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ORIGINAL ARTICLE

A 850-year record climate and vegetation changes in East Siberia (Russia), inferred from geochemical and biological proxies of lake sediments

A. P. Fedotov · V. A. Trunova · I. V. Enushchenko · S. S. Vorobyeva · O. G. Stepanova · S. K. Petrovskii · M. S. Melgunov · V. V. Zvereva · S. M. Krapivina · T. O. Zheleznyakova

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Abstract East Siberia is very sensitive to moisture regimes because it is located in a margin area, where moisture from the North Atlantic is strongly depleted, and the penetration of the East Asian monsoon is weak and rare. In winter months, the Siberian anticyclone strongly blocks and reduces the external influences on the region. The high latitude of the study area at 52°N is probably sensitive to variation in insolation and solar activity. We analysed a 42-cm-long sediment record from Lake Mountain located in East Siberia (Russia) for geochemical, mineralogical, subfossil diatoms, chironomids, and pollen to provide a reconstruction of the climate history of the area for the last 850 years. According to our reconstruction, a clear decrease in summer temperatures occurred in East Siberia after ca. 1400 and we linked this temperature drop with the beginning of the Little Ice Age. The coldest summer occurred about ca. 1570–1700 and 1830–1900. We assumed that the most significant changes of the lake bio-productivity and the catchment area were occurred about ca. 1160–1350, 1350–1590, 1590–1730, 1730–1900 and 1940 to the present. The most dramatic period with unfavorable climate conditions for lake-biota was during 1590–1730.

Keywords Paleoclimate · LIA · Diatoms · Pollen Chironomods · East Siberia

Introduction

Lakes respond sensitively to environmental change and these changes are often reflected in the physical, geochemical and biological features of the sediments accumulating in lakes. Hence, sedimentary research plays a key role in deciphering the effects of increasing environmental pressure on aquatic ecosystems as sediments collect and unify environmental signals of both local and regional change (e.g. Cohen 2003; Battarbee and Bennion 2012). Understanding the variation in lake ecosystems, be it as a result of natural or anthropogenic causes, is of utmost importance if we are to fully comprehend how these ecosystems function and how they can be restored and protected (Smol 2002; Sayer et al. 2010). Many studies have focused on the Late Holocene climatic changes in Europe, Fennoscandia and Canada (Jones et al. 2009 and their references). However, the information on fluctuations of East Siberian climate during the Late Holocene and in particular the last few centuries is still scarce.

During the last 20 years attempts to understand climate patterns of East Siberia during the Pleistocene and Holocene have greatly profited from studies of the sediments of Lake Baikal (BDP Group 2000; BDP-99 Baikal Drilling Project Members 2005). The diatomaceous sediments represent warm interglacial climatic conditions with rich and diverse planktonic/benthic diatom floras, chrysophyte algae, sponges and pollen, whereas the diatom-barren clay represents cold glacial environments devoid of the major primary producers (diatoms, chrysophytes), sponges and pollen (Grachev et al. 1997). The majority of paleoclimate

A. P. Fedotov (✉) · I. V. Enushchenko · S. S. Vorobyeva · O. G. Stepanova · S. K. Petrovskii · S. M. Krapivina · T. O. Zheleznyakova
Limnological Institute of the Siberian Branch of RAS, Ulan-Batorskaya st., 3, Irkutsk 664033, Russia
e-mail: mix@lin.irk.ru

V. A. Trunova · V. V. Zvereva
Nikolaev Institute of Inorganic Chemistry of the Siberian Branch of RAS, Acad. Lavrentiev av., 3, Novosibirsk 630090, Russia

M. S. Melgunov
Institute of Geology and Mineralogy of the Siberian Branch of RAS, Acad. Koptug av., 3, Novosibirsk 630090, Russia

work using Lake Baikal sediments has consisted of single-proxy studies, focusing on e.g. only diatoms, or biogenic silica (e.g. Grachev et al. 1997; Mackay et al. 2000; Khursevich et al. 2001; Battarbee et al. 2005) or pollen (e.g. Kataoka et al. 2003; Demske et al. 2005).

The most detailed records of the Holocene from East Siberia (Lake Baikal area) are generally derived from peat-bog cores, but resolution of the records were from 100 to 250 years per sample (Takahara et al. 2000; Bezrukova et al. 2005, 2013; Shichi et al. 2009). Records from shallow lakes located near Lake Baikal show that the salinity and total alkalinity of lake waters depended on the climatic cycles and water level fluctuations (Tarasov et al. 1994; Sklyarov et al. 2010; Fedotov et al. 2013). The oxygen isotope records of biogenic silica and carbonates have been demonstrated the potential for quantitative reconstruction of past climate changes in the Lake Baikal region (Morley et al. 2004; Kalmychkov et al. 2007; Mackay et al. 2008, 2011; Kostrova et al. 2013). Chironomid-based temperature reconstruction has become a widely used approach for reconstructing past temperature variability (e.g. Brooks 2006; Walker and Cwynar 2006). However, these reconstruction have been performed for northern and arctic parts of East Siberia (Nazarova et al. 2008, 2011, 2013), while quantitatively reconstruct Holocene temperature changes for south part East Siberia remains uncertain (Mackay et al. 2012), and our investigations were intended to fill this gap of East Siberia.

High resolution geochemical proxies from bottom sediments have been widely used as a method for paleoclimate reconstructions (Meyers 1997; Wick et al. 2003; Hall et al. 2004). Boës et al. (2005) constructed grey scale measurements from Lake Baikal thin-sections, thereby deriving very high resolution records of about 20 μm per measurement. Increases in grey scale values were related to concentrations of diatoms in the sediments, and suggested to be indicative of periods of warmer prevailing climate in the Lake Baikal region. Using high-resolution scanning (step scanning up to 1 mm) X-ray fluorescence analysis with synchrotron radiation investigated the distribution of elements in the sediments and reconstructed Holocene climate changes in Lake Baikal area (Goldberg et al. 2005; Phedorin et al. 2007; Fedotov et al. 2013).

The region is very sensitive to moisture regimes because it is located in a margin area, where moisture from the North Atlantic is strongly depleted, and the penetration of the East Asian monsoon is weak and rare. In winter months, the Siberian anticyclone (SH) strongly blocks and reduces the external influences on the region (Kuznetsova 1978; Ding 1990). In addition, the high latitude of the study area at 52°N is probably sensitive to variation in insolation and solar activity. Therefore, even minor shifts of the global climate may cause drastic climate changes within the study area.

In the present study, we used lake sediment sequences to investigate changes in the landscape of East Siberia (Russia) during the past 850 years. Records of this period bear critical information about significant climate changes, including the transition from the Little Ice Age (LIA) to the Recent Warming (RW) and the beginning of anthropogenic global warming. We analyzed pollen, diatom, chironomid, geochemical and mineralogical records from Lake Mountain, situated near Lake Baikal and Lake Khovsgol (Northern Mongolia). We interpret these records in terms of the changing regional temperature, precipitation, soil water-saturation, vegetation and lake bio-productivity. We then discuss the effects of the changing climate on the hydrology of the lake and the landscape.

Regional setting

Lake Mountain is situated in the western part of East Siberia (Russia) at the East Sayan Ridge, approximately 200 km to the south shore of Lake Baikal (Fig. 1). Lake Mountain (51°56′N, 100°45′E) is a small freshwater lake situated at 2,098 m above sea level and is approximately 0.14 km^2. The climate in the Eastern Sayan region is continental, as reflected in the large differences of temperature (Fig. 1). According to monthly precipitation and temperature data representing the 1900–2010 interval, the climate is characterized by the mean January −28 °C and mean July 12 °C temperatures (Fig. 1). The annual precipitation ranges from 220 to 380 mm, with the precipitation largely (60–75 %) accumulating during the summer months (NOAA data-set ftp://ftp.ncdc.noaa.gov/pub/data and grid model from http://climate.geog.udel.edu/).

An earlier climate investigation of the Holocene sedimentary sequences from lakes located near to Lake Mountain was based on pollen, algae, chironomid and isotope geochemistry analyses (Mackay et al. 2012).

Methods

Sampling

A 42.5-cm-long sediment record (LM-01/10) was taken from the central part of Lake Mountain at 14 m water depth using the Uwitec Corer sampler in July 2010.

Lithology

Water content (WC) was determined by weighing wet sediment and the residue after drying at 60 °C, and the sampling resolution was 1 cm. The pH of unbroken

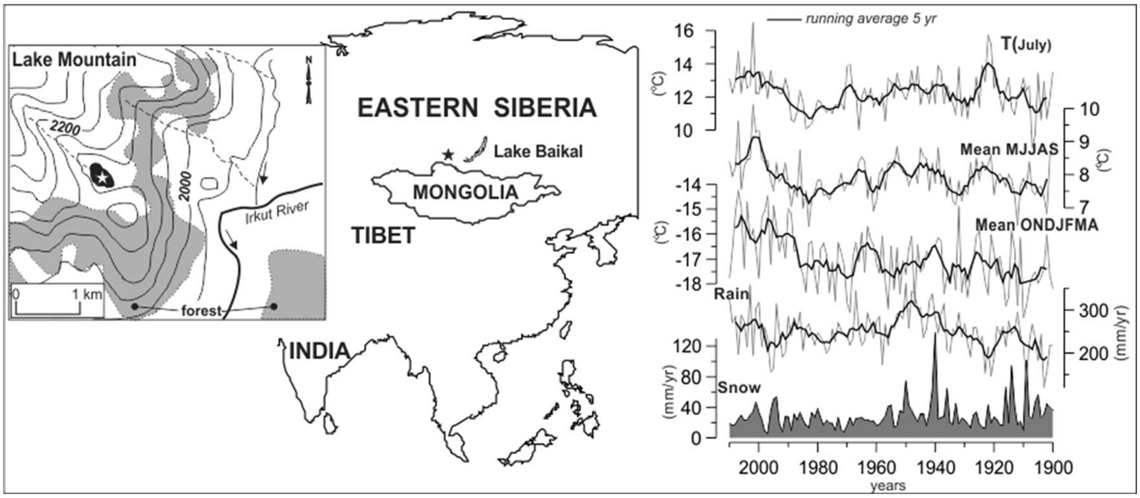


Fig. 1 The locations of Lake Mountain (*asterisk*) and regional climate parameters (Terrestrial Air Temperature: 1900–2008 Gridded Monthly Time Series, Version 2.01, http://climate.geog.udel.edu). $T_{(July)}$ July air temperatures, *Mean MJJAS* the mean value of positive temperatures from May to September, *Mean ONDJFMA* the mean value of negative temperatures from October to April, *Rain* liquid precipitation, *Snow* solid precipitation

sediments along cores was measured with a Testo 206-pH2 pH meter from Testo® after opening the cores.

Dating

Dating of the LM-01/10 core was based on ^{210}Pb and ^{137}Cs chronology. Measurement of ^{238}U, ^{234}Th, ^{226}Ra, ^{137}Cs and ^{210}Pb content in the studied samples was carried out using a high-resolution semiconductor gamma-spectrometry technique. We calculated the depth-age relationship for the uppermost 15 cm using the constant rate of supply (CRS) model (Binford 1990). ^{210}Pb was found in sediment minerals first in a "supported" form by in situ production from ^{238}U or ^{226}Ra decay, at equal activity to its parent nuclides, which is normally assumed to be constant down core. The second fraction, the "unsupported" or excess ^{210}Pb (^{210}Pb$_{excess}$), was of atmospheric origin. Atmospheric ^{210}Pb entered the aquatic environments by direct deposition with rain and also by runoff and erosion from catchment areas. Excess ^{210}Pb concentrations were calculated by ^{226}Ra (Fig. 2).

The elemental composition

The elemental composition of the core was investigated by two methods: X-ray fluorescence spectrometry (XRF-SR) and inductively coupled plasma mass spectrometry (ICP-MS).

X-ray fluorescence spectrometry was undertaken to provide quantitative information on 20 trace elements (from K to U). The scanning analysis was performed at the XRF beamline of the VEPP-3 storage ring (Budker

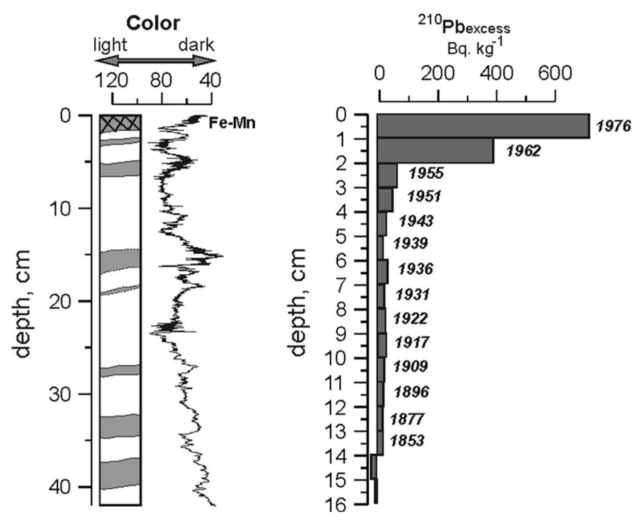


Fig. 2 Color composition (Strati-Signal 1.0.5, Ndiaye et al. 2012) and the depth-age model of the core based on radioactive isotopes ^{210}Pb. *Gray areas* more dark silty clay layers, *Fe–Mn* the upper oxidised layer

Institute of Nuclear Physics, Novosibirsk, Russia). The samples were 22 cm long, 1.2 cm wide and 0.6 cm thick slices of wet sediment placed in aluminium trays. The trays were moved vertically under a 1 mm and SR beam collimated by a cross-slit system (Zolotarev et al. 2001). The spectral responses of the absolute concentrations of elements from their fluorescent peaks were processed using the algorithm of fundamental parameters to overcome problems caused by water content variability in nearly saturated wet sediments (Phedorin and Goldberg 2005). The intensity of each element was normalized within the range between zero (minimal content) and one (maximal content).

ICP-MS was performed essentially as described by Zhuchenko et al. (2008).

Pollen analysis

For pollen analysis, samples of 1 cm^3 were processed with a standard pre-treatment technique, including 10 %-KOH, 30 %-HF solution and heavy-liquid separation without using acetolysis (Anderson et al. 1993; Grachev et al. 1997). Before treatment, a definite amount of polystyrene microspheres micro-spheres (Ogden 1986) was added to the samples for the calculation of the pollen concentration and influx. The extracted pollen and spores were mounted in glycerine oil.

Pollen grains were identified using keys, atlases and a reference collection (Dzyuba 2005; Moore et al. 1991) with a sampling resolution of 1 cm. For each sample, a minimum of 400 terrestrial pollen grains was counted. Identifications were made at 400× magnification.

Chironomid analysis

Samples of 1 cm^3 used for chironomid analysis were immersed in concentrate HF; 24 h later the acid washed out and then the samples washed through a 100-µm sieve with a sampling resolution of 1 cm. The remains of chironomid head capsules were identified by following the method of Pankratova (1970, 1977, 1983) and Makarchenko (1982, 2006). The distribution of modern chironomid assemblages in the Baikal area was taken from Linevich (1981), Erbaeva and Safronov (2009).

Diatom analysis

Siliceous microfossils were quantitatively determined by counting permanent smear slides prepared according to the method described in Grachev et al. (1997). Diatom frustules (from 400 to 1,100 frustules per sample) were identified using keys, atlases and a reference collection (Makarova 1992; Zabelina et al. 1951).

The distribution of organic matter, biogenic silica, quartz and feldspar

The content of the total organic carbon (TOC), biogenic silica (BiSi), quartz and feldspar were investigated using the Fourier-transform infrared (FTIR) technique with KBr (3 mg sample/170 mg KBr) at wave numbers from 700 to 4,000 cm^{-1}. Absorbance bands approximately 2,925–2,856 cm^{-1} were used to calculate the ratio of the TOC in the lake sediments (Liu et al. 2013). The BiSi content was calculated using an absorbance band of 770–850 cm^{-1} (Chester and Elderfield 1968), which the intensity of signal was estimated using the original technique (Stolpovskaya et al. 2006). The quartz content was calculated using the absorbance band at ca. 695 cm^{-1} (Stolpovskaya et al. 2006; Chester and Green 1968). The absorbance band at ca. 645 cm^{-1} (Stolpovskaya et al. 2006; Hecker et al. 2010) was used for calculation of total feldspar content. FTIR analysis was performed with a sampling resolution of 1 cm.

Statistical analyses

Statistical analyses were performed using a principal component analysis (PCA) then redistributed using the varimax rotation method and CONISS analysis using a square-root transformation by the R package rioja (Juggins 2012). Quantitative transfer functions of July air temperature based on chironomid analyses were developed using weighted-averaging (WA) and weighted-averaging partial-least-squares (WA-PLS) calibration techniques. Transfer functions were developed in C2 version 1.7.4 (Juggins 2007).

Results

Lithology

The LM-01/10 core was composed of fine, light brown silty clay, an average WC of 60–70 and a pH of 5.9–6.3. The top of the core (0–1.5 cm) was well oxidised. The bottom part (up to 25 cm) of the core was more enriched with dark layers than the upper part (Fig. 2). Clay silt lamination was more pronounced in the bottom part of the core. There are any lithological evidences about changes in lake level. Based on these lithologic features, we assumed the interval from 0 to 42.5 cm below sediments surface (bss) was formed under lake conditions.

Dating

Excess ^{210}Pb concentrations were negative below depth in 14 cm (Fig. 2). According to the CRS model, the upper part (0–14 cm) of the core was formed after 1853. There were no markers of a hiatus or a change of sedimentation type below 14 cm. If the calculated sediment accumulation ration for the last seven dated points is extrapolated to the layer of 42 cm, in this cause, the layer in 42 cm bss was formed approximately at 1160.

Diatom record

The diatom assemblages of the LM-01/10 core consist of the genera: Aulacoseira (*A. subarctica*, *A. ambigua*, *A.*

lirata, *A. valida*, *A.* sp.1, *A.* sp.2); Cyclotella (*C. tripartita*, *C. ocellata*, *C.* sp.); Pliocaenicus (*P. costatus*); Stephanodiscus (*S. meyeri*, *S. hantzschii*); Tabellaria (*T. flocculosa*, *T. fenestrata*); Ellerbeckia (*E. arenaria* var. *teres*); Orthoseira (*O. dendroteres*). A smooth transition in the characters of the frustules of *C. tripartita* and *C. ocellata* leads to intermediate forms. These forms can not be clearly identified and we combined *C. tripartita* and *C. ocellata* to *Cyclotella*-complex.

Cyclotella-complex and *A. subarctica*, with contents up to 148×10^6 and 23×10^6 frustules g^{-1} dry weight, respectively, was completely dominant over the entire record (Fig. 3). The abundance of other plankton diatom species were $<1 \times 10^6$ frustules g^{-1} except *P. costatus*. More then 50 genera of benthic diatoms was found, when genera contents substantially from one to four species. However, Eunotia and Pinnularia content 13 and 7 species, respectively. Benthic diatoms were evenly distributed along the core, and the mean amount was about 2.8×10^6 frustules g^{-1} dry weight (Fig. 3). Two principle component of the diatom date, PCA-1 and PCA-2 which accounted for 20 and 15 % of the total variance, respectively. The PCA-1 had a large loading for *Cyclotella*-complex and *P. costatus*. The PCA-2 had a large loading for *A. subarctica*.

We divided the diatom records into five local zones (D 1–5). Diatom zone D-1(42–33 cm bss) was formed around ca. 1160–1370. This zone is characterised by low content of plankton and benthic diatoms, only were a high content up to 30×10^6 frustules g^{-1} dry weight of *P. costatus*, as the content of *P. costatus* was up to 0.01×10^6 frustules g^{-1} dry weight in other diatom zones.

A sharp increase in the content of *A. subarctica*, *Cyclotella*-complex was detected in diatom zone D-2 (33–25 cm bss), the abundance of benthic diatoms was also high there. Diatom zone D-3 was formed around 25–11 cm bss. This zone had a highly contrasting pattern of the distribution of diatoms, it was divided into three sub-zone: D-3.1 (25–19 cm bss), D-3.2 (19–16 cm bss) and D-3.3 (16–11 cm bss). A high amount of *Cyclotella*-complex and benthic diatoms and a low amount of the genera Aulacoseira were in D-3.2, as contrast to sub-zones D-3.1 and D-3.3. Diatom frustules of Aulacoseira and *Cyclotella*-complex were abundant in sediment of zone D-4 (11–6 cm bss, 1910–1939). Zone D-5 is characterised by a high amount of *Cyclotella*-complex but the content of *A. subarctica* were decreased.

Chironomid record

A total of 18 taxa of the subfamilies: *Diamesinae*, *Tanypodinae*, *Orthocladiinae* and *Chrinominae* were identified. The most widespread taxa were *Procladius* gr. *choreus, Heterotrissocladius* gr. *marcidus*, *Sergentia baueri*, *Tanytarsus* conf. *medius* and *Stempellina bausei*-type (Fig. 4). Percentages of other chironomid taxa (*Protanypus* spp., *Corynineura arctica*-type, *Orthocladius* sp., *Psectrocladius* gr. *sordidellus*, *Smittia* sp., *Synortocladius* sp., *Thienemanniella* gr. *clavicornis*, *Vivacriocotopus* sp., *Zalutschia* sp., *Chironomus plumosus*-type, *Paracladopelma camptolabis*-type, *Micropsectra* sp., *Shangomyia* sp.) often comprised less than 6 %. Two principle component of the chironomid date, PCA-1 and PCA-2 which accounted for 41 and 22 % of the total variance, respectively. The PCA-1 had a large loading for *H.*gr. *marcidus* and *T.* conf. *medius* with a high negative correlation. The PCA-2 had a large loading for *Sergentia baueri*.

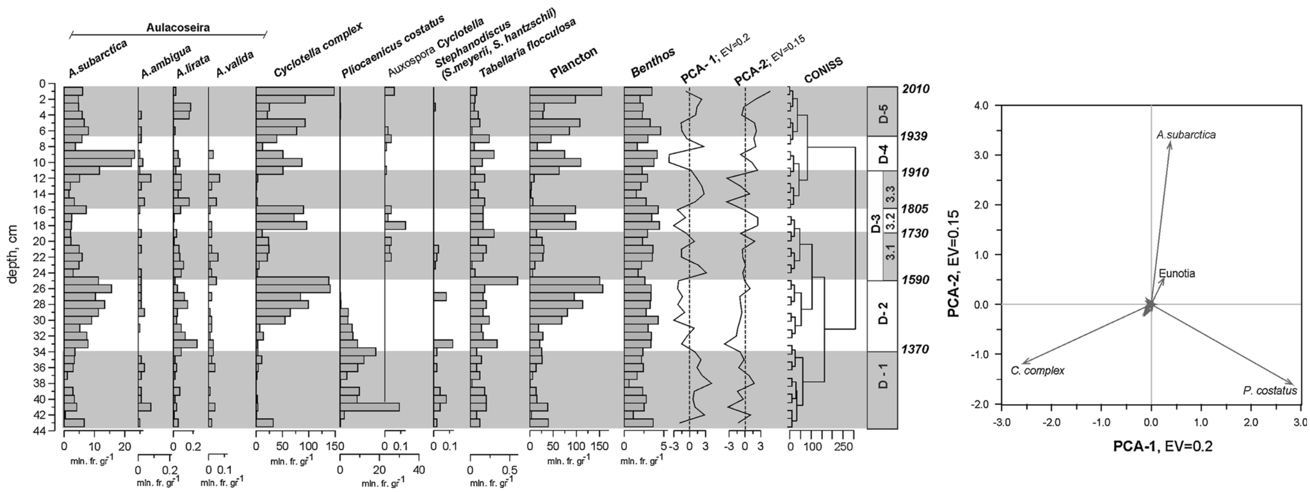


Fig. 3 Biostratigraphical record of selected diatom taxa. *CONISS* a square-root transformation, *PCA 1* and *2* are plots of component scores and variables for the firs two principal components

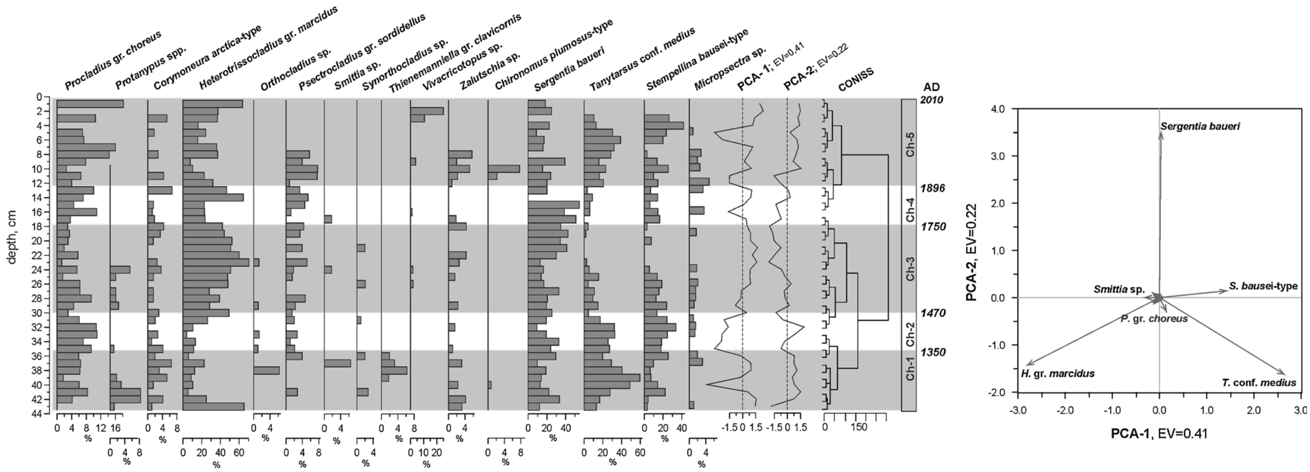


Fig. 4 Biostratigraphical record of selected chironomid taxa plotted as percentages *CONISS* a square-root transformation, *PCA 1* and *2* are plots of component scores and variables for the firs two principal components

We divided the chironomid records into five zones (Ch 1-5). Zone Ch-1 (43–35 cm bss) was dominated by *T.* conf. *medius* (30–60 % abundance). *Procladius* gr. *choreus, Protanypus* spp., and *S. baueri* also were abundant. Zone Ch-2 (35–30 cm bss) *T.* conf. *medius* and *S. baueri* remained abundant in this zone. *Procladius* gr. *choreus*, *S. bausei*-type and *T.* gr. *clavicornis* increased in relative abundance. Zone Ch-3 (30–18 cm bss) *H.* gr. *marcidus* sharply increased in abundance to highest values at between 26–21 cm bss., while *P.* gr. *choreus*, *S. bausei*-type and *T.* conf. *medius* declined. Zone Ch-4 (18–12 cm bss) *S. baueri* remained abundant, *P.* gr. *choreus* and *T.* conf. *medius* moderately increased, while *H.* gr. *marcidus* decreased. Zone Ch-5 (12–0 cm bss) the abundances of *P.* gr. *choreus*, *H.* gr. *marcidus*, *S. baueri*, *T.* conf. *medius* and *S. bausei*-type showed frequent fluctuations.

Pollen record

Two principle component of the pollen date, PCA-1 and PCA-2 which accounted for 50 and 31 % of the total variance, respectively. The PCA-1 had a large loading for *B.* sect. *Nanae* and Poaceae. The PCA-2 had a large loading for *Pinus sylvestris* and *Pinus sibirica.*

The pollen record was divided into four local pollen zones (P 1–5). The pollen assemblage of P-1 (42–34 cm bss) was characterised by a high percentage of tree pollen (Fig. 5). *Pinus sylvestris* (up to 50 %) and *Pinus sibirica* Du Tour (up to 16 %) were the main components of the tree assemblage. *Betula* sect. *Nanae* pollen (up to 13 %) was the main component of the shrub assemblage. Zone P-2 (34–25 cm bss), *P. sylvestris* remained abundant, the abundance of *P. sibirica*, *Betula* sect. *Albae*, *P. obovata* and Cyperaceae pollen increased up to 18, 11, 7 and 1 %, respectively, while *B.* sect. *Nanae* gradually declined in abundance. Zone P-3 (25–12 cm bss) was divided in two sub-zones (P-3.1 and 3.2). This zone showed moderate decline of *P. sylvestris* and *P. sibirica,* distinct decrease of *P. obovata*, while the abundance of *B.* sect. *Nanae,* Poaceae, Cyperaceae and Sphagnales increased gradually in sub-zone P-3.1 (25–20 cm bss) and drastically in sub-zone P-3.2 (20–12 cm bss). Zone P-4 (12–0 cm bss) was characterised by the minimal contents of tree pollen in comparison with other pollen zones. The shrub and herb pollen, and several spore taxa, especially Sphagnales and Polypodiaceae exhibited a tendency to increase of the abundance toward to the zone top.

Elemental composition

Figure 6 shows comparison of the data on elemental composition of sediments by XRF-SR and ICP-MS methods. The distribution of elements obtained there methods was generally comparable, although there is a poor correlation for some elements (e.g., Br, Ti and Zr). A distinct in detected of some elements by these methods was described by Phedorin et al. (2000a) and Zhuchenko et al. (2008). For example, "light" elements (e.g. Br, I, Cu, etc.) are practically inaccessible for determination by ICP-MS, or Ti, Be, Sn, Hf and Zr are poorly extracted from sediment samples at treatment by acids. As result, there are some differences in biplots of PCA-1 and PCA-2 of these methods (Fig. 6). However, the distribution of PCA-1 sample scores for XRF-SR method was correlated ($R^2 = 0.57$) with one for ICP-MS method (Fig. 6). The elemental composition will be closely associated with the distribution of the trace elements typical of the clastic material (e.g., Ga, Y, Nb, Rb, Ti and Th) or autochthonous in origin (e.g., Cu, Br and

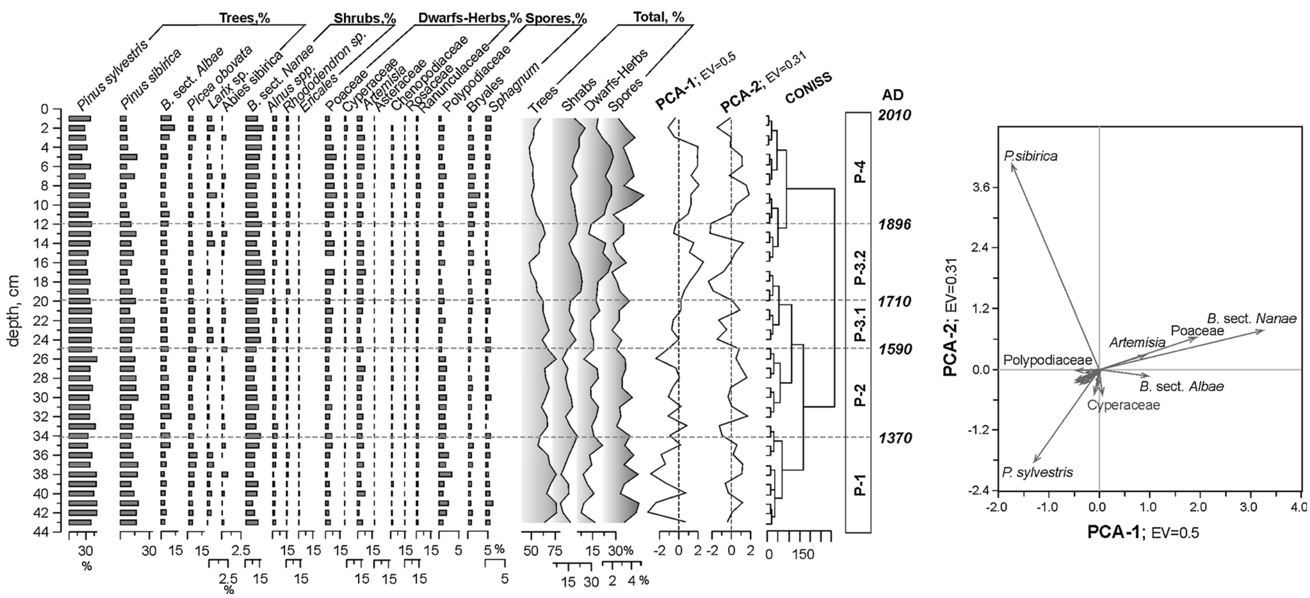


Fig. 5 Biostratigraphical record of selected pollen taxa. *CONISS* a square-root transformation, *PCA 1* and *2* are plots of component scores and variables for the firs two principal components

U) or chemical weathering (e.g., K, Ca, Sr and Mg) (Nesbitt 1979; Pokrovsky et al. 2006).

We divided the elemental record into seven zones (E 1–7) because XRF-SR method was performed with a high sample resolution (1 mm) and "thin fluctuations" in contents of elements was lighted. Transitions between these zones occurred at 36, 32, 25, 19, 14 and 10 cm bss (Fig. 6). Periods with a low content of practically all elements (even zones) were short then periods with a high content of elements (odd zones). There was a tendency when the content of elements increased dramatically during 15–20 years.

FTIR records

According to the distribution of average quartz and feldspar content, we divided mineralogical record into four zones (M 1–4). Zone M-3 (16–6 cm bss) was characterised by maximal contents of quartz and feldspar as contrasted to zone M-4 (6–0 cm bss) when theirs contents were minimal (Fig. 7). A high quartz and feldspar content was found in zone M-1 (42–30 cm bss).

The wavenumber indicates the functional group, which is determined based on absorption. The bands at approximately 2,925 and 2,856 cm^{-1} are typically due to C–H vibrations of CH_3, CH_2 and CH groups. This region also shows a rare, intense band from 1,400–1,470 cm^{-1} (Haberhauer et al. 1998).

The pattern of TOC distribution was more contrast than that of the BiSi (Fig. 7). A high organic content amounts were 6–7.2 % for zone O-3 (24–15 cm bss, ca. 1610–1830) and O-5 (11–6 cm bss, 1909–1939). In contrast, the low amounts (about 4 %) of TOC were detected for zone O-1 (43–35 cm bss) and O-4 (15–6 cm bss). The tendency of BiSi content to increase was occurred from O-4 to the present. The distribution of the BiSi more closely correlated with a distribution of benthic diatoms than plankton diatoms. This was likely due to different sizes of plankton and benthic diatoms. For example, the sizes of plankton dominant *Cyclotella*-complex was 3–15 µm while Eunotia (benthic dominant) was 14–84 µm long and 3–9 µm wide. The distribution of the BiSi and the TOC was closely correlated with distribution of Br and the phosphorus detected with XRF-SR and ICP-MS method, respectively (Fig. 7). Although, there is a disparity between distributions of the TOC and the phosphorus in zone O-6, that should explained by dominance the phosphorus mineral origin there.

Discussion

We compared our obtained records with regional climate records from Asia and Europe and to proxy global climate forcing factors (the temperature anomalies of the Northern Hemisphere and the total solar irradiance).

The intercomparison of biological and non-biological proxies

It is likely that obtained proxies had different time responses to climate changes. For example, it is well-known that diatoms depend on water temperature, the

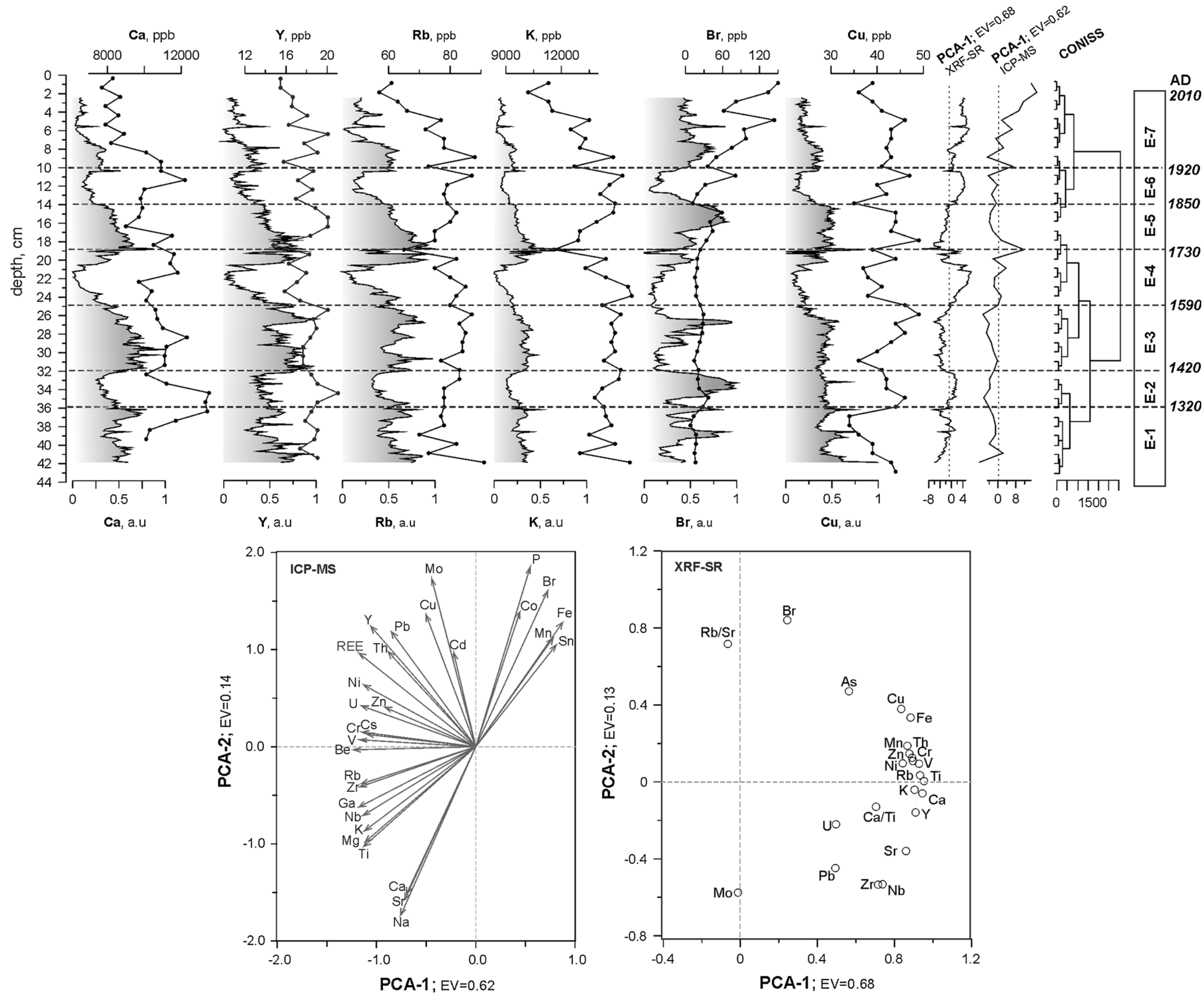
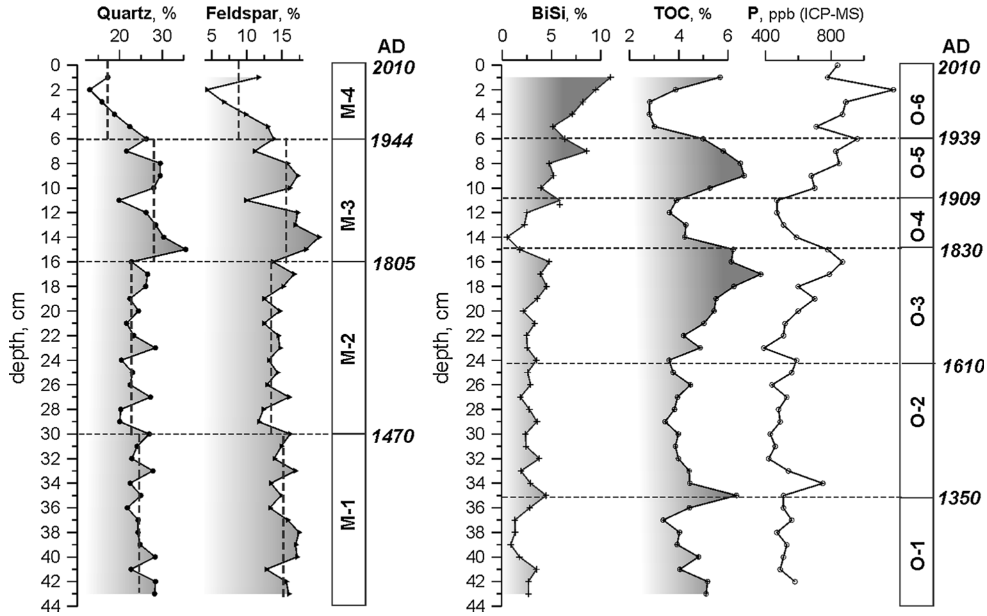


Fig. 6 The distributions of some elements in the sediments of the LM-01/10 core. *ICP-MS method* the upper scale and *black circles*, *XRF-SR* method the bottom scale and *gray shaded*. *CONISS* a square-root transformation; *PCA 1 XRF-SR* and *PCA 1 ICP-MS* are plots of component scores and variables for the first principal components of XRF-SR and ICP-MS method

duration of open and closed water, insolation and the supply of nutrients into water (e.g. Stoermer and Smol 1999). Therefore, diatom records are likely characterised by short time lag in response to changes of these parameters. Chironomid records are sensitive to summer air temperatures (e.g. Heiri et al. 2003; Brooks and Birks 2001), and the time-lag between changes of air temperatures and chironomid taxa is probably minimal. Pollen records are much more difficult for climate reconstructions because vegetation is influenced by temperature, moisture, altitudinal limits, level groundwater, life-span and so on. The time-lag for arboreal taxa is very likely longer than the response of chironomid and diatom records, and this lag probably increases from deciduous to coniferous taxa, whereas the changes of chironomid and diatom taxa occurred earlier (Fedotov et al. 2012a).

It is well-known that chemical weathering is more intensive under warm and moisture conditions. However, the continental climate of East Siberia causes minor chemical weathering and strong physical weathering due to seasonal frost cracking. There was not a strong different between the trace elemental profiles typical of physical (e.g. Rb, Nb, Th or Ti) and chemical weathering (e.g. Ca, K or Sr) (Fig. 6), but there distribution were closely to the distribution of quartz and feldspar. It is most likely that the main part of the trace elements strongly associated with suspended particulates supplied by surface water, when water saturation in the catchment area was high. Direct XRF measurements of modern phytoplankton from bottom sediments have revealed the active accumulation of Br by living organisms in Siberian and Mongolian lakes (Phedorin et al. 2000b, 2008). In addition, it is well known that

Fig. 7 The distribution of quartz, feldspar and organic components (BiSi, TOC and P) along the LM-01/10 core



P, U and Cu easily incorporate into organic materials (Bibi et al. 2010; Zhu et al. 2013; Hou et al. 2014). Therefore, P, Cu, Br and U in lake sediments can be regarded as autochthonous in origin, and a high content of these elements most likely indicated about high like bio-productivity du to an increase in the rate of supply of nutrients into the lake from the catchment area at a high soil water saturation.

It is evident that small lakes with small watersheds exist in a sensitive state of equilibrium with climate and are therefore good indicators of climate change with the minimal time-lag. Therefore, the slightest changes in climate can affect upon bio-productivity of lakes and water saturated soil in the catchment area. Figure 8 likely confirmed this assumption, when changes in biological and non-biological proxies simultaneously occurred. The distribution of the diatoms, TOC, BiSi and elemental zones were most similar, these records also had the highest number of zones. In our opinion, this result may indicate that the lake bio-productivity has primarily depended upon the supply of nutrients into the lake as expressed by elemental records and the water saturation of the catchment area. We assumed that the most significant changes of the lake bio-productivity and the catchment area as to reflect of climate changes were occurred about ca. 1350, 1590, 1730, 1900 and 1940 (Fig. 8).

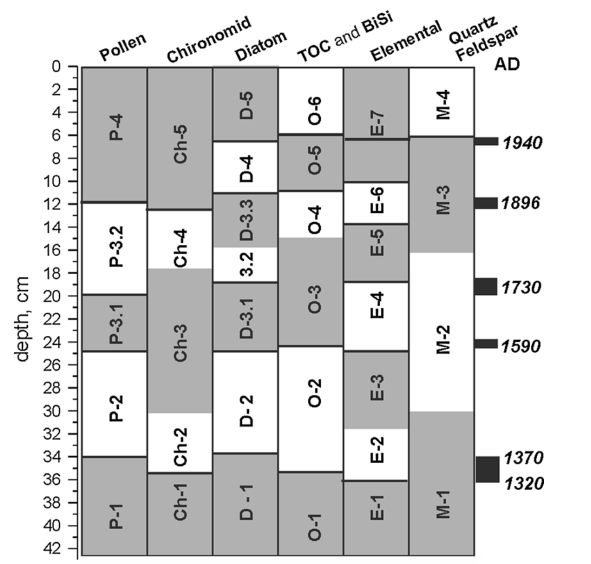


Fig. 8 Correlation between pollen, chironomid, diatom, organic components and BiSi, elemental and mineral zones

Summer air temperature reconstruction

Among various palaeolimnological methods, chironomid analysis of lake sediments as one of the most promising biological methods for reconstructing past temperatures (e.g. Larocque and Hall 2003; Brooks 2006; Walker and Cwynar 2006; Eggermont and Heiri 2012). Temperature reconstruction based on chironomid records relies on the empirical relationship between the taxonomic composition of chironomid assemblages in recently deposited lake sediments and air or lake surface water temperature during the summer months (e.g. Barley et al. 2006; Larocque et al. 2006; Rees et al. 2008).

In present time, Siberian regional mean July air temperatures (T_{July}) were reconstructed using a modern chironomid-based temperature calibration training set only from Yakutia and North-Eastern Russia (Solovieva et al. 2005; Nazarova et al. 2008, 2011, 2013). Unfortunately, a similar calibration training set has not yet been tested for the south part of East Siberia.

We used two approaches to reconstruction July air temperatures. The first approach—the T_{July} was calculated by using a simple WA method based on the ratio of chironomid dominant taxa (*P.* gr. *choreus*, *H.* gr. *marcidus*, *S. baueri*, *T.* conf. *medius* and *S. bausei*-type) in the core LM-01/10. A training set of temperature preferendum for these taxa in the Northern Hemisphere (NH) was calculated according by Linevich (1981), Olander et al. (1999), Larocque et al. (2001), Eggermont and Heiri (2012), Marziali and Rossaro (2013). Temperature optimums of these taxa were characterised by a significant distinction (Fig. 9). For example, *H.* gr. *marcidus* is most abundant at the T_{July} about from +7 to +11.5 °C, in contrast to *T.* conf. *medius* prefers to inhabit at the T_{July} about from +12 to +18.5 °C.

The second approach—a temperature calibration training set was performed by using WA–PLS, when the chironomid composition of each sample from within the upper part of the LM-01/10 formed after 1900 was linked with the regional T_{July}. The data set of the regional T_{July} was obtained from instrumental (weather stations Mondy and Il'chir, ftp://ftp.ncdc.noaa.gov/pub/data) and grid model (temperature data representing the 1900–2010, http://climate.geog.udel.edu). Parameters of WA-PLS were: the r^2 (based on leave one-out cross-validation)—0.97, RMSEP—0.13 °C and maximum bias −0.2 °C.

A range of the T_{July} obtained by WA and WA-PLS were closely similar and mean value of the T_{July}–WA was corresponded with one of the T_{July}-WA-PLS (Fig. 9). The founded range (from +12 to +14 °C) of the T_{July} was similar to that reconstructed for other Holocene records from this area (Mackay et al. 2012). According to our estimates, a clear decrease in summer temperatures occurred in East Siberia after ca. 1400 (33 cm bss), when the T_{July} before ca. 1400 was similar or high in compare to the modern (1900–1950) (Fig. 9). Temperature reconstructions for China (Yang et al. 2002) and West Siberia (Kalugin et al. 2009) also suggest a warm condition during ca. 1160–1400, although some decrease in temperatures occurred about 1300 (Fig. 10). In addition, the level of total solar irradiation (TSI) was also high at the span (Steinhilber et al. 2009). Based on these results, we assumed that the LIA in Central Asia began after 1400. This estimate agreed well with the global temperature record (Mann et al. 2009) but incompletely coincided with some regional reconstructions from Europe and North America (e.g. Jones and Mann 2004; Helama et al. 2009).

The second episode of high regional summer temperatures was occurred about 18–15 cm bss (ca. 1750–1825). The similar increase summer temperature at the end of the 17th/beginning of the 18th century has been observed in Europe (Luterbacher et al. 2004; Xoplaki et al. 2005). In addition, the North Atlantic Oscillation (NAO)—index of summer months was high at the span (Luterbacher et al. 2001). It is very likely that summer temperature rise at the second episode in East Siberia was mostly induced by warm air transfer from the North Atlantic. Fort example, this episode appeared only weakly in temperature records of China (low impact of the North Atlantic transfer), although the level of the TSI was high (Fig. 10).

According to our reconstruction, the coldest summer T_{July} occurred about ca. 1570–1700 and 1830–1900. The beginning of the decrease in summer temperature after 1570 was somewhat unexpected because the level of the TSI was relatively high and the Maunder minimum occurred later; and also European summer temperatures did not show contrasting changes ca. 1600–1700 (Fig. 10). However, empirical reconstructions (the model ECHO-G) of the NH temperature evidenced that annual temperature dramatically decreased after ca. 1550 (Storch et al. 2004) as well as in West Siberia (Kalugin et al. 2009) and China (Anderson et al. 2002; Yang et al. 2002). Central Asian temperature records clearly evidenced that in the last 850 years the regional summer temperature minimum occurred about 1650 (Fig. 10).

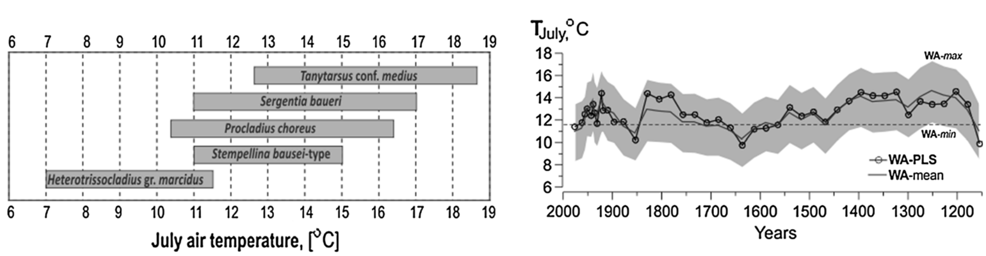


Fig. 9 The reconstruction of summer air temperature. *Left panel* temperature optimums of dominant chironomid taxa from the LM-01/10 core according by Linevich (1981), Olander et al. (1999), Larocque et al. (2001), Eggermont and Heiri (2012), Marziali and Rossaro (2013). *Right panel* quantitative transfer functions of July air temperature based on chironomid analyses were developed using weighted-averaging (WA) and weighted-averaging partial-least-squares (WA-PLS) calibration techniques

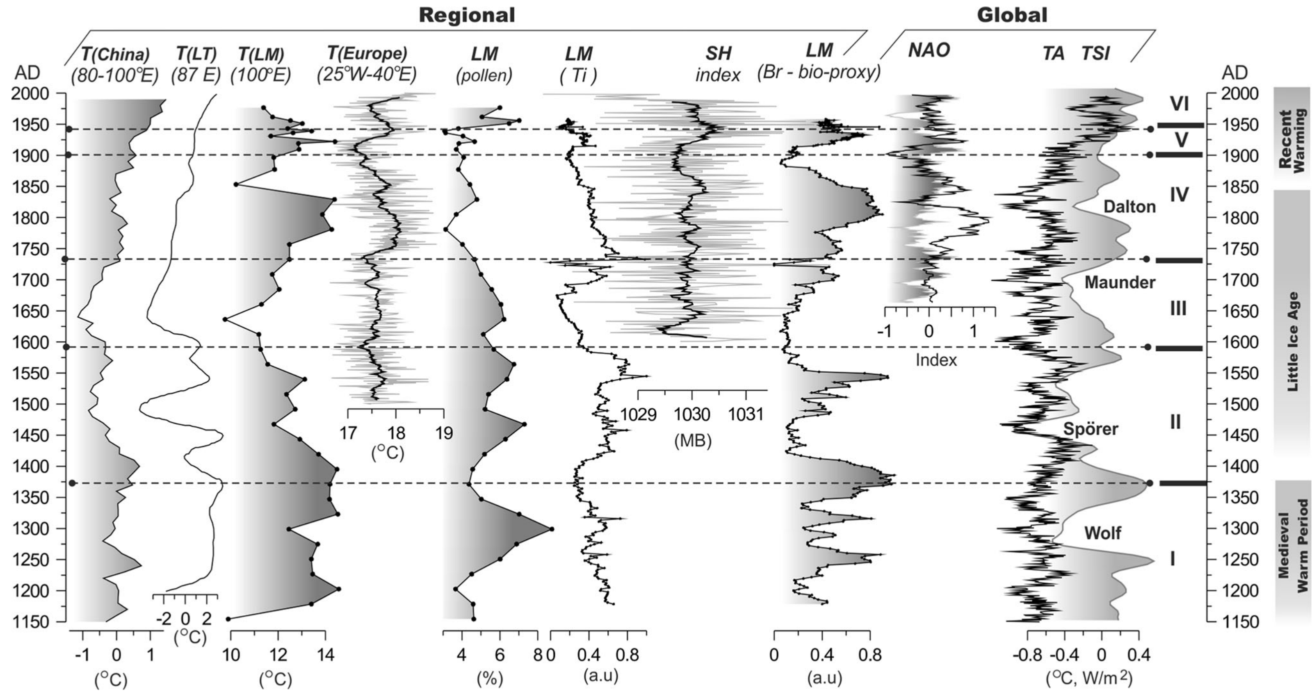


Fig. 10 Relationships between regional records from Central Asia and global record from the Northern Hemisphere and total solar insulation. $T_{(China)}$ China temperature reconstruction (Yang et al. 2002, http://www.ncdc.noaa.gov/paleo/metadata/), $T_{(LT)}$ temperature reconstructiom inferred Teletskoe Lake sediments (Altai region, Western Siberia, Russian) adapted from Kalugin et al. (2009), $T_{(LM)}$ reconstruction July air temperatures based on the distribution of chiromid taxa in the LM-01/10 core, $T_{(Europe)}$ summer surface temperature fields for Europe (Xoplaki et al. 2005, http://www.ncdc.noaa.gov/paleo/metadata/), $LM_{(pollen)}$ the percentage of *P. obovata* and Cyperaceae in the LM-01/10 core, $LM_{(Ti)}$ distribution of titanium along the LM-01/10 core, SH_{index} index activity of the SH, adapted from Cohen et al. (2001) and their running average over 15 points (*black curve*), $LM_{(Br\text{-}bio\text{-}proxy)}$ distribution of bromine along the LM-01/10 core, *NAO* the seasonal NAO with running average over 10 points (Luterbacher et al. 2001, http://www.ncdc.noaa.gov/paleo/metadata/), *black curve* the summer NAO, *grey shaded* the winter NAO, *TA* temperature anomalies in the Northern Hemisphere (D'Arrigo et al. 2006), *TSI* total solar irradiance (Steinhilber et al. 2009, http://www.ncdc.noaa.gov/paleo/metadata/); *I–VI* periods from Fig. 8

According to many reconstructions of global surface temperatures for the NH, a transition from the LIA to the RW was characterised by a sharp increase in annual temperature that occurred ca. 1850–1860 (e.g. Huang 2004; Osborn and Briffa 2006). However, our temperature reconstruction showed that temperature increase was not intense in East Siberia until the1900s. The Dalton minimum was not strong and long in comparison with other periods of low solar activity (e.g. Maunder, Spörer and Wolf minima), however, the decrease of the T_{July} following the Dalton minimum was significant in East Siberia, and did not so notable in China, West Siberia and Europe (Fig. 10). It is likely explained by a contrast response of Lake Mountain under climate changes due to its high altitude position and a strong altitude effect and more duration of the total solar radiance. Liu et al. (2009) have reported a similar assumption that effect of the global increasing temperature is more pronounced at higher-elevations than at lower elevations.

The response of lake bio-productivity and landscapes on climate changes

Teleconnection of the NAO and the Arctic Oscillation (AO) atmospheric circulation and the SH have a strong influence on temperature and precipitation over Eurasia (e.g. An 2000; Gong and Ho 2003; Panagiotopoulos et al. 2005). For examples, the instrumental observations of climate variability in Baikal area and the ice-thermal processes on Lake Baikal for the last 110 year showed a direct relationship between climate changes and the NAO and AO activity, while air and water temperatures also river inflow were gradually reduced due to a decrease in the NAO activity and strengthening of the SH, and vice versa (Shimaraev 2008).

We suggest that distinct changes in the atmospheric circulation and air temperature were represented by six (I–VI) periods in East Siberia over the last 850 years (Fig. 10). However, our climate-proxies showed some distinct results. For instance, *P. obovata* and Cyperaceae prefer settle on the high water saturate soil at mean January temperatures range from −15 to −25 °C (Williams et al. 2006). Thus, an increase of the ratio of *P. obovata* and Cyperaceae is likely interpreted as indicating a high level precipitation and a large supply of surface water enriched by nutrients and clastic matter into the lake. There is a similarity in the distributions of these pollen taxa and "sparingly soluble" tracer elements during 1150–1750 (Fig. 10). However, these records negative correlate with records of lake bio-productivity (e.g. Br, TOC, Figs. 7, 10). Figure 10 shows that high contents of *P. obovata*, Cyperaceae and "sparingly soluble" tracer elements were occurred around: maximums of the SH, positive indices of the winter NAO, low values in the TSI and the T_{July}. It is high probably that changes in combination of these factors can disturb normal lake bio-cycles. For example, the plankton communities of lakes located closely to Lake Mountain have two vegetation periods in late May/early June and September–October (Bondarenko et al. 2002). The first vegetation period can be partly destroyed by the strong winter NAO to produce the increasing in snow depths (the decrease in transparency of the lake ice) and more late spring. For instance, interdecadal winter changes in the NAO activity, which immediately influences ice-break terms of Lake Baikal (Shimaraev 2008). On the other hand, autumn Eurasian snow cover correlates with more intense the SH conditions (Cohen et al. 2001) and high activity of the SH, low levels of the TSI and summer temperatures lead to more early autumn and reducing of the second vegetation period of an algae. Hence, the high ratio of *P. obovata*, Cyperaceae and "sparingly soluble" tracer elements most likely were explained by increased winter precipitation due to prevailing mild and humid winters, as a result of a prevailing positive NAO mode during winter.

Period I (ca. 1160–1370) A beginning of this period was characterised by the high TSI and the T_{July} about +13 °C. A span of humid winters was likely linked with the Wolf minimum (Fig. 10). *Pliocaenicus costatus* was dominant there and did not in other periods. *Pliocaenicus costatus* has mainly been observed in ultra-oligotrophic mountain lakes of East Siberia but it is not often the dominant (Bondarenko et al. 2002; Genkal et al. 2011). The content of TOC, Br and P were moderate high during this span, however, it is likely that saturated of the lake-water by nutrients was an unfavourable for other diatom species. A resembling situation was observed in Lake Bolshoie (East Siberia, Lake Baikal area, 1,800 m above sea level, conductivity 7 µS cm^{-1}), when *P. costatus* is the dominant taxon in the sedimentary diatom assemblage over the last ca. 110 years (Flower et al. 1998). This result may indicate that the intensity of the supply of surface water into the lake during summer and autumn was low. The higher ratio of *Pinus sylvestris* and *P. sibirica* also confirms that dray conditions prevailed at time there.

The TSI significantly increased after the Wolf minimum during ca. 1320–1390, but the growth of annual temperature in the NH was not so significant (Fig. 10). However, our and other regional records clearly evidenced than a respond of Central Asia was more significant, while the T_{July} sharply increased from +12.5 up to +14 °C (Fig. 10).

Period II (ca. 1370–1590) Diatom-productivity was still high but *P. costatus* was gradually replaced by *Cyclotella*-complex and *A. subarctica.* Since ca. 1400 the regional summer temperatures showed a noticeable trend to decreased, when the temperature minimum occurred about the Spörer minimum. We correlated this tendency to cooling with the beginning of the LIA. A tendency to colder climate since ca. 1360 has also been reconstructed from small lake and peat bog sequences from the shore of Lake Baikal (Bezrukova et al. 2006; Mackay et al. 2013).

However, the climate of the LIA seemed to be very unstable with episodes of short warming. For instance, short and drastic episode increasing of summer temperatures and lake bio-productivity was inferred at ca. 1470–1560. At the time, the content of *C. ocellata* and *A. subarctica* was maximal and did closely to the modern. A longer and stronger thermal stratification would favor small, fast-growing planktonic diatoms such as *C. ocellata* (Smol et al. 2005; Genkal et al. 2011). *A. subarctica* usually appears in response to moderate increases in nutrients but is disadvantaged by further enrichment and was particularly abundant in years with short winters (Gibson et al. 2003; Horn et al. 2011). We assumed that a temperature impact on diatoms was more significant then the supply of nutrients.

The availability of pollen of *Pinus sibirica*, *Betula* sect. *Nanae* and Poaceae contrary to a low content of *Picea obovata* indicate the existence of a dray climate from 1370 to 1500. The elemental records from the shallow Lake Shira (East Siberia), located in forest-steppe landscapes ca. 800 km west of Lake Mountain, revealed strong decreasing in annual precipitation during 1437–1603 (Kalugin et al. 2013).

Period III (ca. 1590–1730) It very likely was the most dramatic period with unfavorable condition for lake-biota. A major shift in diatom assemblage composition occurred, with a marked decline in plankton taxa and elemental biomarkers decreased too. For example, *Procladius* is a taxon associated with intermediate/warm climate when in this taxon growth was more closely linked to the attainment of suitable water temperatures in the spring (Dermott et al. 1977; Larocque et al. 2009). In this period, abundance of *Procladius* spp. was minimal for the entire records (Fig. 4).

We founded that colder summer temperatures were about ca. 1610–1670. These changes are similar to changes in temperature regimes reconstructed from lake bottom sediments and soil sequences from other areas of Central Asia (Yang et al. 2002; Kalugin et al. 2009; Zhu et al. 2012; Ge et al. 2013). However, an ice core oxygen isotopic record retrieved from the Belukha glacier (the Siberian Altai, West Siberia), located ca. 800 km west of Lake Mountain, pointed temperature minimum about ca. 1720 (Eichler et al. 2009). Besides, our records were closely corresponded with pollen and charcoal records from the Belukha glacier that during 1600–1700 was driest (Eichler et al. 2011). For example, in our records, the content of tracer elements associating with allochthonous matter (e.g. Ti, Rb or Y) was gradually decreased to extremely low values as result a supply of nutrients into lake was decreased and diatom communities reduced at the time. In addition, a ratio of shrubs increased from the beginning to the end of Period III (Figs. 5, 6, 10). In contras to our records, European summer record and temperature anomalies of the NH was primarily characterised by moderate cold conditions and did not revealed extremely events at ca. 1610–1670. For example, the coldest period of the LIA occurred in Eastern Finland at ca. 1700, when the regional temperatures appear to have been over 1 °C colder than the past 700 years mean.

Deep changes in climate regime of Central Asia was probably explained by a weak summer advection of the NAO and a high activity of the SH occurred on the background of the low TSI (the Maunder minimum) (Fig. 10).

Period IV (1730–1900) The beginning of this period was accompanied by a rapid increase in summer temperatures up to +14 °C at ca. 1775–1825. This temperature tendency was similar to the increase of summer temperatures observed in Europe (Casty et al. 2005; Xoplaki et al. 2005). As a result of the increase in summer temperatures, an increased content of TOC, Br, P, BiSi and diatoms provides evidence that the lake reached of high bio-productivity to ca. 1800 (Fig. 10). Diatom and chironomid records from Alpine lakes showed a notable trend to increasing of lake bio-productivity after 1750 (Gunten et al. 2008).

According to quartz and feldspar records, the supply of fine clastic matter from the catchment area of Lake Mountain also was intensive since ca. 1800 that was very likely associated to increasing in moisture. Decreasing in the abundant of *Pinus sibirica* and *P. sylvestris* from ca. 1730 to ca. 1800 also may be evidence of water enrichment in the catchment area. Significant positives shifting in indices of bio-productivity, the supply of clastic matter and summer temperatures coincided with the global increase of surface temperature in the NH, in particular the drastic increase in the summer NAO and the low winter NAO (Fig. 10). However, according to records from China (Yang et al. 2002), this temperature growth occurred early in China than in Siberia and Europe (Fig. 10). Influence of the East Asia Monsoon on China would be a cause of this discrepancy.

A drastic cooling and gradual decreasing of lake bio-productivity, a notable increasing in a content of coarse clastic material (quartz and feldspar) and an expansion of *P.sibiric* and *sylvestris* occurred at ca. 1850–1900. These changes likely indicated about cold and dry conditions and intensity of soil erosion that were likely linked with the low TSI after the Dalton Minimum and a low activity in the summer NAO. In addition, a winter intrusion of a moderate warm Northern Atlantic transfer in Central Asia was likely blocked due to the high activity of the SH at ca.1840–1875.

According to many reconstructions of global surface temperatures for the NH, a transition from the LIA to the RW was characterised by a sharp increase in annual temperature that occurred ca. 1850–1860 (e.g. Huang 2004; Luterbacher et al. 2004; Osborn and Briffa 2006). However, our evidence shows that prominent changes in climate parameters, vegetation and bio-productivity were not intense until 1900 in East Siberia.

Period V (1900–1940) Regional summer temperatures and moisture showed a clear positive trends from 1910 to 1950 (Fig. 1), and our proxies closely followed by these trends. Lake bio-productivity was high during this span, when plankton diatoms from a minimal abundance to maximal reached during 50 years. Although, *C. ocellata* is a diatom typical of nutrient-poor water, and to prefer very oligotrophic conditions (Anderson et al. 2012; Gurbuz et al. 2003; Cremer and Wagner 2003), in our cause, a high content of *Cyclotella*-complex closely correlate with incrising in summer temperatures and moisture at 1910–1922 and 1934–1944. Major changes in the diatom assemblage of Lake Baikal, with shifts from the autumnal diatom to diatoms that are adapted to growing in the spring under the ice, also occurred since the 1900s (Mackay et al. 2005). Bio-productivity of lakes and chironomid inferred air temperature also notably increased since the early 1900s in arctic Russia (Solovieva et al. 2005).

Beginning with the 1900s, the ratio of hygrophilous vegetation (*Salix* spp., Cyperaceae, Ranunculaceae, Bryales, Lycopodiaceae) notably increased, and it very likely evidenced a high water saturated soil due to permafrost degradation. For instance, significant periods of increased permafrost thawing occurred in West and East Siberia between 1900 and 1938 (Agafonov et al. 2004; Fedotov et al. 2012b).

Period VI (1940 to the present) This period was characterised by decreasing of bio-productivity, however, annual regional temperature and moisture regime was favourable for phytoplankton assemblages. We found two cases for this contradiction: (1) the thickness of the lake snow cover; (2) global anthropogenic forces. Smol et al. (2005) found large increases in planktonic *Cyclotella*-

species in response to lengthening of the summer growing season, which results in reduced ice cover and/or enhanced thermal stratification due to climate warming. For example, the thickness of snow cover less than 10 cm is most likely favourable for vegetation of diatoms in Lake Baikal, because sufficient solar radiation penetrates to drive both photosynthesis and convection (Jewson et al. 2009). In addition, regional snow precipitation dramatically increased since 1936 (Fig. 1), but bio-productivity of Lake Mountain was not (Fig. 10).

A clear trend to acidification and eutrophication of European and Eurasian lakes since the 1940s with the acute acidification phase in the 1960s has been established in multiple studies (e.g., Vitousek et al. 1997; Schöpp et al. 2003; Gottschalk 2011; Moldan et al. 2013; Moiseenko et al. 2013). Anomaly high P and As content in the LM-01/10 formed about the 1960s seems related with anthropogenic forces.

Conclusion

Using geochemical (elemental and mineralogical composition) biological (pollen, diatom, BiSi and chironomid) proxies from a lake sediment core from Lake Mountain (East Siberia), we reconstructed regional climate changes for the last 850 years. Mean July air temperatures were reconstructed using a modern chironomid-based temperature calibration. We compared our obtained records with regional climate records from Asia and Europe and global climate forcing factors (the temperature anomalies of the NH and the total solar irradiance). According to our reconstruction, a clear decrease in summer temperatures occurred in East Siberia after ca. 1400 while the T_{July} similar to the modern (1900–1950) were during 1160–1400. We assumed that a climate condition linked with the LIA started in Central Asia after 1400. The coldest T_{July} occurred about ca. 1570–1700 and 1830–1900. We assumed that the most significant changes of the lake bio-productivity and the catchment were occurred about ca. 1160–1350, 1350–1590, 1590–1730, 1730–1900 and 1940 to the present. The most dramatic period with unfavorable climate condition for lake-biota was during 1590–1730. During the transition from the LIA to the RW, prominent changes in climate parameters, vegetation and bio-productivity were not intense until 1900 in East Siberia.

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