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Climate and environment during the Younger Dryas (GS-1) as reflected by composite stable isotope records of lacustrine carbonates at Torreberga, southern Sweden

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ABSTRACT: Climatic and environmental changes during the Younger Dryas stadial (GS-1) and preceding and following transitions are inferred from stable carbon and oxygen isotope records obtained from the sediments of ancient Lake Torreberga, southern Sweden. Event GS-1 is represented in the sediment sequence by 3.5 m of clay containing lacustrine carbonates of various origins. Comparison of isotopic records obtained on mollusc shells, ostracod valves, and *Chara* encrustations precipitated during specific seasons of the year supports estimates of relative changes in both lake water and mean annual air temperatures. Variations in soil erosion rates can also be estimated from a simple isotope–mass-balance model to separate allochthonous and autochthonous carbonate contributions to the bulk carbonate content of the sediments. The well-known, rapid climatic shifts characterising the Last Termination in the North Atlantic region are clearly reflected in the isotopic data, as well as longer-term changes within GS-1. Following maximum cooling shortly after the Allerød–Younger Dryas (GI-1–GS-1) transition, a progressive warming and a slight increase in aquatic productivity is indicated. At the Younger Dryas–Preboreal (GS-1–PB) transition mean annual air temperature rapidly increased by more than 5°C and summer lake-water temperature increased by ca. 12°C. The subsequent Preboreal oscillation is characterised by an increase in soil erosion and a slight decrease in mean annual air temperature. These results are in harmony with recent findings about large-scale climate dynamics during the Last Termination. Copyright © 1999 John Wiley & Sons, Ltd.

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KEYWORDS: Younger Dryas; GS-1; palaeotemperature estimates; stable isotopes; lake sediments; lacustrine carbonates; Sweden.

Introduction

The termination of the last glaciation was characterised by a series of more or less pronounced climatic oscillations (NASP Members, 1994; Lowe and NASP Members, 1995). Following the proposal by Björck *et al.* (1998), we have chosen to apply the event stratigraphy based on the GRIP $\delta^{18}O$ record to denote these climatic changes. Although GS-1 (Greenland Stadial 1), which was probably the most extensive and severe stadial, may have affected the entire Northern Hemisphere, the effects of the cooling on biota and physical environments were most profound and are most readily detected in northwest Europe, 'downwind' of the North Atlantic. Perturbations in the thermohaline circulation of the North Atlantic (Broecker, 1990) have been proposed as the main mechanism responsible for shifting climatic regimes in Greenland and northwest Europe (Björck *et al.*, 1996). Much of our current knowledge about climate dynamics during the Last Termination is based on data from marine sediments (e.g. Jansen and Veum, 1990; Lehman and Keigwin, 1992), and to an even greater extent on the Greenland ice-core records (e.g. Johnsen *et al.*, 1992; Taylor *et al.*, 1993). Continental climate records, however, are of vital importance for the validation of climate models (Renssen

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and Isarin, 1998) and for the appreciation of impacts on the terrestrial environment derived from possible future ocean circulation changes (e.g. Stocker and Schmittner, 1997). Lake sediment sequences are easily accessible climate archives, with the potential for providing palaeoclimatic data with precise chronological control and detailed temporal resolution for specific intervals.

Southern Sweden has a rich history of lake-sediment studies focusing on the climatic changes accompanying the Late Weichselian deglaciation (e.g. Nathorst, 1870; Holst, 1906; Nilsson, 1935). Proximity to the receding inland ice resulted in well-developed proxy climate records, and GS-1 is commonly represented in sediment sequences by a distinct unit of clay or gyttja clay bracketed by more organic-rich strata deposited during GI-1 (Greenland Interstadial 1) and the early Holocene respectively. Palaeoecological studies have focused mainly on the response of terrestrial and aquatic ecosystems to climate change through stratigraphic analysis of pollen and plant macrofossils (e.g. Berglund, 1966, 1971; Björck and Möller, 1987; Liedberg Jönsson, 1988). Recently such qualitative information has been supplemented by estimates of absolute temperature changes based on fossil Coleoptera (Lemdahl, 1988, 1991; Hammarlund and Lemdahl, 1994). There is also the potential for deriving quantitative palaeoclimatic data from oxygen isotope records obtained on autochthonous lacustrine carbonates, as demonstrated using subfossil ostracods in southern Germany by von Grafenstein *et al.* (1994), although this approach has not yet been applied to characterise the climate of GS-1 in Scandinavia.

Here we present stable carbon and oxygen isotope records obtained from various carbonate components preserved in the sediments of ancient Lake Torreberga, southern Sweden. The sediment sequence spans the later part of the Allerød interstadial (GI-1), GS-1, and the earliest Holocene (early Preboreal). Rapid deposition during this period affords relatively fine temporal resolution, enabling unusually detailed evaluation of short-term trends within GS-1 for comparison with other proxy climate records from the region. Reconstructions of changing aquatic productivity, terrestrial vegetation, and soil development, as well as estimates of changes in lake water and mean annual air temperatures, are presented. Some of these data have been considered in a previous examination of the GS-1–PB transition (Björck *et al.*, 1996). The present study is based on an extended data set enabling us to focus more specifically on GS-1 itself.

In the following text radiocarbon ages are expressed as 'yr BP', whereas calendar ages and time-spans are denoted 'cal. yr BP' and 'cal. yr', respectively (see Taylor *et al.*, 1996).

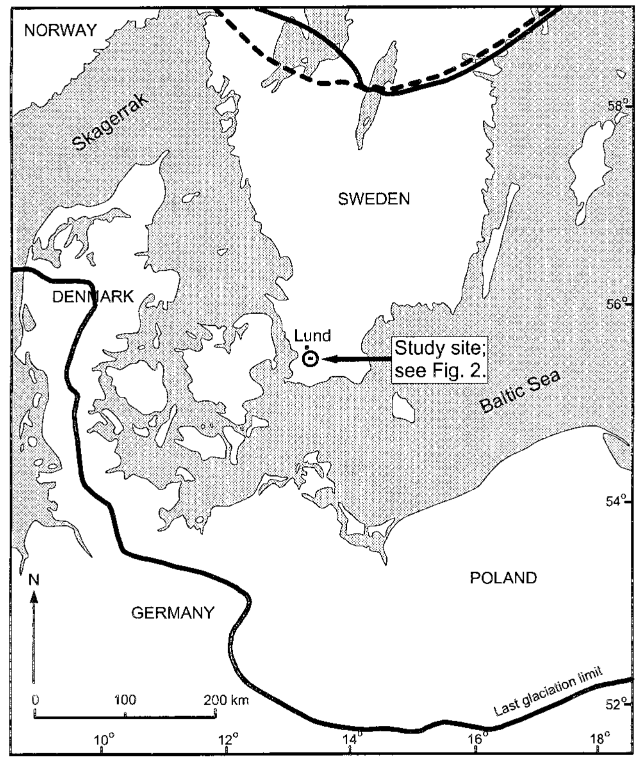

Figure 1 Map of southern Sweden and surrounding areas. The solid line across south central Sweden represents the approximate ice-marginal position at the GI-1–GS-1 transition, and the dashed line refers to the advanced ice margin during GS-1.

Site description

The Torreberga fen (ancient Lake Torreberga; ALT) is situated ca. 10 km south of Lund in southernmost Sweden (55°37′N, 13°14′E; Fig. 1) at an altitude of 7 m a.s.l. The peat deposits (area ca. 2 km^2) are underlain by Late Weichselian and early Holocene lacustrine sediments (Berglund and Digerfeldt, 1970), and the basin is located in an area of hummocky moraine (clayey to silty till), sand and glacial clay (Ringberg, 1980; Ising, 1990). According to Lagerlund and Houmark-Nielsen (1993) the regional retreat of the active ice-margin took place around 14 000 yr BP (17 000–16 500 cal. yr BP), probably followed by the persistence of stagnant ice for some time. The local bedrock, which is composed of Palaeocene limestone, has experienced extensive subsidence along NW–SE directed faults and more than 60 m of Pleistocene deposits have accumulated in the so-called Alnarp Trough that runs through the area, including several units of till and glaciofluvial sediments (Miller, 1977; Ringberg, 1980). The superficial glacial deposits contain both Palaeozoic and Cretaceous–Palaeocene limestone clasts and exhibit bulk carbonate contents in the range of 20–30% (Ringberg, 1980).

Modern mean annual precipitation is ca. 800 mm and the mean annual air temperature is ca. 8°C. January and July mean air temperatures are ca. −1°C and ca. 16°C, respectively.

Material and methods

Fieldwork and subsampling

Cores were obtained from the north-central part of the basin (Fig. 2), close to sampling point MBP 3 of Digerfeldt (1971). Multiple core segments were retrieved with a 1-m-long Russian peat sampler, 10 cm in diameter, and the sediment sequence was described in detail in the field. After correlation in the laboratory, the cores were subsampled into 39 contiguous sections, 30–250 mm thick, taking into account lithostratigraphical boundaries. Three of these sections close to major lithological changes were further separated into upper and lower subsections.

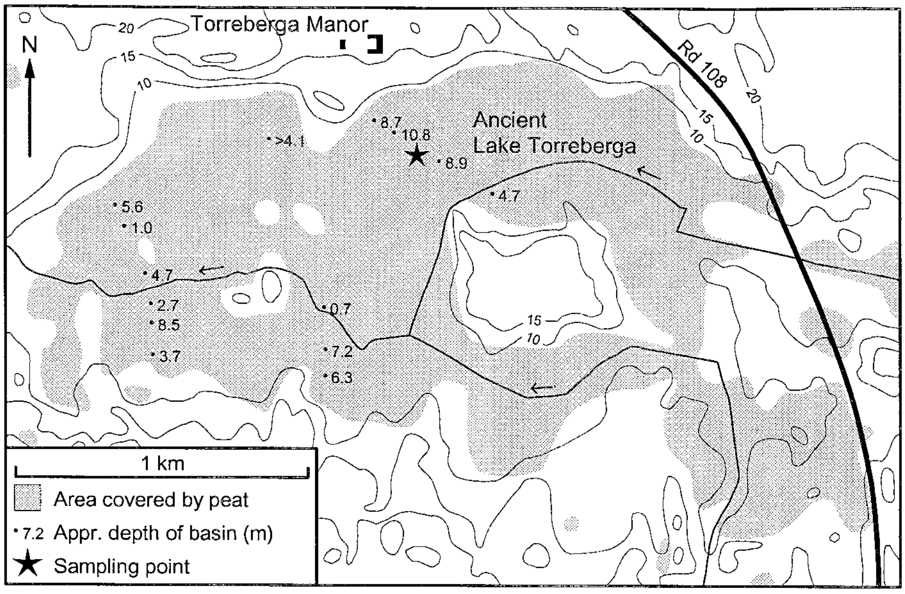

Figure 2 Map of the study area. The distribution of peat was adopted from Ringberg (1980). Approximate depths of the basin refer to total thickness of lacustrine deposits and overlying peat as derived from Berglund and Digerfeldt (1970) and Digerfeldt (1971).

Carbon content

The carbon content of the sediments was determined by temperature-controlled combustion in pure oxygen, with subsequent detection of carbon dioxide by infrared absorption photometry in a Leco RC 412 Multiphase Carbon Determinator. Results are expressed as elemental organic carbon and carbonate carbon contents in percentages of total dry weight, assuming theoretical carbon contents of 12% and ca. 40% for pure calcium carbonate and organic matter, respectively.

Stable isotope analysis

For carbon and oxygen isotope analysis of bulk sedimentary carbonate, small aliquots representing the entire stratigraphical extent of all core sections and subsections were freeze-dried and gently passed through a 125 μm sieve to minimise possible contamination by fragments of molluscs and ostracods, which are effectively retained on this mesh size. Biogenic carbonates, comprising mollusc shell, ostracod valves and macroscopic calcitic encrustations precipitated on Characean algae (*Chara* sp.), were picked from the residue left after washing aliquots of all main core sections through a 200 μm sieve.

$^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios were measured on carbon dioxide evolved by acid dissolution following standard procedures (Buchardt, 1977; McRea, 1950). Analysed samples of carbonate consisted of 20–100 single calcitic valves of adult individuals of *Candona candida* ostracods, 10–50 single calcitic valves of adult individuals of *Cytherissa lacustris*, 4–20 single aragonitic shells of *Pisidium* sp. bivalves, 10–15 calcitic opercula of *Bithynia tentaculata* gastropods, 2–3 mg of *Chara* sp. calcitic encrustations, and 5–30 mg of bulk sediment. Results are expressed as conventional δ-values (per mil deviations from the international PDB standard; Craig, 1957). The analytical reproducibilities for all types of samples are ± 0.03‰ and ± 0.07‰ for $\delta^{13}C$ and $\delta^{18}O$ values, respectively.

Sediment description and age model

The 5.85-m-thick sequence (8.00–2.15 m below the peat surface) was classified into seven lithostratigraphic units (Table 1) partly based on carbon content (see Fig. 4). The lowermost 1.0 m (unit 1) consists of slightly organic clay, grading upward into ca. 0.1 m of clay gyttja containing mollusc shells (unit 2). This unit is overlain by 3.5 m of

Table 1 Lithostratigraphic description

Unit	Depth (m)	Description[a]
	0.00–2.15	Dark brown fen peat, not investigated. Lower boundary gradual
7	2.15–2.23	Dark brown coarse detritus gyttja with mollusc shells. OC 25–42%, CC 0.85–4.8%. Lower boundary very gradual
6	2.23–3.30	Brown to yellowish grey, laminated calcareous gyttja with abundant mollusc shells. OC 2.7–9.4%, CC 7.3–9.8%. Lower boundary rather sharp
5	3.30–3.41	Brownish grey, calcareous clay gyttja. OC 1.3–2.5%, CC 3.6–6.2%. Lower boundary very gradual
4	3.41–5.72	Grey, slightly organic silty clay. OC 0.67–1.2%, CC 2.2–2.9%. Lower boundary rather gradual
3	5.72–6.90	Grey, finely laminated, slightly organic silty clay. OC 0.59–1.1%, CC 2.6–2.8%. Lower boundary rather sharp
2	6.90–6.97	Dark grey clay gyttja with mollusc shells. OC 1.5–2.7%, CC 1.2–2.4%. Lower boundary gradual
1	6.97–8.00	Grey, faintly laminated, slightly organic clay. OC 0.54–0.95%, CC 2.8–3.4%.

[a]OC and CC represent organic and carbonate carbon contents respectively.

slightly organic silty clay (units 3–4), the lower part of which exhibits fine laminations. The clay is overlain, in turn, by ca. 1.2 m of calcareous gyttja with abundant mollusc shells (units 5–6), followed by ca. 0.1 m of coarse detritus gyttja (unit 7) and more than 2 m of peat (not considered in the present study). More detailed descriptions of the stratigraphy of the ALT basin were given by Berglund and Digerfeldt (1970) and Digerfeldt (1971).

An age model based on calibrated radiocarbon ages and inferred calendar ages was constructed for the sediment profile to enable comparison with other climate records. Based on pollen stratigraphy and a set of 13 radiocarbon dates obtained on terrestrial macrofossils (Table 2 and Fig. 3a) within the upper part of the sequence (4.50–2.15 m; Björck *et al.*, 1996) the GS-1–PB transition was placed at 3.41 m, corresponding to the base of the calcareous clay gyttja (unit 5). The age assigned to this transition (11 450–11 390 ± 80 cal. yr BP) by Björck *et al.* (1996) has been revised to 11 500 ± 25 cal. yr BP, based on the recently corrected match between the German oak and pine tree-ring chronologies (Spurk *et al.*, 1998). Calendar ages obtained by matching the radiocarbon dates against this latest tree-ring record are shown in Fig. 3b, excluding the lowermost five dates which lie outside the range of the tree-ring chronology. On the assumption of a constant sedimentation rate above the GS-1–PB transition (ca. 0.9 mm yr^{-1}), the lake became infilled and peat deposition commenced at ca. 10 150 cal. yr BP, corresponding to the upper part of the Preboreal pollen zone (Berglund and Digerfeldt, 1970).

The chronology of the lower part of the sediment sequence is based on lithostratigraphical correlation to nearby cores. Unit 2 can be confidently correlated with a clearly distinguishable and anomalously organic-rich layer (clay gyttja) distributed widely in the basin, as demonstrated by Berglund and Digerfeldt (1970, Fig. 12). By means of detailed pollen stratigraphical studies (Berglund and Digerfeldt, 1970; Ising, 1990), and comparison with the well-established regional pollen-zone stratigraphy for southernmost Sweden (Nilsson, 1935; Berglund, 1966), this unit is correlated to the latest part of GI-1. Thus, the initiation of GS-1 is represented by the shift from clay gyttja (unit 2) to clay (unit 3) at 6.90 m, which can be assigned an age of ca. 12 650 cal. yr BP (Wohlfarth, 1996; Björck *et al.*, 1998), resulting in an average sedimentation rate during GS-1 of ca. 3 mm yr^{-1}. Below the GI-1–GS-1 transition approximate sedimentation rates were inferred from an age model presented by Ising (1990), which is based on pollen stratigraphy and radiocarbon dating of terrestrial macrofossils from a nearby core. These results suggest rapid deposition (ca. 4 mm yr^{-1}) of the clay (unit 1), whereas the clay gyttja (unit 2) was likely deposited at a considerably lower rate. Assuming a sedimentation rate of ca. 1 mm yr^{-1} in unit 2 (similar to the comparable organic-rich sediments of units 5–7) an age of ca. 12 750 cal. yr BP was tentatively assigned to the base of unit 2, whereas the lowermost part of the sequence studied yields an approximate age of 13 000 cal. yr BP.

The age model adopted (Fig. 3b) relies on the assumption of constant sedimentation rates within the respective intervals as defined above, and absence of hiatuses. Although water-level changes were inferred by Berglund and Digerfeldt (1970) based on unconformities and sand layers in the marginal parts of the lacustrine deposits, the sequence presented here from the deepest part of the basin probably represents a continuous Late Weichselian to early Holocene record.

Results and interpretations of the isotope analyses

Records of $\delta^{13}C$ and $\delta^{18}O$ were obtained from the six different carbonate components noted above (Table 3 and Fig.

Table 2 Radiocarbon dates

Sample depth (m)	Laboratory reference	Material analysed	Weight (mg)	$\delta^{13}C$ (‰ PDB)	Reported age (^{14}C yr BP)	Calendar age (yr BP)
2.15–2.20	Ua-4468	Fruits of *Menyanthes trifoliata*	16.2	−26.58	9160 ± 80	10 310 +100/−70
2.20–2.23	Ua-4467	Fruits of *Menyanthes trifoliata*	6.0	−26.70	9305 ± 85	10 540 +150/−230
2.23–2.35	Ua-4466	Fruits of *Menyanthes trifoliata*	14.8	−26.96	9350 ± 90	10 580 +200/−200
2.55–2.65	Ua-4465	Fruits of *Betula alba*, *Menyanthes trifoliata* and *Populus tremula*	4.0	−27.13	9210 ± 105	10 330 +290/−90
2.74–2.80	Ua-4464	Fruits of *Nymphaea alba*	4.9	−25.64	9525 ± 95	10 790 +270/−150
2.90–3.00	Ua-4463	Fruits of *Scirpus palustris*	2.5	−27.78	9805 ± 125	11 280 +90/−280
3.20–3.30	Ua-4462	Fruits of *Betula alba*, *Populus tremula and Scirpus palustris*	6.1	−29.24	9890 ± 85	11 320 +50/−80
3.30–3.41	Ua-4461	Fruits of *Betula nana*, leaves and wood of *Salix herbacea*	3.3	−31.32	9875 ± 110	11 320 +170/−90
3.41–3.60	Ua-4460	Leaves and wood of *Dryas octopetala* and *Betula nana*, undetermined wood	42.6	−29.77	10 145 ± 85	
3.60–3.80	Ua-4458	Leaves and fruits of *Betula* sp., leaves of *Dryas octopetala*	3.0	−30.26	10 310 ± 180	
3.95–4.10	Ua-4457	Leaves of *Betula nana*, *Salix polaris* and *Dryas octopetala*	4.7	−30.63	10 095 ± 90	
4.10–4.25	Ua-4456	Leaves of *Betula nana* and *Dryas octopetala*	3.7	−30.27	10 365 ± 125	
4.25–4.50	Ua-4455	Leaves of *Betula nana*, *Salix polaris* and *Dryas octopetala*	2.1	−31.29	10 510 ± 235	

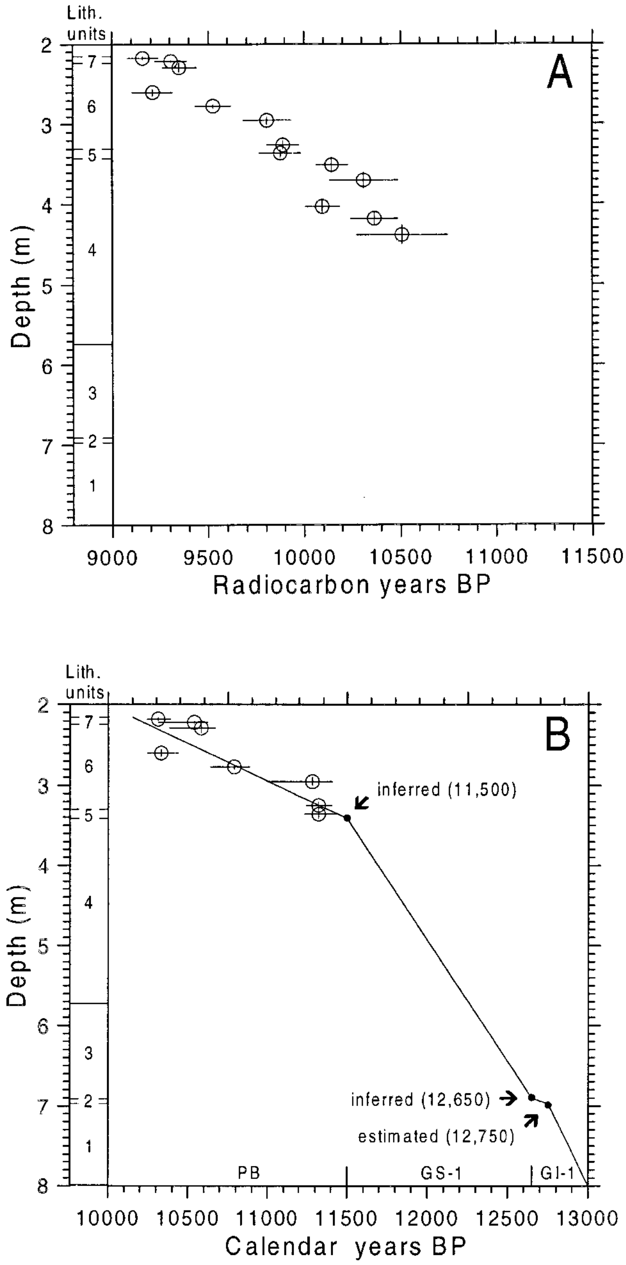

Figure 3 (A) Radiocarbon dates obtained on terrestrial macrofossils from the upper part of the sediment sequence as reported by Björck *et al.* (1996). Horizontal bars represent single standard deviations (see Table 2 for details). (B) Age model of the sediment sequence based on matching of radiocarbon dates younger than 10 000 yr BP against the tree-ring chronology of Spurk *et al.* (1998) and inferred ages of the GI-1–GS-1 and GS-1–PB transitions respectively (see text for details). The base of unit 2 (6.97 m) and the lowermost part of the sequence studied (8.00 m) were tentatively assigned ages of 12 750 and 13 000 cal. yr BP respectively, based on an age model presented for a nearby core by Ising (1990). Constant sedimentation rates were assumed above, between and below the inferred ages.

4). These include 19 samples of *Pisidium* ($\delta_{Pis.}$; 8.00–4.50 m), 19 samples of *Cytherissa lacustris* ($\delta_{Cyt.}$; 7.01–3.30 m), 12 samples of *Candona candida* ($\delta_{Can.}$; 4.50–2.45 m), 10 samples of *Bithynia tentaculata* ($\delta_{Bit.}$; 3.00–2.15 m), 14 samples of *Chara* encrustations ($\delta_{Cha.}$; 3.80–2.23 m), and 36 samples of sedimentary carbonates ($\delta_{Sed.}$; 7.75–2.15 m).

The *Pisidium* records

Bivalves of the genus *Pisidium* are benthic filter-feeders with aragonitic shells. The shell continues to grow throughout the lifetime of the mollusc, with summer being the main season of calcification. The *Pisidium* records span the interval from late GI-1 to mid-GS-1, with both isotope profiles clearly showing evidence of rapid environmental change at the GI-1–GS-1 transition. The record of $\delta^{13}C_{Pis}$ exhibits minimum values immediately following the transition, likely reflecting reduced aquatic productivity in response to the climatic cooling at this stage (e.g. Björck and Möller, 1987; Lemdahl, 1988). Decreased lake productivity is commonly recorded by low $\delta^{13}C$ values in lacustrine carbonates because of diminished selective uptake of ^{12}C from the lake-water DIC (dissolved inorganic carbon) pool by aquatic macrophytes and phytoplankton (McKenzie, 1985). The subsequent increase in $\delta^{13}C_{Pis}$ values may indicate a return to slightly higher levels of productivity during mid-GS-1.

An even more marked oxygen isotope response occurs, with $\delta^{18}O_{Pis}$ values exhibiting an abrupt decrease across the GI-1–GS-1 boundary, following rather stable values during late GI-1. Given well-documented evidence for a substantial decrease in air temperature (e.g. Berglund *et al.*, 1994), this change probably reflects a pronounced decline in lake-water $\delta^{18}O$ inherited from decreased $\delta^{18}O$ of local precipitation. The decrease in $\delta^{18}O_{Pis}$ is partially offset by increased carbonate-water equilibrium fractionation due to reduced summer lake-water temperature. The likely magnitude of the overall temperature sensitivity for $\delta^{18}O_{Pis}$ can be constrained by considering that temperature-dependent shifts in $\delta^{18}O$ of precipitation equivalent to modern spatial isotope-temperature relations of ca. +0.7‰ °C^{-1} (Dansgaard, 1964), if transferred directly to the lake water, would be opposed by carbonate-water equilibrium effects of ca. –0.25‰°C^{-1} (Craig, 1965), resulting in a potential sensitivity in the order of +0.45‰ °C^{-1}. Hence, as a first approximation, if summer water temperatures declined in parallel with air temperatures, then the shift of about –2.5‰ in $\delta^{18}O_{Pis}$ between GI-1 and early GS-1 would be consistent with a drop of more than 5°C in local mean summer temperature.

The *Cytherissa lacustris* records

Cytherissa lacustris is a benthic ostracod with calcitic valves that reproduces and moults throughout the year (Geiger, 1990). This species tolerates bottom water temperatures ranging from 3–23°C (Delorme, 1991). The records of $\delta^{13}C$ and $\delta^{18}O$ for *C. lacustris* largely parallel those for *Pisidium* sp. in the interval of overlap between 7.0 and 4.5 m (latest part of GI-1 to mid-GS-1), suggesting similar isotopic sensitivity of both organisms to environmental change.

Detailed autecological field studies by von Grafenstein *et al.* (in press) have shown that the oxygen isotope composition of *C. lacustris* is influenced by a constant 'vital offset' leading to ^{18}O-enrichment compared with calcite precipitated in isotope equilibrium with the lake water of 1.5 ± 0.2‰. Similarly, the aragonitic shells of *Pisidium* sp. also display a small vital effect, leading to a total offset from calculated equilibrium calcite of 0.9 ± 0.2‰ (von Grafenstein *et al.*, in press), of which ca. 0.6‰ is attributable to aragonite–calcite fractionation (Tarutani *et al.*, 1969).

Closer examination of the data reveals a general tendency for slightly higher $\delta^{18}O_{Cyt}$ values than the corresponding $\delta^{18}O_{Pis}$ values, which is consistent with the expected residual

Table 3 Carbon content and stable carbon and oxygen isotope analysis results[a]

Depth (m)	OC	CC	$\delta^{13}C_{Sed.}$	$\delta^{18}O_{Sed.}$	$\delta^{13}C_{Cha.}$	$\delta^{18}O_{Cha.}$	$\delta^{13}C_{Pis.}$	$\delta^{18}O_{Pis.}$	$\delta^{13}C_{Bit.}$	$\delta^{18}O_{Bit.}$	$\delta^{13}C_{Cyt.}$	$\delta^{18}O_{Cyt.}$	$\delta^{13}C_{Can.}$	$\delta^{18}O_{Can.}$
2.15–2.20	42.0	0.85	−6.28	−8.35					−6.37	−6.44				
2.20–2.23	25.2	4.85	−5.49	−8.73					−6.84	−6.40				
2.23–2.35	9.21	7.09	−4.71	−8.39	−5.71	−7.95			−6.47	−5.96				
2.35–2.45	9.41	6.86	−4.88	−8.34	−5.48	−8.01			−6.27	−5.83				
2.45–2.55	5.30	8.68	−4.90	−8.28	−5.12	−8.38			−6.39	−5.85			−6.43	−2.72
2.55–2.65	6.49	8.53	−4.79	−8.22	−4.89	−8.04			−5.99	−6.16			−6.92	−3.16
2.65–2.74	5.33	8.86	−4.24	−8.26	−4.37	−8.17			−6.36	−5.93				
2.74–2.80	2.66	9.84	−4.24	−8.29	−3.89	−8.37			−6.87	−6.32				
2.80–2.90	4.89	8.91	−4.08	−8.29	−3.81	−8.35			−7.02	−6.58			−6.12	−3.65
2.90–3.00	3.83	9.16	−4.17	−8.22	−3.76	−8.46			−6.53	−6.53			−4.77	−3.05
3.00–3.10	4.31	8.62	−3.74	−8.20	−3.28	−9.05							−6.12	−3.52
3.10–3.20	6.12	7.75	−3.53	−8.32	−4.24	−8.61							$\underline{-5.79}$	$\underline{-3.57}$
3.20–3.30	6.76	8.25	−2.20	−7.78	−1.05	−7.41							$\underline{-5.79}$	$\underline{-3.57}$
3.30–3.39	2.50	6.17	−0.44	−7.32	$\underline{-1.91}$	$\underline{-6.56}$					$\underline{-6.42}$	$\underline{-7.67}$	$\underline{-3.67}$	$\underline{-1.50}$
3.39–3.41	1.28	3.64	+0.50	−7.67	$\underline{-1.91}$	$\underline{-6.56}$					$\underline{-6.42}$	$\underline{-7.67}$	$\underline{-3.67}$	$\underline{-1.50}$
GS-1–PB transition														
3.41–3.43	1.21	2.85	+0.70	−7.48	$\underline{+0.40}$	$\underline{-8.98}$					$\underline{-5.09}$	$\underline{-8.69}$	$\underline{-1.33}$	$\underline{-7.64}$
3.43–3.60	1.16	2.28	+0.88	−5.55	$\underline{+0.40}$	$\underline{-8.98}$					$\underline{-5.09}$	$\underline{-8.69}$	$\underline{-1.33}$	$\underline{-7.64}$
3.60–3.80	0.97	2.50	+1.02	−5.28	+0.38	−9.55					−5.26	−8.33	−2.71	−6.99
3.80–3.95	0.86	2.52	+1.12	−4.16							−4.58	−8.75		
3.95–4.10	0.92	2.56	+1.12	−4.01							−4.70	−8.64	−2.59	−5.33
4.10–4.25	0.96	2.45	+1.00	−4.40							−5.37	−8.85	−4.17	−7.67
4.25–4.50	0.96	2.23	+1.00	−4.81							−4.80	−8.13	−2.75	−5.92
4.50–4.75	0.89	2.24	+0.11	−5.92			−5.39	−8.99			−5.53	−9.31		
4.75–5.00	0.84	2.38					−5.01	−9.31			−5.32	−9.27		
5.00–5.25	0.93	2.35	+0.44	−5.55			−5.02	−9.23			−5.39	−8.76		
5.25–5.50	0.73	2.50					−5.71	−9.45			−4.99	−8.89		
5.50–5.72	0.67	2.65	+1.06	−4.42			−5.36	−9.10			−4.94	−9.49		
5.72–5.92	0.59	2.74					−6.02	−9.90						
5.92–6.12	0.66	2.64	+0.70	−4.43			−5.84	−9.28			−4.76	−8.75		
6.12–6.32	0.64	2.65					−5.27	−10.15			−4.52	−9.54		
6.32–6.52	0.67	2.79	+0.24	−4.47			−5.44	−9.35						
6.52–6.72	0.69	2.80	+1.08	−4.38			−6.81	−9.56						
6.72–6.86	0.74	2.68	+0.17	−4.96			−5.80	−8.81			−6.30	−8.11		
6.86–6.90	1.08	2.65	+0.38	−5.18			−6.71	−8.30			−7.14	−8.30		
GI-1–GS-1 transition														
6.90–6.92	2.07	2.02	−1.37	−6.81			$\underline{-5.24}$	$\underline{-7.02}$			$\underline{-6.50}$	$\underline{-6.76}$		
6.92–6.94	2.72	1.19	−1.75	−6.46			$\underline{-5.24}$	$\underline{-7.02}$			$\underline{-6.50}$	$\underline{-6.76}$		
6.94–6.97	1.54	2.39	−0.16	−5.28			−5.93	−7.16			−6.19	−6.72		
6.97–7.01	0.95	2.82	+0.56	−4.93			−5.34	−6.90			−5.71	−6.46		
7.01–7.25	0.61	2.86	+0.85	−4.56			−5.69	−7.16						
7.25–7.50	0.57	3.07					−5.15	−6.94						
7.50–7.75	0.62	3.24	+1.06	−4.46			−5.09	−6.85						
7.75–8.00	0.54	3.38					−4.53	−7.37						

[a]OC and CC represent organic and carbonate carbon contents (%) respectively. Isotopic units: ‰ (PDB), Sed. = sedimentary carbonates, *Cha.* = *Chara* sp., *Pis.* = *Pisidium* sp., *Bit.* = *Bithynia tentaculata*, *Cyt.* = *Cytherissa lacustris*, *Can.* = *Candona candida*. Underlined values represent measurements on combinations of two consecutive samples.

offset of ca. 0.6‰ observed by von Grafenstein *et al.* (in press), perhaps enhanced by the effects of lower water temperatures, on average, for the longer annual period of *C. lacustris* calcification. More compelling evidence for such seasonality effects, and possibly changes over time, is provided by the carbon isotope data, which reveal a systematic separation in the interval of ca. 6.5–5.5 m, with the lower $\delta^{13}C_{Pis}$ values perhaps indeed reflecting depressed summer productivity during the severe earlier part of GS-1, as speculated above.

Although only a single sample of *C. lacustris* post-dates the GS-1–PB transition, the general trends for both $\delta^{13}C_{Cyt}$ and $\delta^{18}O_{Cyt}$ are consistent with corresponding data from other carbonate components, as discussed below.

The *Candona candida* records

Candona candida is a benthic ostracod with calcitic valves. This species, which tolerates a wide range of water temperatures (0–27°C; Delorme, 1991), exhibits a distinct frequency

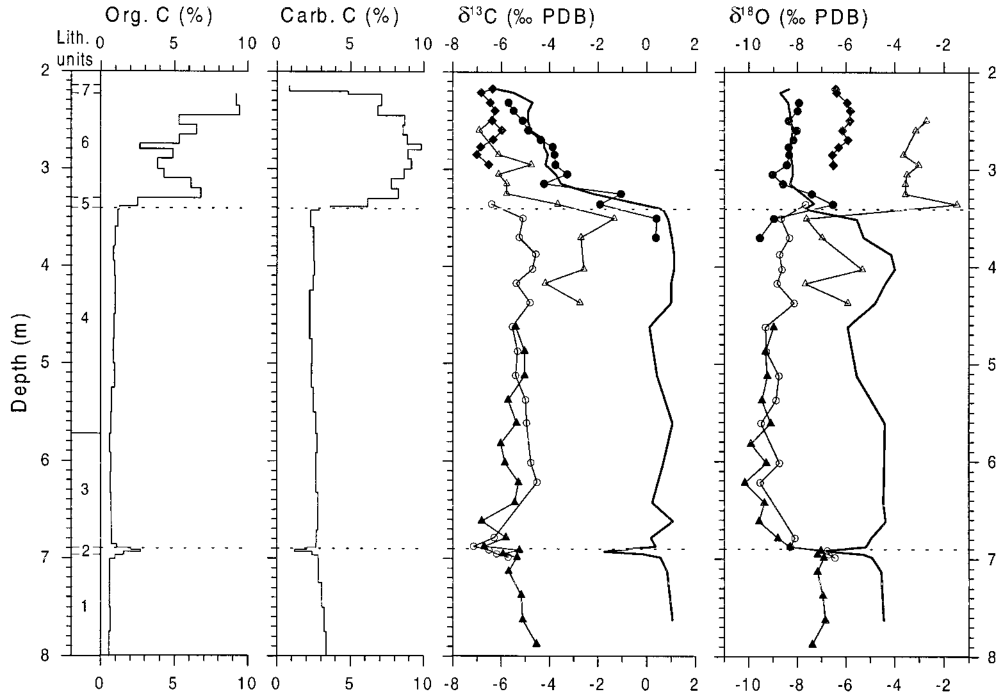

Figure 4 Content of elemental organic and carbonate carbon (expressed as percentages of total dry weight) together with carbon and oxygen isotope records of different carbonate components plotted against sediment depth (Table 3). Filled triangles = *Pisidium* sp. ($\delta_{Pis.}$); open triangles = *Candona candida* ($\delta_{Can.}$); open circles = *Cytherissa lacustris* ($\delta_{Cyt.}$); filled circles = *Chara* sp. ($\delta_{Cha.}$); diamonds = *Bithynia tentaculata* ($\delta_{Bit.}$); solid line = fine-grained sedimentary carbonates ($\delta_{Sed.}$). Numbers in the left-hand column refer to lithostratigraphical units (Table 1). The dashed lines represent important lithological shifts identified as the GI-1–GS-1 and the GS-1–PB transitions respectively.

maximum of adult individuals during the autumn and early winter. Its occurrence in the core includes the later part of GS-1 and the GS-1–PB transition, permitting direct comparison with the upper part of the *C. lacustris* records.

Congruent with the data obtained on other carbonate components, the record of $\delta^{13}C_{Can}$ exhibits a pronounced decreasing trend across the GS-1–PB transition. Similar decreasing trends at this stage have been observed in carbon isotope records of organic material from other sites in southern Sweden (Hammarlund and Keen, 1994; Hammarlund and Lemdahl, 1994) as well as in records of organic material and carbonates from Denmark (Noe-Nygaard, 1995; Hammarlund and Buchardt, 1996). These isotopic shifts evidently reflect a general depletion of ^{13}C in the DIC of lake waters, which can be attributed mainly to the release of ^{13}C-depleted carbon dioxide from terrestrial soils that formed as a result of the climatic warming at the GS-1–PB transition, and from oxidation of sediments with increasing content of organic matter (Hammarlund, 1993, 1994).

A very large positive shift in $\delta^{18}O$ values of *C. candida* carbonate was recorded across the GS-1–PB transition, substantially greater than that recorded for *C. lacustris*. Because relatively constant bottom-water temperatures (ca. 4°C) can be assumed during the autumn and early winter period of *C. candida* calcification, owing to seasonal mixing of the water column, the magnitude of the $\delta^{18}O_{Can}$ shift is probably a close proxy for changes in average lake-water $\delta^{18}O$. This contrasts with $\delta^{18}O_{Cyt}$ (and $\delta^{18}O_{Cha}$; see below) for which sensitivity to changing lake-water $\delta^{18}O$ would be reduced by increased spring and summer water temperatures, exactly analogous to the damping of the isotopic response of *Pisidium* sp. to changing lake-water $\delta^{18}O$ at the GI-1–GS-1 transition. In addition, the much greater increase in $\delta^{18}O_{Can}$ across the GS-1–PB transition compared with either $\delta^{18}O_{Cyt}$ or $\delta^{18}O_{Cha}$ provides clear evidence that this climatic shift caused a substantial increase in seasonal water-temperature variations in ALT. The systematically high $\delta^{18}O$ values of *C. candida* relative to other biogenic carbonate components throughout its record also bears witness to the large vital offset of 2.1 ± 0.3‰ recorded for this ostracod by von Grafenstein *et al.* (in press).

The *Bithynia tentaculata* records

The gastropod *Bithynia tentaculata* is a benthic/epiphytic species feeding on aquatic plants and organic detritus. The calcitic operculum, which is used as a lid during the winter hibernation, exhibits annual growth increments that probably precipitate during the late summer or early autumn prior to hibernation. *Bithynia tentaculata* opercula only occur in the uppermost part of the core, several hundred years after the GS-1–PB transition. The record of $\delta^{13}C_{Bit}$ does not differ significantly from the uppermost values of $\delta^{13}C_{Can}$, which thus suggests that the opercula may have predominantly formed during the autumn.

Calcification of *B. tentaculata* opercula at intermediate autumn temperatures is also supported by the $\delta^{18}O$ values, which fall between those of *Chara* encrustations precipitated in summer (see below) and *C. candida* calcified later in autumn or early winter. Although the possibility of vital effects for *B. tentaculata* cannot be excluded, other gastropods are known to precipitate carbonate in oxygen isotope equilibrium with ambient water (Fritz and Poplawski, 1974). If *B. tentaculata* opercula and *Chara* encrustations (see below) are both produced in equilibrium with lake water, then the systematic $\delta^{18}O$ offset from *Chara* of ca. 2‰ translates into a plausible average summer–autumn water temperature difference during the Preboreal of around 8°C, while the offset from *C. candida* of ca. 3.0‰ suggests an additional late-autumn cooling of 3–4°C, after accounting for the *C. candida* vital effect.

The *Chara* records

Calcitic encrustations commonly form on the stems and leaves of submersed aquatic plants, such as Characean algae, capable of utilising dissolved bicarbonate as a carbon source by means of the 'proton pumping' mechanism (McConnaughey, 1991). Proton pumping in *Chara* sp. seems to be associated with marked kinetic carbon isotope fractionation (Hammarlund *et al.*, 1997), leading to enrichment of ^{13}C in the *Chara* encrustations. Similar to the carbon isotope records obtained on ostracod carbonates, the record of $\delta^{13}C_{Cha}$ also exhibits a substantially decreasing trend across the GS-1–PB transition, presumably reflecting a general depletion of ^{13}C in lake-water DIC. Although the *Chara* encrustations are systematically enriched in ^{13}C by ca. 2‰ with respect to *C. candida*, this is probably related mainly to kinetic effects, rather than seasonal fluctuations in the carbon isotope composition of DIC.

As noted above, a less-pronounced increase occurs for $\delta^{18}O_{Cha}$ than for $\delta^{18}O_{Can}$ across the GS-1–PB transition, which is readily attributable to the influence of temperature-dependent effects on the summer-produced *Chara* encrustations partly counteracting the effect of increased lake-water $\delta^{18}O$ that is faithfully recorded by *C. candida*. Interestingly, both *Chara* and *C. candida* display maximum $\delta^{18}O$ values immediately above the GS-1–PB transition, before stabilising around somewhat lower values. The *Chara* $\delta^{18}O$ record additionally exhibits a local minimum centred at ca. 3.2 m, perhaps reflecting the cool conditions of the short-lived Preboreal Oscillation (discussed further below).

Comparison of the *Chara* encrustration and *C. candida* and *C. lacustris* $\delta^{18}O$ records also permits re-evaluation of the seasonal water-temperature variations during late GS-1 and the Preboreal, as well as providing evidence that weak non-equilibrium oxygen isotope effects probably exist for *Chara* akin to those for carbon isotopes. Although only two samples containing co-existing *Chara* encrustations and *C. candida* valves occur in GS-1, the average $\delta^{18}O_{Cha}$ and $\delta^{18}O_{Can}$ separation is only 2‰, essentially identical to the *C. candida* vital offset of 2.1 ± 0.3‰ documented by von Grafenstein *et al.* (in press). As it is highly unlikely that lake-water temperatures underwent no seasonal changes during late GS-1, it is probable that *Chara* also incorporates a small oxygen-isotope kinetic offset from equilibrium calcite. This is also borne out by the positive $\delta^{18}O_{Cha} - \delta^{18}O_{Cyt}$ difference in the first sample above the GS-1–PB transition, because greater carbonate-water fractionation (induced by lower water temperatures on average during calcification of *C. lacustris*) in combination with the *C. lacustris* vital effect should otherwise lead to a negative offset. Thus, the above estimate of total seasonal lake-water temperature difference of 11–12°C during the Preboreal deduced from the $\delta^{18}O_{Can} - \delta^{18}O_{Cha}$ separation should be considered a minimum, subject to increase according to the magnitude of seasonal lake-water temperature variation during late GS-1, which cannot be quantified adequately with the present data.

The sedimentary carbonate records

The records of $\delta^{18}O$ and $\delta^{13}C$ obtained on fine-grained sedimentary carbonate both reveal rather high and consistent values during late GI-1 and during GS-1, marked by strongly localised minima at the GI-1–GS-1 transition and longer-term minima within mid-GS-1. Within the early Preboreal, both $\delta^{18}O_{sed}$ and $\delta^{13}C_{sed}$ decline, closely following $\delta^{18}O_{Cha}$ and $\delta^{13}C_{Cha}$, respectively. This pattern of variation clearly reflects mixing between one distinct end-member, representing allochthonous local carbonate detritus (marine carbonates of Palaeozoic and Cretaceous–Palaeocene age in the glacial deposits of the catchment) having $\delta^{18}O$ and $\delta^{13}C$, respectively, of about –4‰ and +1‰, and autochthonous carbonate having more depleted ^{18}O and ^{13}C contents. This mixing is evidenced by the combination of general covariance between fluctuations in $\delta^{18}O_{sed}$ and $\delta^{13}C_{sed}$, and the obvious coincidence and consistency of the $\delta^{18}O_{sed}$ and $\delta^{13}C_{sed}$ maxima. The $\delta^{18}O$ value of –4‰ is also consistent with early GI-1 $\delta^{18}O_{sed}$ values of detrital carbonate from Vanstads mosse, an ancient lake situated ca. 40 km east of ALT (Hammarlund and Keen, 1994). The influence of detrital carbonate deposition on lacustrine $\delta^{18}O$ records, as recorded at a site in southeast Denmark, has been discussed by Hammarlund and Buchardt (1996).

Consideration of the various isotopic records reveals the existence of two end-member situations. In the clay deposited during GI-1 (unit 1), both $\delta^{18}O_{sed}$ and $\delta^{13}C_{sed}$ are close to their maxima, indicating minimal autochthonous carbonate contributions, and hence maximum importance of catchment erosion as a sediment source. Conversely, the similarity of the oxygen and carbon isotopic composition of sedimentary carbonate to that of co-existing *Chara* encrustations during the Preboreal suggests maximum contributions from this source of autochthonous material.

Although *Chara* encrustations were not sufficiently well-preserved to permit discrete sampling below 3.8 m depth, disaggregated calcitic encrustations are probably the major contributor of fine-grained autochthonous carbonate throughout the core.

Knowledge of the isotopic composition of the non-detrital component of the sedimentary carbonate for samples having intermediate compositions can be used to estimate the varying allochthonous/autochthonous mixing ratio between these two extremes through simple mass-balance considerations. The proportion of detrital carbonate (DC), which can be used as a proxy for the balance between catchment erosion rate and in-lake production, is given by:

$$\mathrm{DC}\ (\%) = \left[1 - \frac{(\delta^{18}0_{Sed} - \delta^{18}0_{Det})}{(\delta^{18}0_{Cha} - \delta^{18}0_{Det})}\right] \times 100$$

$\delta^{18}O_{det}$ is the oxygen isotope composition of the detrital carbonate component. In the lower part of the core, where $\delta^{18}O_{Cha}$ values are not available, they can be tentatively approximated by $\delta^{18}O$ values obtained on summer-produced *Pisidium* sp., after accounting for the –0.6‰ calcite–aragonite offset (i.e., $\delta^{18}O_{Cha} \approx \delta^{18}O_{Pis} - 0.6$). Within the interval of 3.8–4.5 m interpolation between the lowermost value of $\delta^{18}O_{Cha}$ and the uppermost inferred value was applied. The resulting profile of DC variations is shown in Fig. 5.

Implications for Late Weichselian climatic development in southern Scandinavia

Basic assumptions

The potential of oxygen isotope records obtained on lacustrine carbonates as proxies for past changes in temperature

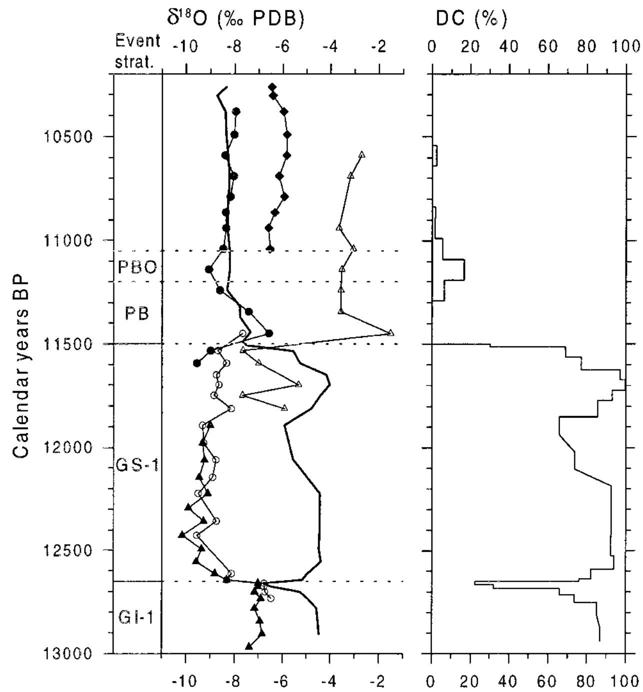

Figure 5 Oxygen isotope records and calculated proportion of detrital versus total carbonate content (DC) of the sediments (see text) plotted against approximate calendar age. Abbreviations in the left-hand column refer to event-stratigraphic units according to Björck *et al.* (1998); GI-1 = Greenland interstadial 1; GS-1 = Greenland stadial 1; PB = Preboreal (early Holocene); PBO = Preboreal oscillation as defined by Björck *et al.* (1996). The age model (Fig. 3b) was derived from a combination of radiocarbon dating and stratigraphical correlation.

may be increased substantially by comparison with isotopic data from carbonates precipitating in the lake studied under present-day climatic conditions. However, at the site studied no modern analogue exists because the ALT basin was infilled during the early Holocene. Furthermore, hydrological information of importance for the interpretation of the isotopic records, such as lake volume and residence time, can be obtained only from available stratigraphic data, catchment topography, and distribution of sediments and peat (Fig. 2). Although the depth (maximum 5–10 m) and volume of ALT were restricted in relation to its area (1–2 km^2), the recharge from the comparatively large catchment (ca. 50 km^2) was probably sufficient to maintain a hydrologically open basin with a short residence time (likely in the range of 0.5–4 months). Non-covariant carbon and oxygen isotope records obtained on the autochthonous carbonate components also point to open-basin conditions (Talbot, 1990). Furthermore, largely parallel $\delta^{18}O$ records and consistent offsets between $\delta^{18}O$ of carbonates representing different seasons of the year suggest that long-term trends are likely related to changes in $\delta^{18}O$ of precipitation rather than by similar variations in lake-water temperature. The seasonal changes in lake-water $\delta^{18}O$ are difficult to estimate and slight variations with time cannot be excluded. However, the amplitude may well have been subdued as a result of extensive input of regional groundwater to the lake and its recharging streams.

The palaeotemperature estimates given below have been obtained by elaborating the well-established relation between water temperature and $\delta^{18}O$ of lacustrine carbonates during equilibrium precipitation (ca. +0.25‰°C^{-1}; Craig, 1965), and applying the modern spatial relation between mean annual air temperature (MAAT) and $\delta^{18}O$ of precipitation in coastal areas (ca. +0.7‰°C^{-1}; Dansgaard, 1964). Only relative temperature changes inferred from oxygen isotope records extending across distinct stratigraphical boundaries are considered, which means that vital offsets can be ignored. Constant ecological preferences in terms of water depth and seasonal growth patterns of the analysed shell-bearing organisms are assumed during the time-span studied.

The GI-1–GS-1 transition

In contrast to several other Late Weichselian lacustrine sequences from the southernmost parts of Sweden (e.g. Liedberg Jönsson, 1988; Hammarlund and Keen, 1994; Hammarlund and Lemdahl, 1994) the GI-1 deposits of ALT are composed mostly of clay, poor in organic material. As pointed out by Berglund and Digerfeldt (1970) this may be due to abundant reworking of clay-rich till surrounding the basin. Although unanimous evidence exists of a gradual climatic cooling during late GI-1 in southern Scandinavia (Berglund *et al.,* 1994; Hammarlund and Keen, 1994), the complacent isotopic records obtained on *Pisidium* within units 1–2 suggest largely constant climatic conditions at this stage. However, the clay gyttja of unit 2 seems to represent a substantial increase in aquatic productivity as reflected by the increased organic carbon content. The substantial decrease in DC values (Fig. 5) also may indicate soil development and decreased erosion rates. A potential enrichment of ^{13}C in DIC and lacustrine carbonates resulting from increased aquatic production may thus have been offset by an increased supply of ^{13}C-depleted carbon dioxide from soil respiration and decomposition of organic detritus (Håkansson, 1985; Hammarlund, 1993, 1994).

The onset of GS-1 involved a pronounced environmental change at ALT. The depletion of ^{18}O in *Pisidium* and *C. lacustris* at the GI-1/GS-1 transition is evidently related mainly to the development of a more negative $\delta^{18}O$ signature of the lake water. Most probably this isotopic shift reflects a general depletion of ^{18}O in precipitation related to a decrease in MAAT. No unambiguous palaeotemperature estimates can be derived due to lowered water temperatures during the summer, which influenced both *Pisidium* and *C. lacustris* carbonates. However, the 2–3‰ decrease in $\delta^{18}O_{Pis.}$ is in close agreement with similar records obtained from shallow-water cores from Lake Ammersee, southern Germany (von Grafenstein *et al.,* 1994) where a decrease in MAAT of ca. 5°C was implied, closely corresponding to the minimum summer-temperature decrease inferred from our data.

GS-1

The distinct cooling at the GI-1–GS-1 transition was succeeded by a continued decrease in MAAT for some time, and a thermal minimum was probably reached during the first 300–400 cal. yr of GS-1. Short and cold summers are indicated by the $\delta^{18}O$ records, consistent with minimum July temperatures (ca. 10°C) inferred from palaeoentomological studies in southern Sweden (Lemdahl, 1988). High DC values indicate extensive soil erosion during the earlier part of GS-1, likely resulting from a very sparse terrestrial vegetation cover (Berglund and Digerfeldt, 1970; Ising, 1990; Berglund *et al.,* 1994). Unstable soils are also indicated by high values of *Cenococcum geophilum* (Liedberg Jönsson, 1988) and humic content (Hammarlund and Keen, 1994; Hammarlund and Lemdahl, 1994) in south Swedish lake sediments from the earlier part of GS-1. In contrast, the aquatic vegetation may have recovered more rapidly as compared with terrestrial ecosystems, as reflected by rising δ^{13}C values shortly after the minimum following the GI-1–GS-1 transition.

During the middle part of GS-1 (5.5–4.5 m) a minimum in DC values coincides with the interval of slightly increased aquatic productivity speculatively inferred from convergence of the $\delta^{13}C_{Pis}$ and $\delta^{13}C_{Cyt}$ records. This may indicate a slight climatic warming (a possible increase in MAAT and longer summers). A more dense terrestrial vegetation cover, as shown by distinctly increasing pollen influx values during mid-GS-1 (Björck and Möller, 1987; Ising, 1990), may also have led to a decrease in soil erosion. During the later part of GS-1 the DC record reaches a second maximum. At this stage, however, the terrestrial vegetation was relatively well developed at ALT and tree-*Betula* was most probably present (Ising, 1990). This means that the extensive detrital input and the deposition of silty clay may reflect an increase in precipitation rather than decreasing temperatures (cf. Liedberg Jönsson, 1988). Although the interpretation of this part of the isotopic records is complicated by changes in species composition, there are no clear indications of an additional climatic cooling or a decrease in aquatic productivity during the later part of GS-1.

The GS-1–PB transition

The increase in $\delta^{18}O_{Can.}$ across the GS-1–PB transition amounts to ca. 6‰, with a difference between average levels above and below the transition of 3.7‰ (Fig. 4). This isotopic shift, which is likely related to an enrichment of ^{18}O in lake water and ultimately precipitation, represents an increase in MAAT of ca. 5.3°C if modern isotope–temperature relations are applied. This estimate, however, should be regarded as a minimum value, taking into account a possible, although probably not very significant, increase in lake-water temperature during the autumn and early winter across the GS-1–PB transition. An increase in water temperature would tend to moderate the isotopic shift, and an even larger increase in MAAT would have to be invoked. The average difference between $\delta^{18}O_{Cha.}$ and $\delta^{18}O_{Can.}$ increases from ca. 2.0‰ to ca. 5.1‰ across the GS-1–PB transition, which could imply an increase in lake-water temperature during the summer of ca. 12°C. This estimate relies on the assumption that the increased isotopic difference is exclusively related to an increase in lake-water temperature during the summer when precipitation of *Chara* encrustations occurs, whereas no major changes in lake-water temperature took place during the growth of adult *C. candida*. However, the record of $\delta^{18}O_{Cha.}$ may have been influenced to some degree by changes in evaporative enrichment of ^{18}O due to slight variations in residence time. An abrupt reduction in the importance of detrital sediment input is reflected in decreasing DC values at the GS-1–PB transition, consistent with the strong increase in limnic productivity indicated by elevated organic content of the sediments. The climatic warming at this stage also led to decreased erosion rates as a result of soil stabilisation by increased terrestrial vegetation cover and permafrost degradation (Björck and Möller, 1987; Berglund *et al.,* 1994).

The Preboreal oscillation

The temporary depletion of ^{18}O in *Chara* encrustations and *C. candida* ostracod valves around 3 m probably reflects a decrease in MAAT succeeding the pronounced climatic warming at the GS-1–PB transition. A slight increase in DC values (Fig. 5), likely reflecting a brief interval of increased soil erosion and reduced productivity (Björck *et al.,* 1996), coincides with the isotopic shifts. On the basis of pollen stratigraphy of the core (Björck *et al.,* 1997) these features can be correlated to the Preboreal oscillation (PBO), a short-lasting cool phase identified in numerous vegetational records from northwest Europe (Behre, 1978; Björck *et al.,* 1997; Hoek, 1997). The subsequently increasing values of $\delta^{18}O_{Cha.}$ and $\delta^{18}O_{Can.}$ are accompanied by a successive increase in $\delta^{18}O_{Bit.}$. Although the magnitude of the PBO is difficult to estimate in terms of a decrease in MAAT due to a possible influence of hydrological change, our data point to a temperature drop of 1–3°C. The higher value is based on the assumption that the ca. 2‰ decrease in $\delta^{18}O_{Cha.}$ and $\delta^{18}O_{Can.}$ recorded from 3.3 to 3.0 m was caused exclusively by a depletion of ^{18}O in precipitation.

The wider perspective

In general, the inferred climatic development at ALT is in good agreement with other proxy records, reflecting changes in ocean and air-mass circulation in the North Atlantic region. The rapidity and extent of the climatic shifts bracketing GS-1 are in accord with both ice-core and marine-sediment records (e.g. Taylor *et al.,* 1993; Haflidason *et al.,* 1995). A thermal minimum during the earlier part of GS-1 followed by a progressive warming throughout the stadial as recorded at ALT has been demonstrated in Greenland on the basis of ice-core δ^{18}O records (Johnsen *et al.,* 1992) and borehole temperature data (Johnsen *et al.,* 1995). Some of the palaeoceanographic data from the eastern margin of the North Atlantic also exhibit similarities with our results. According to Lehman and Keigwin (1992) a severe ocean cooling during the earlier part of GS-1 was followed by slightly warmer conditions during the later part, although other records point to a slight mid-GS-1 warming preceded and followed by cold phases characterised by extensive sea-ice coverage (Koc Karpuz and Jansen, 1992). Furthermore, our data are corroborated by compilations of pollen data from coastal areas of northwest Europe, indicating vegetational changes during the second half of GS-1 that may be related to a slightly more oceanic climate (e.g. Berglund *et al.,* 1994; Hoek, 1997), perhaps associated with a northward shift of the Polar front and increased cyclonic activity.

The regional changes in climate as recorded at ALT are broadly paralleled by the δ^{18}O record from Lake Gosciaz in Poland, which shows evidence of maximum cooling at an early stage of GS-1 (Ralska-Jasiewiczowa *et al.,* 1992). This and other terrestrial climate records from various parts of Europe often exhibit considerable similarities with Late Weichselian ice-core data from Greenland (e.g. Siegenthaler *et al.,* 1984; Goslar *et al.,* 1995; Lowe *et al.,* 1995; von Grafenstein *et al.,* 1998), and as shown by Coope and Lemdahl (1995), the extent of the GS-1 cooling seems to have been fairly uniform across northern Europe. These different lines of evidence suggest a strong ocean–atmosphere coupling during the Last Termination, as demonstrated by Björck *et al.* (1996). Evidently, variations in North Atlantic thermohaline circulation driven by melt-water forcing induced pronounced climatic changes over vast areas of the European continent. This effect may be exemplified by the gradual warming at ALT during the middle and later parts of GS-1, perhaps accompanied by an increase in precipitation, which may be related to a successively increased strength of the North Atlantic conveyor and associated changes in heat flux and sea-ice coverage.

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