

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

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

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

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

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45 *1. Introduction*

46 Comprehensive reconstructions of continental to global-scale temperature variability during  
47 the past millennia have demonstrated the value of paleoclimate proxy data and provided insight into  
48 natural forced and unforced variability, and anthropogenic disturbances (PAGES 2k Consortium,  
49 2013). This information is a key to reduce uncertainties for future climate change (Hegerl et al., 2006).

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

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

65 In this article we present a quantitative reconstruction of cold season climate from varved lake  
66 sediments in NE Poland for the past 1000 years. Information about winter temperature from this area  
67 is very important because of the very strong relationship between Polish and European temperatures  
68 both at the interannual and interdecadal time scales (Luterbacher et al., 2010). Moreover, winter  
69 climate in NE Poland is affected by major atmospheric circulation patterns and modes in the north  
70 Atlantic European domain, especially the North Atlantic Oscillation (NAO) and the East  
71 Atlantic/Western Russia (EA/WRUS) pattern (Luterbacher et al., 2010; Luterbacher et al., 2004). This  
72 is clearly evidenced in snow cover duration (Falarz, 2007), lake ice-cover (Marszelewski and Skowron,  
73 2006; Wrzesiński et al., 2013), documentary and early instrumental data (Luterbacher et al., 2010;  
74 Przybylak et al., 2005; Przybylak et al., 2003) for the late 19<sup>th</sup> and 20<sup>th</sup> century.

75 Poland is also interesting because of its wealth of long high-quality instrumental records back  
76 to AD 1780. For periods prior to instrumental observations, there exist a few successful attempts at  
77 reconstructing air temperature by means of documentary sources, chronicles and meteorological  
78 observations (Przybylak et al., 2005; 2010; Sadowski, 1991), or quantitative reconstructions of air  
79 temperature based on historical sources with high resolution (daily). Such data are available for  
80 shorter discrete periods of several dozens of years (e.g. Bokwa et al., 2001; Limanówka, 2000;  
81 Michalczewski, 1981; Przybylak and Marciniak, 2010; Przybylak et al., 2014). The state of knowledge  
82 of Polish climate before the 16<sup>th</sup> century is generally modest and the existing reconstructions are  
83 uncertain (Przybylak, 2011; Przybylak et al., 2010). Natural paleoclimate records with high (interannual  
84 to subdecadal) resolution that reach further back than AD 1550 are extremely scarce, and only based  
85 on tree-rings widths (Koprowski et al., 2012; Przybylak et al., 2005; Szychowska-Krapiec, 2010). A  
86 synthesis of the longest tree-ring based climate reconstructions for Poland is presented by Zielski et  
87 al. (2010). Most surprisingly, studies using varved lake sediments are missing despite their  
88 demonstrated potential for quantitative seasonal paleoclimate studies (Amann et al., 2014; Bonk et al.,  
89 2015; Larocque-Tobler et al., 2015). Recent work has also highlighted the potential of raised bogs as  
90 paleoenvironmental and paleoclimate archives in Poland (De Vleeschouwer et al., 2009; Gałka et al.,  
91 2014; Lamentowicz et al., 2015; 2009).

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

103 Quantitative methods relating chrysophyte assemblages and air/water temperature during  
104 winter have been developed on high-altitude sites in the Alps, the Pyrenees and the Andes (de Jong

105 and Kamenik, 2011; de Jong et al., 2013b; de Jong et al., submitted for publication; Kamenik and  
106 Schmidt, 2005; Pla and Catalan, 2005). In the Alps, chrysophyte inference models for high altitude  
107 lakes have so far been developed for relatively small geographic regions. Kamenik and Schmidt  
108 (2005) developed a chrysophyte-based temperature transfer function in the Austrian Alps. Results in  
109 their paper showed high correlations between cold-season temperatures (October-May) and  
110 chrysophyte assemblages in modern samples of 29 high altitude lakes. The resulting cold-season  
111 temperature transfer function based on chrysophytes was later applied to varved sediments from Lake  
112 Silvaplana in the Swiss Alps by de Jong et al. (2013a; 2013b). Reconstructions show similar results  
113 when compared with other historical records from the Alps. Moreover the authors found high  
114 similarities to NAO and 'Siberian high pressure' indices for the last millennium. A similar approach was  
115 used by Pla and Catalan (2005) in high-altitude lakes from the Pyrenees. In that study, modern  
116 chrysophyte assemblages were highly correlated to air temperature anomalies along an altitudinal  
117 gradient among the studied lakes. Using a similar methodology as proposed in the present study, de  
118 Jong et al. (submitted for publication) developed a chrysophyte-based inference model to reconstruct  
119 the number of consecutive days with temperatures below 4°C (length of winter) from a training set of  
120 lakes from the south-central Andes which were sampled along an altitude gradient substituting  
121 variations in austral winter temperature.

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

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

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## 147 *2. Study area, material and methods*

### 148 *2.1. Training set data collection*

149 The 50 training set lakes (Fig. 1A) were chosen among more than 2900 low-land lakes of  
150 northern Poland (Jańczak et al., 1996), covering a broad range of morphological, physical, and  
151 chemical parameters (Hernández-Almeida et al., 2014). Sediment traps equipped with thermistors  
152 were deployed in the selected lakes. Field surveys were conducted during October 2011 and  
153 November 2012. After one year of exposure, only 37 sediment traps and thermistors were found. The  
154 distribution of these lakes is along a longitudinal gradient that is related to an E-W winter temperature  
155 gradient of ~4°C during the year of exposure between October 2011 and November 2012 (Fig. 1A).  
156 The spatial extension of the study area is large (52°31'- 54°19' N, 14°37'- 22°53'E), covering more than  
157 700 km from the westernmost to the easternmost lake. Training set lakes are generally small (<20 ha),  
158 below 260 m.a.s.l., between 6 and 44 m deep and slightly alkaline (pH ranging from 6.5 to 9), with  
159 moderate agricultural activity and/or forestry in the catchments. Additional limnological and geographic  
160 characteristics are summarised in Table 1.

161 Sediment traps consisted of PVC-liners with a length of 80 cm and a diameter of 9 cm, 2 tubes  
162 per trap (Bloesch and Burns, 1980), with the openings of the traps approximately 1.5 m above the lake  
163 bottom. Collection of environmental data in each lake was made every three months. Water chemistry

164 (conductivity, oxygen, pH) and turbidity were measured at 2 m water depth. Water samples were  
165 collected at 1 m water depth from each training set lake. Water chemistry analyses were performed at  
166 the University of Gdansk for major ion concentration analysis ( $\text{Ca}^{2+}$ ,  $\text{Mg}^{2+}$ ,  $\text{Na}^+$ ,  $\text{K}^+$ , sulfates, fluorides,  
167 chlorides) as well for total P and N. Analyses were performed using ion chromatography (ICS 1100,  
168 Dionex, USA) equipped with an IonPack AS22 column for anions and an IonPack CS16 column for  
169 cations, and colorimetric method and a Spectroquant NOVA 400 spectrophotometer (Merck),  
170 respectively. Monthly air temperatures for the studied lakes were obtained from the Institute of  
171 Meteorology and Water Management-National Research Institute of Poland. Water temperature in the  
172 epilimnion (2-m depth) was recorded by HOBO U22-001 thermistors at 15-minutes intervals during  
173 sediment trap exposure until trap recovery in November 2012.

174 It is difficult to determine freeze and ice break-up dates with thermistors due to the complexity  
175 of the freezing and thawing process (Gabathuler, 1999; Kamenik and Schmidt, 2005). Ice-cover  
176 formation on a specific lake does not only depend on the air temperature decrease, but also on lake  
177 size, morphology, inflow to the lake and chemical properties of the water. Moreover, a thick ice-cover  
178 insulates the lake from further cooling during exceptionally cold periods. For these reasons we have  
179 chosen as a target climate variable the 'number of consecutive days below 4°C' (DB4°C). This is a  
180 more objective parameter to determine a period of cold conditions in the lake (i.e. length of the winter)  
181 and serves as an indicator of winter severity. Fresh water reaches its maximum density at 4°C and  
182 further cooling reduces convective overturning. DB4°C was calculated from individual thermistor logs  
183 for each lake. In the year of observation (2011/2012) temperatures in the upper water column  
184 exhibited a strong linear correlation with winter air temperature (Dec-Feb) during the study period ( $R=$   
185 0.71). This is attributed to the relatively short ice-cover duration during the year of the training set  
186 experiment: epilimnetic waters were not insulated from air temperatures. In years with very long ice-  
187 cover and very cold winters, the water temperature in the epilimnion is often uncoupled from air  
188 temperature due to the insulating effect of thick ice (Livingstone and Lotter, 1998).

189

## 190 2.2. Reconstruction site

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

194 (41.6 ha) with the maximum depth in the central part of the lake (44.4 m, Fig. 1B). A major outflow is  
195 located in the south-western part, connecting Lake Żabińskie with much larger Lake Gołdopiwo. Two  
196 smaller inflows enter the lake from its southern side and a larger inflow in the northeast comes from  
197 Lake Purwin. Lake Żabińskie is eutrophic, hardwater, dimictic (mixes in early spring and fall) and  
198 anoxic during most of the year. Precipitation and temperature are higher during the summer months  
199 (up to 90 mm and 17°C, respectively) and lower during winter. According to in-situ observations made  
200 during the field surveys, the lake was ice-covered from mid-January 2012 to March 2012.

201         The climate in the Masurian Lakeland is characterized by strong continentality with cold  
202 winters and warm summers. The mean annual air temperature (measured between AD 1971 and  
203 2005) is 6.5°C with the coldest temperatures occurring in January and the warmest temperatures in  
204 July and August. The mean annual precipitation ranges from 550 to 600 mm (Lorenc, 2005). The  
205 region is covered by snow for 1.5 to 3 months in winter.

206         The village of Żabinka, established in AD 1713, is located 0.5 km from the southern shore of  
207 the lake and had about 250 inhabitants in the 19<sup>th</sup> century and 200 in AD 1939. Between AD 1910 and  
208 1920, a restaurant and a recreation place (campsite and beach) were built on the northern shore of  
209 Lake Żabińskie. These infrastructures were used for military purposes during the Second World War,  
210 and recovered for recreation activities during the mid-1950s. New infrastructures were built in the early  
211 1970s and during 1980s due to the increase of activities related to tourism.

212

### 213         2.3. Sediment core and chronology

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

221         The regular succession of two types of laminae along the sediment core is interpreted to  
222 reflect seasonal variations in the sedimentation environment (Zolitschka, 2007): the spring/summer  
223 light layer is mainly composed of autochthonous calcite and diatoms; the dark fall/winter layers consist



224 mainly of decomposed organic matter and clay minerals. The laminae couplets are interpreted as  
225 biochemical varves (Tylmann et al., 2013). Although the varve structure was recognized to be more  
226 complex and included multiple calcite layers within a one-year-deposition, it has a remarkable potential  
227 for a reliable varve chronology and for multiproxy paleoenvironmental reconstructions (Bonk et al.,  
228 2015).

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

240

#### 241 2.4. Sample preparation and cyst analyses

242 Sediment trap and sediment core samples (0.2 g wet sediment) were treated with H<sub>2</sub>O<sub>2</sub> and  
243 HCl for chrysophyte cysts analyses, following the standard diatom procedure (Battarbee et al., 2001).  
244 Samples were then washed with distilled water and filtered with Milipore nylon-filters (63 µm) to  
245 remove large particles. Chrysophyte cysts were analysed using a scanning electron microscope (Carl-  
246 Zeiss EVO40). For the training set samples, a minimum of 300 modern cysts were counted per  
247 sample. For the downcore fossil cyst assemblages, 80 chrysophytes per sample were counted for  
248 samples between AD 2010 and 1898, and 50 cysts per sample until AD 1000. The final reconstruction  
249 was based on 5-yr running means. Identification of cysts followed Duff et al. (1995), Wilkinson et al.  
250 (2002) (labelled as PEARL -Paleoecological Environmental Assessment and Research Laboratory-),  
251 and Huber et al. (2009) (labelled as L). The unpublished new cysts were assigned a new number  
252 using the code 'ZAB' for the fossil and modern samples of Lake Żabińskie, and the code 'TSP'  
253 (Training Set Poland) for the modern samples.

254

## 255 2.5. Statistical analyses

256 Multivariate statistical analyses were used to identify major environmental gradients and to  
257 explore relationships between chrysophyte cysts and environmental variables. Prior to statistical  
258 analyses, environmental variables were explored for normal distribution. Water chemistry variables  
259 (except pH) were log transformed. For the climate related variables, only DB4°C and the seasonal  
260 temperatures showed normal distribution and were not transformed. Chrysophyte data were  
261 expressed as relative abundances. All species present in at least 2 training set lakes with abundance  
262 >1% were retained in the numerical analyses. Chrysophyte abundances were square-root transformed  
263 in order to stabilise variances.

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

281

## 282 2.6. Transfer functions

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

295 Analogue quality was used to diagnose the quality of the reconstruction by evaluating how well  
296 the calibration set of modern samples provided analogues for the fossil core samples (Birks et al.,  
297 1990). The taxonomic distance between each fossil sample and the calibration set was calculated  
298 using a squared chord distance (SCD) as a dissimilarity measure (Overpeck et al., 1985). Fossil  
299 samples with dissimilarity larger than the 10<sup>th</sup> percentile lacked good modern analogues (Birks et al.,  
300 1990).

301

## 302 3. Results

### 303 3.1. Calibration data set and ordination

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

312 Of the 129 taxa encountered in the 37-lake training set, 63 fulfilled the criteria to be included in  
313 the numerical analyses. The gradient length of the first axes was 1.7 standard deviation (SD) units,  
314 indicating that numerical methods based on a linear response model are most appropriate (Birks,  
315 1995). RDA was performed on the biological and environmental dataset (Fig. 2B). The PCA (Fig. 2A)  
316 and high VIFs (>20) revealed that there is strong collinearity among several of the environmental  
317 variables. Of all the environmental variables considered (Table 1), the *ordistep* procedure selected  
318 only six variables for entrance into the final model. (conductivity, K, chlorides, sulfates, DB4°C and  
319 longitude), which explained the variation in the chrysophyte data (28%) almost as well as the complete  
320 set of all environmental variables included in the ordination analyses.

321 A series of partial RDAs run with one explanatory variable at a time indicates that DB4°C and  
322 conductivity are the strongest variables and capture 4.5% and 6%, respectively. This low explained  
323 variance is typical of noisy, heterogeneous datasets which contain many zero values (Jones and  
324 Juggins, 1995). As the data set covers a long geographical transect, some species show scattered  
325 occurrences, limited to a few sites, or centred at a particular region of the training set. Similar results  
326 with low explained variance (<3%), but yet ecologically informative, were also reported in diatom data  
327 sets, for reconstructing salinity in Australia (Saunders 2011). Variance partitioning demonstrates that  
328 relationships between chrysophytes and DB4°C are independent of variables related to water  
329 chemistry (conductivity, chlorides, K, sulfates). The ratios between the first (constrained) and the  
330 second (unconstrained) eigenvalue ( $\lambda_1/\lambda_2$ ) are calculated to assess which variables could be used for  
331 a transfer function development (Birks, 1995). Although DB4°C  $\lambda_1/\lambda_2$  is the highest of all variables  
332 (0.71) and the one of climate interest, values lower than 1 indicate that the training set is affected by  
333 other variables that have an influence on the chrysophyte cyst composition but were not measured.

334

### 335 3.2. Transfer function

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

342 provided the minimal adequate model with highest  $R^2$  (cross-validated) and lowest mean and  
343 maximum bias.

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

350

### 351 3.3. Ordination of fossil cyst assemblages

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

362

### 363 3.4. Downcore reconstruction of DB4°C

364 Using the PLS-2 model, the chrysophyte-DB4°C transfer function was applied to the fossil  
365 chrysophyte assemblages in the sediments of Lake Żabińskie (Figs. 4 and 5). The sediment depths  
366 were converted into calendar ages according to the varve chronology shown in Figure 5.  
367 Reconstructed DB4°C (5-year running mean) fluctuated between 94 and 106 days. The reconstruction  
368 indicated highest DB4°C (> 104) at the beginning of the record (AD 1000-1070). After this date, there  
369 is a decreasing trend toward lower values of DB4°C until AD 1245 (94 DB4°C). Subsequently a rapid  
370 increase toward high DB4°C values (severe winters) is observed for a short period. Afterwards,

371 reconstructed values return to lower DB4°C values until a minimum of 94 DB4°C is reached at AD  
372 1433, followed by an abrupt increase up to 106 DB4°C. The period from AD 1180–1440 coincides with  
373 the warm phase of the Medieval Climate Anomaly (MCA). Higher values of reconstructed DB4°C (long  
374 winters) are recorded until AD 1730. A decrease in DB4°C values is observed between AD 1750 and  
375 1850 (ranging 94-97 DB4°C) interrupted by a short colder period around AD 1800-1820. After AD  
376 1850, values of DB4°C increase until a maximum of 106 DB4°C around AD 1945. From then onwards  
377 there is a steady decreasing trend of DB4°C values (shorter winters).

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

#### 387 *4. Discussion*

##### 388 *4.1. Proxy interpretation and model performance*

389         The final chrysophyte-based transfer function was based on the proxy DB4°C derived from the  
390 thermistor data. Many empirical studies have shown water temperatures to be strongly related to  
391 ambient air temperature in absence of ice-cover (Livingstone and Lotter, 1998; Shuter et al., 1983).  
392 DB4°C was chosen because it showed the highest  $\lambda_1/\lambda_2$  ratio and explains a higher proportion of  
393 variance independently than other variables included in the training set; and it is of climate interest.  
394 Modern data show that there is no strong correlation between DB4°C and water chemistry (Fig. 2A)  
395 and a proportion of the variance of the modern chrysophyte cyst assemblage is explained by DB4°C  
396 independently of the other most important variable (conductivity) (Fig. 2B). Pla and Catalan (2005) and  
397 Kamenik and Schmidt (2005) found a similar orthogonal positioning of the cold-season related  
398 variables (altitude and spring mixing) with respect to variables related to water chemistry, suggesting

399 that a portion of the variation in their cyst assemblages was independent from the chemical  
400 composition of the lake waters.

401 A  $\lambda_1/\lambda_2$  lower than 1 may indicate that there is another secondary gradient, or combination of  
402 them, slightly larger than the first constrained axes that reflects variation in the chrysophyte  
403 assemblages that is unrelated to the DB4°C. Adding more lakes to expand the DB4°C gradient of our  
404 data set would help to overcome these problems. The trend in the residual structure of the PLS-2  
405 model indicates overestimation at sites with low DB4°C values and underestimation at sites with high  
406 DB4°C values. In consequence, the predictive ability of the transfer function is greater for sites with  
407 medium values of DB4°C than for sites falling at the extremes of the environmental gradient. This  
408 'edge effect', however, is inherent to PLS and its unimodal counterpart WA-PLS, both of which utilize  
409 an inverse deshrinking regression (Birks, 1995; ter Braak and Juggins, 1993).

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

419

#### 420 4.2. Relationship between DB4°C and cold-season temperatures

421 Studies converting non-climate variables (cyst assemblages) to climate variables imply that  
422 the relationships between both have remained constant through time. Correlation between the  
423 chrysophyte-based DB4°C reconstruction and the homogenized air temperature series for Lake  
424 Żabińskie (Larocque-Tobler et al., 2015) is highest for January-March temperatures (3-years filter; R=  
425 -0.35,  $p>0.05$ ). Although a better correlation would be expected, this is explained because of distorting  
426 effects of ice-cover. In such periods, the ice-cover insulates the bulk of the water body from further  
427 cooling, and water temperatures are uncoupled from air temperatures (Livingstone, 1993). As the  
428 relationship between DB4°C and air temperatures is not linear but distorted by the presence of ice-

429 cover, and to avoid accumulation of errors during the conversion to winter temperatures, we focus on  
430 the performance of chrysophyte cysts as proxies of changes in DB4°C. As we have shown in the  
431 modern training set data and instrumental data during the 20<sup>th</sup> century, DB4°C has a strong  
432 relationship with winter length.

433

#### 434 4.3. Quality of the reconstruction

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

440 An ordination was carried out on the entire cyst assemblage dataset (Fig. 5). The two first  
441 axes explain 11% and 8.5%, both are significant. This is typical of noisy data sets with a large number  
442 of taxa and many zero values to have a relatively low percentage of variance explained (Bennion,  
443 1994). Comparison between scores of PCA axes with the reconstruction helps to evaluate how  
444 representative the reconstruction is compared to the major ecological changes of the fossil  
445 assemblage as shown in the ordination. Moreover, if anthropogenic effects are first factored out, the  
446 effects of natural variability are likely to be more clearly detected. Reconstructed DB4°C and PCA1 are  
447 very different. PCA1 shows a very abrupt change at AD 1800 (Fig. 5C). This shift in the scores might  
448 reflect an anthropogenic change. Similarity between PCA1 scores and dry bulk density of the  
449 sediments denotes that chrysophyte ecological shifts downcore may be driven by environment  
450 changes in the lake. The shift around AD 1790 is correlated with a large increase in the dry bulk  
451 density of the sediments, reflecting that a large amount of fine material was deposited in the Lake  
452 Żabińskie and may have potentially altered the chrysophyte-environmental variable relationship (Fig.  
453 5D). According to the percentage pollen data from Lake Żabińskie, abrupt forest decline and increase  
454 in ruderal and grassland taxa is observed after AD 1800, which is interpreted to be related to human  
455 impact on vegetation and forest clearance in the area (Wacnik, pers. comm. 2014). We argue that  
456 PCA1 scores reflect the extent to which the fossil cyst community of Lake Żabińskie responded to  
457 anthropogenic stressors, and therefore, ecological changes of the fossil chrysophyte assemblage  
458 shown in PCA1 are not related to climate factors. In contrast, downcore variability of scores of PCA2



459 do not show this abrupt shift at AD 1800, and they are highly correlated to the downcore  
460 reconstruction of DB4°C, indicating that this variable was important as well for the fossil assemblage,  
461 and is not related to anthropogenic changes (Fig. 5A-B).

462

#### 463 4.4. Regional comparison of temperature variability

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

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

487 Although similarities with other winter-season climate records across Europe are evident,  
488 differences exist at some intervals. Following AD 1258, an abrupt cooling for 5-7 years is observed in

489 our DB4°C reconstruction. This is not seen in the Low Countries temperature record (van Engelen et  
490 al., 2001). But this cold spell has been found in the winter temperature reconstruction from Lake  
491 Silvaplana whereby data gaps in the Silvaplana record are interpreted as very cold spells (de Jong et  
492 al., 2013b).

493         Interesting are the differences in the 16<sup>th</sup> century, i.e. at the beginning of the Spörer Minimum:  
494 while the Baltic ice winter severity index (Fig. 6B), the indices for the Low Countries (Fig. 6C) and  
495 other documentary winter climate records from Western Europe show a sequence of very severe  
496 winters in the 1430s followed by moderately cool winters after AD 1450 (Lamb, 2002), records from  
497 Poland show a different pattern: our DB4°C record from NE Poland, winter climate indices for Poland  
498 (Przybylak, 2011; Przybylak et al., 2005) and January-April air temperature for central Poland  
499 reconstructed from tree-ring widths of Scots pine (Przybylak, 2011) show 'normal' or even moderately  
500 warm winters in the 1430s. These were followed by strong cooling in the 1440s and peak cold around  
501 AD 1460. Cold conditions persisted throughout the rest of the 16<sup>th</sup> century. On the other hand, the very  
502 cold conditions in the second half of the 15<sup>th</sup> century did not occur in SE Poland (Szychowska-Krąpiec,  
503 2010). This regional discrepancy during the beginning of the Spörer Minimum and the 15<sup>th</sup> century is  
504 reproduced in independent data sets and seems thus to be real. In Poland, the Maunder Minimum (AD  
505 1645-1715) (Eddy, 1976) does not strike as the very cold phase reported in other European records  
506 (Glaser and Riemann, 2009; Luterbacher et al., 2004). Instead, a brief cooling phase can be  
507 appreciated between AD 1689–1699, that matches with a 11 years cool phase described in the Tatra  
508 Mountains, Poland (Niedźwiedź, 2010).

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

515

#### 516         4.5. Factors controlling winter severity

517         The agreement between short and long winters in Poland (DB4°C) with other European cold-  
518 season temperature records suggests a common large-scale atmospheric circulation mechanism. A

519 first candidate is the North Atlantic Oscillation NAO which represents one of the most prominent  
520 modes of inter-monthly to inter-decadal variability in the Northern Hemisphere and is known to affect  
521 winter climate in Europe north of the Alps (e.g. Hurrell, 1995). Przybylak et al. (2005; 2003) correlated  
522 phases of positive NAO with strong zonal circulation (westerly flow) and increased winter air  
523 temperatures over central Europe and Poland. According to Luterbacher et al. (2010) the NAO index  
524 accounts for more than 50% of the winter temperature variations in Poland.

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

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

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

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

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

#### 561 4.6. Influence of volcanic and solar forcing on winter climate in Poland

562 The question about the causes of the rapid decadal-scale cooling events in Europe during the  
563 last Millennium and during the LIA in particular have been discussed controversially (Hegerl et al.,  
564 2011 and references therein). While anthropogenic influences can be excluded before the 20<sup>th</sup>  
565 century, variability in Total Solar Irradiance (TSI), volcanic aerosols (both external forcings; Crowley,  
566 2000; Fig. 7C-D) or unforced internal (atmosphere-ocean) variability remain potential candidates to  
567 influence seasonal European temperatures and climate. Fig. 7 shows that the severity of winters in NE  
568 Poland (DB4°C) in the pre-industrial period (prior to the 20<sup>th</sup> century) is largely correlated to the  
569 combined solar and volcanic radiative forcing (data from Crowley, 2000) while this relationship is  
570 inverted in the 20<sup>th</sup> century.

571 TSI is generally about  $0.1 \text{ Wm}^{-1}$  higher during the Medieval Climate Anomaly compared with  
572 the LIA and four multi-decadal periods of minimum solar activity are observed: the Wolf (AD 1280-  
573 1350), Spörer (AD 1420-1540), Maunder (AD 1645-1715) and Dalton Minima (AD 1795–1830) (Eddy  
574 and Oeschger, 1993). However, since solar minima coincided with strong negative volcanic forcing in  
575 the past millennium, it is difficult to discriminate the influence of both effects on regional surface  
576 temperatures. The reconstructed DB4°C does show severe winters during the Dalton Minimum but the  
577 other solar minima do not stand out as cold periods, while this is rather the case for the temperature

578 indices from the Low Countries (Fig. 6C). The high DB4°C values (severe winters) at AD 1450-1470  
579 might be partly related to the Spörer Minimum (in addition to a response to volcanic forcing after AD  
580 1450). The severe winters between AD 1795-1820 might reflect the Dalton Minimum, combined with  
581 the effect of the large volcanic eruptions at AD 1809 and 1815-1816 (D'Arrigo et al., 2009). Fig. 7A  
582 shows a systematic offset in the match between the combined S+V forcings and the severity of winters  
583 (DB4°C) between the MCA and the LIA. It seems that the  $0.1 \text{ Wm}^{-1}$  higher TSI during the MCA has not  
584 resulted in significantly shorter winters, suggesting that the influence of TSI on winter lengths in NE  
585 Poland during the past millennium is minimal. This is in line with recent findings for the northern  
586 Hemisphere by Schurer et al. (2014).

587 In the Northern Hemisphere, extratropical volcanic eruptions lead often to positive NAO, and  
588 winter warming in the European sector is likely, but rather a probabilistic than a deterministic feature  
589 (Fischer et al., 2007; Shindell et al., 2004). Nevertheless, European winters are cold during periods  
590 with cumulative strong negative volcanic forcing (Hegerl et al., 2011; Shindell et al., 2003) like, for  
591 instance after the very large eruption AD 1257–58, which was followed by ones in 1269, 1278 and  
592 1286 (Schneider et al., 2009). The AD 1258 eruption is thought to be one of the world's largest  
593 volcanic eruption of the past millennium (Lavigne et al., 2013).

594 Clearest evidence of volcanic effects on the length of winter conditions in Poland is found in  
595 the context of the volcanic eruption at AD 1258, when TSI values were high but volcanic forcing  
596 strong. At this time, among the longest reconstructed DB4°C values are recorded between AD 1259-  
597 1263. Very cold winters struck Europe between AD 1259 and 1262 (Stothers, 2000). Remarkably, all  
598 the (sub) decadal scale periods with severe winters (DB4°C) in NE Poland coincided with periods of  
599 cumulative strong volcanic forcing like AD 1258 and the following years, around AD 1450, around AD  
600 1600, throughout the 17<sup>th</sup> century and the first two decades of the 19<sup>th</sup> centuries and, taking varve  
601 counting uncertainties into account possibly also the events following AD 1060 and AD 1170. Overall,  
602 our findings for Poland agree with the conclusion by Hegerl et al. (2011) that winter climate in Europe  
603 prior to the 20<sup>th</sup> century is significantly influenced by external forcing, whereby cumulative negative  
604 volcanic forcing seems to be the pacemaker for very severe winters and the influence of TSI in the  
605 multidecadal and lower frequency band is minimal. Our findings from Poland do not support the  
606 conclusions by Bengtsson et al. (2006) who postulated that "climate variability in Europe for the 'pre-  
607 industrial' period 1500–1900 is fundamentally a consequence of internal fluctuations of the climate

608 system.". Interestingly, the relationship between external forcing and winter severity in Poland breaks  
609 down in the 20<sup>th</sup> century. We consider two possible explanations: (i) the influence of unforced  
610 variability has increased which was, for instance the case in the severe winters of the 1940s following  
611 the argument of Brönnimann et al. (2004), or (ii) the new combination of forcings including the  
612 anthropogenic GHG forcing in the 20<sup>th</sup> century leads to a new set of dynamic responses of the winter-  
613 time atmospheric circulation in this part of the world.

614

## 615 *5. Conclusion*

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

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

630 The cold-season related DB4°C reconstruction from Lake Żabińskie shows pronounced  
631 decadal and multidecadal variability with a clear shift from warmer MCA to colder LIA conditions at AD  
632 1430-1460. Changing winter climate conditions in Poland were likely controlled by shifts in the  
633 European atmospheric zonal flow (from westerly to easterly), controlled by the NAO, and a higher  
634 influence of continental anticyclonic systems from the Siberian High during the LIA. Comparison with  
635 other European cold-season sensitive reconstructions shows a good agreement, especially during the  
636 prominent warming between AD 1740-1780 and the cooling around AD 1800-1820. On the other  
637 hand, also some discrepancies are observed. These might be attributed to real regional differences

638 between the climates in Poland and other areas in Central and Western Europe, to differences in what  
639 the various proxies indicate (length of the winter, maximum cold, maybe different phenological periods  
640 of winter), or uncertainties in the different proxy reconstructions themselves.

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

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

653

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662

#### 663 *Figure captions*

664

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

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

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

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

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

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

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

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

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697           **References**

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