0.42$). In the sediments, deposited before the expansion of the broad-leaved trees, $\delta^{13}\text{C}$ correlates positively with *Betula* ($r=+0.51$, $n=22$) and negatively with *Picea* ($r=-0.84$).

The correlation coefficient r between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is +0.45.

5.3. Okono

Nine LPAZ are distinguished in the profile. Light grey basal sand and the lowermost part of grey silty marl are attributed to the final Allerod (Fig. 6). Overlying sediments show pollen spectra typical for the Younger Dryas. In Okono we can recognize the YD1 and YD2 units. The latter unit differs by an increase of *Betula* and a decrease of *Picea*. At the upper boundary of the YD 2 unit, greyish white lake marls almost devoid of clastic material begin to accumulate in the lake. At the base of the BO3 unit there is a narrow peak of *Betula*. Above the middle of the BO3 unit there are two samples with almost absent *Ulmus*,

Corylus, *Tilia* and reduced *Alnus*. After the Boreal period the type of sedimentation changes drastically, light brown carbonate gyttja begins to accumulate instead of marl, and *Pediastrum boryanum* increases.

Stratigraphic division of the postboreal part of the core is rather difficult. Pollen spectra in carbonate gyttja show the features of the Atlantic (maximum of Q.m.) as well as characteristics of early Subboreal (abundant *Picea*). Therefore we can distinguish the joint unit of AT and SB1, which is only 40 cm thick. At the upper boundary of this unit the sedimentation type changes, and sedimentation is dominated by dark brown organic gyttja with minor amounts of carbonates. In the uppermost part of the sequence, the unit of SB2+3, SA is distinguished.

OM content varies from 4% to 61% (Fig. 7). Allerod sediments contain 13–15%, and the YD1

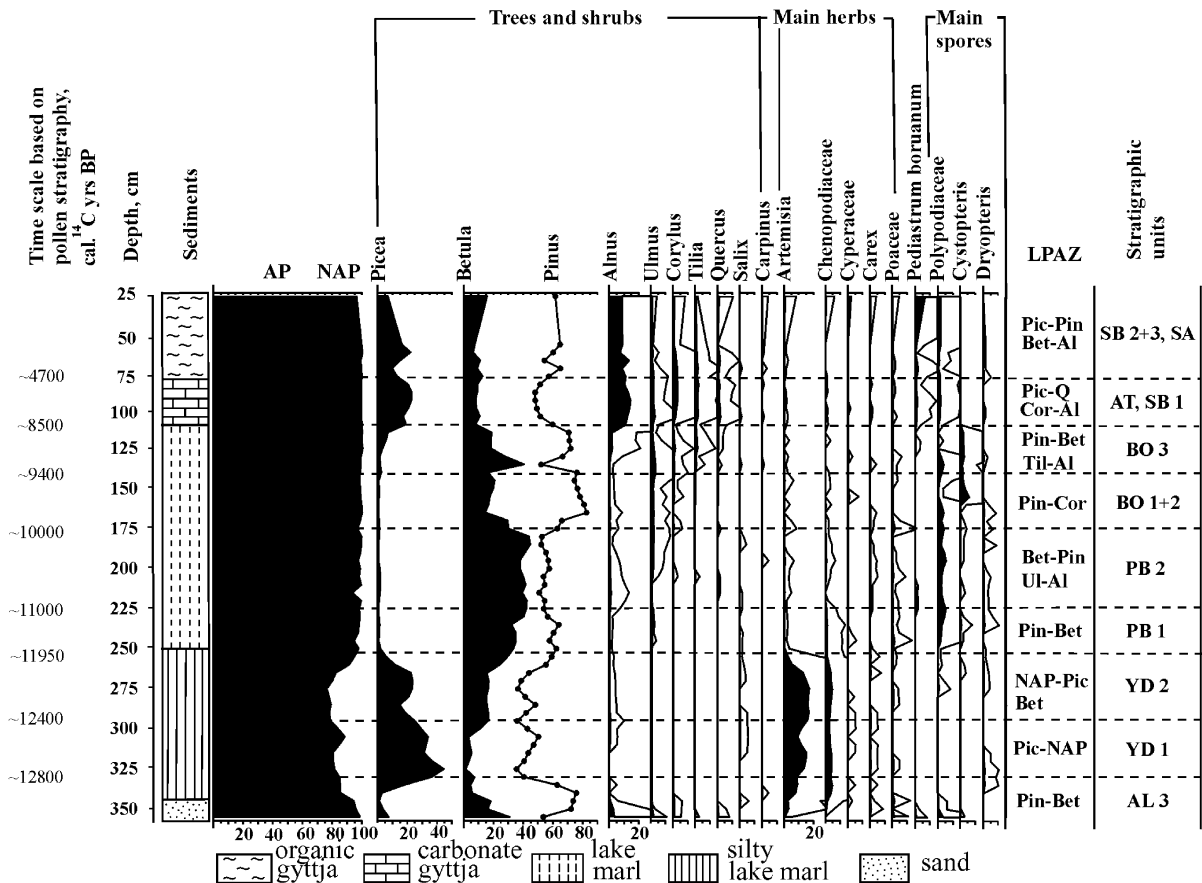


Fig. 6. Simplified pollen diagram derived from lacustrine deposits, Lake Okono.

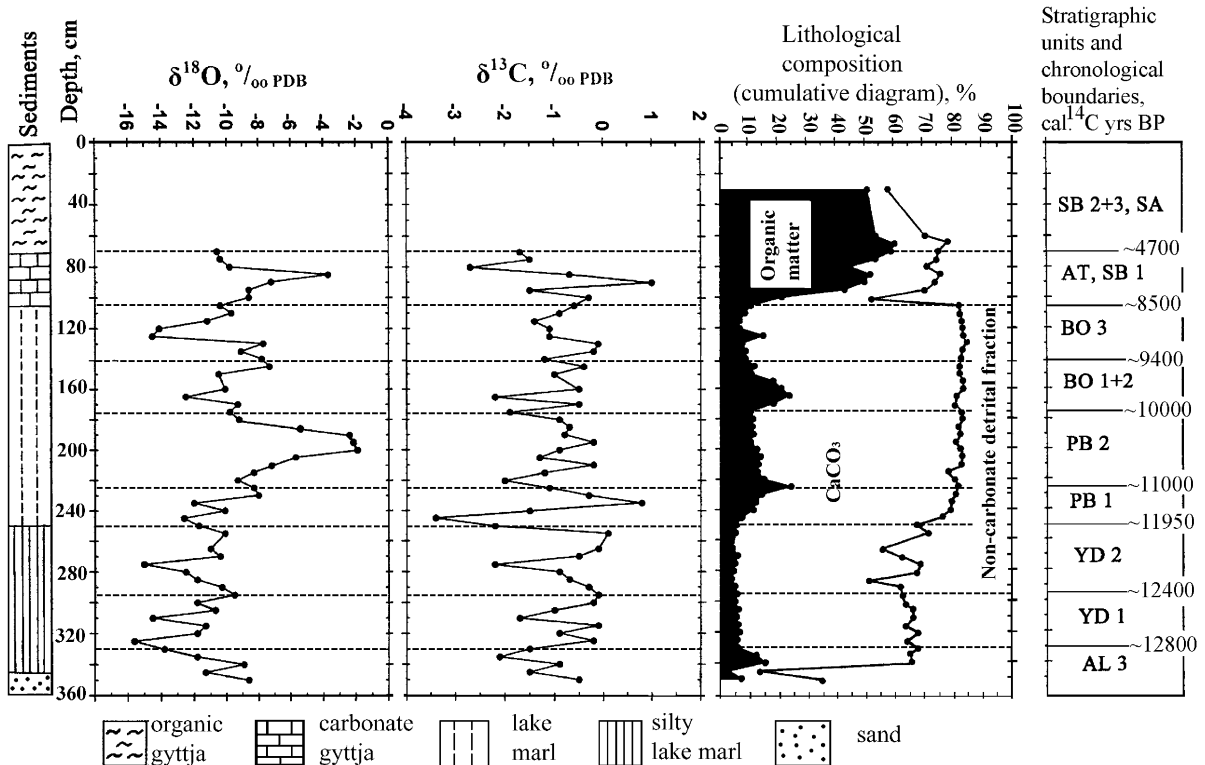


Fig. 7. Stable isotopes in bulk carbonates and lithological composition of lacustrine deposits, Lake Okono.

and YD2 units contain only 4–6% of OM. Throughout the PB1 unit the values increase parallel to a decrease in the detrital fraction. The PB1–PB2 boundary is marked by a peak of increased OM content (24%). In the lower two-thirds of the BO1+2 unit there is a peak of increased OM. The values increase abruptly at the Boreal–Atlantic boundary.

The content of the non-carbonate detrital fraction decreases in the beginning of the PB1 unit. In the PB1–BO3 units, the amounts are between 15% and 18%. Above the Boreal units, detrital fraction abundances rise significantly. At the base of the AT, SB1 unit there is a peak of increased detrital fraction. The amount of CaCO₃ in the ash residue of chemogenic sediments ranges from 50% to 89%, and above the lower third of the PB2 unit to the AT, SB1 unit, from 75% to 89%.

X-ray analyses show that among clastic carbonates (2–3 mm fraction) from glaciofluvial deposits sampled in the lake vicinity, dolomite is essentially more

abundant than calcite (Fig. 8a). Therefore, the amount of total detrital carbonates (dolomite + calcite) in bulk carbonates of Lake Okono can be rather accurately estimated by the presence of dolomite (an undoubted allochthonous carbonate). In the YD1 unit (305 cm, 66% of CaCO₃ in the ash residue, Fig. 8b) calcite dominates, and an admixture of quartz, a minor admixture of dolomite and barely detectable amount of feldspar are present. At the very base of the PB2 unit (220 cm, 76% of CaCO₃, Fig. 8c), against the background of dominant calcite, only a minor admixture of quartz is noted. In sediments with the maximum amount of calcite in a sample from the top of the BO3 unit (120 cm, 84% of CaCO₃, Fig. 8d) calcite only is documented. In carbonate gyttja (85 cm, 52% of CaCO₃, Fig. 8e) calcite and quartz are dominant, and a minor admixture of dolomite is present.

The values of δ¹⁸O vary from –15.7‰ to –1.9‰, and the median value is –10.3‰ (see Fig. 7). In the AL3 unit there is a trend of decreasing values. The base of the YD1 unit is marked by the

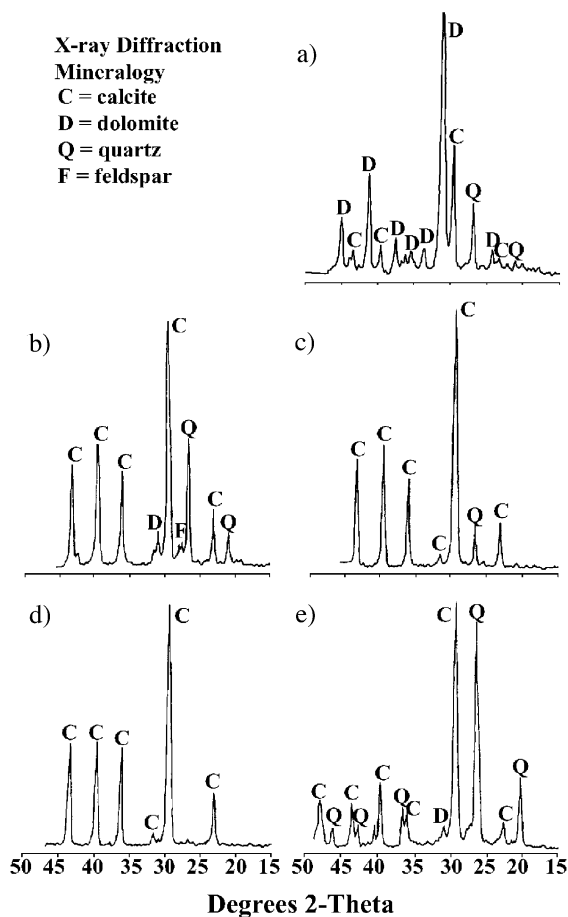


Fig. 8. X-ray diffraction patterns showing the positions (in degrees two theta) and relative intensities of the major minerals in detrital carbonates (2–3 mm fraction) sampled from glaciofluvial deposits in the vicinity of Lake Okono (a), and in sediments of Lake Okono from four selected depths: (b) 305 cm, YD1 unit, 66% of CaCO_3 in the ash residue; (c) 220 cm, PB2 unit, 76% of CaCO_3 ; (d) 120 cm, BO3 unit, 84% of CaCO_3 ; and (e) 85 cm, 52% of CaCO_3 .

sequence's minimum value (-15.7‰). In Younger Dryas sediments there are two strong negative spikes divided by a positive spike. At the top of the YD2 unit there is a shift to higher values. In the PB1 unit a small negative spike can be seen. In the third quarter of the PB2 unit there is a strong positive spike with the maximum value of -1.9‰ . Within Boreal sediments the oxygen curve exhibits a negative spike in the BO1+2 unit, a positive spike around the BO2–BO3 boundary, and a strong narrow negative spike with a very low value (-14.5‰) in the middle of the BO3 unit. At the top of the BO3 unit a trend of increasing

values begins. This trend culminates in a peak with the value of -3.7‰ in the AT, SB1 unit.

The $\delta^{18}\text{O}$ values correlate positively with *Betula* ($r=+0.45$, $n=53$) and negatively with *Picea* ($r=-0.31$) and *Artemisia* ($r=-0.44$).

The minimum, maximum and median values of $\delta^{13}\text{C}$ are -3.4‰ , $+1.0\text{‰}$, and -0.9‰ , respectively. The absolute values are significantly higher than those in the neighbouring Lake Teklits. The most distinct shift to more negative values is noted roughly at the YD–PB boundary. These low values, including the minimum for the sequence, abruptly change to higher ones (with the maximum for the sequence) approximately in the middle of the PB1 unit. One more shift to more positive values is noted roughly in the middle of the BO1+2 unit.

In general, there are no covariance or antipathetic relationships between of $\delta^{13}\text{C}$ and OM content (r is about 0). In the BO1+2 unit, a spike of increased OM corresponds to relatively low $\delta^{13}\text{C}$ values, and a decrease in OM content is coincident with the increase in $\delta^{13}\text{C}$.

Statistical analysis does not show significant correlation between $\delta^{13}\text{C}$ and pollen taxa. The correlation coefficient between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is low ($r=+0.39$).

5.4. Bezymiannoe

Fourteen LPAZ were distinguished in the pollen diagram. Five of them are within 2.8-m-thick carbonate gyttja. Basal light yellow sand is overlain by black noncarbonate peat, attributed to the Oldest Dryas due to abundant NAP, *Betula s. Fruticosa*, the presence of *Pinus s. Strobus*, and due to a date of 13740 ± 850 ^{14}C year BP or 18270–14150 cal. ^{14}C year BP (Fig. 9). In this peat layer, noncalcareous light yellow silty loam devoid of pollen occurs. It is overlain by noncarbonate peat (the PB1 unit). The oldest lake sediments (light brown silty carbonate gyttja) occur at the very top of the PB1 unit. At the lower boundary of the BO1+2 unit, silty carbonate gyttja gives way to brownish grey carbonate gyttja with less abundant terrigenous material. After the Boreal, silty carbonate gyttja accumulates again. By the end of the Atlantic 1, precipitation of carbonates ceases, and black organic gyttja begins to accumulate. In the upper part of the section, within non-carbonate sediments, which were not analyzed for stable isotopes, seven more units are

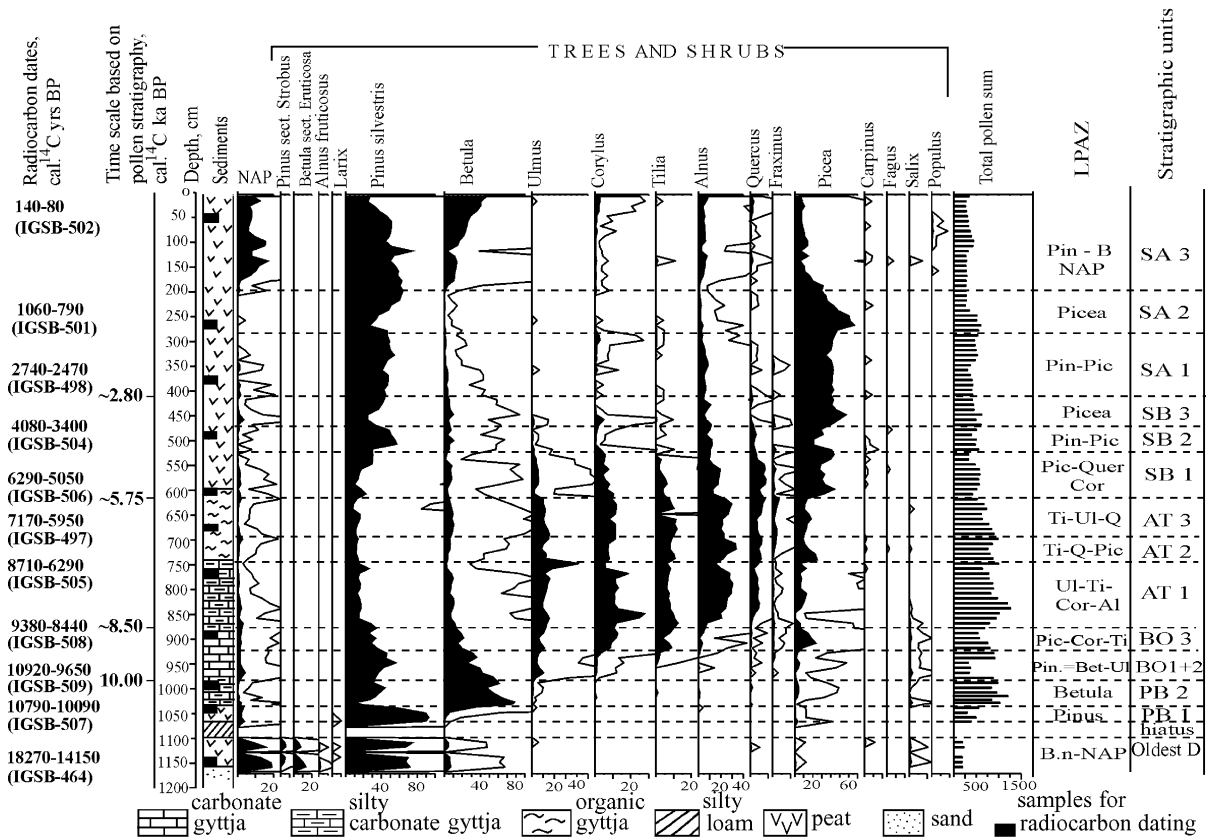


Fig. 9. Simplified pollen diagram derived from lacustrine deposits, Lake Bezymiannoe. All radiocarbon dates were obtained on an organic fraction.

distinguished according to standard palynological criteria. In the beginning of the Subboreal, accumulation of lake sediments is substituted by the accumulation of peat.

Eleven ¹⁴C dates obtained on organic fractions are available from the Bezymiannoe profile. Despite the fact that a significant part of the sequence (including three radiocarbon dates) is represented by carbonate sediments, and the ages might be too old due to the hard water effect, almost all of the dates are in good agreement with the ages estimated on the basis of pollen zonation. The only exception is the date of 9370 ± 90 ¹⁴C year BP or 10790–10090 cal. ¹⁴C year BP (IGSB-507) obtained on an organic fraction from noncarbonate peat (PB1–PB2 transition, 9800 ¹⁴C year BP or ~ 11000 cal. ¹⁴C year BP according to traditional Belarusian chronological scale).

OM content varies from 13.5% to 57% (Fig. 10). In the lowermost sample (the PB1 unit) the content is 37%, and at the PB1–PB2 transition, is 13.5%. At the Preboreal–Boreal transition, OM content increases to 33%. There are peaks of decreased OM at the BO2–BO3 boundary, in the lower third of the AT1 unit, and at the base of the upper third of the AT1 unit. Roughly in the middle of the AT1 unit there is a peak of increased OM content. The maximum content is noted in the uppermost sample of the carbonate section of the profile at the AT1–AT2 boundary. There is a close correlation between OM content and some individual pollen taxa: a positive one with *Corylus* ($r=+0.53$), *Picea* ($r=+0.66$) and a negative one with *Artemisia* ($r=-0.63$) and *Betula* ($r=-0.84$).

The amount of the non-carbonate detrital fraction ranges from 15% to 70%. Minimal amounts are ob-

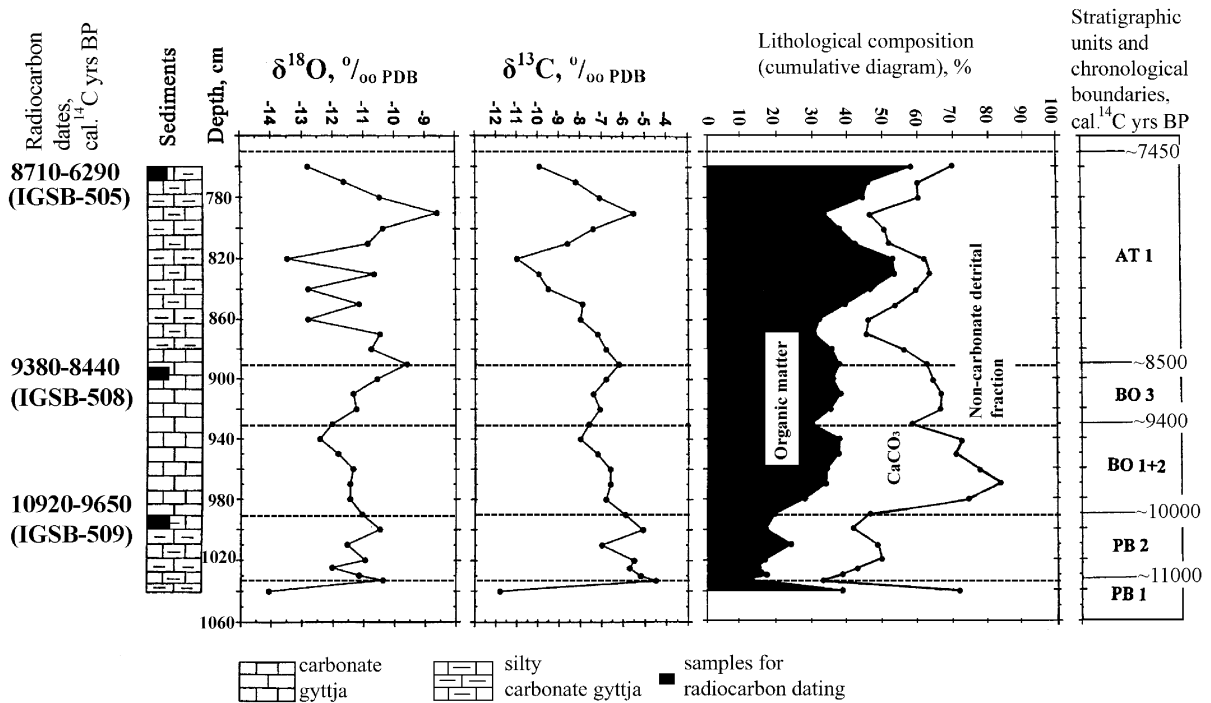


Fig. 10. Stable isotopes in bulk carbonates and the lithological composition of lacustrine deposits, Lake Bezymiannoe.

served in the BO1 + 2 unit. The amount of CaCO_3 in the ash residue falls in the range of 13% to 78%. The sediments, which underlie and overlie the carbonate section, do not react with HCl.

Minimum, maximum and median $\delta^{18}\text{O}$ values are -14.1‰ , -8.6‰ , and -11.3‰ , respectively. The minimum value is consistent with the very top of the PB1 unit (the lowermost sample of the curve). From the top of the PB2 unit, $\delta^{18}\text{O}$ values steadily decrease to the top of the BO1+2 unit, where they exhibit a negative spike. A positive spike is seen at the BO3–AT1 boundary. In the lower half of the AT1 unit there is a series of fluctuations with the general trend of decreasing $\delta^{18}\text{O}$. After a negative peak roughly in the middle of the AT1 unit there is a shift to isotopically heavier carbonates. The maximum for the sequence (-8.6‰) is noted near the top of the AT1 unit.

The $\delta^{18}\text{O}$ values do not show significant correlation with pollen taxa and CaCO_3 content.

Minimum, maximum and median values of $\delta^{13}\text{C}$ are -11.8‰ , -4.5‰ , and -7.1‰ , respectively. In general, changes of $\delta^{13}\text{C}$ are strictly parallel to

those of $\delta^{18}\text{O}$ ($r=+0.69$, $n=31$). Only fluctuations of $\delta^{18}\text{O}$ in the first half of the AT1 unit have no correspondence in the shape of the carbon curve. Values of $\delta^{13}\text{C}$ do not show significant correlation with CaCO_3 (and hence with detrital carbonates) content. There is a close negative correlation between $\delta^{13}\text{C}$ and OM ($r=-0.79$) and positive correlation of $\delta^{13}\text{C}$ with *Betula* ($r=+0.70$) and *Artemisia* ($r=+0.56$). *Picea* and $\delta^{13}\text{C}$ show antipathetic relationships ($r=-0.62$).

6. Interpretation and palaeogeographical synthesis

With a view to interpreting isotope data in terms of palaeoenvironmental changes it seems necessary to consider recent theoretical ideas of stable isotope fractionation in freshwater carbonate precipitating systems, to evaluate theoretically expected values of $\delta^{18}\text{O}$ in Belarusian carbonates, and to understand how palynological and lithological information may help to choose the correct explanation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in the profiles studied.

6.1. Theory of stable isotope fractionation in freshwater carbonate precipitating systems

The precipitation of freshwater carbonates (almost exclusively calcites) is mediated by the photosynthesis of macrophytes, algae and plankton. Under temperate climatic conditions freshwater authigenic calcites usually precipitate in the summer months and in the surface waters (Punning et al., 1988; Jones and Marshall, 2002).

The oxygen isotopic composition of freshwater carbonates is dependent on the isotopic composition and the temperature of the water. The isotopic composition of the water is controlled mainly by mean annual air temperature and evaporation. Dansgaard (1964) found that an increase in average annual temperature by one degree would cause an increase in the $\delta^{18}\text{O}$ of rainfall (and hence of underground and surface waters) by 0.7‰. Dansgaard's regression equation was based on the data obtained from mid- and high northern latitude coastal stations and may not be relevant for the interpretation of isotopic records from continental sites. The availability of data from worldwide monitoring of the oxygen composi-

tion in atmospheric precipitation makes it possible to derive an equation linking $\delta^{18}\text{O}$ and air temperatures for the European part of the former Soviet Union (i.e. for Belarus and adjacent regions). Meteorological and isotope data, which served as the basis for this equation (Fig. 11), were selected from the database published in Rozanski et al., 1993. The slopes of the best-fit lines of Dansgaard's equation and of the equation for the European part of the former Soviet Union are the same, but the intercepts differ by 2‰. It means that at the same mean air temperature, meteoric precipitation in Eastern Europe shows lighter oxygen isotopic composition than at the Atlantic shore, which is caused by the "continental" effect. Data from Belarus are also included in Fig. 11. In Belarus, monitoring of oxygen isotopes in meteoric precipitation has been carried out in Minsk, 53°52'N, 27°32'E, and Brest, 52°07'N, 23°41'E (see Fig. 1) during 44 months from April 1980 to December 1983 (Rozanski et al., 1993). Mean $\delta^{18}\text{O}_{\text{SMOW}}$ in Minsk is -10.79‰ and in Brest is -9.71‰ . It is important that during the monitoring period mean annual air temperatures were 6.8 °C in Minsk and 8.5 °C in Brest. That is 1.5 °C higher than the mean annual

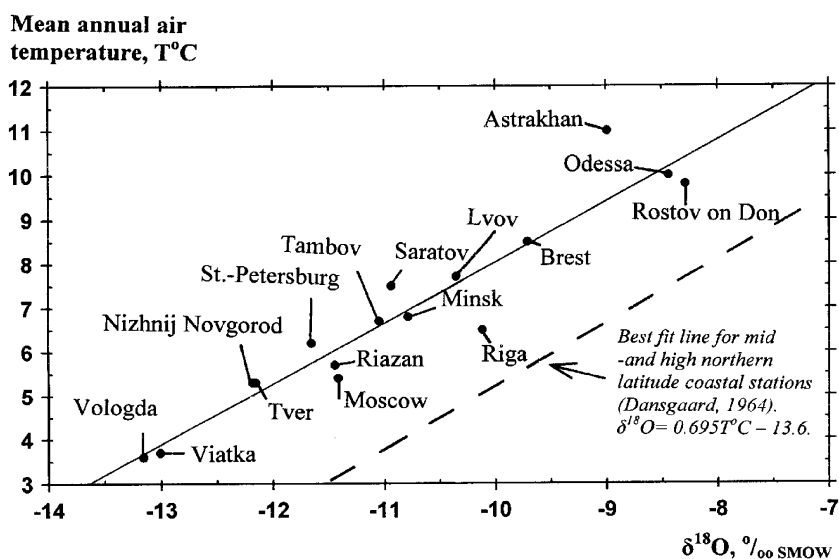


Fig. 11. The ^{18}O content of precipitation for 16 continental stations from the European part of the former Soviet Union. Isotope and meteorological data are selected from the database published in Rozanski et al. (1993). Relationship between $\delta^{18}\text{O}_{\text{SMOW}}$ values in meteoric precipitation and mean annual air temperatures in the European part of the former Soviet Union (i.e. in Belarus and adjacent regions) can be described by the regression equation: $\delta^{18}\text{O} = -15.37 + 0.66T$ °C, $r = 0.953$.

temperatures according to long-term observations (see isotherms in Fig. 1). Using the regression equation from Fig. 11 one may compute that the long-term annual $\delta^{18}\text{O}_{\text{SMOW}}$ in Minsk and Brest should be -11.8‰ and 10.4‰ respectively.

The temperature coefficient for the $\delta^{18}\text{O}$ fractionation between carbonate and water is $\sim -0.24\text{‰}$. That follows from the equation: $T\text{ °C} = 15.7 - 4.36(\delta_c - \delta_w) + 0.12(\delta_c - \delta_w)^2$, where $T\text{ °C}$ = temperature of calcite precipitation, $\delta_c = \delta^{18}\text{O}$ of calcite relative to PDB, $\delta_w = \delta^{18}\text{O}$ of water relative to SMOW (Hays and Grossman, 1991). Using the latter equation it is easy to conclude that from the water with $\delta^{18}\text{O} = -11.8\text{‰}_{\text{SMOW}}$ the calcite with $\delta^{18}\text{O} = -11.8\text{‰}_{\text{PDB}}$ should precipitate at a water temperature of 15.7 °C , which is close to the mean summer temperature of the lake waters in Belarus. Therefore, theoretically, recent carbonates from overflowing systems in Central Belarus (including in the vicinity of the Ptich site) should show values close to $-11.8\text{‰}_{\text{PDB}}$. In closed-basin lakes, the water may be significantly enriched in ^{18}O compared to meteoric precipitation due to evaporation, and calcite precipitates will be isotopically heavier. Kinetic effects (mainly due to high precipitation rates) may be responsible for ^{18}O depletion of the calcite precipitates (Fronval et al., 1995).

The carbon isotopic composition of freshwater carbonates reflects that of dissolved inorganic carbon (DIC). Variations in the ^{13}C content in lake and spring calcites are caused by variations in contributions by several different sources of DIC in the water. Atmospheric carbon dioxide shows a $\delta^{13}\text{C}$ value of -8‰ to -7‰ and calcite in equilibrium with it would show $\delta^{13}\text{C}$ values of approximately $+4\text{‰}$ to $+5\text{‰}$ (Fritz et al., 1987). The equilibrium with atmospheric CO_2 can be achieved only in stagnant waters. ^{12}C rich carbon dioxide is generated in soils due to root respiration and the decomposition of terrestrial organic matter ($\delta^{13}\text{C}$ of C_3 plants = $\sim -27\text{‰}$). This DIC enters the water basin with groundwaters. Calcite precipitated in equilibrium with the soil-derived carbon should show $\delta^{13}\text{C}$ values of -12‰ (Boutton, 1991). Thus, denser vegetation within a drainage basin will cause lower $\delta^{13}\text{C}$ values of lake or spring carbonates. Isotopically light DIC may originate through the release of carbon dioxide by oxidation or anaerobic decomposition (particularly,

due to bacterial sulfate reduction) of organic matter within the water or sediments. The injection of methane (CH_4), which contains extremely light carbon (Galimov, 1968), from nearshore bogs or gleyic soils into a lake may cause precipitation of strongly ^{13}C -depleted calcites. Enrichment of authigenic calcites in ^{13}C occurs due to the photosynthetic removal of ^{12}C from the DIC of lake waters into plants and plankton (McKenzie, 1985), and an increase in lake productivity would cause the precipitation of isotopically heavier calcites. Carbon derived from detrital (mainly marine) carbonates from the catchment and groundwater aquifers shows $\delta^{13}\text{C}$ values fluctuating around 0‰ . During relatively arid climatic phases with longer time of groundwater–rock interaction (Shvartsev, 1978) this isotopically heavy component should be more abundant in the DIC of underground and surface waters.

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of freshwater carbonates may show covariance as well as antipathetic relationships. An anticovariance is a characteristic feature of carbonates from overflowing lakes and of riverine calcareous sediments (Andrews et al., 1994) and is explained as warmer climatic phases (higher $\delta^{18}\text{O}$) consistent with denser terrestrial vegetation and the more active influx of soil-derived carbon in lakes and rivers (lower $\delta^{13}\text{C}$). A positive correlation between oxygen and carbon isotopic compositions is observed mainly in closed lakes. Turner et al. (1983) suggested that with longer residence times (lower lake levels), gas exchange and evaporation combine to enrich the DIC in ^{13}C and the water in ^{18}O . If that is the case, the palaeotemperature signal of $\delta^{18}\text{O}$ is masked by the forcing of hydraulic regime changes. Also, there is another explanation of the $\delta^{18}\text{O}$ – $\delta^{13}\text{C}$ covariance. Drummond et al. (1995) suggested that under colder climatic conditions (lower $\delta^{18}\text{O}$) the season of calcite precipitation is shorter and lake productivity is lower (less active uptake of ^{13}C -depleted DIC from water to aquatic plants and hence lower $\delta^{13}\text{C}$ of lake carbonates). The latter mechanism may be significant in overflowing lakes as well.

High precipitation rates may cause kinetic effects and be responsible for synchronous ^{18}O and ^{13}C depletion of the calcite precipitates. And finally, an admixture of isotopically heavy detrital calcites and dolomites may cause parallel changes in $\delta^{18}\text{O}$ – $\delta^{13}\text{C}$ in bulk freshwater carbonates.

6.2. Lithological and palynological data as a support for the interpretation of Belarusian isotope records

Lithological and pollen data contain certain information about environmental changes and may help to explain changes in isotopic compositions in individual cores.

Increased amounts of OM in the sediments of carbonate lakes (Teklits, Okono, Bezymiannoe) may be linked either to relatively deepwater facies (with low photosynthetic activity of submerged plants and weak carbonate precipitation) or to the sediments deposited during time phases of increased biological productivity in a lake (high trophic status). Therefore, negative correlation between $\delta^{13}\text{C}$ and OM values in the lake cores (as observed in Teklits, Bezymiannoe, and to a lesser extent in Okono) indicates that $\delta^{13}\text{C}$ changes were regulated by changes in the water depth rather than by changes in productivity. In the latter case there would be $\delta^{13}\text{C}$ -OM covariance. In spring tufa Ptich the increased content of OM (mainly peat) should be consistent with the decreased depth of the sedimentation basin and with the decreased discharge of spring water. Thus, the $\delta^{13}\text{C}$ -OM covariance noted in Ptich indicates that isopically heavier carbon in the calcites is characteristic of relatively arid climatic phases.

Increased amounts of detrital fraction may indicate relatively low lake levels (shallow water facies) and/or active erosion in the catchment area (cold or dry climatic phases with poor vegetation cover). As well, analysis of non-carbonate detrital fraction content may help to evaluate the quality of environmental signals contained in the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of bulk carbonates. It is clear that the contents of non-carbonate clastic material are proportional to those of detrital carbonates. As it was shown in the example of Lake Okono, in bulk carbonates containing more than 34% of non-carbonate detrital fraction in the ash residue, allochthonous carbonates are clearly detectable. Therefore, it may be assumed that in such sediments, the stable isotope ratios for bulk carbonates are higher than those for the authigenic phase only.

Correlations between isotopic compositions and individual pollen taxa percentages show that the stable isotope ratios and vegetation dynamics were forced, to some extent, by common factors. The content of pollen taxa may correlate better with $\delta^{18}\text{O}$ (as in

Okono) or with $\delta^{13}\text{C}$ (as in Bezymiannoe). However, significant negative correlation of either $\delta^{18}\text{O}$ or $\delta^{13}\text{C}$ with *Betula* and positive correlation with *Picea* was not revealed. *Artemisia* and NAP show only a negative significant correlation with $\delta^{18}\text{O}$ and positive with $\delta^{13}\text{C}$. Such a pattern of correlations is explained by assuming that *Betula* expanded during warm and/or dry phases (higher $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$), *Picea* dominated during relatively cold and/or humid phases (lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$), and *Artemisia* and NAP were abundant during relatively cold and/or arid climatic phases (lower $\delta^{18}\text{O}$ and higher $\delta^{13}\text{C}$).

6.3. Possible controls on isotope values in the profiles studied

Consideration of average, maximum, minimum values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, as well as a comparison between isotopic, lithological and pollen records enable us to suggest the primary factors controlling isotopic composition changes in each site studied.

In Ptich (see Fig. 2), the median value of $\delta^{18}\text{O}$ (-11.3‰) is close to that which calcite, precipitating from the water with the average isotopic composition of recent atmospheric precipitation in Central Belarus ($-11.8\text{‰}_{\text{SMOW}}$), should show at summer water temperatures (see Section 6.1). All of the absolute values (except for the maximum one) fall into the range between -13.5‰ and -10.1‰ and indicate an insignificant role of evaporative enrichment in ^{18}O of the water. The absence of a significant correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ as well as between $\delta^{18}\text{O}$ and the content of terrestrial organic matter (indicative of the depth of the sedimentation basin) also suggests that in general the $\delta^{18}\text{O}$ changes are independent on hydrological changes. The increased role of the main arboreal species (*Pinus* + *Betula*), which spread at the expanse of non-arboreal species, positively correlates with $\delta^{18}\text{O}$. The minimal value of $\delta^{18}\text{O}$ in the middle of the BO3 unit is consistent with the increased portion of NAP. The maximum value of $\delta^{18}\text{O}$ is registered in the Atlantic, which is traditionally believed to be the warmest period of the Holocene. These results indicate that the changes in oxygen isotope composition of calcites in Ptich might be regulated by changes in $^{18}\text{O}/^{16}\text{O}$ ratios of the water that were linked mainly to air temperature variations.

However, if that is the case, relatively high values of $\delta^{18}\text{O}$ in the YD unit need special explanation. The fact that the amount of detrital (isotopically heavier) carbonate fraction declines, from the bottom of the AL unit to the middle of the PB2 unit, suggests a lighter isotopic composition of Late Glacial authigenic calcites than of Holocene calcites. Relatively cold water in the Younger Dryas also might cause ^{18}O -enrichment of precipitating calcites (see the equation of Hays and Grossman, 1991 in Section 6.1). Moreover, in the glacial melt water channel, where the Ptich site is located, permafrost conditions might be especially severe (compared to the vicinities of lakes), and spring tufa might accumulate only during relatively warm intervals of the stadial.

Absolute values of $\delta^{13}\text{C}$ (-11.6‰ to -8.0‰) and the median -10.1‰ in Ptich are typical for freshwater carbonates precipitated from the flowing water and reflect conditions close to equilibrium with the soil-derived carbon (Andrews et al., 1994). Parallel changes in $\delta^{13}\text{C}$ and in the contents of terrestrial OM, as well as an antipathetic relationships between $\delta^{13}\text{C}$ and dominant arboreal species (*Betula* + *Pinus*) pollen suggest that during dryer phases (with expanding grasses and decreased discharge of spring waters) isotopically heavier carbon from detrital carbonates of the catchment and aquifers entered the water more actively, while the contribution of carbon from the soil organic pool was smaller.

In Teklits (see Fig. 5), the median value of $\delta^{18}\text{O}$ (-13.4‰) and especially light values (to -17‰) in the SB1 unit are lower than those which can be expected for Holocene carbonates precipitated under isotopic equilibrium with $\delta^{18}\text{O}$ of meteoric precipitation, and the neighbouring samples show the values differing by several permil. That makes assume the influence of kinetic effects on the isotopic compositions of the calcites. At the same time, some features (the shifts to more positive values at the Late Glacial–Holocene and Boreal–Atlantic transitions, decreasing $\delta^{18}\text{O}$ after the Atlantic climatic optimum) suggest that main $\delta^{18}\text{O}$ trends might be conditioned by air temperature fluctuations. However, the maximum value in the Atlantic 1 is noted in a carbonate sand layer, deposited during the period of low lake level. Therefore, detrital contamination is partially responsible for this $\delta^{18}\text{O}$ peak. Positive correlation of $\delta^{18}\text{O}$ with *Betula* and negative with *Picea* can be explained by

either temperature or effective moisture controls of the oxygen isotope signals.

The $\delta^{13}\text{C}$ values in Teklits are extremely low ($<-20\text{‰}$) at the bottom and at the top of the sequence in the sediments with increased OM. These low values may indicate anaerobic near-bottom conditions and bacterial sulfate reduction during the stages of increased lake depths or incorporation of methane from the adjacent bogs into lake carbonate precipitating system. However, the median value (-11.9‰) is typical for freshwater carbonates precipitated from the flowing water at equilibrium with the soil-derived carbon (Andrews et al., 1994). Negative correlation between $\delta^{13}\text{C}$ and OM suggests that at higher lake levels (shorter water residence times, higher precipitation–evaporation balance, deepwater sediment facies) the portion of the atmospheric carbon was especially low in DIC of the lake. In Teklits, $\delta^{13}\text{C}$ positively correlates with NAP and negatively with *Picea* which supports the assumption that $\delta^{13}\text{C}$ was controlled by effective moisture and, hence by lake depth and by the water residence time.

In Okono (see Fig. 7), low $\delta^{18}\text{O}$ values for bulk carbonates deposited in the Younger Dryas (and even lower values which may be supposed for pure authigenic phase), the pronounced shifts to more positive $\delta^{18}\text{O}$ values in the end of the Younger Dryas and near the Boreal–Atlantic boundary evidence for the temperature genesis of $\delta^{18}\text{O}$ variations. Positive correlation of $\delta^{18}\text{O}$ with *Betula* and negative with *Picea* might rise due to either temperature or hydraulic control of oxygen isotopic composition, but negative correlation of $\delta^{18}\text{O}$ with *Artemisia* proves the temperature rather than hydrological control. Together with this the median value (-10.3‰) as well as the values $>-8\text{‰}$ around the BO2–BO3 boundary and extremely high values (-2.5‰ to -1.9‰) unexpected for freshwater carbonates at the top of the PB2 unit prove that the $\delta^{18}\text{O}$ values were controlled not only by the air temperature changes (at least at certain time spans). The magnitude of the shift to extremely high $\delta^{18}\text{O}$ values in the PB2 unit is not easily explained, because there is no lithological or palynological evidence of low lake level or arid climatic phase, and hence of active evaporative ^{18}O enrichment of the water. Hypothetically, this strong positive $\delta^{18}\text{O}$ spike may be partly explained by the discharge of the groundwaters essentially oversaturated with respect

to calcite what allowed calcite precipitation not only in summer but during the whole year. According to the equation of Hays and Grossman, 1991 (see Section 6.1), calcite precipitated at water temperature close to 0 °C will be ~ 4‰ heavier than the calcite precipitated at summer water temperature (about 16 °C).

The absolute values of $\delta^{13}\text{C}$ (–3.4‰ to +1.0‰) and especially the maximum one are typical for carbonates precipitated at near equilibrium with the isotopic composition of the atmospheric CO_2 carbon, and are characteristic of the lakes with a long water residence time. Similar values fluctuating around 0‰ were recorded in Belarusian Lake Naroch (Punning et al., 1988) and in many European marl lakes: Lobsigensee, Switzerland (Eicher and Siegenthaler, 1984), Gerzensee and Leysin, Switzerland (Schwander et al., 2000), Vanstads, southern Sweden (Hamarlund and Keen, 1994), Perespilno, Biskupin, Poland (Kuc et al., 1996), and Goscianz, Strazym, Mikolajki, Poland (Kuc et al., 1993). A lack of significant correlation between carbon isotopic composition and the percentages of individual pollen taxa makes us doubt that the $\delta^{13}\text{C}$ changes were caused by regional environmental forcing. Probably, they were controlled by local within-lake processes, such as minor fluctuations of primary carbonate productivity. $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ covariance in the sediments of climatologically unstable Late Glacial time might be connected to changes in photosynthetic activity of macrophytes (photosynthesis was less active during colder phases and $\delta^{13}\text{C}$ were lower) as well as to fluctuations of hydrological regimes (colder phases were consistent with more flowing regimes, and the portion of atmospheric carbon was less in DIC of lake waters). Along with this, the spikes of increased OM (the sediments of deepwater facies) at the PB1–PB2 boundary and in the BO1+2 unit correspond to relatively low $\delta^{13}\text{C}$. That is why we may suppose that the most pronounced changes of lake levels, caused by changes in relative humidity, are reflected in changes of carbon isotopic composition.

In Lake Bezymiannoe (see Fig. 11), bulk carbonates contain abundant detrital fractions, and $\delta^{18}\text{O}$ changes are perfectly parallel to those of $\delta^{13}\text{C}$. That may suggest the essential influence of the isotopic signals of allochthonous carbonates on the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of bulk carbonates. On the other hand, the

median $\delta^{18}\text{O}$ value of –11.3‰ and the values within negative spikes are close to those, which are expected for pure Holocene authigenic calcites. A lack of close correlation between the isotope values and CaCO_3 content, as well as the absence of carbonate minerals from the sediments of late Atlantic, Subboreal and Subatlantic indirectly point to a non-significant portion of detrital carbonates in carbonate gyttja of Lake Bezymiannoe. Therefore, a close $\delta^{18}\text{O}$ – $\delta^{13}\text{C}$ covariance may be explained in the following way. Changes in $\delta^{13}\text{C}$ values varying from –11.8‰ and –4.5‰ mainly reflect the proportion of the soil-derived and atmospheric carbon in the DIC of the lake, and indicate the flow-through regime and lake level changes, and thus the changes in the precipitation–evaporation balance. The fact that lower $\delta^{13}\text{C}$ is coincident with organic rich sediments of relatively deepwater facies, with the deficiency of *Betula* and *Artemisia*, with abundant *Picea*, and thus with more humid climatic phases supports this interpretation. The isotopic signal of $\delta^{18}\text{O}$ duplicates the information contained in $\delta^{13}\text{C}$ variations: phases of active evaporation partially coincident with higher air temperatures (positive spikes) and the phases of shorter water residence times (negative spikes).

Thus, the dominant control on the carbonate precipitating systems might vary from site to site and over different time scales. Nevertheless, with a range of possible factors controlling the change in each site isotope record, common trends, probably caused by the most drastic climate changes, appear in different freshwater sediment records from across Belarus.

6.4. Common features of the isotopic curves from Belarus

In order to emphasize the common features of the isotope records, to enable a more rigorous interpretation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ changes, and to recognize major palaeoclimatic events at the regional scale, we compared all the isotopic curves, plotting the values against time. Chronological boundaries between the periods of the Late Glacial and Holocene, established by pollen zonation and radiocarbon dating, were used as reference marks. Sedimentation rates were assumed uniformly between these levels. We decided not to consider the post-Boreal sections of the isotopic curves for Okono because of very slow sedimentation

rates in the lake after the Boreal. The upper parts of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves derived from AT2 to SB1 sediments from Teklits were also not analysed because they have no counterparts in other profiles.

6.4.1. Correlation between $\delta^{13}\text{C}$ data from freshwater carbonates

The comparison of four $\delta^{13}\text{C}$ curves shows the existence of several common features (or carbon isotope events), which are indicated in Fig. 12 by the letters A–H. Some lithological and palynological peculiarities as well as oxygen isotope peaks, which seem to be helpful for interpretation, are also depicted. At the Late Glacial–Holocene transition, strong shifts from the minimal values (A) to the maximum or almost maximum values (B) are seen in all the cores. These changes in carbon isotope curves may be explained as a transition from high lake levels and increased discharge of springs (event A) to low lake levels and decreased discharge of spring waters (event B). Lithological features support this interpretation of

$\delta^{13}\text{C}$ shifts. Within these shifts, the event A is coincident with an abrupt decrease in the detrital fraction in all of the cores, with negative $\delta^{18}\text{O}$ peaks in all of the lakes, and with peaks of increased OM in Teklits and Bezymiannoe. The event B corresponds to a layer of peaty tufa in Ptich and to positive oxygen peaks in Teklits and Bezymiannoe. Similar hydrological changes, approximately for the same time interval, were recently recognized in Belarus and elsewhere in Europe. Zernitskaya et al. (1998) reconstructed level changes of 25 Belarusian lakes at 500-year chronological slices as the result of the palaeobotanic and lithological analyses of lacustrine sediments. According to these data, predominantly high lake levels were changed to predominantly low ones between 10,500 and 10,000 ^{14}C year BP (between $\sim 12,400$ and 11,250 cal. ^{14}C year BP). In particular, such changes were revealed in Lake Dolgoe, located 3 km to the southeast of the Bezymiannoe site (Zernitskaya et al., 2001). Pazdur et al. (1995) estimated secular changes of the ^{14}C dilution factor of total DIC and thus

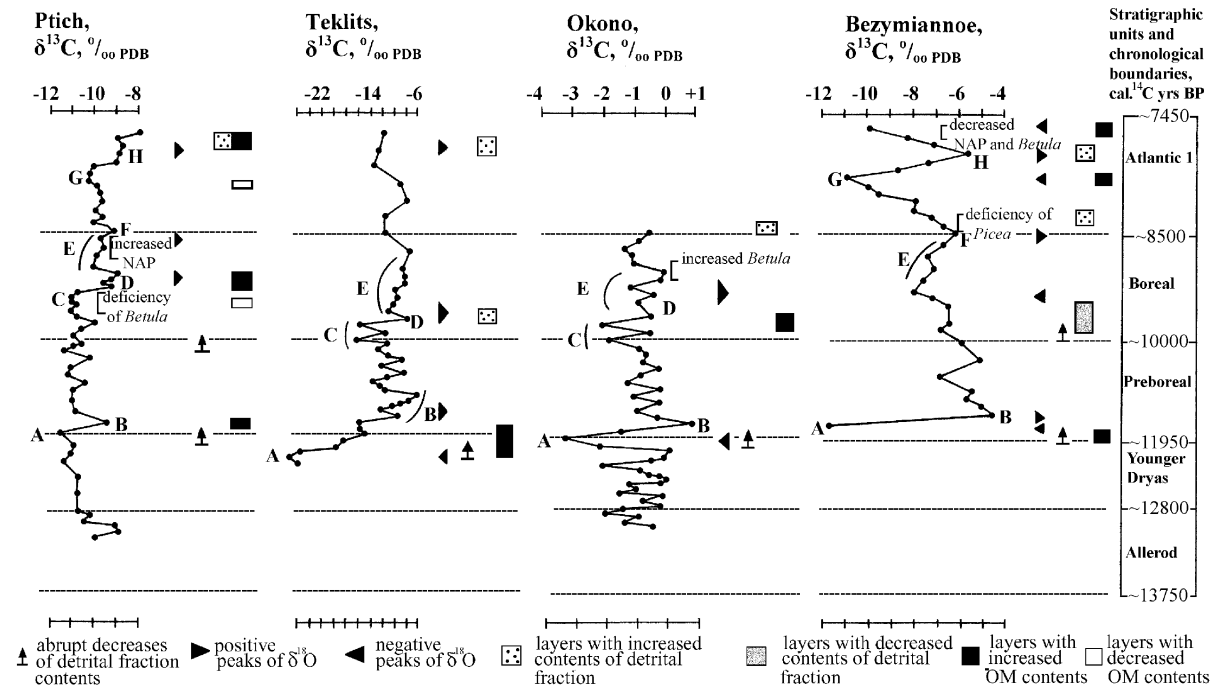


Fig. 12. A comparison between $\delta^{13}\text{C}$ data from freshwater carbonates, Belarus. Basic features of carbon isotope curves (or carbon isotope events) are indicated by the letters A–H. Events A, C, G are attributed to the phases of high lake levels and increased discharges of spring waters, and events B, D, F, and H, to low lake levels and less active discharge of spring waters. The transition from event A to event B is explained as due to the active degradation of permafrost in the region. Trend E reflects monotonically decreasing depths of sedimentation basins.

reconstructed mean levels in Lake Goscianz, Central Poland. According to their reconstruction, the most notable changes of lake depth occurred around the Late Glacial–Holocene transition; at 10,500 ^{14}C year BP ($\sim 12,400$ cal. ^{14}C year BP) the level of Lake Goscianz was very high, and by 9500 ^{14}C year BP ($\sim 10,600$ cal. ^{14}C year BP) the lake became considerably more shallow. Changes of lithological properties of Belarusian lake carbonates at the Late Glacial–Holocene transition also have analogues in other European sites. For instance, increased amounts of humic compounds in marls of the Late Glacial and of the beginning of the Holocene are frequently attributed to the existence and/or degradation of the permafrost and to the influx of reworked terrestrial OM from the active layer of soils into lake basins. Thus, in southern Sweden (Vanstads mosse site) such an event was noted at roughly 10,200 ^{14}C year BP or at $\sim 11,950$ cal. ^{14}C year BP (Hammarlund and Keen, 1994). We probably observe the same event in our carbon isotope curves. At the termination of the Younger Dryas or at the Younger Dryas–Preboreal transition, lake levels rose and discharges of spring waters increased due to the influx of permafrost melt waters. Moreover, the DIC of these waters might have been strongly enriched in ^{12}C , produced by anaerobic decomposition of organic matter in thawing soils as well as by the input of methane into carbonate precipitating systems (note the extremely low $\delta^{13}\text{C}$ near the base of the core from Teklits). The termination of the episode of permafrost decay led to an abrupt drop in the lake level and to decreased discharge of spring waters and hence to a strong enrichment in the ^{13}C of carbonates. Taking into account the very high $\delta^{13}\text{C}$ values (event B in Fig. 12) we suggest that the abrupt fall of lake levels was also caused by an episode of climate aridification.

The next significant quasi-simultaneous change of $\delta^{13}\text{C}$ to more positive values occurred roughly in the middle of the Boreal (the transitions from event C to event D, which are seen in Ptich, Teklits and Okono). The occurrence of event C corresponds to very small amounts of OM, to a deficiency of *Betula* pollen in Ptich, to increased aquatic OM in Okono, and to a smaller portion of the detrital fraction in Bezymianoe. The event D corresponds to increased terrestrial OM in Ptich, positive oxygen peak and abundant detrital fraction in Teklits. Interestingly, in the second

half of the Boreal, monotonic trends of increasing $\delta^{13}\text{C}$ values (event E) are observed in Ptich, Teklits, Bezymianoe and possibly in Okono. Increased NAP in Ptich and *Betula* in Okono are noted in the second half of the Boreal. So, we may attribute this change (from event C to event D and further to event E) in carbon isotopic composition to a fall in lake levels (a climatic aridification). Support for these conclusions comes from the reconstructions of Zernitskaya et al. (1998), who found that at 9000 ^{14}C year BP (~ 10000 cal. ^{14}C year BP), lake levels in Belarus were predominantly high, while at 8500 ^{14}C year BP (~ 9470 cal. ^{14}C year BP), low lake levels prevailed in the region. Thus, during this time interval the shift from high to low lake levels occurred for the second time in the Late Glacial Holocene history. The low lake level phase was recognized in the middle of the Boreal in Lake Dolgoe as well (Zernitskaya et al., 2001). According to the European lake level database (Harrison et al., 1991), at 9000 ^{14}C year BP (~ 10000 cal. ^{14}C year BP) climatic conditions were apparently drier (lower lake levels) than at present in northern and central Europe, and wetter (higher lake levels) than at present in southern, western and northeastern Europe. So, we may suggest a quasi-synchronicity of humid event C (depicted in Fig. 12) with wetter conditions in southern, western and northeastern Europe. However, it should be emphasized that we are not sure of the precise synchronicity between the Belarusian chronological scale and chronological scales used in the European lake level database.

In the younger lacustrine and spring deposits in Belarus (within early Atlantic sections of carbon isotope curves), some synchronous changes of $\delta^{13}\text{C}$ are seen: events F and H (relatively arid phases, low lake levels and decreased discharges of spring waters), and event G (relatively humid phase, higher lake levels and increased discharges of spring waters). Such an interpretation of events F, G, H can be confirmed by lithological data.

Fig. 12 shows that events A–D in different profiles are not absolutely synchronous. On the one hand, this may be explained by imperfections in our correlation model. On the other hand, individual water systems might respond to regional climate changes quasi-synchronously rather than simultaneously. That would be especially relevant for events such as the active

degradation of permafrost as a consequence of rapid warming in the region.

Considering Fig. 12, one should pay attention to the fact that the magnitudes of carbon isotope composition changes are essentially different in four sites. This would be clear if the four profiles were plotted on the same $\delta^{13}\text{C}$ scale. Such a peculiarity can be explained as follows. Individual freshwater carbonate precipitating systems are characterized by different degrees of stability. The same climatic driver could disturb the equilibrium in one lake only slightly (as in the stable system of Okono) but may cause drastic changes in the $\delta^{13}\text{C}$ of DIC in other case (as in Teklits). Moreover, qualitatively the same shifts in $\delta^{13}\text{C}$ in individual basins as responses to a climate change might arise

due to different changes in the proportions of contributions from different sources of DIC. For instance, the shift from lower $\delta^{13}\text{C}$ values to higher ones as a response to a fall in lake levels might occur due to less active sulfate reduction or reduced input of methane (as in Teklits), to an increased amount of DIC derived from detrital carbonates (as in Ptich), or to a higher equilibration with isotopically heavy atmospheric carbon (as in Okono).

6.4.2. Correlation between $\delta^{18}\text{O}$ data from freshwater carbonates

Four Belarusian oxygen curves plotted against time are given in Fig. 13. The most important peaks and spikes (oxygen isotope events) are indicated by the

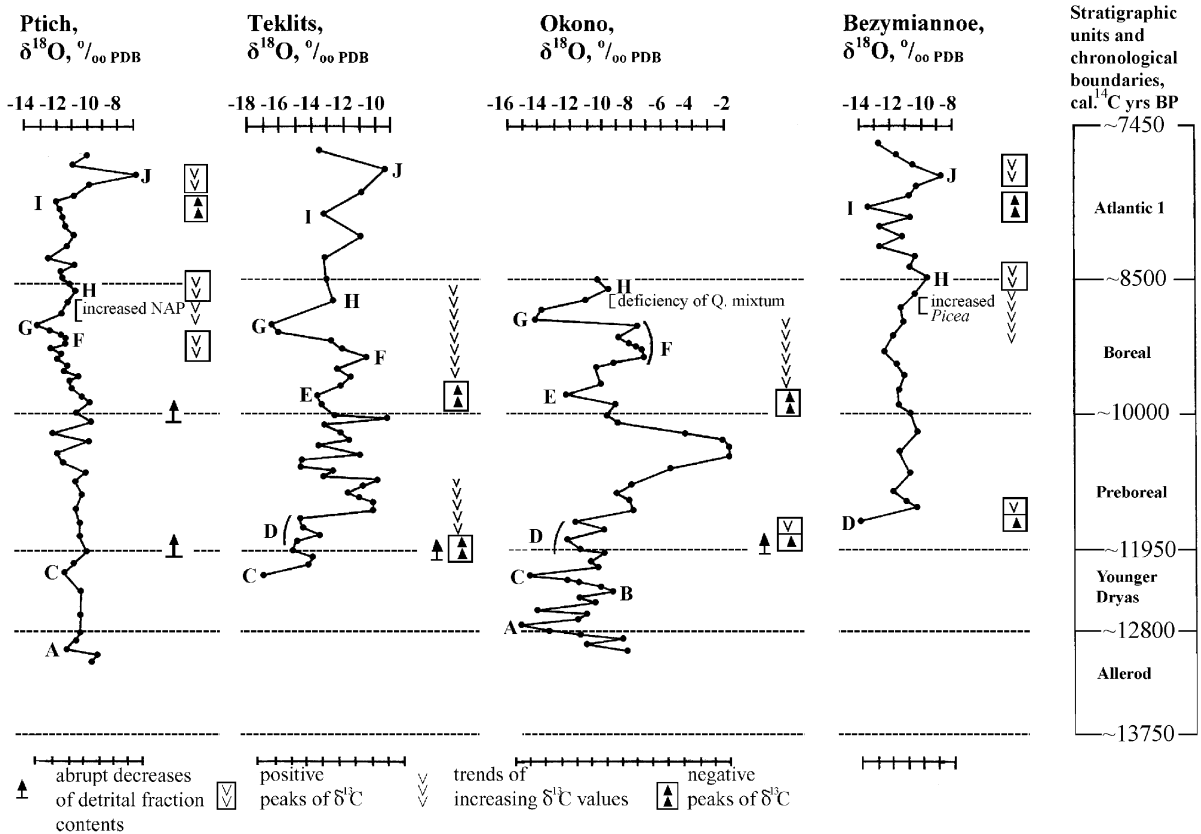


Fig. 13. A comparison between $\delta^{18}\text{O}$ data from freshwater carbonates, Belarus. Basic features of oxygen isotope curves (or oxygen isotope events) are indicated by letters A–J. Events A, C, G, I are interpreted as cooling peaks. Event D is attributed to an influx of melt water in sedimentation basins during a phase of active decay of permafrost. Event E reflects a relatively humid phase, and event F a relatively arid one. Events B, H, J are explained as warming peaks. In the early Atlantic, warmer episodes were consistent with time intervals of a decreased precipitation–evaporation balance.

letters A–J. The main features of carbon isotopic curves, lithological compositions and pollen spectra are also depicted.

The most detailed curve for the Younger Dryas in Belarus is derived from Okono. The structure of the Younger Dryas includes two main negative spikes (A and C) divided by a positive spike (B). This qualitatively resembles that found in the oxygen curves from the Swiss lakes Gerzensee and Leysin, and the GRIP ice core (Schwander et al., 2000), as well as from the lakes Hawes Water, England, Ammersee, Germany (Jones and Marshall, 2002), Goszczan, Poland (Kuc et al., 1996), where lower values of $\delta^{18}\text{O}$ are clearly attributed to lower air temperatures. In all of the above-mentioned curves, the analogues of events A, B, C, as well as the general trend of increasing $\delta^{18}\text{O}$ through the Younger Dryas can be recognized. However, the positive peak (B) in the middle of the stadial is much more pronounced in Okono than in the above-mentioned curves. It has been suggested (Makhnach and Kolosov, 2003) that if air temperature fluctuations reflected in $\delta^{18}\text{O}$ changes in the Younger Dryas in Okono were simultaneous to those revealed in GRIP, Leysin, and Goszczan, the chronology of the Late Glacial in Okono appears to be ~ 400 years old. However, we understand that the synchronicity of climatic shifts in the Late Glacial in different European localities are still to be proven by accurate dating of events in each site and region (Lowe et al., 2001). The structure of the Younger Dryas section of the oxygen curve, qualitatively similar to that found in Okono, may be recognized in Ptich, although the magnitude of events in the latter profile is essentially smaller (see explanation in Section 6.3).

In the beginning of the Preboreal, negative spikes in Teklits and Okono and a low value in Bezymiannoe (event D) are seen. This event can be linked either to the so-called Youngest Dryas, a well known early Preboreal oscillation (Bjorck et al., 1997), or to an episode of discharge of the permafrost melt waters which contained relict isotopic signatures. The latter explanation is in good agreement with $\delta^{13}\text{C}$ and lithological records.

In the rest of the Preboreal, the shapes of the oxygen curves are individual, and the palaeogeographical interpretation of the oxygen curves for the PB2 appears to be debatable.

In Teklits and Okono, the shapes of oxygen curves for the first half of the Boreal strongly resemble each

other (events E, F). Probably, these peaks reflect the changes in water evaporation and hence in flow-through regimes, lake levels and relative humidity: the transition from higher lake levels (E) to lower ones (F). Such an interpretation is in good agreement with carbon isotope data. Event F may also be recognized in Ptich, although it is weakly expressed in the sediments of this overflowing system. In the Boreal section of the oxygen curve from Bezymiannoe, there is no similarity to other curves, although the supposed arid phase (event F in other oxygen curves) corresponds in time to the beginning of a trend in increasing values in the middle of the Boreal in Bezymiannoe.

Synchronous very low (the minimal for the Holocene) $\delta^{18}\text{O}$ values (event G) are recorded roughly in the last third of the Boreal in all of the profiles except for Bezymiannoe, where $\delta^{18}\text{O}$ changes are less dependent on air temperature fluctuations. These minimal $\delta^{18}\text{O}$ values correspond to the absence of *Ulmus*, *Alnus*, *Corylus* and abundant NAP in Ptich, to a deficiency of *Q. mixtum* (*Ulmus*, *Corylus*, *Tilia*) in Okono, and to increased *Picea* in Bezymiannoe. These results are evidence that the $\delta^{18}\text{O}$ negative spikes (G) reflect a short-lived cold event. The trends of increasing $\delta^{18}\text{O}$ values in hydrologically sensitive lake Bezymiannoe, as well as increasing $\delta^{13}\text{C}$ values in the second half of the Boreal in all profiles, suggest that this cold event might occur parallel to decreased relative humidity and might be linked to an episode of increased continentality of climate and to the increased role of easterlies ($\delta^{18}\text{O}$ -depleted meteoric precipitation). The magnitude of peak G supports such an interpretation.

At the very end of the Boreal and Atlantic 1, the synchronous events H, I, J occur in carbonates of the overflowing systems of Ptich and Teklits, as well as in the hydrologically sensitive system of Bezymiannoe. That suggests that in the early Atlantic, warming events (H, J) were consistent with the reduction of effective moisture, and a cooling event (I) was parallel to increased relative humidity.

6.5. Oxygen isotopes of freshwater carbonates from Belarus and their correlation to the GISP2 ice core

In our previous work (Makhnach et al., 2000), we suggested that strong negative $\delta^{18}\text{O}$ peaks, seen in

three Belarusian oxygen isotope curves at the end of the Boreal, may reflect the prominent 8.2 cal. ka cold event that has been documented in Greenland and other archives (Stuiver et al., 1995; Alley et al., 1997; Klitgaard-Kristensen et al., 1998). If that is the case, some other palaeotemperature changes at a supra-regional scale may also be reflected both in Greenland ice cores and in Belarusian isotope records. In order to test this suggestion we compared $\delta^{18}\text{O}$ records from the GISP2 Greenland ice core and from calcareous tufa Ptich. We previously showed that the $\delta^{18}\text{O}$ values of the carbonates that accumulated in the overflowing water system of Ptich from the Preboreal 2 to the Atlantic 1, and that were composed of pure authigenic calcite (almost 100% of CaCO_3 in the ash residue), bear more precise air temperature signals than in other profiles.

Using the 8.2 cal. ka cold event and some other peaks and troughs around it as time markers, we plotted $\delta^{18}\text{O}$ values from Ptich against those from GISP2 in order to achieve the best intuitive matching of the curves. Bidecadal data from GISP2 (Stuiver et al., 1995) were averaged over 100 year intervals. As seen in Fig. 14, all of the trends and spikes of the GISP2 curve have their counterparts in Ptich. Thus we reached close correlation between the $\delta^{18}\text{O}$ values from GISP2 and the corresponding values in the Ptich curve ($r = +0.57$, $n = 31$). However, there is an obvious discrepancy between the time scales of the compared curves, because the ages for GISP2 are indicated in calendar years, and in Ptich, in radiocarbon years. The correlation suggests a 600–1000 year age difference between climatic events in the Ptich valley and what are considered to be comparable

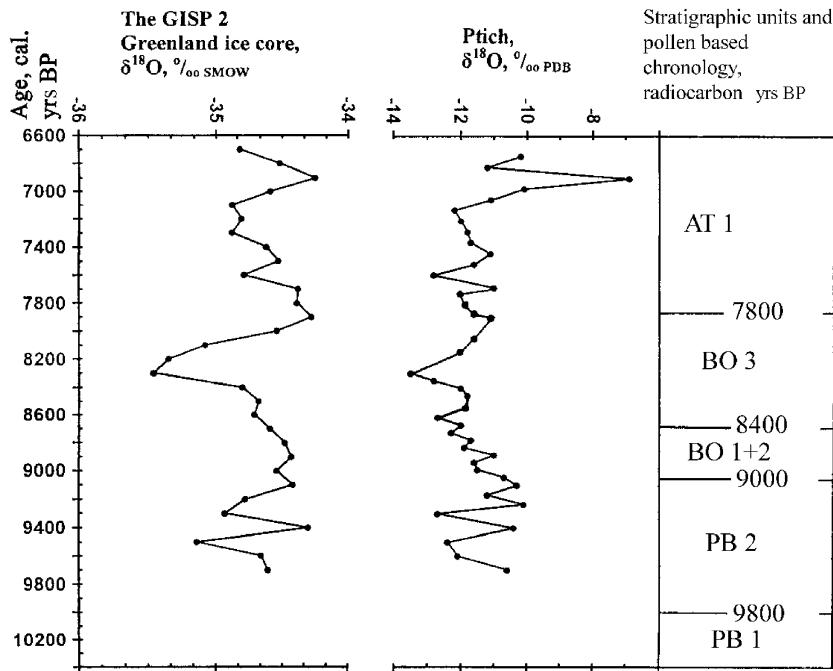


Fig. 14. A comparison between $\delta^{18}\text{O}$ data from the GISP2 Greenland ice core waters and from the authigenic calcites of the calcareous spring tufa, Ptich, Belarus. Bidecadal data from GISP2 (Stuiver et al., 1995) are averaged over 100-year intervals. Note that the ages for GISP2 are indicated in calendar years, and in Ptich, in radiocarbon years. Applying the calibration program Cal-20 (Van der Plicht, 1993) to the age model proposed for Ptich we can roughly compare the two scales in calendar years. The PB–BO boundary in Ptich (9000 ^{14}C year BP = \sim 9990 cal. year BP) corresponds to 9000 cal. year BP in GISP2; the lower boundary of the BO3 in Ptich (8400 ^{14}C year BP = \sim 9410 cal. year BP) corresponds to \sim 8650 cal. year BP in GISP2; the BO–AT boundary in Ptich (7800 ^{14}C year BP = \sim 8500 cal. year BP) corresponds to 7900 cal. year BP in GISP2. Thus, the correlation suggests a 600–1000 year age difference between climatic events in the Ptich valley and what are considered to be comparable events in the GISP2 record.

events in the GISP2 record. It is clear that the causally linked events in Ptich and Greenland may not have happened exactly simultaneously, but if significant time lags had occurred, the signal would probably have been attenuated. Thus, if the correlation between oxygen isotope trends in Ptich and in GISP2 is not a coincidence, we should suggest that the $\delta^{18}\text{O}$ variations in Ptich really reflect air temperature changes at a supra-regional scale, but the chronology of the Early and Mid-Holocene in Belarus is approximately 600–1000 years too old.

7. Conclusions

A comparison of $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ ratios in bulk carbonates with pollen spectra and lithological compositions of four Late Glacial to mid-Holocene spring and lacustrine carbonate deposits of Belarus enabled the recognition of primary factors which controlled $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in carbonate precipitating systems with different hydrological regimes. ^{18}O variations in the overflowing spring system of Ptich mainly reflected the changes in mean annual air temperature. In the carbonates of hydrologically sensitive lake Bezymiannoe, variations in oxygen isotope compositions were closely linked to fluctuations of the water residence time and thus of net evaporation. In lakes Teklits and Okono, $\delta^{18}\text{O}$ trends are conditioned both by temperature and hydraulic regime changes, and the dominant control might vary over different time scales.

Variations in $\delta^{13}\text{C}$ values were controlled by a combination of such factors as water residence time, lake productivity, and vegetation density in the catchment areas. The most drastic changes in sedimentation basin depths and flow-through regimes, particularly those caused by permafrost decay and by fluctuations in precipitation–evaporation balance, were clearly reflected in the carbon isotope curves.

Variations in isotopic composition and in the dynamics of terrestrial vegetation were regulated by more or less common climatic forcing. *Betula* was expanding during relatively warm and/or arid phases (higher $\delta^{18}\text{O}$ and/or $\delta^{13}\text{C}$), *Picea* was spreading during relatively cold and/or humid phases (lower

$\delta^{18}\text{O}$ and/or $\delta^{13}\text{C}$), *Artemisia* and non-arboreal species were abundant during relatively cold (lower $\delta^{18}\text{O}$) or arid phases (higher $\delta^{13}\text{C}$).

A comparison of four pairs of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ curves from Belarus revealed a number of palaeoclimatic events at the regional scale. Two main short-lived cold events interrupted by a warmer interval occurred in the Younger Dryas. A short-term episode of active permafrost decay at the end of the Younger Dryas and at the very beginning of the Preboreal caused an abrupt rise of lake levels followed by their rapid fall. A shift toward a lower precipitation–evaporation balance and lower lake levels took place roughly in the middle of the Boreal. A trend of decreasing temperatures throughout most of the Boreal culminated in a short-lived very cold phase at ~ 8100 ^{14}C year BP (~ 9050 cal. ^{14}C year BP) according to the proposed age model. In the Atlantic 1 warmer phases were coincident with phases of lower precipitation–evaporation balance. An episode of significant short-term warming occurred in the last third of the Atlantic 1.

An intuitive correlation of the $\delta^{18}\text{O}$ record, derived from pure authigenic calcites (PB2–AT1), compared the well-drained system of Ptich with that of the GISP2 Greenland ice core, and allowed good matching of the two datasets. However, the correlation suggests essential age differences between palaeotemperature events in the Ptich valley and what are considered to be comparable events in the GISP2 record. This suggests that the temperature changes in the early to mid-Holocene in Belarus followed supra-regional (probably hemisphere scale) trends, but Belarusian chronology based on pollen zonation appears to be 600–1000 year too old and should be clarified in the future. A precise AMS-dating of marked changes in oxygen isotope curves could be recommended as a way to improve the regional chronological scale.

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