Vegetation, climate and lake changes over the last 7,000 years at the boreal treeline in north-central Siberia

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Abstract

Palaeoecological investigations in the larch forest-tundra ecotone in northern Siberia have the potential to reveal Holocene environmental variations, which likely have consequences for global climate change because of the strong high-latitude feedback mechanisms. A sediment core, collected from a small lake (radius~100 m), was used to reconstruct the development of the lake and its catchment as well as vegetation and summer temperatures over the last 7,100 calibrated years. A multi-proxy approach was taken including pollen and sedimentological analyses. Our data indicate a gradual replacement of open larch forests by tundra with scattered single trees as found today in the vicinity of the lake. An overall trend of cooling summer temperature from a ~2 °C warmer-than-present mid-Holocene summer temperatures until the establishment of modern conditions around 3,000 years ago is reconstructed based on a regional pollen-climate transfer function. The inference of regional vegetation changes was compared to local changes in the lake's catchment. An initial small water depression occurred from 7,100 to 6,500 cal. years BP. Afterwards, a small lake formed and deepened,

probably due to thermokarst processes. Although the general trends of local and regional environmental change match, the lake catchment changes show higher variability. Furthermore, changes in the lake catchment slightly precede those in the regional vegetation. Both proxies highlight that marked environmental changes occurred in the Siberian forest-tundra ecotone over the course of the Holocene.

Keywords

tundra-taiga ecotone; *Larix gmelinii*; palynology; sediment geochemistry; mean July temperature; ordination; WA-PLS; Procrustes rotation

1. Introduction

The globally occurring warming trend is especially pronounced in the arctic region as a consequence of polar amplification (Serreze et al., 2009; Bekryaev et al., 2010; Hinzman et al., 2013) and is expected to accelerated in the future in northernmost Siberia, particularly around the Taymyr Peninsula (IPCC, 2013). To substantiate this prediction it is useful to interpret reconstructions from the past with similar spatial patterns, but few quantitative climate reconstructions are available from northern Siberia.

Reconstruction of past climate requires an understanding of how the climate proxy is temporally and spatially related to climate change. From the ongoing environmental changes we already know that the timing and strength of the various components of the Arctic environmental systems to climate forcing are extremely variable (Lenton, 2012; Hinzman et al., 2013; Pearson et al., 2013). For example, hydrological changes of permafrost lakes may be abrupt but the direction of change varies locally, e.g. rising lake level at one site and increased outflow at a nearby site (Brouchkov et al., 2004; Smith et al., 2005; van Huissteden et al., 2011; Morgenstern et al., 2011; Kanevskiy et al., 2014; Turner et al., 2014). Accordingly, proxies of hydrological changes in thermokarst lakes may respond immediately but change is not linearly related to climate. On the other hand, the vegetation change in response to climate may by uniform, i.e. northward species migration and a boreal forest expansion in times of warming (Naurzbaev and Vaganov; 2000; Elmendorf et al., 2012a, b; Berner et al., 2013; IPCC, 2013). This response to climate variation might be consistent over larger areas but its reaction can be masked regionally (Sidorova et al., 2009; Giesecke et al., 2011; Tchebakova and Parfenova, 2012; Kharuk et al., 2013). At the Siberian treeline, the most reasonable scenarios are leading-edge vegetation-climate disequilibrium at times of climate warming due to restricted larch migration rates and trailing-edge disequilibrium because of persistent forest despite a cold climate. This indicates that a reasonable ensemble of environmental variables needs to be collected to control for the uncertainties originating from the various scales on which processes operate.

Continuous records of millennial-scale environmental changes in northern Siberia are best obtained from lake sediments that can be explored for various parameters. Here, we present results of palynological and sedimentological analyses of a lake sediment core from the southern Taymyr Peninsula (northern Siberia) covering ~7,100 cal. years BP to present. Because pollen is still one of the most reliable climate proxies available for the region, we provide a pollen-based climate reconstruction and assess the obtained results in connection with local hydrological changes as inferred from sedimentological and geochemical parameters.

2. Regional setting

The Khatanga River Region forms part of the Northern Siberian Lowlands and is located between the Taymyr Peninsula to the North and the Putorana Plateau to the South, politically belonging to the Krasnoyarsk Krai of Russia. The studied lake’s catchment is underlain by thick terrigenous and volcanic sediments that are rich in smectite originating from Siberian Trap basalts of the Putorana Plateau (Wahsner et al., 1999; Petrov, 2008; Vernikovsky et al., 2013). Overlying Quaternary periglacial and, to some extent, lacustrine-alluvial deposits are predominately of Putoran origin and therefore basaltic (Peregovich et al., 1999; Shahgedanova et al., 2002). Loadings in the Khatanga River have been reported to comprise up to 80% of the montmorinolit clay mineral smectite (Rachold et al., 1997; Dethleff et al., 2000). The lowland's landscape is homogeneous with low relief. The region was probably not or only locally glaciated during the Last Glacial Maximum but was situated between the glaciers of the Taymyr and Putoran Mountains, hence, periglacial conditions prevailed (Svendsen et al., 2004; Ehlers and Gibbard, 2007). The region is controlled by continuous, very deep permafrost with medium ground-ice content up to 20% by volume (Schirrmeister et al., 2013; Brown et al., 2014) and numerous lakes are found there (Ananjeva and Ponomarjeva, 2001).

The regional climate is dominated by the polar front, which is located close to the coast of the Arctic Ocean during winter. In summer, the region lies within the arctic front. Prevailing winds are from the north-west and south-east (Treshnikov, 1985; MacDonald et al., 2000b; Pospelova et al., 2004). The subarctic climate of the region is continental, having short and mild summers with a mean July temperature around 12.5°C and severe winters with a mean January temperature ~ -31.5°C. Annual precipitation is low, around 250 mm with the most rain falling during the summer month between June and September. Snow cover lasts between 180 and 260 days with up to 80 cm height (Grigoriev and Sokolov, 1994; climate station, established in Khatanga town in 1934, http://www.pogodaiklimat.ru/climate/20891.htm).

The vegetation of the region represents the southern fringe of shrub tundra and is composed of a mosaic of vegetation types (Stone and Schlesinger, 1993; Yurtsev, 1994; CAVM, 2003) with continuous vegetation cover, but locally, for example on drier hilltops, bare soil may be found (Chernov and Matveyeva, 1997). The moss layer is extensive and at least 10 cm thick. The most abundant genera are *Sphagnum, Hylocomium, Aulacomnium, Dicranum*, and *Polytrichum*. The herbaceous and dwarf-shrub layer grows up to fifty centimetres high. Dominating are sedges, such as species of *Eriophorum* and *Carex*, and shrubs, especially *Ledum palustre, Vaccinium* species, *Betula nana*, and *Alnus viridis* subsp. *fruticosa*. This shrub tundra is dotted by stands of *Larix gmelinii* (Abaimov, 2010). In this area, the northernmost “forest islands”, with the regional name *Ary-Mas*, grow as far north as 72°56’N (Bliss, 1981; Tishkov, 2002). The main human impact in the Khatanga River region is commercial reindeer herding, which intensified from the 1960s (Pavlov et al., 1996).

The study site is located at 72.40°N and 102.29°E; 60 m a.s.l. The small lake—given the technical name CH-12—is elliptic in shape with a surface area of around 2.4 hectares and a mean radius of 100 m (Fig. 1). Its maximum depth is 14.3 m. The lake is located in a confined depression on a low-lying plateau in the northern lowlands. It has no inflow streams but drains the surrounding ridges. One small outflow is present on its western side draining into the Novaya River, which is one of the main tributaries of the Khatanga River. Our vegetation surveys within the catchment revealed that the lowgrowing shrub tundra is dominated by Ericaceae dwarf-shrubs (*Cassiope tetragona*, *Vaccinium vitis-idaea* and *V. uliginosum*) while *Betula nana* and *Alnus fruticosa* are more rare and only obtain low growth heights (< 20 cm). *Salix* spp. grow predominantly along the river and lake shorelines. Cyperaceae and Poaceae, as well as herbs such as *Dryas octopetala* ssp. *punctata*, are abundant. Scattered patches of *Larix gmelinii* trees up to 5 m in height occur in the area.

[figure 1]

3. Material and Methods

3.1. Material collection

Fieldwork was undertaken as part of a joint Russian-German Expedition to the Khatanga region in 2011. Sampling took place at a central lake position at 14.3 m depth, where a 131.5 cm-long core with a UWITEC gravity corer extended with a hammer action was deployed. The core was subsampled in Germany at the laboratory of the Alfred Wegener Institute (AWI). To allow for a precise estimation of the sedimentation rate of the investigated lake, a parallel short core of 32 cm was obtained and sliced into 0.5 cm thin samples in the field.

3.2. Age determination

The uppermost 10 cm of the short-core were freeze-dried and sent for radiometric dating of lead and caesium at the *Environmental Radioactivity Research Centre* of the University of Liverpool, UK (Appleby et al., 1991 and 2001). Furthermore, material (moss, wood or leaf remains or bulk sediment) from fifteen samples were freeze-dried and sent to the *Poznan Radiocarbon Laboratory*, Poland, for radiocarbon dating The age-depth model was established using the Bacon package (Blaauw and Christen, 2011 in the R environment version 3.02 (R Core Team, 2013) , in which the calibrated ages before present (cal. years BP) are based on IntCal13 (Reimer et al., 2013).

3.3. Pollen analysis

For pollen analysis, 65 fossil sediment samples of 1.5 ml were retrieved using plastic syringes and prepared following standard procedure (Fægri and Iversen, 1989, HCl, KOH, HF cooking for 2h, acetolysis). Final samples were mounted in water-free glycerine and examined at 400X magnification.

Pollen taxonomic determination was based on a regional reference collection and standard literature (Moore et al., 1991; Reille, 1998; Blackmore et al., 2003; Beug, 2004; Savelieva et al., 2013). Pollen types are given in the text in CAPTIAL letters to facilitate the differentiation between POLLEN TAXA and plant taxa (Joosten and de Klerk, 2002). At least 500 terrestrial pollen grains were counted for each sample. Non-pollen palynomorhps, such as coniferous stomata (Hansen, 1995), were counted alongside the pollen grains.

3.4. Sedimentological (geochemical and granulometric) analyses

There were no signs of hiatuses in the record. At 109–111 cm the sediment was offset, possibly due to the coring process, but no loss of material was indicated in the field or in the laboratory examination. The core description follows initial analyses and picture scan results. The sediment core was opened in the laboratory at AWI Potsdam, and one half was directly transported to the laboratory AWI Bremerhaven to perform line-scanning using the Avaatech XRF scanner using a Rh X-Ray tube at 1 mA and a 10 s count time at 10 kV without a filter, and at 30 kV for heavier elements, with a “PD thick” filter. The resolution of logging was set to 5 mm. This study presents the geochemical results of the aluminium, titanium, silicon, rubidium, strontium, bromine, iron, and manganese counts (252 observations). For statistical analysis we used the log-ratios of the elements (Weltje and Tjallingii, 2008). The relatively heavy element titanium, showed stable count results with low X^2 errors (mean $X^2 = 0.97$). It had the highest correlation to biogenic components, with a Pearson correlation coefficient of 0.72 for total organic carbon (TOC) and 0.69 for total nitrogen (TN). Consequently, titanium could be used to normalise the other elements and counteract the dilution effect of high organic material content to some extent (Löwemark et al., 2011; Shala et al., 2014). Prior to the analysis extreme outliers were excluded, e.g. those from the edges of the core or those around inclusions and at the offset at 109 cm. To allow numerical correlation with other sedimentological proxies the running means of 2 cm window-size of the scanning data were calculated.

The gravimetric water content (WT) was measured for 66 samples of the sediment core to infer the compaction of the sediment calculated as the difference between wet and dry weight of the material. A Vario EL III carbon-nitrogen-sulphur analyser was used to measure total carbon and TN content; and a

Vario MAXC analyser was employed for TOC measurements. Total inorganic carbon (TIC) was calculated as difference between the total carbon and TOC. The elemental ratio of the weight percentages of TOC and TN was calculated to check for possible variation in the sedimentary origin of the organic matter (Meyers and Lallier-Vergés, 1999), hereafter referred to as C/N ratio.

Sediment particle sizes of 65 samples were measured. A minimum of 2.5 g sediment was first treated with 35% hydrogen peroxide for four weeks to remove the organic components. Second, 10% acetonic acid was used to remove calcium carbonate within the remaining sample. Last, the volume percentage of 86 particle size classes between 0.3 and 1000 µm particle diameter were measured with a COULTER LS 200 Laser Diffraction Particle Analyser. The reported volume percentages were calculated from the particle diameter classes: 0.0625–1 mm, 2–62.5 µm, and 0.3–2 µm.

3.5. Data analysis

Pollen percentage calculation was based on the total terrestrial pollen count and pollen concentrations were calculated using *Lycopodium* marker spores (Stockmarr, 1971). Ordination analyses of the pollen data were based only on those 31 taxa that occurred in at least five samples of the core. The stratigraphically constrained cluster analysis (CONISS) was based on the Bray-Curtis dissimilarity matrix (Grimm, 1987), and to assess the significance of the obtained clusters the broken-stick model was used (Bennett, 1996). Principle component analysis (PCA) was based on square-root transformed pollen data. To reconstruct past climate variation, a previously established pollen-climate transfer function for mean July temperature (T_{July}) based on pollen spectra exclusively from lake surface-sediments from northern Siberia (Klemm et al., 2013) was applied to the fossil pollen spectra from CH-12. Fifteen modern surface samples from the Khatanga expedition 2011 were added following the same protocol so that the calibration set consisted of 111 modern spectra in total. The included modern T_{July} data ranges between 7.5 and 18.5°C, this data was retrieved from MODIS satellite imagery from the years between 2007 and 2010. The inclusion of these surface samples into the modern pollen dataset slightly improved the performance of the weighted-average partial least squares model, for which one component was employed, resulting in a root mean square error of prediction of 1.66°C and maximum bias of 4.1°C for T_{July}. The significance of the final reconstructed T_{July} was tested against possible reconstructions derived from random environmental data (using 1000 reconstructions; Telford and Birks, 2011). The complete modern and fossil datasets are available from: *PANGAEA link (follows upon publication)*.

The grain size data was analysed with the end-member modelling algorithm using a W-transformation described in Dietze et al. (2012, accessible through the EMMAGeo R-package). With this approach, the contribution of robust end-members (EM) to all the different size classes as well as the quantitative EM contribution throughout the sediment core can be identified (Weltje, 1997; Weltje and Prins, 2007). The selection of the minimal potential number of end-members was based on a minimal cumulative explained variance of at least 0.9% of the total dataset variance. The value of the mean coefficient of determination (r^2) was used to determine the maximum number of EMs. The robustness of the EMs was tested and the final robust EM and the residual member were calculated. Furthermore, the elementary ratios and the grain size data were jointly analysed to retrieve patterns in the sediment signal of the lacustrine archive via cluster and ordination analyses. The constrained cluster analysis and final ordination followed the same approach as described for the pollen data analysis but employed a Euclidean distance matrix to standardised and log(x+1) transformed data of every second centimetre (Legendre and Gallagher, 2001).

To test whether the sediment signal and the pollen signal followed similar trends over the core, the ordination results of both PCAs, using the first two axes scores, were compared with a Procrustes rotation and associated PROTEST with 1,000 permutations (Jackson, 1995; Wischnewski et al., 2011). The Procrustean superimposition approach scales and rotates the ordination results to check for a maximal fit of a superimposition between ordination results (Gower, 1971; Peres-Neto and Jackson, 2001).

All statistical data analyses were performed in the R environment version 3.02 (R Core Team, 2013) using the analogue (Simpson and Oksanen, 2014), rioja (Juggins, 2014), palaeoSig (Telford, 2015) and vegan (Oksanen et al., 2015) packages.

4. Results

4.1. Age-depth model

The 131.5 cm-long lake sediment core covers the time from 7,100 cal. years BP to the present-day (Fig. 2 and Table 1). $^{210}Pb/^{137}Cs$ results indicate a relatively stable, recent sedimentation rate of about 0.03 cm/a (Table 2). The age-depth model based on radiocarbon dates shows a similar and stable accumulation rate over nearly the whole core of around 0.025 cm/a. However, between the depths of 87 and 61 cm, corresponding to a time between 5,400 and 2,600 cal. years BP, a lower accumulation rate of ~0.01 cm/a is inferred. The comparison of radiocarbon dates based on terrestrial wood and moss samples with nearby bulk samples does not reveal any offset. However, the bulk sediment date of the top part of the sediment, at 5.5 cm, dates to about 1,280 ^{14}C years, whereas radiometric dates of lead and caesium for the uppermost samples show that these sediments are clearly of more recent origin given that the timing of nuclear weapon testing in the 1950s and early 1960s is captured within the core's uppermost three centimetres, the 'true' radiocarbon ages of those samples are most likely affected by nuclear activities (Manning et al., 1990). In the final age-depth model, the radiocarbon result of this upper sample is disregarded.

[figure 2, table 1 and 2]

4.2. Pollen data

All pollen spectra are dominated by shrub pollen of BETULA NANA type and ALNUS VIRIDIS type, and POACEAE and CYPERACEAE contributions are also high throughout the core spectra (Fig. 3). LARIX is present only at low percentages ranging between 0.3 and 9.9% showing a decreasing trend throughout the record. The depth-constrained cluster analyses reveals two significant pollen zones, which were further subdivided on visual inspection. The lower zone (PZ I: 131-53 cm, 7.1-2,200 cal. years BP) is characterised by high LARIX, BETULA NANA type and ALNUS VIRIDIS type, while the upper zone (PZ II 52-0 cm, the last 2,200 years) is rich in POACEAE and CYPERACEAE.

The first PCA-axis (Sup. Fig 1A) explains 70% of the total variance; high 1st axis scores are correlated with high LARIX and ALNUS VIRIDIS type percentages, whereas negative scores are correlated with

POACEAE, CYPERACEAE and PINUS percentages. The second axis explains only 7% of the variance within the dataset and is positively correlated to BETULA NANA type and negatively to ERICACEAE and some herb taxa, such as CHENOPODIACEAE and BRASSICACEAE.

A transfer function-based estimate of July temperature for the upper sample yields 14.5°C, which is in close agreement with the modern satellite-based temperature inference of 14.2°C for the Khatanga region (mean over n=15). The test of the significance of the transfer-function indicated that the pollen-inferred T_{July} reconstruction was statistically significant (p=0.037). The pollen-based climate reconstruction of T_{July} revealed a cooling trend over the last ~7,100 cal. years with an absolute change of about 2 °C. Relative to the overall Holocene cooling trend, periods of variable summer temperature occurred between 1,500 and 1,000 cal. years BP (4 samples) as well as between 900 and 700 cal. years BP (3 samples).

[figure 3, Sup. Fig 1A]

4.3. Sedimentological data

Total organic carbon (TOC) varied between 0.9 and 17.8 wt% and total nitrogen (TN) ranged between 0.1 and 1.5 wt% (Fig. 4). Both element curves show generally similar variations, still C/N varied between 1 and 16. Bromine counts correlated well with the organic components (Pearson correlation index: 0.6–0.65). Over the whole core, the water content varied between 15 and 85 wt%. In the bottom ten centimetres, high values are measured followed by a drop around 120 cm depth and then by a steady gradual increase of the water content towards the surface sediments. The geochemical components expressed as the ratios Al/Ti, Si/Ti, Rb/Sr, and Fe/Mn show relatively small variations throughout the core, with the highest variability in the lower 45 cm (7,100–5,500 cal. years BP, Fig. 4). Iron and manganese show similar trends throughout the core, however Fe shows more variation, particularly since 2,700 cal. years BP.

The minerogenic sediment component mainly consists of fine to medium silts with occasional sections of fine sands with a mean grain size of ~11 µm and maximum sample means of 75 µm. The chosen EM model explains a mean of 79% of the total variance over the sediment core. The model error is largest in the lowermost section of the core. EM1 has its main maximum in the medium-to-fine sand fraction. EM2 displays its maximum at the silt-to-clay transition (Sup. Fig 2A).

Depth-constrained cluster analysis of the various sedimentological datasets reveals a significant split at 115 cm depth (~6,600 cal. years BP). Based on the clustering and visual inspection, the upper zone was further divided into six subzones (Fig. 4). The first and second PCA axes explain 50% and 15% of the variance, respectively (Sup. Fig 3A). The first axis was positively correlated to EM1 and Rb/Sr and negatively correlated to EM2 values and Al/Ti. The second axis separated TOC and C/N, which spanned the positive side, from Fe/Mn on the negative side.

[figure 4, Sup. Figure 2A and 3A]

4.4. Numerical comparison of pollen and sedimentological data

Generally, the sedimentological parameters show higher variability than the pollen data, however the overall trends of the two datasets are significantly correlated as revealed by Procrustes rotation (r=0.49, p<0.001). The goodness of fit between the ordinations is shown in figure 5 with periods of higher agreement having lower residuals. However, a simple inspection of the two cluster analyses shows that the respective clusters of each dataset do not completely overlap. First, the main division of the sediment dataset, which separates the bottom section from the remaining core (the last 6,500 years), is not indicated in the pollen zonation at all. This section has high concentrations of stomata and MENYANTHES TRIFOLIATA. Second, periods of major change in the sedimentological data during the last 6,500 cal. years BP always slightly preceded periods of major change in the palynological data (Fig. 5). For example, major change in the sedimentological data between 2,500 and 2,300 cal. years BP finds a counterpart in the pollen data around 2,200 cal. years BP. Likewise, a sedimentological regime shift recorded for the period between 1,500 and 1,000 cal. years BP may correspond to an abrupt change in the pollen data around 700 cal. years BP.

[figure 5]

5. Discussion

5.1. Assessment of investigated parameters as proxies for regional vegetation and climate, and lake catchment development

With the selection of the study site we aimed at capturing a regional-scale pollen signal. Because CH-12 lacks any inflowing streams, the portion of fluvial pollen input should be minimal; also only a minor proportion of pollen may be introduced to the small lake via slopewash (Crowder and Cuddy, 1973; Fall, 1992). Consequently, most of the deposited pollen grains are of aerial origin. As a function of the lake size, the relevant source area of pollen (RSAP; Sugita, 1994) is expected to encompass an area with a radius of hundreds of metres to a few kilometres. An estimation of its actual size depends not only on lake size but also on surrounding vegetation, namely its composition, spatial structure and openness (Sugita et al., 1999; Bunting et al., 2004, Poska et al., 2011). Today the lake is surrounded by tundra with a high portion of arctic herbs characterised by low pollen productivity. The background pollen loading is high and the spatial scale of vegetation reflected in the pollen source is quite large (Pitkänen et al., 2002; Broström et al., 2005; von Stedingk et al., 2008). The RPSA is possibly above ten to twenty kilometres in radius as suggested by the high value of 25 km published for the modern vegetation in the Khatanga River region (Niemeyer et al. 2015). The RSAP was probably much smaller in times of denser forests during the mid-Holocene compared with today. This theoretical consideration is supported by the observation that PINUS values vary contrarily to LARIX. We regard pine pollen as an indicator of landscape openness, because no modern or fossil presence of pine trees in the regional vegetation is documented. Reported modern and fossil occurrences of *Pinus* are at least 200 km away, east and south of the study site (Hultén and Fries, 1986; Kremenetski et al., 2000). PINUS grains are well known for their long-distance transport particularly in open landscapes (Birks and Birks, 2003; Hicks, 2006, Ertl et al., 2012). Awareness of such changes in landscape openness and RSAP is needed when pollen signals are compared with other environmental variables.

It is well-known that LARIX is underrepresented in the pollen spectra compared to its abundance in the vegetation, because it is a medium-to-low pollen producer and has a low pollen dispersion (Clayden et al., 1996; Binney et al., 2011; Klemm et al., 2013). Being a deciduous tree, its foliage production is high and, therefore the interpretation of pollen records with respect to treeline changes can be aided by *Larix* stomata concentrations in the sediment (Ammann et al., 2014; Birks, 2014). Still the estimation of larch cover remains a challenge, and LARIX percentages of around as little as 0.5% may indicate its local presence in the vegetation (Lisitsyna et al., 2011). Modern sediment studies from northern Siberia indicate that northern larch forests are typically reflected by 2% LARIX in the pollen spectra (Klemm et al., 2013).

The pollen-based quantitative mean July temperature reconstruction is highly correlated to PCA1 and the reconstructed changes are larger than the error ranges. The significance of the T_{July} reconstruction for this core also supports that T_{July} may be the driving force of pollen changes. Therefore, the trend and the absolute temperature offset between the middle and late Holocene can be considered reliable. The absolute values, however, may be rather biased towards the mean of the trainings set (see e.g. ‘edge-effect’ as discussed by Birks et al., 2012). The absolute values are slightly higher than the Khatanga climate station measurements of 12.5°C, because the transfer function is built upon MODIS satellite images deriving from the relatively warm summers between 2007 and 2010 (Klemm et al., 2013).

Lake CH-12’s catchment is without fluvial inflows and well-confined within a few hundred metres of the lake’s edge; consequently the scale captured by sedimentological proxies is relatively local. C/N is indicative of the relative contributions of aquatic and terrestrial organic matter to the lacustrine sediment. The obtained C/N ratios mostly range between 10 and 15 suggesting a mixture of both sources (Meyers and Teranes, 2001). We assume that high C/N values, for example at the bottom of the core, relate to low water levels which cause high amounts of terrestrial material to reach the coring position at the centre of the lake. Based on the C/N ratios we assume that relative TOC content at this lake likewise mirrors the relative changes in organic and minerogenic material supplies but is also affected by the within-lake productivity (Briner et al., 2006). The Fe/Mn ratio is assumed to represent the level of lake-water mixing at the water-sediment interface (e.g. Haberzettl et al., 2007; Och et al., 2012; Naeher et al., 2013; see supplementary material for details).

According to our field observations the sediments within the small catchment are rather homogeneous. Changes in the grain-size composition and selected elemental ratios of the minerogenic component therefore predominately represent variations in the transportation and sedimentation processes in the direct vicinity of the coring position rather than changes in the material source (Dearing and Jones, 2003). The grain-size data of this lake core indicate the occurrence of two main sedimentation regimes within the last 7,100 years. Sections of clay-to-silt sediments, and higher Rb/Sr values, can be assumed to represent times of deep lake conditions, because a large distance between the coring position and the lake shore causes the sedimentation of a rather fine fraction. In contrast, sections of higher grain size variability and high sand contributions represent unstable lake conditions and an influx of less sorted sediment from near-by lake shores. These grain size signals correspond well to changes in elemental ratios, among them Al/Ti that likewise reflects the transport of coarser minerogenic material to the lake centre. (A detailed discussion of the applicability of these ratios is provided in the supplementary material).

5.2.Vegetation and climate change in Arctic Siberia over the last ~7,000 years

Our palynological investigation reveals a general larch forest decline during the last ~7,100 years. The mid-Holocene vegetation was characterised by open *Larix* taiga with *Alnus* shrubs in the understorey. Modern vegetation conditions, i.e. shrub tundra, dominated by sedges and grasses with only sparse *Larix* stands, became established at approximately 2,200 cal. years BP. This observed general Holocene vegetation trend confirms earlier investigations from north-eastern Siberia using pollen and/or macrofossils analyses (e.g. Prentice and Webb, 1998; Hahne and Melles, 1997; Tarasov et al., 1998, 2007; MacDonald et al., 2000a, 2008; Andreev et al., 2011 and references therein) or modelling approaches (Monserud et al., 1998, Kleinen et al., 2011). Our record reveals that the strong turnover occurred between 3,000 and 2,000 years ago; a similar timing of strong change has also been reported from other sites in the Taymyr region (fig. 6) and or throughout most circumarctic environments (Kaufman et al., 2004; Salonen et al., 2011; Luoto et al., 2014).

[figure 6]

5.3. Catchment and lake development

The initial lake development started from a small water-hole in a boggy environment. High terrestrial organic input together with the presence of large macrofossils supports a conclusion of very local sedimentation of plant material into a small wet depression. Additionally, the presence of pollen from the semi-aquatic *Menyanthes trifoliata* is typical for a shallow water-logged environment. Initial lacustrine sedimentation started around 7,000 cal. years BP during the late phase of the regional climate optimum that occurred from 9,000 to 6,800 cal. years BP (Andreev et al., 2011). Thermokarst processes are assumed to be more active in times of warming and accordingly strong thermokarst activity has been reported for Siberia during the early and mid-Holocene (Romanovskii et al., 2004; Grosse et al., 2006). During that time, high temperatures and high humidity together with poor drainage may have promoted the formation of a small water-filled depression at the study site lasting for around 500 years.

The following subsidence of the initial depression may have been rapid due to internal feedback mechanisms (Czudek and Demek, 1970; Murton, 2001). In modern Yakutia, fast subsidence rates of 5–10 cm/a (Brouchkov et al., 2004) and 17–24 cm/a (Fedorov and Konstantinov, 2003) are reported. Our sedimentological data from the period following the initial lake formation show high variability from 6,500 until around 5,200 cal. years BP, indicating processes of a deepening water body and relief formation. Thaw slumps and instable lake margins might have led to a mix of fine and coarse material accumulating in a shallow, well-ventilated lake. Our reconstruction suggests that lake sedimentation stabilised, probably because of the formation of a deeper lake after about 5,200 cal. years BP. Over the last 5,200 years the lake experienced two short-term changes in the sedimentological regime, at about 2,500 cal. years BP and about 1,500 cal. years BP, where strong inputs of unsorted material to the lake basin occurred. Such inputs may indicate either a change in the hydrologic regime of the lake's catchment leading to an increased water inflow from the surrounding slopes or represent the input due to slumps from instable margins.

5.4. Assessment of the reconstruction

The pollen-based climate reconstruction of our study yields a summer temperature change of about 2 °C over the last 7,100 years. This magnitude of Holocene temperature change is in general agreement with other studies from the Taymyr region and throughout northern Siberia (Miller et al., 2010; Andreev et al., 2011) and has been attributed to a decrease in solar radiation in summer over the high-northern latitudes (Berger and Loutre, 1991) and related high-latitude feedback mechanisms (Kerwin et al., 1999; Wanner et al., 2008; Marcott et al., 2013). Some distinct short-scale variations are obvious within the last 2,000 years of the reconstruction (fig. 6). A warm phase around 1,500–1,000 cal. years BP may reflect the Medieval Climate Anomaly (MCA, defined after Mann et al., 2009 between 1,050–750 years ago in northern Europe). A possible MCA is also indicated by tree-ring chronologies from the nearby Khatanga region (Briffa et. al., 2008; McKay and Kaufman, 2014). Also regional lacustrine summer temperature reconstructions based on pollen and diatoms indicate a warm MCA (e.g. Lama Lake: Andreev et al., 2004; Kumke et al., 2004). This warm interval was followed by a rapid cool period in the Northern Hemisphere known as the Little Ice Age (Overpeck et al., 1997; Briffa and Osborn 1999; Briffa, 2000; MacDonald et al., 2008). At Lake CH-12, a cooling is indicated around 900 cal. years BP, as is also found in the 100 km-distant Labaz Lake region (Andreev et al., 2002).

The general similarity in the proxies for local lake and catchment changes and regional vegetation change probably originates from a joint driver, which most likely is climate variation. Earlier studies found that, compared to vegetation changes, changes in the within-lake sedimentation or catchment erosion are captured in sediments mostly with short time-lags (Dearing and Jones, 2003). Other possible factors that would result in similar changes in the proxies are disturbances through, for example, fire, insects, or humans. In this pristine setting human disturbance can be considered minimal, as can major effects from insects (Hauck et al., 2008; Dulamsuren et al., 2010). However, fire is a frequent feature in the forest-tundra ecotone (Berner et al., 2012) and may have affected the study site to some extent. A charcoal analysis, however, was not included in this approach.

This comparison of the environmental development at two spatial scales yielded that the local changes within the lake and its catchment possibly preceded the regional vegetation changes by several decades. However, more detailed inferences about vegetation lag-times are not possible because of the limited temporal resolution of the reconstruction results. Accordingly, only the general trends of pollen-based reconstructed climate, i.e. variations on millennial time-scales are reliable while short-term changes may be biased by lagged responses. Still, we assume that pollen is the most reliable proxy for climate reconstruction because all limnological proxies potentially respond non-linearly to climate change.

6. Conclusions

An overall cooling of summer temperature by about 2 °C since 7,000 cal. years BP was reconstructed by the application of a pollen-based transfer function to a sediment record from a lake located at the present-day northern larch limit on the southern Taymyr Peninsula. This trend is significant and adds to information to the Taymyr region especially due to the good resolution of the lacustrine core for the last 2,000 years. The temperature decrease mainly reflects the density decrease of larch forests supporting the high sensitivity of this ecosystem to climate variations.. Regional vegetation change generally matches the lake system development and is probably driven by climate-related thermokarst processes. However, the sub-millennial scale changes and variability differ for each proxy dataset, i.e. we inferred a lagged vegetation response and a non-linear lake system response to climate. This studies approach combining the regional vegetation signal and the more local lake catchment signal helps to understand the resolution of both reconstructed signals and highlights that a careful consideration of the scale of the reconstruction has to be made.

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