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## Multiple oscillations during the Lateglacial as recorded in a multi-proxy, high-resolution record of the Moervaart palaeolake (NW Belgium)



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### ABSTRACT

This paper presents the results of multi-disciplinary research carried out on the deposits of Moervaart depression, NW Belgium, one of the largest palaeolakes (~25 km<sup>2</sup>) that existed during the Lateglacial interstadial in NW Europe. The multi-proxy study, including physical (organic matter and calcium carbonate, magnetic susceptibility, micromorphological), botanical (pollen, macrofossils, diatoms), zoological (ostracods, molluscs, chironomids) and chemical analyses (stable carbon and oxygen isotopes) has resulted in a detailed reconstruction of the Lateglacial landscape as well of the local conditions that prevailed in the lake itself. A chronology of the record was provided by radiocarbon dating and comparison with radiocarbon dates of the nearby Rieme site. These yielded a good match with the regional biostratigraphy.

During the Lateglacial, vegetation and geomorphology of the landscape in general changed from a tundra landscape to a boreal forest. The vegetation development, however, was interrupted by a number of cold reversals. Three centennial-scale cold oscillations are present in the record: 1) the so-called Older Dryas corresponding to GI-1d in the Greenland ice-cores, 2) a short and pronounced cold event during the early Allerød, which could be correlated to GI-1c2 and 3) a cooling event during the late Allerød probably corresponding to the Intra Allerød Cold Period (IACP) or GI-1b. The latter most likely was responsible for the disappearance of the Moervaart palaeolake.

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## 1. Introduction

The Moervaart region is situated in the sandy lowland of NW Belgium at the southern margin of the Great NW European Plain, which is mainly covered by coversands (Fig. 1a–c). It is an important area for Lateglacial and early Holocene geoarchaeological research, as it is one of the few areas in NW Europe in which high-resolution and high-quality palaeoenvironmental data are available in close spatial relationships with prehistoric hunter-gatherer sites (Crombé and Verbruggen, 2002; Crombé et al., 2011).

Partly building on former palaeoecological studies in the area (Verbruggen, 1979; Denys et al., 1990; Verbruggen et al., 1996), the Moervaart region recently has been the subject of a new multi-disciplinary research project that focussed on understanding the geomorphological, hydrological and ecological evolution of the landscape. Detailed investigations were conducted into the palaeotopography (Werbrouck et al., 2011; Crombé et al., 2013), palaeotopography (De Smedt et al., 2011, 2012; Zwertvaegher et al., 2013) and palaeovegetation (Bos et al., 2013; Demiddele et al., 2016) of the area, which were subsequently combined with extensive radiocarbon and OSL-dating (Derese et al., 2010; Crombé et al., 2012, 2014).

As part of this research, a high-quality lacustrine sediment record was retrieved from the centre of the Moervaart depression, a large, Lateglacial palaeolake bordering the southern side of the Great Coversand Ridge of Maldegem-Stekene (Fig. 1d). This paper compiles the results of a multi-proxy research project carried out on the sediments of this palaeolake (Fig. 1d). The aim of this study is to reconstruct the Lateglacial landscape of the surrounding area, as well as the local conditions that prevailed in the lake itself. As the observed variations in former vegetation patterns are likely linked to climate changes (major and minor) occurring during the Lateglacial and Early Holocene, the results of this study will also be compared with the Greenland ice-core data and previously published climatological, palaeoenvironmental and lake-level data from the NW European coversand region.

## 2. Study area

The Moervaart region is characterised by a complex and dynamic palaeolandscape consisting of four major features (Figs. 1d and 2a,b): (1) an extensive coversand ridge, i.e., the Great Coversand Ridge of Maldegem-Stekene (De Moor and Heyse, 1978; Verbruggen et al., 1996; Crombé et al., 2012), including numerous shallow dune-slacks and ponds; (2) a large but shallow freshwater inland palaeolake, i.e. the Moervaart palaeolake, immediately south of the Great Coversand Ridge (Heyse, 1979, 1983); (3) an anastomosing river system consisting of numerous shallow gullies, connected to the palaeolake; and (4) a deep meandering palaeochannel of a river named the Kale (upper course) or Durme (lower course), which currently runs through the palaeolake area from west to southeast where it joins the river Scheldt.

In the Moervaart region, due to the prevailing western and northwestern winds, longitudinal coversand ridges developed during the Weichselian Pleniglacial and Lateglacial (Heyse, 1979). One of the largest coversand ridges within NW Belgium, the Great Coversand Ridge of Maldegem-Stekene, is over 80 km long and locally 3 km wide. It is characterised by a complex microrelief of small dunes and irregularly elongated depressions and dune slacks (Heyse, 1983; Crombé and Verbruggen, 2002; Bos et al., 2013). Its height varies between ca. 5 m and 15 m above present sea level, although its average relative elevation is only 3–4 m. On the top of this massive sand ridge numerous shallow, closed depressions, forming temporary slacks or ponds, were created as a result of local aeolian erosion (so-called blow-outs). Two of these smaller

depressions situated within the Moervaart region (Rieme and Wachtebeke “Heidebos”) were investigated in detail (Bos et al., 2013; Derese et al., 2010). Besides these shallow dune-slacks, directly south of the Great Coversand Ridge of Maldegem-Stekene, a great number of larger lakes formed during the Lateglacial in depressions when the previously open, northern exit route for surface waters became blocked by the formation of the Great Coversand Ridge (De Moor and Heyse, 1978; Verbruggen et al., 1996). The Moervaart palaeolake was by far the largest within northwestern Belgium (length: ~15 km, width: ~2.5 km, 25 km<sup>2</sup>, Fig. 1d), with the deepest part located in the centre along the northern lakeside (De Moor and Heyse, 1978; Verbruggen, 1971, 2005; Crombé et al., 2013).

Numerous prehistoric sites dating from the Lateglacial (Final Palaeolithic, ca. 14,000–13,000 cal. BP) have been discovered in the Moervaart region concentrated along the lake edge on the southern steep slope of this dune complex (Fig. 2a) (Crombé et al., 2011). Most typical for the Lateglacial is the high density of *Federmesser* Culture sites associated with the Allerød period (Crombé and Verbruggen, 2002; Crombé et al., 2011), forming an almost continuous site-complex of temporary campsites stretching along the Maldegem-Stekene coversand ridge and on some small interfluvial levees (over ca. 15 km, Fig. 2a).

## 3. Material and methods

### 3.1. Fieldwork, sedimentological and physical analyses

An extensive fieldwork survey (Fig. 2b, Bats et al., 2009, 2010; Crombé et al., 2013) allowed reconstruction of the general morphology and succession of the palaeolake sediments in the Moervaart depression. For a more detail study a 70 m long north-south oriented trench was dug across the deepest part (i.e., Moerbeke *Suikerfabriek*, Fig. 2b; 51°10'08"/3°55'43") in which palaeoecological sampling was conducted at three different sites, P3/M3/MB8 (Fig. 3), S2 and S4 (Fig. 4). The most complete and best preserved P3/M3/MB8 master sequence (Fig. 3) was collected at the deepest point. The M3 sequence was used for macroscopic (botanical macroremains and *Bithynia* molluscs) analyses, AMS <sup>14</sup>C radiocarbon dating and oxygen and carbon isotope measurements on ostracod valves; the P3 sequence for microscopic analyses (pollen, chironomids and diatoms); and the MB8 sequence for magnetic susceptibility profiling. Sampling for micromorphological analysis was performed on the sediments directly north of the master sequence (Fig. 3). Two complementary sequences, S2 and S4, retrieved respectively 12 m and 14.5 m north of P3/M3/MB8 (Fig. 1 in Supplement), were sampled for more extensive ostracod and mollusc analyses. These locations were correlated to the master sequence by visually following the layers.

The different layers in the Moervaart trench were visually described according to the following parameters: (1) lithology (colour and grain-size determination), (2) sedimentological features and post-depositional deformation, (3) thickness and bed form and (4) character of lower and upper boundaries of sediment beds (transitional, sharp or erosional) and presence of internal erosional surfaces (Fig. 3, Table 1 and Fig. 2 in Supplement). To aid lithological description, volumetric magnetic susceptibility ( $\kappa$ ) was measured on the MB8 sequence at 2.5 mm intervals with a Bartington MS2 susceptibility meter (Dearing, 1999). Furthermore, organic carbon and CaCO<sub>3</sub> contents were determined on samples with 1 cm resolution by automated loss-on-ignition (LOI) using a PrepASH 229 Precisa following Heiri et al. (2001).

For micromorphological research, samples of the different layers (Fig. 3) were collected in Kubiëna boxes (6 × 9 cm). Samples were oven-dried (50–60 °C), impregnated under vacuum with an

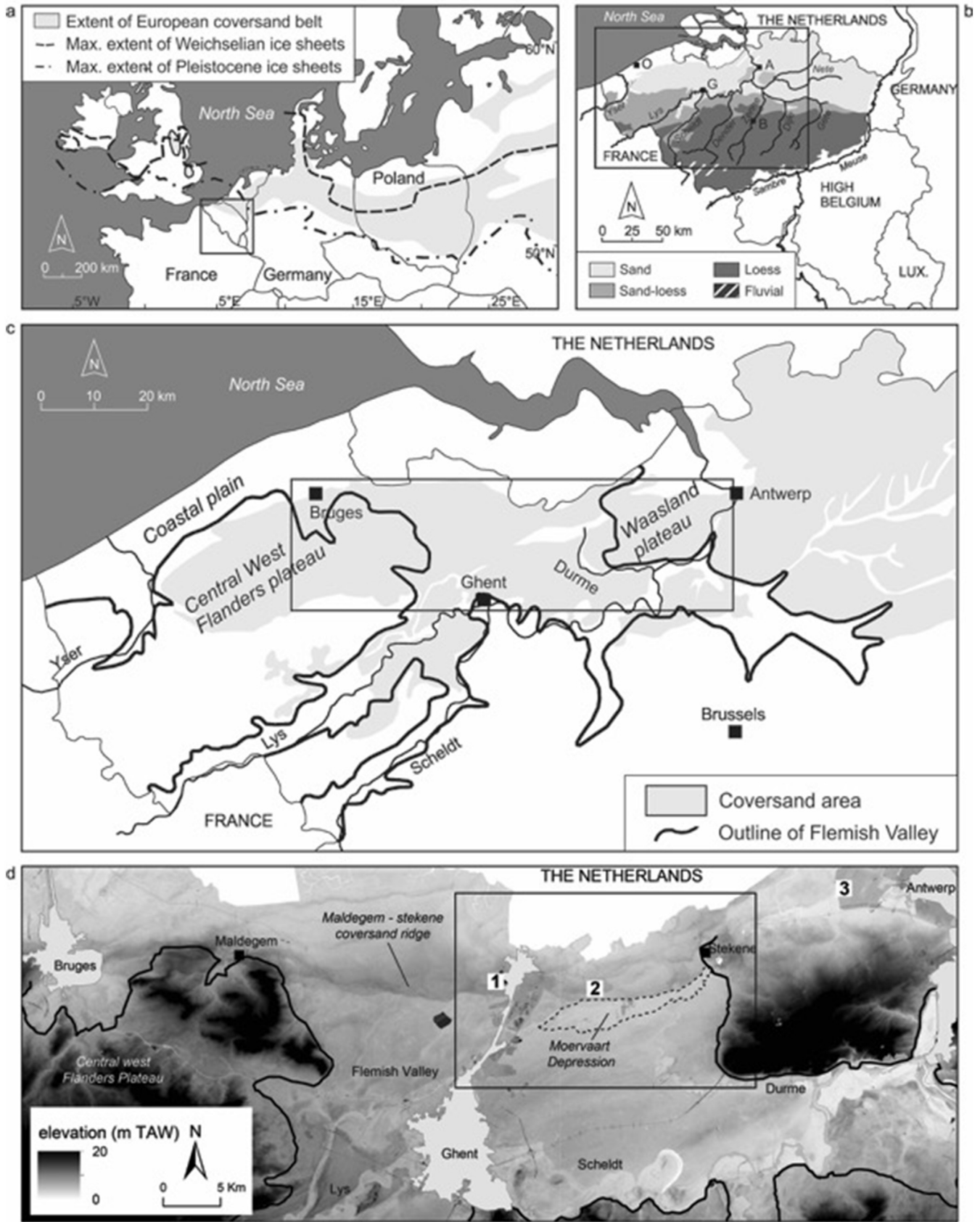
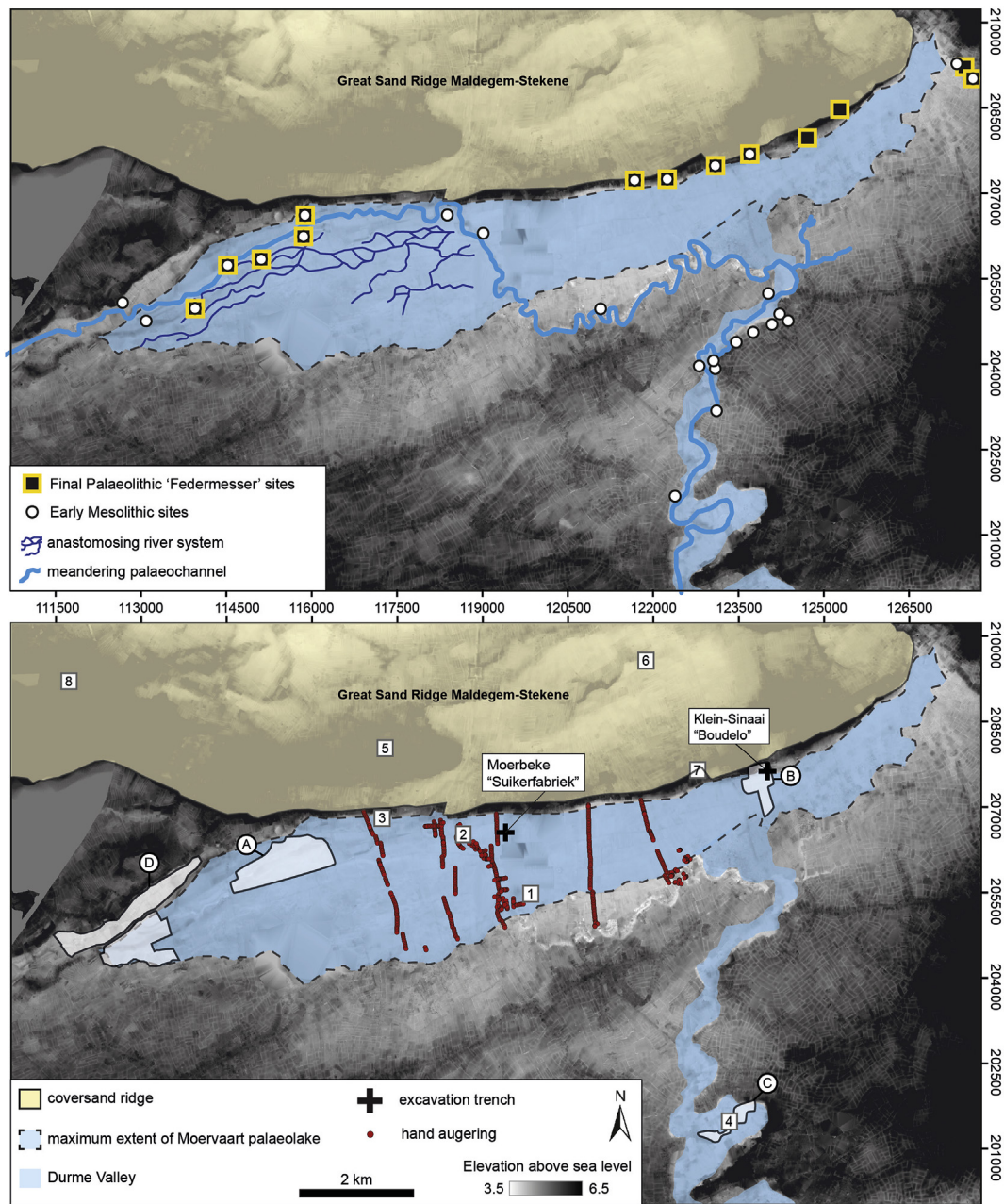


Fig. 1. a. Extent of the European coversand belt. The location of Belgium is indicated with an open square. b. Schematic map of Belgium, showing some of the major rivers and the Pleistocene sedimentation areas in N Belgium. G = Ghent, A = Antwerp, B=Brussels and O=Ostend. c. Extent of the coversand area and major geomorphological units in NW Belgium. The Sandy Flanders study-area is indicated with an open square. d. Simplified geomorphological map of Sandy Flanders (after Derese et al., 2010; Crombé et al., 2011). Nr. 1. Rieme, 2. Wachtebeke "Heidebos", 3. Verrebroek "Dok 2". The Moervaart-study area is indicated with an open square, see also Fig. 2. Figure after Crombé et al., 2013.





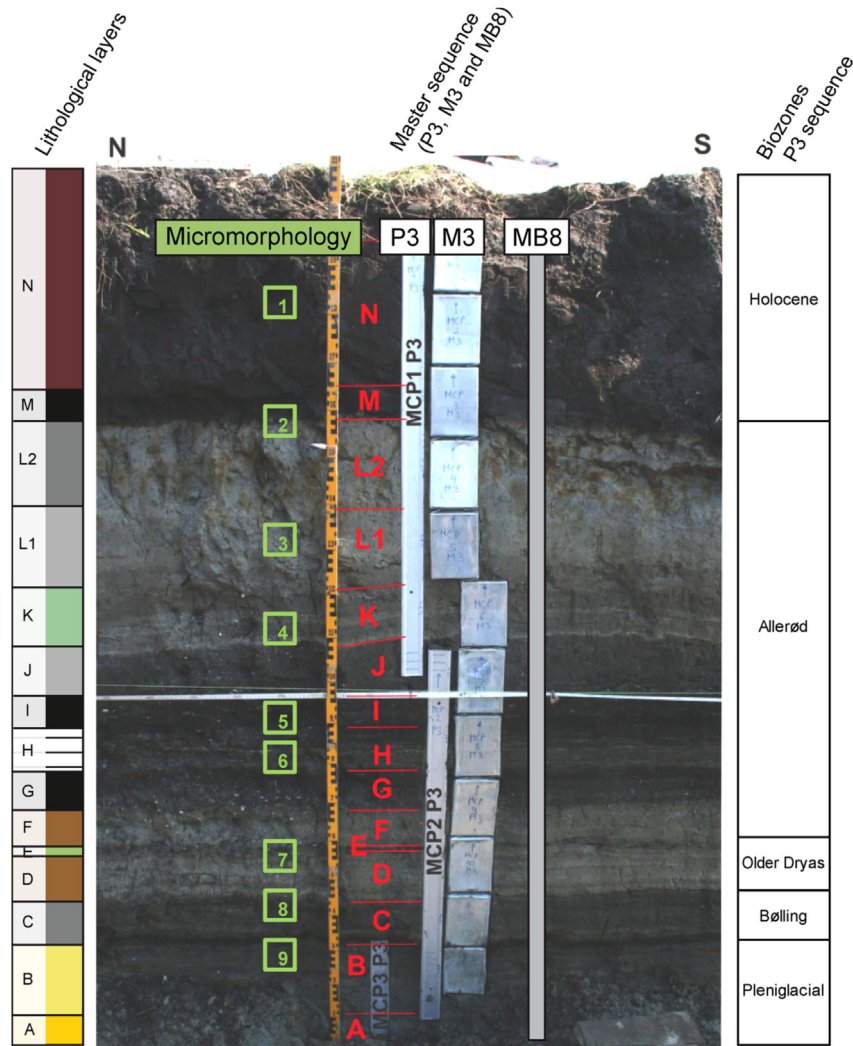
**Fig. 2.** Topography of the Moervaart study area with the maximum extent of the Moervaart depression on the southern edge of the Great Sand Ridge of Maldegem-Stekene. a. Reconstructed course of the anastomosing and meandering palaeochannels in the Moervaart depression, based on data from corings, EMI and DEM (Lidar, AGIV, 2001–2004). Also the distribution of *Federmesser* sites and Early Mesolithic sites along the southern edge of the Great Sand Ridge of Maldegem-Stekene and course of the meandering channel are indicated. b. The different areas where detailed EMI-surveys were conducted are shown (A–D) along with the locations of the mechanical coring campaigns (1–7) and test-pitting (8). In the central part of the palaeolake, manual auger transects are shown along with the location of the two trenches. Figure after Crombé et al., 2013.

unsaturated polyester resin and dried. Subsequently thin sections (25–30  $\mu\text{m}$  thick) were prepared, studied with a polarising microscope and described following Stoops (2003).

### 3.2. Chronology

The chronology of the Moervaart sequence was mainly based on correlation of the regional pollen zones defined in the record with the Lateglacial pollen zonation scheme for the Netherlands and northern Belgium (Hoek, 1997a,c). This was further supported by a number of radiocarbon dates from the record obtained on samples,

only reflecting atmospheric  $^{14}\text{C}$  concentrations. Due to the absence of sufficient datable 'terrestrial' material however only four samples could be selected for dating (Table 1). All  $^{14}\text{C}$  samples were pre-treated at the Royal Institute for Cultural Heritage, Brussels (Belgium) following Van Strydonck and Van der Borg (1990/91). AMS  $^{14}\text{C}$  measurements were performed at the Leibniz-Labor für Altersbestimmung und Isotopenforschung, Kiel (Germany). Radiocarbon dates were converted to calendar years using CALIB 7.1 and INTCAL13 calibration curve (Stuiver and Reimer, 1993; Reimer et al., 2013). Radiocarbon ages are expressed in  $^{14}\text{C}$  years BP, calibrated ages are reported in yrs cal BP.



**Fig. 3.** Detail of the sampled layers collected in the Moervaart master sequence composed of P3, M3 and MB8 and of the micromorphological sampling (green boxes). The biozones based on the pollen data are also indicated.

### 3.3. Proxy analyses

#### 3.3.1. Botanical analyses (pollen & spores, NPP's, macrofossils, diatoms)

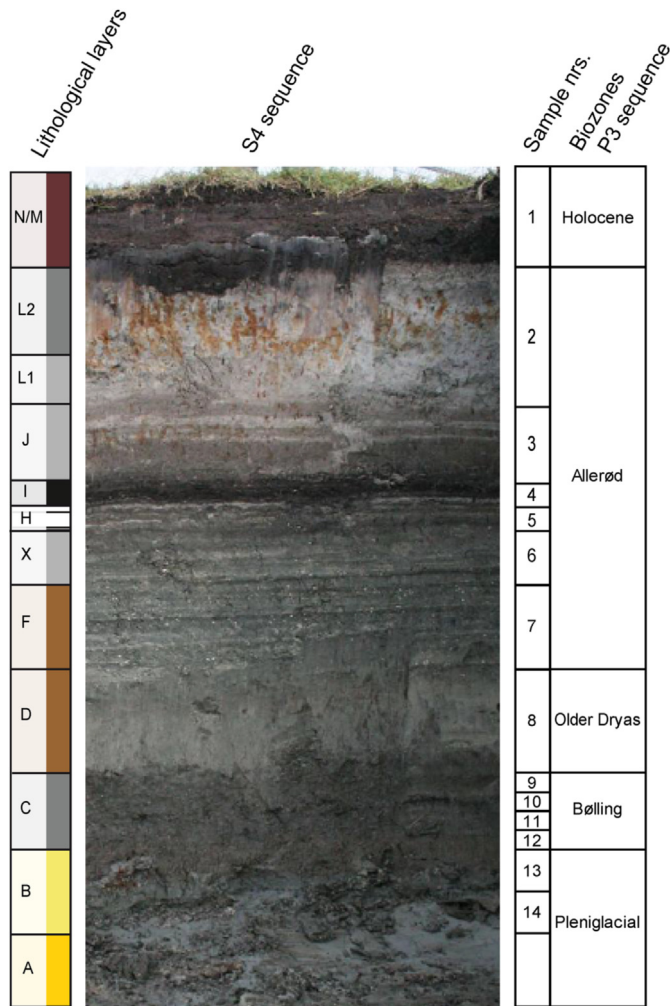
Microfossil samples were collected at 1–3 cm intervals from the P3 sequence (Fig. 3). Only in the sandy, lower part larger (4–5 cm) intervals were used. Samples (1.5–8.2 cc) were prepared following Fægri and Iversen (1989) and Moore et al. (1991) with additional treatment with warm (80 °C) 40% HF and sieving over 150 µm. Residues were mounted in glycerine jelly and sealed. A light microscope (magnification 400x and 1000x) was used for analysis. Pollen and spore types were identified by comparison to modern reference material and identification keys of Fægri and Iversen (1989), Moore et al. (1991), Reille (1992) and Beug (2004). Non-pollen palynomorphs (NPP) were identified using Van Geel (2001) and Van Geel et al. (1981, 1989, 2003). Microfossil taxa were divided into regional and (extra-)local components following Janssen (1973, 1981). Combined AP and NAP totals were employed for percentage calculations. The pollen sum (ca. 300–1000) includes trees, shrubs, Ericales, upland herbs and Poaceae and is directly comparable with the pollen sum of the Dutch and northern Belgium Lateglacial regional pollen zonation scheme (Hoek, 1997a, b; 2001). Pollen and spores of the local aquatic- or marsh

vegetation (including Cyperaceae) and exotic trees (e.g., *Carya*, *Cedrus*, *Picea* and *Abies*) were excluded. The latter were excluded because they originate from older pre-Lateglacial deposits (i.e., Eemian or older).

Plant macroremains were analysed for  $^{14}\text{C}$  samples and collected from the M3 sequence (Fig. 3). Macrofossils were recovered by washing the subsamples over a 125 µm mesh sieve, handpicked from the residue and stored at 4 °C in Eppendorf vials. A Leitz dissecting microscope (magnification of 8–40x) was used for isolation and identification. Plant macrofossils were identified by comparison with modern reference material and identification keys of Berggren (1969, 1981), Anderberg (1994) and Cappiers et al. (2006).

The P3 sequence was furthermore subsampled for quantitative diatom analysis at 2 cm intervals. Samples (wet weight 1 gr.) were prepared following the standard diatom extraction technique (Battarbee et al., 2001). An Optika B 500 Tpl. light microscope (magnification 1000x) was used for analysis. Identifications were made using the identification guide of Hofmann et al. (2013). Diatoms were abundant and well-preserved, however, long pinnate diatoms were often broken and fragmented particularly in the upper layers. Here the counting procedure for broken valves (Schradler and Gersonde, 1978) was followed.





**Fig. 4.** Detail of the location of the mollusc and ostracod samples (sample numbers 1–14) collected from the stratigraphic units at the Moervaart S4 sequence and their reference to the lithological Layers A–N in the P3/M3/MB8 sequence. The biozones based on the pollen data are also indicated. The bottom units 13 and 14 comprise the muddy horizontal trench bottom.

### 3.3.2. Zoological analyses (chironomids, molluscs and ostracods)

The Moervaart P3 sequence (Fig. 3) was analysed for fossil remains of chironomid larvae at 4–8 cm resolution. Samples (0.5–2 cm<sup>3</sup>) were added to 5% KOH, heated to 90 °C for 1 h and passed through 212 and 90 µm mesh sieves. Chironomid head

capsules were manually picked from the residues under a dissection microscope (40x magnification), subsequently mounted in Euparal® and identified under a compound microscope (magnification 400x) using keys of Rieradevall and Brooks (2001) and Brooks et al. (2007). Head capsules were abundant and well preserved with the exception of the basal sands and the upper part of the lake marl (Layer L) in which oxidation traces are visible. Most chironomid taxa in the Moervaart P3 sequence are commonly found in the littoral of relatively warm meso- to eutrophic lakes (Brooks et al., 2007). A chironomid-temperature inference model was used to reconstruct mean July air temperatures. This model is based on a modern calibration dataset consisting of 274 lakes from Norway and the Alps and spans a July air temperature range from 3.5 to 18.4 °C (Brooks and Birks, 2001; Heiri and Lotter, 2010; Heiri et al., 2011). A two component weighted averaging partial least-squares (WA-PLS) regression was used for the model, which, after outlier deletion, had a bootstrapped root mean square error of prediction of 1.40 °C (Heiri et al., 2011). Bootstrapping was used to estimate the sample-specific errors for the fossil samples.

Mollusc and ostracod analysis was carried out on the S4 sequence (Fig. 4, Serbruyns, 2010; Gobert, 2012). Sampling was done continuously and each bulk sample matches one specific stratigraphic unit, therefore sample thickness varies with the thickness of the stratigraphic unit (Fig. 4). Subsamples of 500 ml were collected and air-dried. Once dried, one quarter was taken for malacological analysis. Subsequently, samples were put into a plastic cup, covered with water and gently stirred in order to disintegrate the material. Floating mollusc shells were poured off and left to dry. The residue was passed through 1 mm, 0.5 mm and 150 µm mesh sieves. Organic-rich sediments were first dissolved in Na<sub>4</sub>P<sub>2</sub>O<sub>7</sub> solution before sieving. The mollusc shells were manually picked out of the residues. After mollusc collection, ostracods were handpicked from the contained residues (Gobert, 2012). A Zeiss Stereo Discovery V8 dissection microscope (magnification 60x, 10x) was used for mollusc and ostracod analysis. Identifications were based on comparison to modern reference material and on the standard works for molluscs (Adam, 1960; Gittenberger and Janssen, 2004; Anderson, 2005) and ostracods (Meisch, 2000; Frenzel et al., 2010). Ostracod valves were also identified with the help of a Scanning Electron Microscope (SEM) (Gobert, 2012).

All identifiable mollusc shells and opercula were counted. As some samples contained either shells or opercula, both shells and opercula of *Bithynia tentaculata* were grouped together. For the *Pisidium* species, each valve was seen as an individual. In addition, *Bithynia* shells and opercula were analysed in more detail (1-cm resolution) from the M3 sequence.

**Table 1**

AMS <sup>14</sup>C ages of the dated macrofossil samples of the Moervaart master sequence. KIA: Leibniz Labor, Kiel, Germany. Radiocarbon dates were converted into calendar years by using the CALIB 7.1 computer program and the INTCAL13 calibration curve using the 1 (68.3%) and 2 (95.4%) sigma interval (Stuiver and Reimer, 1993; Reimer et al., 2013).

Depth (cm)	Lab. Nr.	Moervaart sample Nr., Weight	Dated material f. = fruits s. = seeds	δ13C (‰)	Age ( <sup>14</sup> C years BP)	Age (calendar years BP)
186	KIA-46064	1: 1.65 mg C	<i>Schoenoplectus lacustris</i> f. 42×; <i>Carex cf. oederi</i> f. 1×	-25,88 ± 0,13	12,205 ± 60	14,180–14,000 (68.3%) 14,320–13,850 (95.4%)
173	KIA-46065	2: 1.69 mg C	<i>Schoenoplectus lacustris</i> f. 9,25×; <i>Carex cf. rostrata</i> f. 12,3×; <i>Lycopus euopaeus</i> f. 1×; <i>Potentilla palustris</i> f. 2,5×	-26,19 ± 0,12	12,240 ± 60	14,240–14,040 (68.3%) 14,500–13,950 (95.4%)
139,5	KIA-46066	3: 1.28 mg C	<i>Schoenoplectus lacustris</i> f. 13×; <i>Carex cf. pseudocyperus</i> f. 1×; <i>Carex cf. rostrata</i> f. 3×; <i>Carex sp.</i> f. 1×	-19,78 ± 0,10	11,965 ± 60	13,940–13,730 (68.3%) 14,020–13,600 (95.4%)
126,5	KIA-46067	4: 1.03 mg C	<i>Schoenoplectus lacustris</i> f. 14×; <i>Eleocharis palustris</i> f. 1×; <i>Carex cf. rostrata</i> f. 12×	-30,49 ± 0,08	11,630 ± 60	13,550–13,400 (68.3%) 13,580–13,320 (95.4%)

All identifiable adult ostracod valves and carapaces were counted. Valves and complete carapaces of ostracods were counted separately so that each carapace was counted as two valves. Juveniles and heavily fragmented or damaged valves were not included in the counts. Ostracod counts however, were relatively low and variable (Gobert, 2012).

### 3.3.3. Stable carbon and oxygen isotopes

Valves of *Candona candida* recovered from the M3 sequence were used for stable carbon ( $^{13}\text{C}$ ) and oxygen ( $^{18}\text{O}$ ) isotope analysis of ostracod calcite. Valves of *C. candida* were transferred to 100% EtOH, manually cleaned, rinsed thoroughly with 100% EtOH and air-dried on a nitex filter. Subsequently, the valves were transferred to airtight glass vials and reacted with 100% phosphoric acid. The evolved  $\text{CO}_2$  was cryofocussed and -trapped for coupled  $^{18}\text{O}/^{16}\text{O}$  and  $^{13}\text{C}/^{12}\text{C}$  analysis on an isotope ratio mass spectrometer (20–20, Sercon, UK) interfaced with a trace gas unit (ANCA-TGII, Sercon, UK). Replica samples for stable isotope analyses were made for six intervals with a sufficient number of recovered valves.

## 4. Results and discussion

### 4.1. Lithology of the Moervaart sequence

The basal sediments of the Moervaart sequence were formed of grey sands (Layer A) (Fig. 3 and Fig. 2 in Supplement). On top of these sands, sandy gyttja deposits (Layer B) were present. These were followed by a 0.25 m thick stratified layer (Layer C) of humic calcareous gyttja with intercalated thin peaty lenses. The calcareous gyttja Layer B was covered by two layers of laminated lake marl (Layers D and F), separated by a thin humic to peaty marl horizon (Layer E). The layers of lake marl were subsequently covered by two layers of (very) humic calcareous gyttja (Layers G and I). In between these two, a layer of black humic sand interbedded with white sand was present (Layer H). The thickness of this layer increased towards the eastern end of the palaeolake, which was shown in the Boudelo trench more to the east (Fig. 2b). The gyttja Layer I was followed by lake marl deposits (Layers J–L), varying from very humic (Layer J) through clayey (Layer K) to finely laminated with, in the upper part, irregularly oxidised patches (Layer L). Layer L consists of light grey lake marl and can be subdivided into two, Layer L1 and L2, with oxidation features in Layer L2. The lake sediments at the top were covered by a thin layer of peat (Layer M). The discontinuous position of Layer M on top of the lake marl deposits suggests that a hiatus is present in between these layers L and M. Farther to the east, as was shown in the Boudelo trench (Fig. 2b), the peat deposits were ca. 0.5 m thicker and better preserved, thanks to the deposition of overlying Medieval sediments (13<sup>th</sup>–14<sup>th</sup> century). Elsewhere in the Moervaart depression these upper peat deposits have disappeared almost entirely due to extraction and subsequent ploughing (Jongepier et al., 2011).

As observed in the trench, most layer boundaries are normal bedding contacts (transitional to sharp), but five contacts are clearly erosional surfaces: the contacts C/D, D/E G/H, and L2/M. The contact J/I shows in the south of the trench a gully like sandy infill, with many gastropods, 5–12 cm deep and a sharp 7 cm deep V-form wedge at the bottom.

The magnetic susceptibility showed an intrinsically diamagnetic environment dominated by quartz sands and carbonate sediments (marl). Five magnetically distinct units can be observed (Fig. 5): (1) In the basal deposits and the overlying sandy sediments, Layers A–B, display the highest  $\kappa$ . (2) The subsequent series of calcareous deposits (Layers C–I) behave entirely diamagnetically, with the

exception of layers D and G, where a slight  $\kappa$ -increase is measured. (3) Layers J–L contrast sharply with these preceding carbonate deposits with a stronger presence of para- and ferrimagnetic minerals in a third  $\kappa$ -unit. (4) A strong decline in  $\kappa$  marks the peat of Layer M, (5) overlain by the magnetically enriched plough layer (Layer N).

### 4.2. Dating of the Moervaart sequence

Table 2 demonstrates that the four  $^{14}\text{C}$  AMS dates from the Moervaart master sequence concur with the expected age of the regional biostratigraphy by Hoek (1997a, 2001), as well as with the radiocarbon evidence from nearby dune-slacks on the Great Coversand Ridge, e.g. at Rieme. The apparent age reversal of the two lowest dates (KIA-46064 and KIA-46065) is not necessarily problematic; calibrated both dates have an important chronological overlap, which implies that it is statistically not excluded that the uppermost date is older than the lowest one. This correlation with the Lateglacial pollen zonation scheme indicates that the sediments cover the late Pleniglacial to *Pinus*-phase of the Allerød. It further suggests that the Moervaart regional pollen zones can be correlated with the established major climatic and environmental events as recorded in the terrestrial and marine records of the North Atlantic region and in the Greenland oxygen isotope records, particularly GS-2 to GI-1 (Björck et al., 1998; Blockley et al., 2012; Lowe et al., 2001, 2008; North Greenland Ice-Core Project members, 2004; Rasmussen et al., 2014).

### 4.3. Reconstruction of the lake environment in relation to climate

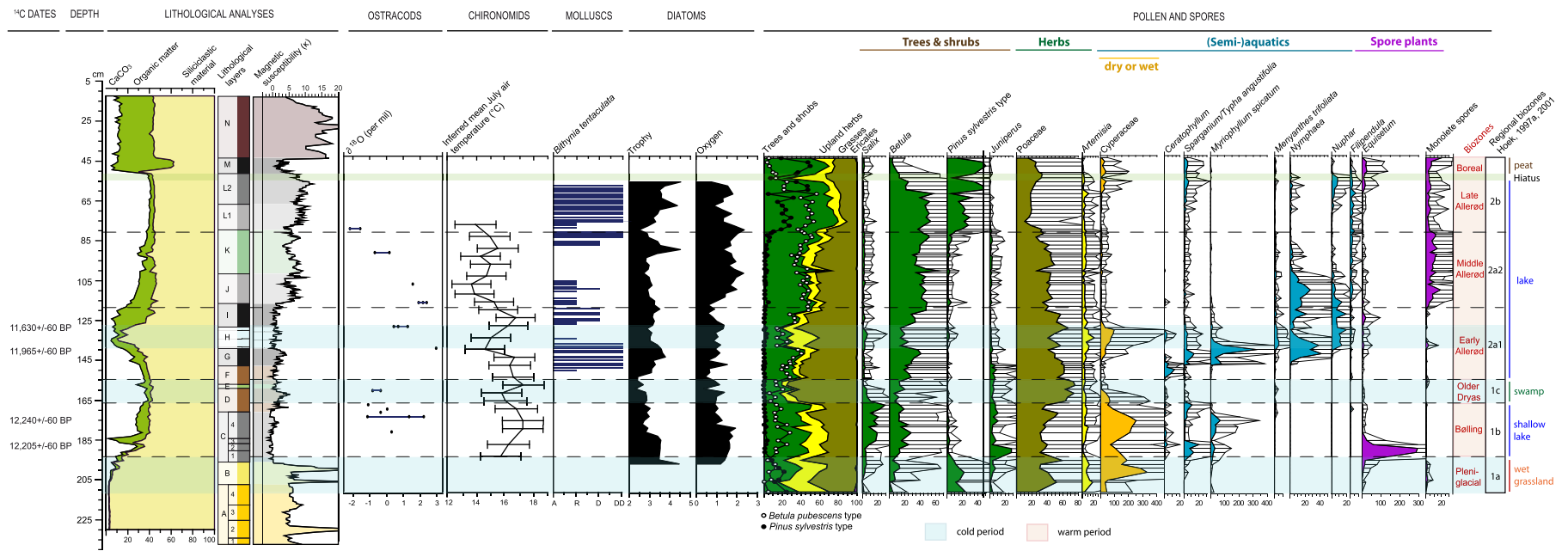
#### 4.3.1. Proxy climate data

From all the different proxies analysed from the Moervaart sequences a range of climatological information could be inferred. From the pollen and macrofossils relative minimum mean July temperatures can be deduced, while the chironomids provide us with data on mean July air temperatures.

The ostracod record can provide a range for mean July and January water temperatures (MOTR approach, see Horne, 2007), although their ranges are rather large (Gobert, 2012). The molluscs can also provide information about drought and relative temperatures (cold-warm). For example, *Acroloxus lacustris*, *Bithynia tentaculata* and *Lymnaea stagnalis* are warm water indicators, while *Pisidium obtusale* and *Valvata piscinalis* prefer cold water (Gittenberger and Janssen, 2004).

The  $\delta^{18}\text{O}$  data, measured on the biogenic carbonate shells of ostracods (i.e., *Candona candida*), can be compared with the oxygen isotope records in the Greenland ice-cores. In addition they also reflect changes in atmospheric circulation, temperature and hydrological change (e.g. Holmes, 1996). Furthermore, the combination of  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  analysis offers insights into hydrological changes (Fig. 4 in Supplement). Relatively large offsets between replicate measurements may suggest hydrologically rather unstable environments and/or high shifts in temperature or trophic state, with consequent large variability in the isotopic composition of valves.

A change to colder climate conditions can also be reflected in evidence for erosional activity in the stratigraphic record. Proof for erosional processes were observed visually in the trench (Table 1 in Supplement). Evidence for periglacial processes was found both in the field (Table 1 in Supplement) and in the sediment thin sections (Fig. 5 in Supplement). Also some plants and fresh-water gastropods found in the sequence may give information about frost or snow cover in winter. Shrubs, such as *Juniperus* (juniper), *Betula nana* (dwarf-birch) and *Arctostaphylos alpina* (alpine bearberry) for example are frost-susceptible and prefer a winter snow-cover (e.g.



**Fig. 5.** Summary overview combining a range of multi-proxy data, e.g., LOI, CaCO<sub>3</sub>, lithology, MS, pollen, diatoms (trophy and oxygen), chironomids (chironomid-inferred temperature), the mollusc *Bithynia tentaculata* and delta <sup>18</sup>O data measured on the ostracod *Candona candida* from the Moervaart P3/M3/MB8 master sequence.



**Table 2**  
List of uncalibrated  $^{14}\text{C}$  dates from the Moervaart palaeolake master sequence, compared with those from the dune-slacks at Rieme (Bos et al., 2013) and the regional biostratigraphy of Hoek (1997a, 2001).

Hoek (1997a, 2001) Age BP Biozones		Bio- stratigraphy	Rieme (Bos et al., 2013)	Moervaart "Suikerfabriek" (this study)
11,250	2b	late Allerød		
	2a2	middle Allerød		
11,500	2a		11,565+/-55	
	2a1	early Allerød	11,625+/-60 11,710+/-55 11,920+/-60	11,630+/-60 11,965+/-60
11,900	1c	Older Dryas		
12,100			12,110+/-60	
	1b	Bølling	12,205+/-60	12,240+/-60
12,450				12,205+/-60
	1a	Pleniglacial		

Iversen, 1954; Lauber and Wagner, 1998). Some gastropod species are also frost-sensitive (e.g., *Bithynia tentaculata*, *Lymnaea stagnalis*) whereas others are frost-resistant (e.g., *Galba truncatula*, *Valvata piscinalis* and *Pisidium obtusale*). For example, *Lymnaea stagnalis* and *Bithynia tentaculata* can tolerate temperatures down to  $-4\text{ }^{\circ}\text{C}$ , whereas *Pisidium obtusale* will freeze at  $-11\text{ }^{\circ}\text{C}$  (Gittenberger and Janssen, 2004).

Precipitation and evaporation are more difficult to reconstruct from biological proxies. Changes in relative lake or groundwater levels may result from changes in humidity caused by changes in temperature, precipitation and evaporation. Regionally synchronous changes in lake levels are assumed to be climatically driven (e.g. Lotter, 1988; Harrison and Digerfeldt, 1993; Magny, 2001). Past lake water levels can be reconstructed with changes in the sediment composition and the distribution of the lake vegetation, such as shifts in the aquatic and semi-aquatic botanical taxa (e.g., Digerfeldt, 1986, 1988; Hannon and Gaillard, 1997; Bos et al., 2006). Furthermore, shifts in the littoral and benthic taxa of mollusc, ostracods and chironomid taxa may point to changes in lake levels. For the diatoms, changes in life-form (benthic vs. tychoplanktonic diatoms) likely reflect changes in water depth, while also the acidity may point to fluctuating levels (Demiddele et al., 2016).

#### 4.3.2. Pleniglacial (corresponding to GS-2)

The multi-proxy data indicate that the last part of the Pleniglacial (equivalent to late GS-2) was cold and dry. The botanical data show that the landscape was partly bare and the vegetation open. The vegetation can be characterised as a steppe-tundra with some shrubs of *Juniperus* and many grasses (Poaceae), sedges (Cyperaceae; local presence of sedges is also indicated by the occurrence of the fungus *Clasterosporium caricinum* sp., Fig. 3c in Supplement), and heliophilous herbs (e.g., *Saxifraga granulata*, *Plantago*, *Amaranthaceae*, *Helianthemum*, *Botrychium lunaria*, *Thalictrum*, Fabaceae, Apiaceae, Brassicaceae, Ranunculaceae and Asteraceae including *Artemisia*) (Fig. 3 in Supplement). Records of *Juniperus* and *Selaginella selaginoides* (lesser clubmoss) suggest a

minimum July temperature of  $7\text{--}8\text{ }^{\circ}\text{C}$  (Kolstrup, 1979, 1980). Also the mollusc data suggest a cold and dry climate (Serbruyns, 2010). In these Pleniglacial deposits (Layers A and B) the only specimen of the ostracod *Tonnacypris tonnensis* was recorded (Fig. 5 in Supplement). This species, known mainly from Pleistocene glacial deposits (Van der Meeren et al., 2009 and references therein), is generally considered a strong indicator of cold climatic conditions (Parfitt et al., 2010). Also other species (*Candona candida* and *C. neglecta*) in these layers show a preference for colder waters (Fig. 5 in Supplement).

#### 4.3.3. Bølling (corresponding to GI-1e)

During the Bølling (GI-1e), mean July and January temperatures increased (Bohncke et al., 1987; Van Geel et al., 1989) and a dwarf shrub tundra developed with pioneer communities of heliophilous herbs and scattered (dwarf) shrubs of *Salix* (willow), dwarf-birch, *Juniperus* and *Hippophae rhamnoides* (sea-buckthorn) (Fig. 3a in Supplement). Due to the amelioration of the climate at the start of the Lateglacial, effective precipitation increased and large amounts of meltwater were released (Bohncke, 1993). During this period, the vegetation in the Moervaart depression changed from a wet grassland with shallow pools into a shallow lake with swamps along the shores. In the shallow water of the lake, aquatic taxa (Fig. 3b,d in Supplement), such as Characeae (stoneworts), *Gloeotrichia* (cyanobacteria), *Ceratophyllum* (probably *C. demersum*, rigid hornwort), *Myriophyllum spicatum* (spiked water-milfoil), *Groenlandia densa* (opposite-leaved pondweed), *Potamogeton pectinatus* (fennel-leaved pondweed), *Ranunculus* subgenus *Batrachium* (water crowfoot), *Schoenoplectus lacustris* (lakeshore bulrush), *Utricularia vulgaris* (common bladderwort) and a large number of algae (e.g., *Botryococcus*, *Pediastrum*, *Spirogyra*, *Tetraedon minimum*, *Zygnema* type and HdV-128) were present, indicating calcareous and mesotrophic conditions.

The rise in the groundwater level during the Bølling (and early Allerød) was probably due to an increase in lateral groundwater input from the calcareous sand ridges (favoured by snow-melt)

associated with high summer temperatures. It, however, also may have been a result of the melting of the surrounding permafrost (e.g., Bos, 1998; Hoek et al., 1999; Hoek and Bohncke, 2002; Bos et al., 2006, 2013). The rise in groundwater level is also recorded in several dune-slacks on the Great Coversand Ridge of Maldegem-Stekene, e.g. at the Rieme sites (Bos et al., 2013; Crombé et al., 2012) and in many other Lateglacial lakes in NW Europe (e.g. Bohncke et al., 1987; Bohncke and Wijmstra, 1988; Denys et al., 1990, 1998; Hoek et al., 1999; Bos et al., 2006; Magny, 1995, 2001). Based on the presence of aquatic and semi-aquatic plants such as *Ranunculus* subgenus *Batrachium*, *Myriophyllum spicatum*, *Schoenoplectus lacustris*, *Typha latifolia* (broadleaf cattail), *Ceratophyllum demersum* and *Groenlandia densa* (Fig. 3d in Supplement) minimum mean July temperatures of 10–16 °C could be inferred (e.g., Iversen, 1954; Bell, 1970; Kolstrup, 1979, 1980; Brinkkemper et al., 1987; Litt, 1994; Isarin and Bohncke, 1999). This compares well with the chironomid-inferred mean July air temperatures between ~16.2 and 17.3 °C for this period. The mollusc data also indicate warmer temperatures, but a still relatively dry climate in comparison to the previous Pleniglacial.

#### 4.3.4. Older Dryas (corresponding to GI-1d)

In the Moervaart multi-proxy record, several short climatic oscillations could be recognised during the Lateglacial. A first climatic oscillation is recorded between 155 and 167 cm depth below surface (=Layers D (without the base), E and base of Layer F). The erosional surface between Layers C/D just prior to the Older Dryas and Layers D/E within, both suggest a change to colder climate. The oscillation is reflected in the pollen assemblage as a short-term decrease in the pollen values of trees and shrubs and a temporary increase of herbaceous pollen, especially grasses (Poaceae, Figs. 5 and 6 and Fig. 3a in Supplement). This suggests a drier and colder climate with more barren ground and a larger abundance of grasses in the vegetation, resulting in the development of a grass-steppe-tundra. Based on the pollen biozones, these deposits appear to correlate with the Older Dryas (sensu Van Geel et al., 1989; Hoek, 1997a). The Older Dryas corresponds with a centennial-scale cold oscillation (GI-1d) between 14,075–13,954 b2k (=14,025–13,904 cal yrs BP) in the oxygen isotope record of the Greenland ice-cores.

The aquatic and semi-aquatic botanical taxa (Fig. 5 and Fig. 3b,d in Supplement) both show a strong decrease in values, indicating lower lake levels probably as result of a drier climate. The low acidity reflected in the diatom data also may point to fluctuating lake levels. The increase in salinity indicated by the diatom *Mastogloia lacustris* furthermore implies a shallower lake (or swamp) caused by drying (see Cumming and Smol, 1993; Laird et al., 1998; Pienitz et al., 2000) (Fig. 6 in Supplement). The decreasing trend in averaged  $\delta^{18}\text{O}$  values at the transition to the Older Dryas broadly corresponds with the isotopic signature of atmospheric precipitation over the northern hemisphere during that time (GRIP members 1993; von Grafenstein et al., 1999), but sample resolution limits further interpretation. The other proxies from the Moervaart sequences also suggest a temporarily colder and drier climate. Among the chironomids *Microtendipes pedellus*-type reaches highest abundances in the early part of the Older Dryas of almost 50%. *Microtendipes* occurs in warmer, low-latitude lakes (Walker et al., 1991) and in the Alps, this taxon is predominantly found in lowland and subalpine lakes (Lotter et al., 1997; Heiri and Lotter, 2010). It often indicates intermediate temperatures in Lateglacial and Holocene deposits in northern Europe (Brooks and Birks, 2001).

The Moervaart botanical and mollusc data (presence of the cold water indicator *Valvata piscinalis*) also suggest lower summer temperatures in comparison to the previous Bølling period. This is

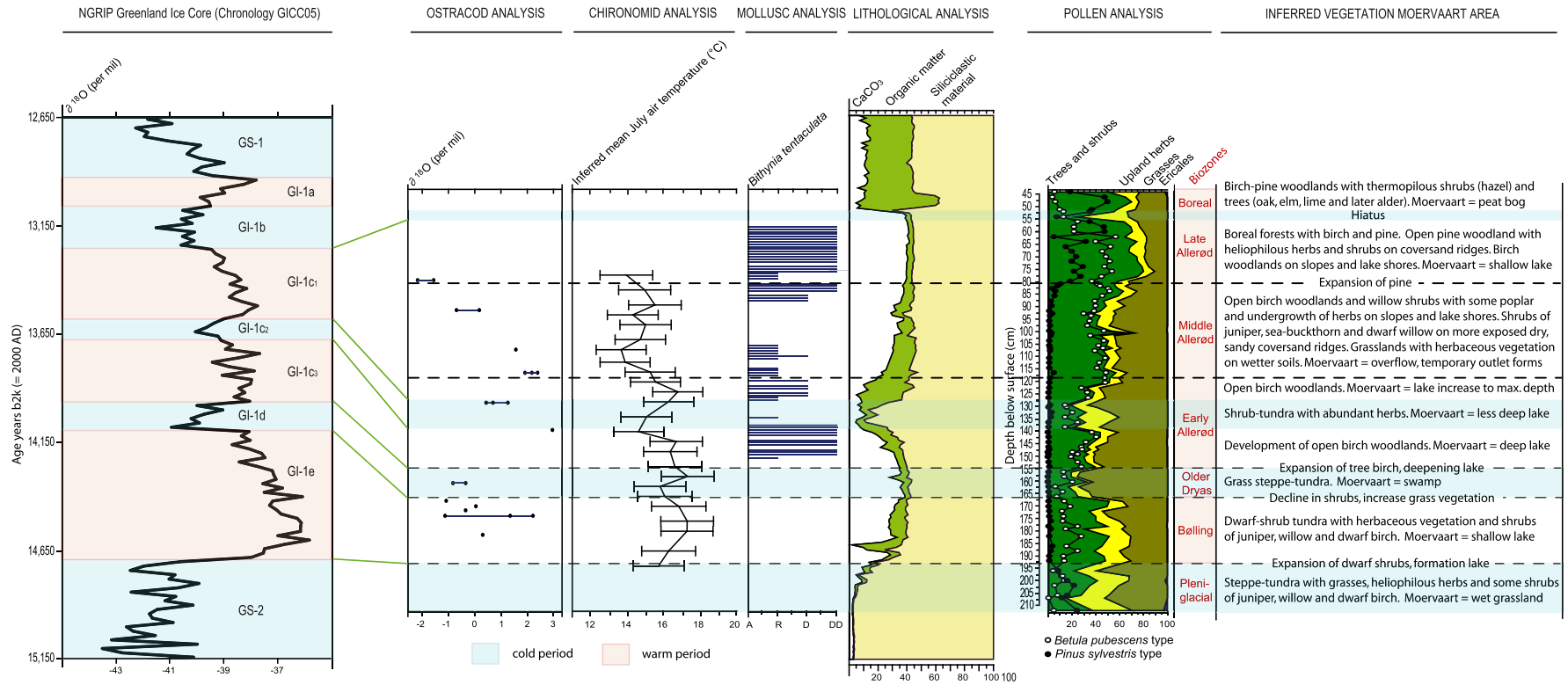
supported by the chironomid-inferred mean July air temperatures that show a minor decrease to 15.8 °C (Figs. 5 and 6). In sediment thin section 7 lenticular microstructures (Fig. 7c and d in Supplement) were observed at the transition from Layer D/E to F (= top of Older Dryas deposits), which are probably related to freeze/thaw cycles and associated ice segregation. Fossil soil wedges or frost cracks that were seen in the field in the top of these Older Dryas deposits indicate mean annual air temperatures below –1 to 0 °C (Maarleveld, 1976; Huijzer and Isarin, 1997) and thus indicate very cold winters. The occurrence of *Juniperus* also suggests the presence of a protecting snow cover in winter (Iversen, 1954). This oscillation is also recorded at the Rieme sites on the Great Sand Ridge of Maldegem-Stekene (Bos et al., 2013), at Snellegem (Denys et al., 1990, 1998) in NW Belgium and many other sites in NW Europe.

#### 4.3.5. Early Allerød (corresponding to GI-1c3)

At the start of the Allerød (GI-1c3), tree birch (*Betula* cf. *pubescens*) immigrated (Fig. 3a,b,d in Supplement), boreal birch forests developed and the vegetation became more diverse. Later also aspen (*Populus tremula*) appears. The woodlands however, remained relatively open. Soils formed and more nutrient-rich habitats developed. From the start of the Allerød, people of the *Federmesser* Culture were present in the Moervaart area. The area provided a very suitable landscape for the *Federmesser* Culture hunter-gatherers with extensive and fertile woodlands and lake edges for wild game hunting, plant gathering and fowling. The Moervaart lake furthermore supplied fresh drinking water. During this period the groundwater level increased again, which resulted in a deepening of the Moervaart palaeolake and the formation of ponds and dune-slacks on the Great Coversand Ridge Maldegem-Stekene (Derese et al., 2010; Crombé et al., 2012; Bos et al., 2013). The deepening of the lake is well reflected in all biological proxies (Figs. 3, 5–9 in Supplement): The (re-)appearance of a large number of marsh taxa, floating-leaved aquatics (e.g., *Nymphaea alba*, white water lily and *Nuphar lutea*, yellow water lily), cyanobacteria and various algae. Amongst the chironomid, mollusc and ostracods, many species appear that are connected with presence of macrophytes (Bos et al., submitted). In the diatom assemblage there is an increase in species richness and number of individuals at the start of the Allerød that relates to a growing habitat diversity in the lake.

Superimposed on the groundwater rise during the Allerød, water level fluctuations connected with short climatic oscillations occurred (see further discussion below). With an increase in organic production, the sediments in the lake changed during the early Allerød from calcareous to humic gyttja. The presence of typical seepage indicators, e.g., *Utricularia*, *Potamogeton mucronatus* (flat-stalked pondweed), *Menyanthes trifoliata* (bog bean), *Hippuris palustris* (mare's-tail), *Myriophyllum spicatum* and Characeae (Weeda et al., 1988; Bos et al., 2005; Van Asch, 2007) demonstrates that local seepage of carbonate-rich groundwater from the surrounding dunes occurred. From the early to middle Allerød, however, the clastic inflow and seepage flow were diluted by authigenic lacustrine organic productivity, due to higher lake water levels and an increased trophic status. At the end of the early Allerød, the Moervaart palaeolake probably reached its largest size and deepest phase. This is reflected in a maximum in the LOI values (Layer I, Figs. 5 and 6) and dominance of a large number of aquatic taxa (e.g., *Myriophyllum spicatum*, *Ceratophyllum*, *Potamogeton* spp., *Nymphaea alba* and *Nuphar lutea*). Many of the recorded diatom, mollusc, ostracod and chironomid taxa are associated with these aquatic plants.

Based on the presence of botanical taxa (*Myriophyllum spicatum*, *Schoenoplectus lacustris*, *Potamogeton mucronatus*, *Typha latifolia*, *Ceratophyllum demersum*, *Carex pseudocyperus*, *Sanguisorba minor*



**Fig. 6.** Correlation of the various multi-proxy data of the Moervaart P3/M3/PB8 master sequence with the  $\delta^{18}\text{O}$  data of the NGRIP Greenland Ice-Core (Rasmussen et al., 2014). Age is in years b2k (= 2000 AD). Short climatic oscillations as recorded in the NGRIP Greenland Ice Core and various multi-proxy data are displayed. The inferred regional vegetation development in the Moervaart area is also shown.



and *Arctostaphylos alpina*) (Fig. 3a,b,d in Supplement) minimum mean July temperatures of 12–15 °C could be inferred (e.g. Iversen, 1954; Kolstrup, 1979, 1980; Brinkkemper et al., 1987; Litt, 1994; Hoffmann et al., 1998; Isarin and Bohncke, 1999). This compares well with the chironomid-inferred mean July air temperatures between ~14.6–16.7 °C for this period (Figs. 5 and 6). Although the palaeolake in the Moervaart depression was a shallower and larger lake ecosystem than commonly explored for chironomid-based palaeotemperature reconstruction, the temperatures are comparable to Interstadial C-IT reconstructions from Klein Ven in the southern Netherlands (Van Asch et al., 2013) and Hijkermeer in the northern Netherlands (Heiri et al., 2007, 2011). Furthermore, warm water indicators such as *Bithynia tentaculata* and *Lymnaea stagnalis* are recorded (Fig. 9 in Supplement). Relative to the other dominant ostracod species in this record, *Herpetocypris reptans* has the highest minimum July and January temperature tolerance (Gobert, 2012), again pointing to possible warmer water temperatures during the early Allerød.

Also the presence of a protecting snow cover in winter must be assumed (e.g. Van der Hammen, 1951; Iversen, 1954; Lauber and Wagner, 1998) as *Juniperus*, *Betula nana* and *Arctostaphylos alpina* are frost susceptible. The snow in winter must have been variable in thickness depending on local differences in relief, while *Arctostaphylos alpina* does not tolerate more than a thin and transient snow cover (Iversen, 1954). The mollusc *Bithynia tentaculata* is also frost susceptible.

#### 4.3.6. Early Allerød oscillation (corresponding to GI-1c2)

The most pronounced short-term climatic oscillation is present between 128 and 140 cm depth below surface and occurred within the early Allerød. This oscillation is clearly visible in the lithology as a gyttja deposit with many sandy layers (Layer H) and very little CaCO<sub>3</sub> and organic material. Furthermore, it is also recognised in the pollen data as a distinct dip in the tree and shrub values and a strong increase in herbaceous pollen, especially grasses, sedges and various herbs. This suggests that the landscape became more open with more barren ground and herb and grass vegetation. The pollen data show that the grasslands also consisted of many herbs such as Ranunculaceae (buttercups), Apiaceae (members of the parsley family), *Plantago* (plantain), *Gentiana* (gentians), *Helianthemum* (rockrose), with on the wetter soils also sedges, *Thalictrum* (meadow-rue) and *Filipendula* (meadowsweet). On the drier, open, sandy soils, also grasses and *Artemisia* (mugwort) probably expanded (Fig. 5 and Fig. 3a and b in Supplement).

In the mollusc data, a temporary decline (sample nr. 5, S4 sequence, Fig. 9 in Supplement) is recorded, while in the ostracod record, the pioneer species, *Candona candida* returns. The warm water molluscs *Bithynia tentaculata* and *Acroloxus lacustris* are absent in this interval, while the cold-water indicators *Pisidium obtusale* and *Valvata piscinalis* are present (compare Gittenberger and Janssen, 2004) (Figs. 5,6 and Fig. 9 in Supplement). This indicates a colder and possibly drier climate, probably with frost, as *Bithynia tentaculata* is frost-sensitive. In addition, there is also evidence for colder conditions from the chironomid and micro-morphological data. The chironomid C-IT-curve shows a ~2 °C temperature decline to ~14.6 °C, while in sediment thin section 6 (transition Layer G/H) lenticular microstructures were observed which, again, may point to freeze/thaw cycles and associated ice segregation.

Due to the more open vegetation and larger barren ground area, more surface erosion occurred during this period. Wind erosion increased and sandy layers were deposited in the lake. This is corroborated by the presence of thin layers composed entirely of rounded sand grains in thin section 6 (Layer H). This layer also becomes thicker towards the eastern end of the palaeolake

(Crombé et al., 2013). The sands were probably blown into the palaeolake from the adjacent Great Sand Ridge by northern winds. The roundness, size and mineralogy (dominance of quartz and some calcium carbonate) of the sand grains are very similar to the deeper, non-decalcified sand layers of the ridge. Most likely these sediments were deposited as lake infill beds on the lake surface during episodes of freezing (Crombé et al., 2013). The erosional contact between Layers G/H also may suggest the start of a colder phase.

This oscillation during the early Allerød, probably corresponds with a centennial scale cold oscillation (corresponding to Greenland Interstadial-1c2) in the oxygen isotope record of the Greenland ice-cores (Rasmussen et al., 2014, Fig. 6) that dates around 13,640 b2k (= 13,590 cal yrs BP). In the Moervaart sequence this oscillation could be dated quite precisely, as both bottom and top were dated (Table 1). From this it can be concluded that the oscillation started slightly after 13,990–13,780 b2k (= 13,940–13,730 cal yrs BP) and ended at 13,600–13,450 b2k (= 13,550–13,400 cal yrs BP), which seems to fit well with the chronology of this event in the Greenland ice-core records. The same oscillation is probably also recorded in several dune-slacks on the Great Sand Ridge of Maldegem-Stekene (Crombé et al., 2012), in particular at the Rieme sites (Bos et al., 2013). At these sites accumulation of organic deposits ceased around ca. 13,550 b2k (= 13,500 cal yrs BP) when water levels dropped and accumulation of sand continued (Table 2). At Rieme, a chironomid sample from the uppermost deposits (~corresponding to the start of GI-1c2) includes a single record of the cold-stenothermic *Micropsectra radialis*-type (e.g. Brodin, 1986), a taxon, which usually indicates lower temperatures. Both the botanical and chironomid assemblages at Rieme indicate declining temperatures, which also suggest correspondence to this drier and/or cooler phase.

The difference between this oscillation and the earlier one (corresponding to GI-1d) is that during the former more heliophilous herbs and less grasses were present in the regional vegetation around the Moervaart lake and that the mean air summer temperatures were probably slightly lower. In contrast to the Moervaart sequence, in the Greenland ice-cores oscillation GI-1d is much more pronounced than GI-1c2 (Rasmussen et al., 2014).

#### 4.3.7. Middle Allerød (corresponding to GI-1c1)

At the transition from the early to middle Allerød, some marked changes occur in the various Moervaart proxies. A decrease of ca. 4‰ in δ<sup>18</sup>O during the middle Allerød (Figs. 5 and 6) suggests a hydrological change in the Moervaart lake towards a lake with and inflow and an outflow. This was possibly due to an increased groundwater discharge or shift to a topographically through-flow lake; both processes would alter the hydrological residence time and shift the isotopic signature towards more depleted values. The decrease of ca. 4‰ in δ<sup>18</sup>O between 119.5 and 79 cm (Figs. 5 and 6) is much too large to be primarily attributed to the change in isotopic signature of meteoric water. Comparison with reconstructed meteoric δ<sup>18</sup>O for lake Ammersee (Von Grafenstein et al., 1999) indicates that this change was much smaller during this time window. Speculation about the proportion that can be attributed to a hydrological change (increased groundwater discharge, or shift to a topographically through-flow lake) or to other factors is tentative, since also changes in seasonality of precipitation and temperature can affect this change.

During this period, also the deposits become less organic and change from humic, calcareous gyttja (Layer J), to clayey, calcareous gyttja (Layer K) into light grey lake marl (Layers L1 and L2) (Fig. 5). This change is also reflected in the magnetic susceptibility values, which show a marked increase to higher values at this level (110 cm, Layer J), suggesting changing biochemical and redox

conditions resulting in the formation of magnetic minerals (Fig. 5). Furthermore, diatoms increase that need more oxygen (i.e. tycho-planctonic species) and changes occur in the chironomid assemblages that also may be attributed to changes in the lacustrine system.

The hypothesis is that at the early/middle Allerød transition a change occurred in the hydrological system of the Moervaart lake. Before and during the early Allerød, the Moervaart lake was a shallow and topographically closed lake (i.e., reflected in the good correlation between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ ). There was a complex inflow delta at the western lake margin, but no topographical outflow. During the early Allerød, the lake level rose to a maximum level (i.e., reflected in the high LOI values, Layer I, Figs. 5 and 6). During this high water level phase at the beginning of the middle Allerød, an outlet was formed, possibly as a result of the incision of the Kale/Durme river (Fig. 2b). Unfortunately the exact dating of the incision is unknown so far, but radiocarbon dates from the basal infilling of the channel clearly indicates that the incision took place before ca. 13,150 b2k (= 13,100 cal yrs BP) (Crombé et al., 2014). Due to the formation and gradual incision of this outlet, the lake water level was also lowered. The establishment of the topographically open lake resulted in opposing trends between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  for the later part of this record (Figs. 5,6 and Fig. 4 in Supplement). The erosional contact between Layers I/J corresponds to the transition from a lake that was at its deepest point to a lake with slightly shallower depth. During this period, channels developed in the lake bottom. At the same level an increase in the magnetic susceptibility signal occurs, possibly indicating influx of erosion material.

During the middle Allerød, some minor fluctuations were observed in the chironomid and *Bithynia* records. In the chironomid record a fluctuation in the C-IT is present between 102.5 and 117.5 cm depth below surface (Layer J), although similarly as for the early Allerød oscillation, variations in CI-T are within the prediction error of the applied chironomid-temperature transfer function (Figs. 5,6). In the other multi-proxy data, however, no distinct changes are observed. The main taxon responsible for this fluctuation seems to be *Corynocera ambigua* (Fig. 8 in Supplement). *C. ambigua* occurs in cold oligotrophic lakes (Brooks et al., 2007) and has an optimum July temperature of ~10 °C in the calibration dataset. Nevertheless, this species is also found in warm, shallow lakes in Denmark (Brodersen and Lindegaard, 1999). The occurrence of this species, therefore, does not necessarily point to colder conditions, even though these might be reconstructed from the record.

Another fluctuation during the middle Allerød is observed in the *Bithynia* curve (Figs. 5, 6). *Bithynia* shells and opercula are absent between 87.5 and 105 cm depth below surface (top Layer J and Layer K). This may suggest a drier and colder phase with lower winter and/or summer temperatures and periods of frost persisting during most of the spring. The micromorphological data also show that at the transition between Layers J and K (thin section 4), horizontal planes (only partially accommodated) were present, which may point to ice segregation. In the regional pollen assemblage, birch values were slightly decreased and herb values (especially *Artemisia*) increased (Fig. 3a in Supplement), which suggests a temporary presence of slightly more herbaceous vegetation in the Moervaart area. In the local pollen assemblage (Fig. 3b in Supplement) this is accompanied by a decrease in the aquatic taxa (especially *Nuphar* and *Nymphaea*). This suggests that a lake level lowering was responsible for this fluctuation and not a change to a colder climate. Also the fact that Layer K is missing from the more littoral S4 sequence, may point to lower lake water levels. Furthermore, mollusc taxa such as *Pisidium milium*, *Pisidium pulchellum* and *Pisidium obtusale* that prefer small surface waters, suggest lower lake water levels (Fig. 9 in Supplement). The absence

of the relatively deeper water species *Bithynia tentaculata* and aquatic taxa was probably related to the decrease in lake water depth as result of a newly formed temporary outlet, as discussed above. At the same time birch settled along the lakeshore, which also may have contributed to a higher evapotranspiration.

#### 4.3.8. Late Allerød (corresponding to GI-1c1)

At the start of the late Allerød (Layer L1), the water level in the Moervaart depression slightly increased again. These somewhat higher water levels during the late Allerød are reflected in the temporary re-occurrence of aquatics such as, *Potamogeton natans*, *P. mucronatus*, *Nymphaea*, *Nuphar*, *Utricularia* and Characeae (Fig. 3b,d in Supplement). The re-occurrence of typical seepage taxa (e.g., *Utricularia*, *Menyanthes trifoliata*, *Potamogeton mucronatus*) suggests that leaching of carbonates from the adjacent coversand ridges increased again.

For the late Allerød, the chironomid-inferred mean July air temperatures show slightly lower temperatures between 13.6 and 15.5 °C compared to the early Allerød (Figs. 5 and 6). This is also suggested by botanical taxa, such as *Schoenoplectus lacustris*, *Potamogeton mucronatus*, *Typha latifolia*, *Carex pseudocyperus* (cyperus segde), *Nymphaea*, *Nuphar lutea* and *Sanguisorba minor* (salad burnet) from which minimum mean July temperatures of 12–13 °C could be inferred (e.g., Iversen, 1954; Kolstrup, 1979, 1980; Brinkkemper et al., 1987; Isarin and Bohncke, 1999). In this interval (Fig. 5 in Supplement), also some indicator species for colder, arid conditions were present in the ostracod record, e.g., *Fabaeformiscandona protzi* is followed by *F. levanderi*, but they were only found in small numbers. *Fabaeformiscandona levanderi* has in past palaeoenvironmental studies been regarded as indicative of more or less arid conditions with increased evaporation rates (e.g. Wetterich et al., 2005). The increase in evapotranspiration was probably related to the immigration of pine and formation of pine woodlands on the sandy coversands and soils nearby.

#### 4.3.9. Late Allerød oscillation (corresponding to GI-1b)

Towards the end of the late Allerød (Layer L2), the terrestriation processes progressed and a swamp with some pools of open water developed. It is recorded as the start of a hiatus in the Moervaart master sequence at 55 cm below surface. The erosional contact between Layers L2/M suggests the start of a colder climate phase. This is also supported by the diatom data that record an increase of cold resistant species (e.g., *Cymbella en Ulnaria*) in the top part of the Moervaart lake sequence (Fig. 6 in Supplement) (Demiddele et al., 2016). Also the presence of the cold-water ostracods *Candona candida* and *C. neglecta* in the upper late Allerød deposits may point to a colder phase (Fig. 5 in Supplement). As a consequence of this cold interval or oscillation the Moervaart lake and surrounding ponds ceased to exist, resulting in a marked decrease or even a hiatus in the human occupation and exploitation of the region which lasted almost 2000 years.

In absence of radiocarbon dates the exact chronology of this oscillation, which most likely was responsible for the permanent disappearance of the Moervaart lake, is difficult to determine. Either it is related to the Intra Allerød Cold Period (IACP; Lotter et al., 1992; Donnelly et al., 2005; Van Raden et al., 2013)/GI-1b dated in GICC05 between 13,311 (= 13,261 cal yrs BP) and 13,099 b2k (= 13,049 yrs BP) (Lowe et al., 2008; Blockley et al., 2012) or to the Younger Dryas/GS-1. There is, however, indirect evidence in favour of an association with the GI-1b-event. A radiocarbon date from the basis of a peat layer in one of the shallow anastomosing gullies, which fed the Moervaart lake, gave a date between 13,150 (= 13,100 cal yrs BP) and 12,846 b2k (= 12,796 cal yrs BP (Crombé et al., 2014). This indicates that the lacustrine system (at least locally) had turned into a swamp already before the start of the

Younger Dryas. A similar observation has been made in a nearby smaller freshwater lake at Snellegem; here peat started to grow on top of the lacustrine sediments during the *Pinus*-stage of the late Allerød (shortly before 12,980 (= 12,930 cal yrs BP) and 12,630 b2k (= 13,580 cal yrs BP) (Denys et al., 1990, 1998).

Similarly temporal lower lake level phases in the final stage of the Allerød, leading to facies changes and/or hiatuses are also recorded in other areas in NW Europe, for example in the southern Netherlands (Hoek et al., 1999; Hoek and Bohncke, 2002; Bos et al., 2006) and in northern France (Deschodt et al., 2009; Magny, 1995, 2001). In Ireland this oscillation was recorded in C-IT and isotope data (Van Asch and Hoek, 2012; Van Asch et al., 2012).

## 5. Conclusions

This multi-proxy study enabled us to make a detailed reconstruction of the Lateglacial landscape along one of the largest palaeolakes within NW-Europe. Radiocarbon dates from the master sequence in combination with  $^{14}\text{C}$  dates of the nearby Rieme sequence provided an absolute chronology for the record, which encompasses the final part of the Pleniglacial up to the late Allerød. Furthermore, there was an excellent match with the regional biostratigraphy.

Former vegetation patterns and hydrology were strongly influenced by climate changes that occurred during the Lateglacial. During this period, vegetation and geomorphology of the landscape in general changed from a tundra landscape to boreal forest with birch and pine. The Moervaart palaeolake itself was formed during the Bølling (corresponding to GI-1e) as the result of a major rise of the groundwater table in the area. This groundwater rise was likely caused by an increase in effective precipitation, probably in combination with an increase in lateral groundwater input from the calcareous sand ridges (favoured by snow-melt) associated to high evapotranspiration in summer. The groundwater level rise continued during the early Allerød. From the start of the Allerød onwards, the area provided a suitable landscape for *Federmesser* Culture hunter-gatherers with extensive and fertile woodlands for hunting and gathering. In addition, the Moervaart lake could provide fresh drinking water. The progressive vegetation development, however, was interrupted several times by a number of cold reversals. Three centennial-scale cold oscillations could be recognised in the multi-proxy record: (1) the so-called Older Dryas or corresponding to GI-1d in the Greenland ice-cores, (2) a short and pronounced cold event during the early Allerød which could be correlated to GI-1c2 and (3) a cooling event during the late Allerød most likely corresponding to the Intra Allerød Cold Period (IACP)/GI-1b or the Younger Dryas/GS-1. As a consequence of the latter, the Moervaart lake and surrounding ponds ceased to exist, resulting in a marked decrease or even a hiatus in the human occupation and exploitation of the region which lasted almost 2000 years.

Of special interest is the early Allerød oscillation, which contrary to the Greenland ice-core record apparently had a greater environmental impact compared to the Older Dryas. This short but abrupt cooling event triggered wind erosion, resulting in the deposition of sandy material into the Moervaart lake and surrounding ponds. The latter were covered by aeolian sediments reaching a thickness between 1 and ca. 3 m. Along the northern Moervaart bank aeolian sands were deposited on top of the lacustrine sediments, indicating that the coversand ridge of Maldegem–Stekene slightly moved southwards. This most likely is connected to the formation of a ca. 10 km<sup>2</sup> wide wind-blow depression on top of the coversand. Whether these aeolian sediments were all deposited during the early Allerød event or represent different phases of wind erosion, including the Younger Dryas, remains to be investigated by means of OSL dates. However, the

presence of *Federmesser* Culture sites on top of these aeolian sediments strongly suggests that the deposition mainly occurred during the Allerød. If this is confirmed by future research, the erosional event of the early Allerød must have had a considerable impact on the environment and human occupation of the Moervaart region.

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## Appendix A. Supplementary data

Supplementary data related to this article can be found at <http://dx.doi.org/10.1016/j.quascirev.2017.02.005>.

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