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Holocene climate and environmental change in north-eastern Kamchatka (Russian Far East), inferred from a multi-proxy study of lake sediments

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ARTICLE INFO

ABSTRACT

A sediment record from a small lake in the north-eastern part of the Kamchatka Peninsula has been investigated in a multi-proxy study to gain knowledge of Holocene climatic and environmental change. Pollen, diatoms, chironomids and selected geochemical parameters were analysed and the sediment record was dated with radiocarbon. The study shows Holocene changes in the terrestrial vegetation as well as responses of the lake ecosystem to catchment maturity and multiple stressors, such as climate change and volcanic eruptions. Climate change is the major driving force resulting in the recorded environmental changes in the lake, although recurrent tephra deposition events also contributed. The sediment record has an age at the base of about 10,000 cal yrs BP, and during the first 400 years the climate was cold and the lake exhibited extensive ice-cover during winter and relatively low primary production. Soils in the catchment were poor with shrub alder and birches dominating the vegetation surrounding the lake. At about 9600–8900 cal yrs BP the climate was cold and moist, and strong seasonal wind stress resulted in reduced ice-cover and increased primary production. After ca. 8900 cal yrs BP the forest density increased around the lake, runoff decreased in a generally drier climate resulting in decreased primary production in the lake until ca. 7000 cal yrs BP. This generally dry climate was interrupted by a brief climatic perturbation, possibly attributed to the 8.2 ka event, indicating increasingly windy conditions with thick snow cover, reduced ice-cover and slightly elevated primary production in the lake. The diatom record shows maximum thermal stratification at ca. 6300–5800 cal yrs BP and indicates together with the geochemical proxies a dry and slightly warmer climate resulting in a high productive lake. The most remarkably change in the catchment vegetation occurred at ca. 4200 cal yrs BP in the form of a conspicuous increase in Siberian dwarf pine (*Pinus pumila*), indicating a shift to a cooler climate with a thicker and more long-lasting snow cover. This vegetational change was accompanied by marked shifts in the diatom and chironomid stratigraphies, which are also indicative of colder climate and more extensive ice-cover.

1. Introduction

To trace and reconstruct long-term environmental changes in aquatic as well as terrestrial ecosystems multi-proxy stratigraphic analysis of lake sediments is a widely used and successful approach (Lotter et al., 1995). Aquatic ecosystems at high latitudes are susceptible to, and commonly respond promptly to climate change. In particular there is a straightforward relationship between climate and aquatic environment in lakes situated near ecotonal boundaries (Smol et al., 2005; Lotter et al., 2010). Many physical properties of lakes, e.g., the duration of ice-cover and thermal stratification, are crucially linked to algal dynamics and community structure as these factors affect the light availability, available habitats and nutrient cycle (Smol et al., 2005; Rühland et al., 2008). Warming trends decrease the length of

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lake ice-cover which prolongs the growth season and expands the zone of available habitats and new substrates for littoral diatom taxa and the development of more diverse diatom communities (Douglas et al., 1994; Smol et al., 2005; Douglas and Smol, 2010).

The Kamchatka Peninsula in eastern Russia is one of the most remote and least studied regions of eastern Asia. With 29 active and 300 extinct volcanoes, Kamchatka is one of the most active volcanic regions in the world. The location at the Pacific Rim contributed to its significant strategic military importance, and a large part of the peninsula was prohibited territory until the collapse of the Soviet Union in the early 1990s. Kamchatka has a sparse population of about 400,000, of which the majority live in the city of Petropavlovsk-Kamchatsky, which also contributes to the preservation of a relatively unexplored and pristine terrain. Atmospheric circulation and climate variability in Beringia today is depending mainly on the strength and position of the Siberian High and Aleutian Low (Mock et al., 1998). In a palaeoclimatic perspective variability in sea surface temperature suggests a close atmospheric coupling between the North Pacific and North Atlantic until 10,000 cal yrs BP, visible as similarities between the NGRIP oxygen isotope record and reconstructed sea-surface temperatures from the western Bering Sea, NW Pacific and Sea of Okhotsk (Max et al., 2012). However, during the last 7000 years the sea surface temperatures in the North Pacific area show more complex variations, suggesting strong regional overprints (Max et al., 2012). Changes in the position and activity of the Aleutian Low, sea ice distribution and summer insolation have systematically influenced Holocene climate in the subarctic North Pacific (Harada et al., 2014). Eastward displacement or increased intensity of the Aleutian Low seems to correspond with increased sea ice extent in the western Okhotsk Sea and the eastern Bering Sea (Harada et al., 2014). Between 10,000 and 6500 cal yrs BP the atmospheric pressure pattern over the Okhotsk Sea shows a northern Aleutian Low mode resulting in expanded easterly distribution of sea-ice (Katsuki et al., 2010). Studies of oxygen isotope data from the Yukon Territory, northwestern Canada, indicate millennial-scale variations in the position and intensity of the Aleutian Low pressure system (Anderson et al., 2005).

Palaeoclimatic reconstructions from the Kamchatka region have been carried out mainly based on terrestrial peat sequences (Dirksen et al., 2013 and references therein) but there are also reconstructions based upon studies of glacial deposits (Barr and Solomina, 2014) and marine sediment records from the Sea of Okhotsk, western Bering Sea and north-western Pacific Ocean (Max et al., 2012). The earliest palynological investigations of the landscape development in Kamchatka date to the late 1960s/early 1970s (e.g., Braitseva et al., 1968, 1973). The postglacial vegetation development starts with the predominance of dwarf birch tundra with grasses and lower shrub communities, and the most distinct shift was an expansion of *Pinus pumila* (Siberian dwarf pine) during the mid-Holocene (Khomentovsky, 2004; Dirksen et al., 2013). Vegetation development in Kamchatka is explained by the influence of climatic factors, but also by the impact of geological factors such as frequent volcanic eruptions (Khomentovsky, 2004). There is a direct impact of volcanic eruptions on vegetation, visible as modified soil properties and nutrient cycles, which favour terrestrial growth (Urrutia et al., 2007). Volcanic events could also affect aquatic ecosystems through direct increases in nutrient input from ashfall on the water surface (e.g., silica, phosphorous, and sulphur), or as input of organic matter from vegetation damage in the surrounding catchment and changed water pH (Harper et al., 1986). Changes in physical conditions of lakes, such as temporarily reduced light penetration, sealing of the sediment–water interface, or burial of macrophytes in the littoral zone could further disrupt the aquatic environment (Urrutia et al., 2007). Such changes of lacustrine ecosystems can, in turn, produce important changes in algal communities and benthic fauna (Urrutia et al., 2007). Numerous studies have investigated the impact of tephras on diatom assemblages in lakes from volcanic areas all over the world, and the results seem to vary between significant correlation to minor registered impacts (e.g., Harper et al., 1986; Hickman and Reasoner, 1994; Barker et al., 2000, 2003; Telford et al., 2004; Hoff et al., 2013) with diatoms in lakes responding only about one half of the time from tephra deposition (Telford et al., 2004). Previous studies concluded that the predominant millennial-scale forcing of the diatom flora relates to climate shifts despite frequent centennial-scale perturbations to tephra influx (Barker et al., 2003; Telford et al., 2004).

Until recently only few palaeolimnological studies of Holocene environmental and climate change have been undertaken in Kamchatka, and most of these have focused on the southern part of the peninsula (e.g., Hoff et al., 2012, 2013; Nazarova et al., 2013). The present investigation was carried out as a part of the Swedish BERINGIA 2005 expedition (Bennett et al., 2006). We present a multiproxy study of a Holocene lake sediment sequence, which includes analyses of pollen, chironomids, diatoms and geochemistry. The study site, Pechora Lake, located in north-eastern Kamchatka, was selected based on its relatively remote distance to active volcanoes and its proximity to the Pacific Ocean, which has a pronounced influence on the regional climate (Ivanov, 2002). The aim of this study is to provide novel insights into Holocene climate and environmental change from the poorly studied area of northern Kamchatka. Our study allows us to interpret both terrestrial vegetation changes and lacustrine responses to catchment maturity and multiple stressors, such as climate change and volcanic eruptions throughout the Holocene. We identify drivers of lake development and outline the causal-link between the aquatic ecosystem response, e.g., succession in species composition and fluctuating primary production, and regional climate change.

2. Setting

The Kamchatka Peninsula is 1250 km long and up to 500 km wide (51–60°N, 156–163°E), covering an area of about 472,000 km^2 located between the Sea of Okhotsk to the west and the Bering Sea to the east (Fig. 1).

Pechora Lake (unofficial name; 59°17.6′N, 163°07.8′E), is a small lake (approximately 50 x 300 m, water depth ca. 4.3 m) located about 2 km from the Bering Sea coast, near the town of Ossora in north-eastern Kamchatka (Fig. 1). The area is scattered with small lakes that consists of depressions in the irregular moraine terrain. The water is circumneutral with pH 6.7 and a conductivity of 16 μS measured on a single occasion during fieldwork in August 2005 at water temperature 16.3 °C. The lake, which is hydrologically open with well-defined inlet and outlet streams, is situated at an altitude of 45 m a.s.l., and the nearby mountains reach altitudes of 200–800 m (Fig. 2). Pechora Lake is partly surrounded by *Sphagnum* peatlands and the present-day vegetation is characterised by bushes of *P. pumila* and *Alnus viridis* (mountain alder), which grow on the higher ground near the lake. *Betula nana* (dwarf birch), *Empetrum nigrum* (crow berry) as well as various herbs and grasses are also common. The slopes of the mountains in the vicinity are covered by tree birches. The nearest active volcano is Shiveluch which is located ca. 300 km south of the study area.

The present-day climate of Kamchatka is influenced by the position of the peninsula at the eastern margin of the Asian landmass and the proximity to the cold Bering Sea and Sea of Okhotsk (Ivanov, 2002). This results in a temperate monsoon climate south of 60°N, controlled by the interaction between the Siberian High and the Aleutian Low pressure systems (Mock et al., 1998). At the north-eastern coast the maritime influence results in strong cyclonic activity which reduces the impact of the monsoon (Ivanov, 2002). The complex topography of the peninsula with two high mountain ranges also affects the local climate resulting in an interior with relatively dry climate and marked seasonal variations in temperature. The coastal margin of the Ossora
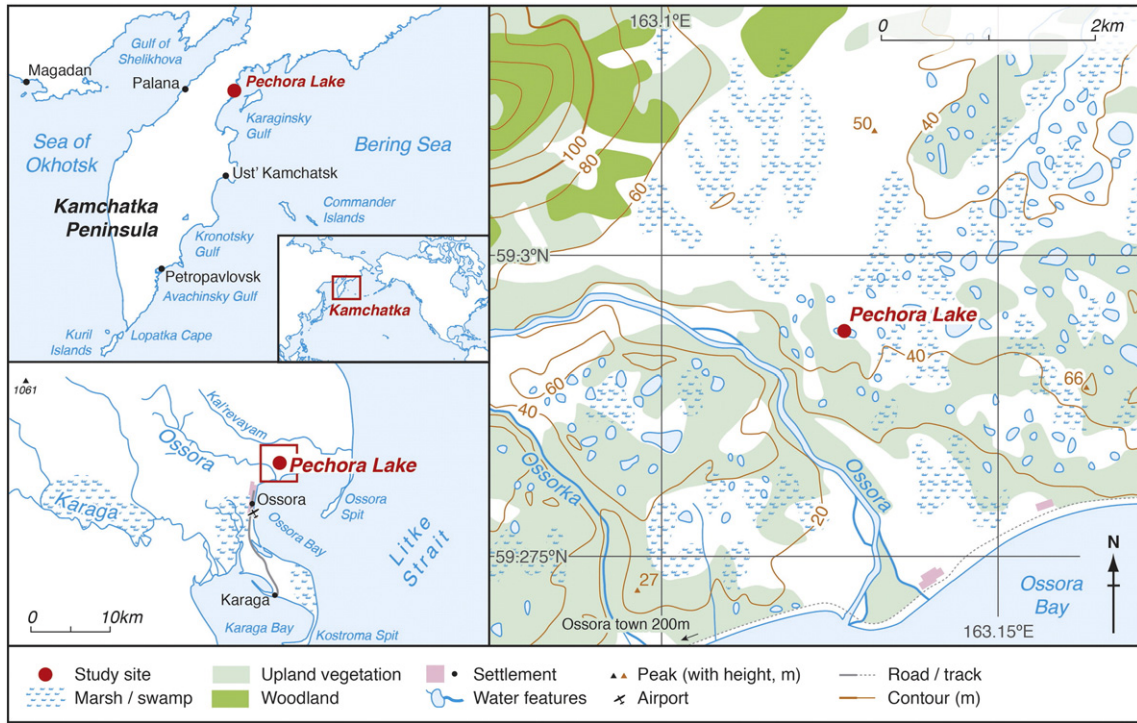
Fig. 1. Map of Pechora Lake and its situation near the north-eastern coast of Kamchatka, Russian Far East.

region is characterised by a cool and maritime climate with cold summers and short growing seasons. The meteorological station at Ossora (WMO ID 32246, 59°15′N, 163°04′E, running since April 2008), displays an average temperature (5-years mean 2010–2014) of the warmest months (July/August) of 13.7 °C/13.6 °C and the coldest months (January/February) − 15.7 °C/− 15.5 °C. Annual precipitation (only available from 2014) is 737 mm yearly mean (http://rp5.ru/). Monthly mean temperatures are under 0 °C between November and April (5-years mean 2010–2014), and precipitation fall as snow between November and May (snow cover data 2012–2014). Maximum snow depth reached 197–83 cm during January to April year 2012 to 2014.

3. Methods

3.1. Coring, sub-sampling and dating

Sediment cores were sampled with a 1-m long, 5-cm diameter Livingstone piston corer (Wright et al., 1984), operated from a rubber boat in the centre of the lake. To stabilise the coring platform the rubber boat was tied to crossing ropes extending across the entire lake. Water depths were measured with a scaled plumb line from the water surface. Multiple holes were cored with 0.5 m depth displacement to assure recovery of a complete stratigraphy. Piston cores were retrieved as 1-m long sediment sections which were wrapped in plastic film, aluminium

Fig. 2. Photo showing Pechora Lake in a westerly direction with the nearby mountains about 12 km away in the background. For scale use the rubber boat and silhouette of two persons standing on the shore.

foil, thick plastic and modified plastic drain pipes, and stored in wooden boxes until sub-sampling was carried out in the laboratory.

A sample at the base of the sediment sequence was sampled for AMS ^{14}C dating directly in the field and measured in the Radiocarbon Dating Laboratory at Lund University. Six additional bulk sediment samples evenly distributed along the stratigraphy were subsampled in the laboratory for AMS ^{14}C dating at Queen's University Belfast. Seven visible tephra layers were subsampled for geochemical analyses to allow for tephrochronology. Further details of the chronological approach are provided by Plunkett et al. (2015–in this issue).

Biological proxy data will be stored in an appropriate database (such as Neotoma) as soon as possible. Until then, data can be obtained on request to the authors.

3.2. Pollen

Pollen samples were taken volumetrically (0.5 cm[3]) at every 4 cm with a plastic syringe with the nozzle removed as described in Maher (1981). Sub-samples for the analysis of pollen and spores were treated following conventional methods using HCl, KOH, HF and acetolysis (Bennett and Willis, 2001). Two *Lycopodium* spore tablets were added to the volume-specific samples (0.5 cm[3]) to enable concentration calculations (Stockmarr, 1971). Microscopic analysis was generally carried out to a minimum count of 500 pollen grains (if possible) of terrestrial trees, shrubs and herbs under a magnification of ×400 for routine identification. Critical types were determined by ×1000 magnification using an oil-immersion (anisole). It was aimed to count at least 100 added *Lycopodium* spores per sample to allow statistical analyses.

The total pollen sum used for determination of relative frequencies includes terrestrial pollen types. Percentages of other taxa (i.e., obligate aquatics, non-pollen palynomorphs, indeterminable and unknown grains) are based upon the total pollen sum plus the count of the taxon in question. The palynological nomenclature generally follows Beug (2004). Stratigraphic zonation of the pollen data was performed using psimpoll 4.27 (Bennett, 2009) and six different methodologies of numerical zonation techniques were tested (Klimaschewski, 2011). Percentage pollen and spore dataset were used and only taxa exceeding a threshold of 5% at any level in the dataset were included into the calculations (Gordon and Birks, 1972; Birks and Berglund, 1979; Birks, 1973; 1986). The datasets were recalculated to the sum of the included taxa (Bennett, 1996). Each method was assessed against a broken-stick-model to determine precisely the number of significant zones in pollen sequences (Bennett, 1996). Optimal splitting by information content, carried out with non-randomised and non-transformed samples of the summarised diagram was selected to be the most suitable method for zonation. The rate-of-change measures the degree of dissimilarity between adjacent samples (e.g., Lotter et al., 1992). We used Chi-squared coefficient 2 available in psimpoll (Bennett, 2009) as a measure of dissimilarity. Principal component analysis (PCA) was used to define plant communities (cf. Seppä and Bennett, 2003) based on square-root transformation with covariance matrices using only taxa exceeding a threshold of a minimum of 5% of the main pollen sum. The reduced datasets were recalculated as proportions of the sum of types that are included (Bennett, 2009). According to Birks and Line (1992) rarefaction analysis is a useful tool to measure the palynological richness of pollen spectra. Changes in the palynological richness are interpreted as reflecting changes in the floristic richness and the mosaic structure of a landscape through time.

3.3. Chironomids and temperature reconstruction

Chironomid samples were analysed at 4–12 cm resolution. The samples (0.41–1.13 g wet weight) were heated to 75 °C in 5% KOH for 5 min, and progressively passed through 212 μm and 90 μm mesh sieves (Brooks et al., 2007). Chironomid head capsules were picked out of the sieving residues under 40× magnification and mounted in Euparal©. A minimum of 50 head capsules were picked for each sample (Heiri and Lotter, 2001). Chironomids were identified with reference to Wiederholm (1983), Rieradevall and Brooks (2001) and Brooks et al. (2007). The stratigraphic zonation of the chironomid data was performed within ZONE version 1.2 (Juggins, 1991) and the statistical significance of the zones was assessed using BSTICK (Bennett, 1996). Detrended correspondence analysis (DCA), a unimodal indirect ordination method (Hill and Gauch, 1980), was performed on the chironomid relative abundance data to summarise compositional changes over time.

Mean July air temperatures were inferred using a chironomid-based temperature inference model (WA-PLS, 2 component, $R^2_{boot} = 0.81$, $RMSEP_{boot} = 1.43$ °C) (Nazarova et al., 2015) based on a modern calibration dataset of 88 lakes from eastern Russia (53°–75°N, 141°–163°E) covering a mean July air temperature range of 1.8–13.3 °C. The reliability of the chironomid-inferred temperature reconstructions was assessed by the following methods. Less reliability was placed on fossil samples in which (1) fewer than 50 chironomid head capsules were extracted, (2) more than 5% taxa were not represented in the modern training set or (3) more than 5% taxa were rare in modern dataset (i.e., Hill's N2 less than 5) (Heiri and Lotter, 2001; Heiri et al., 2003, 2007). (4) Goodness-of-fit to temperature was evaluated by passively positioning the fossil samples on a CCA of the modern training set constrained solely against July temperature (Heiri and Lotter, 2001). Any fossil samples that had a squared residual distance value within the 10th percentile of values in the modern training set were considered to have a poor fit-to-temperature. (5) The modern analogue technique (MAT) was used to detect fossil samples that lacked good analogues in the modern calibration dataset using squared chord distance as a measure of dissimilarity. Samples with a dissimilarity larger than the 95% threshold in the modern data were considered as having no good analogues in the modern calibration dataset (Birks et al., 1990; Birks, 1995, 1998; Velle et al., 2005). (6) The significance of the reconstructions was also evaluated using the palaeoSig package (Telford, 2011) in R (R Development Core Team, 2013) with 999 random reconstructions. Following Telford and Birks (2011), a reconstruction is considered statistically significant if it explains more of the variance in the fossil data than 95% of reconstructions that are derived from random environmental variables. Chord-squared distance and Hill's N2 values were calculated with the program C2 version 1.7.4 (Juggins, 2007). DCA and CCA were performed using CANOCO 4.5 (ter Braak and Šmilauer, 2002). Species data were square-root transformed to stabilise species variance and rare species were downweighted. In the evaluation of goodness-of-fit (4) the CCA scaling focused on inter-sample distances with Hill's scaling selected to optimise inter-sample relationships (Velle et al., 2005).

3.4. Diatoms and pH-reconstruction

Sediment samples were cleaned for diatom analysis following Battarbee et al. (2001) at sample intervals of between 4 and 8 cm and mounted on permanent slides with Naphrax™. Diatoms were identified and counted under an Olympus BX51 light microscope using the Nomarski differential interference contrast with a magnification of ×1000 and oil immersion. At least 300 valves were enumerated in each sample following the protocol of Schrader and Gersonde (1978). Diatom identification followed Krammer and Lange-Bertalot (1986, 1988, 1991a,b), Camburn and Kingston (1986), Lange-Bertalot and Metzeltin (1996), Fallu et al. (2000) and Antoniades et al. (2008). The stratigraphic zonation of the diatom data was performed with CONISS within TILIA version 1.7.16 (Grimm, 1987). Detrended correspondence analysis (DCA), was used on the total diatom assemblage to summarise compositional changes in the diatom community over time and performed using the vegan package in R (Oksanen et al., 2013). Species richness of the fossil diatom assemblages was estimated using rarefaction analysis (Birks and Line, 1992). pH-values were inferred using

AL:PE pH-diatom model (Cameron et al., 1999, WA classical, $R^2_{jack} = 0.78$, $RMSEP_{jack} = 0.37$) since it contains closest modern analogues. The reliability of the pH reconstructions was estimated using the same statistical approach as with the chironomid-based temperature reconstruction (see above): (1) goodness-of-fit to pH was evaluated by passively positioning the fossil samples on a CCA of the modern training set constrained solely against pH (Heiri and Lotter, 2001); (2) the modern analogue technique (MAT) was used to detect fossil samples that lacked good analogues in the modern calibration dataset using squared chord distance as a measure of dissimilarity (Birks et al., 1990; Birks, 1995, 1998; Velle et al., 2005) and (3) the significance of the reconstructions was also evaluated using the palaeoSig package (Telford, 2011) in R (R Development Core Team, 2013) with 999 random reconstructions. Chord-squared distance was calculated with the program C2 version 1.7.4 (Juggins, 2007). CCA was performed using CANOCO 4.5 (ter Braak and Šmilauer, 2002). Species data were square-root transformed to stabilise species variance and rare species were downweighted. In the evaluation of goodness-of-fit (4) the CCA scaling focused on inter-sample distances with Hill's scaling selected to optimise inter-sample relationships (Velle et al., 2005).

3.5. Geochemistry

Biogenic silica was analysed following the wet chemical digestion technique described by Conley and Schelske (2001). Results are presented as wt% SiO_2 of total dry weight.

Sediment samples, generally taken at 4-cm intervals, were treated with 10% HCl to remove potential carbonate material, rinsed with de-ionised water, freeze-dried, and then sieved using a 500-μm mesh. Determination of total organic carbon (TOC) and total nitrogen (TN) contents as well as stable-carbon isotope analyses were performed on the fine-grained fraction by an elemental analyser interfaced with a continuous-flow isotope-ratio mass spectrometer (CF-IRMS) at the University of Waterloo Environmental Isotope Laboratory (UW-EIL). The results are expressed as δ values, representing deviations in per mil (‰) from the VPDB standard, such that $\delta^{13}C_{sample} = 1000[(R_{sample}/R_{standard}) - 1]$, where R is the $^{13}C/^{12}C$ of the sample and standard, respectively. The uncertainty, based on repeated analyses of samples from the same level, gives TOC and TN results that are within ±0.1% and $\delta^{13}C$ within ±0.1‰. Based on the cross-plot test of corresponding TOC and TN data (Talbot, 2001) minor amounts of inorganic N were detected and corrected for to allow calculation of total organic nitrogen (TON), as well as atomic C/N ratios. TOC contents are expressed as dry-weight percentages.

4. Results

4.1. Lithostratigraphy and chronology

Five holes were cored at 427 cm water depth in the centre of the lake, resulting in three full parallel sediment sequences. The coring stopped at 865 cm below the water surface without reaching any underlying minerogenic strata. In the laboratory the most complete sequence, 423 cm long, was selected for radiocarbon dating and all

biostratigraphical, geochemical and lithological analyses. The top of sediment measures 442 cm which means that the uppermost 15-cm part of the sediment sequence was lost during core retrieval. The sediments consist of dark brown homogeneous detritus gyttja, abruptly interrupted at seven levels by macroscopic layers (1–3 cm) of light-coloured volcanic tephra.

An age–depth model based on seven bulk sediment samples (Table 1) was produced using Clam version 2.2 applying a smooth spline of 0.1 (Blaauw, 2010). Tephra layers in Pechora Lake provided valuable time-stratigraphic markers with nearby Lifebuoy Lake, but due to the uncertain ages of the tephra layers, they did not contribute to the Pechora age model (Plunkett et al., 2015–in this issue; Solovieva et al., 2015–in this issue). ^{14}C dates were calibrated using the Northern Hemisphere calibration curve IntCal13 (Reimer et al., 2013) and all dates are given in calibrated years before present (AD 1950), cal yrs BP. The sediment surface was assumed to be of recent age at the time of core collection (AD 2004 ± 5 (−54 BP)). Tephra layers of >1 cm in thickness were excised from the age–depth model as slumps, since they represent abrupt deposition events that interrupted the otherwise smooth sediment accumulation. The age model gave an age for the base of the sequence of 10,065 cal yrs BP and sediment accumulation rates between 7 and 38 yr cm^{-1} (Plunkett et al., 2015–in this issue). The age model indicates moderate (22–38 yr cm^{-1}) sediment accumulation up until c. 5800 cal yrs BP, followed by a period of rapid accumulation (7–20 yr cm^{-1}) to ca. 3900 cal yrs BP and thereafter moderate rate of accumulation (20–35 yr cm^{-1}). The period with increased sedimentation rate coincides with maximum primary production as reflected by low C/N-ratio and high biogenic silica content produced by planktonic diatoms (Fig. 6).

4.2. Pollen

Pollen and spore percentages of selected taxa are shown in Fig. 3. A more complete pollen diagram with concentration data can be found in Klimaschewski (2011). The pollen diagram is, following the models of zonation and PCA, divided into four pollen assemblage zones (PAZ-1 to PAZ-4) described as follows:

PAZ-1 (862–836 cm, ca. 10,000–9200 cal yrs BP). The zone is dominated by pollen of *A. viridis*, *Betula* (birch) and spores of Filicales (ferns), and contains low percentages of herbs as well as aquatics/marsh plants. Towards the upper part of the zone *Betula* pollen and Filicales spore frequencies decrease slightly whereas *A. viridis* increases and becomes more dominant towards the upper boundary. Poaceae (grasses) and *Artemisia* (mugwort) occur at moderate frequencies, *Salix* (willow) and *Juniperus* (juniper) at low frequencies and *P. pumila* pollen are very rare. The concentration of terrestrial taxa is high at the beginning of sedimentation and declines towards the top of PAZ-1. The rate-of-change is relatively low and the average palynological richness within the zone ranges between 13 and 17 taxa.

PAZ-2 (836–756, ca. 9200–6500 cal yrs BP). Pollen of trees and shrubs become more abundant and there is a clear decline of spores.

Table 1

Radiocarbon determination results from Pechora Lake. Radiocarbon dates were calibrated using the Northern Hemisphere calibration curve IntCal13 (Reimer et al., 2013) and calibrated age ranges at 95% confidence intervals. n.d. = no data.

| Lab code	Composite depth (cm)	Material dated	^{14}C	SD	$\delta^{13}C$	Calibrated age 2σ
UBA-8657	557.5–558.5	Bulk sediment	3461	51	−38.1	3594–3850
UBA-8658	581.5–582.5	Bulk sediment	3856	33	−34.3	4157–4409
UBA-8659	673.5–674.5	Bulk sediment	4481	31	−32.7	4980–5289
UBA-8660	749.5–750.5	Bulk sediment	5517	33	−32.7	6279–6397
UBA-8661	781.5–782.5	Bulk sediment	6539	38	−36.6	7337–7559
UBA-8662	813.5–814.5	Bulk sediment	7741	39	−36.3	8434–8590
LuS 6266	863.5–864.5	Bulk sediment	8960	70	n.d.	9824–10,245
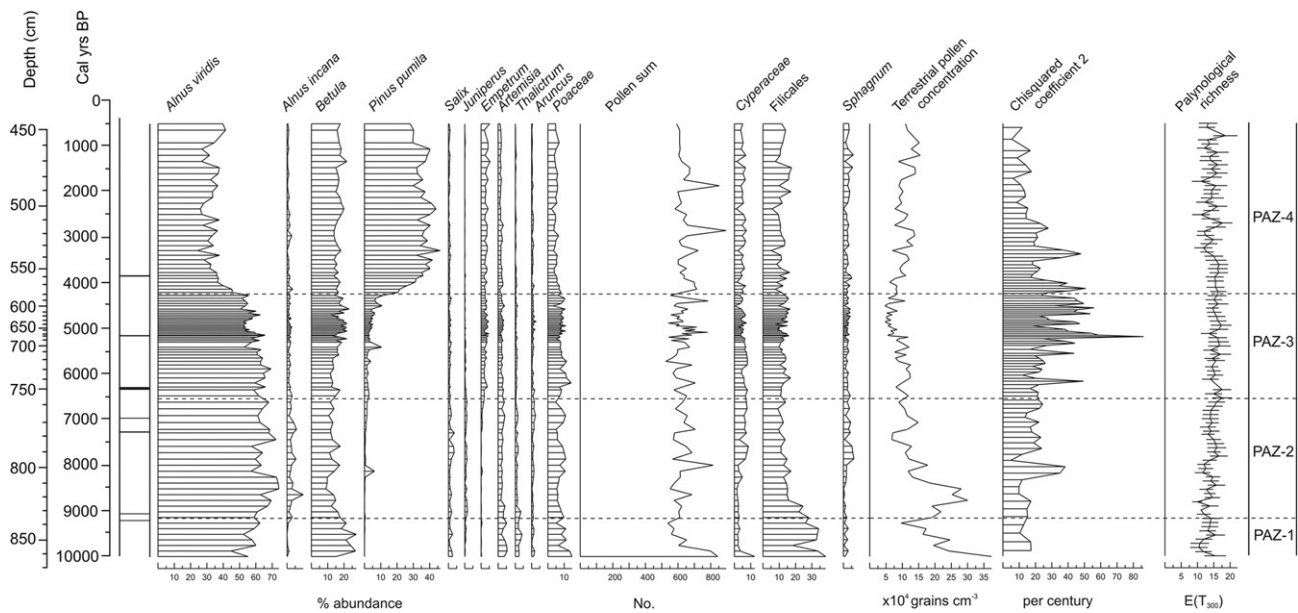
Fig. 3. Relative pollen and spore percentage diagram plotted on a linear age scale (cal yrs BP) and visible tephra layers plotted to the left. Depth scale refers to depth below lake surface in centimetres. Number of counted pollen (pollen sum), terrestrial pollen concentration, rate-of-change (Chi squared coefficient 2), and palynological richness are plotted to the right. The stratigraphy has been divided into four pollen assemblage zones; PAZ-1 to PAZ-4.

A. viridis frequencies are high at the beginning and range between 50 and 60% throughout the zone. *Betula* frequencies remain relatively constant, *Alnus incana*-type shows maximum frequencies and Filicales decreases. The records of *P. pumila* and Ericaceae show low but increasing frequencies near to the upper boundary of zone. Percentages of *Salix*, *Juniperus* and Poaceae remain relatively constant while Cyperaceae (sedges) and *Sphagnum* increase slightly at the middle part of the zone. The average palynological richness is similar to PAZ-1. The notable rise of *A. viridis* clearly shows in the total terrestrial concentration. Towards the upper boundary of PAZ-2 the values decline considerably.

PAZ-3 (756–580 cm, ca. 6500–4200 cal yrs BP). Although this zone exhibits no substantial changes, it is characterised by gently diminishing percentages of *A. viridis*, a slight increase in Ericaceae and an increase in *P. pumila* to 15%. The concentration frequencies of the terrestrial taxa remain relatively constant throughout this zone. In the upper half of PAZ-3 there are some noteworthy peaks in the rate-of-change. The average palynological richness is between 15 and 20 taxa.

PAZ-4 (580–440 cm, ca. 4200–300 cal yrs BP) shows slight increases in the proportions of tree and shrub pollen to the disadvantage of spores. A remarkable increase in the frequency of *P. pumila* from about 15% at the zone boundary to a peak value of 46% at 3300 cal yrs BP was recorded. The frequency of *P. pumila* remains high and a corresponding decrease in *A. viridis* occurs. The total pollen concentration increases slightly throughout the zone. The rate-of-change decreases towards the upper boundary. Palynological richness equals PAZ-3.

4.3. Chironomids and inferred temperature

Chironomid head capsules are well preserved and abundant throughout the core, with 62–177 head capsules counted per sample (Fig. 4). Seventy-two taxa were identified in 59 samples. *Glyptotendipes pallens*-type, *Polypedilum nubeculosum*-type, *Paratanytarsus penicillatus*-type and *Limnophyes* which are generally indicative of relatively warm,

shallow, mesotrophic to eutrophic lakes with abundant aquatic macrophytes are common throughout the sequence. *Demeijerei rufipes* occurs periodically in the sequence at ca. 10,000, 9400 and 5000 cal yrs BP, at 0.8–1.6% abundance. This species feeds on freshwater sponges and bryozoans and although these organisms occur in many freshwater environments *D. rufipes* is rarely recorded as a larval subfossil (Pinder and Reiss, 1983). The chironomid stratigraphy is divided into three assemblage zones (CAZ-1 to CAZ-3) described as follows (Fig. 4):

CAZ-1 (863–675 cm; ca. 10,000–5200 cal yrs BP): The chironomid fauna at the base of the core is initially dominated by *Chironomus anthracinus*-type, Chironomini larvula and *Limnophyes*. *G. pallens*-type, *Tanytarsus mendax*-type and *Zavrelia* increase to peak-abundances between 9500 and 9000 cal yrs BP, and *Cricotopus cylindraceus*-type increases in abundance from ca. 9400 cal yrs BP. These taxa decline from ca. 9000 cal yrs BP and are replaced by increasing abundances of taxa with higher temperature optima such as *P. nubeculosum*-type and *P. penicillatus*-type. *Lauterborniella*, which has one of the warmest optima in the modern east Russian dataset (Nazarova et al., 2015), reaches its peak abundance (3–4%) between 8200 and 7300 cal yrs BP before declining throughout the remainder of the zone. *Sergentia coracina*-type increases in abundance from ca. 8200 cal yrs BP and remains abundant throughout the rest of the zone.

CAZ-2 (675–537 cm; ca. 5200–3400 cal yrs BP): *S. coracina*-type and *P. penicillatus*-type initially decline and are replaced by taxa with cooler temperature optima such as *Tanytarsus lugens*-type and *C. anthracinus*-type.

CAZ-3 (537–445 cm; ca. 3400–450 cal yrs BP): From ca. 4000 cal yrs BP *G. pallens*-type and *P. nubeculosum*-type decrease and are replaced by *Zalutschia zalutschicola*-type together with increasing abundances of taxa indicative of cooler summer temperatures such as *C. anthracinus*-type and *Micropsectra insignilobus*-type. The zone boundary at ca. 3400 cal yrs BP is marked by an increase in the abundance of *Corynocera ambigua* from 1.4% to 10.9%. *Tanytarsus chinyensis*-type which is indicative of warm summer temperatures
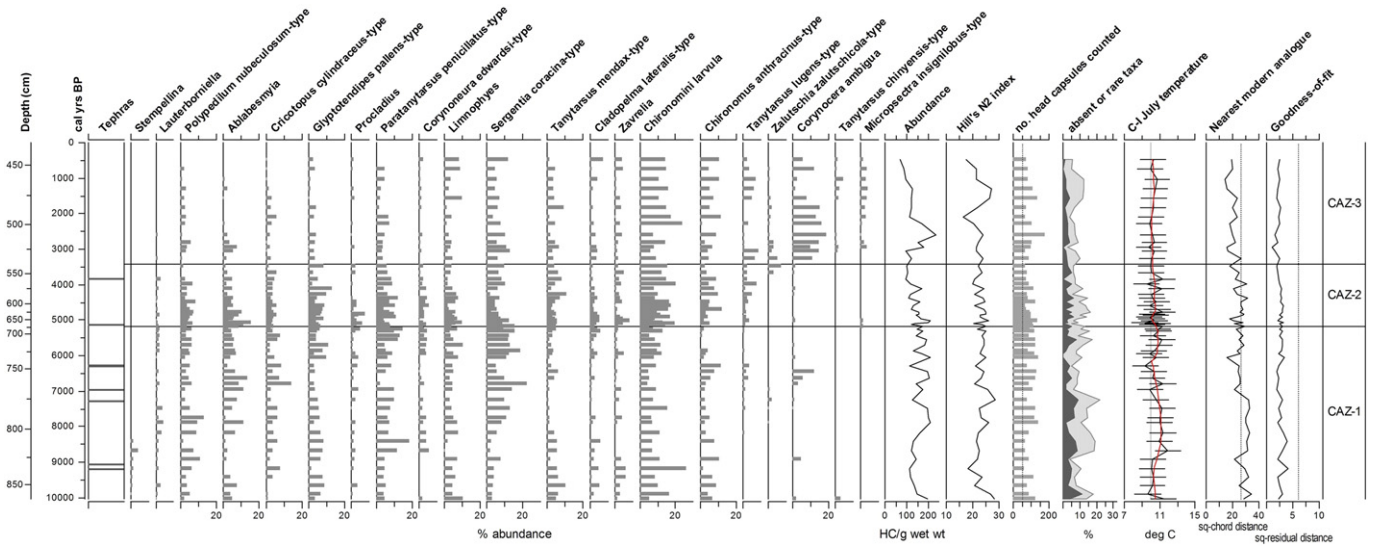
Fig. 4. Chironomid biostratigraphy of selected abundant taxa plotted on a linear age scale (cal yrs BP), with abundance, species richness (Hill's N2 index), total number of head capsules examined and chironomid-inferred mean July air temperatures (C–I T$_{July}$) together with nearest modern analogues for the fossil samples in the training set data, and goodness-of-fit of the fossil samples with temperature. Vertical dashed lines are used to identify samples with no 'good' (5%) modern analogues and samples with 'poor' (0.90) fit with temperature (see text for details). Depth scale refers to depth below lake surface in centimetres. Visible tephra layers are plotted to the left. The stratigraphy has been divided into three chironomid assemblage zones; CAZ-1 to CAZ-3.

in the eastern Russian training set lakes (Nazarova et al., 2015) occurs briefly at ca. 3000 cal yrs BP and between ca. 1600–1000 cal yrs BP at up to 4% abundance.

The chironomid-inferred mean July air temperatures (C–I T$_{July}$) were similar to present-day (ca. 10 °C) between 9900 and 8900 cal yrs BP, before warming by approximately 1.3 °C between 8200 and 7300 cal yrs BP (Fig. 4). From ca. 6700 to 3600 ka C–I T$_{July}$ show a gradually declining trend although individual values fluctuate between 9.3 and 11.3 °C, possibly indicating a period of unstable summer temperatures. Chironomid-inferred July temperatures decline by approximately 0.5 °C from 3800 cal yrs BP, to values similar to present-day.

4.4. Diatoms and inferred pH

Diatom assemblages were enumerated at 69 levels and ca. 700–320 valves were counted at each level. A total of 312 diatom taxa were identified and found to be excellently preserved. The diatom assemblages are dominated by planktonic freshwater taxa within the genus *Aulacoseira* and freshwater periphyton within the genus *Fragilaria* (*Fragilaria* sensu lato, consisting of the genera *Fragilaria*, *Pseudostaurosira*, *Staurosira* and *Staurosirella*). Most taxa are rare and only 22 taxa occur at frequencies exceeding 2% at two or more stratigraphic levels (Fig. 5). There are some clear trends in the diatom stratigraphy, and cluster analysis divides the diagram into six diatom assemblage zones (DAZ-1 to DAZ-6), which are confirmed by visual inspection:

DAZ-1 (862–848 cm, ca. 10,000–9600 cal yrs BP) is dominated by a diverse (maximum species richness) periphytic assemblage with *Staurosira elliptica* in the lowermost part of the zone along with other fragilaroid taxa, such as *Staurosira venter*, *Staurosirella pinnata*, *Fragilaria exigua* and *Pseudostaurosira brevistriata*. There are also maximum frequencies of *Discostella stelligera*, *Stauroneis anceps* var. *gracilis* and *Sellaphora pupula*.

DAZ-2 (848–826 cm, ca. 9600–8900 cal yrs BP) shows a clear shift from a periphytic to a plankton dominated assemblage. The
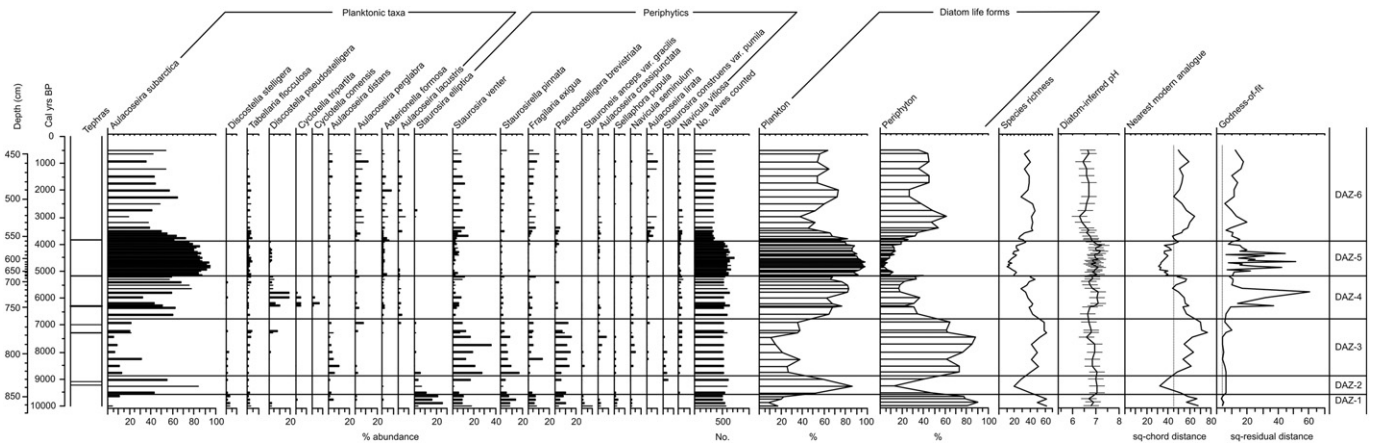

Fig. 5. Diatom stratigraphy of selected taxa in relative percentages plotted on a linear age scale (cal yrs BP) with number of valves counted, summary of diatom life forms, species richness, diatom-inferred pH (D–I pH) together with nearest modern analogues for the fossil samples in the training set data, and goodness-of-fit of the fossil samples with pH. Vertical grey lines are used to identify samples with no 'good' (5%) modern analogues and samples with 'poor' (0.90) fit with pH. Depth scale refers to depth below lake surface in centimetres. Visible tephra layers are plotted to the left. The stratigraphy has been divided into six diatom assemblage zones; DAZ-1 to DAZ-6.

planktonic species *Aulacoseira subarctica* has a peak and reaches ca. 85% simultaneously with a very low abundance, <10%, of *Fragilaria* taxa. The shift results in a very low species richness and a high species turnover, as shown by the DCA axis 1 graph (Fig. 6).

DAZ-3 (826–762 cm, ca. 8900–6800 cal yrs BP) is dominated by periphytic taxa within *Fragilaria* (*S. venter*, *S. pinnata*, *F. exigua* and *P. brevistriata*), and an overall increasing species richness.

DAZ-4 (762–677 cm, ca. 6800–5200 cal yrs BP). *A. subarctica* dominates the zone but planktonic *Cyclotella*/*Discostella* taxa, such as *Discostella pseudostelligera*, *Cyclotella tripartita* and *Cyclotella comensis* exhibit maximum frequencies. Species richness decreases as planktonic taxa reach maximum frequencies in the middle part of the zone.

DAZ-5 (677–560 cm, ca. 5200–3900 cal yrs BP) is totally dominated by *A. subarctica* with over 90% in the middle part and accordingly very low percentages of periphytic taxa as well as low species richness.

DAZ-6 (560–446 cm, ca. 3900–500 cal yrs BP) is still dominated by *A. subarctica* but also by other *Aulacoseira* taxa, most of them planktonic, such as *A. distans*, *A. perglabra*, *A. lacustris* and *A. lirata*. The planktonic species *Asterionella formosa* shows a frequency maximum but also periphytic taxa contribute to a more diverse assemblage with e.g., *S. venter*, *F. exigua* and *P. brevistriata*. In general the species richness is relatively high within the zone.

The diatom-inferred pH (D–I pH) show little change throughout the sediment record fluctuating between a circumneutral pH of 6.3 and 7.2 (Fig. 5). The lower part of the section (DAZ-1 to DAZ-5) show a D–I pH of about 7 whereas the uppermost assemblage zone, DAZ-6 shows a slightly more acidic pH of around 6.6.

4.5. Geochemistry

Biogenic silica (BSi) concentrations range between 3.2 and 17.8 wt.% SiO_2 (Fig. 6). Most concentrations range between 10 and 15 wt.% SiO_2 and are close to the mean (12.1%) for lakes in general (Frings et al.,

2014). The low concentrations of 3.2 to 6.4 wt.% are recorded in connection to visible tephras.

The total organic carbon (TOC) content of the sediments decreases from just below 30% to a low around 18% during the initial millennium, followed by a rise to above 30% during the subsequent ca. 1500 years (Fig. 6). Thereafter, a generally decreasing trend was recorded, stabilising around 18% from ca. 5000 cal yrs BP, but with pronounced lows in the range of 4–9% coinciding with visible tephra horizons. After ca. 3800 cal yrs BP a further decrease to below 15% and slightly more fluctuating TOC values were recorded.

Atomic C/N ratios show variations in the range of 16–21 (Fig. 6). Following an initial decrease, a sequence minimum was recorded at about 9000 cal yrs BP. Distinctly rising trends occurred at ca. 9000–6500 and 3400–1500 cal yrs BP, while relatively low C/N ratios were recorded around 5900 and at ca. 5100–4500 cal yrs BP.

The $\delta^{13}C$ record shows a decreasing trend with some variations in the range of −30.5 to −32‰ during the first two millennia, followed by a persistent rise to above −30‰ at ca. 8400–6500 cal yrs BP (Fig. 6). The remaining part of the record shows minor variations centred around −30‰ but with a general increase during the last four millennia.

5. Interpretations and discussion

5.1. Catchment and lake development and relation to climate change

The overall changes in the diatom and chironomid stratigraphies are not isochronous in Pechora Lake in early Holocene (see assemblage zones in Fig. 6) suggesting that they are not responding to the same external drivers, in opposite of Lifebuoy Lake situated only ca 20 km south of Pechora Lake (Solovieva et al., 2015–in this issue). From about 5000 cal yrs BP however, changes in aquatic ecosystem occur more or less simultaneously. The seemingly non-isochronal development could be a response to the different seasonality of the various proxies. The major changes in diatom composition in Pechora Lake may be characterised by fluctuations between periphyton (mainly in the genera *Fragilaria* sensu lato) and plankton (mainly *Aulacoseira* and *Cyclotella*/*Discostella*) dominated assemblages (Fig. 5). Diatoms,
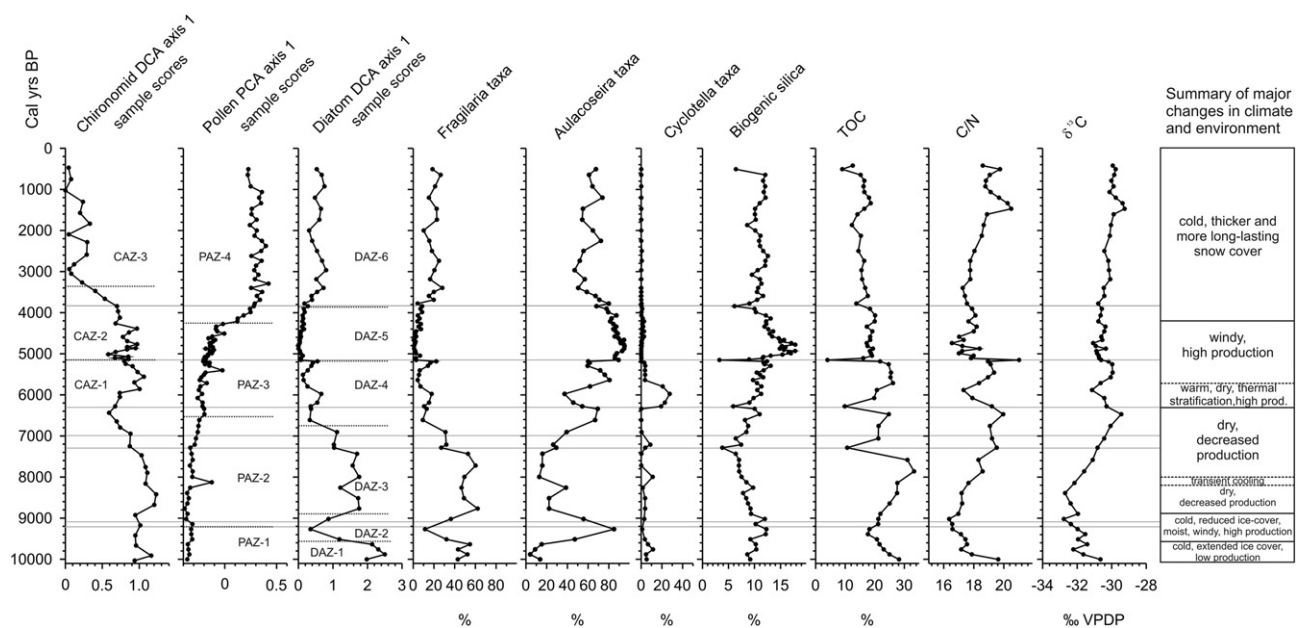

Fig. 6. Chironomid DCA, pollen PCA and diatom DCA axis 1 sample scores divided into assemblage zones (CAZ, PAZ, DAZ) and plotted on a linear age scale (cal yrs BP) together with diatom genera *Fragilaria* (*Fragilaria* sensu lato, consisting of the genera *Fragilaria*, *Pseudostaurosira*, *Staurosira* and *Staurosirella*), *Aulacoseira*, *Cyclotella* (including *Discostella*) and the geochemistry; biogenic silica, total organic carbon (TOC) content, C/N ratio and $\delta^{13}C$. The grey lines represent visible tephra layers. To the right a column showing the summary of major environmental and climate change (production refers to lake primary production).

although not responding directly to fluctuations in water or air temperature, are sensitive to changes in variables controlled by climate as water turbulence, light conditions and nutrient availability (Anderson, 2000). The shifts in diatom life form in Pechora Lake seem to be mainly controlled by indirect climate impact, i.e., ice-cover, windiness and length of growing season which influence available substrates, and correlates with simultaneous changes in aquatic primary production manifested as increased BSi, low C/N-ratio and δ^{13}C (Fig. 6). To clarify the drivers of water chemistry changes in the lake, diatom-inferred pH reconstruction was carried out. Changes in pH may reflect export of dissolved organic carbon from the catchment in response to changes in vegetation or soils (Reuss et al., 2010), or direct and indirect effects of deposition of volcanic ash (Harper et al., 1986). The uppermost reconstructed pH (6.6 about 500 cal yrs BP) is close to the measured pH in the lake (6.7), thus giving certain confidence in the pH reconstruction. But only about a quarter of the fossil samples also have good analogues in the training set (i.e., in DAZ-5 between 674 and 570 cm, about 5200 and 4100 cal yrs BP) with a D–I pH of about 7. However, the results of pH reconstruction from the bottom of the core until 5200 cal yrs BP and from 4100 to 500 cal yrs BP should be interpreted with caution due to the absence of good analogues in the modern training set. The fossil diatom assemblages show a poor fit-to-pH in CCA and the Telford–Birks test also showed that the pH reconstructions were not statistically significant although the test may not provide a robust estimate of reconstruction reliability for the periods with little change in reconstructed variables (Self et al., 2015; Telford and Birks, 2011).

Although the chironomid assemblages show a good fit-to-temperature in CCA, the percentages of taxa absent or rare in the training set are high throughout the sequence with no good analogues present in the modern training set between 8700 and 7200 cal yrs BP and intermittently at other time in the record until 3200 cal yrs BP. Due to the low amplitude of fluctuations in the C–I T$_{July}$ record all the inferred values are within the sample specific errors of prediction of the inference model. Analysis using the Telford and Birks (2011) method also indicated the C–I T$_{July}$ reconstructions were not significant. The Telford–Birks test determines whether the reconstruction explains a larger proportion of the variance in the fossil data than 95% of 999 reconstructions of random environmental variables. If the magnitude of temperature fluctuations is low, the associated variability in the chironomid assemblages may be insignificant in comparison with the natural variability in the training set. Therefore the test may not provide a reliable estimate of reconstruction reliability for time periods with low amplitude climate variability (Self et al., 2015). However, in view of the lack of good analogues, the C–I T$_{July}$ record should be interpreted with caution particularly for the early part of the sequence between 10,000 and 7000 cal yrs BP.

Using the subdivision of the Holocene as proposed by Walker et al. (2012), the development in the Pechora Lake area can be described as follows.

5.1.1. Early Holocene: 10,000–8200 cal yrs BP

As the minerogenic substrate was not reached during the coring, it can only be suggested that sedimentation in the lake was established by about 10,000 cal yrs BP. Our pollen record indicates a relatively stable terrestrial vegetation during the early Holocene, although A. viridis increased slightly at about 10,000–8500 cal yrs BP at the expense mainly of Betula and Filicales (Fig. 3). The diatom record indicates that the lake from the onset of the record had a diverse flora dominated by small benthic Fragilaria species, which are opportunistic and pioneering taxa associated with cold periods and extensive ice-cover (Smol et al., 2005; Rühland et al., 2008). They are also competitive during phases of rapid environmental change (Lotter et al., 2010), and have for example been used as indicators of isolation of lake basins from the sea (Stabell, 1985). Small-celled diatoms have a competitive advantage at low nutrient conditions and taxa within the genus Fragilaria form blooms and out-compete larger forms with slower growth rates

(Lotter et al., 2010). Small Fragilaria often dominate the diatom assemblage in shallow lakes, especially in mountain and tundra regions having short growing seasons (Lotter et al., 2010), and their presence could indicate high availability of shallow benthic habitats (Bennion et al., 2010). The chironomid assemblage was initially dominated by C. anthracinus-type and Chironomini larvulae. The taxonomy of the Chironomini larvula was based on Brooks et al. (2007) which does not differentiate between larvulae from different Chironomini genera. From the core results there is no clear relationship between increases in larvula and any particular Chironomini morphotype. Although the larvula are probably Sergentia or Chironomus species they could be Polypedilum nubeculosum-type or Glyptotendipes pallens-type which are also present in the core and found as early instars in sediment cores. Chironomus have a high oxy-regulatory capacity and are able to tolerate periods of oxygen depletion under ice (Brodersen et al., 2004; Brodersen and Quinlan, 2006). The relatively high TOC content, C/N ratios and δ^{13}C values at this stage may indicate a transition from an initially high supply of detrital organic matter from the surrounding catchment, while the primary production in the lake was relatively low. This is further supported by the occurrence of D. stelligera, which is indicative of low nutrient conditions, while Fragilaria are poor indicators of lake trophic status (Bennion et al., 2010).

Around 9600 cal yrs BP there was a transient shift in the lake lasting for about 700 years indicating less ice-cover and more wind-mixed, turbulent conditions, which resulted in elevated primary production in the lake, especially by diatoms as indicated by an increase in BSi content of the sediments, as well as by decreasing C/N-ratios and δ^{13}C values. Increases in the relative abundances of Glyptotendipes and Cricotopus species at the time may also be indicative of increasing lake productivity (Brodersen and Quinlan, 2006). This development was probably related to increased catchment runoff during relatively moist climatic conditions as indicated by elevated pollen percentages of Filicales and supported by a decreasing trend in δ^{18}O obtained from the same sediment sequence (Hammarlund et al., 2015–in this issue). The increased runoff provided the lake with abundant nutrients in the form of ^{13}C-depleted dissolved inorganic carbon (DIC), which promoted aquatic productivity and led to increased proportions of algal organic matter in the sediments. The dominating diatom life form also indicates a shift towards plankton dominance in response to increased nutrient availability and increased primary production (Bennion et al., 2010). Diatoms benefit from turbulent mixing and their abundance is correlated to changes in the thermal–physical dynamics of the water column (Winder et al., 2009). Species of Aulacoseira are heavily silicified, forming colonies that require turbulence-induced resuspension to remain in the photic zone (Rühland et al., 2008; Lotter et al., 2010). Aulacoseira species are an indirect palaeoenvironmental indicator of the persistence of strong seasonal wind stress and resultant turbulent water column mixing and nutrient upwelling conditions (Wang et al., 2008) and typically dominate in early spring, blooming under the ice (Solovieva et al., 2005). A. subarctica, the overall dominating species in Pechora Lake, occurs in cool and well-mixed water columns. An advantage for the survival of A. subarctica is its ability to form resting cells to tolerate nutrient limitation and periods of hypoxia, and sediments could act as refugia (Lepskaya et al., 2010). A study of a lake in southern Kamchatka revealed that A. subarctica thrived during years when thermal stratification was delayed by cold and windy weather but in the absence of extensive ice-cover (Lepskaya et al., 2010). Although Chironomus spp. remain abundant, increases in taxa with lower oxy-regulatory capacity such as T. mendax-type may indicate a reduction in the duration of ice-cover and hypoxic period (Brodersen and Quinlan, 2006).

The rapid short-term 8.2 ka event was caused by a massive outflow of freshwater from the glacial lakes Agassiz and Ojibway slowing the North Atlantic deep water formation (Barber et al., 1999). There is evidence that this event was global in nature (Walker et al., 2012 and references therein), but it seems to be superimposed on a more long-

term climate deterioration ca. 8500–8000 cal yrs BP which relates to fluctuations in solar input and impact of atmospheric circulation rather than the North Atlantic deep water formation (Rohling and Pälike, 2005). Although manifested by single samples only, our proxy records show some indications of a transient climate response at about 8200 cal yrs BP, visible in the pollen PCA and diatoms DCA axis 1 sample score graphs (Fig. 6) as increased taxa turnover. The pollen data record a single sample peak to c. 6% frequency of *P. pumila* at ca. 8200 cal yrs BP (Fig. 3) indicating a brief period of thicker and more persistent snow-cover, which is consistent with Rohling and Pälike (2005) stating that the 8.2 event is sharp and more evident in winter-biased proxies. In the diatom stratigraphy there are increases in *A. subarctica* and *F. exigua* (Fig. 5) as well as in biogenic silica (Fig. 6). This indicates a slightly elevated primary production in the lake and cold and windy climate conditions. From the chironomids it appears to be part of a longer-term cooling (cf. Rohling and Pälike, 2005) rather than a marked short-term cold spell. The chironomid *P. penicillatus*-type, which is a warm indicator, declines from 18% to 7% and *S. coracina*-type, which is a cold stenotherm, increases from 3% to 6% between ca. 8450 and 8200 cal yrs BP (Fig. 4).

5.1.2. Middle Holocene: 8200 to 4200 cal yrs BP

Previously published pollen records from the Kamchatka Peninsula exhibit vegetation conditions suggesting a continuous period with relatively warm climate assigned to the Holocene thermal maximum (HTM) 8900–5000 cal yrs BP in central areas and between 7000 and 5800 cal yrs BP at coastal sites (Dirksen et al., 2013 and references therein). However, a study of regional glacier fluctuations suggests that the HTM lasted between ca. 6800 cal yrs BP and 4500 cal yrs BP (Barr and Solomina, 2014). None of these studies is based on data from as far north as Pechora Lake but a wider discussion about the HTM in the region is carried out in Brooks et al. (2015–in this issue). In Pechora Lake *A. viridis* and *Betula* dominated the vegetation, while ferns, Cyperaceae and *Sphagnum* were also important components of the local vegetation cover. The frequency of *P. pumila* rose slowly but show persistently low values whereas a continuous decline of *A. viridis* is visible.

Climate warming results in increased thermal lake stratification and reduced nitrogen/phosphorous-ratios, which in turn leads to an altered diatom community structure favouring small-sized, fast-growing plankton, such as *Cyclotella/Discostella* (Winder et al., 2009). *Cyclotella/Discostella* has a competitive advantage during strong stratification compared to heavy and thickly silicified *Aulacoseira* and small *Fragilaria* species (Lotter and Bigler, 2000; Rühland et al., 2008). There is a significant link between an increase in *Cyclotella* species and the 20th century climate warming trend both in Arctic and Alpine settings and at temperate latitudes (Rühland et al., 2008). At Pechora Lake there is a clear peak in *Cyclotella* taxa (including *Discostella*) at ca. 6300–5200 cal yrs BP, reaching almost 30% during the first 500 years, which could be interpreted as a thermally stratified lake and accordingly warmer water temperatures (Fig. 6). In the chironomid record *S. coracina*-type reaches peak frequencies between 7700 and 4900 cal yrs BP. *S. coracina*-type is typically a cold stenotherm (Brodin, 1986). The weighted average (WA) temperature optimum of this taxon in the eastern Russian dataset is 8.8 °C, which is one of the lower temperature optima (Nazarova et al., 2015). Its increasing abundance in parallel with increasing abundances of more thermophilic, littoral taxa may also reflect increased thermal stratification in the lake.

As demonstrated by the generally high C/N ratios, the organic matter content of the sediments is probably dominated by terrestrial organic detritus, although periods of lowered C/N ratios signify temporary increases in aquatic productivity and dilution by lacustrine organic matter. This hypothesis is supported by the negative correlation between BSi content and C/N ratio (Fig. 6). Similarly, increasing $\delta^{13}C$ values, which are also negatively correlated with BSi content (Fig. 6), represent primarily elevated proportions of terrestrial organic matter

in the sediments, possibly in response to increased forest density around the lake. Hence, the trends towards higher C/N ratios and $\delta^{13}C$ values at ca. 9000–6500 cal yrs BP reflect a successively decreased importance of algal production as compared to terrestrial organic matter supply to the lake. This probably occurred in combination with diminished dissolved inorganic carbon (DIC) supply due to decreased runoff in a generally drier climate as compared to the earliest part of the record, which is further supported by the $\delta^{18}O$ record obtained on aquatic cellulose from the same sediment sequence (Hammarlund et al., 2015–in this issue).

The establishment of a slightly warmer and drier climate after ca. 6300 cal yrs BP as indicated by the increase in *Cyclotella* taxa in combination with maxima in C/N ratio and $\delta^{13}C$ is broadly consistent with the onset of the HTM as inferred from regional pollen data (7000–5800 cal yrs BP, Dirksen et al., 2013) and glacier status data (little evidence of advances at 6800–4500 cal yrs BP, Barr and Solomina, 2014). Reconstructed sea surface temperatures show a slight warming trend in the western Bering Sea and a cooling in the NW Pacific and the Sea of Okhotsk during the last 7000 years (Max et al., 2012; Harada et al., 2014). From ca. 6500 cal yrs BP the atmospheric pressure pattern started to approach present-day conditions with a south-westerly mode of the Aleutian Low, resulting in north-easterly winds over the Sea of Okhotsk and a more westerly extension of sea ice (Katsuki et al., 2010).

5.1.3. Late Holocene: 4200 to 400 cal yrs BP

The transition into the Late Holocene at 4200 cal yrs BP is connected to a global climatic reorganisation proposed as the mid/low-latitude aridification event (Walker et al., 2012). The climate became generally drier and cooler, and the Asian monsoon weakened or failed, which coincided with the onset of the modern El Niño Southern Oscillation regime (Fisher et al., 2008; Walker et al., 2012 and references therein). A period of increased sea-ice extent in both the Okhotsk Sea and the Bering Sea at about 4000–3000 cal yrs BP seems to correlate with intensified winter–spring storms in the Gobi Desert and corresponding changes in the intensity and position of the Aleutian Low and the Siberian High (Harada et al., 2014). At Pechora Lake the pollen record indicates a significant change of the vegetation composition around 4200 cal yrs BP dominated by a remarkable increase in the frequency of *P. pumila* (Fig. 3). Probably *P. pumila* expanded across large parts of the area, while *A. viridis* became less abundant. However, birches, grasses, sedges and ferns remained important in the vegetation until the present. *P. pumila* covers huge areas in the Russian Far East from sea level to above the alpine treeline and its northern limit is located approximately at the 12 °C mean July isotherm in Russia (Andreev, 1980) and 10 °C mean August isotherm in Japan (Yanagimachi and Ohmori, 1991). In Kamchatka it occurs on the lower mountain ranges in the north and grows in most areas of the volcanic belts down to Cape Lopatka in the south. *P. pumila* is an important component in the primary succession (Krestov, 2003). It grows in thickets and has a significant ability to withstand extreme environmental conditions such as a short growing seasons, low winter temperatures, wind abrasion and desiccation. With the first frost in autumn the dwarf shrub drops its branches onto the ground and during the winter season they are protected by a deep snow cover (Krestov, 2003; Anderson et al., 2010). The ability to withstand cold conditions may have allowed the survival of *P. pumila* during the last glacial maximum at isolated sites in many areas of its modern distribution range (Kremenetski et al., 1998). *P. pumila* expanded in northern Kamchatka at ca. 4200–4000 cal yrs BP (Pechora Lake, Fig. 3; Lifebuoy Lake, Solovieva et al., 2015–in this issue). Compared to available pollen records from other parts of Kamchatka (e.g., Dirksen et al., 2013 and references therein; Self et al., 2015–in this issue) the expansion of *P. pumila* occurred rapidly at Pechora Lake, from 5 to 30% within ca. 300 years. The expansion phase may be linked to a climatic shift resulting in thicker and more persistent snow-cover, as *P. pumila* is known to have a competitive

advantage over other species in the region under such climatic conditions (Krestov, 2003). This interpretation is supported by stable oxygen isotope data from Pechora Lake where a depletion in ^{18}O is attributed to cooling and increased winter precipitation (Hammarlund et al., 2015–in this issue). The timing of the *P. pumila* expansion coincides with the onset of the Neoglaciation at ca. 4500 cal yrs BP as inferred from the formation of numerous moraines indicating extensive advances of mountain glaciers on the peninsula (Barr and Solomina, 2014). It has been argued for the existence of several small scattered glacial refugia in Kamchatka where *P. pumila* survived the ice age (Khotinsky, 1977) and from which the shrub pine thickets started to spread after the last glacial maximum. *P. pumila* thickets may have been growing in the Pechora Lake catchment throughout the Holocene, either at low abundances or producing only restricted amounts of pollen. It most likely spread from glacial refugia located on the Kamchatka Peninsula rather than from the mainland of Russia, and then spread slowly and possibly erratically across the peninsula, constrained by soils, topography, and perhaps volcanism. However, it is not possible to exclude colonisation of the northern areas of Kamchatka from the mainland of Russia via Priokhot'ye or Chukotka. After the *P. pumila* expansion the compositional turnover of vegetation in the catchment of Pechora Lake became low towards the present.

DCA axis 1 sample scores for the chironomid assemblages decline rapidly after 3900 cal yrs BP indicating a period of high species turnover (Fig. 6) as taxa indicative of warm, productive lakes decline and are replaced by species typical of cooler summer temperatures.

There is a gradual transition between the two uppermost diatom assemblage zones (DAZ-5 and 6), where periphytic taxa, especially in the genus *Fragilaria*, increase on expense of the planktonic taxa (Figs. 5–6). This change, could indicate colder conditions and an increase in winter ice-cover (Smol et al., 2005) and coincides with the significant change in vegetation at about 4200 cal yrs BP attributed to a colder and more snow-rich climate (Hammarlund et al., 2015–in this issue). The general decrease in TOC content and the relatively low BSi content of the sediments indicate lowered aquatic productivity, most likely in response to colder and/or shorter ice-free seasons (Douglas and Smol, 2010) during the Late Holocene, although indirect effects of the *P. pumila* expansion on the lake ecosystem, such as acidification, cannot be excluded. The diatom-inferred pH reconstruction (Fig. 5), although interpreted with caution due to lack of analogues, show a slightly more acidic pH with a decrease from about 7 to 6.6 which could indicate acidification of the lake. The successive increases in C/N ratio and δ^{13}C represent direct effects of increased incorporation of terrestrial organic matter in the sediments, initially as a result of

increased forest density around the lake, and perhaps more recently in response to soil degradation.

5.2. Environmental impacts of tephra deposition events

There does not seem to be any prominent influences of volcanic eruptions on the vegetation in the catchment of Pechora Lake (Fig. 3). However, a minor increase in Poaceae was recorded above one of the tephra horizons (Table 2), which might indicate short-lived vegetational impacts to ashfall, but the resolution of our data does not allow any detailed analysis. Urrutia et al (2007) found considerable increases in Poaceae pollen frequencies in response to tephra deposition, which can be explained by the short life cycle of grasses and their rapid profiting from nutrients supplied by volcanic eruptions. A pollen and macrofossil study of peat sequences with numerous tephra layers in the Kamchatka River lowlands recorded only slight influences at the most of ashfalls on the local vegetation (Andreev and Pevzner, 2001). The original vegetation probably recovered quickly during the subsequent succession as described, for example, by Hotes et al. (2004).

Diatom communities commonly respond abruptly to ashfall, both in terms of abundance and diversity (e.g., Barker et al., 2003). Responses to tephras seem to be stronger among diatoms compared to other palaeolimnological proxies, possibly due to their short life cycles, which allow them to respond quickly to environmental change (Urrutia et al., 2007). Tephra layers may also lead to enhanced preservation of diatom frustules (Harper et al., 1986). In addition, macrophyte growth is inhibited by ashfall as plants become coated with silt, which reduces the area of plant habitat available to epiphytic diatoms (Harper et al., 1986). Planktonic and epipelic (growing on the sediment surface) diatoms show the most rapid responses to tephra input but the proportion of epiphytic species (growing on plants or other algae) decreases, which suggests that their habitats of submerged waterweeds can be partly destroyed by ash deposition (Harper et al., 1986). Laboratory experiments show that epipelic diatoms buried by a 5 cm ash layer migrated up to the surface and are thus likely to be less affected by tephras than epiphytic diatoms (Harper et al., 1986). In the Pechora Lake diatom record the responses to tephra deposition are manifested as occasionally recorded shifts in diatom species composition and generally increased production (Table 2). The two uppermost tephras correspond to diatom assemblage zone boundaries and thus significant changes in species composition (Fig. 6). The most prominent shift in diatom assemblages following a tephra input is the sharp transition from DAZ-4 to DAZ-5 at ca. 5150 cal yrs BP (Table 2), which resulted in diatom life forms completely dominated by plankton and a high

Table 2
Visible tephras in the Pechora Lake stratigraphy and possible correlation to recorded environmental change and impact on different proxies.

Depth (cm)	Tephra age from age model (cal yrs BP)	Thickness (cm)	Response to tephra deposition events recorded by different proxies
557–559	3844	2	Diatoms: coincides with a diatom assemblage zone boundary (DAZ-5 and 6) but is a more gradual change, a succession. No visible response in elemental geochemistry except a slight increase in BSi.
674–676	5154	2	Diatoms and geochemistry: coincides with a diatom assemblage zone boundary (DAZ-4 and 5). A sharp transition with reduced species richness and increase in plankton/*Aulacoseira* taxa and decrease in periphyton/*Fragilaria* taxa. Increase in diatom production (maximum BSi, low C/N-ratio, low δ^{13}C). Chironomids: high abundances of *Ablabesmyia* and Chironomini larvulae. Delay in increases in the abundance of later instar *Chironomus*. Both the concentration of head capsules and species richness decrease.
746–749	6311	3	Diatoms and geochemistry: increase in plankton, especially *Cyclotella* taxa. Decrease in C/N ratio and δ^{13}C indicating increased production. Chironomids: Both the concentration of head capsules and species richness decrease. Pollen: possible transient increase in Poaceae.
768–769	6984	1	Diatoms and geochemistry: very slight increase in BSi indicating increased diatom production. No apparent responses in C/N ratio and δ^{13}C. Chironomids: Peak in the abundance of *Corynocera ambigua*. Both the concentration of head capsules and species richness decrease.
776–778	7287	2	Diatoms: very slight increase in BSi indicating increased production. No apparent responses in C/N ratio and δ^{13}C.
832–833	9083	1	No sample between the tephras so change not possible to separate from tephra below
836–837	9207	1	Diatoms: increase in *Fragilaria* and decrease in *Aulacoseira* taxa but this is probably part of a natural succession. Possible increased primary production as recorded by lowered C/N ratio and δ^{13}C but not evident in the BSi record. Chironomids: Peak in the abundance of *Corynocera ambigua*, very short duration.

production (Fig. 5). Similar sharp transitions occur elsewhere in the diatom stratigraphy in the absence of tephras, e.g., at the two lowermost zone boundaries (Fig. 5). Furthermore, diatom-inferred pH showed unexpectedly none or only very slight change following tephra deposition events, but we need to take into account the varying quality of the reconstruction and the resolution in the data. Not all tephras resulted in diatom assemblage changes, probably as the chemical composition of the tephras determined whether they acted as fertilisers of the lake ecosystem, and as epiphytic taxa only constituted a marginal proportion of the assemblage. In a palaeolimnological study of Lifebuoy Lake situated ca 20 km south of Pechora Lake more substantial impact of tephra input is visible affecting the trophic state of the lake, occasionally acting as fertiliser and the opposite when an ash-layer sealed nutrients in the sediments (Solovieva et al., 2015–in this issue).

In the chironomid stratigraphy (Fig. 4) peaks in the frequency of *C. ambigua* at 8900 and 6800 cal yrs BP occur shortly after tephra deposition events and may indicate short-lived perturbations of the lake ecosystem. However, the highest frequency of *C. ambigua* at ca. 3700 cal yrs BP postdates the deposition of tephra by approximately 500 years and was probably not related to tephra deposition. The lowermost chironomid zone boundary coincides with a tephra layer and was initially followed by high frequencies of *Ablabesmyia* and Chironomini larvulae (Fig. 4). Increases in the frequency of later instars of *Chironomus* are delayed and less prominent, which may indicate high early instar mortality. This could be due to environmental change, such as lowered dissolved oxygen concentrations or increased minerogenic input to the lake. Early instars are also more sensitive to heavy metals than later instars (Timmermans et al., 1992) so the former may be more severely impacted by heavy metals derived from tephra deposition. Urrutia et al. (2007) found generally decreasing concentrations of head capsules just after tephra deposition events but no related changes in chironomid diversity. In Pechora Lake both concentration of head capsules and chironomid species richness decrease in the sample following deposition of the tephras at about 5150, 6300 and 7000 cal yrs BP (Fig. 4). The tephra at 6300 cal yrs BP is the thickest in the sequence, and no change is apparent in concentration of head capsules or diversity following deposition of thinner tephras at 3800, 7000, 9000 and 9200 cal yrs BP.

The effects of tephra deposition on lake ecosystems seem to be transitory and once the tephra input ends, the ecosystem re-establishes, and the conditions that prevailed prior to the ashfall commonly return (Urrutia et al., 2007). Responses of diatom communities to thin tephra deposits appear to be more short-lived, about 150–200 years, and the most long-lasting impacts occur when tephras make significant contributions to the catchment silica pool and/or disrupt the internal recycling of phosphorous (Telford et al., 2004). Harper et al. (1986) found that above ash layers the absolute abundance of diatoms increases, with greater effects of thicker tephras, sometimes lasting up to 300 years. This seem to be consistent with the results from Pechora Lake where the most visible tephra impact are recorded after the 3-cm thick tephra at about 6300 cal yrs BP and the 2-cm tephra at 5150 cal yrs BP (Table 2). At Pechora Lake the duration of the responses to tephra deposition is short, a few hundred years at the most, and the low resolution of our biological proxy records precludes definite separation of tephra-related responses from the simultaneous overprint by climate change.

6. Conclusions

Climate change is the major driving force of the recorded environmental changes at Pechora Lake. Some short-lasting perturbations of the aquatic ecosystem can be attributed to tephra deposition events, however, recurrent ashfall did not seem to affect the long-term lake evolution, which is in agreement with previous studies (Barker et al., 2003; Telford et al., 2004). Our interpretation of the climatic development in north-eastern Kamchatka based on biological and geochemical

proxy records obtained from Pechora Lake can be summarised as follows:

- During the first 400 years of the 10,000-year sediment record the catchment was characterised by poor soils and a vegetation dominated by shrub alder and birches. The climate was relatively cold and the lake exhibited extensive ice-cover during winter and a relatively low primary production.
- At ca. 9600–8900 cal yrs BP the climate remained cold but moist, with reduced ice-cover and strong seasonal wind stress, which resulted in turbulent mixing, increased nutrient availability and enhanced primary production.
- After 8900 the forest density increased around the lake, runoff decreased in a generally drier climate resulting in decreased primary production in the lake until ca. 7000 cal yrs BP.
- A short-term perturbation, which may be attributed to the 8.2 ka event, was recorded at ca. 8200 cal yrs BP, with indications of relatively windy climatic conditions, increased snow cover, decreased duration of ice-cover and slightly elevated primary production in the lake. Also individual chironomid taxa support a slightly cooling.
- The diatom record indicates strong thermal stratification about 6300–5800 cal yrs BP. The diatom and geochemical proxy records indicate a dry and slightly warmer climate which could be assigned to the Holocene thermal maximum, resulting in a thermal stratified lake with high primary production.
- At about 4200 cal yrs BP a notable shift in the catchment vegetation was recorded in Pechora Lake as an increase in Siberian dwarf pine, *P. pumila*, indicating a shift to a cooler climate with a thicker and more long-lasting snow cover. The change in the catchment occurred simultaneously with corresponding shifts in diatom and chironomid assemblages also indicative of colder climate and more extensive ice-cover.

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