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Multiple drivers of Holocene lake level changes at a lowland lake in northeastern Germany

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Many German lakes experienced significant water level declines in the last decades that are not fully understood due to the short observation period. At a typical northeastern German groundwater-fed lake with complex basin morphology, an acoustic sub-bottom profile was analysed together with a transect of five sediment cores, which were correlated using multiple proxies (sediment facies, μ-XRF, macrofossils, subfossil *Cladocera*). Shifts in the boundary between sand and mud deposition were controlled by lake level changes, and hence, allowed the quantification of an absolute lake level amplitude of ~8 m for the Holocene. This clearly exceeded observed modern fluctuations of 1.3 m (AD1973-2010). Past lake level changes were traced continuously using the Calcium-record. During high lake levels, massive organic muds were deposited in the deepest lake basin, whereas lower lake levels isolated the sub-basins and allowed carbonate deposition.

During the beginning of the Holocene (>9700 cal. a BP), lake levels were high likely due to final melting of permafrost and dead-ice remains. The establishment of water use intensive *Pinus* forests caused generally low (3-4 m below modern) but fluctuating lake levels (9700-6400 cal. a BP). Afterwards, the lake showed an increasing trend and reached a short-term high-stand ~5000 cal. a BP (4 m above modern). At the transition towards a cooler and wetter late Holocene, forests dominated by *Quercus* and *Fagus* and initial human impact likely contributed more positively to groundwater recharge. Lake levels remained high between 3800 and 800 cal. a BP, but the lake system was not sensitive enough to record short-term fluctuations during this period. Lake level changes were recorded again, when humans profoundly affected the drainage system, land cover and lake trophy. Hence, local Holocene water level changes reflect feedbacks between catchment and vegetation characteristics and human impact superimposed by climate change at multiple temporal scales.

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The hydrological system is a vulnerable landscape component showing considerable variability in different areas of the world (Allen & Ingram 2002; Gronewold & Stow 2014), which complicates predictions in the context of future climate change (Sherwood & Fu 2014). During the last 30 to 50 years, groundwater-fed lakes in northeastern Germany showed a significant lowering (Germer *et al.* 2011; Kaiser *et al.* 2014a), which was linked to decreasing water availability in response to global climate change (Natkhin *et al.* 2012). However, lake levels have started to rise again since ~AD2008, in contrast to predictions (Natkhin *et al.* 2012). This unexpected development revealed our incomplete understanding regarding regional hydrological responses to global environmental changes, calling for additional palaeohydrological studies.

Depending on the natural configuration of a landscape, different archives provide proxy information on components of past water cycles; for example, lakes provide lake level proxies indicating palaeo-water balances. Morphological-sedimentological features within lake basins provide discrete but absolute lake level reconstructions (Dietze *et al.* 2010, 2013). Continuous reconstructions from profundal lake sediment cores use relationshipships between water depth and certain biological proxies (Hannon and Gaillard, 1997; Korhola *et al.* 2000; Laird *et al.* 2011). Also authigenic carbonate has been used as a relative lake level proxy (Haberzettl *et al.* 2005; Pompeani *et al.* 2012).

Multiple cores collected along transects from littoral to profundal sites enable the reconstruction of lake level changes (Digerfeldt 1986; Magny 2004; Pribyl & Shuman 2014). In the European Alps, high detrital carbonate input into lakes has allowed absolute and highly-resolved lake level reconstructions using changes in precipitated carbonate morphologies related to water depth (Magny 2004). In temperate humid areas with low detrital carbonate input, complementary biological and sedimentological proxies have been related to lake level changes, e.g. macrofossils and the position of the sediment limit, i.e., the upper limit for the predominant deposition of organic-rich mud related to sediment focusing (Digerfeldt 1986; Dearing 1997). The “Digerfeldt approach” has provided reliable lake level reconstructions across the globe (e.g. Punning *et al.* 2004; Shuman *et al.* 2010; Haig *et al.* 2013), but so far has been restricted to shallow lakes <100 ha with simple basin morphologies (e.g. low-angled slopes <4%; Digerfeldt 1986; Pribyl & Shuman 2014).

In the central European lowlands, hydrological reconstructions primarily rely on records from small lakes or peat bogs (Gałka *et al.* 2013, 2014; Słowiński *et al.* 2015). In northeastern

Germany, a high sensitivity of lakes to Holocene hydrological changes has been suggested (e.g. Lampe *et al.* 2009; Kaiser *et al.* 2012, 2014b; Küster *et al.* 2014); however, well-dated lake level reconstructions remained restricted to certain times.

Lake Fürstenseer See, a typical groundwater fed lake with complex morphology has already showed pronounced changes in lake level and size during the instrumental period (Germer *et al.* 2011; Kaiser *et al.* 2014a), suggesting that its sediments may provide a highly sensitive record of past hydrological shifts. The first acoustic sub-bottom profile of northeastern German lakes was analysed together with a transect of five sediment cores across the deepest sub-basin to evaluate the potential of different sedimentological and biological proxies to reconstruct Holocene lake level changes in such a setting. The main aims of this study are to reconstruct i) full Holocene water level amplitudes, ii) the Holocene lake level history and iii) to discuss potential driving mechanisms that caused these past lake level changes.

Study area

Lake Fürstenseer See (53°19’ N, 13°12’ E, ~63 m a.s.l., Fig. 1) formed after the retreat of the Weichselian ice sheet in the direct forefront of the Pommerian ice margin ($W2_{max,}$ 18-20 ka, Lüthgens & Böse, 2011; Fig. 1A). This outwash plain area is characterized by sandy and gravelly glaciofluvial sediments (<1-2% silt and clay). Today, *Fagus* and *Pinus* forests dominate the 33.4 km^2 catchment (63%) followed by 18% agricultural land, lakes (13%), peatlands (5%), and settlements (1%, Kaiser *et al.* 2014a). Within the sub-maritime to sub-continental temperate humid climate zone, Neustrelitz (~5 km west of the lake) has a mean annual temperature (precipitation) of 8.0 °C (584 mm) with most precipitation occurring as summer rainfall (range: 428-814 mm, 1961-1990, Stüve 2010). Main wind directions are from the west (P. Stüve, pers. comm. 2014).

There are no surficial inflows to this naturally closed lake system suggesting groundwater as the primary source of recharge to the lake. This is supported by monthly lake level variability in the period AD1973-2010 co-varied with groundwater recharge (Stüve 2010; Fig. 1D). However, since at least the 17th century and until the establishment of the Müritz national park in AD199, the lake was artificially connected to adjacent lakes (Kaiser *et al.* 2014b). Kaiser *et al.* (2014b) reconstructed a 3 m higher lake stand during the late Medieval time, but full water level amplitudes are unknown for the Holocene.

In September 2011 (water level: 63.71 m a.s.l.), the ~240 ha large dimictic lake had a volume of 17.4 Mio. m³, a perimeter of 19.9 km, a maximum water depth (w.d.) of 25 m (z_{mean} = 6.9 m) and a catchment to lake ratio of 14:1. In the deepest, i.e., south-eastern sub-basin (Fig. 1B), discontinuous water quality observations revealed Secchi depths at 5.2±1.2 m w.d., pH of 8±0.6 and electrical conductivity of 262±23 mS cm^{-1}. The thermocline was established at ~10 m w.d. in summers (1973-2011, Ministerium für Landwirtschaft, Umwelt und Verbraucherschutz Mecklenburg-Vorpommern, pers. comm. 2012). Modern macrophyte surveys have revealed a high water plant diversity dominated by *Chara spec.* (Kirschey & Oldorff 2012).

Material and methods

Field methods

For estimations of present and past lake areas and volumes, bathymetric survey points from October 2002 (Ministerium für Landwirtschaft, Umwelt und Verbraucherschutz Mecklenburg Vorpommern, pers. comm. 2013) were interpolated using triangulation (SAGA-GIS 2.2) supported by near-shore elevations from an airborne laser scanning digital elevation model (1 m horizontal, 0.2 m vertical resolution; Amt für Geoinformation, Vermessungs- und Katasterwesen Mecklenburg-Vorpommern, pers. comm. 2013).

To detect the modern mud-sand boundary (i.e. modern sediment limit), an acoustic sub-bottom profile across the southeastern bay was performed in May 2014 using a SES-96 sediment echo sounder from Innomar Technologie GmbH (frequency 10 kHz, gain 14 dB, sound velocity 1462 m s^{-1}). In the same bay, five lake sediment cores along a 173 m long transect between 10 and 23 m w.d. were recovered in October 2008 (short core GFS10-8) and in September 2011 (four long cores GFS11; Fig. 1B, C, Tab. 1). An Uwitec piston corer recovered 2-m long drives from one borehole at each site, with an overlapping additional short core next to the long drives. All relative water depths mentioned below refer to an absolute level of 63.71 m a.s.l. (September 2011). Cores were split and macroscopically described after Schnurrenberger *et al.* (2003). Sediments from cores GFS11-AB, -C and -E were microscopically inspected using smear slides and thin sections (preparation after Brauer & Casanova 2001).

Geochemical analyses

The sediment composition of the long cores was continuously analysed with an ITRAX μ-X-ray fluorescence (μ-XRF) core scanner (Cr-HE-tube, 40 kV, 40 mA, 1 mm step size, 500 ms exposure time). Nine elements (Si, S, K, Ca, Ti, Fe, Rb, Sr, Zr) with less than 5% zero values were used further. Element intensities from XRF core scanning are influenced by down core changes of physical properties, the sample geometry, enhancement and absorption effects (Tjallingii *et al.* 2007). They are non-linear functions of geochemical concentrations, whereas log-ratios of element intensities are linearly related to log-ratios of concentrations (Weltje & Tjallingii 2008). Hence, element intensities were transformed with a centered log-ratio (clr) using the R-package "robCompositions" (Templ *et al.* 2011) to prevent mis-interpretations, which are further related to the compositional nature of geochemical data.

Composite core stratigraphies were derived from overlapping cores using macroscopically visible boundaries and the Ca record. To derive objective boundaries for the lithological units of the composite cores GFS11-C, -D and -E, a depth-constrained hierarchical cluster analysis (CONISS, Grimm 1987) was performed using the clr-transformed element data and the ratio of incoherent and coherent scattering (Cr_{inc}/Cr_{coh}, representing elements lighter than Al) in the R-package "rioja" (Juggings 2012).

Discrete samples were taken from blackish, greyish and whitish stratigraphic layers to measure total nitrogen (TN) and total carbon (TC) with an elemental analyser (EA3000-CHNS Eurovector). Samples with a volume of one cm^3 (n = 195; see Fig. S1 for sample positions) were freeze-dried, milled and prepared as 5 mg aliquots in Sn capsules. Additional 3 mg aliquots were decalcified in Ag capsules with 20% HCl and dried at 85 °C to measure total organic carbon (TOC). The analytical precision is 0.01%. C/N ratios and amount of carbonate (TIC) could then be calculated.

Biological analyses

Discrete samples from the main lithological units were analysed for macrofossil remains (n = 71) and subfossil *Cladocera* composition (n = 39, Fig. S1). Macrofossil samples were washed on 125 µm sieves and analysed according to Birks (2007). All macrofossil counts were standardised as numbers of fossils per 50 cm^3. Determination of macrofossil remains was based on the literature (Birks 2007; Tobolski 2000; Velichkevich & Zastawniak 2006, 2008) and a reference collection of the Institute of Geography, Polish Academy of Science.

Cladocera samples were heated, stirred 20 minutes in a 10% solution of KOH and washed through a 38-µm sieve following Frey (1986). Samples were then analysed using a transmitted-light microscope at magnifications of 100x to 400x. All counted *Cladocera* remains (headshields, shells, ephippia, postabdomens) and the most abundant body part of each species were considered to represent the number of individuals. A minimum of 200 individuals per sample were identified based on Szeroczyńska & Sarmaja-Korjonen (2007). Subfossil *Cladocera* data was evaluated using a detrended correspondence analysis (DCA) after Hill & Gauch (1980) in the R-package "vegan" (Oksanen *et al.* 2013). The first two axes had axis lengths >1.3 and were chosen to represent the dataset.

Dating and core correlation

AMS ^{14}C dating was performed on 22 samples of terrestrial plant remains at Poznań Radiocarbon Laboratory (Tab. 2). Calibration was done using Oxcal 4.2 (Bronk Ramsey 2008) and the IntCal13 dataset (Reimer *et al.* 2013). Age-depth modelling was performed for the longest continuous core GFS11-D. A Bayesian p-sequence depositional age-depth model was calculated with Oxcal 4.2 using a k-value of 1 cm and allowing sedimentation rate to vary within U(-1,1). This was a rather limited range of freedom, in contrast to the suggestion by Bronk Ramsey & Lee (2013) to use U(-2,2), which introduced unrealistic sedimentation rates (not shown). Sediment cores were correlated using sediment facies and multi-proxy analyses.

Results

Basin morphology and acoustic transect

The bathymetric survey revealed that the lake basin of Lake Fürstenseer See has a complex morphology (shoreline development = 5.5, Hutchinson 1957; Fig. 1A). The two main elongated basins follow the general paraglacial meltwater flow direction from northeast to southwest. These basins contain several sub-basins and are linked by a shallow area that falls dry in times of instrumentally recorded low lake stands. The steepest slopes occur along the eastern shores and southeastern sub-basin. In the southeastern sub-basin, which is the deepest (Fig. 1B, C), the maximum fetch along the prevailing wind direction is up to 1 km (Tab. 1) leading to strong turbulence in the upper water column. During the time period 1973 to 2013 (Fig. 1D) the lake level, area and volume changed interannually by ~2.1% (i.e., 1.3 m), 49% and 23%, respectively.

Sediment cores were located in the southeastern sub-basin, between 81 to 191 m from the shore (GFS11-AB to -E, respectively). Slopes were steepest at site GFS11-AB (22.5%) and shallowest at the deepest site (2.4% at GFS11-E, Tab. 1). Across the core transect, an acoustic sub-bottom profile indicated a clear mud-sand boundary at ~14 m w.d., i.e. the modern sediment limit (Fig. 3). At this depth, the first reflector at the sediment-water interface changed from very high backscatter representative of dense, sandy sediments upslope towards very low, almost transparent impedance contrasts of onlapping reflections that represent fine-grained, water-saturated mud from calm pelagic deposition below the storm wave base. Sediment focusing caused increasing sediment thickness below 14 m w.d. However, the deepest part of the sub-basin was masked by gas, as is typical in such environments (Gilbert 2003; i.e., diffuse reflections directly at the sediment-water interface and high scattering of the acoustic waves within the water column, Fig. 3).

Sediment facies, geochemical and biological results

Along the core transect, different lithological units (LU) were distinguished by visual descriptions after Schnurrenberger *et al.* (2003; all cores) and CONISS-clustering of the μ-XRF-data (mud-dominated cores GFS11-C, -D, -E; Tab. 2). LU were numbered from top to bottom for each core separately. Sand layers and distinct LU-boundaries indicated unconformities. Continuity of sedimentation increased with water depth (Figs 3, S1).

Three main sediment facies were distinguished across the transect: sands, blackish organic muds, and greyish carbonate muds. The 1.9 m long core GFS11-AB at 10 m w.d. primarily consisted of fine to coarse, partly bedded sands with organic detritus and some gravel, except for LU2. This 9 cm thick layer of fine-grained carbonate-rich mud contained 5-10 μm long calcite crystals accumulated with up to 30 μm diameter *Chara spec.* oogonia and thin amorphous organic matter layers (no sand, Fig. S2). At 15.2 m w.d., core GFS11-C contained 3.7 m of dominantly fine-grained, organic- and carbonate-rich muds intercalated with several sand layers in a carbonate mud matrix at 210 to 280 cm sediment depth (LU5). Two fining-upwards sand layers at 165 to 178 cm sediment depth (LU4) may have resulted from erosional events upslope. At 19.8 m w.d., the 8.4 m long core GFS11-D was completely composed of alternating blackish organic mud and greyish calcareous mud sequences and did not contain any sand layers. The 6.3 m long core GFS11-E at 23 m w.d. had similar, but slightly thicker, lithological units to core GFS11-D. In contrast to cores GFS11-AB and -C, cores GFS11-D and -E did not reach the lacustrine sediment base.

Organic muds consisted of decomposed organic matter (sometimes identifiable as aquatic plant remains), varying contents of diatoms, chrysophytes, secondary pyrite and few dispersed sand-sized mineral grains (Fig. S2). Carbonate muds differed from organic muds by the additional appearance of micritic calcite in cores GFS11-C to -E. Finer crystals than the littoral calcites of GFS11-AB-LU2 (Fig. S2) suggest dominantly pelagic carbonate precipitation. Dry densities were 0.24±0.2 and 0.81±0.4 g cm^{-3} for mud and sand facies, respectively.

TOC contents were generally very high (15±9%, n = 195) and C/N was between 10 and 20, but close to 10 in all mud units independent of carbonate content (Fig. S1, Tab. S1), revealing the dominance of authigenic organic matter production (Meyers & Ishiwatari 1993) with terrestrial matter input contributing primarily to sand layers.

The main lithological variation between blackish organic and greyish carbonate muds can be illustrated by the continuous Ca record as a proxy for authigenic carbonate preservation (i.e. high Ca in carbonate muds, Fig. S1; no detrital calcite found in microscopic inspection, Fig. S2). Clr-transformed Ca was positively correlated with TIC (r = 0.76). Ca was negatively correlated with most other elements (e.g. Ti, K, Fe; not shown).

Macrofossil analyses revealed a strong presence of *Chara spec.* oogonia, but other water and wetland plant remains occurred only marginally in few samples. Subfossil *Cladocera* analyses indicated relatively similar species compositions through time, dominated by the open-water turbidity-tolerating species *Bosmina longirostris* (Bjerring *et al.* 2009), the eutrophy-tolerant species *Chydorus sphaericus* (Vijverberg & Boersma 1997) and varying amounts of a few further pelagic and littoral species (Fig. S1, Tab. S2).

Core correlation

Stratigraphic correlation of sediment facies and LU between the cores was supported by sedimentological and subfossil *Cladocera* analyses (Fig. 3). The topmost units LU1 to 3 in cores GFS11-C to -E and LU1/2 in core GFS10-8 had similar facies and subfossil *Cladocera* compositions (Fig. S1). However, the thicknesses of LU1 to 3 increased with water depth due to sediment focusing, allowing the subdivision of LU2 in the two deepest cores (Fig. 3). Layer distortions in LU2b in cores GFS11-D and -E could be related to slumping, possibly explaining the erosional upper boundary and low thickness of LU2 in GFS11-~LUs 4 and 5 showed high-frequency facies shifts in cores GFS11-C to -E (and weakly developed in core GFS11-AB). These shifts were either between carbonate muds and sandy layers (GFS11-C) or between carbonate and organic muds (GFS11-D, -E, Figs 3, S1). A single organic mud layer in core GFS11-C-LU4 and a massive 10 cm thick organic mud in GFS11-D-LU4a correlate with the 9 cm thick carbonate mud of LU2 in shallow core GFS11-AB (Fig. 3). A distinction between lower and upper deposits (with the boundary between LU3 and LU4) was supported by subfossil *Cladocera* assemblages (the first DCA axis site scores showed a trend related to sediment depth; Fig. S1B).

Chronology

The initiation of lake formation was dated to ~13300 cal. a BP using remains of *Arctostaphylos uva-ursi* and *Juniperus communis* found at 365 cm sediment depth at the lacustrine sediment base of core GFS11-~The Allerød age was confirmed by two further dates above the same layer in core GFS11-AB (Poz-61454, Poz-61452, LU5, ~12 800 cal. a BP, Tab. 2). Hence, Lake Fürstenseer See originates from melting dead-ice blocks as suggested by macrofossil remains and C/N-ratios (Słowiński 2010), similar to other lakes in the central European lowlands (Kaiser *et al.* 2012).

The 9 cm thick littoral carbonate muds of GFS11-AB-LU2 had minimum and maximum ages of 4730 and 5300 cal. a BP, respectively (Tab. 3). The short core GFS10-8 records the last 500 years, whereas the long cores GFS11-D and -E contained sediments older than 8945±290 and 4307±212 cal. a BP, respectively (2σ-error, lowermost samples, lake sediment base not reached, Tab. 3). Ages in all cores were in stratigraphic order, except for two too young samples in cores GFS11-D and -E suggesting contamination during core recovery (Poz-55889, Poz-61462, not considered further; Tab. 3, Fig. 4).

As cores GFS11-AB and -C contained several sand layers and distinct boundaries indicating hiatuses and cores GFS10-8 and GFS11-E were rather short, a detailed age-depth-model was only calculated for core GFS11-D (Fig. 4). Major sediment facies shifts had ages of 6350±130 cal. a BP (base of LU4), 3750±100 cal. a BP (base of LU3), 800±70 cal. a BP (base of LU2) and 220±80 cal. a BP (base of LU1; ages and errors rounded). Sedimentation rates in core GFS11-D were highest in the upper part of the core, especially in LU2 (~2.3 mm a^{-1}, ~220-800 cal. a BP) and lowest in LU5 (~0.4 mm a^{-1}, ~6400-9000 cal. a BP; Fig. 4), due to soft mud compaction. If the LU5 sedimentation rate is considered to be constant throughout LU5, then the basal age of LU5 can be linearly extrapolated, indicating a further facies shift at ~9670±290 cal. a BP (base of LU5, Fig. 4).

The age-depth model was confirmed by dates from other cores and correlative LUs (Figs 3A, 4, Tab. 2). For example, the base of LU3 in GFS11-E was also dated to ~3800 cal. a BP in accordance with an age of ~3200 cal. a BP in the lower part of GFS11-C-LU3. Tab. 3 summarizes the facies descriptions of the LUs and their estimated time of deposition based on discrete dating, the detailed age model of GFS11-D and core correlation.

Interpretation and discussions

Proxy interpretation in terms of lake level

Absolute lake level reconstructions from changes in sediment limit. The determination of the modern sediment limit at 14 m w.d. from the acoustic reflections (Fig. 2) was supported by the composition of the topmost facies above and below (sandy LU1 of GFS11-AB *versus* muddy LU1-3 of cores GFS11-C, -D, -E). The presence of fine-grained carbonate muds in 10 m w.d. (GFS11-AB-LU2) indicates that even on slopes as steep as 22% fine sediment was preserved, which is much steeper than previously suggested (Håkanson & Jansson 1983; Dearing 1997). The mud deposition at ~5000 cal. a BP, required the sediment limit and corresponding storm wave base and lake level to rise by at least 4 m. This would be higher than the 3 m Medieval lake high-stand recorded in onshore palaeosol sediment sequences (Kaiser *et al.* 2014b). Lake volume and area would have increased by at least 390 and 310% (Fig. 5). These are minimum estimates due to missing bathymetries of adjacent lakes, the unknown amount of deposited lake sediments after this high-stand and undeterminable changes in surface topography due to human impact. Furthermore, DEM boundaries were exceeded.

In contrast, the deposition of sand layers in 15.2 m w.d. (core GFS11-C) represents littoral erosion in times of a lowered sediment limit (Digerfeldt 1986; Dearing 1997; Pribyl & Shuman 2014). The sands of GFS11-C-LU4/5 were deposited with a fine-grained carbonate matrix between ~3750 and 9670 cal. a BP. Their absolute position at 210 to 280 cm sediment depth suggests that the sediment limit lowered to ~17.3 to 18 m w.d., but remained above 19.8 m w.d. (i.e., no sand layers in core GFS11-D). Hence, a maximum lake level and sediment limit decline of 4 m can be assumed. The lake would have reduced to 53 and 43% of its modern area and volume (minimum estimates based on modern bathymetry) isolating the sub-basins (Fig. 5).

Relative lake level reconstruction from facies variations represented by Ca-record. At ~5000 cal. a BP, when carbonate muds were deposited in 10 m w.d., blackish organic muds accumulated at the deeper positions (tentatively in core GFS11-C-LU4 with poor age control; better resolved in core GFS11-D-LU4a; Fig. 3). This opposing pattern of carbonate preservation was probably related to carbonate dissolution in profundal regions due to stratification (Anderson *et al.* 2005), with the thermocline depth being located between the core sites AB and C.

High organic matter deposition and low carbonate preservation at the deep sites are associated with high lake levels for two additional reasons. First, an up to 4 m higher lake would have inundated a large area of primarily flat vegetated terrain (Fig. 5). A high amount of biomass would have degraded in the littoral zone and been winnowed towards deeper areas. Carbonate that still precipitated in the littoral and/or pelagic zone would have been more easily dissolved within a larger water volume and/or by the pH-reduction due to the extra organic acids in the water column (Håkanson & Jansson 1983; Dean 1999). Second, the supersaturation of Ca needed for carbonate precipitation would have required a much larger amount of Ca in a larger, more slowly warming water column than in a small, rapidly warming lake (Kelts & Hsü 1975; Dean 1999). For these reasons, low carbonate and high organic matter contents have generally been associated with high lake levels and cool, humid climate, e.g. in North American lakes (Anderson *et al.* 2005; Pompeani *et al.* 2012) or in Argentina (Haberzettl *et al.* 2005; Ohlendorf *et al.* 2013). This seems to be also a likely interpretation for groundwater-fed lakes of the central European lowlands.

Along the same line, a shallow lake favors higher rates of carbonate accumulation. Supersaturation and thresholds for carbonate precipitation is more easily reached in a rapidly warming shallow water column with increased authigenic productivity (Kelts & Hsü 1975; Dean 1999). Hence, high Ca contents that occur in LU4/5 and LU2 of the deeper cores GFS11-C, -D and -E point to lower lake levels.

Constraints for Ca-interpretation as palaeohydrological proxy. Some sensitivity constraints should be considered when interpreting the Ca record as a lake level proxy in this oligotrophic lake. First, the preservation of the littoral carbonate muds in GFS11-AB that formed under a 4 m higher lake level at ~5000 cal. a BP could indicate the maximum extent of the lake (Fig. 5).

Alternatively, it could have been just a very high lake stand that could have occurred several times to a similar extent, but one whose sediments survived intense reworking at this steep slope to preserve the record. From the sedimentary evidence alone, it cannot be judged whether other similarly low Ca-phases represent similarly high lake stands.

Second, the black homogenous organic muds of the topmost sediments (LU1) below the modern sediment limit indicate that i) the deposition of organic muds already occurs at lake stands similar to those of today and ii) that the instrumentally-observed lake level range of 1.3 m cannot be traced in the modern lake sediments.

Third, after a long period of strong fluctuations in the early and mid-Holocene down to 4 m below modern lake level, Ca remained low for ~3000 years after ~3800 cal. a BP (LU3). This suggested a generally higher lake level. However, stronger fluctuations, as reported from nearby lakes (Lorenz 2007), were no longer recorded. One reason might be that thresholds for Ca-preservation were not reached, either because the lake was too high or lake level fluctuations in this time were too low to leave an imprint in the sediments. Another reason could be a reduced Ca availability in the groundwater. Bicarbonate primarily originates from local glacial deposits and was ubiquitous after the melt down of the Pleistocene ice sheets, but was reduced in the later stages of the Holocene (as evidenced by low concentrations in modern groundwater, H. Wilke, pers. comm. 2015).

Fourth, a rise in trophic conditions after ~800 cal. a BP, especially during Medieval times, was evidenced by the relative increase in eutrophy-tolerant subfossil *Cladocera* species in calcareous LU2 of the GFS11-C to -E cores (e.g. *Chydorus sphaericus;* Vijverberg & Boersma 1997) that grouped along the first DCA axis (Fig. S1B). Higher trophy would increase algae growth and related carbonate precipitation (Kelts & Hsü 1975; Dean 1999). Establishment of human settlements and forest pasture (*sensu* Segerström & Emanuelsson 2002) in the catchment might have increased nutrient input to the lake causing a renewed higher sensitivity of the carbonate system to lake level changes. Also, stronger lake level fluctuations, which allowed the Ca-preservation-threshold to be crossed, could have been provoked by humans manipulating the local hydrological system, e.g. by drainage measures (Kaiser *et al.* 2014b). In contrast, a lack of a trophy-related forcing of carbonate precipitation before 3800 cal. a BP was evidenced by subfossil *Cladocera* species preferring rather oligotrophic conditions (e.g. *Rynhotalona falcata,* Bjerring *et al.* 2009) in calcareous LU4/5 of cores GFS11-C, -D and -E and GFS11-AB-LU2.

In summary, the Ca record reflects "relative" lake level changes with a high sensitivity of the carbonate system to fluctuating lake levels during the early to mid-Holocene (high-frequency shifts in LU4/5) and during the period of major human impact (LU2). The proxy was less sensitive during deposition of the late Holocene organic muds (LU3, LU1).

Additional classical palaeohydrological proxies related to the Digerfeldt approach were of limited use at this "non-classical" site (e.g. loss-on-ignition residuals as proxy for minerogenic matter was biased by high amounts of diatoms in the sediments). Common water depth sensitive macrofossils and subfossil *Cladocera* (Hannon & Gaillard 1997; Korhola *et al.* 2000) were lacking in the sediments, because of the depth and steepness of the core transect throughout the Holocene and/or the high degree of organic matter degradation. Although the second DCA axis of subfossil *Cladocera* showed a relation to modern core position (i.e., distance to shore; Fig. S1B; highest site scores in the carbonate muds of GFS11-AB-LU2, not shown), it rather represents a habitat association (e.g. *Alonella exigua* strongly connected with macrophytes versus *Leydigia acathocercoides* and *Leydigia leydigi* living on the sediment; Bjerring *et al.* 2009).

Temporal evolution of the lake and regional comparison

With the mechanisms and constraints in mind, the temporal evolution of Lake Fürstenseer See lake levels can, on the one hand, be directly affected by climate, with cooler and/or wetter periods leading to a higher water availability and more effective groundwater recharge because of lower evapotranspiration and vice versa (e.g. Anderson *et al.* 2005). On the other hand, regional vegetation composition as determinable from pollen analyses strongly influences groundwater recharge and, hence, lake levels. Deciduous tree stands (e.g. *Fagus* and *Quercus*) allow a comparably high groundwater recharge due to high stemflows and throughfall (esp. in winter), whereas conifers (e.g. *Pinus*) have a rather negative effect on the local water budget due to high evapotranspiration throughout the year (e.g. Nathkin *et al.* 2012; Guswa & Spence 2012).

The Holocene evolution of Lake Fürstenseer See is compared to European-wide palaeohydrological records of different quality and temporal resolution. For comparability, and if not provided in the original publications, ^{14}C ages of all records were calibrated with IntCal13 (Reimer *et al.* 2013) to cal. a BP.

>9700 cal. a BP. The deposition of massive organic muds in LU6 (Fig. 4) suggests a rather high lake level during the Early Holocene. During a warming climate, organic matter was distributed and deposited in a large lake by an active lake circulation driven by long fetches and carbonate preservation was inhibited. Forests were still sparse and light with dominance of *Pinus, Betula* and *Corylus* (Jahns 2007; Lampe *et al.* 2009; Feurdean *et al.* 2014). Similar lake high-stands were observed in Poland at 11 300-10 500 and 9900-9500 cal. a BP (Pazdur *et al.* 1994; Michczynska *et al.* 2013). High lake levels occurred also in Switzerland and eastern France (11250-11050, 10300-10 000 cal. a BP, Magny 2004; 10 400-9000 cal. a BP, Magny *et al.* 2011), whereas in southern Sweden lake levels reached a minimum at ~10 500 cal. a BP (Digerfeldt 1988; Fig. 6). Pollen-based climate reconstructions suggested rather dry conditions throughout Europe 12 000 to 9000 years ago (Mauri *et al.* 2015), and some German lakes showed lake regressions during this time (Kleinmann *et al.* 2000). However, at Lake Fürstenseer See, rather than humidity or evapotranspiration as drivers, local melting of Younger Dryas permafrost remains and dead-ice could have contributed to groundwater recharge and high lake levels (Pazdur *et al.* 1994; Lorenz 2007; Błaszkiewicz *et al.* 2015).

9700 - 6400 cal. a BP. Between ~9700 and ~6400 cal. a BP the lake was generally low and fluctuating with a few episodes of higher lake stands occurring at ~9600-9500, ~7800 and ~6800 cal. a BP as recorded by finely laminated, carbonate-rich GFS11-D-LU5 sediments intercalated with some organic muds (Figs 3, 6). During this period, sand layers intercalated with carbonate muds at core site GFS11-C, indicating that the sediment limit was up to 4 m lower than today (Fig. 5). Calcium from glacial deposits was widely available in the groundwater, sub-basins were more isolated, and fetches were reduced. This allowed thresholds for carbonate preservation to be easily crossed, reflecting a high sensitivity of the lake system even to minor lake level fluctuations. Although the lake was lower and smaller in extent, it still was at least 3 m deeper than today because the overlying 6 m sediments at site GFS11-D were not accumulated yet. This allowed intermittent anoxic conditions to preserve the fine laminations, which were also found at nearby Lake Drewitzer See. There, similar short term high level episodes during a generally low stand were reflected by few minerogenic peaks during the same period (Lorenz 2007). High-frequency lake level variations were also observed at Lake Hämelsee, but age control needs to be improved to detect comparable high-frequency changes at other German lakes (Kleinmann *et al.* 2000). Northern Central European lakes reached their Holocene water level minima (at some sites 5-7 m below the modern level) especially during the early stages of this period. Afterwards, continuously rising lake levels were proposed (Fig. 6; Pazdur *et al.* 1994; Lampe *et al.* 2009; Kaiser *et al.* 2012). In the Alps, a minimum lake level was observed at 9000-8500 cal. a BP and a series of low levels thereafter (Magny *et al.* 2011). Higher lake levels occurred 9550-9150, 8300-8050 and 7550-7250 cal. a BP (Magny 2004; Fig. 6).

In contrast to low lake levels in north-eastern Germany, southern Swedish and Polish palaeohydrological reconstructions revealed spatio-temporally heterogeneous, but rather high lake levels peaking ~8000 years ago (Digerfeldt 1988; Pazdur *et al.* 1996; Edvardsson *et al.* 2012; Fig. 6). Also pollen-based climate reconstructions suggested that Central Europe experienced its wettest period during this time period (Mauri *et al.* 2015). Hence, again a regional driver for generally low lake levels in northeastern Germany should be considered: regional pollen records indicated that forests dominated by *Pinus*, with some *Betula, Corylus* and *Quercus* were fully established on sandy sites (Jahns, 2007; Lampe *et al.* 2009). *Pinus,* in particular, would consume most of the available water throughout the year and affect groundwater recharge negatively (Kaiser *et al.* 2012; Guswa & Spence 2012).

6400 - 3800 cal. a BP. Between ~6400 and ~3800 cal. a BP, several shifts from carbonate to organic muds suggest intense water level fluctuations with four periods of high stands at ~6400-6100, ~5800-5600, ~5100-4900 and ~4600-4500 cal. a BP (Fig. 6). The preservation of the littoral carbonate muds in GFS11-AB that formed under a 4-m-higher lake level at ~5000 cal. a BP indicates the likely maximum extent of the lake in a short time period (Fig. 5). However, there is no local onshore evidence for such high levels during the mid-Holocene so far (Kaiser *et al.* 2014b), perhaps because only “catastrophic” high-stands that move enough sediment to create, e.g. beach ridges, can be identified in the area. Regional lakes reached their maximum extents or similar lake levels to today (Kaiser *et al.* 2012). In Poland, high lake levels were reconstructed for the periods 6500-5600, 4900-4600, and 4400-3500 cal. a BP (Gałka & Apolinarska 2014; Gałka *et al.* 2013, Fig. 6), when European-wide flooding was also observed (Macklin *et al.* 2006), within the age-uncertainties supported by high-stands of the GFS11-D-record (Fig. 6). Between ~4400 and 3800 cal. a BP, carbonate muds at site GFS11-D and slump events recorded by the fining-upwards sand layers of core GFS11-C indicate lake levels ~3 m lower than today (Fig. 6), in agreement with dry bogs in northwestern Germany

(Eckstein *et al.* 2009). Although western Alpine lakes had been rather low since ~6000 cal. a BP (Magny 2004), the transition to cooler and wetter conditions probably caused higher lake levels in west-central Europe (6350-5900, 6500-5200, 4850-4800, and 4150-3800 cal. a BP, Magny 2004, Fig. 6).

A reason for the strong lake level fluctuations visible in GFS11-D-LU4 and interpreted, for example, from southern Swedish lakes during this time (Digerfeldt 1988), might be the transition from the mid-Holocene thermal maximum towards the onset of neoglaciation. The initiation of this transition was recognized as period of rapid climate change (Mayewski *et al.* 2004). However, the tendency towards generally higher lake levels might also reflect transitional change in forest composition from *Pinus* to dominance of *Quercus*, *Ulmus* and *Fagus* (Jahns 2007; Lampe *et al.* 2009), which root water directly to the ground, improving groundwater recharge (Guswa & Spence 2012).

3800 - 800 cal. a BP. The most prominent sediment facies shift at ~3800 cal. a BP suggests an abrupt shift in lake system behavior. Low Ca values indicate that the lake level was high, at least similar to today. The organic muds were well mixed suggesting an intense lake-internal circulation across the sub-basins. No high-frequency lake level fluctuations were recorded for almost 3000 years, because the thresholds for carbonate preservation were not exceeded.

Within the age uncertainties, this prominent facies and lake system shift might represent a response to the global climate shift at the onset of neoglaciation (~4200 cal. a BP; Mayewski *et al.* 2004; Wanner *et al.* 2008). An associated increase in wetness since the mid-Holocene was related to a gradual decrease in northern hemisphere summer insolation and was reflected, e.g. in increased fluvial activity in Germany and Europe (Macklin *et al.* 2006; Hoffmann *et al.* 2008), and rising, but strongly fluctuating water levels in Sweden (Hammarlund *et al.* 2003; Edvardsson *et al.* 2012), in the Alps (Magny 2004), Ireland (Swindles *et al.* 2010) and Great Britain (Charman *et al.* 2006). In Poland, lake levels were high during the periods 3000-950 cal. a BP (Gałka & Apolinarska 2014), 3400-300 cal. a BP (Gałka *et al.* 2014) and 2550-700 cal. a BP (Gałka *et al.* 2013; Fig. 6).

A prominent shift in forest composition in northern central Europe (i.e. *Pinus*-*Quercus* forests substituted by *Carpinus* and *Fagus*) during the Early Bronze Age was probably associated with the initiation of wood extraction and forest grazing since 3800 or 3500 cal. a BP (Jahns, 2007; Ralska *et al.* 2003; Küster *et al.* 2014; Fig. 6). Hence, higher groundwater recharge from increased landscape openness, stemflow and throughfall (Guswa & Spence 2012) could have sustained high lake levels under a generally cooler and wetter climate.

800 cal. a BP to present. The lake system showed more fluctuations in carbonate deposition after ~800 cal. a BP, a time when humans started to affect the landscape more severely in multiple ways, such as changes in the drainage systems and/or the land cover, leading to locally diverse changes in the hydrological system (Jahns 2007; Kaiser *et al.* 2014b). Carbonate was accumulated, suggesting that lake levels lowered and that the lake system was again sensitive to lake level changes. This sensitivity shift may have been caused by a higher lake trophy, as subfossil *Cladocera* assemblages suggest (see above, Fig. 3E). The facies shift was accompanied by the highest sedimentation rates of the record, between ~800 and ~630 cal. a BP (LU2b-distortions, Fig. 3, 4; i.e. late Slavic and early/mid-Medieval time). The most intense human alteration of forests and drainage systems was reconstructed for the Medieval time period (~AD1290) and to a lesser extent for the 16th and 19th century, when water mills and glasswork industries were established (Küster *et al.* 2014; Fig. 6). Lake trophy at this time was rising, allowing the thresholds for carbonate preservation to be crossed again.

Hence, carbonate muds representing lower lake levels were preserved from the periods ~800-700 and ~500-200 cal. a BP, whereas organic muds were accumulated between ~700-500 cal. a BP. The latter is in agreement with onshore evidences that indicated a lake level rise of up to 3 m compared to today for the period AD1250-1400 (Kaiser *et al.* 2014b; Fig. 6), during the onset of the cold and wet interval of the Little Ice Age (Büntgen *et al.* 2011; Wanner *et al.* 2011). Similar water level fluctuations occurred at many lakes in the region (Kaiser *et al.* 2012). In northern Polish peatlands, the period between ~650/750 and 50 cal. a BP was characterized by hydrological instability (Lamentowicz *et al.* 2009) and even drought (Marcisz *et al.* 2015). In the western Alps, in contrast, a period of high lake level was observed in 750-650 and after ~450 cal. a BP (Magny 2004), maybe related to a different circulation regime.

In the 18th century, the lake shifted again to dominantly organic mud deposition (LU1, Fig. 6) and the modern lake system was established. The shift coincides with a period when forest management strategies changed to conservation rather than exploitation (Messner 2009), which reduced soil erosion and lake productivity and, therefore, the lake's sensitivity to record lake level changes. Vegetation changed from a rather heterogeneous structure caused by the multiple ways humans used forests on a poor soil (Segerström & Emanuelsson 2002) to a dominantly deciduous forest used for hunting (Messner, 2009). This may have increased groundwater recharge again and established the current level of Lake Fürstenseer See. The recent fluctuations of 1.3 m, however, did not leave an imprint in the sediments.

Multiple drivers of Holocene lake level fluctuations

Commonly, records of lake level fluctuations are interpreted as evidence for climatic changes. However, additional local and regional factors can superimpose climatic effects and complicate interpretation of such records. At Lake Fürstenseer See, a major driver for the early Holocene lake high stand until ~9700 cal. a BP was the final melting of local remains of dead-ice and permafrost, which formed during the Younger Dryas (Pazdur *et al.* 1994; Błaszkiewicz, 2002; Kaiser 2004; Błaszkiewicz *et al.* 2015). The locally high water availability, as evidenced by massive organic muds, might have delayed the subsequent early- to mid-Holocene low levels (3 to 4 m below modern, ~9700-6400 cal. a BP; evidenced by sand and carbonate deposition). Early Holocene lake low stands were also observed regionally (Kaiser *et al.* 2012) suggesting a response to the establishment of *Pinus* forests on sandy substrates starting at ~11 200 cal. a BP (Theuerkauf *et al.* 2014). *Pinus* contribute negatively to groundwater recharge due to their high evapotranspiration (e.g. Guswa & Spence 2012).

During the course of the Holocene, a general trend of rising lake levels was observed at Lake Fürstenseer See and other northeastern German lakes (Kaiser *et al.* 2012). This reflects the water level response to both: i) insolation-driven changes to a cooler and wetter mid to late Holocene climate (Magny 2004; Wanner *et al.* 2008) and ii) a gradual transition of vegetation towards the dominance of tree species (e.g. *Fagus*, *Quercus*, Ralska *et al.* 2003; Jahns 2007) that favoured higher groundwater recharge (Guswa & Spence 2012).

Until ~3800 cal. a BP, the GFS11 records showed additional multi-decadal to centennial-scale sediment facies changes representing water level fluctuations within the range of ~8 m. These might reflect superimposed short-term changes of moisture availability (Digerfeldt, 1988; Pazdur *et al.* 1994), which was suggested as main driver for high-frequency western European palaeohydrological changes in the few studies that compiled several regional water level records (Magny 2004; Charman *et al.* 2006; Swindles *et al.* 2010). These studies linked the position and strength of the moisture-providing westerly storm track to solar variation and North Atlantic atmospheric circulation.

After a ~3000 year long high-stand period of low sediment sensitivity to further lake level fluctuations, strong human impact that altered the local drainage system and land cover interacted with climate and vegetation effects in driving lake level fluctuations (Kaiser *et al.* 2012; Michczynska *et al.* 2013; Kaiser *et al.* 2014b).

Hence, Holocene lake level changes were driven by several causes acting at different time-scales. Local causes influencing groundwater recharge such as melting of permafrost remains and vegetation composition interacted with long and short term climatic changes. In addition to catchment-specific processes, regional climatic and eco-hydrological feedbacks, multifaceted human impact also played an important role in driving local to regional water budgets, particularly since Medieval times. To better disentangle the involved drivers at various temporal scales, much more highly-resolved palaeohydrological archives and proxies need to be investigated and compiled following Magny (2004) and Charman *et al.* (2006).

Conclusions

At a northeastern German lake with complex basin morphology, the combined approach of determining modern and past sediment limits from acoustic sub-bottom profiling and sediment facies along a transect of sediment cores allowed an effective estimation of the absolute lake level amplitude and relative fluctuations during the Holocene, even in a steep and deep sub-basin. The variability of carbonate preservation in the sediments (represented by the Ca record) is interpreted as a proxy for relative lake level changes with high Ca contents representing smaller than modern lake extents. However, this proxy relation was not stable through time. Sensitivity for recording lake level changes in the sediments seemed to depend on either the intensity of lake level fluctuations, the mean lake level upon which high-frequency fluctuations occur and/or the abundance of Ca in the groundwater system. Hence, the fluctuations during the instrumental period were not recorded within the lake sediments.

The lake was strongly fluctuating at a level 3 to 4 m lower than today during the early and mid-Holocene. Shifts of up to 4 m higher lake stands occurred during the late mid-Holocene. These extremes represent a total natural water level amplitude of 8 m. This clearly exceeded the modern range of 1.3 m lake level fluctuations during the last 40 years and occurred during a time when human impact on the lake and catchment hydrology was negligible. These strong fluctuations might be the result of differential groundwater recharge from feedbacks between vegetation composition and variable climatic background conditions. Further intense fluctuations with up to 3 m higher lake levels happened during the late Holocene, when humans diversely affected the drainage system, forest structure and lake trophy. If current forest and water management intends to reach quasi-natural conditions within the Müritz national park, the long Holocene record suggests that i) future lake water level changes could exceed the so-far-observed amplitudes and variability and ii) the hydrological response to climate change strongly depends on forest and vegetation composition.

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Figures

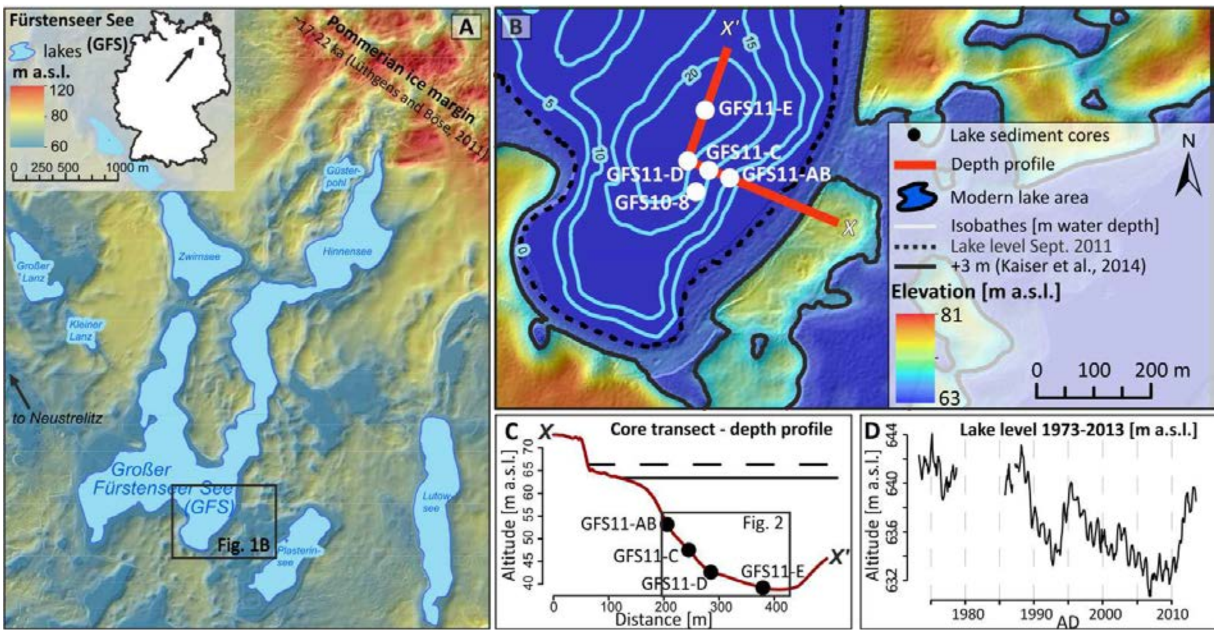

Fig. 1: A. Map of Lake Fürstenseer See within the Müritz national park, Germany. B. Location of the sediment core transect in the southeastern bay. C. Bathymetric profile across the core transect. D. Monthly lake level data from reading of a graduated rod for the period 1973-2013 (Stüve 2010; P. Stüve, pers. comm. 2014).

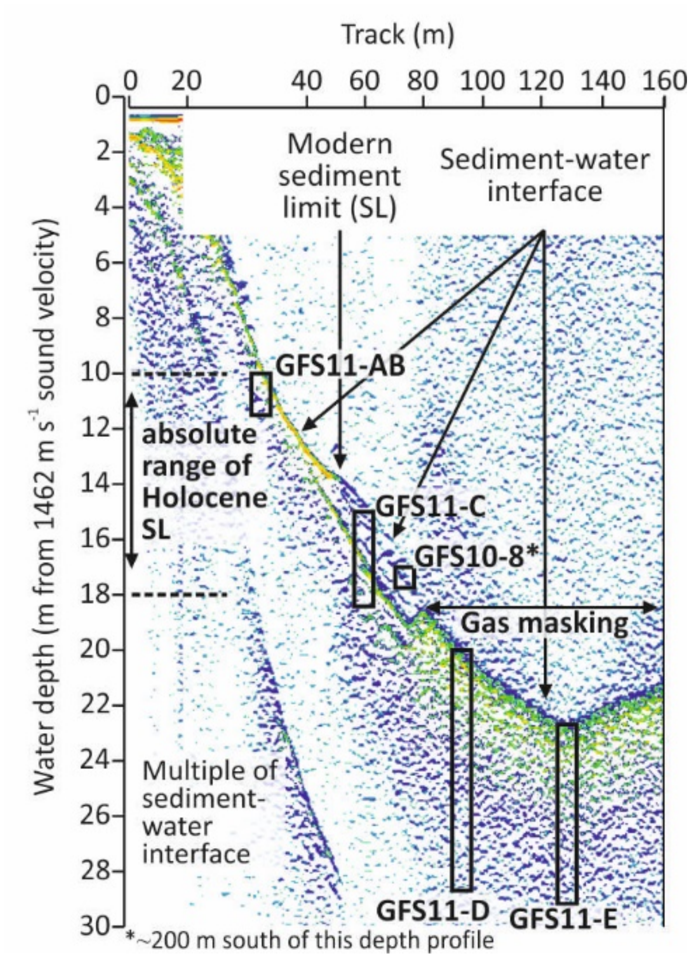

Fig. 2: Acoustic sub-bottom depth profile along the core transect in the southeastern bay of Lake Fürstenseer See. Blue to green colours indicate high transparency and low impedance contrasts as typical for water-saturated pelagic sediment, yellow to red colours indicate high backscatter from dense, sandy sediments.

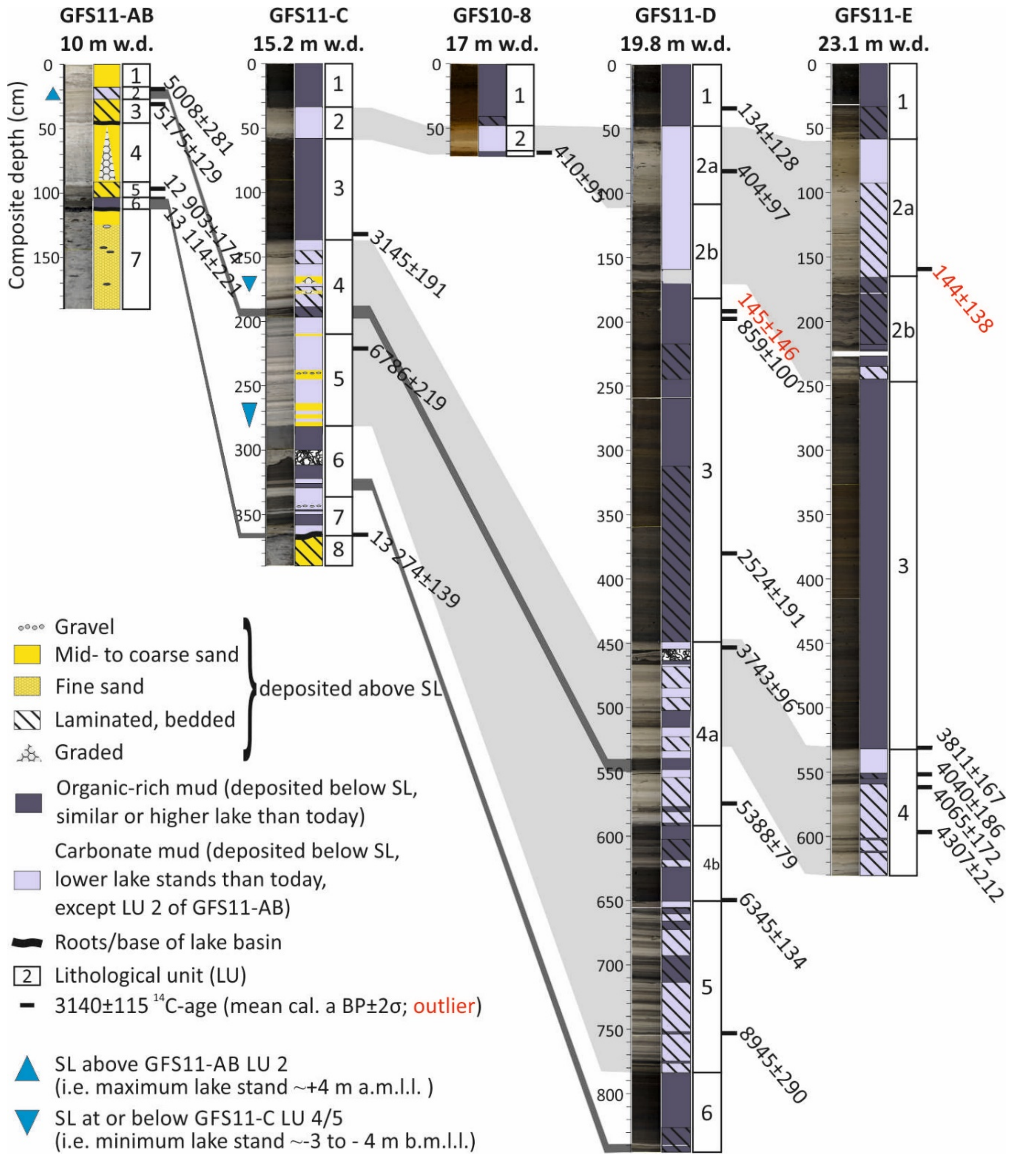

Fig. 3: Correlated stratigraphies of Lake Fürstenseer See composite cores including mean dates with their 2σ-ranges (see Tab. 2). Grey bars show correlative lithological units. SL and blue triangles indicate the assumed sediment limit position.

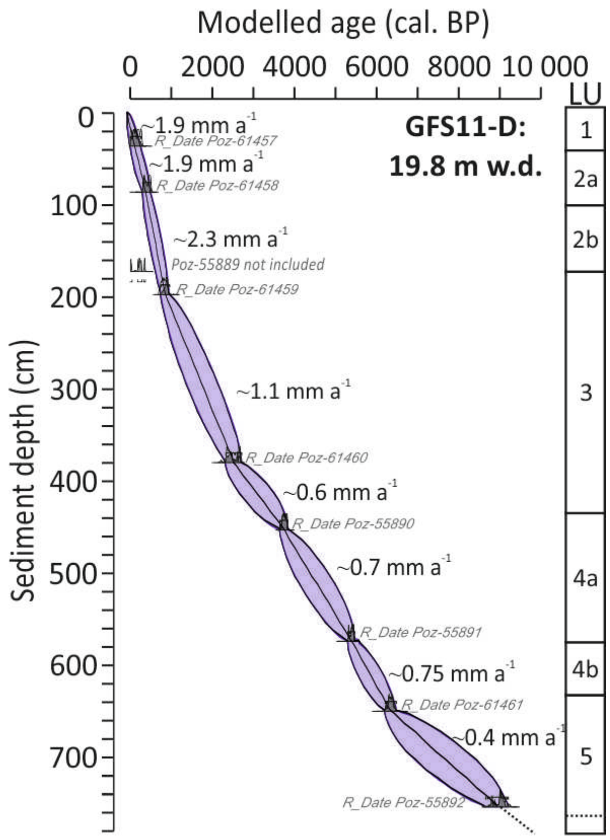

Fig. 4: Age-depth-model of core GFS11-D including sedimentation rates and lithological units (LU) for comparison. Dotted line marks linear age-depth-extrapolation to base of LU5.

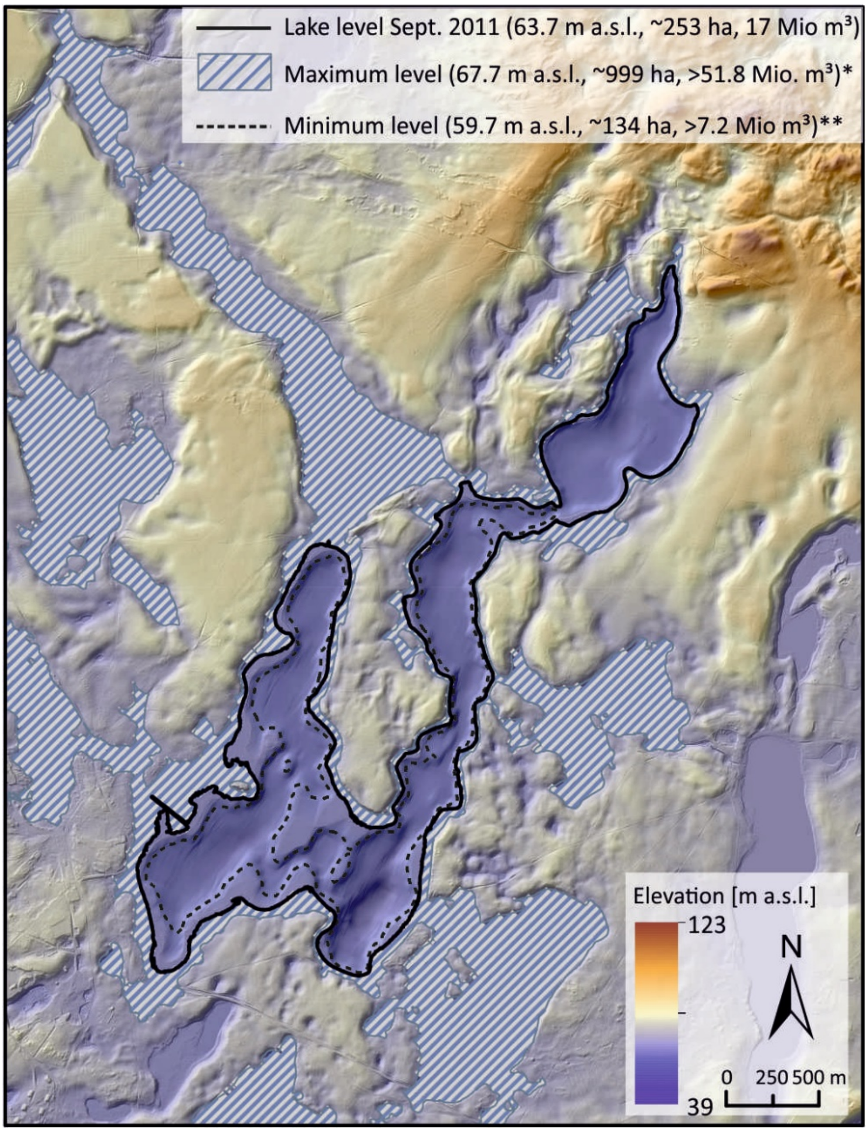

Fig. 5: Modern and reconstructed past extensions of Lake Fürstenseer See using a bathymetric survey (Ministerium für Landwirtschaft, Umwelt und Verbraucherschutz Mecklenburg-Vorpommern, 2013) and an airborne laser scanning digital elevation model (Amt für Geoinformation, Vermessungs- und Katasterwesen Mecklenburg-Vorpommern, 2013). *minimum estimate because no bathymetric data from adjacent lakes available, DEM boundaries exceeded and potential biases from human-induced morphological changes. **minimum estimate due to use of modern bathymetry.

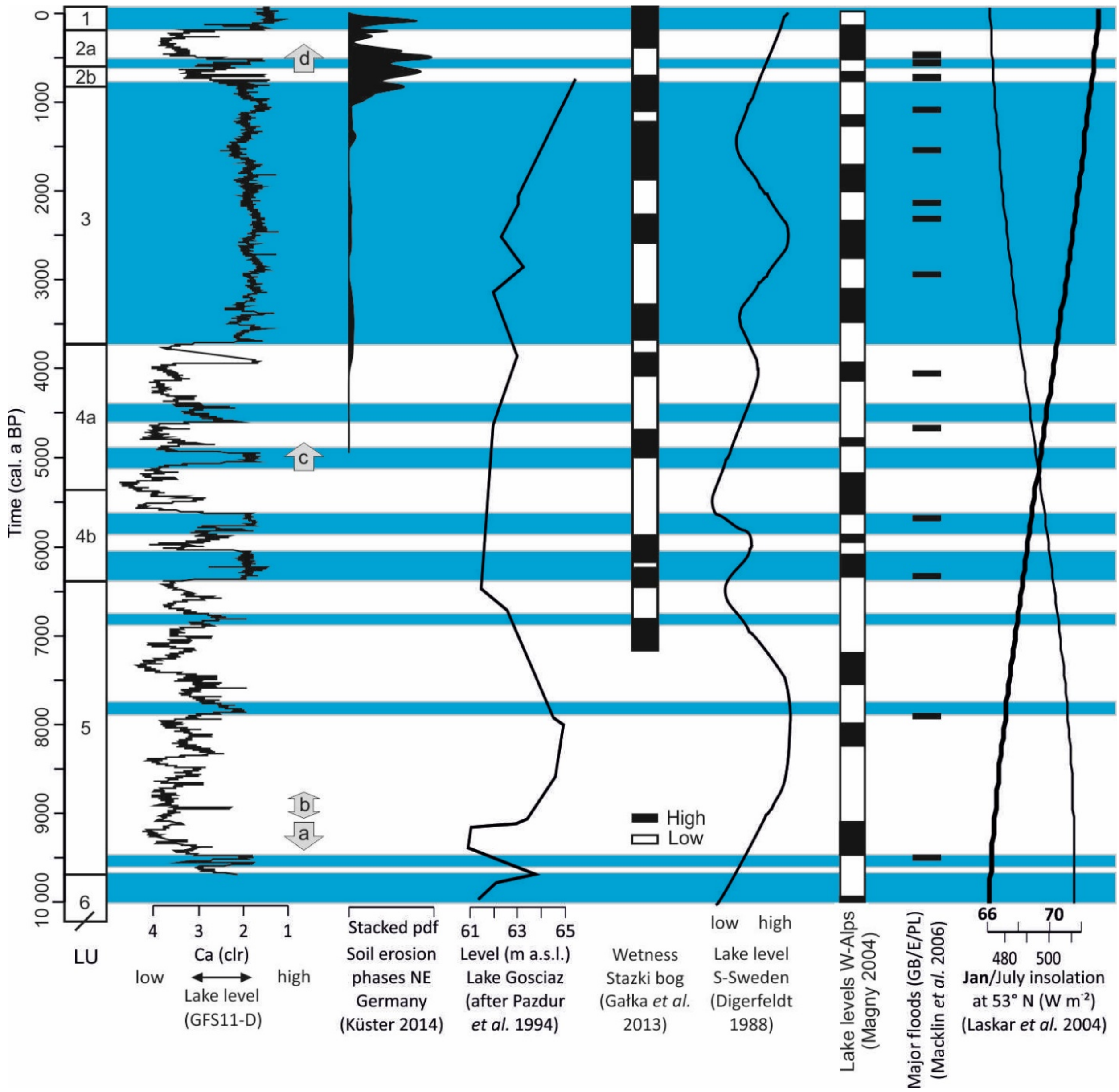

Fig. 6: Comparison of clr-transformed Ca record (reversed axis) and lithological units (LU 1 to 6) of core GFS11-D with regional soil erosion phases (as proxy for human impact), European proxies for wet phases (lake levels and floodings) and seasonal insolation. a = 4 m low stand reconstructed from sand layers in GFS11-C; b = peat in 90 cm below present lake level (Kaiser *et al.* 2014b); c = 4 m high stand reconstructed from littoral carbonate muds in GFS11-AB; d = 3-m high-stand reconstructed by Kaiser *et al.* (2014b). Blue bars in background indicate high-stand phases of Lake Fürstenseer See.

Tables

Table 1. Location of sediment cores in Lake Fürstenseer See and basinmorphometric attributes. Slope was measured using a 2 m buffer around the core location. Maximum fetch was generally from rare northerly wind directions, why a maximum fetch of the prevailing wind directions (SW to NW, P. Stüve, pers. comm. 2013) was calculated additionally.

Core name	Latitude (N) WGS84	Longitude (E) WGS84	Water depth (m)	Core length (m)	Slope (%±σ)	Distance to shore (m)	Maximum fetch (m)	Maximum fetch in prevailing wind direction (m)
GFS11-AB	13° 10' 11.6	53° 18' 2.5	10	1.9	22.5±0.8	81	1624	972 (NW)
GFS11-C	13° 10 9.6	53° 18' 2.9	15.2	3.9	14.6±0.8	117	1594	930 (NW)
GFS11-D	13° 10' 7.6	53° 18' 3.4	19.8	8.5	8.5±1.0	156	1567	890 (NW)
GFS11-E	13° 10' 9.1	53° 18' 6.3	23.1	6.3	2.4±0.2	191	1470	1094 (WNW)
GFS10-8	13° 10' 8.42	53° 18' 1.67	17	0.7	14.9±0.1	118	1625	955 (NW)

Table 2. AMS radiocarbon dates from a transect of sediment cores at Lake Fürstenseer See. Calibration of dates using INTCAL13 (Reimer *et al.* 2013) in Oxcal 4.2. Cores are sorted after water depth. Date Poz-55889 was excluded from age-depth-modelling of core GFS11-D (Fig. 4).

Core name	Lithological unit	Composite sediment depth (cm)	Represented sediment (cm³)	Material	Mean ^{14}C-age (a BP)	Error (a, 1σ)	Min. cal. age (cal. a BP, 2σ)	Max. cal. age (cal. a BP, 2σ)	Mean cal. age (cal. a BP)	Error (cal. a BP, 2σ)	Lab.Code/ Reference
GFS11-A0	AB-LU2	19.25	1	*Pinus sylv.* bark	4350	80	4727	5288	5007.5	280.5	Poz-61451
GFS11-B1	AB-LU2	31.9	0.5	*Pinus sylv.* bark	4510	35	5046	5303	5174.5	129	Poz-55886
GFS11-A1	AB-LU2	97	1	*Larix decid.* cone	11040	100	12 729	13 076	12 902.5	174	Poz-61454
GFS11-A0	AB-LU2	104	1	Charcoal	11270	100	12 893	13 334	13 113.5	221	Poz-61452
GFS11-C1	C-LU3	131.15	10.5	*Pinus sylv.* bark	2960	60	2954	3335	3144.5	191	Poz-61455
GFS11-C1	C-LU5	220.8	2.3	*Pinus sylv.* bark	5960	80	6567	7005	6786	219	Poz-61456
GFS11-C2	C-LU7	365.5	0.5	*Arctostaphylos uva-ursi* seeds, *Pinus sylv.* needle, *Juniperus communis* needle	11430	60	13 135	13 413	13 274	139	Poz-55888
GFS10-8	LU2	69	0.5	plant remains	370	35	315	505	410	95	Poz-37472
GFS11-D0	D-LU1	36	1	*Pinus sylv.* bark	145	60	6	262	134	128	Poz-61457
GFS11-D1	D-LU2a	85.9	1	leaf remains	385	60	307	501	404	97	Poz-61458
GFS11-D0	D-LU3	191.7	0.5	*Pinus sylv.* needle	180	25	0	291	145.5	146	Poz-55889
GFS11-D1	D-LU3	197.2	1	Leaf remains	905	60	759	958	858.5	100	Poz-61459
GFS11-D2	D-LU3	379.7	19.5	*Pinus sylv.* bark	2365	60	2334	2715	2524.5	191	Poz-61460
GFS11-D2	D-LU4a	453	0.5	*Pinus sylv.* bark	3485	30	3647	3838	3742.5	96	Poz-55890
GFS11-D3	D-LU4a	573.8	0.5	*Pinus sylv.* bark	4645	35	5309	5466	5387.5	79	Poz-55891
GFS11-D3	D-LU4b	649.8	1	*Pinus sylv.* bark	5530	70	6212	6479	6345.5	134	Poz-61461
GFS11-D4	D-LU5	753.85	0.5	*Pinus sylv.* needle	8100	60	8655	9234	8944.5	290	Poz-55892
GFS11-E1	E-LU2a	159.4	1	*Pinus sylv.* bark	110	60	6	281	143.5	138	Poz-61462
GFS11-E3	E-LU3	531.34	1	*Pinus sylv.* bark	3535	60	3644	3978	3811	167	Poz-61464
GFS11-E3	E-LU4	552.34	1	*Pinus sylv.* bark	3685	60	3854	4225	4039.5	186	Poz-61465
GFS11-E4	E-LU4	561.76	2.3	*Pinus* sylv. bark, betula seed	3710	60	3893	4236	4064.5	172	Poz-61466
GFS11-E4	E-LU4	596.76	1	*Pinus sylv.* bark, betula seeds	3890	70	4095	4518	4306.5	212	Poz-61467

Table 3. Descriptions of lithological units (LU) in each core of the transect from Lake Fürstenseer See. Stratigraphy after visual descriptions (all cores), CONISS clustering using □-XRF data of cores C, D, E as well as microscopic descriptions for cores C,D,E and AB unit 2. Chronology estimated after dates from Tab. 2 and correlation with core GFS11-D, where age-depth-model could be calculated (Fig. 4).

Core	LU	Composite depth (cm)	Description	Estimated age (maximum cal. a BP)
GFS11-AB	1	0-18	Massive medium-to-coarse sand, wavy lower boundary	modern- ~4800
	2	18-27	Laminated carbonate mud with *Chara spec.* oogonia and thin layers of amorphous organic matter, wavy lower boundary	~4800-5200
	3	27-46	Laminated coarse to fine sand with carbonatic sand layers, diffuse lower boundary	~5200-12 600
	4	46-91	Normally graded sand, distinct lower boundary	~5200-12 600
	5	91-103.4	Laminated fine sand, wavy lower boundary	~12 900-13 100
	6	103.4-110.9	Massive dark brown sandy organic mud, wavy lower boundary	~13 300
	7	>110.9	Heterogenous massive sand and gravel layers, in-situ Arcto-staphylos uva-ursi and *Juniperus communis* roots on top	Deglaciation
GFS11-C	1	0-33.3	Blackish wet homogenous organic mud, distinct lower boundary	<300 (~AD1750)
	2	33.3-58.3	Grey carbonate mud with greyish-brown upper part, wavy lower boundary	~300 - ~900 (~AD1100)
	3	58.3.-136.7	Blackish, homogenous organic mud, diffuse lower boundary	~900 - ~3800 (~1700 BC)
	4	136.7-209.8	Alternating greyish to blackish laminated carbonate mud with two normally-grading sand layers	~3800 - ~6400 (~4400 BC)
	5	209.8-281.3	Laminated carbonate muds in different shades of grey with several thin sand layers	~6400 - ~9400 (~7500 BC)
	6	281.3-336.3	Blackish massive organic mud with well-preserved lacustrine organic matter, more carbonatic in the lower part	>9400
	7	336.3-366.6	Greyish to blackish calcareous mud with layers of yellowish-amorphous gel-matrix, slightly sandy	~13 300 at the base
	8	>366.6	Heterogenous massive sand and gravel layers, in-situ *Arcto-staphylos uva-ursi* and *Juniperus communis* roots on top	Deglaciation
GFS11-D	1	0-48.1	Blackish wet homogeneous organic mud, diffuse lower boundary	<300 (~AD1750)
	2a	48.1-108.9	Greyish-brown to grey finely laminated carbonate mud, diffuse lower boundary	~300 - ~630 (~AD1450)
	2b	108.9-181.9	Dark-greyish finely laminated carbonate mud with some organic-rich layers, distorted, diffuse lower boundary	~630 - ~900 (~AD1100)
	3	181.9-448.9	Blackish organic mud with faint laminations and distinct lower boundary	~900 - ~3800 (~1700 BC)
	4a	448.9-591.6	Laminations of carbonate mud in different shades of grey with layers of blackish partly massive organic mud, partly distinct boundaries	~3800 - ~5400 (~3200 BC)
	4b	591.6-650.4	Blackish organic mud with fine greyish laminae	~5400 - ~6400 (~4400 BC)
	5	650.4-782.9	Strongly alternating blackish to light-greyish laminae	~6400 - ~9670 (~7500 BC)
	6	782.9-845.4	Blackish homogenous organic mud with a thin grey carbonate layer	>9670
GFS11-E	1	0-58.3	Blackish wet homogenous organic mud	<300 (~AD1750)
	2a	58.3-164.9	Brownish-grey massive carbonate mud with faint laminations, diffuse lower boundary	~300 - ~630 (~AD1450)
	2b	164.9-247.1	Heterogenous, partly distorted, dark greyish carbonate mud with some organic-rich layers, partly finely laminated, diffuse lower boundary	~630 - ~900 (~AD1100)
	3	247.1-532.6	Blackish organic mud with faint laminations and distinct lower boundary	~900 - ~3800 (~1700 BC)
	4	532.6-630.3	Laminations of carbonate mud in different shades of grey with layers of blackish partly massive organic mud, partly distinct boundaries	~3800 - ~4500 (~2600 BC)
GFS10-08	1	0-48	Blackish homogenous organic mud, distinct lower boundary	<300 (~AD1750)
	2	48-68	Grey carbonate mud with greyish-brown upper part, wavy lower boundary to organic mud	~300 - ~470 (~AD1580)

Supporting Information

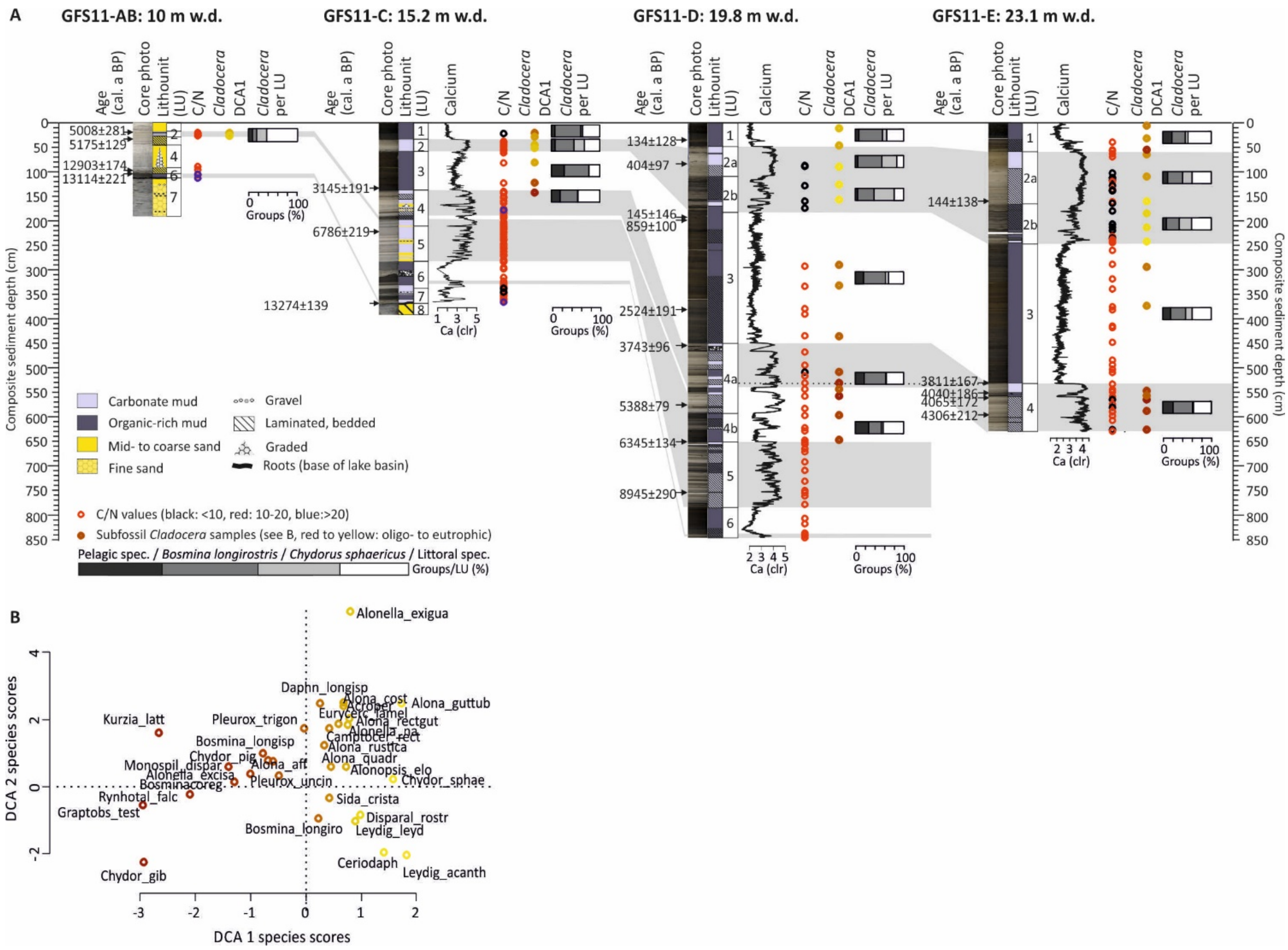

Fig. S1: A. Transect of Lake Fürstenseer See composite cores including results of μ-XRF-Ca records, C/N-ratios and subfossil *Cladocera*. Dates include their 2σ ranges (see Tab. 2). Grey bars show correlative lithological units. See legend and text for further descriptions. B. Species scores of subfossil *Cladocera* DCA with main trend along the first axis representing trophic conditions. Second axis represents proximity to water plants. Colours refer to sites scores (samples) of the first axis in the columns "*Cladocera* DCA1"in (A).

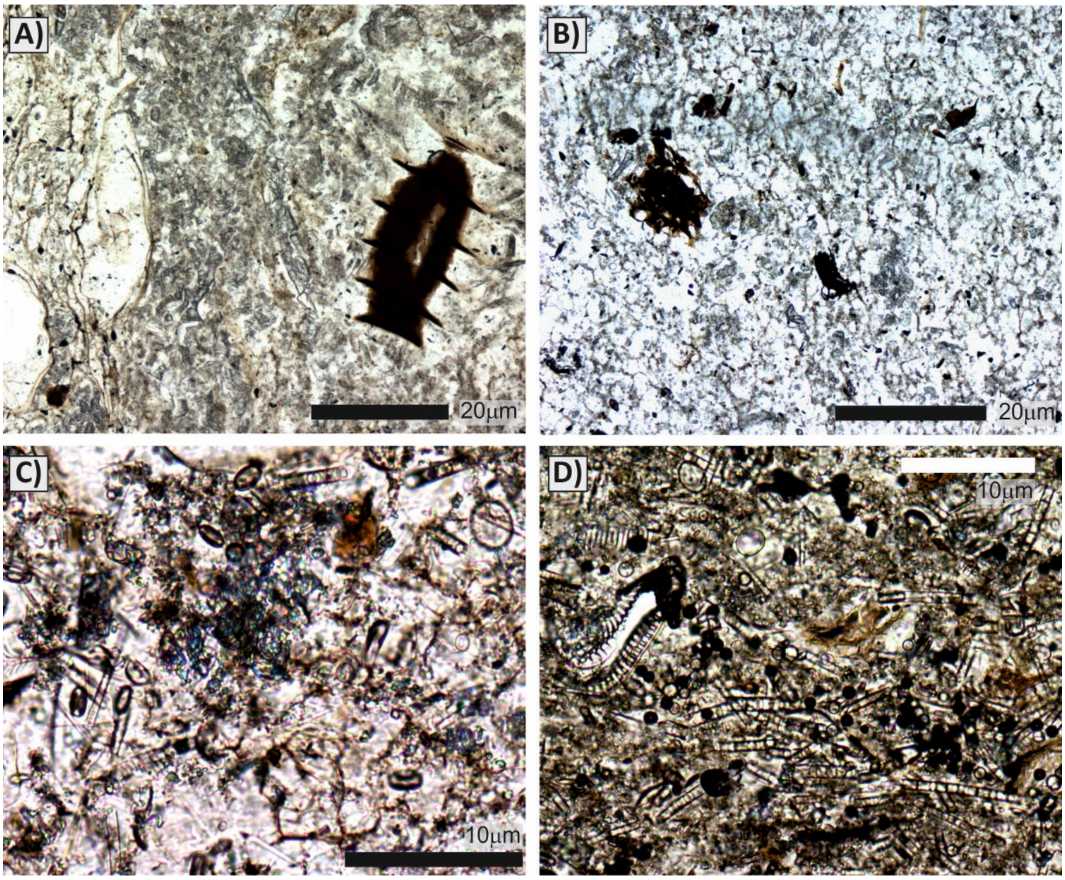

Fig. S2: Microscopic images from cores at Lake Fürstenseer See. A. Laminated carbonatic mud with few organic remains and *Chara spec.* oogonia (GFS11-AB: ~22 cm sediment depth, LU2). B. Decomposed organic mud with plant macro remains and little carbonate (GFS11-C: ~90 cm sediment depth, LU3). C. Diatom-rich sediment with carbonate clumps (GFS11-E: ~75 cm sediment depth, LU2a). D. Diatom- and organic-rich sediment with post-depositional pyrite (GFS11-E: ~555 cm sediment depth, LU4).

Table S1. Average values of discrete geochemical analyses sorted after cores and lithological unit (LU). SD is the respective standard deviation for at least 3 samples of the parameter to the left. Minimum and maximum values of a certain parameter were marked in bold-italics and bold, respectively. Grey bars mark blackish organic mud layers, other units were dominated by carbonate-rich muds.

LU	Dry density (g/cm³)	SD	Water content (%)	SD	TN (%)	SD	TC (%)	SD	TOC (%)	SD	TIC (%)	SD	CaCO3 (%)	SD	C/N	SD
A2	0.6	0.2	61.1	7.4	0.3	0.2	13.0	3.8	4.2	2.0	**8.7**	2.8	**72.7**	23.0	12.3	0.9
C2	0.1	0.1	88.1	4.6	1.7	0.8	21.4	6.9	17.5	8.1	3.9	1.6	32.8	13.2	10.5	0.3
C3	***0.1***	0.0	**92.7**	2.5	**2.9**	0.6	**32.6**	6.2	**31.2**	5.8	1.3	1.3	11.1	10.4	11.0	0.6
C4	0.4	0.5	74.0	21.5	1.0	0.5	15.6	7.0	11.9	5.5	3.7	2.0	30.8	16.7	13.0	1.0
C5	0.6	0.2	59.3	11.1	0.4	0.3	8.6	4.8	5.8	3.1	2.8	2.5	23.5	20.7	13.7	1.5
C6	0.3	0.1	77.2	3.8	1.5	0.8	18.5	7.6	16.6	8.1	1.9	1.4	15.8	11.7	11.3	0.5
C7	**0.8**	0.4	***54.7***	14.0	***0.3***	0.1	***4.9***	1.6	***3.4***	1.2	1.5	0.9	12.6	7.6	14.0	1.4
D2a	0.2	0.0	84.5	1.3	1.1	0.1	15.4	0.3	8.8	1.4	6.5	1.8	54.5	14.7	***8.2***	1.8
D2b	0.1	0.0	90.8	1.3	2.5	0.4	25.3	3.8	24.4	3.2	0.9	1.0	7.6	8.6	9.8	0.3
D3	0.1	0.0	92.1	0.5	2.6	0.2	29.5	3.3	28.3	3.5	1.1	0.9	9.3	7.8	10.8	0.4
D4a	0.1	0.0	87.7	3.1	1.9	0.6	23.9	4.0	20.7	6.3	3.2	2.5	26.4	20.9	10.7	0.8
D4b	0.1	0.0	88.7	2.5	2.5	0.7	30.8	4.8	29.3	6.6	1.5	2.1	12.3	17.7	11.7	0.6
D5	0.2	0.0	84.8	2.6	1.8	0.4	25.6	3.1	22.2	4.6	3.5	1.7	29.1	14.4	12.1	0.6
D6	0.3	0.2	77.5	13.3	1.7	1.1	19.7	10.6	18.1	11.6	1.6	2.4	13.5	19.6	12.5	2.2
E1	0.2	0.1	90.3	0.6	1.9	0.4	21.9	1.9	20.1	3.1	1.8	1.6	15.0	13.7	10.6	1.4
E2a	0.3	0.1	85.3	2.3	1.0	0.2	14.9	2.4	10.6	2.3	4.4	2.1	36.2	17.3	10.5	2.1
E2b	0.2	0.0	86.7	2.4	1.7	0.4	20.4	3.1	16.5	3.6	3.9	2.0	32.7	16.7	9.6	1.0
E3	0.1	0.0	89.6	1.7	2.1	0.3	22.9	2.5	22.5	2.8	***0.4***	0.5	***3.1***	4.3	10.8	0.3
E4	0.2	0.0	84.4	2.8	1.5	0.5	20.3	2.8	15.3	4.1	5.0	2.8	41.3	23.4	10.2	1.7

Table S2. Mean abundances of subfossil *Cladocera* in different units of cores GFS11. Abundances in %. Littoral species: *Acroperus harpae, Alona affinis, Alona costata, Alona guttata tuberculata, Alona quadrangularis, Alona rectangula/guttata, Alonella excisa, Alonella exigua, Alonella nana, Alonopsis elongata, Camptocercus rectirostris, Chydorus globosus, Chydorus piger, Eurycercus lamellatus, Monospilus dispar, Pleuroxus trigonellus, Pleuroxus uncinatus, Rynhotalona falcata, Sida cristallina*. Pelagic species: *Bosmina (E.) coregoni, Bosmina (E.) longispina, Daphnia longispina-group.*

Core - lithological unit	Pelagic	Littoral	Bosmina longirostris	Chydorus sphaericus	n
AB-LU2	9.0	61.8	8.3	20.8	2
C-LU1	7.2	35.2	53.6	3.9	2
C-LU2	7.7	31.5	38.7	21.9	3
C-LU3	26.5	22.2	51.3	0.0	1
C-LU4	18.1	44.9	36.2	0.7	1
D-LU1	7.6	34.6	48.8	9.1	2
D-LU2a	5.4	13.8	50.0	30.8	1
D-LU2b	5.1	19.3	34.0	41.6	2
D-LU3	17.3	30.7	45.2	6.7	3
D-LU4a	18.5	35.5	44.7	1.3	4
D-LU4b	28.0	42.1	26.3	3.7	2
E-LU1	19.8	50.8	29.5	5.9	3
E-LU2a	12.5	48.8	38.7	24.4	3
E-LU2b	10.0	54.2	35.8	33.9	3
E-LU3	18.5	45.2	36.2	15.7	2
E-LU4	21.7	37.2	41.1	2.0	5