Lateglacial and Holocene environmental history of the central Kola region, northwestern Russia revealed by a sediment succession from Lake Imandra

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Lake Imandra at the southwestern border of the Kola Peninsula (NW Russia) is the largest lake in the Russian part of the European Arctic. However, the lake is located in the most heavily populated arctic region with up to 500 000 people living within the catchment area (Rzevsky 1997; Rumyantsev et al. 2012). Anthropogenic landscape transformation since the early AD 1920s has led to significant changes in the water and land ecosystems, which makes this site a prime location to study human impacts vs. a natural background state (Doncheva & Kalutskov 1977; Chidzikov 1980; Moiseenko & Yakovlev 1990; Kravtsova 1995; Innes & Oleksyn 1999; Rigina et al. 2000; Moiseenko et al. 2002, 2009a, b; Voinov et al. 2004; Daugwalter & Kashulin 2015, 2018).

The environmental history studied here is complex, in particular due to a distinct anthropogenic impact. While repeated glacialiations of the Kola region strongly influenced the lake throughout the Quaternary, a sediment succession from Lake Imandra was investigated by a hydro-acoustic survey followed by sediment coring down to the acoustic basement. The sediment record was analysed by a combined physical, biogeochemical, sedimentological, granulometrical and micropalaeontological approach to reconstruct the regional climatic and environmental history. Chronological control was obtained by 14C dating, 137Cs, and Hg markers as well as pollen stratigraphy and revealed that the sediment succession offers the first continuous record spanning the Lateglacial and Holocene for this lake. Following the deglaciation prior to c. 13 200 cal. a BP, the lake’s sub-basin initially was occupied by a glacialfluvial river system, before a proglacial lake with glaciallacustrine sedimentation established. Rather mild climate, a sparse vegetation cover and successive retreat of the Scandinavian Ice Sheet (SIS) from the lake catchment characterized the Bolling/Alleröd interstadial, lasting until 12 710 cal. a BP. During the subsequent Younger Dryas chronzone, until 11 550 cal. a BP, climate cooling led to a decrease in vegetation cover and a re-advance of the SIS. The SIS disappeared from the catchment at the Holocene transition, but small glaciers persisted in the mountains at the eastern lake shore. During the Early Holocene, until 8400 cal. a BP, sedimentation changed from glaciallacustrine to lacustrine and rising temperatures caused the spread of thermophilous vegetation. The Middle Holocene, until 3700 cal. a BP, comprises the regional Holocene Thermal Maximum (8000–4600 cal. a BP) with relatively stable temperatures, denser vegetation cover and absence of mountain glaciers. Reoccurrence of mountain glaciers during the Late Holocene, until 30 cal. a BP, presumably results from a slight cooling and increased humidity. Since c. 30 cal. a BP Lake Imandra has been strongly influenced by human impact, originating in industrial and mining activities. Our results are in overall agreement with vegetation and climate reconstructions in the Kola region.
Glaser 2007). Also the timing of the deglaciation in this region is not well known. A few existing surface exposure dates give maximum ages that have a wide range between 23 and 16 cal. ka BP (Stroeven et al. 2016). Consequently, the time of deglaciation until today is based on the interpolation between two prominent end moraines in the Karelian region and their equivalents on the interpolation between two prominent end moraines in the Karelian region and their equivalents on the

Ernest part of the Kola Peninsula, northwestern Russia (latitude 67°21′–68°04′N, longitude 31°52′–33°27′E; Fig. 1A). The lake has a surface area of 812 km² (880.4 km² including islands), and an indented shoreline of more than 2500 km (Rzewsky 1997; Rumyantsev et al. 2012). The catchment area of 12 300 km² equals roughly 15 times the lake surface (Fig. 1B). The relief is relatively flat over large parts of the catchment, with only 8% of the area exceeding 500 m above sea level (m a.s.l.) comprising parts of a mountain range with highest elevations up to 1200 m a.s.l. occurring in the east. The lake is fed by 1379 small tributaries, which partly flow through other smaller lakes. The only outlet is the Niva River, which drains to the south into the Kandalaksha Gulf of the White Sea (Moiseenko & Yakovlev 1990).

The present-day lake is dammed and the lake level is subject to seasonal fluctuations between 126.7 and 128.3 m a.s.l. The lake has average and maximum water depths of 14 and 67 m, respectively, and a volume of ~11 km³ (Elshin & Kupriyanov 1970). Lake Imandra is subdivided into three relatively isolated sub-basins, which are connected only via narrow straits (salmas). Lake Bolshaya Imandra (= Big Imandra) in the north (Fig. 1B – ②) is the main focus of this study. It has a surface area of 311.6 km², a maximum depth of 67 m, and contains, with a volume of 4.6 km³, the largest amount (42.2%) of water (Rumyantsev et al. 2012). Lake Ekostrovskaya Imandra in the central part (Fig. 1B – ③), with an area of 352.2 km² and a maximum depth of 42 m, and Lake Babinskaya Imandra in the west (Fig. 1B – ⑩), with an area of 148.7 km² and a maximum depth of 43.5 m, share the rest of the volume. Bolshaya Imandra, parts of Ekostrovskaya Imandra and the Niva River have a north–south orientation and are located in a deep north–south depression that separates the Kola Peninsula from the continental part of Fennoscandia (Dauwalter et al. 1999a). Babinskaya Imandra and the main parts of Ekostrovskaya Imandra, in contrast, have a west–east extension and are considered to be formed by glacial erosion (Herdendorf 1982).

Human-controlled open pit mining, smelters, and power plants led to the pollution of the lake and its catchment over the last century (Kraitsova 1995; Rigina 2002; Voinov et al. 2004; Moiseenko et al. 2018), which had a strong impact for example on the water chemistry of Lake Imandra (Rigina et al. 2000; Moiseenko et al. 2002; Dauwalter & Kashulin 2015). Until the early AD 1920s the lake was, in its natural state, an oligotrophic water body typical of arctic regions and characterized by a low ion concentration between 24–30 mg L⁻¹, water transparency of up to 8 m and pH values around 6.4–7.2 (Moiseenko & Yakovlev 1990; Moiseenko et al. 2002). From the AD 1920s onward, the lake experienced a rapid phase of eutrophication. The sum of ions rose to up to 82 mg L⁻¹, water transparency decreased down to less than 1 m, the pH shifted to more alkaline values (up to 8.1) and the aquatic habitat changed accordingly (Kravtsova 1995; Moiseenko et al. 1996, 2009b; Dauwalter et al. 1999b; Rigina 2002). Since the AD 1990s, environmental constraints have helped to improve the water quality. As a result, Lake Imandra has
reached a stable state, with the water transparency increased to several metres and pH values decreased to around 6.4–7.7, but still high ion concentration prevail (Moiseenko et al. 2009b, 2018).

The climate at Lake Imandra is a result of the influence of the North Atlantic Current and fairly mild considering its latitudinal position (Rumyantsev et al. 2012). Meteorological stations around the lake today record mean annual air temperatures between −0.6 and −1.0 °C, with monthly averaged July and January temperatures of 12.9 and −12.8 °C, respectively, and only 71–74 days per year with temperatures exceeding 10 °C (Zyuzin & Demin 2007; Merkel 2019). Between early November and end of May, Lake Imandra is covered by ice, which reaches thicknesses of 60–90 cm (Elshin & Kupriyanov 1970). The ice cover prevents wind-induced mixing of the water body, resulting in reduced gas exchange with the atmosphere. In April, microbial activity during organic material decomposition partly depletes the oxygen in the bottom waters and at the sediment surface. During the ice-free season wind- and temperature-induced mixing causes widely uniform temperatures throughout the water column (Dauwalter et al. 1999a; Ingri et al. 2011; Rumyantsev et al. 2012). Long-term observations that started in the AD 1960s suggest changes in regional climatic conditions, such as increasing temperatures and decreasing precipitation, which affect the duration of lake-ice cover, the length of the vegetation-growth period, and changes in vegetation (Johansen et al. 2005; Demin & Zyuzin 2006; Beck et al. 2007; Rubel & Kottek 2010; Ogureeva et al. 2012; Hogda et al. 2013; Mathisen et al. 2014; Marshall et al. 2016; Park et al. 2016).

The Imandra basin is located in the northern Taiga zone characterized by spruce–birch and pine–birch plant associations (Vasilevskaya 2014). The shorelines today are covered by sub-arctic birch (Betula pubescens) and various types of spruce–pine (e.g. Pinus) forests with mosses (Pleurozium schreberi), dwarf shrubs (Vaccinium myrtillus L., Vaccinium vitis-idea L., Vaccinium uliginosum L., Empetrum hermaphroditum Hagerup), and shrubs (mainly Betula nana L., but also Salix sp.). Particularly moist conditions result in low heat availability, and hence low evaporation, leading to widespread wetlands (Elshin & Kupriyanov 1970; Zyuzin & Demin 2007). The adjacent Khibiny Mountain massive at the eastern shore, however, exhibits an altitudinal transition in vegetation cover from coniferous forests at lower altitudes towards sub-arctic forests and tundra near the highest elevations (Orlova et al. 2012).

The bedrock in the area of Lake Imandra consists of crystalline formations of Archean and Early
Proterozoic age (Fig. 2; Mitrofanov et al. 1995). At
the northern shores of Bolshaya Imandra, granodi-
orites of Late Archean age crop out and at the
eastern and western shores, sedimentary rocks and
volcanic rocks are present, which are mostly covered
by Quaternary deposits. The lake basin is also directly
connected to the Khibiny Massif in the east, which
formed as result of Early Carboniferous intrusions of
alkali magma and is uniquely rich in ore deposits
and rare-earth elements (Radchenko et al. 1994; 
Pripachkin et al. 2013).

Material and methods

Hydro-acoustic survey and coring

In summer 2017, a hydro-acoustic survey was conducted
along a total of 80 km of profile lines in the central part of
the northern basin of Lake Imandra (Fig. 1C; Bolshaya
Imandra). For high-resolution reflection seismic a
Micro-GI airgun (2 × 0.1 L) and a 50-m-long Geometrics
GeoEel digital streamer (Geometrics, USA) with 32
channels and a spacing of 1.5625 m was used. Simultane-
ously, a parametric echo-sounding system (Innomar
SES-2000 compact, Germany), operated at 10 kHz, was
used. Measurements were conducted from a floating
platform (UWITEC, Austria) with an outboard engine
using a GPS device with an accuracy of a few metres for
navigation. The IHS Kingdom software© version 2016.1
was used for interpretation of the data sets. Depth
conversion was conducted assuming a constant seismic
velocity of 1430 m s⁻¹.

Based on the hydro-acoustic data, coring site Co1410
was selected at ~23 m water depth (67°42′56.76″N, 
33°5′6.42″E, Fig. 1C), where two sediment successions
with lengths of 7.98 and 8.65 m, respectively, were
recovered only a few metres apart. Coring was conducted
from a floating platform using a gravity-corer to retrieve
undisturbed near-surface sediments and a percussion
piston-corer with a 2-m-long core barrel for deeper
sediments (all UWITEC, Austria). After recovery, the
core segments were cut into up to 1-m-long sections and
kept at low temperatures and in the dark during
transport and storage.

Fig. 2. Simplified geological map of the
Lake Imandra area after Mitrofanov et al. 
(2001). The red line marks the lake’s catch-
ment area and the black box highlights the
study area shown in Fig. 1C.
Core description, XRF-scanning, and development of the core composite

In the laboratories of the University of Cologne, the sediment cores were split lengthwise and the core halves were described for sedimentary structures, grain size (finger probe), colour (Munsell Soil Colour Chart), consistency, carbonate content (reaction with hydrochloric acid (HCl)) and sediment composition (smear-slide microscopy). While one core half was stored as an archive, the working half was used for non-destructive magnetic susceptibility (MS) measurements at a resolution of 1 mm using a multi-sensor core logger (MSCL) equipped with a Bartington MS2E point sensor (Geotek 2016). High-resolution line-scan photography and subsequent X-ray fluorescence (XRF) scanning were carried out at 2-mm resolution with an ITRAX core scanner (Cox Analytical Systems, Sweden), equipped with a Cr-tube and a Si-drift detector in combination with a multi-channel analyser. Voltage and amperage were set to 30 kV and 55 mA, respectively, and the exposure time to 20 s. Prior to the measurement, the sediment surface was carefully flattened and covered with an Ultra-Polyester® thin-film (1.5 μm) (Chemplex, USA). The detected peak area intensities are given in total counts per second (cps) and can be used as semi-quantitative estimates of relative concentrations of the detected elements (Croudace et al. 2006). Post-processing of the data was conducted using the QSpec 6.5 software (Cox Analytical Systems, Sweden). Compared to wave dispersive XRF analyses (WD-XRF), XRF scanning is less time consuming, but more susceptible to inaccuracies resulting from variations in the sediment’s water content, surface structure, grain size, and porosity (Tjallingii et al. 2007; Hennekam & de Lange 2012) as well as a matrix effect, which has a greater impact on the data as organic matter (OM) content increases (Bertin 1975; Beckhoff et al. 2006). In order to compensate for these inaccuracies, the data are standardized to the element aluminium (Al), which is most suitable considering its geochemical behaviour (Löwemark et al. 2011).

Based on line-scan images, core description, and selected XRF data (Ca counts and Si/Ti ratio) the core segments were merged using the Corelyzer 2.0.4 software (National Lacustrine Core Facility LacCore and Continental Scientific Drilling Coordination Office, CSDCO). All depths refer to the spliced 8.46-m-long master record. A gap between 6.92 and 7.54 m results from missing data was conducted using the QSpec 6.5 software (Cox Analytical Systems, Sweden). Compared to wave dispersive XRF analyses (WD-XRF), XRF scanning is less time consuming, but more susceptible to inaccuracies resulting from variations in the sediment’s water content, surface structure, grain size, and porosity (Tjallingii et al. 2007; Hennekam & de Lange 2012) as well as a matrix effect, which has a greater impact on the data as organic matter (OM) content increases (Bertin 1975; Beckhoff et al. 2006). In order to compensate for these inaccuracies, the data are standardized to the element aluminium (Al), which is most suitable considering its geochemical behaviour (Löwemark et al. 2011).

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Subsampling and sedimentological analyses

The core composite Co1410 was continuously subsampled in 2-cm sections. Subsamples were freeze-dried and water content was calculated from the weight loss after freeze-drying. Subsequently, aliquots were taken for biogeochemical, granulometric and palynological analyses.

For biogeochemical analyses, aliquots of 0.5 g were ground to <63 μm and homogenized in a planetary mill (Fritsch, Germany). Bulk sediment concentrations of total nitrogen (TN) and total sulphur (TS) were determined using a Vario Micro Cube combustion CNS elemental analyser (Elementar Analysensysteme GmbH, Germany), whereas total carbon (TC) as well as total inorganic carbon (TIC) were measured using a DIMATOC 2000 (Dimatec Analysentechnik GmbH, Germany). Total organic carbon (TOC) was calculated from the difference between TIC and TC.

For grain-size analyses of the lithogenic material, aliquots of 1.5 g were treated with 15% hydrogen peroxide (H₂O₂), 10% HCl and 1 M sodium hydroxide (NaOH), in order to remove organic material, calcium carbonate and biogenic silica, respectively (Gee et al. 2002). The pretreated samples were then dispersed in a solution of about 60 mL deionized water with 1.2 g of sodium pyrophosphate (Na₄P₂O₇, 2.5 g L⁻¹, Graham’s salt), using a shaker for at least 12 h. The grain-size distribution of the remaining material was determined in three parallel runs with a laser diffraction particle size analyser LS 13 320 (Beckman Coulter GmbH, Germany), which was equipped with an aqueous liquid module (ALM) and an auto prep station (APS).

Statistical analyses

Processing of the grain-size data was carried out with the GRADISTAT v8 software (Blott & Pye 2001) based on averaged grain-size distributions of a triplicate measurement of each sample and the method by Folk & Ward (1957). On the processed data end-member modelling was carried out using the AnalySize algorithm (Paterson et al. 2016) run with the MatLab software (The MathWorks Inc. 2019).

Multivariate principal component analysis (PCA) was applied to selected XRF (Si, P, K, Ca, Ti, Mn, Fe, Rh, Sr, Zr, Si/Ti, Fe/Mn and Rb/Sr), biogeochemical (TN, TOC and TS) and end-member (EM1, EM2 and EM3) data, in order to create comparable data sets for different sections of the core composite and to explore the grouping of subsamples according to their similarities (Zitko 1994; Jolliffe 2002; Bro & Smilde 2014). PCA was performed using the Palaentological Statistics (PAST 3 v. 3.25) software (Hammer et al. 2001).

Pollen analysis

Pollen analyses were carried out at the St. Petersburg State University, Russia. Pollen and spores were extracted from 87 samples with 8–10 cm spacing in the upper part (0–5.16 m) and 2–4 cm in the lower part (5.16–7.82 m) of core Co1410. Analyses followed stan-
standard protocols (Savelieva et al. 2019), which included sample preparation with HCl, NaOH and hydrofluoric acid (HF) (Berglund et al. 1986), before they were mounted with glycerol on slides. In each sample between 70 and 600 terrestrial pollen grains were counted and identified using a stereomicroscope at 400× magnification. Taxonomic identifications followed Kupriyanova & Alyoshina (1972), Kupriyanova & Alyoshina (1978), Moore et al. (1991), Savelieva et al. (2013) and the modern pollen reference collection of the St. Petersburg State University. Calculated pollen percentages refer to the total sum of terrestrial pollen. Spore percentages are based on the sum of pollen and spores. Lycopodium tablets were used to determine pollen taxa concentrations (Stockmarr 1971).

**Cladocera analysis**

Sediment samples were prepared for cladoceran analyses at the Kazan Federal University, Russia using the methods described in Korhola & Rautio (2001). In total, 43 samples with a spacing of 2 cm in the upper 0.50 m and 4–14 cm down to 5.10 m were analysed. No samples have been analysed in the lowermost part of the core between 5.10 and 8.45 m. Each sample (0.5–0.6 g of dry sediment) was heated for approximately 30 min at 75 °C in 10% potassium hydroxide (KOH). In the next step the pretreated sediment mixture was sieved using a 50-μm mesh size. The sediment retained on the sieve was transferred with distilled water into a small 12.5-mL vial. A few drops of ethanol (C₂H₆O) and safranin-glycerine solution were added to prevent fungal growth and to stain the cladoceran remains. The chitinous remains of cladocerans (i.e. headshields, shells, post abdomens, postabdominal claws, and ephippia) were identified with reference to subfossil (Frey 1959, 1973; Sзероzyiska & Sarmaja-Korjonen 2007) and modern (Flößner 2000; Kotov et al. 2010) cladoceran identification keys, and their nomenclature follows Kotov et al. (2010). The most abundant body part of each species was chosen to represent the number of individuals. Each sample contained between 221 and 294 individuals, from which the percentages for all cladoceran taxa were calculated.

**Chronostratigraphical analyses**

In order to identify the time markers AD 1900 and AD 1886, according to the event-stratigraphical approach used for nearby Baltic Sea sediments (Moros et al. 2017), mercury (Hg) and radionuclide measurements were carried out at the Leibniz Institute for Baltic Sea Research in Warnemünde (IOW), Germany. The activities of artificial ¹³⁷Cs radionuclides were measured by gamma spectrometry with a CANBERRA BE3830 broad energy Germanium detector (Moros et al. 2017). Mercury measurements (Leipe et al. 2013) were carried out on subsamples between 20 and 100 mg with a DMA-80 analyser (MLS Company).

Moreover, radiocarbon (¹⁴C) dating on organic macro-remains was carried out on 12 samples. Wherever applicable, macro-remains were directly sampled from the sediment core after core opening and during subsampling. In most cases, however, they had to be isolated from subsamples of 1–4 g, which were sieved using mesh widths of 125, 63 and 36 μm. In the isolated fractions, larger organic remains were picked out by hand and smaller ones isolated by density separation. For the latter, a heavy-density solution consisting of sodium polytungstate (calibrated to 2.3 g cm⁻³) was added to the samples. After 15 min of centrifugation at 2500 rpm, the lighter material floating at the top was pipetted off the solution. This step was repeated three times. Finally, the extracted material was rinsed using deionized water to remove and dissolve any residual sodium polytungstate. The isolated bulk organic and terrestrial plant remains were treated according to the standard protocol of organic material pretreatment of organic material and graphitization (Rethemeyer et al. 2013). Samples were measured by accelerator mass spectrometry (AMS) using the radiocarbon dating method of the Cologne-AMS centre (Dewald et al. 2013). For the age-depth model a Bayesian-model approach was employed. In the upper part tie points from the sediment–water interface, ¹³⁷Cs and Hg measurements, and radiocarbon dating were included in the age-depth model. In the lower part, two tie points were derived for the Bolling/Allerød to Younger Dryas and for the Younger Dryas to Holocene boundaries according to Lohne et al. (2013), which are clearly reflected by the palynological data of core Co1410 (see below). The age-depth calculation was performed using the ‘rBacon’ library (Blauw & Christen 2011; Blauw et al. 2020) for R (R Development Core Team 2019). The software calibrates radiocarbon dates to calendar years before present (AD 1950) using the IntCal13 calibration curve for terrestrial Northern Hemisphere material (Reimer et al. 2013a, b), while negative ¹⁴C dates (modern/post-bomb) are calibrated by the NH1 post-bomb calibration curve for the Northern Hemisphere (Hua et al. 2013), all with an uncertainty range of 2σ. Pollen-derived ages were calculated with an estimated error of ±100 a, whereas the ¹³⁷Cs and Hg markers, and the sediment surface were calculated with errors of ±10 a and ±1 a, respectively.

The model was calculated by adding two boundaries (at 0.32 and 5.16 m), where prominent lithological changes suggest the memory of the age-depth calculation should be reset. Moreover, two mass movement deposits (6.50–6.70 and 7.92–8.46 m) were included with an infinite sedimentation rate. Additionally, paying attention to a prominent change in the lithology below 5.16 m, the accumulation mean and shape were adopted to a more rapid sediment accumulation. Based on the computed age-depth model the sedimentation rate was
estimated for those parts of the core that suggest a continuous sedimentation, using the same software package. The probability range was reduced in order to avoid unrealistically high or low sedimentation rate estimates.

**Results and discussion**

**Sediment architecture**

Sediments in the vicinity of the coring location Co1410 are characterized by a rather even reflector marking the sediment–water interface, but more rugged sub-bottom reflectors (Fig. 3). The top of the diffuse blue reflector marks the acoustic basement, which occurs at a depth of ~8 m below lake floor at site Co1410 and was penetrated by the sediment corer by about 0.5 m. The blue reflector shows short-distance relief changes in the order of 1–2 m in the surrounding ~100 m, but higher-amplitude changes of about 10 m beyond. The basement is overlain by a subparallel sediment unit of 2.0–2.5 m thickness that is acoustically stratified in more flat areas and widely transparent in steeper areas. This unit is bordered at the top by a strong acoustic impedance contrast (green reflector), which indicates sediments with lower densities above. These sediments are acoustically less stratified, with faint stratification visible in areas with a rather flat green reflector, but massive internal structure in areas with a steeper green reflector. Furthermore, the sediments differ from the underlying sediments by distinct thinning in the areas where the blue and green reflectors are elevated, leading to a strongly reduced relief at their top (reflector 3). Hence, sedimentation between the green and orange reflectors was predominant in depressions and widely levelled the lake floor around the coring site. The orange reflector finally is draped by a thin (<0.5 m) layer of acoustically transparent sediments, bordered by the modern lake floor (red reflector) in a water depth of ~23 m at coring site Co1410.

**Lithostratigraphy and biogeochemistry**

Based on the visual core description as well as the grain-size, XRF, and water content data the 8.46-m-long core composite Co1410 was subdivided into four main sediment units (I–IV), some of which are composed of subunits (Fig. 4).

**Unit I (8.46–7.85 m) – glacifluvial deposition.** – The lowermost part of the core consists of lithogenic sediments, with particle sizes varying between silt and very coarse gravel (Fig. 4). The diverse, partly very coarse grain size excluded representative XRF-scanning (Croudace et al. 2006) and water content determination and the almost pure lithogenic composition prevented elemental and micropalaeontological analyses (Figs 4, 5). Hence, the genetic interpretation of this sediment unit mainly relies on the core description, the shape of macroscopic grains and the hydro-acoustic characteristics.

The top of Unit I corresponds to the blue reflector, which marks the acoustic basement (Fig. 2). This is explained by the almost pure lithogenic composition and rather coarse grain size of Unit I. The lack of acoustically stratified sediments underneath suggests that Unit I either is rather thick or bordered at the base by impenetrable material such as glacial till or bedrock. In the latter cases, Unit I would most likely represent initial glacifluvial deposition that formed during the deglaciation of the basin. A glacifluvial formation of Unit I is supported by the scarcity of OM and the absence of clay. The latter excludes other depositional processes or formations such as a debrisflow or till. Moreover, the grain-size distribution from silt to gravel indicates a highly variable fluvial transport energy and likely redeposition of glacially supplied material.

Further evidence pointing to a glacifluvial deposition is provided by the shape and composition of the material, which indicate different transport distance and provenance, respectively. Angular nepheline syenites represent a proximal sediment supply to the coring site from the adjacent Khibiny Mountains at the eastern shore (Mitrofanov et al. 1995; Mitrofanov & Zozulya 2002), whereas subangular to rounded granodiorites as well as other magmatic and sedimentary rocks indicate a more distal sediment supply from the west (Fig. 2; Radchenko et al. 1994; Slabunov et al. 2006). A glacifluvial deposition is also supported by the different appearance of the boundaries of sedimentary units I and II in the two sediment cores, recovered just a few metres apart at site Co1410 (Fig. 6). Whilst one core exhibits a very sharp boundary between gravelly sand of Unit I and clayey silt of Unit II, the other core shows a more gradual transition, with a distinct fining upward succession over a few decimetres.

**Unit II (7.85–5.16 m) – glaciolacustrine deposition.** – Most parts of Unit II comprise laminated sediments, which are characterized by thin (3–35 mm) interbedding of grey silty fine sand and almost pure light greenish-grey clay. These rhythmic layers are interpreted as clastic varves, which were deposited as glaciolacustrine sediments in a proglacial lake. Some coarse-grained, gravel-sized particles, which are sporadically present in the sediment, probably represent ice-rafted debris. A high concentration of lithogenic material is indicated by elevated Ti/Al and K/Al ratios (Fig. 4; Kylander et al. 2011).

The absence of bioturbation is further evidence of clastic varve deposition during most of Unit II (e.g. Gromig et al. 2019). This also includes hostile conditions for endobenthic organisms, probably due to light limitation by lake-ice and snow cover during winter and to a high suspension load of the lake water in summer, and a
very low concentration of nutrients. The latter is indicated by the absence of terrestrial or aquatic macrophyte remains and very low amounts of finely dispersed OM, as revealed by the lowest recorded TOC (<0.2%) and TS (<0.02%) values of the sediment succession Co1410 (Fig. 4). Strong fluctuations in the TOC/TN ratio (2–27) suggest that OM experienced repeated and short-term changes in the proportions of terrestrial and aquatic sources (Meyers et al. 1995), and distinct decomposition. Pollen grains contribute to terrestrial OM supply; however, their low total concentration (<10 000 pollen grains g\(^{-1}\)) and low arboreal pollen (AP) to non-arboreal pollen (NAP) ratio (Fig. 5) indicate a sparse, herb-dominated vegetation cover of the landscape (Favre et al. 2008).

Throughout the clastic varve deposition of Unit II, differences in sediment composition reflect some gradual changes in the environmental conditions. Sediments of Subunit IIa are rather coarse-grained deposits, also reflected by low Rb/Sr ratios and EM1, which reflect a low clay and fine silt content. Medium silt or coarse silt and sand are represented by high Zr/Rb and EM2 or EM3, respectively (Figs 4, 7; Kalugin et al. 2007; Kylander et al. 2011). Rather high EM1 and EM2 as well as low EM3 might be associated with a high meltwater supply, which may occur in the proximity of a retreating ice sheet. A relatively high amount of Pinus pollen along with an enhanced AP/NAP ratio especially in the lower part of the Subunit (Fig. 5), might be a consequence of aeolian transport.

In Subunit IIb, between 6.70 and 6.50 m, the clastic varve deposition is interrupted by dark and massive, well-sorted coarse silt to fine sand (Fig. 4). As a consequence of the high proportion of coarse sediments,
this Subunit is demarcated from the other Subunits in the PCA (Fig. 7). It is also displayed in the hydro-acoustic profile as an acoustically transparent layer about 1 m above the blue reflector (Fig. 3), indicating its origin in a mass movement process (Lebas et al. 2019). The formation of this Subunit by a mass movement event is supported by a sharp, likely erosive lower and transitional upper boundary. The most probable mechanism of sediment deposition is a grain flow of pre-sorted sediment, for instance originating from a delta (Sauerbrey et al. 2013).

In the lowermost part of Subunit IIc, decreasing EM2 and medium silt contents at the expense of increasing EM1 values (Figs 4, 7) are probably attributed to a persisting proximity of the ice margin. Moreover, a broad increase in Artemisia pollen (Fig. 5) suggests that a sparse, herb-dominated vegetation established in a glacial setting. In the middle part of Subunit IIc, highest input of fine-grained material as reflected in EM1, Rb/Sr ratios and water content (Håkanson 1977; Kalugin et al. 2007; Kylander et al. 2011), indicates an initially high glacial meltwater supply, whereas in the upper part of Subunit IIc decreasing EM1 and Rb/Sr ratios suggest a decrease in transport energy, probably as a result of glacier retreat (Fig. 4). This is supported by a decreasing mean grain size (d50), MS, EM3, Zr/Rb ratios (Peck et al. 1994; Cohen 2003; Reynolds et al. 2004) and gradual decrease in varve thickness over the course of the upper part of Subunit IIc. Furthermore, a simultaneous decrease of Artemisia pollen indicates less favourable conditions for the growth of herbs (Fig. 5).

During deposition of Subunit IID, the grain-size spectrum shows a slightly coarser composition and little variation, as reflected in the grain-size end members as well as the Rb/Sr and Zr/Rb ratios (Fig. 4). This is likely attributed to elevated but rather constant meltwater supply from glaciers. A decreasing amount of Artemisia pollen (Fig. 5) indicates more favourable growing conditions for the vegetation in the catchment and a slight increase in the total pollen concentration (Fig. 5) probably points to lower sedimentation rates (Fig. 8). This correlates with the distinctly increasing biogenic accumulation at the expense of the lithogenic sediment.
supply, reflected by increasing TOC and TS values and decreasing Ti/Al and K/Al ratios (Figs 4, 7).

**Unit III (5.16–0.32 m) — lacustrine deposition.** — The impedance contrast of the green acoustic reflector (Fig. 3), marking the boundary between higher densities of Unit II and lower densities of Unit III, is predominantly controlled by a further increase in biogenic accumulation, rather than by grain size, which remains relatively constant at the transition. Higher OM accumulation is reflected by a higher TOC (~6%) and TS (~0.4%) in Unit III, along with lower Ti/Al and K/Al ratios (Fig. 4) and higher pollen concentrations (up to 530 000 pollen grains g⁻¹; Fig. 5). Enhanced productivity in Unit III is confirmed by higher Br/Al ratios, as a result of the capacity of Br to be independent of matrix effects and to form strong covalent bonds with organic molecules (Davies et al. 2015). Furthermore, significantly higher Si/Ti ratios along with constant Si counts suggest a higher in-lake productivity of diatoms (Brown et al. 2007; Melles et al. 2012), which is confirmed by a high abundance of diatoms in the microscopic smear slides.

The increased biogenic accumulation along with decreased lithogenic sediment supply during the deposition of Unit III indicates lacustrine conditions in the absence of a significant sediment supply from glaciers in the catchment. Brownish sediment colours and low Fe/Mn ratios suggest complete oxygenation of the water column (Fig. 4; Cuven et al. 2011). The appearance of varying amounts of dark brown to black streaks is related to reducing conditions of soils in the catchment. This is derived from an anti-correlation of Fe/Mn ratios and Fe count peaks, most obvious in the lower part of the unit (Fig. 4), which results from the leaching of Mn in the soils and formation of the black manganese oxides (e.g. MnO₂) in the lake under oxygenated bottom-water conditions (Mackereth 1966; Engstrom & Wright 1984; Renberg 1985; Bryant et al. 1997).
Throughout Unit III, the strongest changes in sediment composition occur in Subunit IIIa. There, all proxies indicating biogenic productivity (TOC, S, Br/Al, Si/Ti, pollen and diatoms) show strong increases. This reflects an increase in biogenic accumulation, which according to simultaneously slightly increasing TOC/TN ratios originates from terrestrial rather than from lacustrine OM (Meyers et al. 1995; Silliman et al. 1996). Within the pollen assemblages, an increase in Pinus pollen at the expense of Artemisia pollen along with an increase in the AP/NAP ratio (Fig. 5) indicate the emergence of a tree-prevalent landscape (Favre et al. 2008). Coarser grain sizes (Figs 4, 7) suggest an increased transport energy, although the decrease in the relative amount of lithogenic sediment supply, as reflected in reduced K/Al and Ti/Al ratios, can be attributed to increased soil stabilization. The cladoceran littoral-benthic chydorid assemblages, which comprise Alona affinis, Alona quadrangularis, Alonella nana and Alona intermedia, suggest the prevalence of relatively cool waters (Harmsworth 1968; Whiteside 1970; Korhola & Rautio 2001; Sarmaja-Korjonen 2002; Nevalainen et al. 2015; Bledzki & Rybak 2016). Indications for a fresh, oligotrophic water body come from the presence of the dominant taxa Bosmina (Eubosmina) cf. longispinulae, coregoni and Alona affinis as well as a low taxa diversity (Hofmann 1996). A low planktonic/littoral-benthic ratio of the cladocerans (Fig. 5), in particular in the lower part of Subunit IIIa, reveals widespread areas with shallower water depth, which likely are attributed to a low lake-level. This explains well the sediment becoming concentrated in deeper parts of the lake, which is evident in hydro-acoustic profiles for the lower part of Unit III (Fig. 3).

Subunit IIIb, which corresponds to an acoustically transparent layer at ~4.50 m at coring site Co1410 (Fig. 3), differs from Subunit IIIa by a lighter sediment colour and reduced occurrence of dark streaks. Lithogenic sediment supply, according to elevated Ti/Al and K/Al ratios, is slightly increased, whereas biogenic sedimentation, as revealed by lowered TOC and Br/Al values, is reduced (Figs 4, 7). A slight decrease in Pinus pollen as well as a distinct increase in the Cladocera Alona affinis towards the top of Subunit IIIb is likely attributed to less favourable conditions for vegetation growth and more favourable for cold-adapted cladocerans (Kattel & Sirocko 2011).

Subunit IIIc shows low Ti/Al and K/Al values, and a high level of organic deposition, predominantly reflected in high TOC and Si/Ti values (Figs 4, 7). This indicates reduced lithogenic sediment supply, which is supported by a denser vegetation cover in the catchment, as inferred from highest abundances of Pinus pollen, total pollen concentration, and AP/NAP ratios in the sediment record Co1410 (Fig. 5). A distinct decrease in the fine fraction, represented by EM1 (Figs 4, 7), may be associated with a strongly reduced glacial meltwater supply. The appearance of Alonella exigua and Alona guttata/Coronatella rectangula (Sarmaja-Korjonen et al. 2003; Nevalainen et al. 2013) as well as the appearance of Alonella excisa, Alona rusticata and Pleuroxus truncatus (Hann & Warner 1987) and the decrease of Alona intermedia and Alonella nana (Nevalainen et al. 2013) indicate favourable conditions for cladoceran development, which allowed for the appearance of many new taxa and maximal cladoceran species diversity (Hofmann 1987).

Subunit IIId shows only small differences to Subunit IIIc. Slightly reduced Si/Ti ratios can be at least partly explained by reduced Si concentrations (Fig. 4) and indicate lower aquatic productivity or slightly increased input of lithogenic material (Fig. 7). The latter is supported by a slight decrease of Pinus pollen, AP/NAP ratios, and total pollen concentration (Fig. 5), and increasing Zr/Rb and Zr/Al ratios, which indicate enhanced aeolian supply (Müller et al. 2001). However, an elevated Br/Al ratio (Mayer et al. 2007; Ziegler et al. 2008), and a decrease in the TOC/TN ratio suggest higher aquatic productivity or a reduced terrestrial OM supply.

Unit IV (0.32–0.00 m) – lacustrine deposition. – A prominent change in sediment colour and geochemical
Fig. 7. PCAs of Units II, III and IV showing different trends in response to changing environmental conditions and processes based upon geochemical data and grain-size distributions. Scatterplots of loading 1 (x-axis) vs. loading 2 (y-axis) are shown in the small plots in blue colour for each unit and the significances of component 1 (x-axis) and component 2 (y-axis) are specified for each unit.
composition is visible at the transition to sediments of Unit IV, and is also reflected in the low-impedance orange reflector close to the lake floor (Fig. 3). Despite its low thickness of only 0.32 m at coring site Co1410, Unit IV can clearly be divided into two Subunits of different composition and genesis.

Unit IVa consists of massive light olive grey, silty clay. Pronounced increases of MS and EM1 values, Ti/Al and K/Al ratios, and Ti counts (Figs 4, 7) are attributed to enhanced input of lithogenic material of particularly fine grain size. The constant occurrence of the Cladocera *Chydorus* spp., which is the most toxic-tolerant taxon, in combination with the disappearance of e.g. *Alonopsis elongata* and the appearance of *Leptodora kindtii*, indicates a high degree of eutrophication (Nevalainen et al. 2013, 2015). The disappearance of the cladocerans *Alona affinis* and *Acroperus harpae*, as indicators for cold-water and sub-arctic conditions (Harmsworth 1968), may reflect rising temperatures in the water body. Significantly reduced Si/Ti and Br/Al ratios as well as TOC values are probably ascribed to dilution effects as a result of enhanced lithogenic input (Leng et al. 2012); however, they might also reflect a decrease in biogenic accumulation.

Compared to the distinct transition at the lower boundary of Subunit IVa, its change to Subunit IVb is
more gradual. Subunit IVb covers the topmost 0.20 m of core Co1410 and is characterized by greyish brown to light olive brown, massive to weakly stratified organic mud. Most proxies change in the direction of Unit III values, thus suggesting a modification of the lake’s ecosystem. For instance, somewhat decreasing values of MS, EM1, Ti/Al, K/Al and Ti counts in Subunit IVb, simultaneous with increasing grain size Δ50 and EM2 (Figs 4, 7), suggest lower input of fine-grained lithogenic sediment from the catchment. Additionally, aeolian input probably became reduced, as suggested by a decrease in Zr/Al (Fig. 4). Increasing TOC and Si/Ti point to a decreasing influence of lithogenic dilution, but may also represent an increase in in-lake bioproductivity. This may have caused short-term oxygen depletion of the bottom water, as indicated by a distinct peak in TS (Fig. 4). Persistent eutrophication is indicated by the replacement of Cladocera (Fig. 4). Persistent eutrophication is indicated by the replacement of Cladocera Bosmina (Eubosmina) by Cladocera Bosmina (Bosmina) longirostris and Leptodora kindtii (Fig. 5; Bjerring et al. 2009; Brancelj et al. 2009).

Furthermore, taxa common in cold northern lakes, like Paralona pigra, Alonopsis elongata, Rhyncchotalona falcata and Eurycercus spp., are or become absent (Nevalainen et al. 2013). This, together with the overall lowest species diversity in the sediment succession, confirms the persistence of rather exceptional environmental conditions, which are suitable only for specialized taxa (Goulden 1964; Hofmann 1978; Korhola & Rautio 2001).

Chronology
The Bayesian age-depth model for the core composite Co1410 is based on radiocarbon dates as well as tie points from the undisturbed sediment surface, 137Cs and mercury data, and the pollen stratigraphy (Table 1; Fig. 8).

For the age-depth calculation, glacial-fluvial Unit I has been treated as an event layer (Fig. 8), because we assume a very rapid deposition in a fluvial system during deglaciation. Accordingly, the presumed grain-flow deposit incised into the glaciolacustrine Unit II (Subunit IIb) was treated as an event layer, originating in a mass movement. The age-depth model for the remainder of Unit II is based on pollen-derived ages of the Bølling/Allerød and Younger Dryas chronozones (Fig. 5), which refer to the chronozone boundary dated in northern Europe by Lohne et al. (2013) to 12 710±52 cal. a BP. The ages suggest, first, that the grain-flow deposition was associated with only little if any erosion, since the base of the Bølling/Allerød interstadial is clearly preserved in the lower part of Unit II. Second, the ages indicate that the radiocarbon dates obtained from Unit II are clearly too old, being biased by old organic carbon supplied from the catchment during deglaciation (Abbott & Stafford 1996; Björck et al. 2001). The decreasing offset of pollen and radiocarbon-derived ages towards the top of the unit probably reflects a reduced admixture of old organic carbon in the very low-productive lake during this period.

The sedimentation rates derived from the age-depth model show distinct variations. During the formation of the glaciolacustrine deposits of Unit II, sedimentation rates varied between 1.8 and 3.0 mm a⁻¹ (Fig. 8). For the lacustrine deposits of Unit III the model shows much lower sedimentation rates between 0.5 and 0.9 mm a⁻¹ and relatively constant sedimentation, probably because of the deduced retreat or absence of glaciers in the catchment. The rapid and strong increase in sedimentation rates of Unit IV to values between 5.4 and 5.6 mm a⁻¹ indicates a distinct shift in the environmental conditions.

Climatic and environmental history

Lateglacial (>13 200–c. 11 550 cal. a BP)

The glacioulvial and glaciolacustrine deposits of Units I and II in core Co1410 (Figs 3–7) were formed during the Lateglacial between >13 200 and 11 550 cal. a BP (Fig. 8). According to the age-depth model, Unit I (Fig. 9A) and Subunit IIa (Fig. 9B) are attributed to the Bølling/Allerød chronozone, when relatively warm climatic prevailed (Björck et al. 1998). In contrast, Subunits IIb, IIc and IIId (Fig. 9C) are associated with the Younger Dryas chronozone, when relatively cool climatic conditions are reported (Lohne et al. 2013). According to the chronostratigraphical data derived from the pollen core Co1410, the last ice-sheet coverage of the Imandra basin has to pre-date 13 200 cal. a BP. This is in good agreement with large-scale ice-sheet reconstructions for the LGM, which indicate that the SIS had inundated the Imandra region and reached its maximum extent around 20–18 cal. ka BP several hundred kilometres further to the east (Svendsen et al. 2004; Demidov et al. 2006; Hughes et al. 2016).

Moraines occurring in the vicinity of Lake Imandra are, however, mainly attributed to a local mountain glaciation in the Khibiny Mountains rather than to the SIS (Kolka & Korsakova 2005; Hättestrand et al. 2008; Kolka et al. 2008; Yevzerov & Nikolaeva 2008), and a few exposure ages derived recently by cosmogenic nuclide dating show a wide range between 16 and 11 ka BP (Stroeven et al. 2016). Partly ice-free conditions in the region already in the Bølling/Allerød chronozone are also indicated by studies of ice sheets and mountain glaciations in the Khibiny Mountains to the southeast of the coring location (Yevzerov & Nikolaeva 2008). The lake sediment cores hitherto retrieved in the region do not reach back to the time of initial deglaciation (Solovieva & Jones 2002; Solovieva et al. 2005; Olyunina et al. 2008; Ilyashuk et al. 2013; Tolstobrova et al. 2016; Shilova et al. 2019). As a consequence, reconstructions of the ice coverage in the Imandra region were for a
long time based on large-scale interpolations, which indicated a deglaciation during the Younger Dryas chronozone (Lundqvist & Saarnisto 1995; Kleman et al. 1997; Boulton 2001; Svendsen et al. 2004; Hughes et al. 2016).

The time frame for the deglaciation proposed by these studies partly agrees with the results from core Co1410 from Lake Imandra. The initial, relatively early ice retreat is supported by the Rugozero and Kalevala moraines in the Karelian region and their equivalents at the northern coast of Kola Peninsula, which reflect the retreat of the SIS (Ekman et al. 1991; Rainio et al. 1995). The onset of continuous sediment accumulation in these regions, which suggests prevailing ice-free conditions, is dated to 14–13 cal. ka BP (Snyder et al. 1997, 2000; Kremenetski et al. 2004; Korsakova et al. 2016). This provides information about the timing of deglaciation of the Imandra region and thus suggests that the history reflected by core Co1410 starts only a few centuries after deglaciation of the site.

The depositional setting of the initial sediments of Unit I at coring site Co1410 most likely can be described as a braided river system, which was embedded within one of the larger, north–south orientated meltwater channels that developed in the Imandra region during the Lateglacial (Kleman et al. 1997; Stur 2006a, b). These channels are up to several hundred metres wide and between a few decimetres and several metres deep (Kolka & Korsakova 2005; Kolka et al. 2008). The spillway is cut perpendicular to the flow direction in the west–east bounding hydro-acoustic profile (Fig. 3). There, the relief of the braided river at the end of the glaciﬂuvial deposition may be reﬂected by the irregular shape of the blue reﬂector at the top of Unit I. This irregularity might be attributed to short-term and small-scale changes in braided river deposition (Williams & Rust 1969; Bridge et al. 2009), and is the reason for the different transitions between sediments of Unit I and Unit II (Fig. 6), whereof gradual transitions are attributed to riverbeds and abrupt changes to riverbanks.

Still during Bølling/Allerød times, glaciﬂuvial sedimentation at coring site Co1410 passed over to glaciolacustrine sedimentation in a proglacial lake (Figs 8, 9). Pollen assemblages suggest a sparse vegetation cover and rather mild climate (Fig. 10), which is also documented further northward in coastal areas (Snyder et al. 2000; Corner et al. 2001). During the Younger Dryas chronozone, climate cooling and decreasing precipitation led to a decrease in vegetation cover and a re-advance of the SIS in Lake Imandra’s catchment, which is reﬂected in changes of pollen, grain size, and MS (9; 10). Similar developments were reconstructed for the Younger Dryas along nearly all ice-marginal positions around the SIS (Lundqvist & Saarnisto 1995; Rainio et al. 1995; Demidov et al. 2006; Mangerud et al. 2011; Putkinnen et al. 2011). The grain-ﬂow deposit incised into the glaciolacustrine sediments of core Co1410 (Subunit IIb; Figs 4, 7) may have resulted from a high sediment supply during deglaciation, as observed for sub-recent glacier-induced mass movement deposits in the northern part of the Khibiny Mountains (Shilova et al. 2019).

The deglaciation of the Kola Region during the Lateglacial was associated with a marine transgression. This is reﬂected in brackish or marine sediments, which frequently overlay glacial diamictons along the northern coast of the Kola Peninsula and in the White Sea area (Corner et al. 1999; Snyder et al. 2000; Grøsfjeld et al. 2006; Korsakova et al. 2016). Even less than 30 km southwest of site Co1410 diatom analyses suggest a marine ingression during Lateglacial times (Tolstobrova et al. 2016). However, there is no such evidence at coring site Co1410. The Lateglacial sediment at this site lacks

| Sample ID | Composite depth (m) | Fm | C(μg) | δ13C (‰) | 14C age (a BP) | Calendar age (cal. a BP) | Error (a) |
|-----------|---------------------|----|-------|----------|---------------|-------------------------|-----------|
| Surface   | 0                   |     |       |          |               | –68                     | 1         |
| Co1410_26-28 | 0.27               | 137Cs | 0.840±0.008 | 27 | 1398±81 | 1299 | 122 |
| Co1410_40-42 | 0.41               | Hg  | 0.522±0.006 | 28 | 5227±92 | 6033 | 180 |
| Co1410_628-630 | 0.41               | Radiocarbon (OMR) | 0.414±0.004 | 31 | 7080±86 | 7787 | 160 |
| Co1410_516-518 | 5.17               | Pollen | 0.369±0.002 | 997 | –33.7 | 8017±51 | 9023 | 322 |
| Co1410_631-632 | 0.27               | Radiocarbon (T) | 0.296±0.003 | 31 | 9769±92 | 11231 | 166 |
| Co1410_40-42 | 0.41               | Radiocarbon (OMR) | 0.287±0.003 | 36 | 10032±93 | 11540 | 291 |
| Co1410_516-518 | 5.17               | Pollen | 0.287±0.003 | 36 | 10032±93 | 11540 | 291 |
| Co1410_628-630 | 0.29               | Pollen | 0.203±0.006 | 147 | 1.3 | 12792±231 | 15105 | 830 |
| Co1410_40-42 | 0.41               | Radiocarbon (OMR) | 0.122±0.005 | 156 | –10.4 | 16869±360 | 20434 | 917 |
| Co1410_628-630 | 0.27               | Pollen | 0.106±0.005 | 272 | –5.4 | 18003±374 | 21733 | 910 |
marine shell fragments, and very low S contents (Fig. 4) clearly argue against brackish or marine conditions (Berner & Raiswell 1984).

**Early Holocene (c. 11 550–8400 cal. a BP)**

The change from glaciolacustrine to lacustrine conditions in Lake Imandra is reflected by the onset of Unit III deposition (Figs 4, 5), which is dated to c. 11 550 cal. a BP (Fig. 8) and about 100 years delayed compared to supra-regional records (Stroeven et al. 2006) as well as the NGRIP ice-core record (Walker et al. 2009; Rasmussen et al. 2014). However, this age is in good accordance with regional records to the west of Lake Imandra and at the northern coast of Kola Peninsula, which show the same characteristics (Snyder et al. 2000; Solovieva & Jones 2002; Kremenetski et al. 2004), and coincides very well with the Younger Dryas chronozone; D = retreat of the SIS from the palaeo-catchment area until the beginning of the Early Holocene under the presence of mountain glaciers and establishment of forests around the lake; E = completely ice-free conditions in the catchment during the Middle Holocene and migration of the tree line to high altitudes; F = build-up of small mountain glaciers in the Late Holocene and migration of the tree line downward; G = retreat of the mountain glaciers nowadays and sediment deposition under the influence of anthropogenic activities in the catchment area. The block diagrams are based on ASTER Global Digital Elevation Model (GDEM) Version 3 (NASA/METI/AIST/Japan Space systems 2001), including ASTER Global Water Bodies Database (ASTWBD) Version 1 (NASA/METI/AIST/Japan Space systems 2019) to correct elevation values of water body surfaces.

**Fig. 9.** W–E trending schematic block diagrams illustrating the environmental history in the vicinity of the coring site Co1410 (yellow circle). White colour indicates the SIS and local mountain glaciation in the west and east, respectively. Lake bathymetry and sediment architecture are derived from the seismic profile presented in Fig. 3 and suggest the different stages of sediment deposition. A and B = ice retreat and transition from glaciifluvial sedimentation in a braided river setting to glacial lacustrine sedimentation in a proglacial lake during the Bolling/Allerød interstadial; C = glacial re-advances during the Younger Dryas chronozone; D = retreat of the SIS from the palaeo-catchment area until the beginning of the Early Holocene under the presence of mountain glaciers and establishment of forests around the lake; E = completely ice-free conditions in the catchment during the Middle Holocene and migration of the tree line to high altitudes; F = build-up of small mountain glaciers in the Late Holocene and migration of the tree line downward; G = retreat of the mountain glaciers nowadays and sediment deposition under the influence of anthropogenic activities in the catchment area. The block diagrams are based on ASTER Global Digital Elevation Model (GDEM) Version 3 (NASA/METI/AIST/Japan Space systems 2001), including ASTER Global Water Bodies Database (ASTWBD) Version 1 (NASA/METI/AIST/Japan Space systems 2019) to correct elevation values of water body surfaces.
tation, suggesting that the ice mass mainly responsible for this, the SIS, had retracted from the lake’s catchment and mountain glaciers are strongly reduced in size (Figs 4, 9, 10). However, the latter may have persisted, as indicated by a delayed onset of full lacustrine conditions in small lakes in the Khibiny Mountains (Olyunina et al. 2008; Shilova et al. 2019). Changes in the cladoceran assemblages in core Co1410 are characteristic of the transitions from glacial to postglacial conditions (Korhola et al. 2001; Bledzki & Rybak 2016). Furthermore, they suggest that the lake level of Lake Imandra at the onset of the Early Holocene was low (Fig. 10), coincident with lake-level drops in other large basins on the Kola Peninsula (Shvarev 2003; Olyunina et al. 2008).

Pollon data of core Co1410 suggest that rather sparsely covered landscapes were successively replaced by denser vegetation in a warming and moister climate during the course of the Early Holocene (Fig. 10). This change is also recorded in other areas of northwestern Russia (Snyder et al. 2000; Arslanov et al. 2001; Wohlfarth et al. 2004; Ilyashuk et al. 2013; Nikolaeva et al. 2015), and matches with a pronounced shift to higher surface temperatures in the Norwegian Sea around 9300 a BP (Koc et al. 1993), when moist and warm North Atlantic air masses replaced high Arctic air masses (Hammarlund et al. 2002; Jones et al. 2004).

In Subunit IIIa (Fig. 8; 11 550–8850 cal. a BP; Fig. 10) more short-term climate variations are indicated by some of the high-resolution biogeochemical data. Peaks in proxies reflecting lithogenic supply (e.g. Ti/Al, K/Al, Zr/Al; Figs 4, 10) are probably attributed to cooling events, such as those recorded in a number of North Atlantic marine, terrestrial and ice-core records at 11 400, 10 400–10 300 and c. 9400 cal. a BP (Björck et al. 1997, 2001; Bond et al. 1997; Nesje et al. 2001; Yu & Wright 2001; Rasmussen et al. 2014). These climatic variations may also have led to unstable redox conditions in Lake Imandra, as suggested by short-term variability in redox-sensitive geochemical proxies (e.g. Fe/Mn; Fig. 10). Such changes, although detected with limited age control, were also determined in other studies in the Kola region and attributed to a transitional phase towards more stable lacustrine conditions (Snyder et al. 1997; Corner et al. 1999, 2001).

In Subunit IIIb (Fig. 8; 8850–8400 cal. a BP) the sedimentological and palynological proxies (e.g. Ti/Al, Si/Ti, TOC, grain size, Pinus; Fig. 10) indicate cooler climatic conditions. A potential correlation of this cooling with the 8.2 ka cooling event (Alley et al. 1997; Magny et al. 2003; Rohling & Pälike 2005) would imply a dating uncertainty in the order of several hundred years in this part of the record. The 8.2 ka event was identified in a sediment record to the east of Lake Imandra by a chironomid-based temperature reconstruction (Fig. 10; Ilyashuk et al. 2013), but is not documented in the multiproxy study on a lake sediment core to the west (Jones et al. 2004). The latter supports the interpretation of a decreasing significance of Atlantic climatic signals from Norwegian coastal areas to the central Kola region, which was deduced from pollen records (Korhola et al. 2004; Solovieva et al. 2005; and Seppä et al. 2007).

**Middle and Late Holocene (8400–30 cal. a BP)**

The lacustrine Subunits IIIc and IIIId in core Co1410 were deposited in the periods 8400–3700 and 3700–30 cal. a BP (Fig. 8), which are assigned to the Middle and Late Holocene, respectively.

The Middle Holocene was, according to the pollen and cladoceran data of core Co1410 (Figs 5, 10), characterized by a particularly warm climate, which led to a dense vegetation in the catchment of Lake Imandra and probably the disappearance of glaciers in the adjacent Khibiny Mountains (Fig. 9E). The Holocene Thermal Maximum (HTM) recorded between 8000 and 4600 cal. a BP at Lake Imandra is slightly delayed compared to other reconstructions of the HTM in the central Kola region, which are based on micropalaeontological analyses that confine this period to 9000–5000 cal. a BP (Solovieva & Jones 2002; Kremenetski et al. 2004; Ilyashuk et al. 2013). However, it is within the overall range of HTM reconstructions in northwestern Europe, for instance in southern Finland from 7500–4000 cal. a BP (Ojala et al. 2008), in the northeastern European Russian Arctic from 8000 to 3500 cal. a BP (Salonen et al. 2011), in Latvia from 8000 to 4500 cal. a BP (Heikkila & Seppä 2010) and in north Sweden from 9500 to 3400 cal. a BP (Meyer-Jacob et al. 2017). The differences in the timing of the HTM could be a result of supra-regional shifts; however, the lack of obvious spatial trends rather suggests inaccuracies in the climate reconstructions and chronological uncertainties.

Rather small variations in almost all physical and geochemical proxies in the Middle Holocene deposits of core Co1410 (Figs 4, 10), compared to the Early Holocene, suggest prevailing stable climatic conditions. This supports the suggestion that the climate in Fennoscandia during the HTM was controlled by a higher insolation and rather stable anti-cyclonic conditions during summer as compared to today (Seppä & Birks 2001; Bakke et al. 2005; St. Amour et al. 2010).

For the Late Holocene, the proxies of core Co1410 suggest a slight cooling, with a somewhat reduced vegetation cover, increased input of lithogenic material, including aeolian supply, and an increase in aquatic production. The cooling may have been associated with an increase in humidity, as indicated by the development of widespread wetlands in the area simultaneous with retreating forests (Hammarlund et al. 2004; Weckstrom et al. 2010; Valiranta et al. 2011). Micropalaeontological investigations of lake archives in the vicinity of Lake Imandra also document wetter conditions during this period (Jones et al. 2004; Solovieva et al. 2005; Ilyashuk et al. 2013).
et al. 2013; Nikolaeva et al. 2015). Both the cooling and the increased humidity may have their origin in a weakening of the prevailing anti-cyclonic circulation patterns (Seppä & Birks 2001, 2002; Wanner et al. 2008), and may have supported the reactivation of local mountain glaciers in the vicinity of the lake, such as those occurring today in the Khibiny Mountains (Fig. 9; Demin & Zyuzin 2006; Khromova et al. 2014). A reoccurrence of glaciers was also reconstructed for the Late Holocene in northern Sweden (Snowball & Sandgren 1996; Nesje et al. 2001).

**Anthropogenic influence (since 30 cal. a BP)**

The lacustrine deposits of Unit IV in core Co1410 reflect the human impact on Lake Imandra’s catchment during the past c. 100 years.

Subunit IVa reflects enhanced input of lithogenic material of particularly fine-grained particles along with an increase in aeolian sediment supply and a change in cladoceran assemblages (Fig. 10). These changes can be traced back to human impact in the course of an intensification of land use since the end of the 19th century, and especially to industrial and mining activities in the catchment area as well as direct wastewater dumping into the lake since the early AD 1920s (Dauwalter et al. 1999b; Moiseenko et al. 2002, 2009b; Rigina 2002; Voinov et al. 2004). Enhanced sediment supply is obvious from studying aerial photographs taken in the AD 1960s that show the distribution of a sediment fan in Bolshaya Imandra originating from the mouth of Belaya River (Kravtsova 1995). Sediment deposits in the southeastern part of Bolshaya Imandra, in the vicinity of the river mouth, exceed a thickness of 8 m (Chidzikov 1980).

Fig. 10. Sedimentary units and selected physical property, geochemical, pollen, and cladoceran data from core Co1410 plotted against calibrated age (y-axis). Note that pollen, cladoceran and end-member calculations are presented in % and partly share one x-axis. The data are compared to the local Scandinavian Ice Sheet (SIS) and mountain glacier histories reconstructed here (bluish shadings) as well as the local July temperature from Ilyashuk et al. (2013), the LOESS-smoothed record (span = 0.20, order = 1; grey line), and the $δ^{18}O$-derived annual temperature from the GISP2 ice-core (Cuffey & Clow 1997; Alley 2000) using the calibration from Cuffey et al. (1995) (violet line).
The anthropogenically induced modifications of the lake water characteristics are also evident in various studies on fishes and smaller organisms (Chidzikov 1980; Moiseenko et al. 2002, 2009a; Lukin 2013).

The transition from Subunits IVa to IVb reflects the reduction of fine-grained lithogenic sediment input and aeolian supply. The level of pollution and trophic state is also reduced but remains elevated compared to the natural state. These changes are associated with enhanced environmental protection and reduced industrial production (Fig. 7; Voinov et al. 2004), as well as the build-up of a dam in the AD 1960s, which according to aerial photographs has efficiently reduced industrially induced sediment input (Rigina 2002). Besides shifts in the anthropogenic impact, pollen spectra in core Co1410 (Fig. 10) indicate a warmer and drier climate that is also recorded since the AD 1950s by long-term meteorological observations (Demin & Zyuzin 2006; Marshall et al. 2016), and an upward shift of the tree line of around 100 m in the Khibiny Mountains (Fig. 9; Demin & Zyuzin 2007; Mathisen et al. 2014), as well as changes in the growing seasons in Lake Imandra’s catchment (Karlsen et al. 2006; Hogda et al. 2013).

Conclusions

Hydro-acoustic data and different proxies from an 8.46-m-long sediment succession with a basal age of 13 200 cal. a BP from Lake Imandra allow us to draw the following conclusions concerning the regional environmental and climatic history:

- The hydro-acoustic data show the existence of large channels, which can be attributed to meltwater channels that formed large glacial-flooding systems during the initial deglaciation of the area prior to 13 200 cal. a BP. According to the sediment core data, glaciolacustrine deposits filled these channels successively until c. 11 550 cal. a BP and drape the bottom morphology. Subsequently, lacustrine sediments levelled the underlying topography until recent times.

- Pollen assemblages indicate the initial deglaciation during the Bølling/Allerød chronozone. At least one glacial advance is documented during the Younger Dryas chronozone. Another, smaller advance of glaciers in the Khibiny Mountains in the Early Holocene is indicated in the physical properties of grain size and MS, as well as the geochemical proxies Rb/Sr and Zr/Rb. They reflect changing energetic levels in the lake basin in dependence on the proximity of the ice margin or the existence of small ice caps in the mountain regions.

- With the onset of the Holocene at c. 11 550 cal. a BP, greyish glaciolacustrine sediment deposition changed to brownish lacustrine sedimentation and the pollen assemblages indicate a shift from cold glacial to warmer interglacial temperatures. Increasing diversity of the cladoceran assemblages and higher amounts of TOC can be attributed to a higher productivity in the lake and its catchment, respectively.

- A short-term cold event can be reconstructed from a decrease of TOC and Si/Ti between 8800 and 8400 cal. a BP, which can be attributed to the 8.2 ka cooling event within the limits of our age-depth model. The Holocene Thermal Maximum (HTM) is recorded between 8000 and 4600 cal. a BP and indicated by the highest concentration of Pinus pollen and highest AP/NAP ratios of the sediment succession. During the later Holocene between 4600 and 30 cal. a BP, a slight cooling and increased aeolian activity is documented in a decrease of Pinus pollen and higher Zr/Al ratios, respectively.

- During the last century, the dramatic influence of industrial activity in the catchment area is documented in a sharp lithological shift. A rapidly increasing sediment supply is a consequence of mining activities and wastewater dumping. Biogeochemical proxies and cladoceran assemblages indicate high levels of pollution and eutrophication of the water column.

- Our results show that biogeochemical and micropaleontological-based climatic and environmental history reconstruction agrees well with other studies from the Kola Peninsula highlighting vegetation and climatic changes in this area. Moreover, the record from core Co1410 specifies the deglaciation pattern for the central Kola Peninsula, which previously was based on wide interpolations and only a few dates. Within the uncertainty of the age model, the Holocene palaeoenvironmental history recorded in Lake Imandra is in general agreement with observations made in nearby Lakes Chuna and Kupal’noe. However, all lakes react slightly differently to climatic events and seem to be influenced by local effects. Size and geographical position should be taken into consideration when studying lake archives and comparing them with each other.

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Author contributions. – ML, GF, VK, BW and MMDe measured the study ML, LS, LF, AC, MMo, NK, and NN performed the analyses. ML wrote the paper with the figures with input from all co-authors.

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