High-resolution late-glacial chronology for the Gerzensee lake record (Switzerland): δ18O correlation between a Gerzensee-stack and NGRIP

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Oxygen-isotope variations were analyzed on bulk samples of shallow-water lake marl from Gerzensee, Switzerland, in order to evaluate major and minor climatic oscillations during the late-glacial. To highlight the overall signature of the Gerzensee δ18O record, δ18O records of four parallel sediment cores were first correlated by synchronizing major isotope shifts and pollen abundances. Then the records were stacked with a weighting depending on the differing sampling resolution. To develop a precise chronology, the δ18O-stack was then correlated with the NGRIP δ18O record applying a Monte Carlo simulation, relying on the assumption that the shifts in δ18O were climate-driven and synchronous in both archives. The established chronology on the GICC05 time scale is the basis for (1) comparing the δ18O changes recorded in Gerzensee with observed climatic and environmental fluctuations over the whole North Atlantic region, and (2) comparing sedimentological and biological changes during the rapid warming with smaller climatic variations during the Bølling/Allerød period. The δ18O record of Gerzensee is characterized by two major isotope shifts at the onset and at the termination of the Bølling/Allerød warm period, as well as four intervening negative shifts labeled GI-1e2, d, c2, and b, which show a shift of one third to one fourth of the major δ18O shifts at the beginning and end of the Bølling/Allerød. Despite some inconsistency in terminology, these oscillations can be observed in various climatic proxies over wide regions in the North Atlantic region, especially in reconstructed colder temperatures, and they seem to be caused by hemispheric climatic variations.

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1. Introduction

The transition from the last glacial to the present interglacial (Termination 1, –15–10 kyr BP) is characterized by several major and minor climatic oscillations which had strong impacts on marine and terrestrial environments (Karpuz and Jansen, 1992; Ammann et al., 1994; Lowe et al., 1994; Hughen et al., 1996; Brauer et al., 2000; Litt et al., 2003). In lacustrine settings, stable isotope analysis of carbonates is one of the best methods to identify and reconstruct those terrestrial climatic and environmental changes. The oxygen-isotope composition of inorganic lake carbonate is mainly influenced by (1) the δ18O of precipitation in the watershed, (2) the water temperature, (3) the hydrologic balance of the lake, and (4) biological activity within the lake (Leng and Marshall, 2004; Bernasconi and McKenzie, 2007). Whatever the exact causes affecting the isotopic composition of lake carbonates, major and minor fluctuations in δ18O are observed during the late-glacial in various lacustrine archives of the Northern Hemisphere (e.g. von Grafenstein et al., 1999; Yu and Wright, 2001) and seem to dominantly reflect variations in summer temperatures.

The three most prominent and therefore most extensively studied oxygen isotope shifts are the abrupt increase from low δ18O during the Oldest Dryas (GS-2) to less negative δ18O values during the Bølling/Allerød (B/A) interstadial (GI-1), followed again by low δ18O during the Younger Dryas (GS-1) and a sharp increase at the onset of the Holocene (Riezebos and Slotboom, 1984; Siegenthaler et al., 1984; Eicher, 1987; Lotter et al., 1992; Schwander et al., 2000; Rasmussen et al., 2006; Andrč et al., 2009). The strong similarity between δ18O and several other climate proxies in terrestrial, marine, and ice-core records suggests that these shifts are climate-controlled and thus synchronous over wide regions (Hughen et al., 1996; Benson et al., 1997; Hoek and Bohncke, 2001). This has already been suggested by pioneering work of Siegenthaler et al. (1984), showing a very similar pattern in the oxygen-isotope record of lake marls from Lake Gerzensee, Switzerland and in the Dye 3 Greenland ice-core (Siegenthaler et al., 1984).
Since the development of analytical techniques enables additional and more highly resolved data sets, several minor fluctuations in \(\delta^{18}O\) could be observed. The present study focuses on the B/A \(\delta^{18}O\) record with its internal fluctuations. Three to four decadal to centennial climatic oscillations during this period were recognized in the Greenland ice-core (Björck et al., 1998; Brauer et al., 2000; Lowe et al., 2008) and in marine records (Karpuz and Jansen, 1992; Asioli et al., 1999; Fletcher et al., 2010) of the North Atlantic region. In terrestrial records, these oscillations have been observed in Europe, e.g. in various lakes in Switzerland (Ammann, 2000; Lotter et al., 2000; von Grafenstein et al., 2000), southern Germany (von Grafenstein et al., 1999), Great Britain (Brooks and Birks, 2000a; Jones et al., 2002; Marshall et al., 2002; Lang et al., 2010), and Scandinavia (Paus, 1988; Paus, 1989; Brooks and Birks, 2000b). Several records in North America also reflect 3 to 4 short-lived environmental shifts during the B/A, mainly due to colder temperatures (Whittington et al., 1996; Nolan et al., 1999; Yu and Wright, 2001; Marshall et al., 2002; Yu, 2007).

It is remarkable that the cold phases of different duration and amplitude are all well reflected in terrestrial response over wide regions. This leads to the basic assumption that the warming and cooling during the late-glacial interstadial were hemispheric and that their recording in the oxygen isotopes of the two distant archives of Greenland ice and Gerzensee lake marl was synchronous and without delays. Quasi-simultaneous occurrence of major changes is indeed likely since the climatic conditions in Greenland and on the European continent are synoptically dependent, e.g. by the North Atlantic Oscillation (Hurrell et al., 2003). Also, a mechanism that would lead to a very similar but time-shifted climatic pattern at different geographical sites, is unlikely, since a climatic pattern delayed by more than the synoptic time scale would be dampened and altered.

Thus, because of the expected synchronicity of these climatic shifts in the North Atlantic region, \(\delta^{18}O\) fluctuations can be used as time markers and can be correlated not only within one archive but also between different records from different regions. This approach was used for establishing an age-depth model for the new high-resolution Gerzensee record, because building a radiocarbon-based chronology was not possible due to the absence of terrestrial macrofossils in the studied cores. In addition, precise conventional radiocarbon dating would be difficult because a plateau of constant radiocarbon age during the early part of the Bölling (Ammann and Lotter, 1989; Reimer et al., 2009) prevents a precise age assignment by \(^{14}\)C-dating. In addition, annual laminations and tephra layers (except for the Laacher See Tephra) to further constrain the age are not present during the period of interest.

The NGRIP \(\delta^{18}O\) record on the Greenland Ice Core Chronology 2005 (GICC05) (Rasmussen et al., 2006) was used for building the age-model for the B/A. This record seems most suitable for a comparison with the Gerzensee record, since (1) no European \(\delta^{18}O\) record is well enough established to meet the conditions necessary for high-resolution correlation during the late-glacial period, (2) the Greenland ice cores are some of the best absolutely dated archives for the period of interest, and (3) the NGRIP record was adopted by the North Atlantic INTIMATE group as a regional stratotype (Lowe et al., 2008). The resulting high-precision chronology of the lake marl in Gerzensee is the basis for further studies (this volume) that compare sedimentological and biological changes and to evaluate response times and mechanisms and reconstruct environmental development during times of rapid climatic change.

2. Study site and analytical methods

2.1. Lake Gerzensee

Gerzensee is a small kettle-hole lake on the Swiss Plateau at 603 m asl (46°49’56.95”N, 7°33’00.63”E) (Fig. 1). The lake is situated on a ridge separating the valleys of the Aare and the Gürbe rivers. Its low catchment relief protects the lake from major erosional input. The small catchment area of 2.6 km\(^2\) (Lotter et al., 2000) is underlain by the till of the Aare glacier over Miocene Molasse sandstone. Today the lake has a small artificial inflow from the North, a lake surface of 25.16 ha and a maximum water depth of 10.7 m. The surroundings of Gerzensee became ice-free about 17–18 kyr BP (Preusser, 2004; Ivy-Ochs et al., 2008). Because the water level was higher during the late-glacial than today (Eicher, 1979) late-glacial shallow-water carbonates occur underneath about 1 m of soil in today’s reed zone. This late-glacial marl formed a littoral sub-aquatic terrace, comprised mainly of authigenic carbonates (both inorganically precipitated and bio-induced) with some molluscan shell debris but very little organic matter and detrital material. Four parallel cores were recovered on the eastern shoreline of Lake Gerzensee in the years 1976 (labeled: GE III), 1992 (GEAB), 2000 (GEJK), and 2008 (GEM) (Fig. 1, Table 1).

2.2. Analytical methods

The sampling resolution for continuous stable-isotope analyses ranged between 5 and 0.5 cm (see Table 1 for details). For all cores the sampled material was freeze-dried, and shells were carefully removed from the sediment before analysis to minimize biogenic calcite contributions to the isotope signal of the inorganic calcite (Fig. 2). The term “bulk carbonate” (\(\delta^{13}C_{\text{bulk}}\)) is here used for the fine-grained inorganic lake marl. Carbon and oxygen isotopic compositions were measured in the bulk carbonate of four different cores of the Gerzensee marl (Table 1). Resulting values are reported in the conventional delta notation with respect to VPDB.

For core GEM, the analyses were made in the isotope-geochemistry laboratory of ETH-Zürich on a Thermo Fisher Scientific GasBench II coupled to a Delta V mass spectrometer. About 350 µg of powdered sample was placed in 12 ml vacutainers, flushed with helium, and were reacted with 5 drops of 10% phosphoric acid at 72 °C. The instrument was calibrated with the international reference materials NBS 19 (\(\delta^{13}C = +1.95\,\text{‰}\), \(\delta^{18}O = -2.2\,\text{‰}\)) and NBS 18 (\(\delta^{13}C = -5.05\,\text{‰}\), \(\delta^{18}O = -23.1\,\text{‰}\)). The analytical reproducibility based on repeated measurement of an internal standard was better than ±0.08‰.

For the cores GEJK, GEAB, and GE III each sample of <50 mg of carbonate was reacted with 95% phosphoric acid (\(\text{H}_3\text{PO}_4\)) at a constant temperature of 50 °C for 1 h to produce CO\(_2\), which was then analyzed for its \(^{18}O/^{16}O\) ratio with a mass spectrometer ( Finnigan MAT 250) at the University of Bern (Siegenthaler and Eicher, 1986).

The scale was calibrated and checked regularly with international standards available through the International Atomic Energy Agency (IAEA). The analytical precision is about 0.02‰ for \(\delta^{18}O\). Long-term stability is controlled by an internal laboratory standard (marble) with −2.85‰ on the VPDB scale. Over several decades the values of the internal standard are within 0.1‰. The carbonate content was generally around 80–90% with a minimal value of 45%.
Table 1
List of sediment cores indicating the depth intervals and resolution of the individual stable-oxygen records used in this study. *1: Laacher See Tephra (LST) as reference level was set to 272 cm core depth.

<table>
<thead>
<tr>
<th>Core name</th>
<th>Coring date</th>
<th>Analyzed interval (cm)</th>
<th>Sampling interval (cm)</th>
<th>Sampling resolution (cm)</th>
<th>Coring method</th>
<th>Thickness start Belling–LST (cm)</th>
<th>Thickness LST-start YD (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GEM</td>
<td>Sept. 2008</td>
<td>70.5–406.5</td>
<td>70.5–406.5</td>
<td>0.5</td>
<td>Uwitec piston corer (5.8 cm)</td>
<td>73</td>
<td>16</td>
</tr>
<tr>
<td>GEJK</td>
<td>Sept. 2000</td>
<td>270–414</td>
<td>270–336</td>
<td>0.5</td>
<td>Streif modified Livingstone corer (8 cm)</td>
<td>102</td>
<td>– no data –</td>
</tr>
<tr>
<td>GEAB</td>
<td>Oct. 1992</td>
<td>169–364</td>
<td>191–201</td>
<td>0.5</td>
<td>Streif modified Livingstone corer (8 cm)</td>
<td>92</td>
<td>19</td>
</tr>
<tr>
<td>GE III</td>
<td>June 1976</td>
<td>112–414</td>
<td>245–275</td>
<td>2.5</td>
<td>Streif modified Livingstone corer (8 cm)</td>
<td>77</td>
<td>20</td>
</tr>
</tbody>
</table>

3. Method for establishing a high-resolution core chronology

The Gerzensee chronology was established by applying an “event stratigraphy” approach in order to correlate the Gerzensee lake marl record with the well-dated $\delta^{18}$O isotopic records from Greenland ice cores. Whittaker et al. (1991) describe event stratigraphy as the procedure of correlating strata on the basis of geologically short-lived events such as volcanic eruptions or climate excursions, which are expressed in a great variety of records and therefore can be used for correlation of different proxies (e.g. pollen, insects, and isotopes) and localities.

To highlight the overall signature of the Gerzensee $\delta^{18}$O marl record and to eliminate sampling or analytical artifacts, a stack of four parallel $\delta^{18}$O records was established that was then matched to the NGRIP $\delta^{18}$O record.

Fig. 3 shows the workflow in which (a) the original $\delta^{18}$O data sets from GEJK, GEAB and GE III were correlated to a common depth scale (i.e. the GEM depth scale), (b) then mathematically resampled to the same resolution, and (c) combined to a weighted stack. This stack was then (d) visually correlated to NGRIP $\delta^{18}$O and partially fine-tuned (e) by a Monte Carlo method. Finally, (f) a chronology for each original $\delta^{18}$O core record was determined by using the now established relationships of the original depth to the common depth scale (established in step b) and of the common depth scale to NGRIP-age (established in step e).

For the above-described procedure, AnalySeries 2.0 (Paillard et al., 1996) was extensively used. For resampling (steps b and f), simple linear interpolation was applied. For matching two or more records the lineage function was utilized (steps a and d).

In the following each step is discussed in detail.

(a) Correlation of four sediment cores

The four individual sediment cores GEM, GEJK, GEAB, and GE III were first correlated by synchronizing the most prominent isotopic shifts at the beginning and end of Bølling/Allerød interstadial (Fig. 4, tie point 1, 2, 11, 12). The records were fine-tuned during the B/A by the pollen abundances of Betula, Pinus, Salix, Gramineae and Juniperus, which were correlated whenever they were available (Fig. 4, tie points 4–13). The tie points are mainly based on minima and maxima of the pollen abundances. The higher the sampling resolution and number of available records the smaller the uncertainty in the position of tie points. Thus, the uncertainty of the position of the tie points is mostly less than 2 cm, where all four records are available. However, tie points 6 and 7 are defined mainly by synchronizing the gradients in Pinus and Betula abundances of only two available records (GE III and GEJK), which leads to an uncertainty within a few centimeters regarding the exact position of these tie points.

Fig. 2. Original $\delta^{18}$O records on individual depth scales. The Laacher See Tephra (LST, yellow vertical line) was set as a common reference level to 272 cm core depth. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 3. Schematic workflow for establishing the age-depth model for the Gerzensee sediments. For details see “Section 3. Method for establishing a high-resolution core chronology”.

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A distinct 6–8 mm thick volcanic ash layer visible in all four cores, stratigraphically and chemically identified as the Laacher See Tephra (LST), is used as an additional independent marker and set as a reference level to 272 cm core depth corresponding to the center of the tephra layer (Fig. 4, tie point 3). As the δ18O record from GEM had the highest resolution, a constant sampling resolution of 0.5 cm over the investigated time interval, the GE III, GEAB, and GEJK cores was transferred on the GEM depth scale (Fig. 5a). Thus, the stack is dominated by the δ18O signal of GEM and partially GEJK.

The weighted sample mean (=weighted stack) is:

\[ \bar{x}_w = \frac{\sum_{i=1}^{n} (w_i \times x_i)}{\sum_{i=1}^{n} w_i} \]

with:

- \( \bar{x}_w \): weighted sample mean
- \( w_i \): allocated weight for the given data
- \( x_i \): observed value for a given data

The standard error of the weighted stack has been estimated by using the equation:

\[ s_w = \sqrt{\frac{\sum_{i=1}^{n} w_i (x_i - \bar{x})^2}{(n-1) \sum_{i=1}^{n} w_i}} \]

with:

- \( s_w \): standard deviation of the weighted stack (eq. 1.105a, p. 160 in Sachs, 1997).

Considering the small degree of freedom (number of stacked records – 1) in the standard error calculation we multiply \( s_w \) by a factor of 100 to figure legend, the reader is referred to the web version of this article.

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obtained with Eq. (2) by Student’s t value for p = 0.317 (confidence level corresponding to one standard deviation of a normal distribution). This results in a consistent confidence interval throughout the record.

Because the relative weighting of each single record may vary within the stack, we checked that the basic characteristics of the stacked record did not change. By comparing the weighted stack with the normal averaged stack, it was assured that there were no artificial steps or distinct signatures at the positions where there are changes in the relative weighting of the stack (Supplements 1).

(d) Correlation to NGRIP

The chronology was established by matching the Gerzensee and Greenland isotopic records. This relies on the assumption that changes in $\delta^{18}$O in Greenland and Europe occurred simultaneously. For better comparison with terrestrial, marine, and archeological records that are expressed in radiocarbon years before AD 1950, the GICC05 timescale (Rasmussen et al., 2006), which is originally in years b2k (before 2000 AD), was converted to years BP (before 1950 AD). In the Greenland ice core record the age and depth values refer to the bottom of each interval. However, the depths in the Gerzensee isotope data refer always to the middle of each sampling interval. Therefore, the NGRIP ages calculated for the middle of each interval were used for matching the two records.

First, the prominent $\delta^{18}$O-shifts at the beginning and end of B/A were correlated. Then, three minor oscillations visible clearly in both, the NGRIP and Gerzensee $\delta^{18}$O records, were correlated (dashed red lines in Fig. 6). Applying this approach, one needs to be aware that the resulting timescale for the Gerzensee record is not independent but is based on a correlation to NGRIP (NGRIP-dating-group, 2006) and GICC05 (Rasmussen et al., 2006).

(e) Monte Carlo simulation

For detailed correlation between the two $\delta^{18}$O records of Gerzensee and Greenland, we applied a Monte Carlo method, as described in Schwander et al. (2000). This method is a stochastic technique that matches two time series by randomly shifting the time points of one series until the best correlation between the two series is reached. Constraining parameters used in this study are: a minimum of correlation coefficient of 0.4, a total of 50 points for the correlation (size of window), and a range of slope between two records of 0.4 and a maximum age shift of ±25 years.

First, we selected a subset of each time series, including the two distinct major shifts in $\delta^{18}$O at the beginning and end of the B/A, resulting in subsets of 183 $\delta^{18}$O values for Gerzensee (i.e. between ca. 12,800 and 14,800 BP) and 124 for NGRIP record (i.e. between ca. 12,500 and 15,000 BP). Then the age window of the lake subset was placed at one end of the ice core subset, and a preliminary age from the previously established visual match was assigned to the first data point of the Gerzensee subset. To find the best correlation of the two records, all points of the window, except for the first one, varied randomly within meaningful limits (Schwander et al., 2000). A significant correlation was considered when the correlation coefficient ($r$) exceeded a threshold of 0.4. Once the simulation determined a group of values with the highest correlation coefficient, the window was shifted one data point, with the first point fixed to the age of the best correlation point, and the randomization process repeated. In this manner the window was subsequently shifted until the end of the record was reached. The resulting ages of the new time scale are the averaged values from the best correlations of each window position. In order to achieve a high level of accuracy, the simulations were performed on 75,000 trials per step and a total of 15 runs through the time series.

Two main factors may significantly affect the results of the Monte Carlo simulation: (1) the total width of the windows, which is the number of correlation points and (2) the maximum age shift of each point of the window. Thus, sensitivity tests for both variables were performed by first varying the width of the window (4, 8, 25, 50 correlation points) and then the maximum age shift (10, 50, 100 years), while keeping all the other parameters constant. A maximum age shift of ±25 years and a maximum of 50 correlation points were chosen to avoid possible overturning (Supplement 2). All other parameters were set as in Schwander et al. (2000). Finally, to account for the direction in which the window moved through the record, the analysis was run for both directions of the moving window (each in 10 replicates).

In principle, the Monte Carlo method can be expected to be less subjective than visual matching. However, it can be observed that the differences between the results generated by the visual matching and the Monte Carlo method are marginal and within uncertainties (less than ±15 years) (Fig. 6), supporting the objectivity of the performed visual match. Nevertheless, one must be aware of the limits of such matching approaches. In fact, it is often possible to achieve a good correlation of very

Fig. 6. Correlation between the NGRIP and stacked Gerzensee $\delta^{18}$O records on GICC05-timescale (converted to years before 1950 AD). A) Local isotope zonation (GRZ bulk = Gerzensee oxygen isotope of bulk sediment without shell debris) and detailed modified Greenland terminology (GI = Greenland Interstadial, YD = Younger Dryas) divide the records into periods of higher and lower $\delta^{18}$O values. Dashed lines mark the visually assigned tie points between the NGRIP and the stacked Gerzensee record. The result of the fine-tuning of the visual match by applying a modified Monte Carlo method is shown for the period of 12.8-14.8 kyr BP. The age difference between the visual correlation and the correlation by the Monte Carlo method is shown in B).

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different records if no constraints are imposed on stretching and compressing the time axis. This can easily lead to overtuning, that is matching of signals that are not causally linked. A subtle choice of constraints is therefore the basis of any time-series matching.

(f) Resampling for individual chronologies

The relationship between the original depth scales and the common GEM depth scale (established in step b) and the relation of the GEM-depth scale to NGRIP age (established in step e) were used to determine the relation between the original depth and the NGRIP age. This leads to individual chronologies for each sediment core (Fig. 7).

4. Results

Fig. 6 shows the high-resolution chronology for the δ18O Gerzensee stack established by correlating δ18O of Gerzensee and NGRIP. The temporal resolution of the Gerzensee stack, which is strongly based on the spatial resolution of 0.5 cm in the GEM and GEJK cores, is 8.2–14 years during the period of interest. This confirms that the comparison with the NGRIP record, which has a temporal resolution of 20 years, is appropriate. Although the Laacher See Tephra is not found in the Greenland ice record, its age can be determined from our isotope correlation to 13,034 yr BP.

Signal noise, erratic sedimentation features (such as small scale redeposition of sediment), and possible sampling artifacts (such as small shell fragments within a bulk sample) can be minimized by stacking the four available data sets. Since the stacked oxygen isotope record of the lake carbonate is smoother than each original data set and mainly reflects overall features, it best represents 13 local isotope zones. GRZ b̄ulk 1 being the oldest (Fig. 6). The δ18O-values abruptly rise from about −9‰ before 14,685 BP (GRZ b̄ulk 1) in −95 years (GRZ b̄ulk 2) to −5.8‰ (GRZ b̄ulk 3) and then continuously decrease during the following −1500 years (GRZ b̄ulk 4–10). There is a final peak of −6.3‰ at about 12,880 years (GRZ b̄ulk 11) before the abrupt drop in δ18O during −150 years (GRZ b̄ulk 12) leading to values of around −9‰ (GRZ b̄ulk 13).

A significant feature of the stacked δ18O record is the pattern of short-term oscillations imposed on the long-term trend. Three distinct fluctuations with pronounced negative isotopic shifts can be identified in the stacked isotopic record, as well as in all individual records. GRZ b̄ulk 6 at a sediment depth of 322–314 cm (14,044–13,908 yr BP) shows an abrupt drop from about −7 to −8‰ in δ18O values. GRZ b̄ulk 8 at a sediment depth of 298–293 cm (13,624–13,522 yr BP) shows a slightly gradual decline towards a final distinct depression, which is followed by a sharp increase in δ18O from about −7.5 to −8.4‰. GRZ b̄ulk 10 at a sediment depth of 282.5–269 cm (13,274–12,989 yr BP) shows a gradual low from about −7.2 to −8.3‰ in δ18O. Thus, GRZ b̄ulk 6 records the largest δ18O shift, whereas the oscillation in GRZ b̄ulk 10 lasts the longest. All three observed distinct minor oscillations reach down more than halfway to earlier stadial δ18O values. An additional minor oscillation (GRZ b̄ulk 4, 336–326 cm, 14,439–14,183 yr BP) can be observed in the stacked isotopic record with only a small decrease in δ18O values of about 0.5‰.

Comparison of the individual original data sets shows that the sediment thickness from the beginning of the Bølling to the LST ranges between 73 and 102 cm in the four cores (Table 1). The individual records of GEM, GEJK, GEAB, and GE III (Fig. 7) show similar sedimentation rates mostly ranging between −3 and 6 cm/100 yr during B/A. Relatively low sedimentation rates occur in the Bölling period. Intermediate sedimentation rates characterize the middle and late Allerød, with a slight increase after the deposition of the LST. Increased sedimentation rates occur in the first half of the Allerød in all four cores, however with especially high rates of up to 15 cm/100 years in core GEJK between −13.6–13.8 kyr BP. This observation indicates some variability in the sedimentation rates during the late-glacial, even though the cores were recovered within a small area. Small-scale lateral heterogeneity in lacustrine settings was documented earlier by Lotter et al. (1997). This period of high sedimentation rate lies within a longer period of rather low-to-intermediate lake levels (Magny, this volume). Whether these variations in sedimentation rate during low lake level arise from small-scale increased calcite production in shallower water or from patchy redistribution of carbonates on the shallow platform remains unclear.

5. Discussion

5.1. Gerzensee lake marl

δ18O of inorganic carbonates in small/medium open lakes dominantly reflect δ18Owater during calcite precipitation (see Ito, 2001; Leng and Marshall, 2004; Teranes et al., 1999 for review). Authigenic calcite inorganically precipitates in late spring/early summer in the epilimnion because of changes in lake water chemistry (increased temperature, algal activity, and pH). However, inorganic carbonates may also be formed as an extra-cellular by-product during photosynthesis. For example, Characeae and other micro and macrophytes actively remove bicarbonate from the lake system, leading to the formation of inorganic calcite, which typically encrusts the plants (von Grafenstein et al., 2000). For the Gerzensee lake marl record it has been shown that the fine-grained carbonate matrix is composed of a mix of authigenic calcite and of disaggregated calcitic encrustations (von Grafenstein et al., 2000) that commonly form on the
stems and oogonia of the aquatic macrophyte Chara, as also seen in other lakes (Apolinarska and Hammarlund, 2009). It is debated in the literature to what extent calcite encrustations precipitate in isotopic equilibrium with the lake water. Especially for the oxygen isotopes of encrustations the opinion differs between isotopic equilibrium (Ito, 2001) and disequilibrium (Pelechaty et al., 2010). However, the $\delta^{13}C$ consistently is shown to be enriched in calcite encrustations. During sample preparation, biogenic carbonate (i.e., ostracods, gastropods) was removed, but it was not possible to discriminate between authigenically formed carbonate and fine particles of Chara encrustations.

Disregarding the precise formation of the inorganic calcite, the similarity of the lake records and the excellent correlation with the NGRIP record validate the $\delta^{18}O_{\text{bulk}}$ proxy for reconstructing climatic changes. In addition, $\delta^{18}O_{\text{bulk}}$ has in general the same signature as the $\delta^{18}O$ of discrete Chara encrustations and molluscan Pisidium shells, which have been observed to represent mainly the $\delta^{18}O_{\text{water}}$ in which the calcite precipitates. This is closely related to $\delta^{18}O_{\text{precipitation}}$ and thus to mean annual air temperature (von Grafenstein, this volume).

A large input of detrital carbonates potentially in the calcite precipitates. This is closely related to $\delta^{18}O_{\text{water}}$ influencing the isotopic changes. In addition, $\delta^{18}O_{\text{water}}$ might be enriched in calcite encrustations. During the calcite precipitates, which have been observed to represent mainly the $\delta^{18}O_{\text{water}}$ in which the calcite precipitates. This is closely related to $\delta^{18}O_{\text{precipitation}}$ and thus to mean annual air temperature (von Grafenstein, this volume).

**5.2. Terminology of events**

Application of the concept of synchronous large-scale climatic changes makes it possible to correlate and compare different late-glacial chronostratigraphic units of various records in the North Atlantic region. A chronostratigraphic unit is a body of sediment strata representing the sediments deposited during a specific interval of geologic time (Mangerud et al., 1974). Thus, chronostratigraphic units in theory are of same age in different regions. However, due to uncertainties in the original studies or misinterpretations by later scientists, as well as regional differences, much confusion exists about the use and implication of the various terminologies during the late-glacial period (Litt et al., 2003; de Klerk, 2004) (Tables 2, 3). Some confusion probably was caused when proxy records were temporally not resolved high enough and therefore missed short-term climatic oscillations of decadal- to centennial-scale duration. Another source for confusion arises from using the same name for chronozones, which represent same time intervals, and biozones, which represent same biological assemblages and may not be synchronous over wider distances. Some earlier studies also use terms such as “Middle Dryas” (Bock et al., 1985) or “Earlier/Late Dryas” (van Geel et al., 1989) instead of the conventional terms “Oldest/Older/Younger Dryas”, which might have led to additional confusion.

Table 3 shows several examples of chronological terms used for naming events and periods during the late-glacial (also called Termination 1) in the North Atlantic region. The stadial (GS-2) preceding the late-glacial

### Table 2

<table>
<thead>
<tr>
<th>Local isotope zonation GRZi bulk</th>
<th>Classical terminology used in Swiss lakes</th>
<th>Modified Greenland terminology</th>
<th>Gerz stack age (yr BP)</th>
<th>Gerz stack depth (cm)</th>
<th>GEJK depth (cm)</th>
<th>Lowe et al. 2008 GICC05 (yr BP)</th>
<th>INTIMATE Greenland terminology</th>
</tr>
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<tr>
<td>13</td>
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<td></td>
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<tr>
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<td>338.5</td>
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<td>336.25</td>
<td>361.5</td>
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<td></td>
</tr>
</tbody>
</table>

*1: High sampling resolution enables definition of transitions as separate zones.
*2: Differs from Greenland terminology, since transitions (GRZ ibulk 2 and 12) were added as separate zones.
*3: Also known as IACP (Lehmann and Keigwin, 1992), parallel to Killarney Oscillation in North America (Levesque et al., 1993).
*4: Laacher See Tephra.
*6: GI-1e was separated into sub-zones (this study), term “Inter Bølling Cold Period” (IBCP) was avoided since it has been used inconsistently (e.g. Karpuz and Jansen, 1992; Hughen et al., 1996).
interstadial is commonly called Oldest Dryas but Pleniglacial in northern and northwestern Germany. The stadial (GS-1) succeeding the late-glacial warm period is consistently called Younger Dryas. The late-glacial interstadial itself (GI-1) is named Meiendorf/Bølling/Allerød in northern and northwestern Germany, whereas in most other areas of the North Atlantic region it is consistently called Bølling/Allerød period. However, the definition of the described boundary between the Bølling and the Allerød varies in different records (see Table 3).

Two to four cold phases during the late-glacial interstadial are recognized in various records of the North Atlantic region (Table 3). They can be correlated with the Greenland ice core records. The earliest cold oscillation during the Interstadial (e.g. GRZ ibulk 4) lies within GI-1e and is named Bølling Cold Period I (BCP I) in the Norwegian Sea (Karpuz and Jansen, 1992) but is not described in many other regional stratigraphic schemes, but rather be used as a standard against which regional stratigraphic schemes are compared as suggested by the North Atlantic INTIMATE group (Lowe et al., 2008). However, it seems that the Greenland terminology is widely used, and introducing new local terms would rather increase confusion in cases where a solid correlation to the Greenland chronology can be established. The highly resolved Gerzensee marl record allows for the description of additional fluctuations, and the determination of the transitions at the onset and termination of the B/A interstadial as separate zones. Therefore, the terminology was slightly modified at the B/A transitions (Fig. 6). This facilitates further studies (this volume) on the timing and response mechanisms of environmental changes to climatic change.

5.3. Age of Laacher See Tephra

A unique time marker at the end of the Bølling/Allerød warm period is the tephra layer of the phonolitic Laacher See eruption in the Eifel region, Germany. The Laacher See Tephra (LST) is widespread in central and northern Europe (Bogaard and Schmincke, 1985). In Gerzensee, it can be stratigraphically correlated in all four Gerzensee cores as a gray layer of 6 to 8 mm thickness with discrete volcanic glass shards, which have been identified chemically by Walter-Simonnet et al. (2008).

The isochronous nature of the LST allows the new Gerzensee chronology, established by event-stratigraphical correlation with NGRIP, to

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<table>
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<tr>
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<th>Eifel region, N-Germany</th>
<th>N-America, Cariaco basin</th>
<th>N-America</th>
<th>Norwegian Sea</th>
<th>N-Atlantic</th>
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<td>Killarney Oscillation</td>
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</table>

Table 3
Overview of different terminologies used for synchronous intervals in various records of different areas in the North Atlantic region. GI = Greenland Interstadial, GS = Greenland Stadial, IACP = Inter Allerød Cold Period, BCP = Bølling Cold Period.
To calculate the age of the LST in Soppensee, Switzerland, to 12,975 (INTEX09). (2010) determine a calendar age of 12,735 BP, taking the combined age estimates of two models, the P_Sequence (2006), this age is well within range of previous studies. (Brauer et al., 2008), but well within error of tree ring age estimation in the NGRIP ice cores into account (about ±150 yr, Rasmussen et al., 2009). Friedrich et al. (1999) yields a calibrated age range of 12,769–13,098 cal BP (two sigma range). Another method to determine the age of the LST is by dendrochronology, by which the observed 192-year difference between the Laacher See Eruption and the onset of the Younger Dryas (Kaiser, 1993) is combined with the recently determined onset of the YD (~12,760 cal yr BP in Hua et al., 2009). This would result in an age for the LST of approximately 12,952 cal yr BP. The LST is also found in numerous European lakes, among them varved lake records, revealing slightly younger ages for the LST. An independent varve chronology from the Meider Maar and Holzmaar, Germany, dates the LST at 12,880 (±120) yr cal BP (Brauer et al., 1999). Blockley et al. (2008) calculate the age of the LST in Soppensee, Switzerland, to 12,975–12,743 cal BP, taking the combined age estimates of two models, the P_Sequence by depth and by varve spacing (INTCAL 04). Hajdas and Michczynski (2010) determine a calendar age of 12,735–12,878 cal BP for LST in Soppensee (INTCAL09). The age of the LST in our study of 13,034 cal yr BP is slightly older than that suggested by the latest studies from Soppensee (Blockley et al., 2008; Hajdas and Michczynski, 2010) and Meider Maar (Brauer et al., 2008), but well within error of tree ring age estimation (Friedrich et al., 1999). However, taking the error range of layer counting in the NGRIP ice cores into account (about ±150 yr, Rasmussen et al., 2006), this age is well within range of previous studies.

5.4 High-resolution paleoclimatic implication

Paleoclimatic interpretation from δ18O of lacustrine carbonates from lakes is challenging, since direct calibration with temperature or δ18O in precipitation is difficult. Changes in δ18O of inorganic lake carbonates can be attributed to (1) changes in water temperature, (2) changes in history/source of the water, (3) variations in biological activity, or (4) changes in hydrology of the lake (Leng and Marshall, 2004; Bernasconi and McKenzie, 2007). Von Grafenstein et al. (2000) postulated that the hydrology at Gerzensee and thus the evaporative enrichment of δ18O_water was rather constant, with approximately 2% during most of the Holocene and late-glacial. Thus, we exclude hydrological changes and assume mostly climatic causes to be the dominant factor controlling changes in δ18O bulk-Schmid (2011) applied clumped isotope analyses to the Gerzensee late-glacial carbonates and found an increase of 2.6% in δ18O_water at the onset of the Bølling and a decrease of 2.2% at the transition from the BA to the YD. These oxygen-isotope shifts in the lake water seem not to correlate with clumped-isotope temperatures, which would reflect water temperatures during calcite precipitation. However, δ18O_water seems to be directly related to changes in δ18O_precipitation, mainly reflecting mean annual air temperatures. Thus, the δ18O bulk can be used as a proxy for annual air temperature change in the region around Gerzensee (Schmid, 2011). This observation is in accordance with the temperature reconstructions and interpretation of von Grafenstein (this volume, 2000) and Lotter et al. (2000, 2012).

With a sampling resolution down to 0.5 cm the resulting data set yields a temporal sampling resolution of ~10 years during the Bølling/Allerød. This permits a detailed comparison between the δ18O record from Gerzensee and the records from other late-glacial archives, as well as for the reconstruction of abrupt climatic changes.

Comparison of the four Gerzensee δ18O records shows some similarities and differences (Fig. 7). The three conspicuous δ18O oscillations (GRZ ibulk 6, 8, and 10) are visible in all records, even at low sampling resolution (GE III). However, GRZ ibulk 6 has a different structure in GEAB, which might be due to variations in the shallow-water sedimentation or a sampling artifact. GRZ ibulk 8 has slightly different structures in all four cores, but all show a gradual decline towards a final distinct depression, which is followed by a sharp increase in δ18O. GRZ ibulk 10 has in all four cores a wide gradual δ18O Depression. The oldest oscillation GRZ ibulk 4 is less pronounced than the other ones and might be an overestimation of noise in these records. However, other proxies such as for lake-level fluctuations (Magny, this volume) at Gerzensee as well as other archives (Karpuz and Jansen, 1992; Marshall et al., 2002) in the North Atlantic region show a distinct cold-temperature oscillation during the early part of the interstadial.

These century-scale cold events during a general cooling trend of the Bølling/Allerød warm period in the Gerzensee δ18O stacked record are similar to several other records in the North Atlantic region, despite differences in archives, geographical settings, or the proxies used. However, not always all four oscillations can be observed, most likely due to insufficient stratigraphic resolution. In many records only GI-1b and -1d are described, since those are the most pronounced cold oscillations (Table 3).

Friedrich et al. (2001) observe several depressions in tree ring width from trees in Germany, Switzerland, and Italy that most probably reflect colder temperatures in southern and central Europe during growing season. The floating tree ring record shows decadal scale cold periods, i.e. a period of 25–39 years, assumed to be similar to GI-1d (Aegelsee Oscillation) and a period of ~60 years equivalent to GI-1b (Gerzensee Oscillation). According to our chronology, GI-1d and GI-1b resulted in cooling periods of centennial scale, i.e. ~135 and 285 years, respectively. The discrepancy between the tree ring and lake marl records may arise from inconsistent definitions of the event boundaries and different sensitivities of the systems to temperature change. In addition, the tree-ring record also contains depressions not reflected in the Gerzensee lake marl δ18O record, which might reflect specific local terrestrial conditions or be a result of higher resolution of the tree-ring record.

The three prominent oscillations during B/A have also been observed in the gray-scale record of marine sediments from the Carico basin north of the Venezuelan coast. This record is interpreted as a proxy for a change in the trade wind strength in the tropics in response to changes in the temperature gradient between high latitudes and the tropical North Atlantic (Hughen et al., 1998). Whatever the cause, the minor cooling events during the B/A appear to affect both low and high latitude regions in the same manner. However, Marshall et al. (2002) note a relative difference in magnitude of temperature values of the events at different sites. In the terrestrial archives (lakes and ice), all records show event GI-1d (Aegelsee Oscillation) as the largest, while in marine records GI-1b (Gerzensee Oscillation) is the most pronounced. Also Renissen and Isarin (2001) point out, that the magnitude of temperature change expected during the late-glacial period strongly depends on the location of a site in Europe (the latitude, altitude, and proximity to an ice sheet). It is therefore quite possible that such minor oscillations are recorded more strongly at sites closer to ecotonal boundaries (Heegaard et al., 2006) located at higher latitudes or altitudes (Brooks and Birks, 2000a, 2000b; Heiri et al., 2007).

The origin of these widespread short-term climate oscillations and the role of internal climatic variability, solar activity, volcanism and other forcings are still under debate. Considering the length and abruptness of the δ18O depressions during B/A, several geological and modeling studies suggest a disruption of the oceanic thermohaline circulation by meltwater discharge from the continental ice sheets as a likely mechanism for abrupt cooling of the North Atlantic.
region during the late-glacial and Holocene (Clark et al., 2001; Teller et al., 2002; Nesje et al., 2004; Donnelly et al., 2005; Renssen et al., 2007; Fleitmann et al., 2008; Kleiven et al., 2008). Changes in the North Atlantic thermohaline circulation would have a strong influence on the atmospheric circulation and moisture transport pathways, as discussed by several authors (e.g. Brauer et al., 2000; Yu, 2007). Björck et al. (1996) and Thornalley et al. (2011) state that the minor oscillations during the late-glacial and early Holocene coincide with ocean ventilation minimum phases. However, clear evidence for the origin of the freshwater input and interrelation of freshwater outbursts and cooling events is sparse and not yet consistent. Stanford et al. (2006), Bard et al. (1996), and Clark et al. (1996) suggest that the Meltwater pulse 1A (MWP-1A) coincides with the abrupt cooling of one of the short-term cooling events, whereas several other authors predate the MWP-1A concurrent to the sharp late-glacial warming (Kienast et al., 2003; Weaver et al., 2003; Nesje et al., 2004; Deschamps et al., 2012). Donnelly et al. (2005) present evidence from North America that a meltwater discharge may have played an important role in triggering the last cold oscillation of the B/A by reducing the thermohaline circulation. Nesje et al. (2004) show that three centennial-scale climatic deteriorations during the B/A were probably linked to three corresponding oceanic rerouting events which caused reduction in thermohaline circulation and cooling in the North Atlantic region. However, further evidence for the source and mechanisms of freshwater discharges are needed.

In addition, Yu (2007) noted that the general trend during the B/A warm period is different among various records. While the δ18O at Crawford Lake and the Greenland ice cores show a declining trend during the B/A warm period, the records from Gerzensee, White Lake, Ammersee and Cariaco show a plateau-like Bölling–Allerød warm period (Yu and Eicher, 1998). These trans-Atlantic similarities and differences hint at the existence of a strong spatial gradient in climatic changes presumably because of changes in moisture source (Yu, 2007). This is in accordance with Magny (this volume) who observes generally higher lake levels during periods of low δ18O in the lake carbonates, which might imply changes in the precipitation pattern coinciding with a shift in the thermocline of the lake water.

6. Conclusions

A large number of terrestrial records are available from the late-glacial period (Termination 1, ~15–10 kyr BP), which give highly detailed pictures of past environmental change. Nevertheless, the Greenland oxygen-isotope records are so far regarded as the best-dated and most-detailed high-resolution climate proxy for the North Atlantic region. Our new developed stacked oxygen isotope record from Gerzensee serves as an additional terrestrial record from continental Europe during the Bølling/Allerød period with an exceptional resolution. The δ18O dataset highlights the overall common features of

Since the stacked record is more representative for the lake record than each single stack it mainly reflects overall features highlighting the presence of three distinct and one less expressed δ18O-depression during the B/A. Since the Greenland terminal is widely and consistently used, this terminology was slightly modified and transferred to the Gerzensee records. Thus, the observed δ18O-depressions correspond to GI-1e2, GI-1e1, and 1b. The last three distinct oscillations show a change of about 1‰ and thus have amplitudes of about one fourth to one third of the δ18O shift at both terminations in and out of the Bölling/Allerød period. These oscillations have been shown to be parallel to many other records due to hemispherical climate variations probably caused by freshwater discharges into the Atlantic. Further studies are needed to verify the timing, distribution, and scale of feedbacks of these centennial- to decadal climatic variations in the North Atlantic region during the warming of the late-glacial period.

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References


Teller, J.T., Leverington, D.W., Mann, J.D., 2002. Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. Quaternary Science Reviews 21 (8–9), 879–887.


