



Climatic and environmental changes during the Weichselian Lateglacial Interstadial in the Weerterbos region, the Netherlands

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Lake sediment records from the Weerterbos region, in the southern Netherlands, were studied to reconstruct summer temperature and environmental changes during the Weichselian Lateglacial Interstadial. A sediment core obtained from a small lacustrine basin was analysed for multiple proxies, including lithological changes, oxygen isotopes of bulk carbonates, pollen and chironomids. It was found that the oxygen isotope record differed strongly from the other proxies. Based on a comparison with three additional lake sediment records from the same region, it emerged that the oxygen isotope records were strongly affected by local environmental conditions, impeding the distinction of a regional palaeoclimate signal. The chironomid-inferred July air temperature reconstruction produced inferred interstadial temperatures ranging between -15° and 18°C , largely consistent with previously published results from the northern part of the Netherlands. A temporary regressive phase in the pollen record, which can be tentatively correlated with the Older Dryas, preceded the expansion of birch woodland. Despite differences between the four pollen records from the Weerterbos region, a comparable regressive vegetation phase that was possibly the result of a shift to drier conditions could be discerned in all of the profiles. In addition, a temporary temperature decline of -1.5°C was inferred from the chironomid record during this regressive phase. The multi-proxy approach used here enabled a direct comparison of inferred changes in temperature, vegetation and environmental conditions at an individual site, while the multi-site approach provided insight into the factors influencing the pollen and isotope records from these small-scale depressions.

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Terrestrial vegetation experienced major changes during the Weichselian Lateglacial (c. 14.7–11.7 cal. ka BP) in large parts of the northern circum-Atlantic region (e.g. Cwynar *et al.* 1994; Peteet *et al.* 1994; Walker *et al.* 1994; Hoek 1997a, b). In many regions of NW Europe, climatic amelioration at the onset of the Lateglacial Interstadial led to a change from polar desert towards birch and pine forests (e.g. Walker *et al.* 1994; Hoek 1997a, b). Subsequently, the Younger Dryas cold phase was marked by a change to more open woodland or shrubland and *Empetrum* heaths in Britain and continental NW Europe. Later, closed forests (re-)established as a result of warming at the transition to the Holocene.

Numerous pollen records from the Netherlands have been published, which enabled the development of a biostratigraphical zonation scheme for the Lateglacial period (Hoek 1997a, b). These records show clear variations in vegetation development related to regional differences in the abiotic landscape. However, all records reflect a progressive vegetation development, culminating in open birch and pine forests during the later part of the interstadial. Furthermore, one or more short-lived regressive phases are often apparent within the Lateglacial Interstadial. One of these regressive phases, the Older Dryas, is usually recorded as a tem-

porary decline in *Betula* prior to the widespread development of birch forests. In addition, multiple records show one or two short-lived declines in the arboreal pollen component during the succeeding Allerød. These fluctuations may be correlated with cold events such as Greenland Interstadial (GI) 1b and 1d, which are recorded in the oxygen isotope records of the Greenland ice cores (Björck *et al.* 1998; Lowe *et al.* 2008). In central Europe these cold events have been correlated with the Aegelsee and Gerzensee oscillations, events identified in Swiss oxygen isotope records (Lotter *et al.* 1992). However, the lack of a common dating framework and the large errors often associated with radiocarbon dates from Lateglacial deposits impede a direct correlation between these records in many parts of the European continent. Furthermore, the influence of local vegetation on pollen records may seriously hamper the identification of regional vegetational fluctuations.

An attempt to correlate pollen records from the southern Netherlands with the Greenland ice-core records based on oxygen isotope wiggle-matching has been published by Hoek & Bohncke (2001). Based on these results, Hoek & Bohncke (2001) hypothesized that the cold event GI-1d occurred after the expansion of birch forest. In contrast, a first chironomid-inferred

reconstruction of Lateglacial summer temperatures for the northern Netherlands (Heiri *et al.* 2007, 2011) indicated that a cold oscillation (GI-1d) immediately preceded the expansion of birch forest. These results could reflect a lag in the expansion of birch forests in the northern Netherlands compared with the southern part (Heiri *et al.* 2007). In addition, results from both sites indicated that a second cold oscillation (GI-1b) coincided with the expansion of pine. However, negative excursions in the oxygen isotope records from the southern Netherlands, correlated with GI-1d and GI-1b, were not very pronounced. Thus, uncertainty remains with respect to the relative timing of environmental changes throughout the Netherlands in relation to climatic changes during the Lateglacial Interstadial.

In this study, lake sediment records from the Weerterbos region were studied to allow a comparison of summer temperature and environmental changes during the Lateglacial Interstadial. A sediment core from Klein Ven, a small lacustrine basin, was analysed for lithological composition, pollen, oxygen and carbon isotopes of lacustrine carbonates, and chironomid assemblages. The chironomid assemblage data allow the reconstruction of mean July air temperatures in the region. The study of multiple proxies in the same sediment sequence enables a direct comparison of the response of the various indicators without uncertainties

in chronology (e.g. Lotter 2003). Furthermore, the pollen and isotope records from Klein Ven are compared with other records from the Weerterbos region. A comparison of these different pollen records shows whether fluctuations in the Klein Ven record are apparent at the local level (within ~2.5 km) and allows local (cf. Jacobson & Bradshaw 1981) and regional changes in vegetation development to be distinguished. Finally, the comparison of oxygen isotope records of these sites provides insight into the factors influencing the isotopic composition of the lacustrine deposits within the different depressions.

Study area and methods

The study site is located in the Weerterbos area, in the southeastern Netherlands (Fig. 1). This region consists of aeolian coversand deposits of Pleistocene age (Schokker & Koster 2004). In the Weerterbos region, a number of infilled circular basins occur within the coversand deposits, north of a large coversand ridge in the source area of the Sterkselse Aa river (Fig. 1; Hoek & Joosten 1995). These depressions were presumably formed as a result of melting ground-ice lenses (open-system pingos) in the Lateglacial Interstadial. In 1993, lithological cross-sections were cored through a

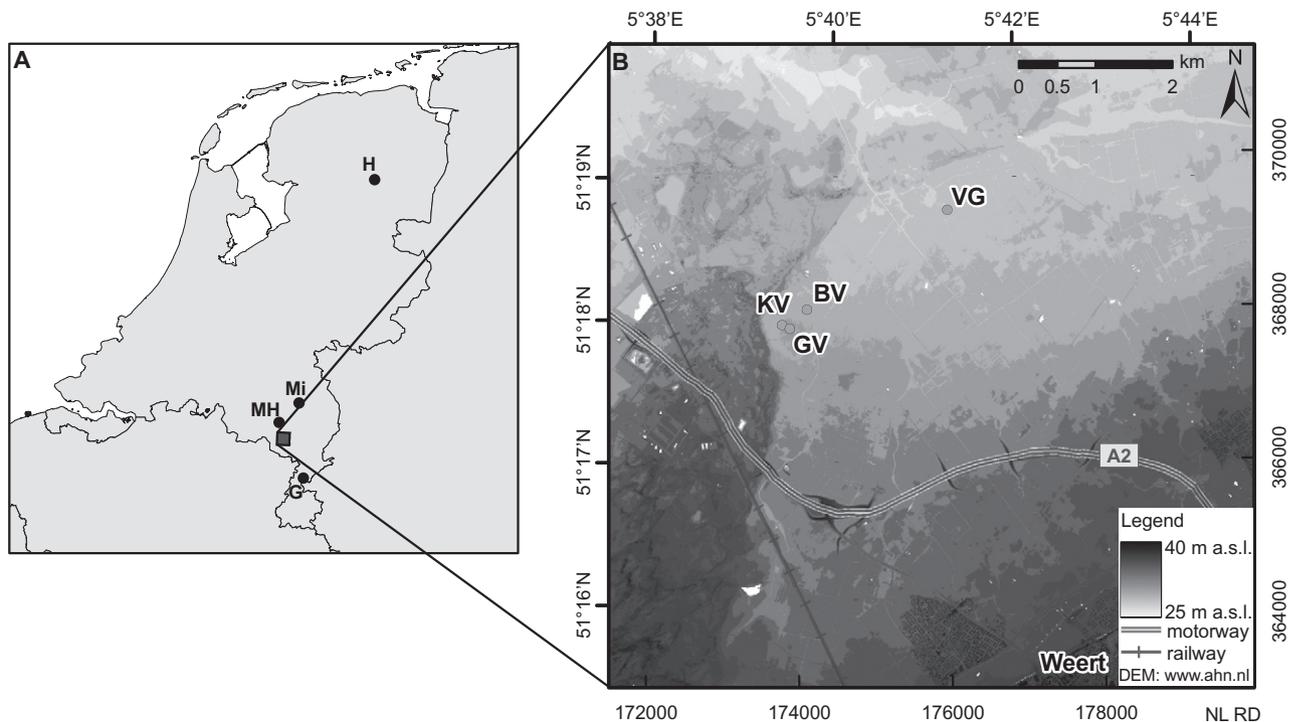


Fig. 1. A. Map of the Netherlands showing the study area (square). Other sites mentioned in the text are also indicated. H = Hijkermeer (Heiri *et al.* 2007); Mi = Milheeze (Bos *et al.* 2006); MH = Mierlo-Ven Hoenderboom (Zagwijn 1971; Hoek 1997b); G = Gulickshof (Hoek *et al.* 1999). B. Digital elevation map showing the four locations in the Weerterbos region. KV = Klein Ven; GV = Groot Ven; BV = Berkenven; VG = Vliegersgat. The city Weert is located on a coversand ridge south of the Sterkselse Aa valley.

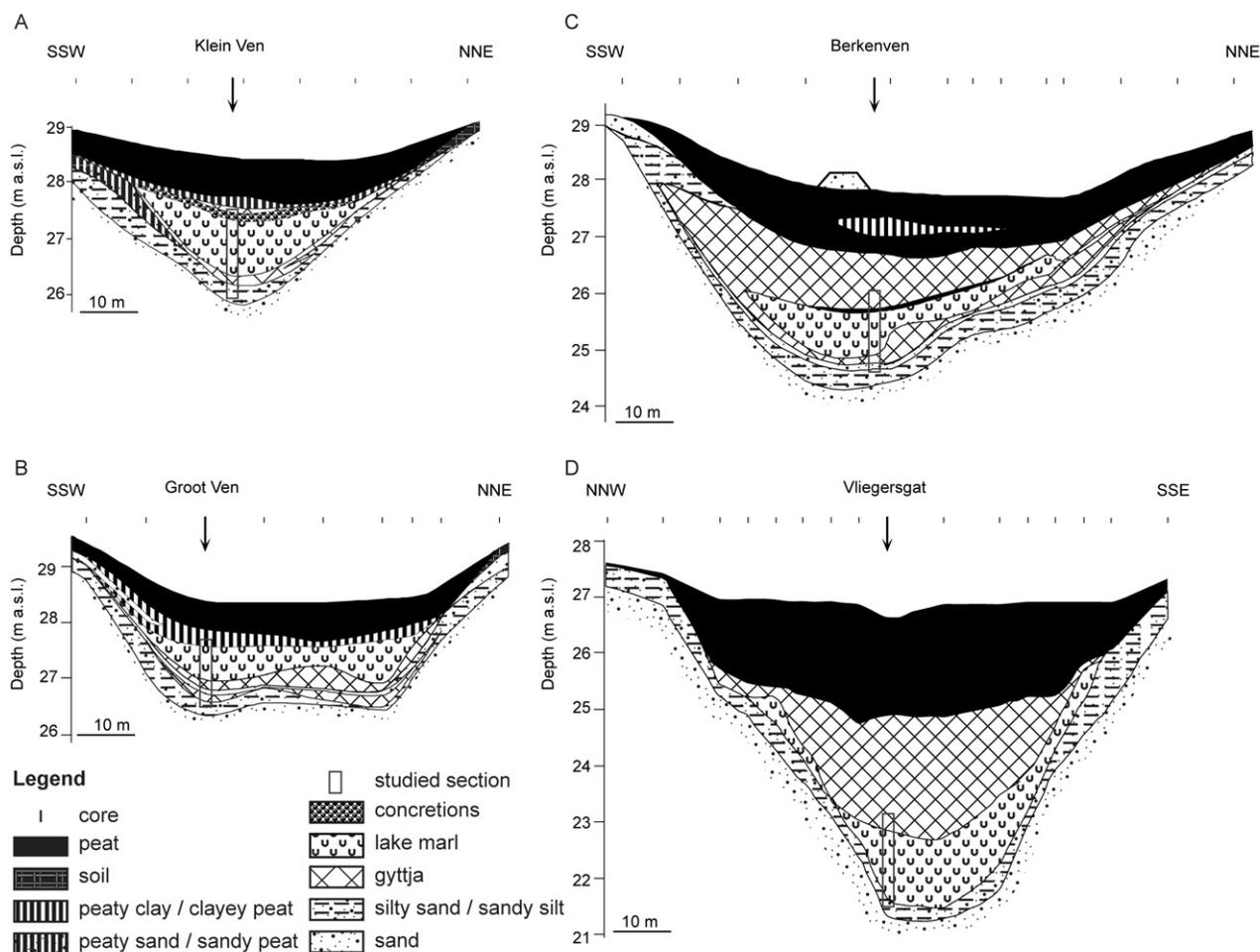


Fig. 2. Lithological cross-sections of the Klein Ven basin (A) and three other basins in the Weerterbos region: Groot Ven (B), Berkenven (C) and Vliegersgat (D). The locations of these basins are depicted in Fig. 1. Also shown are the locations of the sediment cores. The Klein Ven, Groot Ven and Berkenven cross-sections are modified from Hoek & Joosten (1995).

number of these depressions (Hoek & Joosten 1995). Preliminary palynological results showed that their infill consists of calcareous gyttja deposits of Lateglacial age. These deposits presumably formed as a result of seepage of carbonate-rich groundwater after melting of the ground-ice lenses. One of these basins is Klein Ven (KV), which has a diameter of ~70 m and is filled with lake sediments with a maximum thickness of ~2.5 m (Fig. 2; Table 1). An adjacent basin is Groot Ven (GV, Fig. 2), which has a diameter similar to that

of Klein Ven, and a maximum fill of less than 2 m (Table 1). Palynological and isotope results from (Weerterbos) Groot Ven have been published by Hoek & Bohncke (2001). Other nearby basins are Berkenven (BV) and Vliegersgat (VG) (Fig. 2; Table 1), from which cores have previously been analysed for oxygen isotopes and pollen (unpublished data). Pollen and isotope results from the Groot Ven, Berkenven and Vliegersgat records are used for comparison with the KV record. Comparison of these four pollen records is

Table 1. Morphometric characteristics of the four Lateglacial depressions in the Weerterbos area that are examined in this study.

Site	Distance from Klein Ven (m)	Present-day elevation (m a.s.l.)	Maximum thickness of fill (m)	Diameter (m)
Klein Ven	NA	~28.5	~2.5	~70
Groot Ven	~100	~28.5	<2	~70
Berkenven	~350	~28.5	~3	~80
Vliegersgat	~2600	~27.0	~5	~80

NA indicates not applicable.

based predominantly on abundance changes in *Betula*, *Salix*, *Juniperus* and *Pinus*. The original pollen counts of the Berkenven pollen diagram were low (on average 50–100). Therefore, pollen samples of this record have been recounted to obtain a more reliable pollen sum of 300.

In 1999, a sediment core was obtained from the Klein Ven depression, before the lake sediment was partly dug away for fen restoration purposes. Based on the exploratory cross-sections, a location near the centre of the former lake was selected for sampling (Klein Ven (KV): 51°17'54"N, 05°39'22"E). Duplicate sediment cores were obtained from this location, using a 6-cm-diameter piston corer. These sediment cores were then stored at 4°C for future research. In 2008, the cores were visually correlated based on lithology, and sampled for carbonate and organic matter content, stable oxygen and carbon isotopes of bulk carbonates, pollen, chironomids and AMS radiocarbon dating.

Samples for carbonate and organic matter content were obtained at 1–10 cm intervals. Total carbonate content was measured with the Scheibler method following Hoek & Bohncke (2001). After the samples were dried and ground, 0.2 g was treated with a 5% HCl solution, and the volume of produced CO₂ was measured. Results are expressed as weight percentage CaCO₃. Total carbonates in the Klein Ven core presumably consist of a combination of calcite (CaCO₃), siderite (FeCO₃) and possibly other carbonates. Because siderite is heavier than calcite, the presence of siderite leads to an underestimation of the weight percentage of total carbonates.

The organic matter content was determined by loss on ignition (LOI). Dried samples (~1 cm³) were combusted at 550°C for 1 h to burn all organic matter while minimizing the loss of carbonates (Dean 1974). This procedure also minimizes the loss of siderite, which is apparently present in the record and which starts to decompose at 425°C (Weliky *et al.* 1983).

Untreated bulk carbonates were analysed for the oxygen and carbon isotopic composition ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) following Siegenthaler & Eicher (1986). The cores were analysed at 2–6 cm intervals. Samples of 0.5 to 10 mg were digested in phosphoric acid at 90°C and analysed on an Isocarb-SIRA-24 mass spectrometer. The analytical precision of the internal standard was 0.1‰ for $\delta^{18}\text{O}$ and 0.05‰ $\delta^{13}\text{C}$. Oxygen isotope records from bulk carbonates are influenced by a range of environmental conditions (e.g. Leng & Marshall 2004). Nevertheless, variations in lake marl $\delta^{18}\text{O}$ records often reflect climatic changes (Eicher 1987; Lotter *et al.* 1992; Yu & Eicher 1998) that, for the Lateglacial period, can be matched to the Greenland ice-core $\delta^{18}\text{O}$ records (e.g. Schwander *et al.* 2000). Distortion of the isotopic signal owing to the input of detrital carbonates (Kolstrup & Buchardt 1982; Ammann 2000; Magny *et al.* 2006) is not expected for our record, as no calcareous bedrock is

present in the area (Hoek & Bohncke 2001), while the minerogenic fraction of the lake deposits consists of decalcified aeolian sand.

Pollen samples (0.3 cm³) were obtained at 2–8 cm intervals from the same levels as the isotope samples. Pollen samples were treated following the standard methods described by Faegri & Iversen (1989) and mounted in glycerine jelly. To estimate pollen concentrations, a known amount of *Lycopodium* spores was added to the samples. Pollen types were identified using Beug (2004), Faegri & Iversen (1989), Moore *et al.* (1991) and the NEPF Vols I–VIII (Punt 1976; Punt & Blackmore 1991; Punt & Clarke 1980, 1981, 1984; Punt *et al.* 1988, 1995, 2003). A pollen sum of at least 300 was targeted. For the pollen sum, pollen of local vegetation (including Cyperaceae) and of the thermophilous trees *Corylus*, *Quercus*, *Tilia*, *Ulmus* and *Alnus* were excluded. Pollen zones were visually distinguished based on changes in the main pollen taxa.

As terrestrial plant macrofossils are usually scarce in these types of lacustrine deposits, and the regional biostratigraphy is well known, the Weerterbos sequences have not been radiocarbon-dated before. The chronostratigraphy of these records has been established based on correlation of the pollen records with the biostratigraphical framework of Hoek (1997a) (Table 2). From both the KV and the GV sequence we managed to obtain some terrestrial macrofossil remains for AMS radiocarbon dating to support this general chronostratigraphy. Samples from the KV record were dated at the Centre for Isotope Research, University of Groningen, and samples from the GV record at the Leibniz Laboratory for Radiometric Dating and Isotope Research, Christian-Albrechts-Universität, Kiel. These radiocarbon dates were calibrated with the program OXCAL 4.1 (Bronk Ramsey 2009) and the INTCAL09 calibration curve (Reimer *et al.* 2009). The identification of Lateglacial tephra layers in the sequences could potentially provide an additional age control for these records; tephra shards of the Vedde Ash have been

Table 2. Lateglacial and early Holocene pollen assemblage zones (PAZs) and biostratigraphical zonation scheme for the Netherlands.

PAZ ¹	Biostratigraphy ²	¹⁴ C age onset (a BP)
B	Boreal	9150
5	Late Preboreal	9500
4	Preboreal	10 150
3	Younger (Late) Dryas	10 950
2b	Pine phase Allerød	11 250
2a	Birch phase Allerød	11 900
1c	Older (Early) Dryas	12 100
1b	Bølling	12 450

¹Ages of PAZ 1–5 are derived from Hoek (1997a), whereas the age for the onset of zone B is from van Geel *et al.* (1981).

²van Geel *et al.* (1981, 1989).

found in the northeastern Netherlands (Davies *et al.* 2005). Therefore, the BV and VG records, which contain deposits of Younger Dryas age, were sampled for tephra analysis. However, no tephra shards were recovered from the sequences.

Chironomid analysis was performed at 2–12 cm intervals. Samples of 1–2 cm³ were taken from the same levels as the isotope and pollen samples. Pre-treatment consisted of heating the samples to 90°C in a 5% KOH solution, followed by successively sieving through a 212- μ m and a 90- μ m mesh. Head capsules were then picked out under 40 \times magnification and mounted on microscope slides in Euparal©. Identification of the chironomid taxa was based on Wiederholm (1983), Moller Pillot (1984a, b), Rieradevall & Brooks (2001) and Brooks *et al.* (2007).

Chironomid-inferred July air temperatures were reconstructed using a chironomid–temperature inference model described by Heiri *et al.* (2011). This model is based on combined regional calibration data sets from Norway (Brooks & Birks 2000, 2001, unpublished data) and the Alpine region (Heiri *et al.* 2003; Heiri & Lotter 2005, 2010; Bigler *et al.* 2006). The calibration data set consists of chironomid assemblage data from 274 lakes that cover a July air temperature range from 3.5° to 18.4°C. A two-component weighted averaging partial least-squares (WA-PLS) regression was used to develop a chironomid–temperature inference model. The model has a root mean square error of prediction of 1.55°C and a coefficient of determination (r^2) of 0.84, as calculated using bootstrapping (9999 bootstrap cycles) (Heiri *et al.* 2011).

Performance of the model was evaluated by calculating for the fossil samples the closest modern analogue, the goodness-of-fit to temperature, and the percentage of taxa rare in the calibration data set (Birks *et al.* 1990, 2010; Birks 1998), following Heiri *et al.* (2007) and van Asch *et al.* (2012). All calculations, except the percentages of taxa rare in the calibration data set, were based on square-root-transformed percentages. The closest modern analogue of the fossil samples was assessed using squared Chi²-distances. Fossil samples with distances larger than the 2nd or 5th percentiles of all distances of modern samples in the calibration data set were considered to have ‘no close’ or ‘no good’ analogue, respectively. The goodness-of-fit was determined using a Canonical Correspondence Analysis (CCA) of the training set samples with temperature as the only constraining variable, into which the fossil samples were added passively. Fossil samples with a residual distance to axis 1 that exceeded the 90th or 95th percentile of all residual distances of the modern samples were classified as having a ‘poor’ or ‘very poor’ fit with temperature, respectively. Chironomid taxa with a Hill’s N2 below 5 were classified as rare in the modern calibration data set. Sample-specific errors were estimated for the fossil samples by bootstrapping. WA-PLS and squared Chi²-

distances were calculated using the program C2 (Juggins 2003). The CCA was executed with the program CANOCO version 4.51 (ter Braak & Šmilauer 2002).

Results

Results from the Klein Ven core

Lithology. – In the KV record, four lithological zones have been distinguished (KV-1–4) (Fig. 3). The basal deposits consist of sand (KV-1). The transition to the overlying gyttja and lake marl deposits is visible in core KV at a depth of 1.96 m. On top of the basal sand, ~10 cm of slightly organic gyttja is present, with an organic matter content of 6–8% (KV-2A), and a carbonate content of 0%. This is overlain by lake marl (1.85–1.07 m). The lake marl sequence can be divided into two sections. The first section (KV-2B: 1.85–1.48 m) is brownish grey with carbonate content ranging between 40 and 65%. In the upper section of the lake marl (KV-2C: 1.48–1.07 m) the carbonate content is higher, with values ranging between 60 and 85%. Furthermore, the colour of the deposits changes to light grey, and fluctuations in the carbonate content can be observed within the lake marl. On top of the lake marl deposits, an orange-brown layer with iron concretions (KV-3) is present. These concretions presumably result from postdepositional soil formation processes. The overlying clayey gyttja deposits (KV-4) indicate that lacustrine conditions re-established.

The Klein Ven pollen diagram. – The KV pollen diagram can be divided into distinct pollen assemblage zones, based on variations in the main pollen taxa (KV-p1–4) (Fig. 4; Table 3). KV-p1 (1.96–1.48 m) is characterized by relatively high percentages of Non-Arboreal Pollen (NAP) (~50%). The Arboreal Pollen (AP) consists mainly of *Betula* and *Salix*. Maximum Poaceae values of 55% are reached at the base of this zone, followed by a gradual decline. Herbaceous taxa, such as Asteraceae, *Galium*-type, *Helianthemum*, *Rumex* and *Thalictrum*, are also present in this zone. Furthermore, *Empetrum* pollen grains are sporadically recorded. The aquatic vegetation consists mainly of *Potamogeton*, which reaches maximum values of ~300%. In addition, the algae *Botryococcus* and *Pediastrum* reach high values. In the upper part of this zone (KV-p1B: 1.55–1.48 m), a temporary increase in *Salix* and decrease in *Betula* can be observed. Herbs, such as *Galium*-type and Rosaceae, also increase. In addition, Cyperaceae increase and reach maximum percentages of ~200% in this subzone, while the terrestrial pollen concentration decreases.

The transition to KV-p2 (at a depth of 1.48 m) is characterized by an increase in *Betula* pollen. Furthermore, this transition is marked by a decline of Cyperaceae, and the decreasing trend of Poaceae continues.

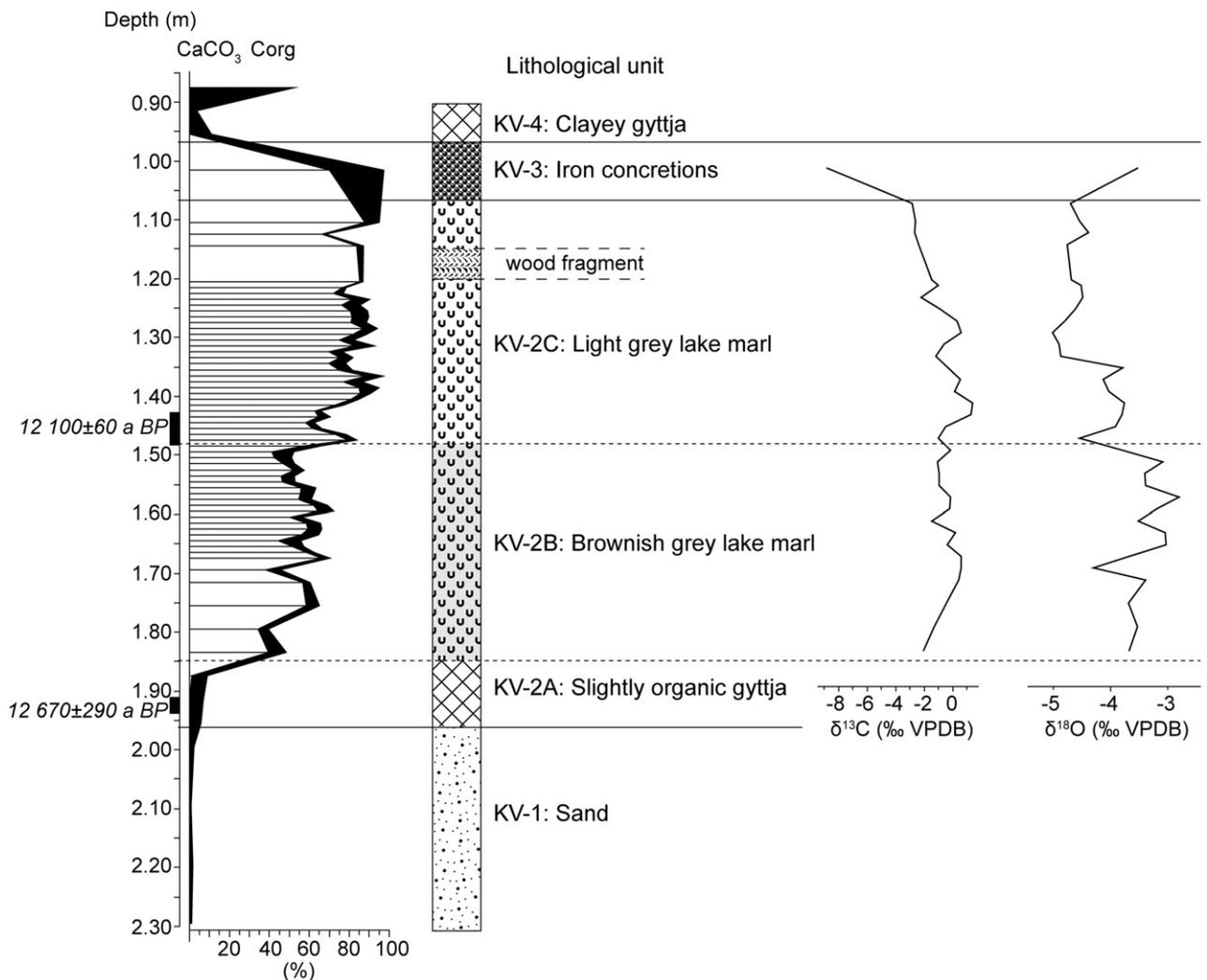


Fig. 3. Lithology, carbon and oxygen isotopes of the Klein Ven core. AMS radiocarbon dates from the core are also shown.

In addition, the algae *Pediastrum* and *Botryococcus* strongly decline at the transition to KV-p2. *Potamogeton* also strongly declines, while oospores of Characeae are more abundant in KV-p2 than in KV-p1. In the first part of this zone (KV-p2A: 1.48–1.25 m), maximum values of both *Juniperus* and *Artemisia* are reached, and *Salix* maintains relatively high values. This subzone is further marked by a number of oscillations in the percentages of the AP (mainly *Betula*). In addition, herbaceous taxa, such as *Galium*-type, *Rumex* and *Thalictrum*, persist. In the next subzone (KV-p2B: 1.25–1.07 m), *Betula* values reach a maximum of 75%, while *Salix*, *Juniperus* and the herbaceous taxa decline. *Nymphaea* appears in the aquatic vegetation.

The subsequent zone (KV-p3: 1.07–0.97 m) coincides with the layer of iron concretions (lithological unit KV-3). In this zone, *Betula* values drop, whereas *Pinus* increases. This coincides with an increase in *Empetrum* pollen. Furthermore, pollen from thermophilous trees, such as *Corylus* and *Alnus*, appears. The upper pollen

sample is located within the clayey gyttja deposits (KV-p4). This sample shows maximum values of *Pinus* pollen (~70%) and an increase in thermophilous taxa, such as *Corylus* and *Alnus*.

Age assessment of the Klein Ven record. – The pollen zones distinguished in the KV record are correlated with the general biostratigraphical zonation of Hoek (1997a), as shown in Table 3. KV-p1A reflects a relatively open landscape with local stands of tree birch, which can be correlated with the Bølling (PAZ 1b: 12 450–12 100 ^{14}C a). This is in good agreement with a basal ^{14}C date of 12 670 ± 290 ^{14}C a (Table 4). Subsequently, a temporary decline in *Betula* and expansion of *Salix* are observed in the KV record (KV-p1B). This temporary decline in *Betula* is often observed before the expansion of birch forest in the Netherlands and probably corresponds to the Older Dryas (PAZ 1c: 12 100–11 900 ^{14}C a), although this change is not very distinct in the KV record.

Table 3. Changes in the main pollen taxa in the KV record compared with the regional pollen assemblage zones of Hoek (1997a) as shown in Table 2. ↑ = increase; ↓ = decrease; NA = not available, see text for further explanation.

Local PAZ	Depth (m)	Main pollen taxa	Regional PAZ Hoek (1997a)
KV-p4	0.94–0.97	<i>Pinus</i> , <i>Corylus</i> , <i>Betula</i> <i>Pinus</i> ↑, <i>Betula</i> ↓, <i>Corylus</i> ↑	B
KV-p3	0.97–1.07	<i>Betula</i> , <i>Pinus</i> <i>Pinus</i> ↑, (<i>Empetrum</i> ↑, <i>Corylus</i> ↑)	NA
KV-p2B	1.07–1.25	<i>Betula</i> <i>Betula</i> ↑, <i>Juniperus</i> ↓, <i>Salix</i> ↓, NAP ↓	2a
KV-p2A	1.25–1.48	<i>Betula</i> , <i>Salix</i> , <i>Juniperus</i> <i>Betula</i> ↑, Poaceae ↓, <i>Juniperus</i> ↑, Cyperaceae ↓	2a
KV-p1B	1.48–1.55	Poaceae, <i>Salix</i> , <i>Betula</i> , Cyperaceae <i>Salix</i> ↑, <i>Betula</i> ↓, Cyperaceae ↑	1c
KV-p1A	1.55–1.96	Poaceae, <i>Betula</i> , <i>Salix</i>	1b

The onset of zone KV-p2 is marked by a strong rise in *Betula*, which appears to reflect the onset of the Allerød (PAZ 2a; Hoek 1997a), even though subzone KV-p2A could alternatively be considered as part of the Older Dryas, based on the ^{14}C date of $12\,100 \pm 60$ ^{14}C a for the base of this subzone (Table 4). During subzone KV-p2A birch values fluctuate, and the relatively high percentages of *Juniperus*, *Salix* and herbaceous taxa, such as *Artemisia*, *Galium*-type and *Thalictrum*, indicate that these taxa were part of the vegetation. At the transition to subzone KV-p2B, *Juniperus*, *Salix* and the herbaceous taxa strongly decline, and birch woodland developed.

The overlying layer with iron concretions cannot be directly correlated with the general biostratigraphical zonation of Hoek (1997a). The observed increase in *Pinus* pollen coincides with an increase in *Empetrum*

and thermophilous taxa. It therefore seems unlikely that these deposits correspond to the pine phase of the Allerød (PAZ 2b). The pollen spectrum from the overlying gyttja deposits points to a landscape that was dominated by pine forest, while hazel and birch trees were also present. Furthermore, low percentages of well-preserved pollen of the thermophilous trees *Quercus*, *Ulmus* and *Alnus* are recorded. This indicates a Boreal age (younger than 9150 ^{14}C a) for these gyttja deposits. Thus, a hiatus seems to be present in the record, encompassing the later part of the Allerød to the Boreal. It appears that soil formation during this period resulted in the underlying layer with iron concretions. This can also explain the observed mixture of pollen of *Empetrum* and *Corylus* in this unit characteristic of the Younger Dryas and the end of the Late Preboreal biozones, respectively (Hoek 1997a).

Table 4. AMS radiocarbon dates based on selected plant macrofossils of the Klein Ven and Groot Ven cores.

Site	Depth (m)	Laboratory code	^{14}C age (a BP)	Calibrated age (cal. a BP $\pm 1\sigma$)	Analysed material
Klein Ven	1.43–1.48	GrA 48441	$12\,100 \pm 60$	13 858–14 024	<i>Betula</i> leaf fragments, twigs; 1 <i>Carex</i> cf. <i>rostrata</i> fruit; 3 <i>Scirpus lacustris</i> fruits
Klein Ven	1.91–1.94	GrA 48414	$12\,670 \pm 290$	14 207–15 500	1 <i>Betula nana</i> male catkin scale, 1 female catkin scale, leaf fragment; <i>Salix</i> twig fragments; <i>Salix/Betula</i> leaf fragments; 1 <i>Eleocharis palustris/lunigumis</i> seed, 2 <i>Carex</i> cf. <i>aquatilis</i> fruits
Groot Ven	0.725–0.75	KIA 11105	$10\,970 \pm 80^1$	12 703–12 942	<i>Betula</i> 4 fruits, 1 fruit scale, leaf fragments; 1 <i>Salix</i> catkin; moss fragments
Groot Ven	1.03–1.05	KIA 11106	$12\,040 \pm 90$	13 795–13 992	5 <i>Betula</i> fruits; 1 <i>Carex</i> sp. fruit; moss fragments
Groot Ven	1.27–1.29	KIA 11107	$14\,740 \pm 220^2$	17 638–18 475	<i>Betula</i> 1 fruit, 5 bud scales; <i>Betula nana</i> leaf fragments; moss fragments
Groot Ven	1.45–1.47	KIA 11108	$11\,810 \pm 310^3$	13 317–14 015	<i>Betula</i> leaf fragments; 2 <i>Carex rostrata</i> fruits; 2 <i>Carex nigra</i> fruits; 4 <i>Juncus</i> seeds; moss fragments

¹Date is considered too young, which may be explained by root penetration and downward displacement of younger material resulting from postdepositional soil formation processes.

²Date is considered too old; plant remains are possibly reworked.

³Date is considered too young, which may be the result of the small sample size or could indicate contamination with younger material.

Oxygen and carbon isotopes of the Klein Ven core. – The basal sand and overlying organic gyttja, as well as the clayey gyttja at the top of the sequence, did not contain measurable concentrations of carbonates. Consequently, no isotopic analyses were attempted for these parts of the sequence. The oxygen isotopic composition of the lake marl ranges between \sim –3 and \sim –5‰ (Fig. 3). At the base of the lake marl deposits, a gradual increase can be observed. Maximum values of \sim –3‰ are reached between 1.66 and 1.50 m. Subsequently, $\delta^{18}\text{O}$ values decrease to \sim –4‰ between 1.50 and 1.35 m. Then, $\delta^{18}\text{O}$ values continue to decrease, and minimum values of \sim –5‰ are reached at a depth of 1.30 m. Values remain low towards the top of the lake marl deposits. In the top sample (at a depth of 1 m), values again increase to \sim –3.5‰. This sample is located in the layer with a high abundance of iron concretions (KV-3; Fig. 3).

Carbon isotopic composition of the lake marl ranges between +1.3 and \sim –3‰ (Fig. 3). No significant changes are observed in the lake marl deposits, although minor fluctuations in the $\delta^{13}\text{C}$ values are recorded. In the layer containing iron concretions, $\delta^{13}\text{C}$ drops to \sim –9‰. The distinct change in the oxygen and carbon isotope records observed in the layer with iron concretions is probably the result of postdepositional soil formation processes.

Chironomid assemblages of the Klein Ven record. – Head capsules are abundant throughout the core (Fig. 5),

although head capsule preservation is poor in the upper part of the lake marl owing to soil formation. In the organic gyttja at the base of the lake infill (lithological unit KV-2A), the chironomid assemblage is dominated by *Chironomus plumosus*-type, which frequently is an early colonizer after large-magnitude environmental changes in Norway (Brooks *et al.* 2007). Other important taxa (>5%) include *Paratanytarsus penicillatus*-type, *Tanytarsus pallidicornis* type-1 and *Corynoneura edwardsi*-type. In the overlying lake marl (lithological unit KV-2B), *C. plumosus*-type strongly decreases. This zone is dominated by *Tanytarsus glabrescens*-type, *T. pallidicornis* type-2 and *Microtendipes pedellus*-type. At the top of this unit, a temporary decrease of *T. glabrescens*-type can be observed, while *Parakiefferiella bathophila*-type, *Paratanytarsus penicillatus*-type and *Einfeldia pagana*-type increase. In the overlying part of the lake marl (KV-2C), *M. pedellus*-type declines, whereas *T. glabrescens*-type and *T. pallidicornis* type-2 remain abundant. Other taxa increase in this zone, such as *Ablabesmyia*, *Tanytarsus lactescens*-type and *Psectrocladius*. An expansion of *Dicrotendipes nervosus*-type can be seen towards the top of this unit.

Chironomid-inferred mean July air temperature reconstruction for the Klein Ven record. – The reconstructed chironomid-inferred mean July air temperatures (C-IT) of Klein Ven range between \sim 15° and \sim 18°C (Fig. 6), although most samples vary between

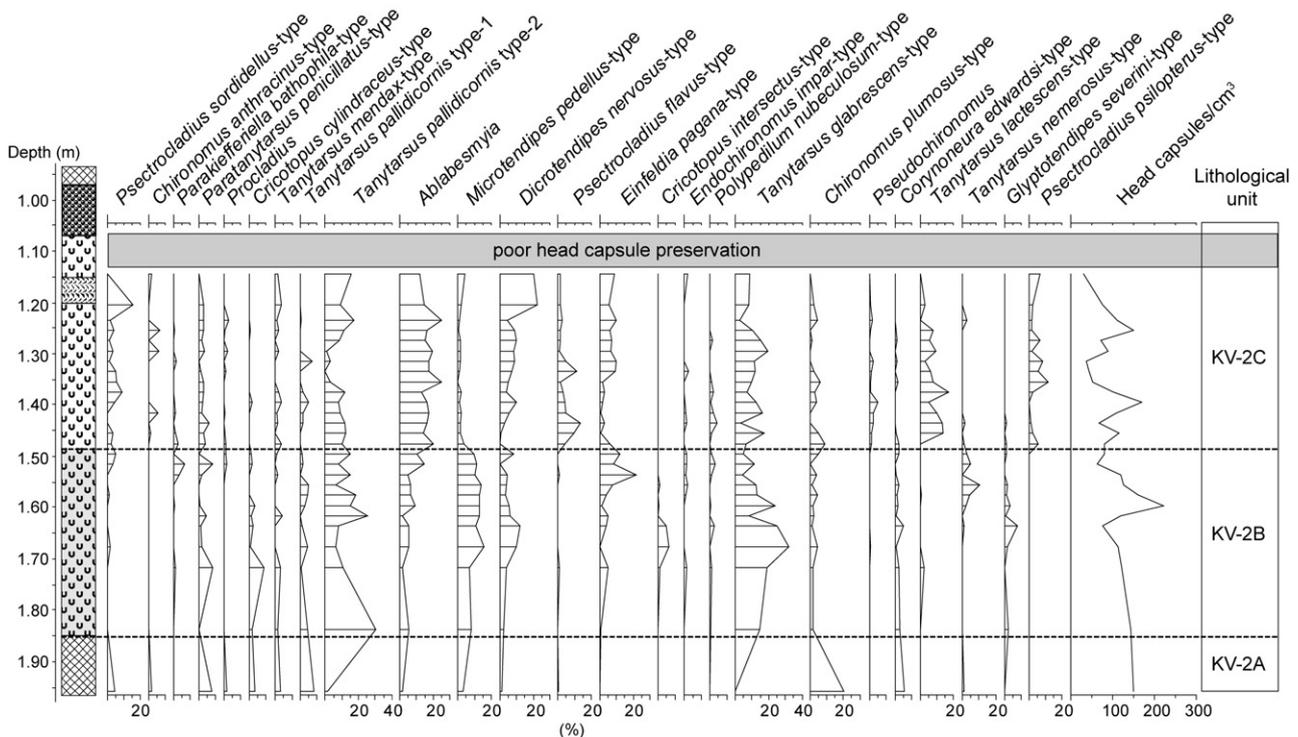


Fig. 5. Percentage diagram of the main chironomid taxa of the Klein Ven core. Taxa are arranged according to their optimum temperature in the calibration data set (Heiri *et al.* 2011), with taxa having the coolest optimum temperature shown on the left.

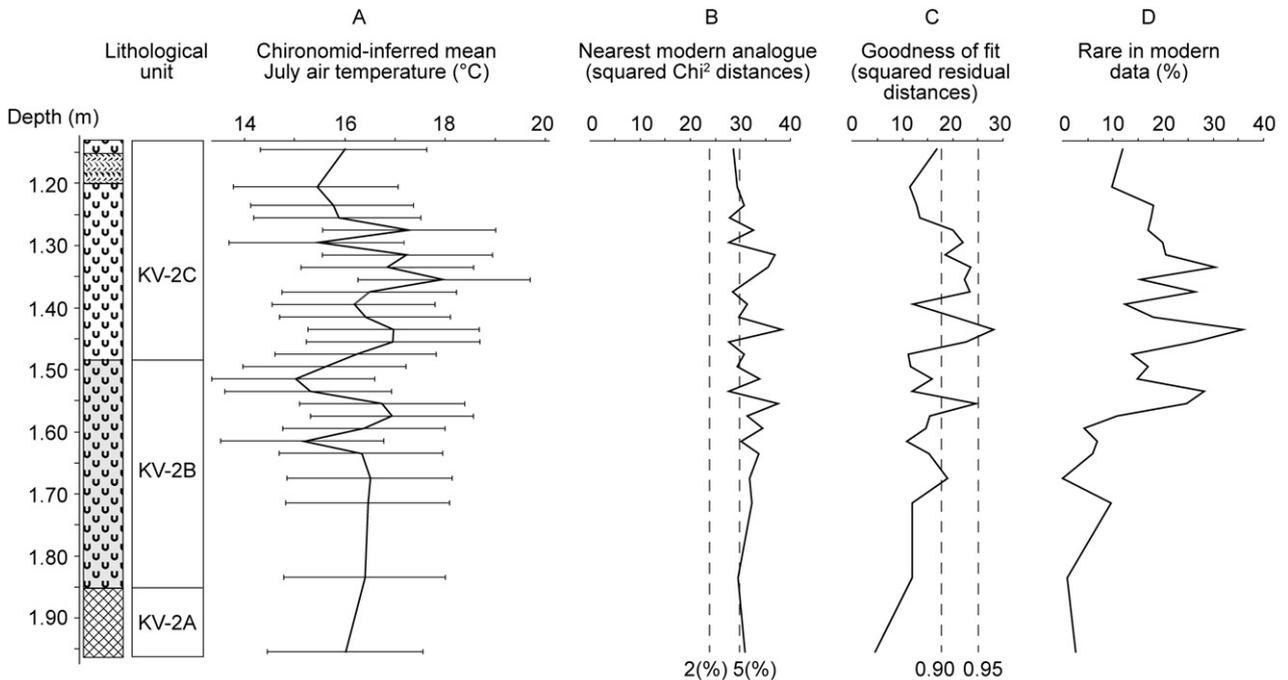


Fig. 6. Chironomid-inferred mean July air temperature reconstruction of the Klein Ven core with sample-specific prediction errors (A), nearest modern analogues for the fossil assemblages in the calibration data set (B), goodness-of-fit of the fossil assemblages with temperature (C), and percentage of chironomid taxa in the fossil samples that are classified as rare in the calibration data set ($N_2 < 5$) (D). Dashed lines in (B) indicate the 2nd and 5th percentiles of all squared χ^2 distances of the modern samples in the calibration data set, which are used to identify fossil samples with 'no close' or 'no good' modern analogues, respectively. Dashed lines in (C) show the 0.90 and 0.95 quantiles of all residual distances of the modern samples to axis 1 in a CCA with temperature as the only constraining variable, which are used to identify fossil samples with a 'poor' or 'very poor' fit with temperature, respectively.

15° and 17°C. A gradual increase in C-IT is visible in the basal part of the lake marl. This is followed by a cold oscillation between 1.55 and 1.48 m, with a decrease in reconstructed temperatures of ~1.5°C. Furthermore, a gradual decreasing trend is visible towards the top of the profile, which starts at a depth of ~1.26 m.

The size of the sample-specific errors of the KV record varies between 1.51° and 1.70°C, which is comparable with or slightly larger than the overall root mean square error of prediction (RMSEP) of the transfer function of 1.55°C. The record consists for a large part of taxa that are well represented in the calibration data set. The percentage of taxa rare in the modern calibration data varies from 0 to 35%, with higher values being reached in the upper half of the lake marl sequence. All chironomid samples from the KV record consist of assemblages that have no close or good analogue in the calibration data set. However, WA-PLS regression is able to perform relatively well in no-analogue situations (Birks 1998). In comparison, only one sample has a very poor fit with temperature, although the residual distance of this sample is only slightly higher than the 95th percentile of the residual distances of the samples in the modern calibration data set, which was used as the cut-off level to identify samples with a very poor fit.

Inter-site comparison

Palynological records from the Weerterbos region. – The four pollen records from the Weerterbos region all reflect the development of birch forest during the Lateglacial Interstadial, although substantial differences can be observed between the records (Fig. 7). The Bølling section (PAZ 1b) of the GV record is largely comparable to the Bølling section in the KV record, with *Betula* values ranging between ~30 and 50%. In contrast, in the BV and VG records *Betula* values rapidly increase to 50 and 60%, respectively, and the presumed Bølling deposits are only relatively thin in these records. Furthermore, these two records show relatively high *Salix* values at the base, which probably reflects the local presence of *Salix* during the formation of these lakes.

A subsequent dip in the *Betula* percentages and an increase in *Salix*, comparable to KV-p1B (presumed Older Dryas biozone, PAZ 1c), are visible in the BV and VG records. No clear regressive phases can be discerned in the GV record, although a minor dip in the percentages of *Betula* and an increase in *Salix* can be observed at a depth of ~1.18 m, which may correspond to KV-p1B. In the BV and VG records, this temporary decline in birch is followed by a distinct increase in *Betula* values up to ~60 and ~75%, respectively. This

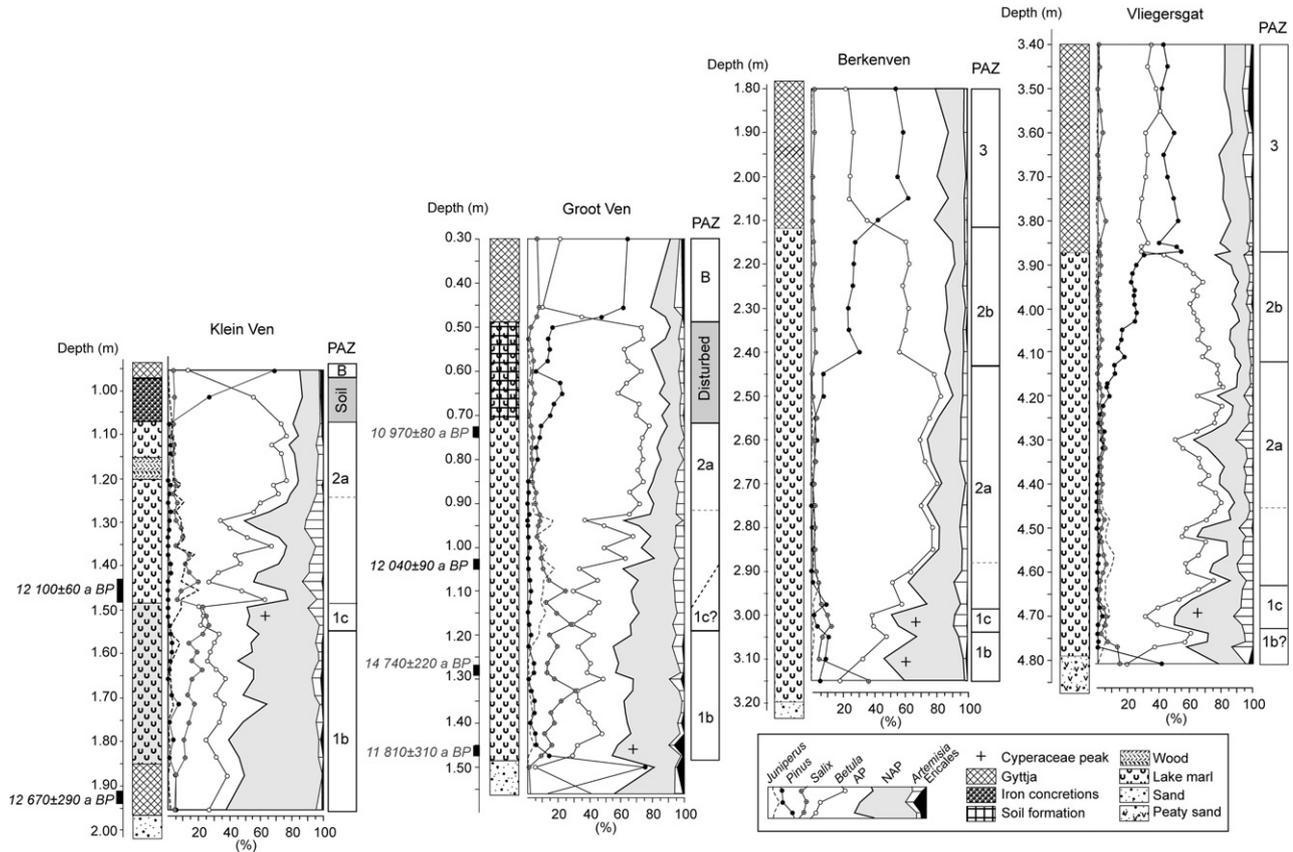


Fig. 7. Comparison of the lithology and main pollen taxa of cores KV, GV, BV and VG. GV has previously been published by Hoek & Bohncke (2001). These records have been correlated with the Lateglacial and early Holocene biostratigraphical zones (PAZs; Table 2). The position where both *Salix* and *Juniperus* diminish during zone 2a is indicated in the PAZ columns as a horizontal dashed grey line, which may be synchronous at the four sites (see text for further explanation). Radiocarbon dates of the KV and GV records are also shown; dates in grey are considered unreliable (Table 4).

presumably marks the onset of the Allerød (PAZ 2a), and is comparable to the transition to KV-p2 in the KV record. In contrast, no distinct increase in *Betula* values is discerned in the GV record, although birch values gradually increase. The increase in *Betula* values in the KV record is followed by a zone with distinct fluctuations in the *Betula* percentage (KV-p2A). This subzone is further characterized by maximum *Juniperus* values and the presence of *Salix* and herbaceous taxa. Similar subzones are visible in the BV and VG records, although fluctuations in the *Betula* percentages are less pronounced in these records. In the BV and VG records, a subsequent decline in *Salix* and *Juniperus* can be observed at a depth of ~2.88 and ~4.45 m, respectively. In the GV record, a decline of *Salix*, *Juniperus* and herbaceous taxa is visible at a depth of ~0.92 m.

In the BV and VG records, the expansion of birch woodland is followed by an expansion of pine that corresponds to the pine phase of the Allerød (PAZ 2b). Subsequently, these two records indicate an opening of the vegetation cover during the Younger Dryas (PAZ 3). An expansion of pine can also be observed in the

upper part of the GV record, which may correspond to the pine phase of the Allerød, although pollen of thermophilous tree taxa are also recorded in this section (Hoek & Bohncke 2001). Furthermore, this section shows indications of soil formation. Root penetration and the downward displacement of younger material associated with soil-forming processes can also explain the relative young date at the top of the GV sequence (Table 4).

Oxygen isotope records from the Weerterbos region. – Oxygen isotope records from the Weerterbos region vary between ~-2 and -7‰ (Fig. 8), comparable to the Gulickshof record (Hoek & Bohncke 2001). Nevertheless, comparison of the four Weerterbos records clearly shows significant differences in $\delta^{18}\text{O}$ values, trends and fluctuations. The GV and KV records, which are located within ~100 m of each other, are largely similar, although the basal and top sections vary significantly between the two records (Fig. 8A). The other two records are distinctly different from the KV and GV records. Furthermore, the strongest fluctuations are

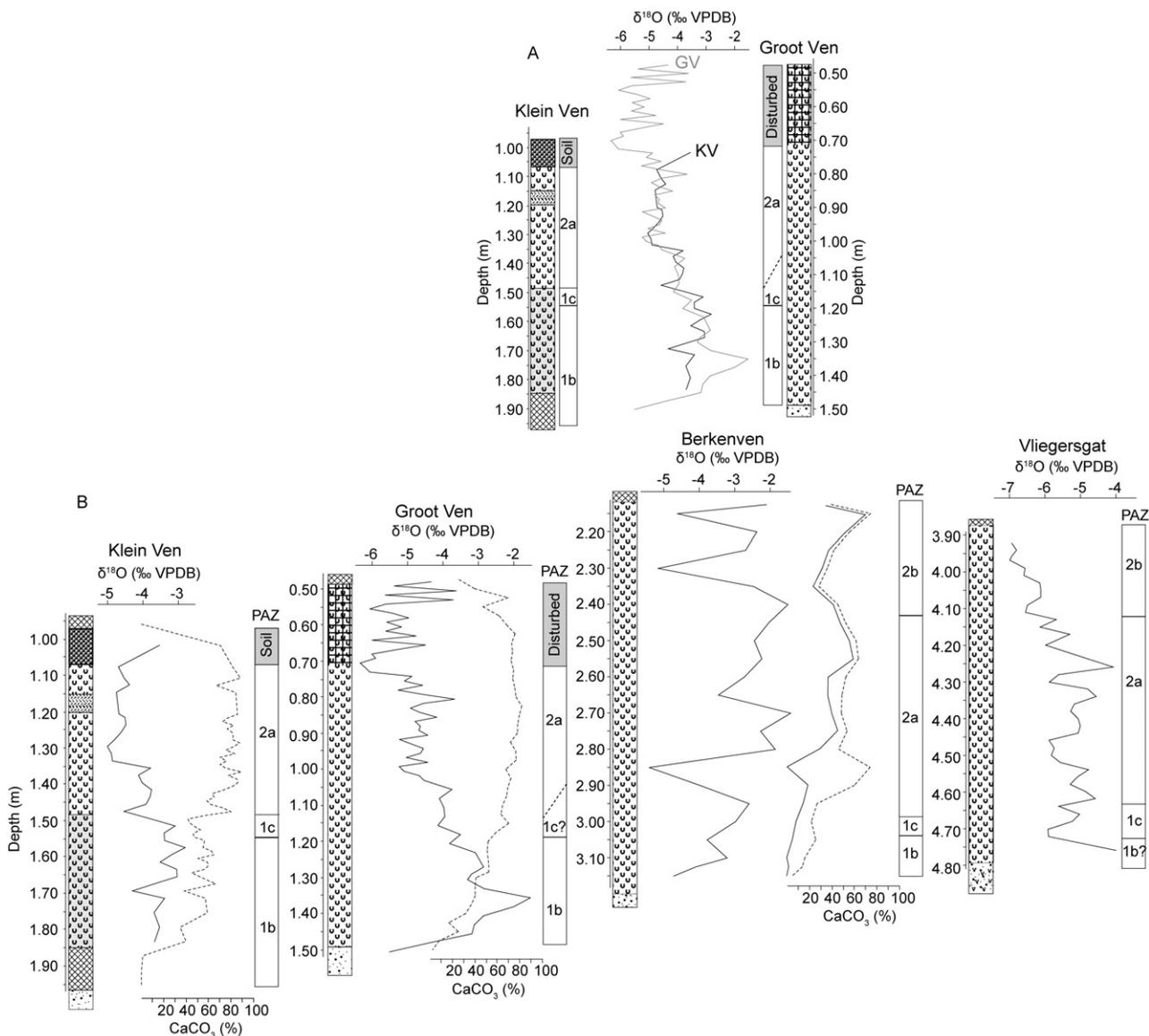


Fig. 8. A. Comparison of the KV and GV oxygen isotope records. B. Carbonate content, $\delta^{18}\text{O}$ and biostratigraphical zones of KV, GV, BV and VG. The siderite content of the BV core was also determined (solid line), and is shown in addition to total carbonates (expressed as CaCO₃; dashed line). Carbonate content results were not available for core VG.

recorded in the BV record. For example, this record shows a negative shift of more than 3‰ at a depth of 2.85 m. For this record, a distinction has been made between siderite (FeCO₃) and other carbonates (presumably CaCO₃). It is apparent that the negative shift in the $\delta^{18}\text{O}$ record coincides with a decrease in the siderite content.

Discussion

Factors influencing the oxygen isotope records

As the strong negative shift in the BV $\delta^{18}\text{O}$ record coincides with a decrease in the siderite content (Fig. 8), it

appears that the isotope shift is caused by differences in isotopic fractionation between calcite (Kim & O'Neil 1997) and siderite (Zhang *et al.* 2001). It seems plausible that variations in the relative importance of siderite also caused isotopic variation in the other records. In addition, differences in the importance of biogenic and inorganic carbonates possibly led to changes in the isotopic composition of bulk carbonates (Leng & Marshall 2004). Furthermore, it seems that Characeae were more abundant in the upper part of the Klein Ven lake marl sequence than in the lower part. This may also explain part of the isotopic variation in the record, as it appears that *Chara* encrustations have a more negative isotopic composition than shells (Hoek *et al.* 1999).

Evaporation is another factor influencing the isotopic composition of water in these small lakes. Relatively more evaporation would have taken place in the shallowest KV and GV lakes. Consequently, these two records will be more isotopically enriched than the BV and VG records owing to the preferential evaporation of the lighter isotopes. Furthermore, carbonate-rich groundwater entering the lakes may already have had a $\delta^{18}\text{O}$ composition that deviates from rainfall $\delta^{18}\text{O}$, thereby further complicating the identification of a climatic signal from the oxygen isotope record.

This comparison of four oxygen isotope records from the Weerterbos region shows that it is difficult to deduce a common regional signal, as the records strongly differ in $\delta^{18}\text{O}$ values, trends and fluctuations. The results imply that the isotopic signal in these small lakes and ponds is dominated by the effects of local environmental conditions, such as groundwater influx, aquatic vegetation and evaporation, rather than directly reflecting the $\delta^{18}\text{O}$ of past of precipitation.

Chironomid assemblages, environmental and temperature changes

The chironomid assemblages found in core KV are indicative of warm to intermediate temperatures and mesotrophic to eutrophic conditions (Brooks *et al.* 2007). These taxa are currently absent from the coldest alpine and arctic lakes (Brooks & Birks 2000; Heiri & Lotter 2010; Heiri *et al.* 2011). A number of taxa present in the KV record are predominantly found in lowland lakes in the Alpine region, for example *Tanytarsus glabrescens*-type, *Tanytarsus lactescens*-type, *Chironomus plumosus*-type, *Polypedilum nubeculosum*-type, *Glyptotendipes* and *Pseudochironomus* (Lotter *et al.* 1997; Heiri & Millet 2005; Heiri *et al.* 2007; Heiri & Lotter 2010). Other taxa are currently found over a wide temperature range, such as *Ablabesmyia*, *Paratanytarsus penicillatus*-type and *Psectrocladius sordidellus*-type (Lotter *et al.* 1997; Brooks & Birks 2001; Heiri & Lotter 2010).

The chironomid assemblages of lithological unit KV-2B (brownish grey lake marl) are dominated by *Tanytarsus glabrescens*-type, *T. pallidicornis* type-2 and *Microtendipes pedellus*-type, while KV-2C (light grey lake marl) is marked by a strong increase in *Psectrocladius*, *Ablabesmyia* and *Tanytarsus lactescens*-type (Fig. 5). The increase of *T. lactescens*-type and decrease of *M. pedellus*-type may be related to changes in lithology (increase in the carbonate content), as the former is often abundant in carbonate lakes, while the latter is associated with silty sediments (Brooks *et al.* 2007). Throughout the record, chironomid taxa occur that are often associated with macrophytes, such as *Ablabesmyia*, *Psectrocladius*, *Paratanytarsus*, *Dicrotendipes* and *Corynoneura* (Brodersen *et al.* 2001; Brooks *et al.* 2007; Langdon *et al.* 2010). Variations in the chirono-

mid assemblages between lithological units KV-2B and KV-2C, such as the increase of *Ablabesmyia* and multiple *Psectrocladius* taxa in unit KV-2C, may be related to changes in the aquatic vegetation, as observed in the pollen record (Fig. 4). It therefore appears that the change in the chironomid fauna between lithological units KV-2B and KV-2C was not driven by temperature. Consequently, it seems plausible that average chironomid-inferred temperatures for lithological units KV-2B and KV-2C are not substantially different (-16.0° – 16.5°C).

A cold oscillation is inferred from the KV chironomid record between 1.55 and 1.48 m (Fig. 6). In this section, no pronounced changes occur in the abundance of chironomid taxa that are associated with macrophytes. The reconstructed temperature decrease coincides with a large increase in Cyperaceae pollen, which points to a concomitant decrease in lake water depth (see also next section). This inferred decrease in lake water depth could also have affected the chironomid fauna. If, however, this fluctuation was associated only with drier conditions, this should lead to an apparent increase in chironomid-inferred air temperatures, as a decrease in lake water depth would probably result in an increase in lake water temperature, and hence in an increase in chironomid-inferred air temperatures. We therefore postulate that the inferred change to drier conditions coincided with a temporary decrease of summer temperatures, as inferred from the chironomid record.

The KV chironomid-inferred temperatures are similar to a chironomid-inferred temperature reconstruction from Hijkermeer in the northern part of the Netherlands (Heiri *et al.* 2007, 2011). For both chironomid records, reconstructed Interstadial July temperatures vary predominantly between $\sim 15^{\circ}$ and $\sim 17^{\circ}\text{C}$. Slightly higher temperatures are reached in the KV record ($\sim 0.5^{\circ}\text{C}$) than in the Hijkermeer record, which may reflect a north–south temperature gradient, similar to the present-day temperature gradient (KNMI 2011). Reconstructed temperatures are also similar to beetle-inferred summer temperatures of $\sim 15^{\circ}\text{C}$ for the early part of the Interstadial in the central part of the Netherlands, and to slightly higher temperatures of $\sim 18.5^{\circ}\text{C}$ for a site in Belgium (Coope *et al.* 1998). Furthermore, Renssen & Isarin (2001) reconstructed similar minimum mean July temperatures of $\sim 14^{\circ}$ – 15°C during the early part of the Interstadial for the Netherlands, based on palaeobotanical evidence. As the KV chironomid-inferred temperatures are consistent with previous results, and especially with the Hijkermeer chironomid-inferred temperatures, it seems possible to reconstruct plausible chironomid-inferred summer temperatures from a small depression, such as Klein Ven, using a calibration data set that is based on generally larger and deeper lakes (Lotter *et al.* 1997; Brooks & Birks 2000).

Temperature and environmental changes during the Lateglacial Interstadial in the Weerterbos region

A comparison of the multiple proxies in the KV record shows similarities in the general trends of the KV total arboreal pollen, carbonate content and C-IT, while the trend of the $\delta^{18}\text{O}$ record strongly differs from the other records (Fig. 9). The Weerterbos $\delta^{18}\text{O}$ records appear to be predominantly influenced by local environmental conditions (see above). Therefore, the $\delta^{18}\text{O}$ records are not discussed further.

The Klein Ven basin formed during the Bølling, when landscapes were still relatively open, although local stands of birch were present. Chironomid-inferred mean July temperatures of $\sim 16.0^\circ\text{--}16.5^\circ\text{C}$ are reconstructed for this period (Fig. 9). The presumed Bølling deposits are relatively thin in the BV and VG records, which suggests that these two lakes formed in the later part of the Bølling. High *Betula* values at the base of the VG sequence, similar to those in the nearby site Mierlo Ven Hoenderboom (Zagwijn 1971; Hoek 1997b; Fig. 1), may also indicate that this basin started to fill in the later part of the Bølling. The later formation of the slightly deeper BV and VG basins, compared with

the KV and GV basins, can be explained by progressively deeper melting of the permafrost, as a result of which lakes with a larger depth formed later.

A temporary decline in *Betula* pollen in the KV record (KV-p1B) coincides with a minor decrease in carbonate content (Fig. 9). Comparable fluctuations are identified in the other three Weerterbos pollen records (Fig. 7), as well as in pollen records from Gulickshof (Hoek et al. 1999) and Milheeze (Bos et al. 2006) (Fig. 1). A temporary change to drier conditions can be inferred from a peak in Cyperaceae pollen in the KV record, which appears to reflect a decrease in lake level, resulting in an increase in the importance of riparian vegetation in the record. This event in the KV record was tentatively correlated with the Older Dryas, which is often considered to be a drier period (e.g. van Geel & Kolstrup 1978; Bohncke & Wilmstra 1988; Bohncke 1993). The temporary regressive vegetation phase coincides with a temporary decrease in the reconstructed chironomid-inferred temperatures in the KV record (Fig. 9). This oscillation is similar to a cold oscillation in the Hijkermeer record from the northern Netherlands (Heiri et al. 2007, 2011). At both sites, a temporary temperature decline to $\sim 14.5^\circ\text{--}15^\circ\text{C}$

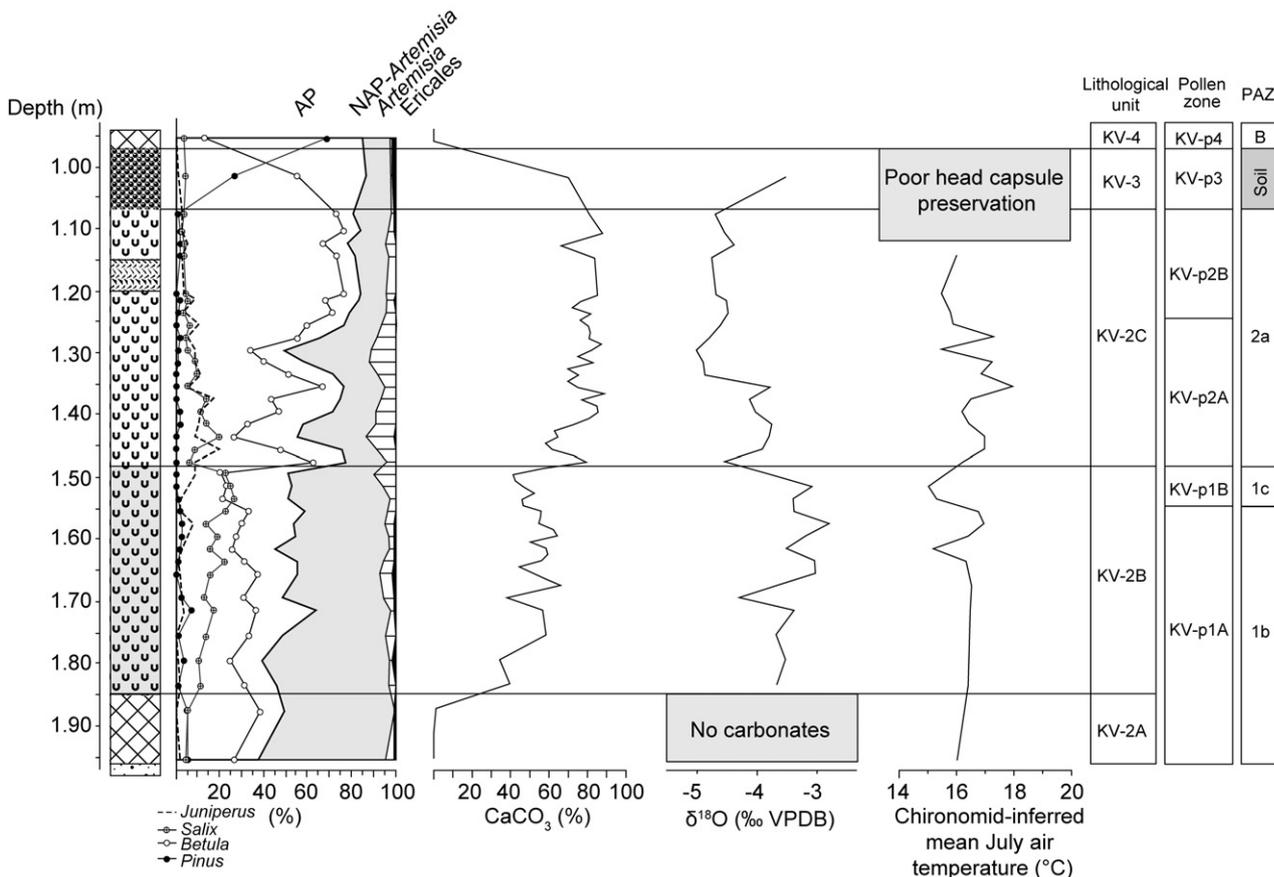


Fig. 9. Comparison of the main pollen taxa, carbonate content, carbonate $\delta^{18}\text{O}$ and chironomid-inferred July temperatures from the Klein Ven core.

coincides with maximum *Salix* values and a minor decrease in *Betula* percentages, and it precedes the expansion of birch forest and maximum *Juniperus* percentages in both records. This cooling might correspond to GI-1d (Björck *et al.* 1998; Lowe *et al.* 2008), although this cannot be confirmed by independent chronological data.

Reconstructed mean July air temperatures for this event well exceed the temperatures of $\sim 10^{\circ}\text{--}12^{\circ}\text{C}$ needed for the development of birch forest (Iversen 1954; Paus 1995). Therefore, it is more probable that the inferred change to drier conditions negatively affected birch, as both *B. pubescens* and *B. pendula* are intolerant of drought (Atkinson 1992), although a decrease in the growing season could also result in a temporary decline of the tree regeneration rate.

Although a regressive phase similar to the widely recognized Older Dryas is identified in all four Weerterbos records, substantial differences were observed among the four pollen records (Fig. 7). These differences reflect the influence of local vegetation (cf. Jacobson & Bradshaw 1981) on the pollen records from the small-scale Weerterbos depressions. For example, the presumed Older Dryas is least distinct in the records from the shallowest KV and GV depressions. In addition, all records show multiple fluctuations that cannot be confidently matched between the records. The presence of multiple fluctuations in these pollen records further emphasizes the difficulty in distinguishing the presumed Older Dryas biozone from local fluctuations in sediment cores obtained from small depressions. Furthermore, often only a limited number of radiocarbon dates are available for such records, and the errors associated with the radiocarbon dating of Lateglacial deposits impede the detailed correlation of short-lived regressive vegetation phases from the different pollen records. Despite these uncertainties, a direct comparison of the multiple proxies in the KV record suggests that the observed temporary regressive vegetation phase in the Weerterbos region was linked to climatological changes, instead of reflecting local environmental change, as it coincided with a decrease in the chironomid-inferred temperatures (Fig. 9).

Subsequent to zone KV-p1B, the KV record reflects the expansion of birch forest in the Weerterbos region, while reconstructed chironomid-inferred temperatures for this phase are comparable to those in the presumed Bølling section of the record. It therefore appears more plausible that the expansion of birch was facilitated by a change to wetter conditions, which may have been the result of an increase in the effective precipitation at the transition to the Allerød (e.g. Bohncke 1993; Hoek & Bohncke 2002; Bos *et al.* 2006). It appears that *Juniperus*, *Salix* and herbaceous taxa persisted in the first part of the Allerød in the Weerterbos region. It seems plausible that the decline of *Juniperus*, *Salix* and herbaceous taxa marks the development of birch woodland in the

area, and this is approximately synchronous in the four Weerterbos records (Fig. 7). Consequently, the decrease of *Salix* and *Juniperus* in the BV and VG records may also coincide with the transition to KV-p2B in the KV record, although this cannot be confirmed owing to a lack of chronologies in these records. The birch increase is followed by an expansion of pine, as indicated in the pollen records from the slightly deeper BV and VG basins that continued to receive sediment throughout the Lateglacial and early Holocene. Meanwhile, a decrease in lake level (Hoek & Bohncke 2002) and the resulting soil formation during the Preboreal probably caused the occurrence of a hiatus in the records from the shallower KV and GV records.

Conclusions

The combination of lithological, oxygen isotope, pollen and chironomid analyses of the Klein Ven sequence enabled a direct comparison of changes in temperature, vegetation and environmental conditions during the Lateglacial Interstadial. Furthermore, the comparison of four sequences from the Weerterbos region provided insight into local and regional factors influencing the oxygen isotope and pollen records. Local differences among the four isotope records presumably reflect variations in groundwater influx, aquatic vegetation and evaporation, thereby impeding the distinction of a regional palaeoclimatic signal. The four pollen records reflect the development of birch forest during the interstadial. However, chronological uncertainties and the influence of local vegetation patterns on pollen assemblages made it difficult to correlate these records. In contrast, reconstructed chironomid-inferred mean July air temperatures of $\sim 15^{\circ}\text{--}18^{\circ}\text{C}$ from the small Klein Ven basin seemed reliable, because these were similar to the Hijkermeer record from the northern Netherlands.

A temporary regressive vegetation phase is identified in the KV pollen record, which suggests a temporary shift to drier conditions, and possibly corresponds to the widely recognized Older Dryas. Despite variations among the pollen records from the Weerterbos region, a similar regressive phase can be discerned in all four records. Based on the KV chironomid-inferred temperature reconstruction, this phase also coincides with a temporary decrease of summer temperatures, which implies that this vegetation change was associated with climatological changes. These results suggest that a temporary cold dry oscillation preceded the development of open birch woodland during the Lateglacial in the southern Netherlands, which is similar to previous results from the northern Netherlands.

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