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Late glacial and Holocene palaeoenvironmental changes in the Rostov-Yaroslavl' area, West Central Russia

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Abstract

Three lake sediment sequences (lakes Nero, Chashnitsy, Zaozer'e) from the Rostov-Jaroslavl' region north of Moscow were studied to provide information on palaeoclimatic and palaeoenvironmental changes during the past 15,000 cal yr. The multi-proxy study (i.e., pollen, macrofossils, mineral magnetic measurements, total carbon, nitrogen and sulphur) is chronologically constrained by AMS ^{14}C measurements. Lake Nero provided the longest sedimentary record back to ca. 15,000 cal yr BP, while sediment accumulation began around ca. 11,000 cal yr BP in the two other lakes, possibly due to melting of permafrost. Limnic plant macrofossil remains suggest increased lake productivity and higher mean summer temperatures after 14,500 cal yr BP. While the late glacial vegetation was dominated by *Betula* and *Salix* shrubs and various herbs, it appears that *Betula* sect. *Albae* became established as early as 14,000 cal yr BP. Major hydrological changes in the region led to distinctly lower lake levels, starting 13,000 cal yr BP in Lake Nero and ca. 9000 cal yr BP in lakes Chashnitsy and Zaozer'e, which are situated at higher elevations. These changes resulted in sedimentary hiatuses in all three lakes that lasted 3500–4500 cal yr. Mixed broad-leaved – coniferous forests were widespread in the area between 8200 and 6100 cal yr BP and developed into dense, species-rich forests between 6100 and 2500 cal yr BP, during what was likely the warmest interval of the studied sequences. Agricultural activity is documented since 500 cal yr BP, but probably began earlier, since Rostov was a major capital by 862 A.D. This apparent gap may be caused by additional sedimentary hiatuses around 2500 and 500 cal yr BP.

Introduction

Late Quaternary palaeoenvironmental change in regions adjacent to the North Atlantic is relatively well known from numerous multi-disciplinary and high-resolution studies. However, information from continental areas situated further to the east is still limited, although results from a number of recent investigations are now gradually emerging. Those with a focus on the westernmost part of European Russia addressed, for example, the extent of the Scandinavian ice sheet during the Weichselian/Valdaian (Larsen et al. 1999; Svendsen et al. 1999; Lunkka et al. 2001; Saarnisto and Saarinen 2001), periglacial climatic and environmental conditions during the last glaciation (Hubberten et al. 2004), and Late Weichselian and Holocene climatic and environmental conditions (e.g., Borisova 1997; Khotinsky and Klimanov 1997; Tarasov et al. 1998; Arslanov et al. 1999; Tarasov et al. 1999; Wohlfarth et al. 1999, 2002, 2004; Elina et al. 2000; Subetto et al. 2002; Velichko et al. 2002).

Information from areas that were situated south of the maximum extent of the Eurasian Ice Sheet is scarce and mainly restricted to studies published in Russian. Compilations of these studies indicate that continuous permafrost reached 53° N and that scattered permafrost may have extended as far south as 46° N during and after the Last Glacial Maximum (LGM) (Velichko et al. 2002). Maps illustrating vegetation types during the LGM (Frenzel et al. 1992) suggest that periglacial tundra and steppe occupied the central part of the East-European Plain. Similar results were obtained by Tarasov et al. (2000), who reconstructed the distribution of LGM biomes in northern Eurasia. Although the hypothesis of LGM refugias for broad-leaved, temperate tree species within the Volga River valley, the Southern Urals and the southern part of the Middle Russian Upland (Grichuk 1984) could not be corroborated by the biome reconstruction of Tarasov et al. (2000), studies by Velichko (1984) and Velichko et al. (2002) hint at the existence of boreal trees south of the ice sheet and in habitats with more favourable microclimatic conditions. However, the recent overview by Hubberten et al. (2004), which summarises the climatic and environmental development in periglacial areas of the northwestern Russian Plain, indicates arid and cold conditions with annual temperatures of below −6 °C close to the former ice margin and treeless Arctic and subarctic vegetation during the early part of the LGM and between 18,000 and 15,000 cal yr BP.

Cold, arid conditions and treeless vegetation have also been reconstructed for the late glacial period (14,000–12,000 cal yr BP) for the westernmost part of European Russia, from areas formerly covered by the Scandinavian ice sheet (Wohlfarth et al. 1999, 2002; Subetto et al. 2002; B. Wohlfarth unpublished data). These findings, however, contrast with those obtained farther to the northeast, close to the Ural Mountains at 66° N and 59 ° E, where plant macrofossil evidence suggests that tree *Betula* and scattered stands of *Picea* were present by about 12,500 cal yr BP and that mean July temperatures may have been as high as 13 °C (Väliranta et al. in press). These results agree well with those of Khotinsky and Klimanov (1997), who argued that *Picea* was present in the central Russian Plain during the late glacial period. Stagnant ice and local climatic conditions, as well as the closeness to the Baltic Ice Lake, may have been responsible for the diverging environmental development in the northwestern part of European Russia (Wohlfarth et al. 2002).

The early presence of *Picea* and *Betula* in northeastern European Russia led to their rapid expansion at the beginning of the Holocene (Väliranta et al. in press). Several recent studies northwest of the Ural Mountains (Väliranta et al. in press; Kultti et al. 2004; Sarmaja-Korjonen et al. 2004) indicate that the early Holocene was marked by a climatic optimum (10,000–6000 cal yr BP) with summer temperatures 3–4 °C higher than present and possibly accompanied by increased moisture (Kultti et al. 2003). This increase in summer temperatures occurred at about the same time as in the western part of European Russia (Wohlfarth et al. 1999, 2002; Subetto et al. 2002), and may be linked to increased summer insolation (Kutzbach et al. 1993). Age estimates for the termination of the thermal maximum in northeastern European Russia differ somewhat, but it seems that cooler conditions prevailed between 6300 and 5000 cal yr BP and were followed by gradual cooling, accelerated paludification and permafrost aggradation (Väliranta et al. in press; Kultti et al. 2003, 2004; Sarmaja-Korjonen et al. 2004). Although the temperature trend during the Holocene was broadly similar between different regions of northernmost European Russia, different regions likely experienced different degrees of humidity. While higher-than-present humidity has been reconstructed for northeast European Russia during the early Holocene (Kultti et al. 2003), dry conditions prevailed on the Kola Peninsula (Solovieva and Jones 2002). A similar development has also been observed for northern Finland,

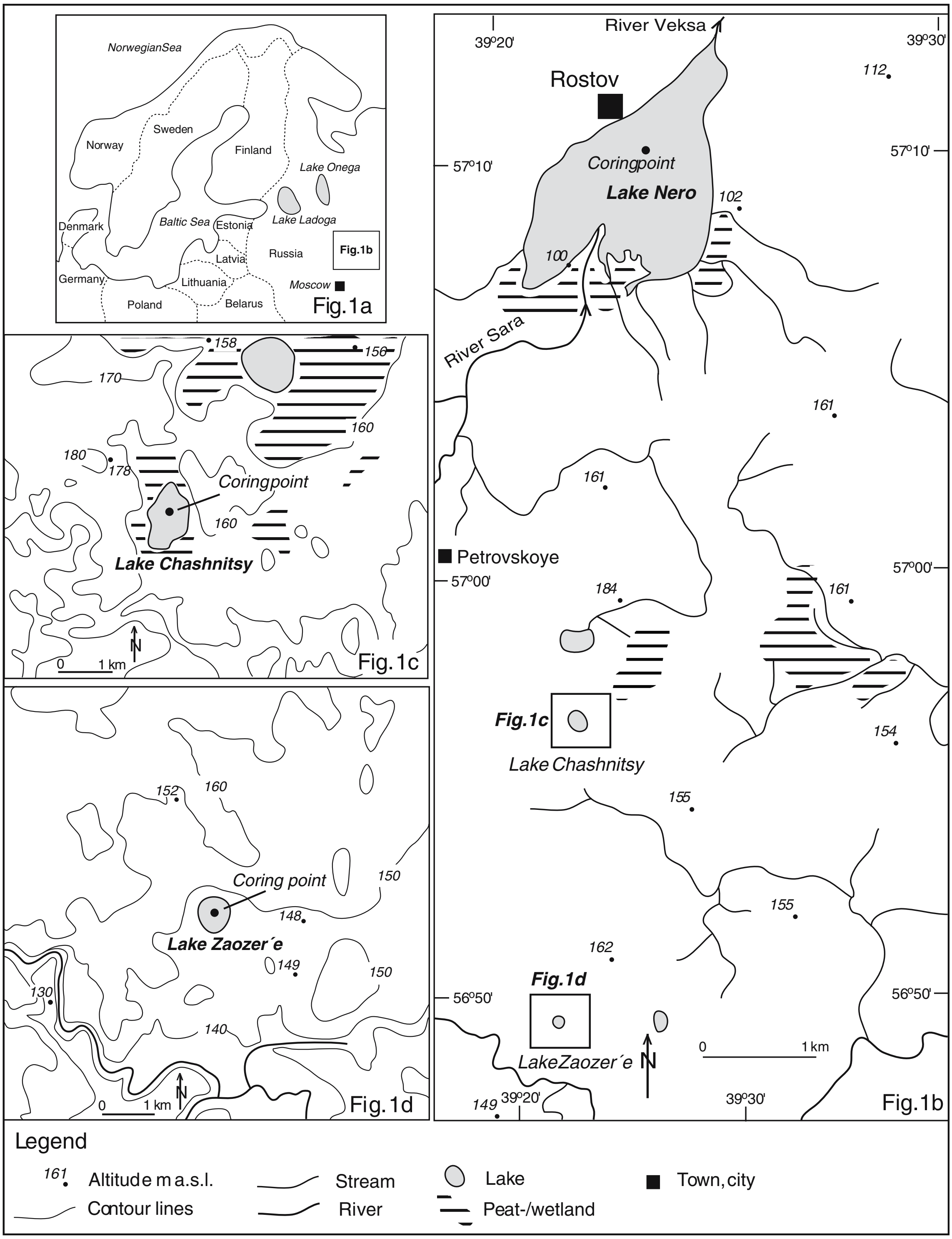

Figure 1. Location of the study area north of Moscow (a) and of the three investigated lakes in the Rostov region (b). Detailed maps of lakes Chashnitsy and Zaozer'e are shown in (c) and (d).

where low lake levels and a dry period prevailed 10,000 to 5000 cal yr BP and was followed by increasing lake levels around 5000 cal yr BP (Väliranta et al. 2005). The mid-Holocene increase in humidity is also seen on the Kola Peninsula (Solovieva and Jones 2002) and in northeastern European Russia (Sarmaja-Korjonen et al. 2004).

The present study is part of a larger project addressing late glacial and Holocene environmental changes along a NW–SE transect in westernmost European Russia (Wohlfarth et al. 2002, 2004). This paper focuses on the Rostov-Yaroslavl' area, which is located ca. 200 km north of Moscow (Figure 1) and forms the southeastern portion of our transect. We present a multi-disciplinary investigation of lake sediment sequences from lakes Nero, Chashnitsy and Zaozer'e covering the late glacial and Holocene time period and attempt to reconstruct the palaeoenvironmental development of this area using pollen, macrofossil, mineral magnetic and geochemical analyses. However, the existence of several sedimentary hiatuses makes it difficult to determine, for example, when trees first colonised the area. We also discuss how the observed changes relate to those reconstructed from other parts of western European Russia.

Study sites

Lake Nero (57°10′25″ N, 39°25′36″ E) is situated in the southern part of the Rostov Lowland and lakes Chashnitsy (56°56′29″ N, 39°22′55″ E) and Zaozer'e (56°49′41″ N, 39°21′20″ E) in the eastern part of the Borisoglebsk Highland of the East European Plain (Table 1, Figure 1a–d). During the penultimate glaciation, the area was covered by the Saalian/Moscowian ice sheet (Goretskii et al. 1982). The hummocky moraine landscape, which has been modified by post-glacial erosion, varies in elevation from 93 to 180 m a.s.l. Lake Nero was selected for the following reasons: (i) it had been the target of a number of earlier investigations and had been used to reconstruct relative changes in lake level (see compilation in Tarasov et al. 1996); however, the low resolution of the published records and the poor dating control (bulk sediment dates) led to large uncertainties in the reconstructions; (ii) pollen stratigraphy from such a large lake could provide a regional picture of past vegetation changes, which could be compared easily to the existing regional pollen stratigraphy for this part of Russia (e.g., Khotinsky 1977); (iii) the present investigation could form a pilot study for future investigations of the underlying Eemian and Weichselian (Valdaian) sediments.

Lake Nero is fed by numerous small rivers and drains northwards via River Veksa into River Volga (Figure 1b). Its water budget is mainly controlled by surface inflow (70%), while groundwater and direct precipitation account for 21 and 9%, respectively (Rokhmistrov 1970). Aquatic vegetation covers about two thirds of the lake, with a *Typha*-*Phragmites*-*Carex* zone along the shore and a sub-merged aquatic zone including *Myriophyllum* and *Potamogeton* to ca. 1.2 m depth (Bogachev et al. 1959). Floating bogs and peatlands occur in the southern part of the lake. The Quaternary stratigraphy of the lake basin has been

Table 1. Details on the three studied lakes according to Potashev (1959), Atlas Yaroslavskoi oblasti (1964), Gunova (1972a) and Aleshinskaya et al. (1986, 1987).

	Lake Nero	Lake Chashnitsy	Lake Zaozer'e
Coordinates	57°10′25″ N 39°25′36″ E	56°56′29″ N 39°22′55″ E	56°49′41″ N 39°21′20″ E
Altitude	93 m a.s.l.	163 m a.s.l.	147 m a.s.l.
Surface area	51.7 km^2	0.7 km^2	0.3 km^2
Catchment area	1190 km^2	–	–
Maximum water depth	4 m	> 6 m	18 m
Bedrock	Jurassic and Triassic clays, glauconitic sands and marls		
Quaternary deposits	>130 m of glacial, fluvial and lacustrine sediments		
Natural vegetation	Cool, mixed forests with dominance of *Picea abies*, *Betula* sect. *Albae*, *Alnus* spp., *Pinus sylvestris*, *Populus tremula*; presence of *Quercus robur*, *Tilia cordata*, *Ulmus* spp.		
Mean temperatures	−11 °C (January), 18 °C (July)		
Annual precipitation	520–600 mm		

Table 2. Lithostratigraphic description of the sediment sequences in lakes Nero, Chashnitsy and Zaozer'e.

Sediment depth below water column (m)	Unit	Lithology
A. Lake Nero		
1.525–1.74	1	Light-brown to grey algae gyttja, loose, gLB
1.74–1.90	2	Brown grey algae gyttja with light brown to grey spots, loose, gLB
1.90–3.32/3.35	3	Dark brown algae gyttja, slightly lighter coloured between 265–332/335 cm), homogenous and jelly, vsLB (wavy, erosive)
3.32/3.35–4.15	4	Brown-beige clayey silty calcareous algae gyttja, vgLB
4.15–4.68	5	Brown-beige clayey silty calcareous algae gyttja, vgLB
4.68–5.39	6	Grey clayey silty calcareous algae gyttja, rare shell fragments, gLB
5.39–5.455	7	Grey, silty gyttja, slightly sandy, abundant shells and shell fragments, coarse organic material, sLB
5.455–5.47	8	Blueish silt (fresh colour: white) with vivianit and large shells, sLB
5.47–5.56	9	Dark brown to black peaty gyttja/gyttja peat with roots, grainy structure, gLB
5.56–5.59	10	Brown-grey clayey silty gyttja with dark brown organic horizons (1 mm thick), gLB
5.59–5.62	11	Dark brown silty gyttja, gLB
5.62–5.73	12	Brown-grey sandy silt or sandy silty gyttja, between 566.5–567 cm: 2–3 mm thick dark organic horizons, gLB
5.73–5.79	13	Alternating layers of light brown silty sand and brown-grey sandy silt with organic remains, LB
5.79–5.82	14	Brown-grey sandy silt with organic material, gLB
5.82–5.88	15	Light brown slightly silty sand (coarser than below) with some organic material, gLB
5.88–5.90	16	Brown-grey silt with fine sand and organic material, gLB
5.90–5.94	17	Grey silty fine sand, gLB
5.94–6.31	18	Grey homogenous clay, very viscous
B. Lake Chashnitsy		
3.26–5.765	1	Dark brown Phragmites-rich coarse detritus gyttja, loose, gLB
5.765–5.985	2	Dark brown algae-rich gyttja, loose, gLB
5.985–6.415	3	Dark brown Phragmites-rich gyttja, loose, sLB
6.415–6.515	4	Dark brown algae-rich gyttja, loose, gLB
6.515–6.84	5	Brown Phragmites-rich coarse detritus gyttja, loose, gLB
6.84–6.875	6	Dark brown compact gyttja/?peat, gLB
6.875–6.98	7	Dark brown coarse detritus gyttja or peaty gyttja, compact, sLB
6.98–6.995	8	Dark brown gyttja, compact, sLB
6.995–7.015	9	Dark brown coarse detritus gyttja or peaty gyttja, compact, sLB
7.015–7.10	10	Dark brown silty clayey fine detritus gyttja, compact, gLB

Table 2. Continued

Sediment depth below water column (m)	Unit	Lithology
7.10–7.16	11	Dark brown silty sandy clayey gyttja, compact, vsLB
7.16–7.23	12	Brown-grey sandy silty clay with plant material, gLB
7.23–7.30	13	Greyish-yellowish silty sandy clay, gravel (~3 mm) and pebbles, gLB
7.30–7.40	14	Greyish-brown sandy silty clay, some plant material, sLB
7.40–8.05	15	Blueish-grey slightly silty clay with plant material, oxidises quickly
C. Lake Zaozer'e		
3.07–4.20	1	Brown clayey algae gyttja, gLB
4.20–4.335	2	Dark brown clayey gyttja with abundant moss layers, gLB
4.335–4.535	3	Dark brown clayey detritus gyttja, sLB
4.535–4.98	4	Blackish-brown drift gyttja/gyttja peat, wood remains between 470–473 cm, gLB
4.98–5.92	5	Brown sedge peat, gLB
5.92–5.935	6	Brown detritus gyttja, sLB
5.935–5.94	7	Black algae gyttja, sLB
5.94–5.965	8	Brown detritus gyttja, sLB
5.965–5.975	9	Blackish-brown to black organic gyttja, sLB
5.975–5.99	10	Light brown detritus gyttja, sLB
5.99–6.005	11	Blackish brown to black sandy gyttja, sLB
6.005–6.13	12	Dark brown detritus gyttja, vsLB
6.13–6.195	13	Brownish-grey silty clayey gyttja with dark brown organic-rich inclusions, vsLB (mixture of minerogenic and organic material)
6.195–6.24	14	Brownish-grey silty gyttja clay, gLB
6.24–6.265	15	Brownish-grey, slightly organic sandy silt with gravel (< 7 cm), vsLB
6.265–6.72	16	Grey to yellowish-grey sandy clayey silt or sandy silty clay, massive and compact with some organic remains; uppermost 3 cm rich in sand; stone (8 cm length) at 625–633

LB – lower boundary; g – gradual; s – sharp; vs – very sharp.

intensively studied (Gunova 1972a, b; Aleshinskaya 1973, 1974; Aleshinskaya and Gunova 1975, 1976; Aleshinskaya et al. 1986, 1987). Thermoluminesence dates on a 129.5-m-long sediment core drilled on a shoreline terrace 200 m north of the modern lake suggest that laminated clay sedimentation started in a deep proglacial lake sometime before 152,000 ± 16,000 yr BP, after the final retreat of the ice sheet (Gunova 1972a, b). Several sediment cores recovered from the centre of the present lake were analysed for pollen and diatoms, and according to the regional pollen stratigraphy, covered the late glacial and Holocene (Khotinsky 1977); however, the few ^{14}C measurements on bulk material only provide chronological control for the mid-Holocene (Tarasov et al. 1996).

Materials and methods

Coring was performed from ice in March 1999 with a Russian corer (7.5 cm ∅, 1 m length). Unfortunately sediments could not be taken in the centre of lakes Chashnitsy and Zaozer'e because the water depth there exceeded the length of the coring rods. The sediment cores were wrapped in plastic and placed in PVC tubes for transport to the laboratory, where they were stored at 4 °C.

Prior to sub-sampling, the core surfaces were carefully cleaned and the lithostratigraphy was described in detail (Table 2). Mineral magnetic measurements were performed on contiguous sub-samples to reveal changes in sediment composition and minerogenic run-off and included saturation isothermal remnant magnetisation (SIRM) and mineral magnetic susceptibility (χ). SIRM was measured with a Molspin 'Minispin' magnetometer using a DC field of 1T. Susceptibility measurements were made with a Digital Voltmeter Kopparbridge KLY-2. After the measurements were made, the samples were dried at 45 °C to calculate mass specific S.I. units.

For assessing the organic matter content of the sediments, the samples used for mineral magnetic measurements were dried at 105 °C overnight, crushed to powder and analysed in a Carlo Erba instrument (NCS 2500), where total carbon (TC) and total nitrogen (TN) were simultaneously determined. For the determination of total sulphur (TS), between 100 and 300 mg of sediment were reacted to extract sulphur. Sulphur in sediments exists mainly in the form of sulphide (mainly pyrite), sulphate, or organic sulphur in the residue. A strong oxidising agent is required to ensure that all sulphur is converted to the sulphate form. We used 14 M HNO_3 and Br_2 (Krouse and Tabatabai 1986), capable also of converting native sulphur to sulphate. The dissolved sulphate obtained from the extraction was converted to $BaSO_4$ by adding $BaCl_2$. The $BaSO_4$ samples were dried and analysed gravimetrically. C/N atomic ratios, which are used to discriminate between terrestrial and aquatic organic matter sources, were obtained by multiplying C/N mass ratios by 1.167 (Meyers and Teranes 2001).

Sub-samples for macrofossil remains were taken contiguously, while avoiding lithological boundaries and spaced between 3 and 5 cm. Samples were sieved under running water (mesh size 0.125 mm), and remains were identified using a dissecting microscope. A total of 20 AMS ^{14}C measurements were performed on wood, bark, charcoal and leaf fragments (Table 3). The selected plant material was dried immediately at 105 °C after sieving and identification. Sample pre-treatment followed the standard procedures at the Ångström Laboratory, Uppsala University and the Radiocarbon Dating Laboratory, Lund University. Calibration of the ^{14}C measurements was performed with OxCal 3.5 (Bronk Ramsey 2000).

Sub-samples for pollen (1–2 cm^3) were prepared by the methods outlined in Berglund and Ralska-Jasiewiczowa (1986) and *Lycopodium* tablets with a known number of spores were added to each sample to estimate pollen concentration. Pollen keys and illustrations in Moore et al. (1991) and Reille (1992), as well as pollen reference collections at the Department of Geology, Lund University and the Department of Geography, Moscow State University, were used for identification. The pollen percentage diagram was sub-divided into local pollen assemblage zones (LPAZ) using sum-of-squares cluster analysis (Grimm 1987).

Results and interpretation

Lake Nero (57°10′25″ N; 39°25′36″ E)

Lacustrine and telmatic proxy-records

The lowermost clay, silt and sand layers (units 18–15; 6.31–5.82 m; Table 2A), which were

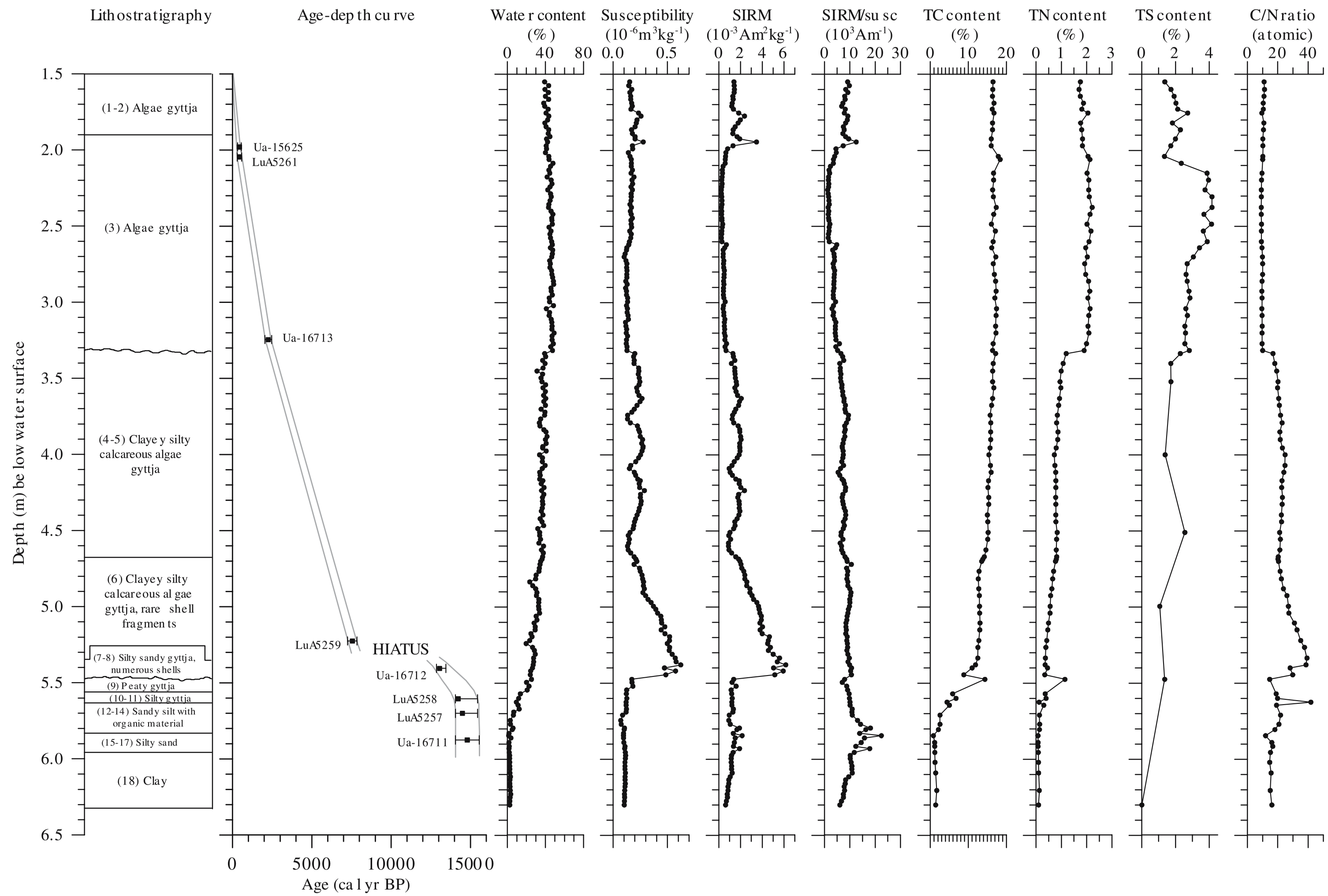

Figure 2. Lithology, age-depth curve and geochemistry for Lake Nero. Wavy line – sharp or erosive boundary; see Table 1A for detailed lithostratigraphic descriptions and Table 3 for details on the AMS ^{14}C measurements.

Table 3. Radiocarbon dates and calibrated ages for the three studied lake-sediment sequences.

Depth (cm)	Material	Lab No.[a]	Radiocarbon age (^{14}C yr BP $\pm 1\sigma$)	Calibrated age (cal yr BP $\pm 2\sigma$)
Lake Nero				
202–194	Bark	Ua-15265	490 ± 70	430 ± 120
207–202	Wood fragments	LuA5261	425 ± 90	430 ± 140
327–322	Wood fragments	Ua-16713	2265 ± 75	2255 ± 215
525–520	Wood fragments	LuA5259	6700 ± 155	7555 + 295/−305
542–539	Wood fragments	Ua-16712	11,100 ± 130	13,050 + 400/−200
562–559	Bark, leaf fragments	LuA5258	12,290 ± 100	14,225 + 1225/−175
573–567	Wood fragments	LuA5257	12,440 ± 110	14,500 + 950/−450
590–585	Wood fragments	Ua-16711	12,530 ± 150	14,800 ± 750
Lake Chashnitsy				
406–407	Wood fragments	Ua-15269	3805 ± 70	4190 + 110/−210
556.5–551.5	Wood, bark fragments	Ua-15268	4510 ± 75	5175 + 175/−325
606–600	Wood, bark fragments	Ua-16710	8225 ± 80	9190 + 210/−180
660–655	Wood fragments	Ua-16709	9185 ± 90	10,330 + 250/−140
687.5–684	Wood fragments	Ua-15267	9435 ± 90	10,650 + 450/−250
713–710	Wood fragments	Ua-16708	9725 ± 125	11,010 + 540/−410
740–735	Wood fragments	Ua-15266	9760 ± 90	11,170 + 180/−420
Lake Zaozer'e				
453.5–449	*Betula* sp., bark	Ua-16716	4585 ± 65	5225 + 275/−275
491–486	Wood, bark fragments	LuA5263	8265 ± 95	9275 + 205/−255
545–541	Wood, bark fragments, charcoal	LuA5262	8730 ± 90	9690 + 460/−190
590–585	Wood fragments	Ua-16715	9380 ± 110	10,575 + 525/−375
617.5–613	Charcoal	Ua-16714	10,085 ± 95	11,625 + 725/−475

[a]Ua – Uppsala, Ångström Laboratory; LuA – Lund Radiocarbon Dating Laboratory.

deposited before ~14,800 cal yr BP, have low χ and SIRM values, and SIRM/χ ratios of 10–20 (Figure 2). The latter likely indicate presence of bacterial magnetite and reduced bottom-water conditions. TC and TN contents are low and the C/N ratio is about 15 (Figure 2), suggesting low productivity and a mixed aquatic/terrestrial origin for the organic matter. This assumption is corroborated by scarce telmatic and limnic macrofossils of *Carex* sp., *Batrachium* sp., *Myriophyllum alterniflorum*, *Callitriche hermaphroditica*, *Potamogeton praelongus*, *Potamogeton filiformis*, *Valvata piscinalis*, and *Candona* spp. (Figure 3).

In the overlying sandy silt (units 14–12, 5.82–5.62 m; Table 2A) and silty gyttja (units 11–10, 5.62–5.56 m) TC increases and the C/N ratio attains 20 (Figure 2). Limnic macrofossils are more diverse and include, in addition to the species present in the underlying sediments, *Hippuris vulgaris*, *Potamogeton obtusifolius*, *Pisicola geometra*, *Erpobdella* sp. and *Pisidium* sp. (Figure 3). The shift towards higher lake organic production around 14,800 cal yr BP may have been favoured by increased input of terrestrial organic matter into the lake, as indicated by higher C/N ratios. Several of the limnic plant and animal remains and more frequent finds of *Carex* sp., as well as high pollen percentages of aquatic taxa (e.g., *Myriophyllum* and *Equisetum*) and Cyperaceae in LPAZ Ne-8 (5.82–5.56 m) (Figure 4), suggest the development of a shallow, eutrophic lake between ca. 14,800 and 14,200 cal yr BP with an extensive telmatic vegetation zone.

The peaty gyttja/gyttja peat (unit 9, 5.56–5.46 m; Table 2A) has low mineral magnetic values and rapidly rising TC and TN contents (Figure 2). In the overlying silt and silty, sandy gyttja (unit 8–7; 5.46–5.39 m) χ and SIRM increase distinctly, TC attains 12% and the C/N ratio is 30–40 (Figure 2). Limnic macrofossils include *Hippuris vulgaris*, *Potamogeton filiformis* and *Pisicola geometra*, and abundant *Valvata piscinalis*, *Pisidium* sp. and *Candona* spp. (Figure 3). Sediments in units 9–7 suggest a gradual lowering of the water level between 14,200 and 13,500 cal yr BP, accumulation of predominantly terrestrial organic matter and the subsequent development of a shoreline close to the coring site, where *Valvata piscinalis* shells accumulated. The sharp boundary between units 9 and 8/7, corresponding to ~13,500 cal yr BP, and the abrupt changes seen in several parameters suggest a hiatus, which is also indicated by the

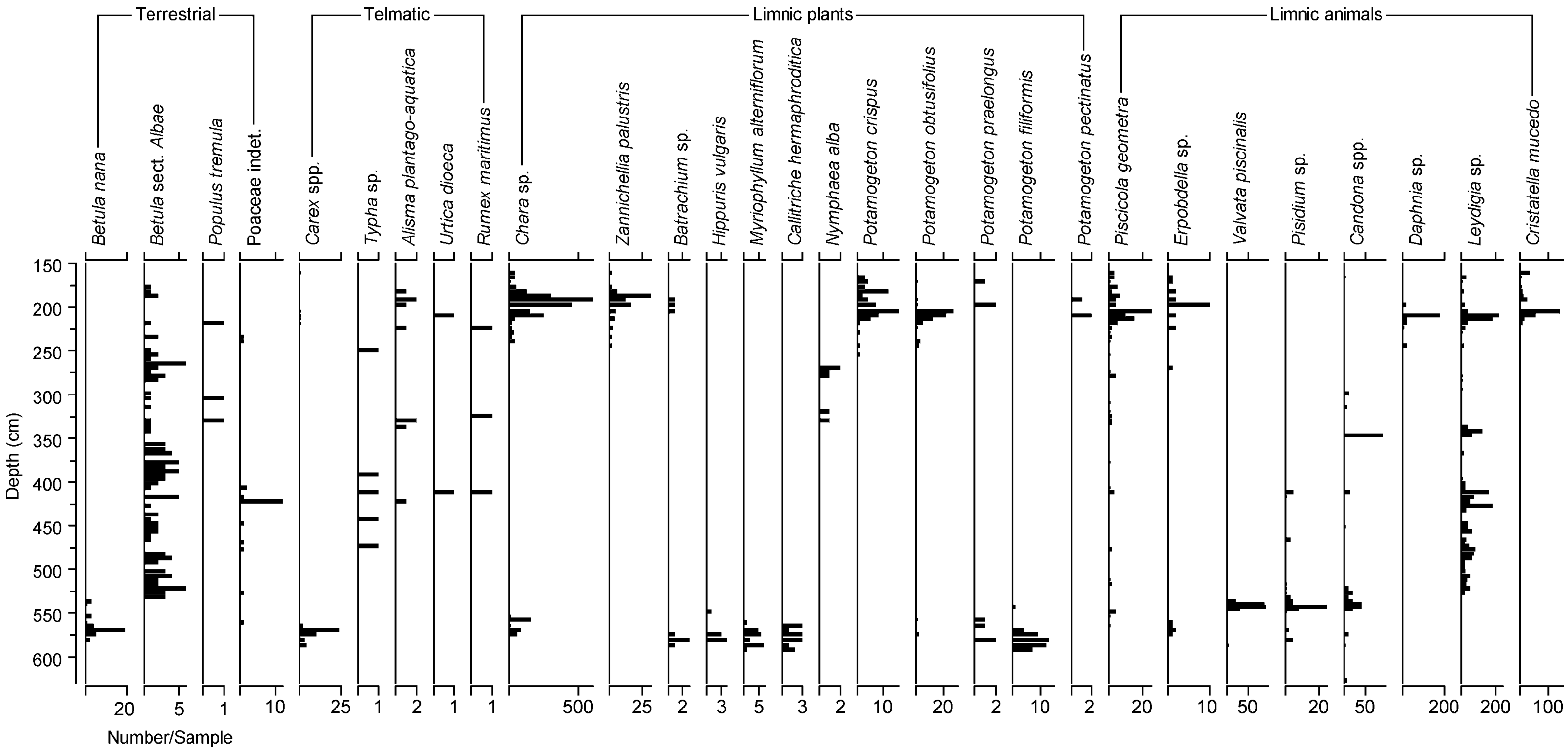

Figure 3. Macrofossil concentration diagram for Lake Nero. Note changes in scale. Units: number of macrofossils/sample.

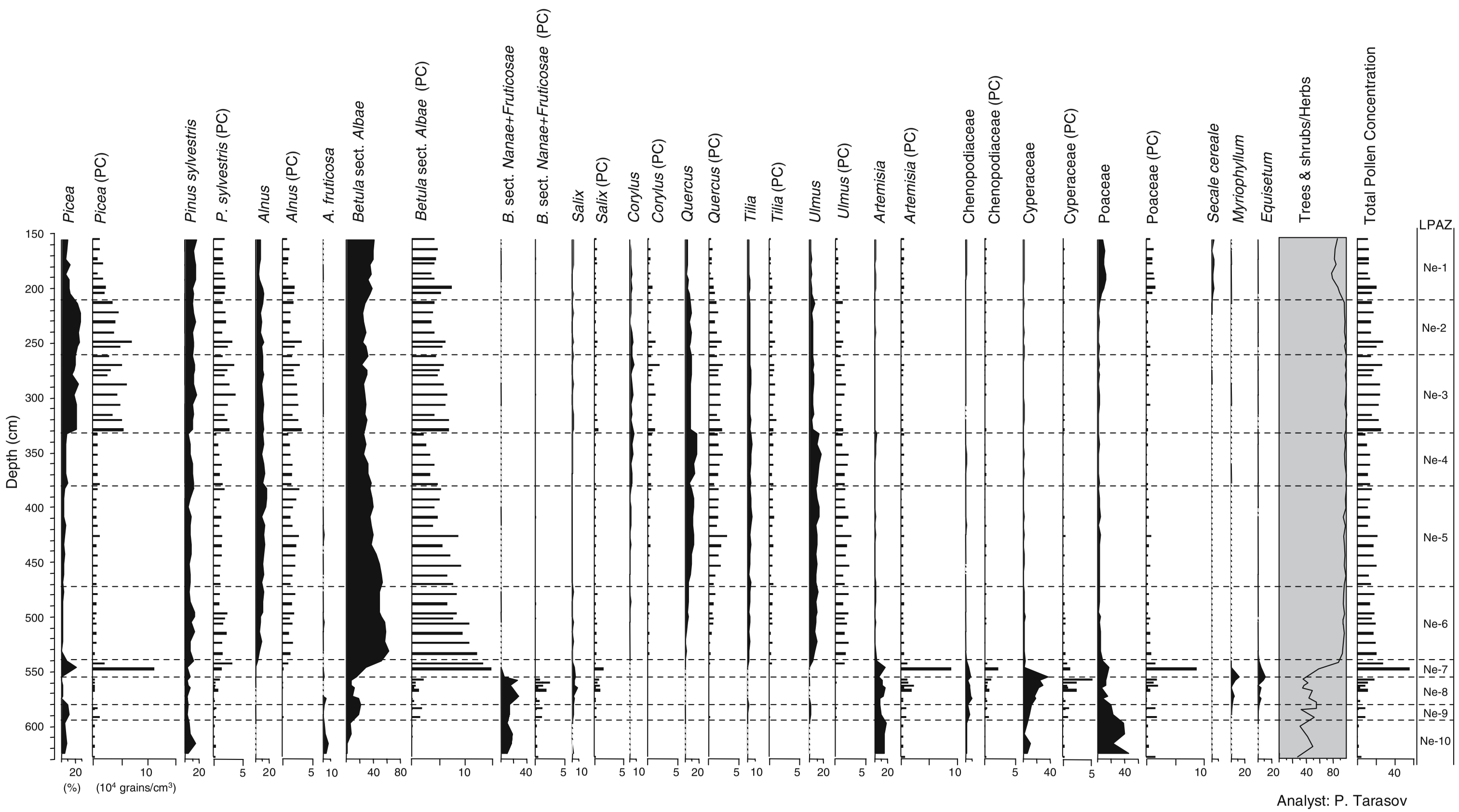

Figure 4. Simplified pollen percentage and concentration (PC) diagram for Lake Nero. Note change in scale for Total Pollen Concentration.

pollen stratigraphy (see below). However, based on the two age estimates of 13,050 cal yr BP (Ua-16712) at 5.42–5.39 m (unit 7) and of 7550 cal yr BP (LuA5259) at 5.25–5.20 m (unit 6) (Table 3, Figure 2), this hiatus could also be located at the boundary between units 7 and 6, where a marked shift towards renewed limnic conditions can be observed. On the other hand, the wood fragments of sample Ua-16712 may be reworked, since they derive from the silty, sandy gyttja containing numerous *Valvata piscinalis* shells (unit 7). The uncertainty connected to sample Ua-16712 allows for two possible scenarios: (1) The subsequent rise in lake level, which can be inferred from the limnic fauna in the overlying unit 6, eroded the underlying layers, which could thus be missing between units 7 and 6; and (2) the lake gradually dried out (unit 9) and a shoreline formed close to the coring site (unit 8/7) between 13,500 and 13,000 cal yr BP, where reworking of older minerogenic and organic matter occurred. Given that the most distinct changes in several parameters (i.e., macrofossil remains, pollen, mineral magnetics, TC and TN content) occur at the boundary between units 9 and 8/7, we favour scenario 2 as a working hypothesis.

χ and SIRM values gradually decline in the silty calcareous algae gyttja of unit 6 (5.39–4.68 m; Table 2A), but have fluctuating values in the clayey, silty calcareous algae gyttja of units 5–4 (4.68–3.32/3.35 m) (Figure 2). TC and TN increase slightly at the transition between units 6 and 5, coincident with a decrease in the C/N ratio to 20, but have stable values thereafter. The TS content increases, fluctuating between 1.5 and 2.5%. However, the low temporal resolution of the measurements in units 6–4 does not allow any further conclusions regarding the comparably high TS content to be drawn. The scarce telmatic and limnic macrofossils are dominated by the cladocera *Leydigia* sp. (Figure 3), indicating a shallow lake phase. An approximate age estimate for units 6–4 is 8200–2500 cal yr BP (Figure 2).

The transition from unit 4 to the algae gyttjas of units 3–1 (3.32/3.35–1.525 m) is erosive Table 2A) and marked by an abrupt decline in χ and SIRM values and in the SIRM/χ ratio. TC has stable values over this transition, while the TN content increases sharply and the C/N ratio decreases from 20 to 10. TS values rise shortly before the transition to unit 3 and are around 2.5–3% (Figure 2). Taken together these changes may be evidence for the presence of another hiatus, which may also be inferred from the pollen stratigraphy (Figure 4). Mineral magnetic parameters remain low until 2.00 m, when SIRM and SIRM/χ increase. A marked shift is observed in the TS curve, which displays an increase from 2.5% at 2.70 m to 4% between 2.60 and 2.10 m, and a gradual decline to values of 1% thereafter (Figure 2). The TS values in units 3–1 are decoupled from the TC values i.e., there is no linear correlation between TC and TS, which may indicate an external source for the available sulphur. Anoxic conditions must have prevailed in the lake between $\sim$1500 and $\sim$500 cal yr BP, given the high TS content, the concomitant dissolution of magnetic minerals and the scarcity of limnic and telmatic macrofossils. *Chara* sp., *Zannichellia palustris*, *Potamogeton crispus*, *Potamogeton obtusifolius*, *Pisicola geometra*, *Daphnia*, *Leydigia* sp. and *Cristatella mucedo*, which are especially abundant between 2.00 and 1.50 m (Figure 3), indicate eutrophic conditions during the past ca. 500 years.

Terrestrial proxy records

The pollen spectra of LPAZ Ne-10 (6.31–5.95 m; $\approx$ unit 18) are characterised by high percentages of non-arboreal pollen (NAP) (i.e., Poaceae, *Artemisia*, Cyperaceae, and Ranunculaceae) and shrub pollen (*Betula* sect. *Nanae*/*B.* sect. *Fruticosae*, *Alnus* subsp. *fruticosa)* (Figure 4). Arboreal pollen (AP) taxa are mainly represented by *Picea* and *Pinus*. However, pollen concentration values are low for all taxa before 14,800 cal yr BP (Figure 4), which could be explained by the high sedimentation rate of the minerogenic sediments and/or by the fact that much of the pollen is long-distance transported. Concentration values for *Picea*, *Pinus*, *Betula* sect. *Albae*, and *Betula nana*, as well as for grass and herb pollen, increase in LPAZ Ne-9 (5.95–5.82 m; units 17–15), where sparse pollen grains of *Ulmus*, *Tilia* and *Quercus* were observed also. However, only the local presence of *Betula nana* can be corroborated by macrofossils in unit 15 (Figure 3). Sediment units 17–15 are composed of silty, sandy sediments, thus the occurrence of *Ulmus*, *Tilia* and *Quercus* pollen and the rise in pollen concentrations for *Picea*, *Pinus*, and *Betula* sect. *Albae* in LBP Ne-9 likely indicate reworking of pollen from older sediments, rather than the actual presence of these tree species in the lake's catchment.

Generally high pollen percentages and concentrations for *Betula nana*, *Salix*, *Artemisia*, Chenopodiaceae, Cyperaceae, Poaceae and aquatic taxa (e.g., *Myriophyllum* and *Equisetum*), especially in the upper part of LPAZ Ne-8 (5.82–5.56 m ≈ units 14–10) (Figure 4) suggest a gradual shallowing of the lake, possibly starting around 14,800–14,500 cal yr BP.

LPAZ Ne-7 (5.56–5.39 m), which extends over sediment units 9–7, starts with a distinct increase in pollen and concentration values for *Betula* sect. *Albae*, *Picea*, *Artemisia*, Chenopodiaceae and Poaceae at 5.50 m (≈ unit 9) and is followed by increasing values for *Pinus*, *Tilia* and *Ulmus* pollen, and an abrupt decrease in NAP concentrations at 5.40 m (≈ units 8–7). This distinct change in pollen assemblages approximately coincides with the boundary between units 9 and 8/7, where lithostratigraphic parameters and the age-depth curve (Figure 2) indicate a hiatus of ca. 5000 years i.e., between ~13,000 and ~8200 cal yr BP, due to a marked lake-level lowering. Also, plant macrofossils remains of *Betula nana* disappear completely at 5.40 m and are replaced by *Betula* sect. *Albae* (Figure 3), which shows that tree birch grew locally. The high concentration values for *Betula* sect. *Albae* at 5.50 m (≈ unit 9) suggest that these colonised the area and may have grown on the former lake floor around 14,000 cal yr BP. However, chronological and lithological problems in the form of a hiatus make it difficult to determine exactly the first occurrence of trees in the lake's catchment. The large increase in AP percentages and the appearance of pollen of broad-leaved taxa at 5.40 m (≈ units 8–7) likely occurred after the hiatus i.e., around 8200 cal yr BP.

The pollen spectra in LPAZ Ne-6 (5.39–4.73 m ≈ unit 6) are characterised by high AP percentages and concentrations, mainly of *Betula* sect. *Albae*, but also including *Pinus*, *Alnus*, *Ulmus*, *Tilia* and *Quercus*, whereas values for *Picea* and NAP are low (Figure 4). The occurrence of *Tilia* and *Quercus* pollen reflect increased summer temperatures from ~8200 cal yr BP onwards. LPAZ Ne-5 (4.73–3.80 m ≈ unit 5) and LPAZ Ne-4 (3.80–3.32 m ≈ unit 4) have similar pollen assemblages as LPAZ Ne-6, except that *Corylus* and *Quercus* percentage and concentration values increase, while those for *Betula* sect. *Albae* decrease. This suggests that the composition of the mixed broad-leaved – coniferous forests became more diverse between ~6100 and ~2500 cal yr BP.

At the beginning of LPAZ Ne-3, or about 2500 cal yr BP (3.32–2.60 m ≈ lower part of unit 3), pollen percentages and concentrations for *Pinus*, *Alnus*, *Betula* sect. *Albae* and particularly for *Picea* increase distinctly (Figure 4) and macrofossils of *Populus tremula* appear (Figure 3). In accordance with the lithological evidence (see above), the distinct change in pollen assemblages very likely indicates a hiatus at 3.32/3.35 m of unknown duration. In LPAZ Ne-2 (2.60–2.10 m ≈ upper part of unit 3, or about 500 cal yr BP) the main pollen taxa have rather stable percentages and concentrations.

The uppermost pollen zone LPAZ Ne-1 (2.10–1.525 m ≈ top of unit 3, units 2–1) is characterised by a decrease in pollen percentages and concentrations for *Picea*, *Quercus*, *Tilia* and *Ulmus* and an increase in Poaceae pollen percentages and concentrations. Pollen of *Secale cereale*, *Rumex acetosa/acetosella*, Polygonaceae, *Urtica*, *Fagopyrum* and *Linum* demonstrate agricultural activity close to the site over the last 500 years. However, since the settlement of Rostov (Figure 1b) is known from the Russian chronicles since 862 A.D. and was a very important Russian city by the 10th century, the absence of taxa indicating agricultural activity before ca. 500 cal yr BP may reflect another sedimentary hiatus.

Lake Chashnitsy (56°56′29″ N; 39°22′55″ E)

The minerogenic bottom sediments (8.05–7.16 m, units 15–12; Table 2B) have low mineral magnetic and low TC and TN values, while TS increases slightly at 7.50 m (Figure 5). The C/N ratio is initially around 10, but rises to 20 in units 14–12, indicating that the organic matter source changed from predominantly aquatic to terrestrial. The minerogenic sediments and their low TC content likely indicate rapid sedimentation and the continuous occurrence of radicells or rootlets (Figure 6) points to a rather shallow basin. Although speculative, we assume that melting of permafrost led to the formation of the lake basin around 11,200–11,000 cal yr BP.

The sandy clayey gyttja of unit 11 (7.16–7.10 m) has a sharp, erosive contact to unit 12 and is overlain by fine detritus gyttja (7.10–7.015 m, unit 10) (Table 2B). TC, TN and TS rapidly attain 30,

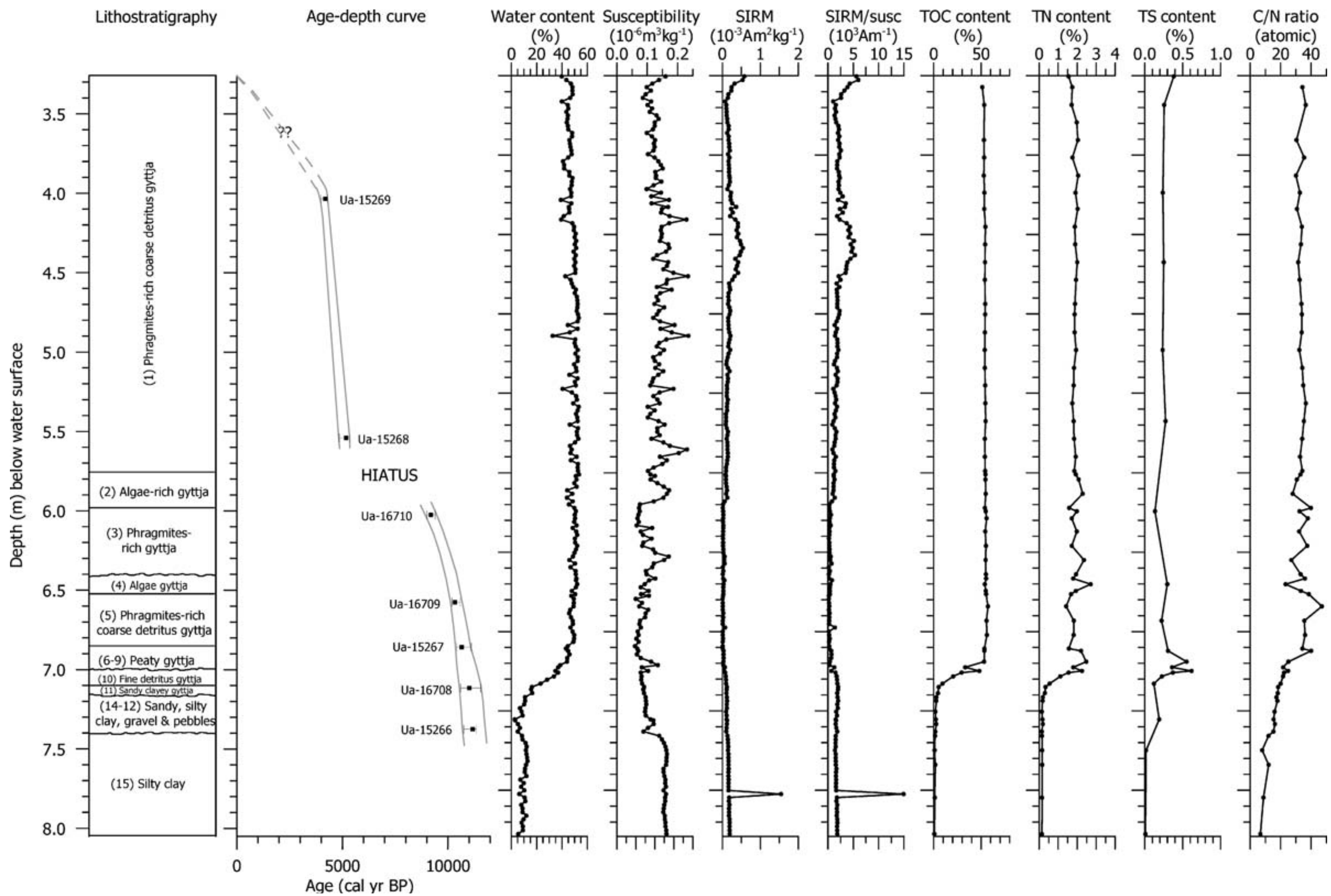

Figure 5. Lithology, age-depth curve and geochemical parameters for Lake Chashnitsy. Wavy line – sharp or erosive boundary; see Table 1C for detailed lithostratigraphic descriptions and Table 3 for details on the AMS ^{14}C measurements.

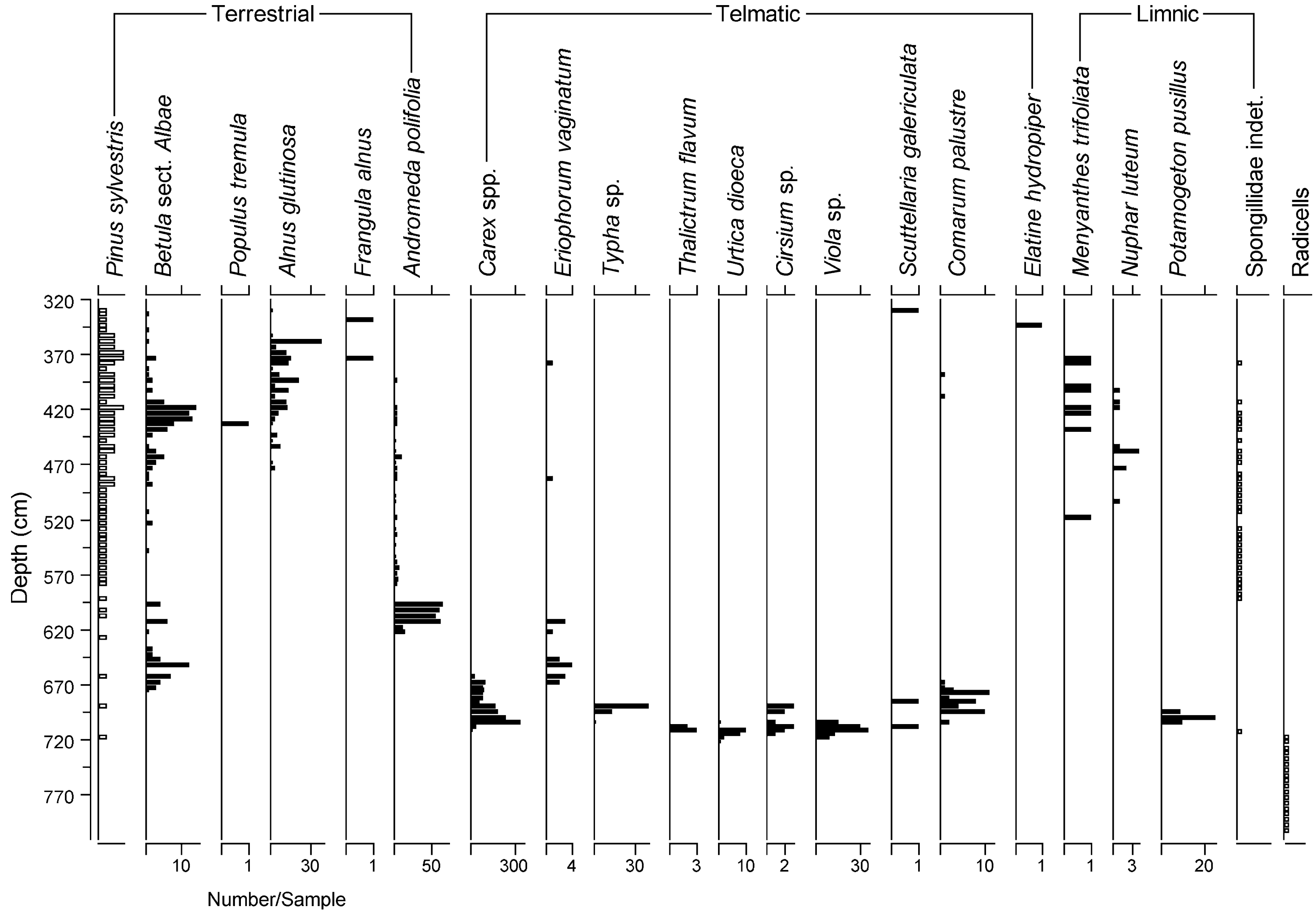

Figure 6. Macrofossil concentration diagram for Lake Chashnitsy. Note changes in scale. Hollow bars indicate relative abundance (short, rare; intermediate, common; long, abundant). Units: number of macrofossils/sample.

1.5 and 0.7%, respectively in unit 10 (Figure 5). Telmatic macrofossils become more frequent and include *Carex* sp., *Typha* sp., *Thalictrum flavum*, *Urtica dioica*, *Cirsium* sp., *Viola* sp. and *Comarum palustre*, while limnic macrofossils are only comprised of *Potamogeton pusillus* (Figure 6). The sharp, erosive boundary between units 12 and 11 might suggest a short hiatus, although mineral magnetic parameters, TC, TN and TS values and the age-depth curve give no indications for abrupt changes. The increasing organic content of the sediments shows that minerogenic input into the lake gradually decreased. Higher input of telmatic and terrestrial organic material, as indicated by a C/N ratio of around 20 and rapid burial of organic matter, could have led to anoxic conditions, which may also be inferred from the higher TS values in unit 10. The increase in TS values coincides with an increase in TC values and we therefore speculate that the majority of the sulphur in the sediments may be derived from a biogenic source. The dominance of telmatic macrofossils likely suggests shallow water at the coring site.

Units 9–6 (7.015–6.84 m; Table 2B) span the interval between ca. 10,800 and ca. 10,600 cal yr BP, and are composed of alternating layers of compact detritus gyttja/peaty gyttja and gyttja with sharp lower boundaries (Table 2B, Figure 5). The sharp contact between units 10 and 9, the slight increase in χ and the concomitant decrease in SIRM and SIRM/χ, as well as fluctuations in TC, TN and TS could indicate another short hiatus. Throughout units 9–6, mineral magnetic parameters remain low, TC attains 50%, TN ~2.5% and TS ~0.6%. A C/N ratio of 40 suggests that the sediments contain mainly terrestrial organic matter (Figure 5), which is also corroborated by the dominance of telmatic (*Carex* sp., *Typha* sp., *Cirsium* sp., *Comarum palustre)* over limnic (*Potamogeton pusillus*) plant remains (Figure 6). Higher TS values in units 9–6 could be explained by enhanced organic matter input into the lake, which could have led to the establishment of anoxic pore water regimes. Anoxic conditions may also be inferred from the low mineral magnetic values, which could indicate dissolution of mineral magnetic particles. The rapidly changing sediment lithology likely reflects fluctuations in water level and the compactness of the sediments may even point to occasional desiccation. The recorded telmatic plant macrofossils are today common along lake shores and on peat – and wetlands and suggest that the lake may have been very shallow and/or partly overgrown.

Units 5–1 (6.75–3.26 m) are made up of alternating layers of loose, *Phragmites*-rich (units 5, 3, 1) and algae-rich gyttjas (units 4, 2) (Table 2B) and encompass the time interval between ca. 10,600 cal yr BP and the present (Figure 5). χ values are low throughout, but a minor increase in SIRM and SIRM/χ values is observed at 5.95 m, between 4.50 and 4.00 m and in the uppermost 14 cm. TC is stable at ~50%, TN fluctuates around 2% and TS around 0.2–0.4%. A C/N ratio of 30–40 shows that the organic material is mainly derived from terrestrial matter sources (Figure 5). *Carex* sp. and *Comarum palustre* remains are sparse between 6.75 and 6.30 m, while *Eriophorum* is more frequent between 6.70 and 6.15 m and the terrestrial *Andromeda polifolia*, an evergreen dwarf shrub that grows on wet moss or peat, increases distinctly between 6.30 and 5.95 m (Figure 6). This, together with the high C/N ratio, is interpreted as a gradual overgrowing of the lake and an expansion of wetlands. At 5.95 m, at the transition to unit 2, *Andromeda polifolia* remains decrease dramatically and Spongillidae appear. These and the appearance of telmatic remains (*Eriophorum vaginatum*, *Comarum palustre*, *Elatine hydropiper)* and more frequent limnic (*Nuphar luteum*, *Menyanthes trifoliata*) remains from 5.20 m upwards testify to a return to open water conditions and therefore a rise in lake level around 5300 cal yr BP (Figure 6). As shown by the age-depth curve, a long hiatus must be present in the sequence somewhere between 6.00 and 5.50 m either at the transition between units 3 and 2, in unit 2 or at the transition between units 2 and 1 (Figure 5). The sediment lithology and the low χ values give no indications for a hiatus, but minor changes in SIRM and SIRM/χ and the decrease in the C/N ratio at 5.95 m may. At the same level, *Andromeda polifolia* macrofossils decrease dramatically and the site returned to open water conditions, as shown by the presence of Spongillidae. It is therefore likely that the hiatus is situated between units 3 and 2 and that it was caused by a decrease in water level. Sedimentation resumed after a water-level rise ca. 5300 cal yr BP.

Terrestrial plant macrofossils include the first finds of *Pinus sylvestris* at 7.18 m, of *Betula* sect. *Albae* at 6.80 m, of *Andromeda polifolia* at 6.30 m and of *Alnus glutinosa* at 5.30 m. *Populus tremula* and *Frangula alnus* remains are scarce (Figure 6). The vegetation surrounding the site may have initially been composed of *Pinus* > 10,600 cal yr BP and *Pinus-Betula* forests between 10,600 and ~9000 cal yr BP and also included *Populus tremula* and *Frangula alnus* shrubs or small trees after 5300 cal yr BP. Wetlands with *Andromeda polifolia* expanded over the site after 10,000 cal yr BP, but were likely restricted to the surrounding area after 5300 cal yr BP.

Lake Zaozer'e (56°49′41″ N; 39°21′20″ E)
Mineral magnetic values, TC, TN and TS content are low in the sandy clayey silt (6.72–6.24 m, units 15–16) and silty gyttja clay (6.24–6.195 m, unit 14) (Table 2C, Figure 7). The C/N ratio is slightly above 10 indicating that the organic material is largely of limnic origin. Only scarce *Phragmites australis* and radicells were observed among the macrofossils (Figure 8). Deposition of these sediments likely occurred before ~11,700 cal yr BP (Figure 7) in a shallow basin, possibly as a consequence of permafrost disintegration.

Sharp (erosive) lower boundaries, low mineral magnetic values and rising TC and TN content characterise the silty clayey gyttja (unit 13) and detritus gyttja (unit 12) between 6.195 and 6.005 m (Table 2C). *Carex* sp., *Phragmites australis* and radicells are present in small numbers. Although the lithology could indicate a hiatus between individual layers, stable mineral magnetic parameters and the gradually rising TC and TN content imply that these would have been of short duration only. Limnic macrofossil remains are absent, but a C/N ratio of 10–20 shows that the sediment organic matter is of mixed aquatic and terrestrial origin. The rise in TC and TN content ca. 11,700 cal yr BP is interpreted as a transition from a shallow lake low in productivity to a lake with higher organic production and telmatic vegetation.

The transition to the overlying sandy gyttja (6.005 m, unit 11) and gyttja of units 10–6 (6.00–5.92 m) occurred around 11,100 cal yr BP (Figure 7). The sharp boundary, the abrupt decrease in SIRM/χ, and the rapid rise in TC, TN, TS values and in the C/N ratio could indicate a hiatus (Table 2C, Figure 7). A C/N ratio of around 30 points to increasing input of terrestrial organic matter. This and the increase in *Phragmites australis* and radicell remains (Figure 8) suggest that the lake level in the already shallow lake decreased further between ca. 11,100 and 10,500 cal yr BP. Anoxic conditions may be inferred from higher TS and low mineral magnetic values. Rapid burial of organic matter could have led to anoxia, which in turn could have caused dilution of iron magnetite in the sediments. The initial increase in TS coincides with rising TC and TN values and may therefore be linked to a biogenic source.

In the sedge peat of unit 5 (5.92–4.98 m) mineral magnetic parameters are low, TC has stable values and the TS content decreases from maximum values of 0.7–0.2% (Figure 7). Although most of the TS in the lower part of unit 5 may be derived from biogenic sulphur sources, the decoupling of the TC and TS values also indicates contributions from an external sulphur source. The C/N ratio fluctuates at 30–40, but starts to rise again at 5.30 m and attains 80 at 5.00 m. Macrofossils include abundant *Carex* sp., *Phragmites australis* and radicells as well as scarce *Comarum palustre* and *Lycopus europaeus* (Figure 8). Together the different proxies show that the former shallow lake became overgrown and changed into peatland between ca. 10,500 and 9300 cal yr BP. Anoxic conditions prevailed as indicated by the initially high TS content and the dissolution of magnetic minerals.

χ and SIRM are initially low in the drift gyttja/gyttja peat of unit 4 (4.92–4.535 m), but increase slightly around 4.60 m, coincident with a marked increase in TN values and a decline in the C/N ratio to 20 (Figure 7). *Phragmites australis* and radicells are present with scarce remains up to 4.85 m, *Polytrichum* sp. throughout the whole unit and the first *Daphnia* sp. at 4.60 m (Figure 8). These changes suggest a rather abrupt shift in depositional environment from peatland to open water (Table 2C, Figure 7). Coincident with this shift, anoxic conditions diminished. As indicated by the age-depth curve (Figure 7), a major hiatus must be present in the sediments somewhere between 4.90 and 4.50 m and it seems likely that it is related to the distinct changes seen at 4.60–4.50 m. The presence of a hiatus makes it impossible to determine the age of the upper boundary of unit 4; however, assuming a constant sedimentation rate, it could tentatively be placed at around 9000 cal yr BP.

Units 3 (clayey detritus gyttja) and 2 (clayey gyttja) between 4.535 and 4.20 m have increasing SIRM values and a higher SIRM/χ ratio. TC and TN contents decrease, TS is stable and the C/N ratio is around 10 (Figure 7). Telmatic macrofossils disappear completely and the assemblage is

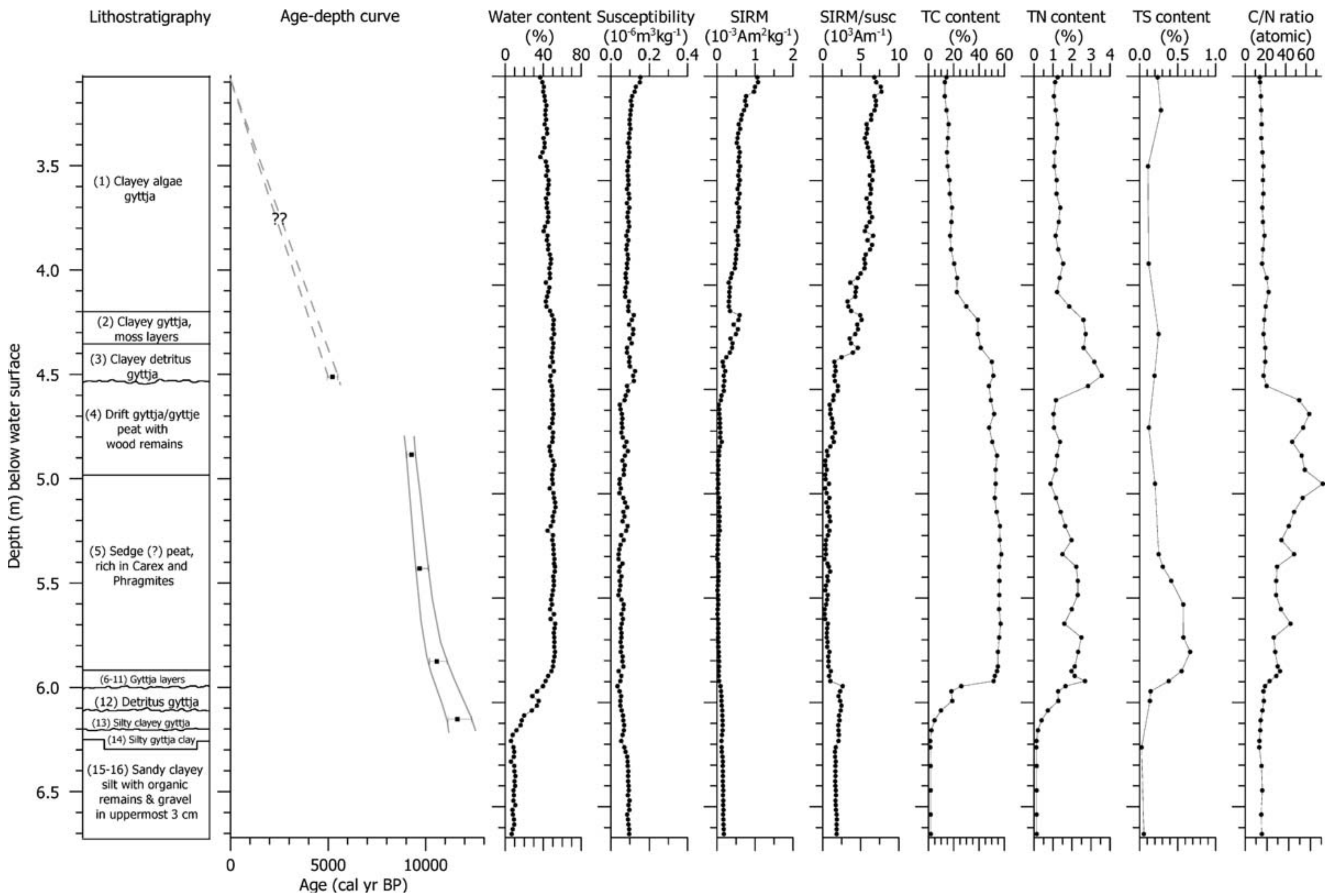

Figure 7. Lithology, age-depth curve and geochemical parameters for Lake Zaozer'e. Wavy line – sharp or erosive lower boundary; see Table 1B for a detailed lithostratigraphic description and Table 3 for details on the AMS ^{14}C measurements.

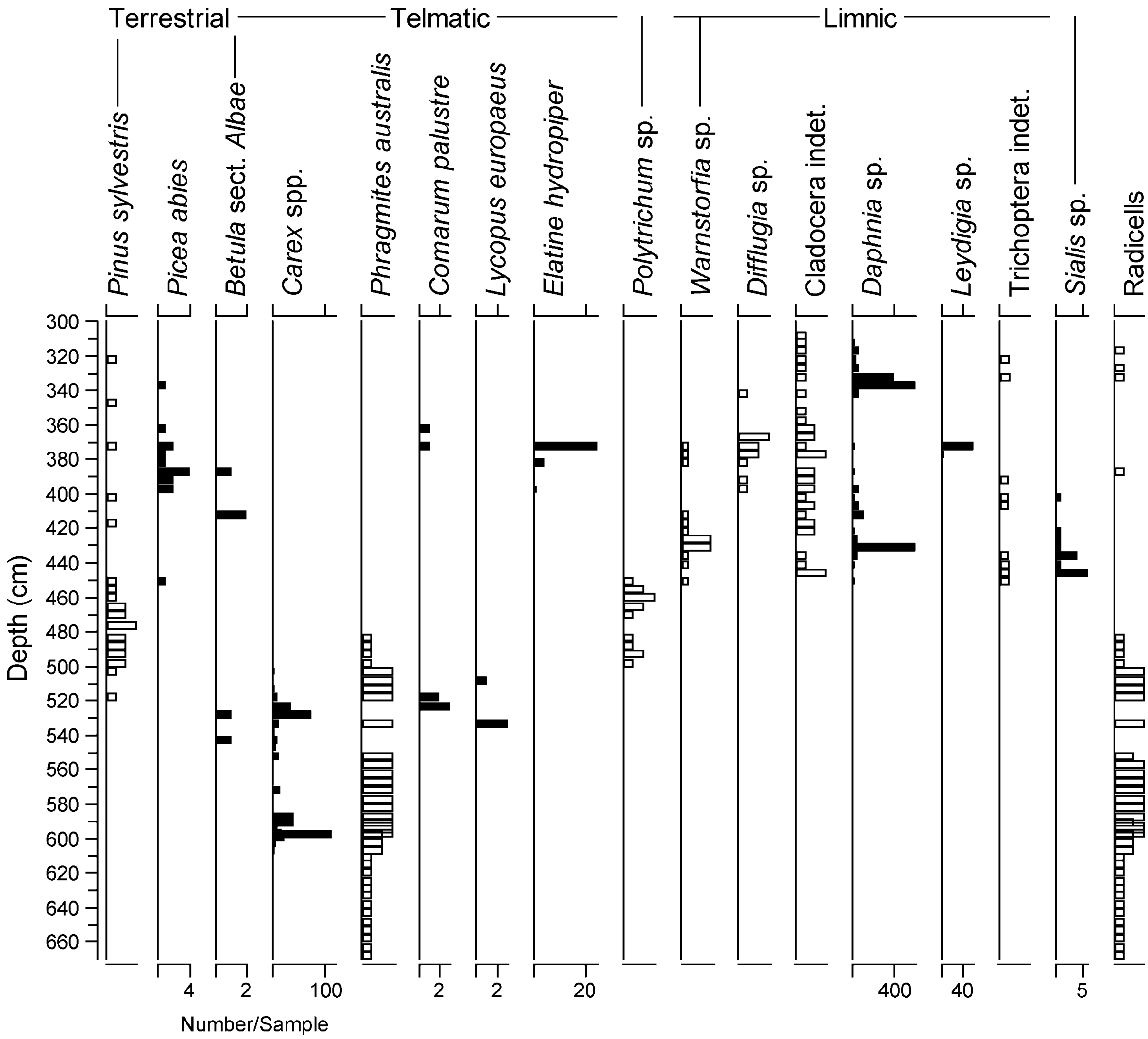

Figure 8. Macrofossil concentration diagram for Lake Zaozer'e. Note changes in scale. Hollow bars indicate relative abundance (short, rare; intermediate, common; long, abundant). Units: number of macrofossils/sample.

dominated by limnic species such as *Warnstorfia* sp., Cladocera, *Daphnia* sp., Trichoptera, and *Sialis* sp. (Figure 8). *Daphnia* sp. remains peak at 4.30 m. Together, the proxy records indicate lacustrine deposition from about 5300 cal yr BP onwards with little or no input of terrestrial organic matter.

The uppermost unit is a clayey algae gyttja (unit 1, 4.20–3.07 m). χ is stable, but SIRM increases slightly towards the top of the sequence, indicating presence of detrital magnetite (Figure 7). TC and TN contents are stable and the C/N ratio is around 10, indicating a limnic organic matter source. Limnic macrofossils dominate, but the few telmatic species (e.g., *Elatine hydropiper)*, together with some radicells (Figure 8) may indicate an expansion of the shoreline flora and possibly a slight lowering of the water level. Assuming a constant sedimentation rate, deposition of this unit occurred within the last 4000 years.

Terrestrial plant macrofossils are scarce throughout the whole sequence and include the first finds of *Betula* sect. *Albae* at 5.45 m (unit 5), *Pinus sylvestris* at 5.25 m (unit 5) and *Picea abies* at 4.53 m (unit 3) (Figure 8). The vegetation surrounding the site was composed of *Pinus* and tree birch from ca. 10,000 cal yr BP onwards, and included *Picea abies* by 5300 cal yr BP.

Discussion

The scarcity of terrestrial plant macrofossils and the presence of a long-lasting hiatus in Lake Nero only allow for a generalised reconstruction of the tree vegetation. Late glacial radiocarbon dates on wood fragments (Table 3) support the presence of woody vegetation in the catchment; however, the vegetation was likely dominated by *Betula* and *Salix* shrubs and various herbs between 15,000 and 13,000 cal yr BP. Based on the increase in pollen concentrations, we assume that *Betula* sect. *Albae* started to colonise the area surrounding the site about 14,000 cal yr BP (Figure 4). It is possible that *Picea abies* and *Pinus sylvestris* also became established around this time but with only a single record of fossil pollen assemblages it is difficult to attribute the increases in the pollen of these types to their arrival in the area. Due to a sedimentary hiatus, no pollen data are available from Lake Nero for the interval between 13,000 and ~8200 cal yr BP. At nearby Lake Chashnitsy, however, the local presence of *Pinus* and *Betula* sect. *Albae* can be inferred from plant macrofossils at >11,000 and 10,600 cal yr BP, respectively (Figure 6).

The interpretation of the *Picea* curve is more problematic. The isolated peak in *Picea* pollen concentrations between 14,000 and 13,000 cal yr BP may result from long-distance dispersal and/or redeposition from older sediments. It is possible, however, that it became established at this time along with tree birch. On the basis of macrofossil evidence, it is clear that *Picea* was present in the Valdai Highlands, about 400 km to the east of the Rostov-Jaroslavl' region, as early as 12,000 cal yr BP (Subetto 2003; B. Wohlfarth unpublished data), and to the north in eastern Russia Karelia by 10,750 cal yr BP (Wohlfarth et al. 2004). *Picea* was also present by 12,500 cal yr BP in north-eastern Russia, close to the Ural Mountains (Väliranta et al. in press).

Mixed deciduous – coniferous forests were common between 8200 and 6100 cal yr BP, and became dense and species-rich between 6100 and 2500 cal yr BP. The local presence of *Betula* sect. *Albae*, *Alnus glutinosa*, *Populus tremula*, *Frangula alnus*, *Picea abies* and *Pinus sylvestris* from 5300 cal yr BP onwards is demonstrated by macrofossil remains (Figures 3, 6 and 8). Noteworthy is the single *Picea* needle in Lake Zaozer'e sediments that date to around 5200 cal yr BP; this is consistent with the mid-Holocene expansion of *Picea* that occurred throughout Fennoscandia and adjacent areas (Giesecke and Bennett 2004). Agricultural activity in the area, which is documented in the pollen record for the last 500 years, probably led to an opening of the forests and to a decline in tree species (e.g., *Picea*, *Corylus*, *Quercus*, *Tilia* and *Ulmus*). Moreover, human activity associated with the settlement, which is reported to have been at the same place as modern Rostov since 862 A.D. and which became an important centre at the beginning of the 10th century, undoubtedly influenced the local vegetation over the last 12 centuries.

The comparably high TS values, which are observed in Lake Nero sediments, particularly between ca. 1500 and 500 cal yr BP (500–1500 A.D.) and at least in part attributed to external sulfur sources, may possibly be linked to human activity. It is known that land use changes lead to erosion of sulphide-rich soils and to increased TS content in lake sediments (e.g., Holmer and Storkholm 2001). Given the relatively high TS content in the Lake Nero sediments, we speculate that much of the sulphur could actually be derived from sulphide-rich soils in the catchment, and that input of sulphur to the lake increased in concert with enhanced land use. Other external sulphur sources are derived from mining and industrial activities. Ek et al. (2001) demonstrated, for example, that sulphur concentrations in lakes close to the Falun copper mine in Sweden increased in concert with the beginning of copper production ca. 1000 years ago. The town of Rostov had been an important economical, political and religious centre and some of the large church bells for the famous Ouspenie Cathedral were manufactured in 1654 A.D. This industrial activity could have been an additional source of pollution.

The bioclimatic limits of the tree species (Sykes et al. 1996) provide some information about palaeoclimatic conditions in the study area. The inferred presence of *Betula* trees by 14,000 cal yr BP and the dominance of forest vegetation since 11,000 cal yr BP, if not earlier, suggest that the moisture index (α), calculated as the ratio of actual to potential evapotranspiration (Sykes et al. 1996), was above 0.65. The continuous presence of tree *Betula* (i.e., *B. pendula*) and *Pinus* (i.e., *P. sylvestris*) requires $\alpha > 0.7$, while *Picea* needs $\alpha > 0.85$

(Sykes et al. 1996). In the study area, modern α values are close to 1.0 (Leemans and Cramer 1991). Lower *Picea* pollen percentage and concentration values between ~8200 and 2500 cal yr BP could be interpreted as a decrease in the abundance of *Picea* in the regional vegetation and therefore possibly decreased moisture availability ($\alpha \sim 0.85$ or even slightly lower). However, the presence of *Picea* macrofossils at 5200 cal yr BP in Lake Zaozer'e sediments suggests that spruce trees survived the relatively dry mid-Holocene in habitats with locally humid conditions, for example, around lakes and in river valleys. The interpretation of the pollen data in terms of winter temperatures is rather difficult. The presence of *Quercus* in the regional vegetation after 8200 cal yr BP shows that mean temperatures of the coldest month were above −16 °C (Sykes et al. 1996). *Betula* sect. *Albae* was present as early as 14,000 cal yr BP, but its local presence is only confirmed through macrofossils since 10,600 cal yr BP. It requires a minimum GDD* (i.e, the sum of growing-degree-days, which has a close linear correlation with mean July temperature) of 700 °C, which suggests that the sum of effective temperatures was above this limit between 14,000 and 13,000 cal yr BP and during the past ca. 10,600 cal yr. Temperate deciduous taxa were continuously present after ca. 8200 cal yr BP. Highest pollen concentrations for *Quercus*, which requires a minimum GDD* of 1100 °C, are registered in LPAZ Ne-5, Ne-4 and Ne-3 (Figure 3) i.e., between ~6000 and ~2500 cal yr BP and imply that this may have been the warmest interval of the period of record.

Summer temperature estimates based on macrofossil remains of *Myriophyllum alterniflorum* and *Potamogeton filiformis* suggest that mean July temperatures (MMJT) were higher than 8–10 °C (O. Bennike unpublished data) between ~14,800 and ~13,000 cal yr BP and increased to > 12 °C from 10,600 cal yr BP onwards, given the presence of *Pinus sylvestris* (Iversen 1954). MMJT of > 13 °C are indicated by macrofossil remains of *Frangula alnus*, *Nuphar luteum* and *Menyanthes trifoliata* (Kolstrup 1980; Brinkkemper et al. 1987) from ca. 5300 cal yr BP onwards.

Low lake productivity, high sedimentation rates and reduced bottom water conditions may have prevailed in the largest of the three studied lakes, Lake Nero, > 14,800 cal yr BP. A distinct shift towards higher lake organic productivity is observed around ~14,800 cal yr BP (Figure 9). This, together with increased input of terrestrial plant material and the development of an extensive telmatic vegetation zone, suggests regionally warmer temperatures. This broadly corresponds to the beginning of the warming seen in ice core, marine and terrestrial records at around 14,700 cal yr BP (Johnsen et al. 1992; Björck et al. 1996, 1998; von Grafenstein et al. 1999; Walker et al. 1999; Walker 2001; Björkman et al. 2002). The gradual lowering of the water level in Lake Nero between 14,000 and 13,000 cal yr BP, the formation of a shoreline close to the coring site and evidence for a long-lasting hiatus between ~13,000 and ~8200 cal yr BP (Figure 9) shows that the level of the lake decreased considerably and remained low for several thousand years. Although a rise in water level is only documented after ~8200 cal yr BP, it is possible that this rise occurred earlier, eroding the underlying sediment. Thus, the estimate of ~8200 cal yr BP should be regarded as a minimum age for renewed lacustrine conditions at the coring site.

Assuming that the lake surface and surface inflow were similar to today around 14,000 cal yr BP, the level of the lake would have to decrease by ~5.5 m to allow for the formation of a shoreline close to the coring site (Figures 2 and 9). A water level lowering of ~5.5 m for a lake as large as Nero and persistent low lake levels between ~13,000 and ~8200 cal yr BP could have been caused by a number of factors, such as melting of permafrost, erosion of the threshold, aridity, and/or regional hydrological changes.

The present water budget of Lake Nero is mainly controlled by surface inflow via River Sara and other small tributaries (Figure 1b) and only to a minor extent by groundwater. Consequently, decreased inflow via River Sara would lead to a lowering of the water level. Decreased flow of River Sara, on the other hand, may have been linked to a lowering of the groundwater table in the region.

Distinct lake-level lowering is also reconstructed for the two higher elevation lakes; however, there the low lake-level phase occurred between ~9000 and ~5500 cal yr BP (Figures 1b–d and 9) i.e., several thousand years later than in Lake Nero. Lakes Chashnitsy and Zaozer'e formed > 11,000 cal yr BP and were initially shallow productive lakes. In both basins the water level

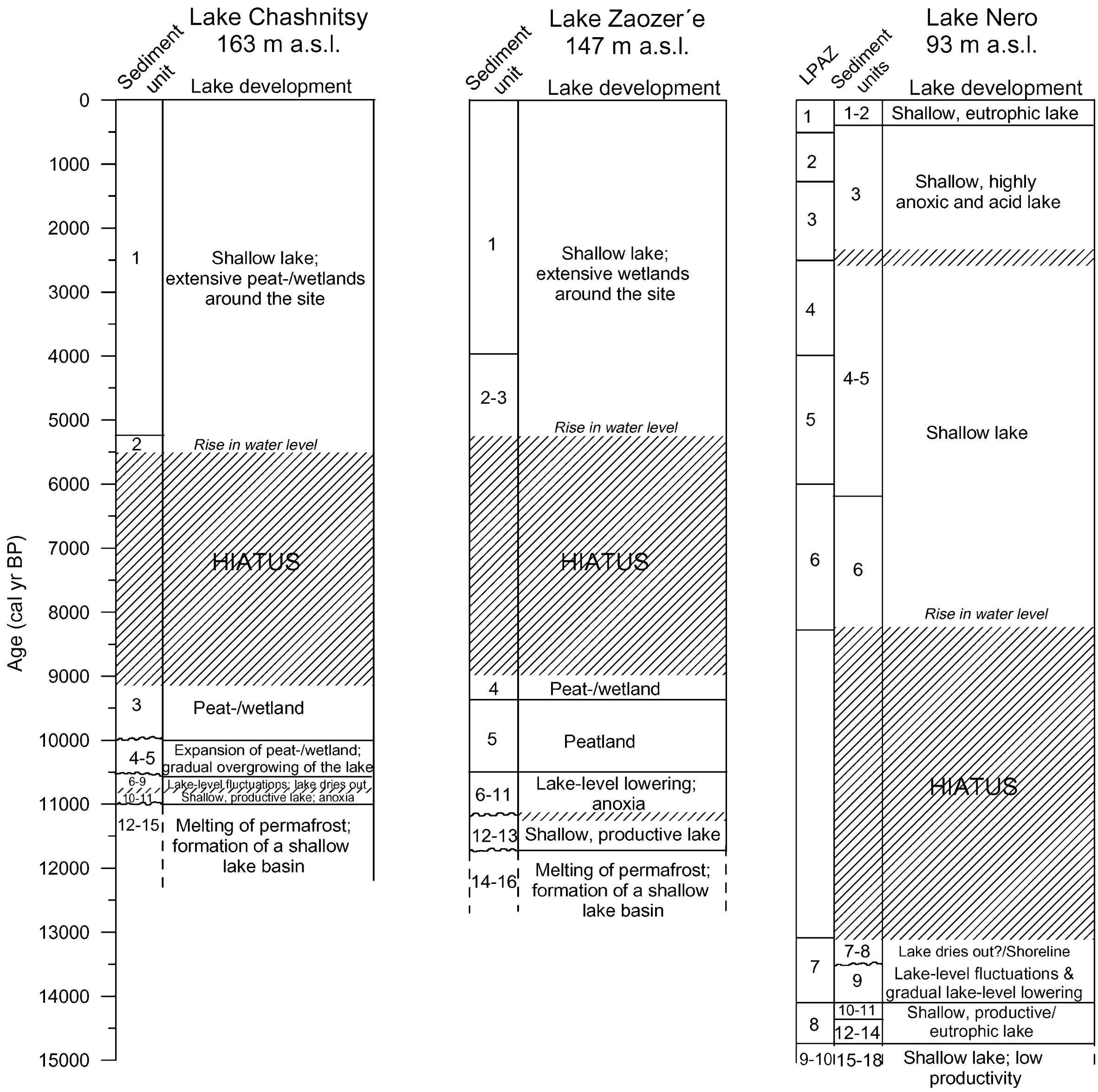

Figure 9. Summary diagram of environmental changes in the Rostov-Yaroslav'l area. Cross hatching indicates a hiatus.

decreased between ~11,000 and ~10,500 cal yr BP and peatland expanded over the coring sites (10,500–9000 cal yr BP). This development implies a water-level lowering of ca. 6 m in Lake Chashnitsy (Figures 5 and 9) and of ca. 4.5 m in Lake Zaozer'e (Figures 7 and 9). A renewed rise in water level and a return to open water conditions at the coring sites is documented only at ~5300 cal yr BP. Both lakes are closed basins and the observed hydrological changes therefore could not have been caused by variations in inflow and/or erosion of the threshold, but could have been linked to disintegration of permafrost, higher evaporation and/or changes in groundwater level.

This leads to a number of questions: were the observed changes in the three lake basins linked with each other, although the timing of the lake-level variations does not seem compatible and the higher elevation lakes are closed basins? Or, were the observed lake-level fluctuations the result of different mechanisms? To answer these questions, we explore and discuss two different scenarios:

Scenario (1) – Permafrost hypothesis - 'thaw lakes':

During the LGM continuous permafrost occupied the study region, and gradually started to disintegrate during the late glacial interstadial and early Holocene. It is possible that infilling of the Lake Nero basin 14,800 cal yr BP resulted from a combination of increased temperature and moisture. The formation of the shallow lake may have enhanced disintegration of permafrost through thermal erosion along the lake margin, which in turn could have led to erosion of the threshold and rapid drainage of the lake basin. However, the water budget in Lake Nero is mainly controlled by inflow through River Sara and it seems unlikely that this inflow ceased completely and restarted again around 8200 cal yr BP, without invoking major regional hydrological changes.

Lakes Chashnitsy and Zaozer'e, on the other hand, are closed basins, and could have been initiated by surface ponding on permafrost during the earliest part of the Holocene (Figure 9). Gradual degradation of the surrounding permafrost may have initially increased lake size, but the ongoing disintegration of permafrost due to warmer air temperatures likely led to rapid sub-surface drainage of the basins. Thaw lakes and thaw lake cycles have been described from present-day permafrost areas (e.g., Hinkel et al. 2003; Yoshikawa and Hinzman 2003; Smith et al. 2005 and references therein), but their formation has also been demonstrated for the early Holocene (Ritchie et al. 1983; Burn 1997). Some thaw lakes may drain completely, but often lakes only drain partially, leaving a residual lake within the older basin (Hinkel et al. 2003). Vegetation is established in drained or partially drained basins. Infilling of lakes Chashnitsy and Zaozer'e started around 5500 cal yr BP, which broadly corresponds to the initiation of mid-Holocene cooling and increase in humidity described for northernmost Finland and European Russia (e.g., Solovieva and Jones 2002; Sarmaja-Korjonen et al. 2004; Väliranta et al. 2005).

Scenario (2) – Groundwater change hypothesis:

Regionally lower groundwater levels caused decreased river inflow to Lake Nero around 13,000 cal yr BP and gradually decreased the water level in the lake. When the groundwater lowering reached a critical threshold, lake basins situated at higher elevations may have become influenced, as is indicated by the presence of hiatuses of shorter duration around ~11,100, ~11,000 and ~10,700 cal yr BP and the start of the long-lasting hiatus at ~9000 cal yr BP in lakes Chashnitsy and Zaozer'e (Figure 9). Before or around ~8200 cal yr BP, the groundwater level started to rise and increased inflow to Lake Nero led to a rise in lake level. The higher groundwater table affected the higher elevation lakes Chashnitsy and Zaozer'e ca. 3000 years later. Given scenario 1, major hydrological changes must have occurred in the region between 13,000 and 8200 cal yr BP leading to a marked lowering of the groundwater level. The occurrence of two additional hiatuses in Lake Nero around 2500 cal yr BP and possibly before 500 cal yr BP, suggest that fluctuating groundwater conditions may be rather normal for this region.

It is possible to explain the hydrological changes reconstructed for lakes Chashnitsy and Zaozer'e with the thaw-lake concept and the known mid-Holocene climatic changes reconstructed for European Russia. The development seen in Lake Nero may be more compatible with scenario (2). However, continental-scale syntheses of lake data used to reconstruct changes in late Quaternary regional climates of northern Eurasia identify the late-glacial period as a drier-than-present phase (e.g., Harrison et al. 1996). Drier conditions along the margin of the Fennoscandian ice sheet are likely linked to the development of glacial anticylconic circulation over the ice sheet and the associated stronger easterly winds (Kutzbach et al. 1993; Harrison et al. 1996). Conditions approach those similar to today with the disintegration of the ice sheet in the early Holocene about 9000 cal yr BP and this coincides with the return to wetter conditions at Lake Nero. Clearly there is a need for more studies in the region, with better temporal resolution to decipher which scenario applies and to better understand the underlying causes of the observed lake-level fluctuations.

Conclusions

Chronological and lithological problems in the form of hiatuses in lakes Nero, Chashnitsy and Zaozer'e make it difficult to determine the timing and nature of late Quaternary palaeoenvironmental changes in the Rostov-Jaroslavl' region north of Moscow. In Lake Nero, which is the largest of the three lakes, a shallow, productive lake developed around 15,000 cal yr BP. Lithological and macrofossil evidence indicate lake-level fluctuations between 14,000 and 13,500 cal yr BP, followed by a low lake-level phase and possibly a dried-out lake between 13,000 and 8200 cal yr BP. These relatively dry conditions are likely linked to anticylconic circulation over the Fennoscandian ice sheet and/or regional changes in groundwater. Lakes Zaozer'e and Chashnitsy were formed ca. 11,000 cal yr BP, possibly in response to melting of permafrost. After a shallow lake phase with increased organic productivity, lake levels started to decrease and portions of the basins became overgrown around 9000 cal yr BP. The 'thaw-lake' concept and known mid-Holocene climatic changes reconstructed for European Russia may account for the drainage of lakes Zaozer'e and Chashnitsy. The beginning of a new lake phase is registered ca. 5500 cal yr BP. Further research may help refine and explain the reconstructed fluctuations in lake level.

Due to the hiatuses in the three lake-sediment sequences, the palaeovegetation cannot be continuously reconstructed and determining the initial establishment of trees in the lake's catchment is particularly problematic. Pollen percentages and concentrations suggest that tree *Betula* was established in the region around 14,000 cal yr BP. The local presence of *Pinus sylvestris* and tree *Betula* can be inferred from plant macrofossils in sediments that date to >11,000 and 10,600 cal yr BP, respectively, implying that forests dominated by *Betula* and *Pinus* occupied the area and that minimum mean July temperatures were above 12 °C (Iversen 1954) by the beginning of the Holocene. Mixed broad-leaved – coniferous forests were widespread in the area between 8200 and 6100 cal yr BP and developed into dense, species-rich forests between 6100 and 2500 cal yr BP, during what was likely the warmest interval of the studied sequences. *Picea abies* was certainly growing locally by 5200 cal yr BP but may have been present during the late glacial period; further reserach is needed to clarify the late glacial and early Holocene palaeoecology of this conifer in the region. Agricultural activity is documented since 500 cal yr BP, but probably began earlier.

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