A chrysophyte-based quantitative reconstruction of winter severity from varved lake sediments in NE Poland during the past millennium and its relationship to natural climate variability

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Abstract

Chrysophyte cysts are recognized as powerful proxies of cold-season temperatures. In this paper we use the relationship between chrysophyte assemblages and the number of days below 4°C (DB4°C) in the epilimnion of a lake in northern Poland to develop a transfer function and to reconstruct winter severity in Poland for the last millennium. DB4°C is a climate variable related to the length of the winter. Multivariate ordination techniques were used to study the distribution of chrysophytes from sediment traps of 37 low-land lakes distributed along a variety of environmental and climatic gradients in northern Poland. Of all the environmental variables measured, stepwise variable selection and individual Redundancy analyses (RDA) identified DB4°C as the most important variable for chrysophytes, explaining a portion of variance independent of variables related to water chemistry (conductivity, chlorides, K, sulfates), which were also important. A quantitative transfer function was created to estimate DB4°C from sedimentary assemblages using partial least square regression

(PLS). The two-component model (PLS-2) had a coefficient of determination of $R^2_{cross} = 0.58$, with root mean squared error of prediction (RMSEP, based on leave-one-out) of 3.41 days. The resulting transfer function was applied to an annually-varved sediment core from Lake Żabińskie, providing a new sub-decadal quantitative reconstruction of DB4°C with high chronological accuracy for the period AD 1000-2010. During Medieval Times (AD 1180 – 1440) winters were generally shorter (warmer) except for a decade with very long and severe winters around AD 1260 – 1270 (following the AD 1258 volcanic eruption). The 16th and 17th centuries and the beginning of the 19th century experienced very long severe winters. Comparison with other European cold-season reconstructions and atmospheric indices for this region indicates that large parts of the winter variability (reconstructed DB4°C) is due to the interplay between the oscillations of the zonal flow controlled by the North Atlantic Oscillation (NAO) and the influence of continental anticyclonic systems (Siberian High, East Atlantic/Western Russia pattern). Differences with other European records are attributed to geographic climatological differences between Poland and Western Europe (Low Countries, Alps). Striking correspondence between the combined volcanic and solar forcing and the DB4°C reconstruction prior to the 20th century suggests that winter climate in Poland responds mostly to natural forced variability (volcanic and solar) and the influence of unforced variability is low.

Keywords: Last millennium; Poland; Lake sediments; Chrysophytes; Cold-season; External forcings

1. Introduction

Comprehensive reconstructions of continental to global-scale temperature variability during the past millennia have demonstrated the value of paleoclimate proxy data and provided insight into natural forced and unforced variability, and anthropogenic disturbances (PAGES 2k Consortium, 2013). This information is a key to reduce uncertainties for future climate change (Hegerl et al., 2006).

Although the number of suitable proxy data sets has increased in recent times, some important challenges remain. Winter season reconstructions using paleolimnological records are scarce, as only few proxies, such as diatoms and chrysophytes, are sensitive to cold conditions in the lake via the amount or duration of ice-cover (e.g. Kamenik and Schmidt, 2005; Rühland et al., 2015; Weckström et al., 2014). Even for Europe, where the data base is arguably very good, most of the proxy series used for millennial-long regional climate reconstructions record predominantly the summer season (e.g. PAGES 2k Consortium, 2013 and references therein; Trachsel et al., 2010). However, at least for Europe it is very well known that the structure of summer temperature variability differed very much from the winter season (de Jong et al., 2013b; Luterbacher et al., 2004) implying that the picture about past climate variability as concluded from the summer season is incomplete.

Moreover, the proxy sites are often not evenly distributed in space. Using pseudo-proxies and climate model data Pauling et al. (2003) and Küttel et al. (2007) have shown for Europe that the location of predictor sites matters : while most of the proxy sites are located in Western Europe, sites in Central Europe, the Baltic and SE Scandinavia are missing; but information from these areas would be very helpful.

In this article we present a quantitative reconstruction of cold season climate from varved lake sediments in NE Poland for the past 1000 years. Information about winter temperature from this area is very important because of the very strong relationship between Polish and European temperatures both at the interannual and interdecadal time scales (Luterbacher et al., 2010). Moreover, winter climate in NE Poland is affected by major atmospheric circulation patterns and modes in the north Atlantic European domain, especially the North Atlantic Oscillation (NAO) and the East Atlantic/Western Russia (EA/WRUS) pattern (Luterbacher et al., 2010; Luterbacher et al., 2004). This is clearly evidenced in snow cover duration (Falarz, 2007), lake ice-cover (Marszelewski and Skowron, 2006; Wrzesiński et al., 2013), documentary and early instrumental data (Luterbacher et al., 2010; Przybylak et al., 2005; Przybylak et al., 2003) for the late 19th and 20th century.

Poland is also interesting because of its wealth of long high-quality instrumental records back to AD 1780. For periods prior to instrumental observations, there exist a few successful attempts at reconstructing air temperature by means of documentary sources, chronicles and meteorological observations (Przybylak et al., 2005; 2010; Sadowski, 1991), or quantitative reconstructions of air temperature based on historical sources with high resolution (daily). Such data are available for shorter discrete periods of several dozens of years (e.g. Bokwa et al., 2001; Limanówka, 2000; Michalczewski, 1981; Przybylak and Marciniak, 2010; Przybylak et al., 2014). The state of knowledge of Polish climate before the 16th century is generally modest and the existing reconstructions are uncertain (Przybylak, 2011; Przybylak et al., 2010). Natural paleoclimate records with high (interannual to subdecadal) resolution that reach further back than AD 1550 are extremely scarce, and only based on tree-rings widths (Koprowski et al., 2012; Przybylak et al., 2005; Szychowska-Krąpiec, 2010). A synthesis of the longest tree-ring based climate reconstructions for Poland is presented by Zielski et al. (2010). Most surprisingly, studies using varved lake sediments are missing despite their demonstrated potential for quantitative seasonal paleoclimate studies (Amann et al., 2014; Bonk et al., 2015; Larocque-Tobler et al., 2015). Recent work has also highlighted the potential of raised bogs as paleoenvironmental and paleoclimate archives in Poland (De Vleeschouwer et al., 2009; Gałka et al., 2014; Lamentowicz et al., 2015; 2009).

Chrysophytes (Chrysophicea and Synophicea group) are excellent candidates to provide information about winter climate. These golden brown freshwater algae have demonstrated skills in paleolimnological reconstructions due to their sensitivity to different environmental and climatic factors, and their abundance and good preservation in lake sediments (Smol, 1995). Although water chemistry is known to be one of the major factors controlling chrysophyte distribution (e.g. Cumming et al., 1992; Cumming et al., 1993; Dixit et al., 1989; Duff et al., 1997; Hernández-Almeida et al., 2014; Pla and Anderson, 2005; Pla et al., 2003; Zeeb and Smol, 1995), other studies have shown that temperature is also an influential factor. Cronberg (1973, 1980) reported blooms of chrysophytes beneath winter ice in Scandinavian lakes and observed some species only producing cysts after a specific temperature threshold was crossed. This suggests that some cysts forms may be useful paleoecological indicators of winter temperatures.

Quantitative methods relating chrysophyte assemblages and air/water temperature during winter have been developed on high-altitude sites in the Alps, the Pyrenees and the Andes (de Jong and Kamenik, 2011; de Jong et al., 2013b; de Jong et al., submitted for publication; Kamenik and Schmidt, 2005; Pla and Catalan, 2005). In the Alps, chrysophyte inference models for high altitude lakes have so far been developed for relatively small geographic regions. Kamenik and Schmidt (2005) developed a chrysophyte-based temperature transfer function in the Austrian Alps. Results in their paper showed high correlations between cold-season temperatures (October-May) and chrysophyte assemblages in modern samples of 29 high altitude lakes. The resulting cold-season temperature transfer function based on chrysophytes was later applied to varved sediments from Lake Silvaplana in the Swiss Alps by de Jong et al. (2013a; 2013b). Reconstructions show similar results when compared with other historical records from the Alps. Moreover the authors found high similarities to NAO and 'Siberian high pressure' indices for the last millennium. A similar approach was used by Pla and Catalan (2005) in high-altitude lakes from the Pyrenees. In that study, modern chrysophyte assemblages were highly correlated to air temperature anomalies along an altitudinal gradient among the studied lakes. Using a similar methodology as proposed in the present study, de Jong et al. (submitted for publication) developed a chrysophyte-based inference model to reconstruct the number of consecutive days with temperatures below 4°C (length of winter) from a training set of lakes from the south-central Andes which were sampled along an altitude gradient substituting variations in austral winter temperature.

Although chrysophytes have been tested in high-altitude areas, the potential of these organisms as proxies of cold-season temperatures has been rarely explored in low-altitude regions. Brown et al. (1997) developed a temperature-based model using chrysophytes from 49 low-altitude lakes from northwestern Canada. Although the inference model was weak (R^2_{cross} = 0.23), results indicated that chysopyhtes may be used as supplement to other paleotemperature reconstructions. In the absence of temperature gradients with elevation (e.g. training sets in mountains), sampling along very large geographic areas is required in order to cover a significant thermal gradient in space (latitude or longitude). Poland is a good candidate to implement this kind of training set design because it is relatively flat (only 1% of the Polish area is above 1000 m), has a high portion of land covered by many lakes (Jańczak et al., 1996) and there is a significant longitudinal thermal gradient (Lorenc, 2005). Moreover, many of the lakes have varved sediments that allow for a precise chronology of paleoclimate reconstructions (Tylmann et al., 2013).

In this paper we develop and apply a quantitative chrysophyte cyst-based inference model of 'number of consecutive days below 4 °C' (DB4°C). DB4°C is a climate variable that is highly correlated to cold-season temperatures and winter length and is, thus an indicator of winter severity. A training set from 37 lakes in north-eastern Poland was developed to explore the potential of chrysophyte cysts as a proxy for cold-season conditions in this region. The resulting transfer function was then applied to a lake sediment core to provide a paleoclimate reconstruction. We selected Lake Żabińskie for the reconstruction because it is included in the training set, has continuously varved sediments that provide an excellent age control and has sufficiently thick varves for high-resolution sampling (Bonk et al., 2015; Tylmann et al., 2013). The reconstruction covers the period AD 1000-2010. Our goal is to provide a winter severity record at sub-decadal resolution (5-year) for Poland. Comparisons with other temperature-sensitive winter season records are made to study teleconnections with atmospheric circulation patterns and natural forcing factors over Central Europe.

2. Study area, material and methods

2.1. Training set data collection

The 50 training set lakes (Fig. 1A) were chosen among more than 2900 low-land lakes of northern Poland (Jańczak et al., 1996), covering a broad range of morphological, physical, and chemical parameters (Hernández-Almeida et al., 2014). Sediment traps equipped with thermistors were deployed in the selected lakes. Field surveys were conducted during October 2011 and November 2012. After one year of exposure, only 37 sediment traps and thermistors were found. The distribution of these lakes is along a longitudinal gradient that is related to an E-W winter temperature gradient of ~4°C during the year of exposure between October 2011 and November 2012 (Fig. 1A). The spatial extension of the study area is large (52º31'- 54º19' N, 14º37'- 22º53'E), covering more than 700 km from the westernmost to the easternmost lake. Training set lakes are generally small (<20 ha), below 260 m.a.s.l., between 6 and 44 m deep and slightly alkaline (pH ranging from 6.5 to 9), with moderate agricultural activity and/or forestry in the catchments. Additional limnological and geographic characteristics are summarised in Table 1.

Sediment traps consisted of PVC-liners with a length of 80 cm and a diameter of 9 cm, 2 tubes per trap (Bloesch and Burns, 1980), with the openings of the traps approximately 1.5 m above the lake bottom. Collection of environmental data in each lake was made every three months. Water chemistry (conductivity, oxygen, pH) and turbidity were measured at 2 m water depth. Water samples were collected at 1 m water depth from each training set lake. Water chemistry analyses were performed at the University of Gdansk for major ion concentration analysis (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, sulfates, fluorides, chlorides) as well for total P and N. Analyses were performed using ion chromatography (ICS 1100, Dionex, USA) equipped with an IonPack AS22 column for anions and an IonPack CS16 column for cations, and colorimetric method and a Spectroquant NOVA 400 spectrophotometer (Merck), respectively. Monthly air temperatures for the studied lakes were obtained from the Institute of Meteorology and Water Management-National Research Institute of Poland. Water temperature in the epilimnion (2-m depth) was recorded by HOBO U22-001 thermistors at 15-minutes intervals during sediment trap exposure until trap recovery in November 2012.

It is difficult to determine freeze and ice break-up dates with thermistors due to the complexity of the freezing and thawing process (Gabathuler, 1999; Kamenik and Schmidt, 2005). Ice-cover formation on a specific lake does not only depend on the air temperature decrease, but also on lake size, morphology, inflow to the lake and chemical properties of the water. Moreover, a thick ice-cover insulates the lake from further cooling during exceptionally cold periods. For these reasons we have chosen as a target climate variable the 'number of consecutive days below 4°C' (DB4°C). This is a more objective parameter to determine a period of cold conditions in the lake (i.e. length of the winter) and serves as an indicator of winter severity. Fresh water reaches its maximum density at 4°C and further cooling reduces convective overturning. DB4°C was calculated from individual thermistor logs for each lake. In the year of observation (2011/2012) temperatures in the upper water column exhibited a strong linear correlation with winter air temperature (Dec-Feb) during the study period (R= 0.71). This is attributed to the relatively short ice-cover duration during the year of the training set experiment: epilimnetic waters were not insulated from air temperatures. In years with very long ice-cover and very cold winters, the water temperature in the epilimnion is often uncoupled from air temperature due to the insulating effect of thick ice (Livingstone and Lotter, 1998).

2.2. Reconstruction site

Lake Żabińskie (54°07'54.5'' N; 21°59'01.1'' E; 120 m a. s. l) occupies a glacially eroded depression formed during the Vistulatian glaciation, surrounded by a low-land, hilly-moraine landscape typical of the Masurian Lakeland, northeastern Poland (Szumański, 2000). It is a relatively small lake

(41.6 ha) with the maximum depth in the central part of the lake (44.4 m, Fig. 1B). A major outflow is located in the south-western part, connecting Lake Żabińskie with much larger Lake Gołdopiwo. Two smaller inflows enter the lake from its southern side and a larger inflow in the northeast comes from Lake Purwin. Lake Żabińskie is eutrophic, hardwater, dimictic (mixes in early spring and fall) and anoxic during most of the year. Precipitation and temperature are higher during the summer months (up to 90 mm and 17°C, respectively) and lower during winter. According to in-situ observations made during the field surveys, the lake was ice-covered from mid-January 2012 to March 2012.

The climate in the Masurian Lakeland is characterized by strong continentality with cold winters and warm summers. The mean annual air temperature (measured between AD 1971 and 2005) is 6.5°C with the coldest temperatures occurring in January and the warmest temperatures in July and August. The mean annual precipitation ranges from 550 to 600 mm (Lorenc, 2005). The region is covered by snow for 1.5 to 3 months in winter.

The village of Żabinka, established in AD 1713, is located 0.5 km from the southern shore of the lake and had about 250 inhabitants in the 19th century and 200 in AD 1939. Between AD 1910 and 1920, a restaurant and a recreation place (campsite and beach) were built on the northern shore of Lake Żabińskie. These infrastructures were used for military purposes during the Second World War, and recovered for recreation activities during the mid-1950s. New infrastructures were built in the early 1970s and during 1980s due to the increase of activities related to tourism.

2.3. Sediment core and chronology

Overlapping sediment cores were collected from the deepest part of Lake Żabińskie (42.6 m) using a 90-mm diameter UWITEC gravity and a piston corer. The cores were then split lengthwise, cleaned and macroscopically correlated using distinct laminae. A composite sediment profile of 348.3 cm was obtained in this way. After initial description and photographic documentation, half-core A was used for the chronology and half-core B was sampled at annual resolution (varve-by-varve) for destructive analyses. Samples for the study of fossil chrysophyte assemblages were taken from every varve.

The regular succession of two types of laminae along the sediment core is interpreted to reflect seasonal variations in the sedimentation environment (Zolitschka, 2007): the spring/summer light layer is mainly composed of autochthonous calcite and diatoms; the dark fall/winter layers consist mainly of decomposed organic matter and clay minerals. The laminae couplets are interpreted as biochemical varves (Tylmann et al., 2013). Although the varve structure was recognized to be more complex and included multiple calcite layers within a one-year-deposition, it has a remarkable potential for a reliable varve chronology and for multiproxy paleoenvironmental reconstructions (Bonk et al., 2015).

The calendar-year chronology was established by performing microscopic analyses of thin sections and varve counting. The preparation for thin section analysis followed the procedure of Lotter and Lemcke (1999) with sediment blocks placed in liquid nitrogen, freeze-dried and subsequently impregnated with Araldite©2020 epoxy resin. After that, thin sections were prepared and inspected under a petrographic microscope. Varves were counted manually repeated times by three different persons using the CooRecorder software. Based on three independent counts the uncertainty of the varve chronology was estimated according to the following procedure: (1) varves indicated in all three countings were added to the chronology without increasing the uncertainty; (2) varves missed in one counting were added to the chronology and the uncertainty was also increased by 1 year in the 'minus direction' (toward younger ages); (3) varves missed in two countings were not added to the chronology but the uncertainty in the 'plus direction' (toward older ages) was increased by 1 year.

2.4. Sample preparation and cyst analyses

Sediment trap and sediment core samples (0.2 g wet sediment) were treated with H_2O_2 and HCl for chrysophyte cysts analyses, following the standard diatom procedure (Battarbee et al., 2001). Samples were then washed with distilled water and filtered with Milipore nylon-filters (63 µm) to remove large particles. Chrysophyte cysts were analysed using a scanning electron microscope (Carl-Zeiss EVO40). For the training set samples, a minimum of 300 modern cysts were counted per sample. For the downcore fossil cyst assemblages, 80 chrysophytes per sample were counted for samples between AD 2010 and 1898, and 50 cysts per sample until AD 1000. The final reconstruction was based on 5-yr running means. Identification of cysts followed Duff et al. (1995), Wilkinson et al. (2002) (labelled as PEARL -Paleoecological Environmental Assessment and Research Laboratory-), and Huber et al. (2009) (labelled as L). The unpublished new cysts were assigned a new number using the code 'ZAB' for the fossil and modern samples of Lake Żabińskie, and the code 'TSP' (Training Set Poland) for the modern samples.

2.5. Statistical analyses

Multivariate statistical analyses were used to identify major environmental gradients and to explore relationships between chrysophyte cysts and environmental variables. Prior to statistical analyses, environmental variables were explored for normal distribution. Water chemistry variables (except pH) were log transformed. For the climate related variables, only DB4°C and the seasonal temperatures showed normal distribution and were not transformed. Chrysophyte data were expressed as relative abundances. All species present in at least 2 training set lakes with abundance >1% were retained in the numerical analyses. Chrysophyte abundances were square-root transformed in order to stabilise variances.

Principal components analysis (PCA) was used to detect the major gradients in the environmental data (ter Braak, 1987). Significance of the axis was tested by the broken stick model (Jackson, 1993). Detrended correspondence analysis (DCA) (Hill and Gauch, 1980) was used to estimate the compositional gradient lengths along the first DCA axes, and the use of unimodal or linear numerical techniques for modelling the relationship between the chrysophyte cysts and environmental variables (Birks, 1995; ter Braak, 1987). Multivariate ordination techniques were used to examine the relationship between the taxa and the environmental data, and to identify redundant environmental variables and samples with unusual species composition. Because the species data was very heterogeneous, we used Hellinger's transformation (Legendre and Gallagher, 2001). Weighted correlations and variance inflation factors (VIF) were used to identify the variables which were intercorrelated. On this basis some variables were deleted from subsequent analyses. The choice of the best subset of predictors was tested using Akaike's information criterion (using the function *ordistep* in the VEGAN package) with significant variables ($p<0.05$), which explain the greater proportion of variation in the chrysophyte cyst community, . Individual RDA was performed to identify how much independent variation of the chrysophyte data was explained by the remaining variables. Ordinations were performed using R (R Development Core Team, 2009) with the add-on packages VEGAN (Oksanen et al., 2006).

2.6. Transfer functions

To establish transfer functions, the following models were tested: Weighted averaging (WA) classical and inverse deshrinking, weighted averaging – partial least squares (WA-PLS), partial least squares (PLS), and a modern analogue technique (MAT) as implemented in the computer program C2 (Juggins 2003). The minimal adequate inference model was identified as having the highest coefficient of determination (R^2), lowest mean and maximum bias and root mean square error of prediction (RMSEP), all based on leave one out cross-validation (Birks, 1995). The number of components in PLS and WA-PLS was chosen in function of the reduction of the RMSEP. To be considered 'useful', a component should give a reduction in prediction error of 5% or more of the RMSEP for the simplest one-component model (Birks, 1998). Outliers were identified as samples having an absolute residual (observed – predicted) higher than the SD of the environmental variable of interest and a low influence on the model indicated by Cook's D. The chrysophyte transfer function was generated by the software C2 (Juggins, 2003).

Analogue quality was used to diagnose the quality of the reconstruction by evaluating how well the calibration set of modern samples provided analogues for the fossil core samples (Birks et al., 1990). The taxonomic distance between each fossil sample and the calibration set was calculated using a squared chord distance (SCD) as a dissimilarity measure (Overpeck et al., 1985). Fossil samples with dissimilarity larger than the 10^{th} percentile lacked good modern analogues (Birks et al., 1990).

3. Results

3.1. Calibration data set and ordination

The PCA of the environmental data determined two major gradients related to the two first axis, both being significant according to the broken stick model: PCA 1 (eigenvalue 0.36) is related to water chemistry (major ions, nutrients and conductivity), and PCA 2 (eigenvalue 0.11) is related to climate variables (seasonal air temperatures and DB4°C) and longitude (Fig. 2A). Water chemistry variables are strongly correlated in the lakes of the training set. In the same way, climate related variables are also correlated among each other. The strongest correlation is found between longitude, winter temperatures (Dec-Feb) and DB4°C, reflecting the relationship with regional air temperature. This is also an ecological gradient (E-W) across northern Poland (Fig. 1A).

Of the 129 taxa encountered in the 37-lake training set, 63 fulfilled the criteria to be included in the numerical analyses. The gradient length of the first axes was 1.7 standard deviation (SD) units, indicating that numerical methods based on a linear response model are most appropriate (Birks, 1995). RDA was performed on the biological and environmental dataset (Fig. 2B). The PCA (Fig. 2A) and high VIFs (>20) revealed that there is strong collinearity among several of the environmental variables. Of all the environmental variables considered (Table 1), the *ordistep* procedure selected only six variables for entrance into the final model. (conductivity, K, chlorides, sulfates, DB4°C and longitude), which explained the variation in the chrysophyte data (28%) almost as well as the complete set of all environmental variables included in the ordination analyses.

A series of partial RDAs run with one explanatory variable at a time indicates that DB4°C and conductivity are the strongest variables and capture 4.5% and 6%, respectively. This low explained variance is typical of noisy, heterogeneous datasets which contain many zero values (Jones and Juggins, 1995). As the data set covers a long geographical transect, some species show scattered occurrences, limited to a few sites, or centred at a particular region of the training set. Similar results with low explained variance (<3%), but yet ecologically informative, were also reported in diatom data sets, for reconstructing salinity in Australia (Saunders 2011). Variance partitioning demonstrates that relationships between chrysophytes and DB4°C are independent of variables related to water chemistry (conductivity, chlorides, K, sulfates). The ratios between the first (constrained) and the second (unconstrained) eigenvalue (λ_1/λ_2) are calculated to assess which variables could be used for a transfer function development (Birks, 1995). Although DB4°C λ_1/λ_2 is the highest of all variables (0.71) and the one of climate interest, values lower than 1 indicate that the training set is affected by other variables that have an influence on the chrysophyte cyst composition but were not measured.

3.2. Transfer function

The final regression and calibration model was a Partial Least Square model (PLS) (ter Braak and Juggins, 1993), because linear methods are more appropriate for short ecological gradients. A transfer function was developed for DB4°C, because this variable is correlated with longitude and hence with Dec-Feb air temperature (Fig. 2A) which is of great paleoclimatic interest. DB4°C explains a significant amount of variation in the cyst assemblages (4.5%), even when effects of other important variables (conductivity, pH or nutrients) were removed. The Linear PLS model with 2 components provided the minimal adequate model with highest R^2 (cross-validated) and lowest mean and maximum bias.

The two-component PLS model (R^2_{cross}= 0.58; RMSEP= 3.41 DB4°C) (Fig. 3A) was significantly improved compared with the one-component model in terms of RMSEP (5.8 % change of DB4°C). Six lakes (TRZ, BRO, SUR, LEK, SZOS, SZE) were identified as outliers because they had absolute residuals (observed – predicted DB4°C) higher than the SD of 7.6 days, and low Cook's D. The residuals (Fig. 3B) show a slight trend with observed DB4°C, indicating an overestimation at sites with low DB4°C and an underestimation at sites with high DB4°C.

3.3. Ordination of fossil cyst assemblages

An ordination was carried out on the entire cyst assemblage dataset (Fig. 4). A total of 85 chrysophyte cysts types with maximum abundance >2% and a presence at least in 2 samples were identified in the 920 sediment core samples corresponding to 1010 years. Preservation was good and no clear signs of corroded cysts were observed. To obtain the gradient length of the dataset, a DCA was performed, which resulted in a gradient length of 1.5 SD. Therefore, a PCA was applied because it assumes a linear response of the species to the environmental variable of interest. The eigenvalues of the two PCA axes accounted for 19% of the variance in the fossil chrysophyte data. Figure 5 shows that the pattern of PCA 2 scores is highly similar to the reconstructed values of DB4°C, while the scores of PCA 1 are different but resemble the pattern of dry bulk density with low variability until the abrupt shift between AD 1800-1850.

3.4. Downcore reconstruction of DB4°C

Using the PLS-2 model, the chrysophyte-DB4°C transfer function was applied to the fossil chrysophyte assemblages in the sediments of Lake Żabińskie (Figs. 4 and 5). The sediment depths were converted into calendar ages according to the varve chronology shown in Figure 5. Reconstructed DB4°C (5-year running mean) fluctuated between 94 and 106 days. The reconstruction indicated highest DB4°C (> 104) at the beginning of the record (AD 1000-1070). After this date, there is a decreasing trend toward lower values of DB4°C until AD 1245 (94 DB4°C). Subsequently a rapid increase toward high DB4°C values (severe winters) is observed for a short period. Afterwards, reconstructed values return to lower DB4°C values until a minimum of 94 DB4°C is reached at AD 1433, followed by an abrupt increase up to 106 DB4°C. The period from AD 1180–1440 coincides with the warm phase of the Medieval Climate Anomaly (MCA). Higher values of reconstructed DB4°C (long winters) are recorded until AD 1730. A decrease in DB4°C values is observed between AD 1750 and 1850 (ranging 94-97 DB4°C) interrupted by a short colder period around AD 1800-1820. After AD 1850, values of DB4°C increase until a maximum of 106 DB4°C around AD 1945. From then onwards there is a steady decreasing trend of DB4°C values (shorter winters).

The range of DB4ºC values in the calibration dataset covers the range of the reconstruction (Fig. 4), ensuring that no linear extrapolation was required for the reconstruction period. Results of dissimilarity analyses are illustrated in Figure 4. In general, there are larger dissimilarities (in squared chord distance) between sediment-core chrysophyte assemblages and the Polish calibration dataset during periods with lower DB4°C values. Analogue matching analysis indicated that the upper part of the core (second half of 20th century) had a better analogue with the modern calibration samples than downcore (Fig. 4). However, despite this weak analogy, reconstructed DB4°C values show a significant negative correlation ($R=-0.35$, $p<0.05$) to cold-season January-March temperatures during the period where instrumental observations are available.

4. Discussion

4.1. Proxy interpretation and model performance

The final chrysophyte-based transfer function was based on the proxy DB4°C derived from the thermistor data. Many empirical studies have shown water temperatures to be strongly related to ambient air temperature in absence of ice-cover (Livingstone and Lotter, 1998; Shuter et al., 1983). DB4°C was chosen because it showed the highest λ_1/λ_2 ratio and explains a higher proportion of variance independently than other variables included in the training set; and it is of climate interest. Modern data show that there is no strong correlation between DB4°C and water chemistry (Fig. 2A) and a proportion of the variance of the modern chrysophyte cyst assemblage is explained by DB4°C independently of the other most imporant variable (conductivity) (Fig. 2B). Pla and Catalan (2005) and Kamenik and Schmidt (2005) found a similar orthogonal positioning of the cold-season related variables (altitude and spring mixing) with respect to variables related to water chemistry, suggesting that a portion of the variation in their cyst assemblages was independent from the chemical composition of the lake waters.

A λ_1/λ_2 lower than 1 may indicate that there is another secondary gradient, or combination of them, slightly larger than the first constrained axes that reflects variation in the chrysophyte assemblages that is unrelated to the DB4°C. Adding more lakes to expand the DB4°C gradient of our data set would help to overcome these problems. The trend in the residual structure of the PLS-2 model indicates overestimation at sites with low DB4°C values and underestimation at sites with high DB4°C values. In consequence, the predictive ability of the transfer function is greater for sites with medium values of DB4°C than for sites falling at the extremes of the environmental gradient. This 'edge effect', however, is inherent to PLS and its unimodal counterpart WA-PLS, both of which utilize an inverse deshrinking regression (Birks, 1995; ter Braak and Juggins, 1993).

The number of consecutive days with low water temperatures is an ecologically important variable for chrysophytes, because cyst-production is linked to low-temperature environments (Adam and Mahood, 1981) or takes place even under the ice (Cronberg, 1973). Rybak (1986) and Duff and Smol (1991) identified some chrysophyte cysts as typical of cold waters. The number of DB4°C affects the cyst assemblages, the timing and magnitude of cyst production during an entire year and shifts the percentage of cold and warm water species. The relationship with winter water temperatures agrees with other chrysophyte-based studies that found similar relationships between different variables related to cold-season temperatures and chrysophyte assemblages (de Jong and Kamenik, 2011; de Jong et al., 2013a; Kamenik and Schmidt, 2005; Pla and Catalan, 2005).

4.2. Relationship between DB4°C and cold-season temperatures

Studies converting non-climate variables (cyst assemblages) to climate variables imply that the relationships between both have remained constant through time. Correlation between the chrysophyte-based DB4°C reconstruction and the homogenized air temperature series for Lake Żabińskie (Larocque-Tobler et al., 2015) is highest for January-March temperatures (3-years filter; $R = -0.35$, $p>0.05$). Although a better correlation would be expected, this is explained because of distorting effects of ice-cover. In such periods, the ice-cover insulates the bulk of the water body from further cooling, and water temperatures are uncoupled from air temperatures (Livingstone, 1993). As the relationship between DB4°C and air temperatures is not linear but distorted by the presence of ice-cover, and to avoid accumulation of errors during the conversion to winter temperatures, we focus on the performance of chrysophyte cysts as proxies of changes in DB4°C. As we have shown in the modern training set data and instrumental data during the 20th century, DB4°C has a strong relationship with winter length.

4.3. Quality of the reconstruction

Transfer functions have been recently under the focus of the scientific community, urging proper validation and assesment of the reliability of the models applied (e.g. Juggins, 2013; Telford et al., 2004; Telford and Birks, 2011). One of the tests proposed is based on the similarity between the ordination axes that summarize the maximum variance in the dataset, and the reconstructions (Juggins, 2013).

An ordination was carried out on the entire cyst assemblage dataset (Fig. 5). The two first axes explain 11% and 8.5%, both are significant. This is typical of noisy data sets with a large number of taxa and many zero values to have a relatively low percentage of variance explained (Bennion, 1994). Comparison between scores of PCA axes with the reconstruction helps to evaluate how representative the reconstruction is compared to the major ecological changes of the fossil assemblage as shown in the ordination. Moreover, if anthropogenic effects are first factored out, the effects of natural variablity are likely to be more clearly detected. Reconstructed DB4°C and PCA1 are very different. PCA1 shows a very abrupt change at AD 1800 (Fig. 5C). This shift in the scores might reflect an anthropogenic change. Similarity between PCA1 scores and dry bulk density of the sediments denotes that chrysophyte ecological shifts downcore may be driven by environment changes in the lake. The shift around AD 1790 is correlated with a large increase in the dry bulk density of the sediments, reflecting that a large amount of fine material was deposited in the Lake Żabińskie and may have potentially altered the chrysophyte-environmental variable relationship (Fig. 5D). According to the percentage pollen data from Lake Żabińskie, abrupt forest decline and increase in ruderal and grassland taxa is observed after AD 1800, which is interpreted to be related to human impact on vegetation and forest clearance in the area (Wacnik, pers. comm. 2014). We argue that PCA1 scores reflect the extent to which the fossil cyst community of Lake Żabińskie responded to anthropogenic stressors, and therefore, ecological changes of the fossil chrysophyte assemblage shown in PCA1 are not related to climate factors. In contrast, downcore variability of scores of PCA2 do not show this abrupt shift at AD 1800, and they are highly correlated to the downcore reconstruction of DB4°C, indicating that this variable was important as well for the fossil assemblage, and is not related to anthropogenic changes (Fig. 5A-B).

4.4. Regional comparison of temperature variability

Although cold-season records are scarce in this region, we have compared our reconstruction with other records that are related to winter climate in the Baltic and across Europe (Fig. 6). Since our reconstruction is based on 5-yr running means, the main strength of the cold season temperature time-series from Lake Żabińskie (Fig. 6F) is in the decadal to multi-centennial spectral range.

Detailed comparison of our cold-season climate record from NE Poland with the other existing chrysophyte cyst-inferred cold season temperature reconstruction from the Eastern Swiss Alps (Lake Silvaplana; de Jong et al., 2013a; 2013b; Fig. 6E) is hampered by data gaps in the latter. Despite their locations in different geographic regions of Europe, both reconstructions show some common features: the very cold period following AD 1460, the prominent warming around AD 1740-1780, and the cooling phase in the early and mid-20th century followed by a warming trend from 1950 onwards. Although a bit earlier, the pronounced winter warming in the 18th century was also recorded in the multiproxy reconstruction by Luterbacher et al. (2004) who used mostly early instrumental data, documentary records, and a few ice core and tree ring datasets from Greenland and Siberia, respectively (Fig. 6D). A similar pattern on decadal and multi-decadal scales is also observed between our DB4°C reconstruction and documentary data from the Low Countries (van Engelen et al., 2001) (Fig. 6C) and the decadal winter climate series from instrumental and documentary sources from Central Poland (Przybylak, 2011; Przybylak et al., 2005, Fig. 6A). A consistent pattern between the Low Countries and the western Baltic in winter is reasonable because both areas are controlled by the NAO determining the severity of winters (Schmelzer and Holfort, 2011). Also the Baltic ice winter severity index values derived from documentary sources (Hagen and Feistel, 2005; Koslowski and Glaser, 1999; Schmelzer and Holfort, 2011) shows the mid-18th century warmth bracketed by severe winters around AD 1700 and 1800 (Fig. 6B). The Baltic ice winter severity index also shows the exceptionally cool period of the 1940s.

Although similarities with other winter-season climate records across Europe are evident, differences exist at some intervals. Following AD 1258, an abrupt cooling for 5-7 years is observed in our DB4°C reconstruction. This is not seen in the Low Countries temperature record (van Engelen et al., 2001). But this cold spell has been found in the winter temperature reconstruction from Lake Silvaplana whereby data gaps in the Silvaplana record are interpreted as very cold spells (de Jong et al., 2013b).

Interesting are the differences in the 16th century, i.e. at the beginning of the Spörer Minimum: while the Baltic ice winter severity index (Fig. 6B), the indices for the Low Countries (Fig. 6C) and other documentary winter climate records from Western Europe show a sequence of very severe winters in the 1430s followed by moderately cool winters after AD 1450 (Lamb, 2002), records from Poland show a different pattern: our DB4°C record from NE Poland, winter climate indices for Poland (Przybylak, 2011; Przybylak et al., 2005) and January-April air temperature for central Poland reconstructed from tree-ring widths of Scots pine (Przybylak, 2011) show 'normal' or even moderately warm winters in the 1430s. These were followed by strong cooling in the 1440s and peak cold around AD 1460. Cold conditions persisted throughout the rest of the 16th century. On the other hand, the very cold conditions in the second half of the 15th century did not occur in SE Poland (Szychowska-Krąpiec, 2010). This regional discrepancy during the beginning of the Spörer Minimum and the 15th century is reproduced in independent data sets and seems thus to be real. In Poland, the Maunder Minimum (AD 1645-1715) (Eddy, 1976) does not strike as the very cold phase reported in other European records (Glaser and Riemann, 2009; Luterbacher et al., 2004). Instead, a brief cooling phase can be appreciated between AD 1689–1699, that matches with a 11 years cool phase described in the Tatra Mountains, Poland (Niedźwiedź, 2010).

While the major features of Polish winter climate seem to be reproducible and spatially consistent, major uncertainties and inconsistencies remain in the details. This is partly attributable to reconstruction uncertainties but also, likely to a large degree, attributable to the diversity of the proxies used for comparison here. Our DB4°C reconstruction is sensitive to the length of the winter while other reconstructions are more sensitive to low temperatures, which is not the same in terms of climate. Therefore, it is unclear whether or not a better match among the different records should be expected.

4.5. Factors controlling winter severity

The agreement between short and long winters in Poland (DB4°C) with other European cold-season temperature records suggests a common large-scale atmospheric circulation mechanism. A first candidate is the North Atlantic Oscillation NAO which represents one of the most prominent modes of inter-monthly to inter-decadal variability in the Northern Hemisphere and is known to affect winter climate in Europe north of the Alps (e.g. Hurrell, 1995). Przybylak et al. (2005; 2003) correlated phases of postitive NAO with strong zonal circulation (westerly flow) and increased winter air temperatures over central Europe and Poland. According to Luterbacher et al. (2010) the NAO index accounts for more than 50% of the winter temperature variations in Poland.

Figs. 6F and H show the comparison between the NAO reconstruction by Trouet et al. (2009) and the winter severity (DB4°C) in NE Poland for the past millennium. Trouet et al. (2009) concluded persistently positive winter NAO conditions during the MCA (AD 1000-1450) and a strong shift around AD 1450 towards a regime with high variability and negative values during the Little Ice Age (LIA). Based on this comparison and on the current relative importance of NAO as a factor controlling winter temperatures in Central Europe, we propose that the shorter winters in Poland prior to AD 1450 were related to the predominance of enhanced westerly airflow (typical for positive NAO) over NW Europe during winter.

A rapid and strong drop from positive to very negative NAO values took place at AD 1450 (Trouet et al., 2009), which is also mirrorred in our DB4°C reconstruction (Fig. 6F). This shift in the NAO index has been interpreted as enhanced influence of easterly anomalies during the LIA (Trouet et al., 2009; 2012). During this period, other large-scale circulation phenomena such as the Siberian High (reconstructed by Meeker and Mayewski, 2002) might have gained more importance in the study area. Although the Siberian High has been observed to extert a large influence on air temperatures especially in Asia (Gong and Ho, 2002) and the Pacific (D'Arrigo et al., 2005), also eastern parts of Europe are under the influence of this anticyclonic field (Katsoulis et al., 1998). De Jong et al. (2013b), for example, found evidence of strong influence of the Siberian High on central Alpine winter temperatures during the LIA, implying that the influence of the Siberian winter anticyclone extended over Central Europe as far as the eastern Swiss Alps.

The very strong cooling (long winters) in the second half of the 15^{th} century and persistently severe winter conditions in Poland until ca AD 1700 coincided with high pressure conditions of the Siberian High (Fig. 6G, Meeker and Majewski, 2002) and transport of cold continental air to eastern Poland. Also in the 18^{th} and 19^{th} centuries the DB4°C and Siberian High index are remarkably similar.

Shorter winters between AD 1740-1790 and around AD 1850 match with minima of the air pressure in the Siberian High index suggesting again a stronger influence of Westerly flow.

The period with very long winters in the 1940s is the most remarkable feature of the 20th century winter climate in Poland. In fact, the decade of the 1940s shows the longest winters (highest DB4°C) for the past 900 years. This period corresponds to a NAO negative phase and weak polar vortex as the result of a long-lasting El Niño event, which is thought to have led to cold conditions in Northern and Central Europe (Brönnimann et al., 2004; Fischer et al., 2008). According to this scenario, it is not unlikely that the very long reconstructed DB4°C (long winters) in Poland were related to negative NAO conditions (Dickson et al., 2000) combined with prolonged blocking over Central and Eastern Europe with southward penetration of cold arctic air. Cold conditions in the 1940s were also recorded by the Ice Winter Severity and Winter Thermal indices for Poland (Fig. 6A-B). Moreover, northeastern Poland experienced a lot of snow between 1940-1950 (Falarz, 2004, 2007) which may have further delayed ice break-up (high number of DB4°C) in Lake Żabińskie.

4.6. Influence of volcanic and solar forcing on winter climate in Poland

The question about the causes of the rapid decadal-scale cooling events in Europe during the last Millennium and during the LIA in particular have been discussed controversially (Hegerl et al., 2011 and references therein). While anthropogenic influences can be excluded before the 20th century, variability in Total Solar Irradiance (TSI), volcanic aerosols (both external forcings; Crowley, 2000; Fig. 7C-D) or unforced internal (atmosphere-ocean) variability remain potential candidates to influence seasonal European temperatures and climate. Fig. 7 shows that the severity of winters in NE Poland (DB4°C) in the pre-industrial period (prior to the 20th century) is largely correlated to the combined solar and volcanic radiative forcing (data from Crowley, 2000) while this relationship is inverted in the 20th century.

TSI is generaly about 0.1 Wm^{-1} higher during the Medieval Climate Anomaly compared with the LIA and four multi-decadal periods of minimum solar activity are observed: the Wolf (AD 1280-1350), Spörer (AD 1420-1540), Maunder (AD 1645-1715) and Dalton Minima (AD 1795–1830) (Eddy and Oeschger, 1993). However, since solar minima coincided with strong negative volcanic forcing in the past millennium, it is difficult to discriminate the influence of both effects on regional surface temperatures. The reconstructed DB4°C does show severe winters during the Dalton Minimum but the other solar minima do not stand out as cold periods, whilethis is rather the case for the temperature indices from the Low Countries (Fig. 6C). The high DB4°C values (severe winters) at AD 1450-1470 might be partly related to the Spörer Minimum (in addition to a response to volcanic forcing after AD 1450). The severe winters between AD 1795-1820 might reflect the Dalton Minimum, combined with the effect of the large volcanic eruptions at AD 1809 and 1815-1816 (D'Arrigo et al., 2009). Fig. 7A shows a systematic offset in the match between the combined S+V forcings and the severity of winters (DB4°C) between the MCA and the LIA. It seems that the 0.1 Wm^{-1} higher TSI during the MCA has not resulted in significantly shorter winters, suggesting that the influence of TSI on winter lengths in NE Poland during the past millennium is minimal. This is in line with recent findings for the northern Hemisphere by Schurer et al. (2014).

In the Northern Hemisphere, extratropical volcanic eruptions lead often to positive NAO, and winter warming in the European sector is likely, but rather a probabilistic than a deterministic feature (Fischer et al., 2007; Shindell et al., 2004). Nevertheless, European winters are cold during periods with cumulative strong negative volcanic forcing (Hegerl et al., 2011; Shindell et al., 2003) like, for instance after the very large eruption AD 1257–58, which was followed by ones in 1269, 1278 and 1286 (Schneider et al., 2009). The AD 1258 eruption is thought to be one of the world’s largest volcanic eruption of the past millennium (Lavigne et al., 2013).

Clearest evidence of volcanic effects on the length of winter conditions in Poland is found in the context of the volcanic eruption at AD 1258, when TSI values were high but volcanic forcing strong. At this time, among the longest reconstructed DB4°C values are recorded between AD 1259-1263. Very cold winters struck Europe between AD 1259 and 1262 (Stothers, 2000). Remarkably, all the (sub) decadal scale periods with severe winters (DB4°C) in NE Poland coincided with periods of cumulative strong volcanic forcing like AD 1258 and the following years, around AD 1450, around AD 1600, throughout the 17th century and the first two decades of the 19th centuries and, taking varve counting uncertainties into account possibly also the events following AD 1060 and AD 1170. Overall, our findings for Poland agree with the conclusion by Hegerl et al. (2011) that winter climate in Europe prior to the 20th century is significantly influenced by external forcing, whereby cumulative negative volcanic forcing seems to be the pacemaker for very severe winters and the influence of TSI in the multidecadal and lower frequency band is minimal. Our findings from Poland do not support the conclusions by Bengtsson et al. (2006) who postulated that “climate variability in Europe for the ‘pre-industrial’ period 1500–1900 is fundamentally a consequence of internal fluctuations of the climate system.”. Interestingly, the relationship between external forcing and winter severity in Poland breaks down in the 20th century. We consider two possible explanations: (i) the influence of unforced variability has increased which was, for instance the case in the severe winters of the 1940s following the argument of Brönnimann et al. (2004), or (ii) the new combination of forcings including the anthropogenic GHG forcing in the 20th century leads to a new set of dynamic responses of the winter-time atmospheric circulation in this part of the world.

5. Conclusion

Our analyses of sediment trap chrysophyte assemblages from lakes across a climatic and ecological gradient in Northern Poland indicates that chrysophyte cysts can provide useful and reliable quantitative estimates of past changes in cold-season climate and lengths (severity) of winters. Ordination analyses demonstrated that relationships between chrysophyte assemblages and DB4°C are statistically significant and independent of morphological (area, depth) and chemical (conductivity, major ions, nutrients) conditions in the lake. This Polish chrysophyte cyst training set has enabled the development of a robust transfer function for reconstructing a cold-season climate-sensitive variable, i.e. the number of consecutive days with water temperatures below 4°C (DB4°C).

The application of the transfer function to an annually varved sediment core from Lake Żabińskie (NE Poland, Masurian lake land) yielded a reconstruction of DB4°C for the past 1000 years at a resolution of 5 years. Comparison of the ordination axes of the fossil cyst data, dry bulk density and the reconstruction indicated that land-use changes were important in the lake, especially between in the early 19th century, but that the reconstruction of DB4°C is independent of these anthropogenic changes.

The cold-season related DB4°C reconstruction from Lake Żabińskie shows pronounced decadal and multidecadal variability with a clear shift from warmer MCA to colder LIA conditions at AD 1430-1460. Changing winter climate conditions in Poland were likely controlled by shifts in the European atmospheric zonal flow (from westerly to easterly), controled by the NAO, and a higher influence of continental anticyclonic systems from the Siberian High during the LIA. Comparison with other European cold-season sensitive reconstructions shows a good agreement, especially during the prominent warming between AD 1740-1780 and the cooling around AD 1800-1820. On the other hand, also some discrepances are observed. These might be attributed to real regional differences between the climates in Poland and other areas in Central and Western Europe, to differences in what the various proxies indicate (length of the winter, maximum cold, maybe different phonological periods of winter), or uncertainties in the different proxy reconstructions themselves.

Striking correspondences between the DB4°C reconstruction from NE Poland and the combined solar and volcanic forcing during the past Millennium suggest that winter climate in NE Poland is strongly influenced by natural forced variability. Persistent strong volcanic forcing, particularly a series of strong events seems to produce very long winters at sub-decadal to decadal scales. The influence of TSI in the multidecadal and lower frequency band seems to be minimal. We do not find support for the idea that climate variability in Poland during the 'pre-industrial' period was fundamentally a consequence of internal atmosphere-ocean variability.

During the 20th century the situation is reverse and winter severity (DB4°C) is negatively correlated with the combined solar and volcanic forcing, suggesting that the combination of forcing factors in the 20th century leads to a different response of winter climate in NE Poland to (natural and anthropogenic) forcing factors, or unforced variability (e.g. ENSO and its influence on extratropical climate) plays a stronger role as it is suggested for the 1940s.

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Figure captions

Figure 1. Study area and geographical settings. A) Spatial distribution of winter temperatures (Dec-Feb) during the year of sediment trap exposure (Oct. 2011- Nov. 2012). Dots show the locations of the

lakes where sediments traps were deployed. Black dots show the lakes were thermistors were recovered. These lakes are included in this study. The yellow star indicates the location of Lake Żabińskie, used in the reconstruction. B) Catchment topography of the Lake Żabińskie area and bathymetry.

Figure 2. Multivariate analyses. A) PCA of the environmental data. Grey dots correspond to the sites. B) RDA of the dataset with all environmental variables. Sites are displayed as grey dots, species as white dots.

Figure 3. Relationship between observed and predicted DB4°C. A) Observed versus estimated DB4°C, and B) observed versus residuals based on PLS-2 regression.

Figure 4. Chrysophyte cyst stratigraphy. Plot showing the 21 most abundant species in the sediments of Lake Żabińskie, the similarity index between the training set and fossil assemblages, and the downcore reconstruction of DB4°C (black line 5-yr. filter).

Figure 5. Age-depth model, downcore fossil ordination and dry bulk density. Varve chronology for the period AD 1000-2010. A) Chrysophyte based-DB4°C reconstruction (grey line, 5-yr filter; black line 21-yr. filter). Scores of the two main axis of the ordination (PCA) of the fossil dataset: B) PCA 2 scores (red line, 5 year filter; black line 21 yr. filter); C) PCA 1 scores (green line, 5-yrfilter; black line 21-yr. filter). D) Dry bulk density (g/cm^3).

Figure 6. A) Polish Winter Thermal Index (Przybylak et al., 2005) for the period AD 1780-1999 (yellow line) and Winter Thermal Index for AD 1400-1800 (orange line) (Przybylak, 2011). B) Baltic ice winter severity index (Koslowski and Glaser, 1999; Schmelzer and Holfort, 2011). C) Winter (Dec-Feb) temperature in the Netherlands (van Engelen et al., 2001). D) Dec-Feb Temperature multiproxy reconstructions for Europe (Luterbacher et al., 2004). E) Chrysophyte based Oct-May temperature reconstruction from Lake Silvaplana (de Jong et al., 2013b). F) Chrysophyte based-DB4°C reconstruction (grey line, 5-yr filter; black line 21-yr. filter). G) Reconstruction of Siberian High air pressure (Meeker and Mayewski, 2002). H) Reconstruction of the NAO (Trouet et al., 2009).

Figure 7. DB4°C variations (this study) in response to forcings. A) combined solar and volcanic forcing (green, 11 years smoothed) (Crowley, 2000), compared to the chrysophyte based-DB4°C reconstruction (grey line, 5-yr filter; black line 21-yr. filter), B) volcanic forcing (blue), C) Total Solar Irradiance (orange), and D) GHG (red).

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