

# Technical Report

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### **The Last Interglacial-Glacial cycle (MIS 5-2) re-examined based on long proxy records from central and northern Europe**

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# Abstract

A comparison is made between five Late Pleistocene terrestrial proxy records from central, temperate and northern, boreal Europe. The records comprise the classic proxy records of La Grande Pile (E France) and Oerel (N Germany) and more recently obtained records from Horoszki Duże (E Poland), Sokli (N Finland) and Lake Yamozero (NW Russia). The Sokli sedimentary sequence from the central area of Fennoscandian glaciation has escaped major glacial erosion in part due to non-typical bedrock conditions. Multi-proxy studies on the long Sokli sequence drastically change classic ideas of glaciation, vegetation and climate in northern Europe during the Late Pleistocene. The focus of this review is on pollen, lithology and macrofossil- and insect-based temperature inferences. The long records are further compared with recent proxy data from nearby sites and with the fastly accumulating high-resolution proxy data from the ocean realm. The comparison allows a re-examination of the environmental history and climate evolution of the Last Interglacial-Glacial (LI-G) cycle (MIS 5-2). It shows that environmental and climate conditions during MIS 5 (ca. 130–70 kyr BP) were distinctly different from those during MIS 4-2 (ca. 70–15 kyr BP). MIS 5 is characterized by three long forested intervals, both in temperate and northern boreal Europe. These mild periods were interrupted by two relatively short cold and dry intervals (MIS 5d and 5b) with mountain-centered glaciation over Fennoscandia. Millennial scale climate events were superimposed upon these longer lasting climate fluctuations. MIS 4-2 shows open vegetation both in central and northern Europe. It includes two glacial maxima (MIS 4 and 2) with sub-continental scale glaciation over northern Europe and dry conditions in strongly continental eastern European settings. Climate oscillations of millennial scale dominated during MIS 3. Summer temperatures approaching present-day values are recorded for various warming events. Mild climate conditions in early MIS 3 at around ca. 50 kyr BP were accompanied by large-scale deglaciation of the Fennoscandian Ice Sheet. Ice-free conditions with *Betula*-dominated vegetation (including tree birch) persisted over large parts of Fennoscandia, possibly interrupted by glaciation, into the middle part of MIS 3 to ca. 35 kyr BP. Overall, MIS 5 was mostly mild with warmest or peak interglacial conditions at the very start during MIS 5e. MIS 4-2 was mostly cold with most extreme or peak glacial conditions in the closing phase during MIS 2. This points to a subdivision of the LI-G cycle into an early, overall mild half and a late, overall cold half, each with duration of ca. 60 kyr. The compilation favors a definition of MIS 5 as an interglacial complex similar as in the original marine oxygen isotope stratigraphy. Additionally, the reviewed data reveals much restricted ice cover during MIS 3 and indicates that climate variability during the LI-G cycle was mostly in terms of changes in degree of continentality possibly due to changes in sea ice cover.

# Contents

<b>1</b>	<b>Introduction</b>	7
<b>2</b>	<b>Methods</b>	9
<b>3</b>	<b>Late Pleistocene proxy records from central and northern Europe</b>	11
3.1	La Grande Pile (France)	11
3.1.1	Environmental reconstructions for MIS 5	11
3.1.2	Comparison with other records	14
3.1.3	Environmental reconstructions for MIS 4-2 and the Late Glacial-Holocene (MIS 1)	16
3.2	Oerel (Germany)	17
3.2.1	Environmental reconstructions for MIS 5	17
3.2.2	Comparison with other records	19
3.2.3	Environmental reconstructions for MIS 4-3	21
3.3	Horoski Duze (Poland)	21
3.3.1	Environmental reconstructions for MIS 5	23
3.3.2	Comparison with other records	24
3.3.3	Environmental reconstructions for MIS 4-3	24
3.3.4	Comparison with other stratigraphies/records	25
3.4	Sokli (Finland)	26
3.4.1	Environmental reconstructions for MIS 5	28
3.4.2	Comparison with other records	29
3.4.3	Environmental reconstructions for MIS 4-2	29
3.4.4	Comparison with other records	30
3.5	Lake Yamozero (Russia)	31
3.5.1	Environmental reconstructions for MIS 5	33
3.5.2	Environmental reconstructions for MIS 4-1	33
<b>4</b>	<b>Discussion</b>	35
4.1	Limitations of proxy records and the value of multi-proxy comparisons	35
4.1.1	Pollen records	35
4.1.2	Insect and macrofossil records	35
4.1.3	The sediment record	36
4.1.4	Comparison with marine records	37
4.2	Late Pleistocene environmental history and climate evolution of central and northern Europe	39
4.2.1	MIS 5 (ca. 130–70 kyr BP)	39
4.2.2	MIS 4-2 (ca. 70–15 kyr BP)	42
4.3	Stratigraphic implications	45
	<b>Acknowledgments</b>	47
	<b>References</b>	49

# 1 Introduction

Knowledge of past changes in climate and understanding of climate driving mechanisms and feedbacks are fundamental for the accurate prediction of future climate change. In the recent geological past, well-pronounced climate fluctuations occurred during the Last Interglacial-Glacial (LI-G) cycle between ca. 130 and 15 kyr ago (Marine Isotope Stages (MIS) 5-2). These Late Pleistocene climate fluctuations involved major re-organizations in the atmosphere, oceans, cryosphere and biosphere, allowing the build-up of large continental ice sheets at high northern latitudes and displacing vegetation belts over distances of up to several thousands of kilometers. Geological proxy-based climate records are important tools for unraveling past climate changes and are available from a large variety of natural archives, including the Greenland and Antarctic Ice Sheets, deep-sea sediments, and long terrestrial lake and loess sequences. During the last two decades, research has focussed on the peak interglacial warmth of MIS 5e, the millennial scale climate variability of MIS 3, the Last Glacial Maximum (LGM; MIS 2) and the Younger Dryas cooling. Knowledge of the climate evolution over an entire interglacial-glacial cycle, however, is fundamental for the understanding of the full range of climate drivers and feedback mechanisms.

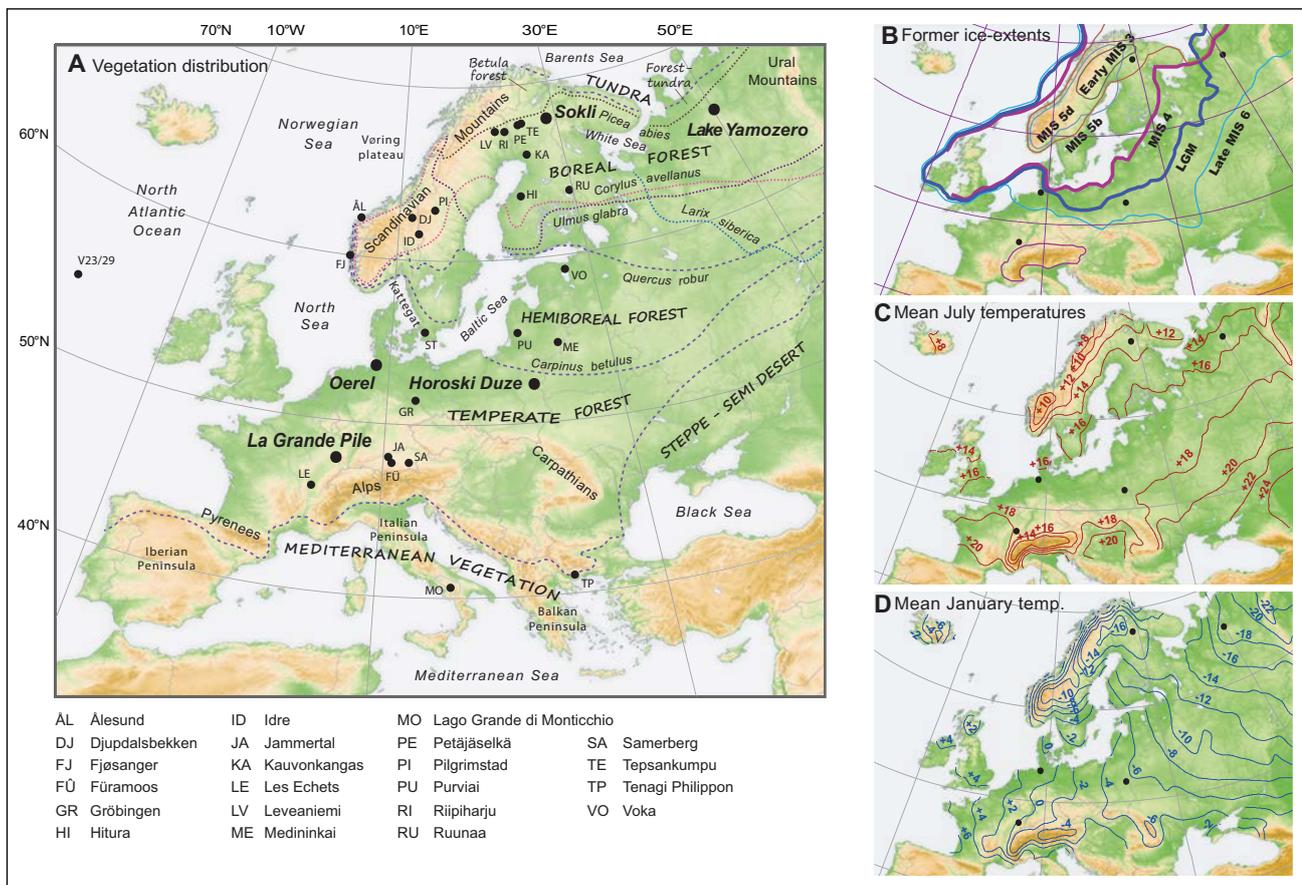
A wealth of proxy-based climate records for the LI-G cycle is available for Europe. The climate in Europe is furthermore under the influence of the North Atlantic Ocean, a region that has been identified as a major driving component in Late Pleistocene climate changes. However, long climate records that extend beyond the present interglacial period (Holocene) are scarce in the area of Fennoscandian glaciation in northern Europe as a result of glacial erosion. Here, highly fragmented, and often poorly-dated, bio(pollen)- and litho(till fabric)-stratigraphic sections have been correlated, and put in apparent stratigraphic sequence, in order to reconstruct climate changes during the Late Pleistocene (e.g. Mangerud 1991, Lundqvist 1992, Donner 1995). Long pollen records on the northern European mainland have been used as templates against which local and regional bio-stratigraphic schemas were correlated. As clearly outlined in Donner (1996), these bio-stratigraphic correlations are fraught with uncertainties, caused by e.g. the long distance of correlation (including an area with a great range of climatic conditions) and truncations of beds resulting in incomplete interstadial or interglacial sequences. In addition, bio-stratigraphic correlations were based on the assumption that Weichselian (MIS 5d-2) vegetation and climate gradients over central-northern Europe were similar as present or steeper, influenced by e.g. the presence of an ice sheet, and that conditions were similar or warmer than today only during the Eemian (MIS 5e). Furthermore, the reconstruction of the Late Pleistocene history of the Fennoscandian Ice Sheet (FIS) has been driven for an important part by correlation with the marine oxygen isotope record which was taken as a proxy for global ice volume. The poorly-dated stratigraphic evidence from Fennoscandia has hampered transect studies over Europe to be extended into high latitudes. These transect studies allow the exploration of climate driving and feedback mechanisms.

A for Fennoscandia unusual long and continuous sediments sequence covering the last 140 kyr has been found at Sokli in northern Finland (Helmens et al. 2000, 2007a; Figure 1-1). The sediments here have been preserved due to non-typical local bedrock conditions combined with limited glacial erosion under the central area of Fennoscandian glaciations (Helmens et al. 2007a, b). Lacustrine and fluvial sediment intercalations of early MIS 3 (around ca. 50 kyr BP) and MIS 5d-c age (ca. 110–90 kyr BP) have been subjected to detailed multi-proxy analyses. Similar studies are in progress on up to 9 m thick gyttja deposits of MIS 5e (around ca. 125 kyr BP) and Holocene age (last ca. 10 kyr) and a lacustrine intercalation dated to MIS 5a (ca. 85–75 kyr BP). Climate parameters are being inferred quantitatively using pollen, plant macrofossils and chironomid (aquatic insect) remains. Sokli was glaciated during MIS 5b, MIS 4 and late MIS 3/2.

Results from the Sokli studies have been surprising in various ways, drastically changing classic concepts of glaciations, vegetation and climate in Fennoscandia during the Weichselian. So far, the data has revealed a highly dynamic FIS with ice-free conditions and present-day summer temperatures at Sokli during early MIS 3 (Helmens et al. 2007b, Engels et al. 2008, Helmens and Engels 2010). These conditions contrast sharply with earlier reconstructions that assumed ice cover over major part of Fennoscandia during this time period. Additionally, Sokli records strong continental climate conditions, instead of glaciation, for MIS 5d, and boreal interglacial conditions, instead of

arctic conditions, for MIS 5c (Väliranta et al. 2009, Engels et al. 2010, Helmens et al. 2012). Weaker-than-present climate gradients over central-northern Europe are reconstructed for MIS 5c (Väliranta et al. 2009).

The results from Sokli combined with other recently obtained proxy records from eastern and northern Europe prompt to a re-examination of the LI-G cycle. In this paper, the Sokli proxy data is compared in detail with long proxy records from central, temperate and northern, boreal Europe. These include the classic records of La Grande Pile in France (e.g. Woillard 1978, Guiot et al. 1989, de Beaulieu and Reille 1992) and Oerel in Germany (e.g. Behre and Lade 1986, Behre et al. 2005) and more recently obtained records from Horoski Duże in Poland (Granoszewski 2003) and Lake Yamozero in northwestern Russia (Henriksen et al. 2008) (Figure 1-1). The review focuses on lithology, biotic proxies (mostly pollen), quantitative climate inferences based on plant and insect remains, and dating. Furthermore, comparison is made with other recently obtained proxy data from central and northern Europe and with the fastly accumulating data from the marine realm including the subpolar North Atlantic Ocean. The aim of the review is a re-examination of the climate evolution of the LI-G cycle.



**Figure 1-1.** Location maps showing sites mentioned in the text in perspective to **A**: present-day distribution of latitudinal vegetation belts over Europe and northern distribution limits of selected plant taxa discussed in the text; **B**: extents of the southwestern portion of the North Eurasian Ice Sheet (including the Fennoscandian and British Ice Sheets) during the main cold stages of the last ca. 140 kyr; **C**, **D**: present-day isotherms of mean July and January temperatures, respectively, over central and northern Europe. The limit of *Larix siberica* on map A corresponds to its western distribution limit. The boundary between hemi-boreal and temperate forests in Fennoscandia corresponds with the southern distribution limit of *Picea abies*; the occurrences of *P. abies* in the Alps, Alps foreland and the Carpathians are not shown. Map B is based on Svendsen et al. (2004), Carr et al. (2006) and Mangerud et al. (2011). The ice extent during MIS 3 corresponds to the maximum extent of ice-cover over northern Fennoscandia during an interstadial dated at ca. 53 kyr ago (Helmen and Engels 2010). Also shown is the maximum limit of Pleistocene glaciation of the Alps (Ehlers and Gibbard 2004). Note that the glacier distribution in other mountain areas is not shown on the map. Temperature isotherms on maps C and D are according to Hijmans et al. (2005).

## 2 Methods

This review study focuses on long pollen records from central and northern Europe that cover a substantial part of the Late Pleistocene (last ca. 130 kyr). Pollen data from five long records, combined with lithology, dating and July temperatures inferred from plant macrofossils or insect remains, are presented in similar ways in Figures 3-1 and 3-3 to 3-6. Pollen sums are based on the total of terrestrial pollen with the exception of the Horoszki Duże record in Figure 3-4 where *Phragmites* (Poaceae) is excluded from the sum (Granoszewski 2003) and the Lake Yamozero record which has a sum excluding Cyperaceae (Figure 3-6; Henriksen et al. 2008). Note that the shrub *Betula nana* is separated from tree *Betula* only in the pollen records shown in Figures 3-4 and 3-5. The sites are situated north of the major south European mountains ranges of the Pyrenees, Alps and Carpathians, and to the east respectively west of the Scandinavian and Ural mountains in northern Europe (Figure 1-1). They cover a latitudinal range of 48–68° N and a longitudinal range of 7–50° E. Figure 1-1 further shows other sites mentioned in the text as well as the present-day distributions of vegetation types and plant taxa discussed in the report (A), inferred glacier-extents over northern Europe during MIS 6-2 (B) and present-day patterns of mean July (C) and mean January temperatures (D).

The description follows the marine oxygen isotope stratigraphy, bearing in mind the various degrees of diachroneity between terrestrial stratigraphic units and marine stages (e.g. Kukla et al. 2002a, Sánchez-Goñi et al. 1999). Marine, ice core and terrestrial stratigraphies are summarized in Figure 4-1.

## 3 Late Pleistocene proxy records from central and northern Europe

### 3.1 La Grande Pile (France)

The La Grande Pile pollen record represents one of several long proxy records spanning the LI-G cycle that have been obtained from fairly complete sedimentary sequences in France in west-central Europe. The peat bog of La Grande Pile is situated at an elevation of 360 m on the southwestern slope of the Vosges mountains in north-eastern France (lat. 47°44' N, long. 6°30' E; Figure 1-1). The glacial depression is located 2.5 km up-valley from terminal moraines formed during the penultimate glaciation (Linexert Glaciation, MIS 6) and 4 km down-valley from a moraine-line dated to the early part of the last glaciation (Lanterne Glaciation, MIS 4-2; Seret et al. 1990). Detailed pollen studies on multiple cores from the basin-infill, together with lithological data, are presented in Woillard (1975, 1978). Because of loss of Woillard's original pollen counts, new analysis on a previously not studied borehole (core GP XX) was made by de Beaulieu and Reille (1992). The pollen stratigraphy and inferred vegetation that are summarized below is according to Woillard (1978) and de Beaulieu and Reille (1992).

The up to ca. 20 m thick La Grande Pile sediment sequence consists mostly of gyttja and silty clay, overlying glacio-lacustrine sediment and till, and with Holocene peat at the top (Figure 3-1). The pollen record shows that the classical central European temperate Eemian Interglacial (locally named the Lure Interglacial; Seret et al. 1990) was followed by two more temperate intervals named St. Germain I and St. Germain II. The latter are defined together with the Lure as the Last Interglacial Complex. The temperate intervals of the Last Interglacial Complex are separated by the cold and dry Melisey 1 and Melisey 2. The Last Glacial period was characterized by an overall evolution towards a drier and colder climate. Originally, correlation of the Last Interglacial Complex at La Grande Pile with the Eemian Interglacial and Early Weichselian Brørup and Odderade Interstadials in the northwest European mainland stratigraphy, the latter interstadials being characterized by boreal conditions, was conceived as problematic (e.g. Mangerud et al. 1989, Behre 1989). It is now widely agreed that St. Germain I and II correlate to the Brørup and Odderade, respectively, and broadly with MIS 5c and MIS 5a in the marine record, and that the Eemian/Lure correlates to MIS 5e (Figures 3-1 and 4-1). The Last Glacial at La Grande Pile corresponds to MIS 4-2. <sup>14</sup>C dating places the Goulotte, Pile, Charbon and Grand Bois Interstadials in the time interval ca. 62–29 (uncalibrated) kyr BP (Woillard and Mook 1982).

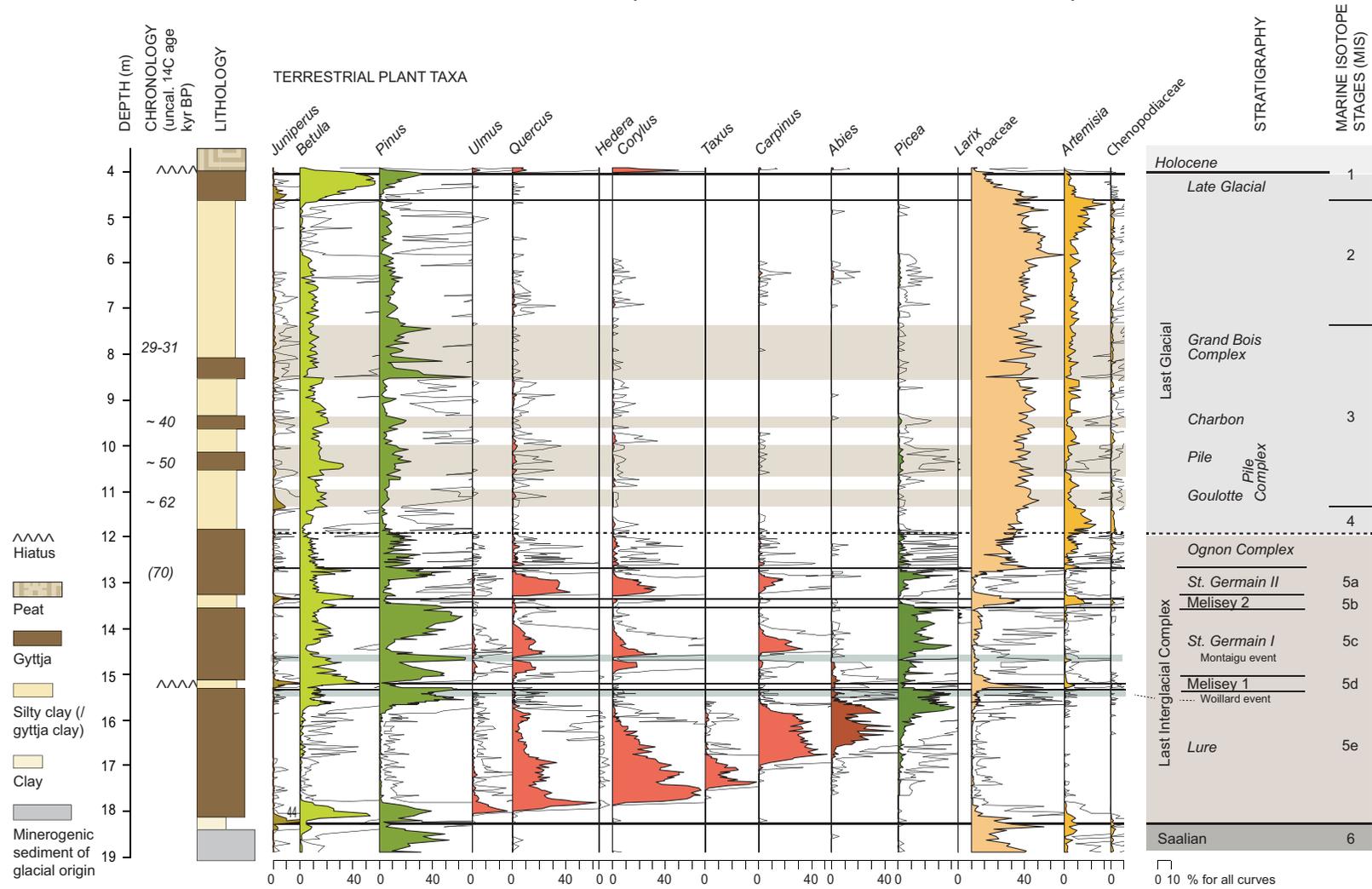
#### 3.1.1 Environmental reconstructions for MIS 5

The environment at la Grande Pile at the start of the lake record, i.e. at the end of the Penultimate Glacial (MIS 6), was characterized by open conditions with grasses and heliophytes (e.g. *Artemisia*, *Chenopodiaceae*) as important elements (Figure 3-1). A slight climate improvement, indicated by an extension of *Pinus* and shrubs, is tentatively correlated with the Zeifen oscillation of southern Germany (Jung et al. 1972).

The Lure (MIS 5e) shows a distinct interglacial vegetational succession. First, heliophytes, *Juniperus* and rapidly expanding trees (*Betula* and *Pinus*) were abundant. This was followed by a major expansion of deciduous tree taxa such as *Ulmus* and *Quercus*. The latter were then replaced by forest vegetation types with *Corylus*. Following a phase with abundant *Taxus*, which marks the end of a climate that was more temperate than the present one (indicated by the presence of *Hedera* and *Ilex*), a general climatic deterioration is inferred with the establishment of *Carpinus* forest, followed by *Abies* forest, *Picea* expansion and finally the establishment of boreal *Pinus* and *Betula* forests; Poaceae increased in significance as *Picea* expanded. A two-fold partition of the *Pinus* curve (showing increases in *Picea*, and even *Abies*) suggests the existence of a moderate climatic oscillation (Woillard event) near the end of the Lure with increased precipitation. With the exception of the earliest *Juniperus* and herb-dominated part of the Lure that is recorded in gyttja-clay, gyttja accumulated in the La Grande Pile depression during the Lure interglacial interval.

St. Germain I (MIS 5c) and St. Germain II (MIS 5a) are both characterized by a vegetation succession rather similar to the Lure one. Following an initial phase with abundant heliophytes (including *Ephedra* during St. Germain II), *Juniperus* and *Betula*, as well as *Pinus* during St. Germain I, deciduous

La Grande Pile (E France; 47°73' N, 6°50' E, 360 m.a.s.l.)



**Figure 3-1.** Simplified chronology, lithology and pollen record of the La Grande Pile sedimentary sequence in eastern France (for location see Figure 1-1). Chronostratigraphy and marine stages are given to the right. Based on Woillard (1978), Woillard and Mook (1982), Seret et al. (1990), and de Beaulieu and Reille (1992). The hiatus during Melisey 1 is according to Turner (2002a).

vegetation developed with first *Quercus* and *Corylus* and then *Carpinus*. The end of the temperate phase is marked by the disappearance of the deciduous forest and the occurrence of conifer forest with *Picea*, *Pinus* and *Larix* (and Poaceae). In contrast to the Lure, *Picea* was present throughout major part of St. Germain I and II suggesting cooler conditions (Guiter et al. 2003). A short climatic regression (Montaigu event) is identified during the early part of St. Germain I; the event, which followed a phase of intense forest dynamics, was characterized by the total disappearance of deciduous taxa and rise of *Pinus* (and *Artemisia*). Similar as during the Lure, gyttja was deposited during most of St. Germain I and II, with the exception of the initial phases with heliophytes, *Juniperus* and *Betula/Pinus* that are recorded in silty clay and gyttja-clay.

The intervening Melisey 1 and 2 (MIS 5d and MIS 5b, respectively) are recorded in thin silty-clay layers, in which the enrichment of silt suggests the accretion of loess (Seret and Woillard 1976). The pollen record shows the development of open vegetation types with Poaceae and steppe species (e.g. *Artemisia*, Chenopodiaceae, *Helianthemum*, *Ephedra*), and subarctic-subalpine plant species such as *Thalictrum*, *Selaginella* and *Bortyichium*, clearly indicative of a cold and dry climate. *Artemisia* attains especially high values during Melisey 2 suggesting that this stadial was colder than Melisey 1. The shrub *Juniperus* shows an increasing trend during the latter part of both cold intervals. Turner (2002a) notes the possible existence of a hiatus in the sedimentary record explaining the particularly short record for Melisey 1 at La Grande Pile. Stratigraphic hiatuses associated with the horizon of Melisey 1 are also registered at e.g. Les Echets in south-eastern France (de Beaulieu and Reille 1984; Figure 1-1), which according to Turner (2002a) suggests a widespread episode of increased aridity within the stadial resulting in lake-level fall and basin disturbance. Melisey 1 shows the first recording of pollen of the ericaceous dwarf shrub *Bruckenthalia*. This shrub presently occurs in the high mountains of continental southeastern Europe, where it endures extreme temperature changes in both winter and summer.

Using an extensive modern calibration set that covered Europe, North Africa and Siberia, Guiot et al. (1989) made quantitative estimates of mean annual temperatures and total annual precipitation values based on the pollen data from La Grande Pile as well as from Les Echets. In agreement with the botanical evidence, the reconstructions show a clear succession of two relatively warm and humid periods after the Lure corresponding to St. Germain I and II, with the Lure appearing as the warmest and most humid episode. The closing phases of the Lure and St. Germain I and II are reconstructed as cold and humid, followed by the even colder and drier Melisey 1 and 2.

An analysis of insect remains (Coleoptera) in the La Grande Pile sediments was made by Ponel (1995). St. Germain I and II yielded Coleopteran assemblages rather similar to those recorded in the second part of the Lure, i.e. including both deciduous- and conifer-dependent taxa. The Lure revealed an additional early phase in which the beetle fauna is characterized by species dependent on deciduous trees. The early Lure deciduous forest also seems to have had a poorly developed herbaceous stratus, probably due to the density of trees, in contrast to a distinct herbaceous stratum and an opening-up of the environment that is reconstructed during St. Germain I. By calibrating the climate signal of the new pollen data of de Beaulieu and Reille (1992), Guiot et al. (1992) inferred a maritime climate for the early part of the Lure and a more continental one during the later part; St. Germain I and II are inferred as mainly continental. Crosscheck between the pollen and palaeo-entomological data suggests that annual mean temperatures were at most 2–4°C higher at La Grande Pile during the Lure climate optimum than at present (Guiot et al. 1993).

Ponel (1995) showed that the Melisey 1 and 2 sediments are decidedly poor in tree-dependent Coleoptera but do not contain any really cold-adapted taxa. It is suggested that the Melisey 1 and 2 climate deteriorations were probably not severe enough to permit the incoming of really cold-adapted beetles. Cold-adapted tundra beetle species are commonly found in the sediments dated to the end of the Penultimate Glacial (MIS 6) and to the Last Glacial (MIS 4-2; see under 3.1.3). Similarly, Pons et al. (1992) noted that the rapid establishment of trees during both St. Germain I and II suggests that Melisey 1 and 2 did not last very long and were not very arid nor cold, allowing the persistence of nearby plant refugia.

### 3.1.2 Comparison with other records

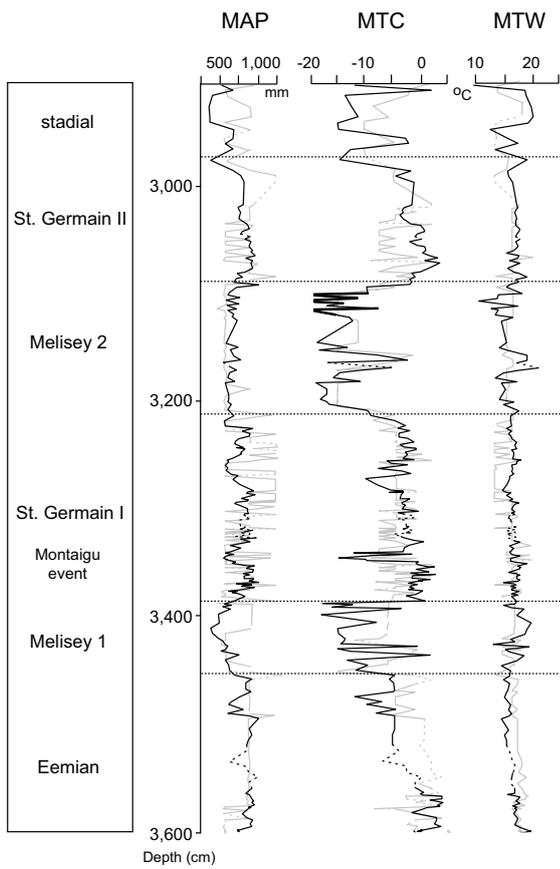
A quantitative reconstruction of January temperatures and annual precipitation for the Lure/Eemian (MIS 5e) based on five high-resolution pollen records from France, including La Grande Pile, and using the 'best modern analogue' statistical method, was made by Cheddadi et al. (1998). The transition from the Penultimate Glacial into the Lure/Eemian is reconstructed as a significant increase in winter temperatures of up to some 20°C and annual precipitation values by some 400 mm. Highest winter temperatures (to about 4°C; compared to present-day values at ca. 0°C in Figure 1-1) and most humid conditions are centered in the early part of the interglacial interval during the mixed oak forest phase with *Quercus* and *Corylus* as dominant trees. With the spread of the *Carpinus* forest, particularly winter temperatures start a distinctly declining trend in the reconstructions. A reconstruction of seasonality changes averaged over 17 pollen records from Europe, using a multi-method approach, depicts a distinct decrease in seasonality during the earliest part of the Eemian and increasing seasonality following the early Eemian climate optimum (Brewer et al. 2008).

Klotz et al. (2004) combined two different climate reconstruction methods, a modern analogue technique and a mutual climate range approach, on high-resolution pollen records covering the late Lure/Eemian – St. Germain II (MIS 5e–5a) from the northern Alpine foreland, including the France Les Echets pollen sequence and the pollen sequences from Jammertal (Müller 2000), Füramoos (Müller et al. 2003) and Samerberg (Grüger 1979) in southern Germany (at elevations ranging from 200 to 662 m.a.s.l.; Figure 1-1). All four pollen records show three forested intervals. However, the proportions of thermophilous taxa during St. Germain I and II are greatly diminished at the more continental easterly sites. In all records the proportion of certain non-arboreal pollen (NAP), including *Artemisia*, is higher during the stadial interval correlated to Melisey 2 than during Melisey 1, suggesting open steppe-like ecosystems (compared to open steppe-tundra-like ecosystems during Melisey 1) and more continental climate conditions during Melisey 2.

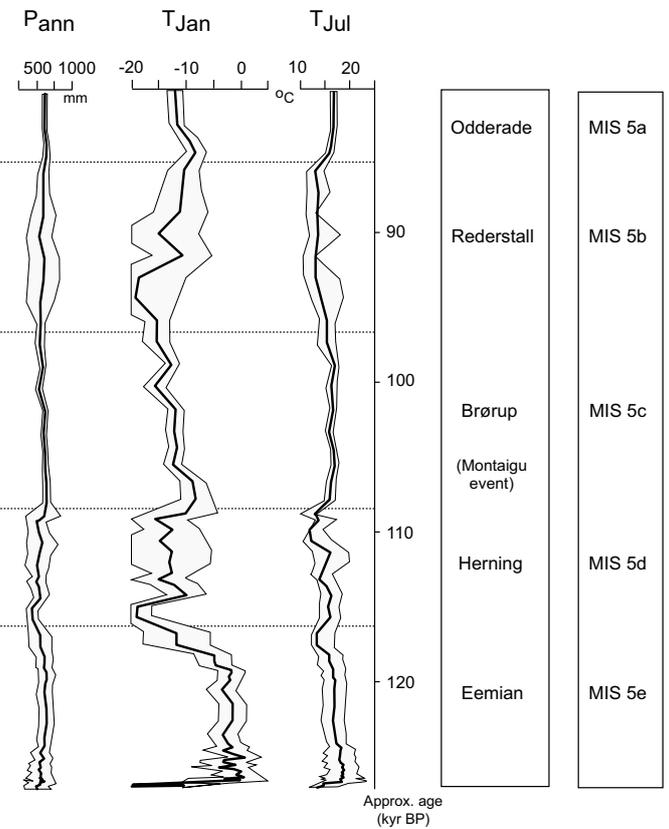
The quantitative reconstructions by Klotz et al. (2004) clearly depict the considerable climate ameliorations at the onset of St. Germain I and II which were mostly in terms of increasing winter temperatures and precipitation values (Figure 3-2A). The reconstructions also suggest that the optima in both mean temperature of the warmest month (MTW) and mean temperature of the coldest month (MTC) were somewhat higher during St. Germain II compared to St. Germain I. MTW and MTC during St. Germain I and II averaged around ca. 16–17°C and ca. –3 to –10°C, respectively (compared to present-day values up to ca. 18°C and down to –2°C in Figure 1-1); the lowest winter temperatures are recorded at the continental eastern sites. A distinct climate evolution is reconstructed across the cold Melisey 1 and 2 that included, at first, an intensification of continentality as MTC fell to an average minimum of about –15°C during Melisey 1 and –17°C during Melisey 2 and MTW remained stable or even increased. Following an intra-stadial amelioration, which is mostly documented by a rise in winter temperatures, another reduction in winter temperatures with mostly stable or decreasing summer temperatures is observed. Similarly, an increase in continentality, accompanied with reduced precipitation values, is reconstructed for the Montaignu cooling event.

Guiot et al. (1989, 1992) identified the cold but humid final parts of the Lure and St. Germain I and II interglacial intervals (with *Picea*, *Pinus* and *Betula* forests) as possible glacial accretion periods correlating to the increases in global ice volume recorded in the earlier parts of MIS 5d, 5b and 4, respectively. Lagging of the terrestrial pollen stratigraphy behind the marine isotope stratigraphy was also noted by e.g. Kukla et al. (2002a) and is supported by direct land-sea comparisons in which pollen and isotope analyses are combined on marine cores (Turon 1984, Sánchez-Goñi et al. 1999, Shackleton et al. 2003). The latter indicates that the start of MIS 5d, which is interpreted to correspond to a substantial accumulation of ice in high-northern latitudes, occurred several thousands of years before the demise of temperate and conifer forests in Iberia and France (end of the Lure). It is the onset of significant ice-rafting in the North Atlantic at ca. 110 kyr ago during North Atlantic Cold event (NAC) 24 (Oppo et al. 2006) that is thought to have led to the elimination of tree populations in SW Iberia through the disruption of the thermohaline circulation (Tzedakis 2003). In the central European Alpine foreland, U/Th dates of peat from different sites indicate that also here interglacial conditions persisted until at least 115 kyr ago. These dates are from deposits that are attributed to the period of cooling towards the end of the Eemian Interglacial (Preusser 2004). The stadial intervals Melisey 1 and 2 are correlated with NAC 24 and 21, respectively, and the moist Woillard event and the Montaignu cooling event are correlated with NAC 25 and 23, respectively (e.g. Kukla 2000, Kukla et al. 2002b, Tzedakis 2003, Klotz et al. 2004, Rousseau et al. 2006; Figure 4-1).

**A LES ECHETS (France; 200 m.a.s.l.)**



**B GRÖBERN (Germany; 100 m.a.s.l.)**



**Figure 3-2. A:** reconstructed mean temperatures of the coldest (MTC) and warmest month (MTW) and mean annual precipitation (MAP) using a modern analogue technique (black lines), and the average of the upper and lower limit of the probability intervals using the mutual climate range approach (grey lines), for the late Eemian (late MIS 5e) to St. Germain II (MIS 5a) at Les Echets in southeastern France (Figure 1-1; from Klotz et al. 2004). **B:** mean probabilistic climate reconstructions (black line), with standard deviation (dashed line), for the Eemian and Early Weichselian (MIS 5e - early MIS 5a) at Gröbern in central Germany (Figure 1-1; from Kühl et al. 2007). Note that the depth scale of the Gröbern record has been adjusted according to age based on correlation with the NGRIP ice core (NGRIP Members 2004) and varve counts at Bispingen (Müller 1974). The Les Echets climate record is based on pollen data and the Gröbern climate reconstructions are inferred from pollen combined with macrofossils.

### 3.1.3 Environmental reconstructions for MIS 4-2 and the Late Glacial-Holocene (MIS 1)

Following a brief interval with steppic conditions, Woillard (1978) distinguishes two climatic improvements. The Ognon I and II oscillations are represented by *Betula-Pinus* forest, including *Larix* and *Picea*, and are recorded in gyttja and glyttja-clay, respectively (Figure 3-1). De Beaulieu and Reille (1992) only recognize Ognon I and consider the occurrences of temperate taxa (*Corylus*, *Quercus*) above Ognon I as redeposited. A significant climate amelioration with *Pinus-Picea* forest vegetation, following St. Germain II, is also registered to the east in the Füramoos record of southern Germany (Müller et al. 2003). The warming at La Grande Pile and Füramoos is detected before the main opening of vegetation that characterizes the upper part of both records (see below). The Ognon I at La Grande Pile and the warming event in the Füramoos record have been correlated to the Dürnten Interstadial defined by Welten (1982) in Switzerland (Guiter et al. 2003, Müller et al. 2003). Müller et al. (2003) consider the persistence of forest vegetation in central Europe during the Dürnten Interstadial as incompatible with the cold climate of MIS 4 and have suggested inclusion of the interstadial into the uppermost part of MIS 5; the interstadial is correlated with North Atlantic Warm event (NAW) 20 (Figure 4-1). De Beaulieu and Reille (1992) also question the inclusion of Ognon I in the Last Glacial period.

The La Grande Pile proxy records show a significant change at the transition to the Last Glacial (MIS 4-2). Pollen of heliophytes, including *Artemisia*, become continuously registered, Poaceae pollen values remain high, and temperate species such as *Corylus* are represented by low values only (Figure 3-1). Similarly, a clearly dramatic change at the transition to the Last Glacial is indicated in the Coleoptera record. Tree-dependent beetles disappear almost totally, only a few isolated individuals of willow-dependent taxa persist, and standing-water species increase in number; also characteristic is the appearance of a number of very cold-adapted Coleoptera species (Ponel 1995). Mostly silty clay accumulated in the La Grande Pile depression.

Following the Ognon Complex, high amounts of steppe species, indicative of arid, particularly rough conditions, are recorded that seem to correspond with MIS 4. The climate reconstruction by Guiot et al. (1989) shows very low mean annual temperatures, i.e. significantly lower than during Melisey 1 or 2.

Between ca. 62 and 29 (uncalibrated) kyr BP (MIS 3), relatively high arboreal pollen (AP) values (especially *Betula* in the early part and *Pinus* latter on) are registered. Pollen values for steppe species are generally relatively low and a rich aquatic flora is recorded. The Pile Interstadial Complex (Goulette and Pile Interstadials) shows open *Betula-Pinus* forest with *Picea*, *Larix* and possibly *Quercus*, and *Juniperus* at the start. A peak in *Betula* and increased pollen percentage values of *Pinus* characterize the Charbon climatic improvement. The Grand Bois Interstadial is represented by open *Pinus-Betula* forest with some *Juniperus* and the presence of *Hedera*. These climatic improvements alternate with steppic conditions. Detailed radiocarbon dating by Woillard and Mook (1982) places the Charbon and Pile proper at around 40 and 50 (uncalibrated) kyr BP, respectively, whereas the base of the Pile Interstadial Complex is placed at 62 kyr BP by assuming a constant sedimentation rate; the Grand Bois Interstadial Complex is dated between ca. 29 and 31 kyr BP (Figure 3-1).

The interstadials during MIS 3 are interpreted as increased expansion or blossoming of woodstands or shrubs in a still open vegetal environment. Insect studies in the U.K., however, do show summer temperatures as high as today during the most temperate episodes of MIS 3 (e.g. Coope et al. 1997, Vandenberghe et al. 1998). According to Guiter et al. (2003), these favorable thermal conditions did not allow the return of forest in northwestern Europe, probably because of the short duration of the warming intervals. The presence of insects in the La Grande Pile sediments suggest a short warming episode with mean July temperatures approaching 20°C during the Charbon Interstadial (Guiot et al. 1993, Ponel 1995). This warming event possibly correlates with the Upton Warren event in the U.K. (Coope et al. 1997).

The highest pollen values of *Artemisia* and Poaceae (cold and dry maximum) are reached during late MIS 2 (Figure 3-1). MIS 2 at La Grande Pile also shows a rise in arctic-alpine Coleoptera species, the complete disappearance of running-water beetles, and an impoverishment of ubiquitous species. In agreement with the pollen and palaeontological data, the linking of pollen with sedimentology (e.g. organic carbon content, periglacial loess content, smectite content) led Seret et al. (1992) to conclude that the climate during MIS 2, as well as MIS 4, at La Grande Pile was coldest and driest, whereas it was less cold and more humid during MIS 3.

The reappearance of pioneer shrubs (particularly *Juniperus*) was the preliminary step to the first arboreal dynamics with *Betula* and *Pinus* that characterized the Late Glacial. The dynamics are interrupted by a hiatus in the sedimentary sequence. The uppermost pollen assemblage shows the expansion of *Corylus* during the Boreal and Preboreal (Holocene).

## 3.2 Oerel (Germany)

The Oerel site in northern Germany (lat. 53°59' N, long. 9°04' E, elevation 10 m.a.s.l.; Figure 1-1) represents a long terrestrial record of five organic beds separated by layers of sands that extends in age from the Eemian into the Middle Weichselian (MIS 5-3). The sediment sequence is up to ca. 20 m thick. Three forested intervals are recorded in the lower part of the sequence and are correlated with the Eemian Interglacial (MIS 5e) and Early Weichselian Brørup and Odderade Interstadials (MIS 5c and 5a, respectively). Two intervals with shrub tundra in the upper part of the record are dated by means of <sup>14</sup>C dating to the time interval ca. 58–48 (uncalibrated) kyr BP (early MIS 3) and named the Oerel and Glinde Interstadials. The Oerel depression occurs directly inside the Lamstedter Staffell thrust moraine ridge of Saalian age (MIS 6); the site is outside the Last Glacial Maximum (LGM; MIS 2) ice extent, moraines of which are found ca. 75 km to the northeast. Research in the paludified depression started in 1960 and included detailed pollen analysis (Behre and Lade 1986, Behre 1989), radiocarbon dating (Behre and van der Plicht 1992), and analysis of plant macrofossils and insect remains (Behre et al. 2005) on the standard profile OE 61 from the deepest part of the basin. In the latter paper, Coleoptera are used to infer temperature values using the Mutual Climatic Range (MCR) method. In the present review, minimum July temperatures are additionally inferred based on the rich macrofossil record at Oerel following e.g. Iversen (1954) and Kolstrup (1980). The botanical and Coleoptera data summarized below follows Behre (1989) and Behre et al. (2005).

### 3.2.1 Environmental reconstructions for MIS 5

Sedimentation initiated in a kettle hole. These early Eemian (MIS 5e) sediments, however, are lacking in the standard profile which starts in the temperate phase with the *Corylus* peak (Figure 3-3). Next, pollen of *Tilia* and *Taxus* increase and this is followed by the *Carpinus* phase. The subsequent *Abies* peak (normally in NW Germany at 4–6%) is missing indicating a hiatus at the transition from lacustrine to telmatic deposition. The later part of the Eemian shows the spread of *Pinus*. *Picea* is well-represented throughout the *Carpinus* phase, showing diminishing percentages as *Pinus* spreads. A peak in Ericales (*Caluna* and *Empetrum*) to 60% in the upper part of the late Eemian peat deposit is ascribed to an over-representation of Ericales pollen from the local raised bog (Behre et al. 2005). The peat is overlain by sterile sands of the early Herning Stadial (MIS 5d). At the Gröbern site in central Germany (Kühl et al. 2007; Figure 1-1), peak values for *Pinus* at the end of the Eemian are followed by a peak in *Betula* accompanied by increased values for Poaceae.

Major part of the Eemian deposit at Oerel consists of a fine-detritus gyttja (Figure 3-3). A reduction of the lake area with an advancement of reed and fen vegetation is recorded in the *Carpinus* phase. Subsequently, a mire developed (with deposition of fen peat) turning into a raised bog (*Sphagnum* peat). The aquatic plant species *Najas minor* is an indicator of warm summer temperatures during the *Carpinus* phase. *N-minor* requires a minimum mean July temperature of 18°C (Figure 3-3). A rather species-rich Coleoptera fauna is recorded in the lower part of the late Eemian fen peat deposit and indicates MTW and MTC of ca. 16–19°C and –1 to –5°C, respectively (compared to present-day values at ca. 16°C and 1°C in Figure 1-1).

Tree *Betula* rapidly spread during the early Brørup (MIS 5c) and Odderade (MIS 5a). Lacustrine sedimentation initially prevailed. As the reforestation proceeded with the local spread of *Pinus* (increasing transpiration and reducing run-off), the lake dried out and peat formation begun, eventually changing the local environment at the site into a raised bog. The vegetation development during both forested intervals was rather similar. However, the *Betula* phase lasted much longer in the Brørup, and both *Picea* and *Larix* were more extensively represented. The expansion of the latter conifer trees took place during the Brørup in the late *Betula* phase, whereas in the Odderade this did not take place until the late *Pinus* period. Thermophilous deciduous tree pollen, especially *Quercus* and *Corylus*, reach low percentage values only. Within the Brørup *Betula* period, a setback in vegetation development is recorded that is correlated with the Montaigne phase at La Grand Pile in France. Like in the late Eemian, sterile sands overly the late Brørup and Odderade peat beds.



The wetland plant *Typha* indicates minimum mean July temperatures of 15°C during the early Brørup *Betula* phase. Mosses that are encountered, such as *Helodium blandowii* and *Meesia longiseta* (and later *Drepanocladus tundrae* as well as the aquatic insect *Corynocera ambigua*), currently have a specific northern distribution. According to the authors, the early Brørup species composition can best be explained by strong continental conditions with warm summers and cold winters. MCR reconstructions on Coleoptera for main part of the Brørup indicate MTW and MTC in the range of ca. 15–19°C and –8 to –14°C, respectively; macrofossils of the aquatic plant *Najas flexilis* indicate minimum mean July temperatures of 15°C. Furthermore, a number of Coleoptera species are found that today occur in temperate regions. The major part of the Odderade is dominated by cold stenotherm insects. MCR reconstructions indicate 8–12°C (MTW) and –11 to –22°C (MTC). In the lowermost part of the Odderade *Pinus* phase, however, values of 19°C (TWM) and 2–4°C (TCM) are reconstructed, and minimum mean July temperatures of 15°C during the preceding *Betula* phase are indicated by *Typha*. Several macrofossils of *Betula humilis* (an eastern-continental species) and, also during the Brørup, of *Picea omorika* (Serbian spruce) are recorded.

The Early Weichselian Herning (MIS 5d) and Rederstall Stadials (MIS 5b) are represented by sterile fluvial sands and fine gravel with on top organic mud deposited under shallow lacustrine conditions. The organic muds in the upper part of both stadials are interpreted to represent slight improvements of climate compared with the preceding phases with minerogenic sedimentation. Pollen assemblages indicate open vegetation with Poaceae (percentages up to 40%), *Artemisia* and *Juniperus*. The dwarf shrub *Betula nana* is additionally represented in the macrofossil record. Pollen of temperate trees and *Picea*, which particularly occur in the lower parts of the limnic sequences, are interpreted as redeposited. The Rederstall sands reach the greatest thickness of the two Early Weichselian sand layers; additionally, pollen of *Artemisia* as well as of other typical heliophytes like *Helianthemum*, *Armeria* and *Gypsophila* are more frequent in the Rederstall organic sediment compared to the Herning organic bed. It is concluded that during the Herning, climatic deterioration probably was not as severe as during the Rederstall. The Coleoptera assemblage in the Rederstall mud is of a boreal character; the presence of the aquatic plants *Cerastium demersum* and *Stratiotes aloides* indicates relatively high summer temperatures at the closing phase of the stadial (minimum 12–15°C). *C. demersum* and *S. aloides* tolerate very low January temperatures. For both stadials, the presence of macrofossils of *Typha* indicates minimum mean July temperatures of 15°C.

### 3.2.2 Comparison with other records

Climate reconstructions based on a multivariate probability approach employed to pollen and macrofossils of Eemian – early Odderade (MIS 5e–5a) age at the Gröbern site of central Germany (Kühl et al. 2007; Figures 1-1 and 3-2B) show that the highest degree of continentality during the Herning Stadial (MIS 5d) occurred during its earliest part. The transition from the Eemian to the Herning at Gröbingen included a decrease in mean January temperatures of some 20°C (to ca. –19°C, compared to present-day values at ca. –1°C in Figure 1-1). July temperatures during the Herning, however, remained high enough (i.e. around 15°C) to support boreal trees. It is concluded that a decrease in precipitation (although difficult to reconstruct) in combination with lowering in January temperatures most probably were major factors driving vegetation changes during the Eemian/Early Weichselian transition in central Germany (Kühl et al. 2007). The reconstructions show that winter temperatures already dropped in the late part of the Eemian (MIS 5e). A high degree of continentality is also reconstructed at the very beginning of the Rederstall Stadial (MIS 5b). The reconstructions by Kühl et al. (2007) further show that during the Eemian, highest July temperatures to ca. 19°C (i.e. ca. 2°C higher than at present) and lowest precipitation values were reached in the earliest part of the Eemian. Mean July and January temperatures of ca. 17°C and around ca. –12°C, respectively (the latter compared to around –3°C during most of the Eemian), are inferred for the Brørup (MIS 5c) and Odderade (MIS 5a) at Gröbingen.

In the northwestern-most portion of the present-day temperate zone in Europe (Figure 1-1), sediment beds at Stenberget in southern Sweden and Fjøsanger in south-western Norway have been correlated using pollen data with the Eemian Interglacial (MIS 5e) and the Early Weichselian (MIS 5d-a).

Thick layers of Pleistocene deposits are found in Skåne (southern Sweden) in tectonic depressions situated in-between horst ridges. The traditional stratigraphy here, based on a century of intense research, was revised by Berglund and Lagerlund (1981). At the chrono-stratigraphic key site

Stenberget (160 m.a.s.l.; Figure 1-1), organic beds of the Stenberg Formation occur interbedded between the Late Saalian Ramslid Till and an upper till bed correlated with the Late Weichselian Dalby Till. The Eemian (locally named the Romele Interglacial) is thought to be represented by its earlier and later parts, with a ca. 8–9 kyr long hiatus representing the temperate phase. Because of the hiatus, however, the correlation with the Eemian at Stenberget is not totally proven (Robertsson 2000). Sedimentation initiated in a lake that filled a kettle hole similar as at Oerel. These early Eemian heavy silty clays and clay gyttja contain a pollen assemblage in which *Betula* as well as *Pinus* are well-represented, as are Poaceae and heliophytes such as *Artemisia*, Chenopodiaceae and *Helianthemum*. Next follows a 1 m thick peat layer with wood pieces of e.g. *Betula*, *Pinus* and *Alnus* in its lower part. High percentages of *Pinus* pollen in this lower wooded fen deposit are replaced by *Picea* and *Alnus* pollen in the upper sedge (*Carex*) fen deposit. The peat is further characterized by high fern spore percentages (Polypodiaceae). Thermophilous taxa, mostly *Corylus*, are represented by low percentages only. A thin layer of heavily disintegrated peat separates the wooded fen bed from the sedge fen deposit suggesting the occurrence of a hiatus in the sedimentary sequence. The hiatus is interpreted to represent the interglacial climatic optimum with reduced regional humidity. The upper part of the interglacial sequence consists of sandy gyttja that is interlayered/fingered with solifluction sediment containing abundant frost-cracked flintstones. Vegetation during this latest part of the Eemian is interpreted as wooded taiga, with first *Picea* and *Pinus* followed by *Betula* and *Larix*, and *Artemisa* throughout the zone, and with a boreal to sub-arctic, humid climate. The humid climate might have made the clayey soils waterlogged and unstable, producing solifluction sediment as suggested in Andersen (1975). The sandy gyttja contains low pollen percentages of *Corylus* as well as of *Carpinus*. It gradually grades into a massive silt and sand layer of possible niveo-fluvial origin, deposited under arctic conditions with, according to the authors, deep permafrost. Throughout the major part of the early and late Eemian sequence, insect fauna assemblages are quite homogenous, characterized by species which today have a sub-arctic to boreal distribution. In the upper part of the late Eemian sandy gyttja bed, three Coleoptera fauna assemblages are distinguished. They indicate a small pond surrounded by damp and mossy ground, with a slightly colder climate than at Stenberget today, followed by distinct climate deterioration to sub-arctic and finally arctic conditions.

Organic silt with pollen floras indicating boreal woodland of *Betula*, *Pinus*, *Picea* and ferns, with some heathlands and *Sphagnum* mires, are found overlying the niveo-fluvial bed at Stenberget (Berglund and Lagerlund 1981). These Slätteröd Interstadial sediments are <sup>14</sup>C dated on bulk material at 58,050 ± 600 (uncalibrated) yr BP and on alkali extract as being older than 51,500 yr BP. The Slätteröd Interstadial is correlated with the Early Weichselian Odderade or Brørup Interstadials (MIS 5a or 5c), and the Jämtland Interstadial of central Sweden (Lundqvist 1967). The contact with an uppermost solifluction layer (1–5 m thick) is sharp and partly erosive; this layer contains abundant frost-weathered stones and shows periglacial features such as fossil ice wedges. The paleo-environment is interpreted as polar desert. The solifluction layer is cut off by till.

Marine sediments found at Fjøsanger in south-western Norway (Figure 1-1) have been interpreted to represent a continuous Eemian – Early Weichselian sequence (Mangerud et al. 1981). The location of the marine sediments on a steep mountain slope at some 30 m above the present-day sea level requires younger neotectonic uplift. The sediments have been studied in great detail using foraminifera, molluscs, pollen and sediment characteristics (e.g. mineralogy). The pollen sequence starts with an open vegetation of *Betula*, *Juniperus* and *Artemisia* that gradually changes into a *Betula* forest with ferns. *Pinus* forest with *Quercus*, as well as *Juniperus* and ferns, is subsequently replaced by a dense broad-leaved forest of *Quercus*, *Corylus*, *Alnus* and *Ilex*. The latter is interpreted to represent the Eemian Interglacial (MIS 5e) optimum with a climate similar to or warmer than during the Holocene climate optimum. Replacement of the temperate forest by coniferous *Picea-Pinus* forest marks the end of the interglacial succession. Pollen of *Carpinus* which are recorded at Stenberget, and particularly at Oerel, during the late Eemian are absent from the pollen record at Fjøsanger, suggesting a similar northern distribution limit for *Carpinus* than the present one (Figure 1-1). The upper part of the pollen record obtained from the marine sediments at Fjøsanger shows relatively high representations of Poaceae and Ericales. For part of the sequence, marine fauna indicates a more favourable climate while re-establishment or expansion of *Betula* is suggested in the pollen diagram (Fana Interstadial). The Fana Interstadial deposit occurs interlayered in-between glaciomarine sediment, whereas glacial till is found both at the bottom (MIS 6) and at top of the Fjøsanger sequence. The upper Bøne Till is correlated with the Rederstall Stadial (MIS 5b) and the Fana Interstadial with the Brørup Interstadial (MIS 5c; Mangerud et al. 2011).

### 3.2.3 Environmental reconstructions for MIS 4-3

Sterile sands of the Schalkholz (MIS 4) and Ebersdorf Stadials (early MIS 3) are directly covered by peat that accumulated at Oerel during the Oerel and Glinde Interstadials. Multiple radiocarbon datings on *Sphagnum*, as well as on brown moss peat (Oerel) and sedge peat (Glinde), give ages of 58–54 (uncalibrated) kyr BP for the Oerel and about 51–48 kyr BP for the Glinde Interstadial (early MIS 3; Figure 3-3). A mesotrophic mire developed into a raised bog during the Oerel. According to the authors, this shows that net precipitation must have been such that the formation of ombrotrophic vegetation was facilitated. Sedge peat overlain by fine detritus gyttja and then silt with fine sand and thin humus layers comprise the ca. 0.3 m thin Glinde deposit.

Both interstadial pollen assemblages record shrub tundra vegetation with *Betula nana*, *Salix* and, during the Oerel, *Juniperus*, as well as a rich herb flora. High NAP values are in part due to over-representation of pollen (Ericales and Cyperaceae) from local wetland vegetation (Figure 3-3). Macro-remains of trees were not encountered in the sediments and the vegetation is interpreted as tree-less. Heliophytes including *Artemisia*, *Helianthemum*, *Armeria* and Campanulaceae are more frequent in the Glinde indicating a more open vegetation than during the Oerel.

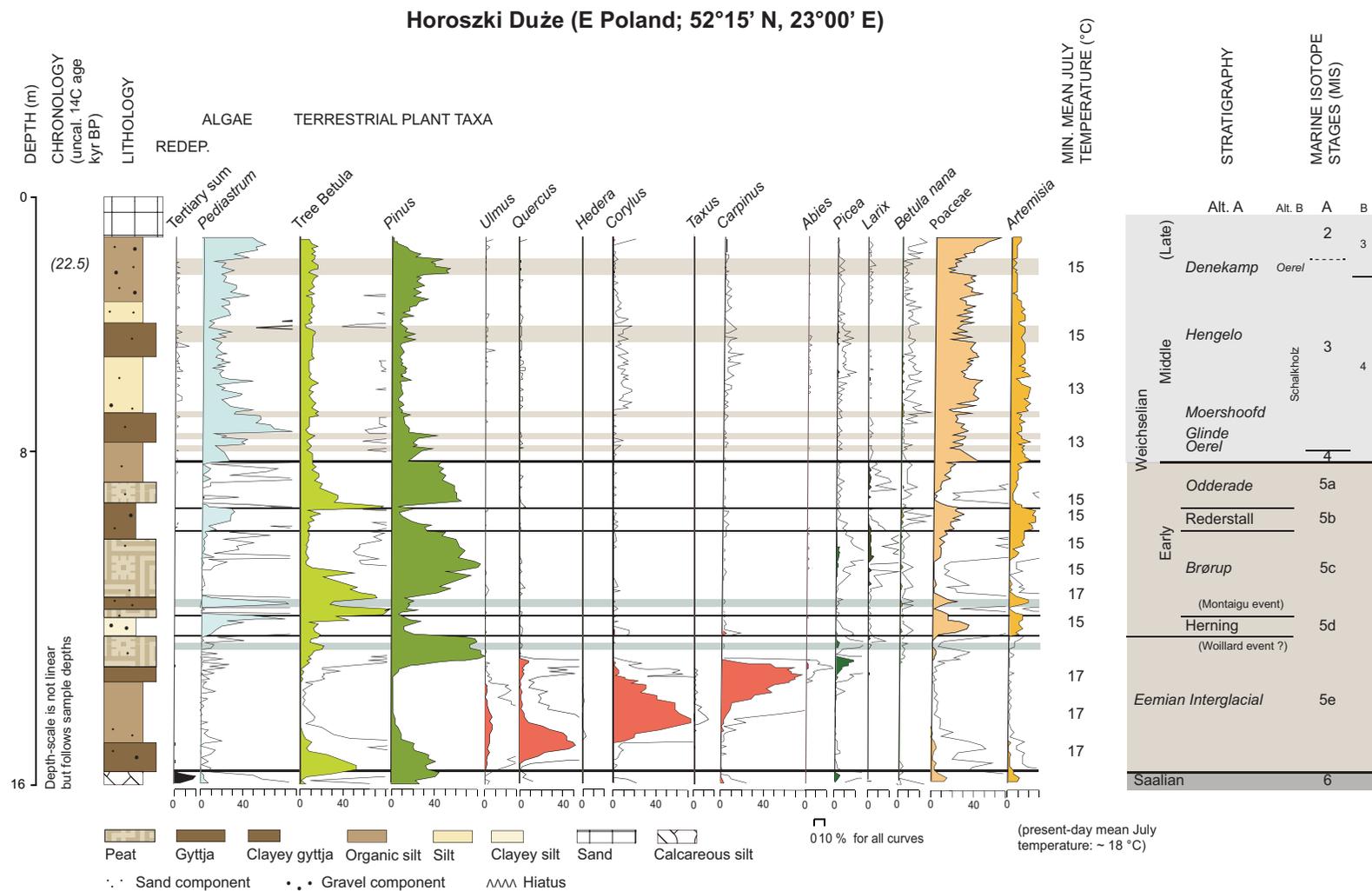
Macrofossils of the wetland plant *R. sceleratus* indicate minimum mean July temperatures of 15°C during both interstadials (Figure 3-3). MCR-inferred temperature values are at 10–11°C (TWM) and –10 to –18°C (Oerel and Glinde, respectively; TCM). During the Oerel Interstadial, the occurrence of the insect *Agonum consimile* is notable. This exclusively northern palaeartic species is currently found in the birch region of Scandinavia. Furthermore, *Leistus rufescens* and *Patrobis atrorufus* need shaded habitats. Obligate tree-dependent insect species, however, were not found and it is argued that shading shrubs might have been sufficient in this case.

## 3.3 Horoszki Duże (Poland)

The plain of the northern European mainland is rich in paleolakes with sediments deposited during various stages of the Pleistocene. A long lake record that comprises a complete sequence covering the Eemian and Early Weichselian (Early Vistulian) (MIS 5), and extends into the Pleniglacial (Plenivistulian; MIS 4-2), has been recovered from the Horoszki Duże site in eastern Poland (lat. 52°15' N, long. 23°00' E; Figure 1-1). The palaeolake lies in a glacial depression carved out in the ravined section of the Bug valley of neotectonic origin. A first paleobotanical study was made by Bitner (1954) and a detailed study including high-resolution palynological, macrofossil and lithological analyses was carried out by Granoszewski (2003). The pollen stratigraphy and inferred vegetation that are summarized below is according to Granoszewski (2003).

The 15 m thick biogenic lake sediment sequence (Figure 3-4) is underlain by sandy and silty sediment with gravel and is capped by a 1 m thick sand layer. According to Granoszewski (2003), the regional stratigraphy unequivocally places the organic rich sediment with temperate forest pollen floras near the base of the Horoszki Duże record in the stratigraphic position of the Eemian Interglacial (MIS 5e). Following the forested Brørup (MIS 5c) and Odderade Interstadials (MIS 5a), the chronology of the Pleniglacial part of the record (MIS 4-2) is less clear. Two alternative chrono-stratigraphic interpretations are presented. In the “A” variant it is assumed that increases in AP values in the upper part of the pollen sequence record a sequence of interstadials corresponding to the Oerel, Glinde, Moershoofd, Hengelo and Denekamp Interstadials (MIS 3). In alternative “B”, major part of the silty sediments overlying the Odderade would correspond with the Schalkholz Stadial (MIS 4) and only the highest increase in AP (mostly *Pinus*) in the uppermost part of the sequence records the Oerel Interstadial (early MIS 3). The only radiocarbon date of 22,500 ± 1,000 yr BP obtained on peaty silts from the uppermost part of the Horoszki Duże lake sequence is considered a minimum age only, probably being contaminated by young organic matter.

Granoszewski (2003) employs a cautious approach in inferring past temperatures from the macrofossil record at Horoszki Duże. Plant macrofossils, however, have been widely used in Europe to infer climatic conditions for the Late Pleistocene (e.g. Aalbersberg and Litt 1998, Bos et al. 2001). Certain plants (indicator plant species) require a specific minimum mean July air temperature in order to flower and reproduce (e.g. Iversen 1954, Kolstrup 1980). Isarin and Bohncke (1999), for instance, use fossil remains of the wetland plant *Typha* to explore regional climate patterns over



**Figure 3-4.** Simplified lithology and pollen and spore record of the Horoszki Duże sedimentary sequence in eastern Poland (Figure 1-1) based on Granoszewski (2003). Pollen of *Phragmites* (*Poaceae*) are excluded from the total sum. The uncalibrated 14C age to the left is considered to be a minimum age. Chrono-stratigraphy and marine stages, with two alternative correlations, are given to the right. Alternative A is favored in the present report and is therefore high-lighted in the figure; possible correlations with the Woillard and Montaigu events of the La Grande Pile record are also according to the present review. Temperatures are inferred from indicator plant species identified by Granoszewski (2003) in the macrofossil record. Note that the 10x exaggeration line for the *Pinus* pollen curve is missing.

central Europe during the Younger Dryas. Their present-day distribution map clearly depicts the strong relationship between the northern limit of *Typha latifolia* and the July isotherms of 13°C (in maritime Europe) to 15°C (more continental sites). Referring to Iversen (1954), Granoszewski (2003) e.g. quotes that ‘a lake-type body of water may have had a specific microclimate different from the general one’. Present-day distribution maps for indicator plant species in northern Europe (e.g. Väiliranta et al. 2009), however, clearly show that these plants do not occur outside their main distribution areas, even though local factors, e.g. around a small lake surface, might create warmer conditions than the regional climate. Because of the strong relationship between indicator plant species and regional climate, the rich macrofossil record at the Horoszki Duże site is used here to explore past climate conditions.

### 3.3.1 Environmental reconstructions for MIS 5

The first pollen zone of late Saalian age (Wartanian Glaciation; MIS 6) shows relatively high pollen percentages for *Betula nana*, *Artemisia*, Chenopodiaceae and Poaceae, and other steppe elements such as *Helianthemum* and *Centaurea* types, reflecting an open vegetation with only a small proportion of trees (e.g. *Larix*) in the vicinity of the lake (Figure 3-4). High amounts of Tertiary sporomorphs in the calcareous silt, and pollen of e.g. thermophilous trees, probably result from erosion processes in a soil not fully stabilized by a continuous plant cover (cf. Mamakowa 1989).

A complete interglacial plant succession is recorded in organic silts and gyttja overlain by peaty sediment. The vegetation succession can be correlated with the forest history at the Eemian Interglacial (MIS 5e) stratotype in The Netherlands (Zagwijn 1961) and is characterized by the spread of trees and hazel in the order: *Betula-Pinus*, *Ulmus*, *Quercus*, *Fraxinus*, *Corylus*, *Taxus*, *Tilia*, *Carpinus*, *Picea-Abies*, and *Pinus* (-*Betula*). Pioneer *Betula* (-*Pinus*) forest with open plant communities including Poaceae, *Artemisia* and the shrub *Betula nana* marks the lower boundary of the Eemian succession at Horoszki Duże. A strong expansion of riverine forests and the first appearance of *Hedera helix* towards the end of the *Quercus-Fraxinus-Ulmus* zone are interpreted as indicating a change from a continental to more maritime climate regime. The maximum spread of *Corylus* marks the beginning of the climatic optimum and the presence of *Trapa* and *Tilia tomentosa* in the following deciduous multi-species forest indicates a warm climate with moderate rainfall, though higher than today. Relatively high winter temperatures persisted until the *Carpinus* peak (presence of e.g. *Hedera helix*). Macrofossils of the aquatic plant *Najas marina* indicate minimum mean July temperatures of 17°C through this early and middle part of the Eemian at Horoszki Duże (Figure 3-4; compared to the present-day mean July temperature value of ca. 18°C in Figure 1-1).

A deterioration of the climate in the late Eemian is indicated by peak occurrences in *Picea* and *Abies* followed by *Pinus* forest in which all thermophilous trees (including *Abies*) ceased to occur. *Larix* was present in the final boreal forest and the amount of both herbs (e.g. Poaceae, *Artemisia*) and shrubs (e.g. *Betula nana*) increased. The *Pinus* curve is tripartite, including an increase in tree *Betula* and *Picea*, possibly indicating a short-lived, cold (and moist) oscillation. Although not noted upon by Granoszewski (2003), the latter most probably corresponds to the Woillard event at La Grande Pile (Figure 3-1). Organic content (determined by loss-on-ignition, LOI) gradually rises through the Eemian to ca. 50%; peaty sediment started to accumulate during the *Picea/Abies* peak. LOI eventually decreases to 10% as NAP increases and gets diverse and the vegetation shows features of forest tundra.

During the Brørup (MIS 5c), *Betula* forest started the forest development. This was followed by *Betula-Pinus* forest in which the presence of *Humulus lupulus* points to a relatively mild boreal climate. The initial expansion of tree *Betula* was interrupted by a brief cold oscillation. High proportions of *Artemisia*, Poaceae and Chenopodiaceae indicate increased continental climate conditions; tree *Betula* and *Larix*, however, remained present in the open vegetation. Although not mentioned by Granoszewski (2003), the cold oscillation most probably corresponds to the Montaignu event at La Grande Pile (Figure 3-1). In the *Pinus-Picea-Larix* (-*Betula*) forest during the late Brørup, increasing contributions of light-demanding plants, particularly Poaceae, *Artemisia* and Chenopodiaceae, point to a thinning of the forest and final formation of forest tundra of a continental type. LOI attains values of 80% during the Brørup. The accumulation of peaty sediment is interpreted as indicating a fall in lake level due to a ‘drying’ effect of the forest (see under 3.2.1.). LOI drops to ca. 20%, and the sediment is sandy, during the cold oscillation; the algae *Pediastrum*, which is well-represented during the cold phases of the Horoszki Duże sequence (Figure 3-4), reaches peak values and might

indicate improved light conditions in the lake water. The Odderade (MIS 5a) starts with a brief *Betula* phase followed by the establishment of boreal *Pinus-Betula* forest (climate optimum) and then *Pinus* (*-Betula*) forest with *Larix* and *Picea*. An increasing contribution of steppe elements is registered through major part of the Odderade. Peat accumulated in the Horoszki Duże depression during the early part of the Odderade (LOI to ca. 70%), changing to organic silts in the latter part (LOI dropping to 10–20%).

According to Granoszewski (2003), the forest communities that developed during the Brørup and Odderade at Horoszki Duże were probably open park-like in character, at least in places. During major parts of the Brørup, a rich macrofossil flora with *Carex elata*, *R. sceleratus*, *Bidens tripartita*, *Typha* and *Rumex maritimus* points to minimum mean July temperatures of 15°C; *N. marina* indicates values of 17°C during the mild *Betula-Pinus* forest phase (Figure 3-4). During the initial *Betula* phase of the Odderade, the presence of macrofossils of *R. maritimus*, *Typha*, *C. elata* and *R. sceleratus* indicate minimum mean July temperatures of 15°C.

The Herning (MIS 5d) and Rederstall Stadials (MIS 5b) are characterized by a dominance of open plant communities showing a considerable diversity, including shrub tundra, steppe-type communities and damp meadows. High proportions of Poaceae, *Artemisia* and Chenopodiaceae, and the presence of e.g. *Ephedra fragilis* type, would indicate a continental climate. Macrofossils of tree *Betula*, *Larix*, and *Picea*, however, indicate the local presence of these trees. The presence of forest tundra in the latter part of the Herning Stadial is suggested by increasing pollen values for tree *Betula*. The Herning Stadial shows the first recording of pollen of the subalpine Balkan species *Bruckenthalia spiculifolia* (see under 3.1.1.). *B. spiculifolia* is near-continuously represented at Horoszki Duże from the latter part of the Brørup onwards. Of aquatics/wetland species, macrofossils of *R. maritimus* in the Rederstall sediments, combined with those of *R. sceleratus*, *B. tripartita* and *Typha* in the Herning, indicate mean July temperatures of 15°C during the latter parts of both stadials. Silty sediment accumulated in the depression during the Herning and Rederstall. Characteristic are also the high abundances of *Pediastrum* spores in the sediment, a relatively high representation of *Sphagnum* spores, and the inferred extensive reedswamp communities with *Phragmites* along the former lake.

### 3.3.2 Comparison with other records

During MIS 5, AP values (particularly *Betula*) are generally higher and NAP (particularly *Artemisia*) lower at more westerly located sites in central Europe such as Gröbern (Kühl et al. 2007) and Füramoos (Müller et al. 2003) compared to Horoszki Duże. Additionally, *Betula* shows a distinct peak both in the beginning and closing phases of the Eemian at Gröbern. This pattern is ascribed by Caspers and Freund (2001) to more continental conditions over eastern Europe during the Eemian and Early Weichselian compared to west Europe. The Horoszki Duże site has a sub-continental climate also at present (Figure 1-1).

### 3.3.3 Environmental reconstructions for MIS 4-3

A marked change is recorded at the transition to the Middle Weichselian (MIS 4; Figure 3-4). Poaceae show high pollen values in the silty sediments in the upper part of the Horoszki Duże record, *Artemisia* is near continuously registered, and tree *Betula* pollen values are overall low (less than 20%). Furthermore, *Pediastrum* and *Phragmites* have near continuous records. AP frequencies show that the site was situated outside the northern line of closed forest for a considerable time. The almost continuous presence of *Armeria maritima* suggests a thin snow cover (Kolstrup 1990). NAP, however, reflects a high diversity of habitats. Macrofossils of aquatic plants, including *Myriophyllum spicatum*, *M. verticillatum*, *Ranunculus flammula* and *Batrachium*, indicate minimum mean July temperatures of 13°C; the fern *Pteridium aquilinum* points to values of 14°C.

Three zones are distinguished. LOI is still relatively high (up to ca. 15%) in the lower zone. Organic silts followed by silty-sandy gyttja were deposited in the lake. Vegetation is inferred as steppe-tundra. During two intervals, however, small stands of tree *Betula* (macrofossils present), *Pinus* and *Larix* occurred. A correlation is suggested with the Oerel and Glinde Interstadials (early MIS 3; according to chrono-stratigraphic variant “A”), or, alternatively, with the lowermost part of the Schalkholz Stadial (MIS 4; “B” in Figure 3-4).

During the middle zone, mostly silts accumulated in the depression, LOI drops to less than 10%, and AP shows minor fluctuations only. Although the significance of steppe-tundra had further increased, individual *Betula*, *Pinus* and *Larix* trees were probably consistent elements of the local flora. Initially, a rich composition of steppe communities with *Ephedra distachya* and *E. fragilis* types is recorded; later on, steppe elements slightly decline, *Pinus* pollen increase, needles of *Larix* as well as macro-remains of tree *Betula* are found, and *R. maritimus* and *Typha* indicate minimum mean July temperatures of 15°C (Figure 3-4). A correlation is proposed with the Moershoofd and Hengelo Interstadials (MIS 3), or, alternatively, this zone represents the upper part of the Schalkholz Stadial (MIS 4).

Mostly peaty silt with sand was deposited during the upper zone. This zone represents the maximum spread of *Pinus* (wood present), and overall lowest *Artemisia* values, but communities of open habitats continued to dominate. The reappearance of *Betula* cf. *humilis* macrofossils and fragments of *Alnus* wood are further signs of amelioration towards a warmer and moister climate. *R. sceleratus* points to minimum mean July temperatures of 15°C. This phase was followed by the complete recession of *Pinus* and absolute dominance of communities of open areas; steppe communities developed strongly. The Denekamp (late MIS 3) or Oerel Interstadial (early MIS 3) is thought to be recorded in the initial part of the zone.

### 3.3.4 Comparison with other stratigraphies/records

The variant “A” chrono-stratigraphy of Granoszewski (2003) is favored in the present review paper. The interstadial intervals are recorded before the main opening of the vegetation, with *Artemisia*/Poaceae steppe biomes (uppermost part of pollen diagram in Figure 3-4), which can be expected to correlate with MIS 2. This implies, however, that MIS 2, as well as MIS 4, is barely represented in the sediment sequence. This is possibly due to subaerial exposure under very dry climate conditions.

The Denekamp and Hengelo Interstadials, and the Moershoofd Interstadial complex, have been originally defined in The Netherlands (van der Hammen et al. 1967, Zagwijn 1974). The interstadials, which show shrub tundra to tundra vegetation, are interpreted to correspond to phases of relatively high rainfall and temperature (Ran 1990, van der Hammen 1995). July temperatures up to ca. 13–15°C have been reconstructed by Kolstrup and Wijmstra (1977). The Denekamp is dated to around ca. 30 (uncalibrated) kyr BP and the Hengelo to ca. 38 kyr BP (Ran 1990). These ages roughly correspond to those for the Grand Bois Interstadial Complex (ca. 29–31 kyr BP) and the Charbon Interstadial (around ca. 40 kyr BP), respectively, at La Grande Pile (Woillard and Mook 1982). The Moershoofd Interstadial Complex (dated to broadly around ca. 48 kyr BP; Zagwijn 1961, Ran 1990) most probably corresponds to the Glinde and Oerel Interstadials at Oerel (48–58 kyr BP; Behre and van der Plicht 1992) and the Pile Interstadial Complex at La Grande Pile (ca. 50–62 kyr BP; Woillard and Mook 1982). Van der Hammen (1995) correlates the Denekamp and Hengelo with Greenland Interstadials (GIS) 8 and 11, respectively. The Denekamp, Hengelo, Glinde and Oerel Interstadials are aligned to GIS 8, 12, 14 and 16, respectively, by Dansgaard et al. (1993; Figure 4-1).

Near and along the Baltic Sea coast north of Horoszki Duże, interstadial deposits dated to MIS 3 and studied by pollen analysis have been described from Estonia and Lituania (Voka and Purviai/Medininkai sites, respectively, in Figure 1-1). Water-lain sediments filling a palaeo-incision near Voka in Estonia have been dated by IR-OSL to the time interval 39–33 kyr BP (Molodkov et al. 2007). Detailed pollen data shows two relatively mild interstadials at around ca. 38 and at ca. 37–35.5 kyr BP. For the older stadial, and around ca. 34 kyr BP, forest tundra with *Pinus* (and *Picea* and *Larix* at ca. 38 kyr BP) is reconstructed. During the younger interstadial, at least two wetter and warmer climate phases are recorded with open coniferous forest of *Pinus* (including *P. sibirica*) and *Picea*. The open forested intervals alternate with periods of steppe-tundra vegetation (Molodkov et al. 2007). A palaeo-lacustrine basin near Venta in north-western Lituania, including pollen-bearing fluvial and lacustrine sediments at the Purviai outcrop dated by <sup>14</sup>C to ca. 33.5 (uncalibrated) kyr BP, is described in Satkunas et al. (2009). Despite a significant contribution of reworked or long distance-transported pollen, interstadial conditions are reconstructed of park tundra vegetation with *Betula*, *Pinus* and some admixture of scattered *Picea* and *Juniperus*; additionally, pollen of the more warm-demanding wetland plant *Typha* is recorded (indicating a minimum mean July temperature of 15°C). The interstadial is correlated with the youngest phase of forest tundra at Voka, and tentatively with the Denekamp Interstadial (Satkunas et al. 2009). At Medininkai in eastern Lituania, a climate

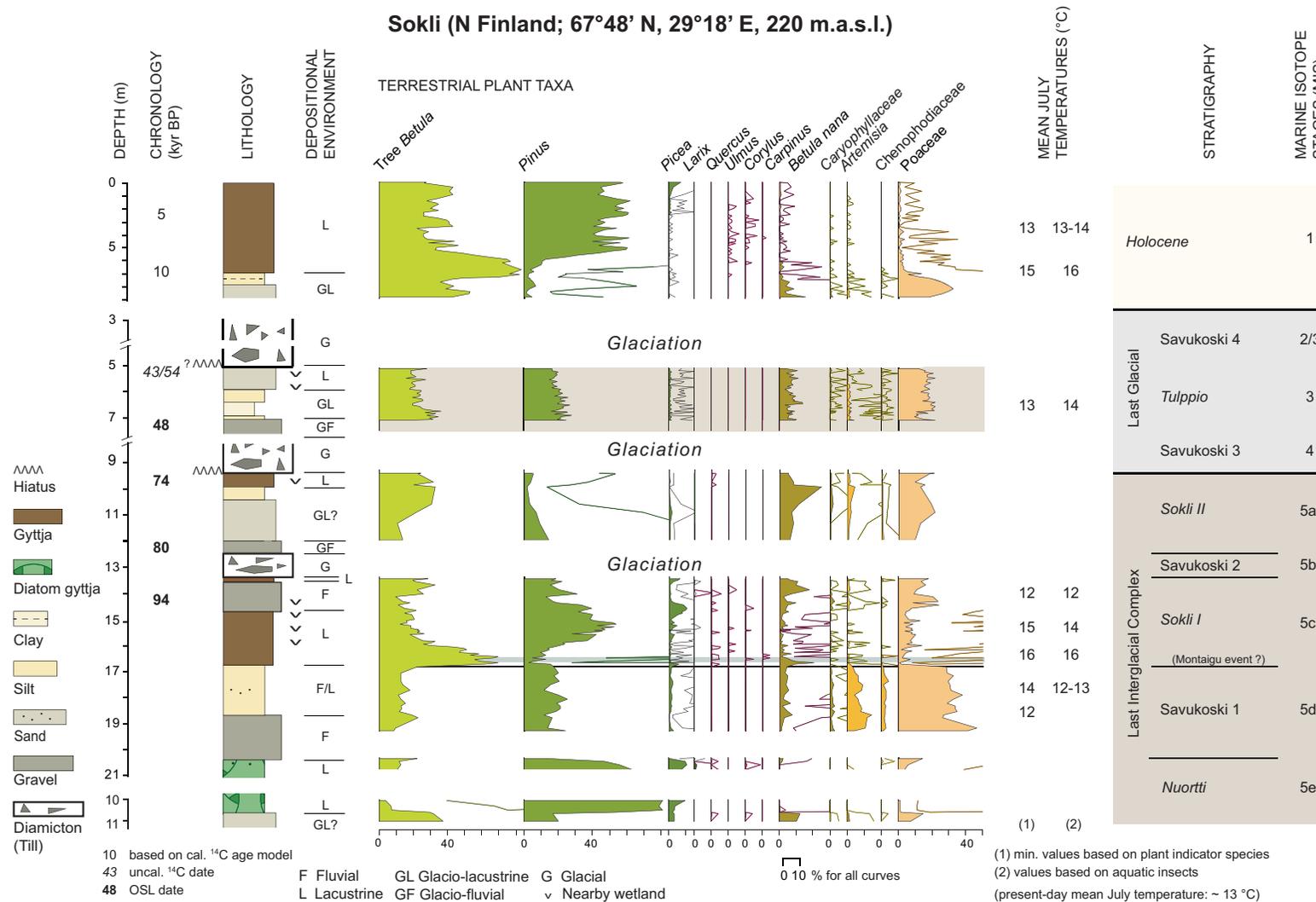
amelioration characterized by AP to 60% (mainly *Betula*) as well as macrofossils of *Pinus sylvestris* and *Betula humilis* is dated by U/Th to ca. 42 kyr BP and correlated with the Hengelo Interstadial (Satkunas et al. 2003).

The sites in Estonia and Lituania are presently situated in the hemiboreal forest consisting of *Pinus* and *Picea* mixed with temperate tree taxa (Figure 1-1). The Late Pleistocene of Lituania has been extensively studied (see Satkunas et al. 2009, and references therein) and includes the Merkinė (Eemian) Interglacial with broad-leaved forests and the Jonionys 1 and 2 Interstadials (correlating to Brørup and Odderade) showing *Pinus-Picea-Larix* and *Pinus-Picea-Betula* forests, respectively. The Jonionys 3 Interstadial, characterised by *Betula* (*B. nana*) forest-tundra vegetation, is the main climate-stratigraphic event of MIS 3 age traceable in sections of Lituania; it is correlated with the Oerel Interstadial.

### 3.4 Sokli (Finland)

The Sokli basin in north-eastern Finland (lat. 67°48'N, long. 29°18'E; elevation 220 m.a.s.l.; Figure 1-1) holds an up to ca. 30 m thick sediment sequence that spans the last ca. 140 kyr. Current vegetation is northern boreal forest with *Betula*, *Pinus* and *Picea*. The Sokli basin has been carved out by glacial erosion in relatively soft deeply weathered rocks of a Palaeozoic magma-intrusion (Sokli Carbonatite Massif). Erosion seems to have occurred during the late Saalian (MIS 6) when the ice-divide zone of the Fennoscandian Ice Sheet (FIS) was situated over the northernmost part of Finland (Hirvas 1991). It is the protected location of the Late Pleistocene sediments in the small basin (ca. 2 km<sup>2</sup>) combined with limited glacial erosion under this central area of Late Weichselian (MIS 2) glaciation that are thought to be major factors for the preservation of this for Fennoscandia unusual long and continuous sediment sequence (Ilvonen 1973, Helmens et al. 2007a). A reconnaissance study on the entire sequence, including lithology, palynology and <sup>14</sup>C and TL dating, was made by Helmens et al. (2000), following earlier work by Ilvonen (1973) and Forsström (1990). With the recovery of continuous boreholes from the central part of the Sokli basin in 2002 and 2010, fine-grained sediment intercalations have been subjected to high-resolution, multi-proxy analyses (pollen, macrofossils, diatom, chironomids, geobiochemical properties). Climate parameters have been reconstructed using the transfer function approach and indicator plant species. So far, analyses have been completed on deposits dated to early MIS 3 (Helmens et al. 2007b, 2009, Engels et al. 2008, Bos et al. 2009), MIS 5d-c (Väliranta et al. 2009, Engels et al. 2010, Helmens et al. 2012) and the Holocene (Salonen et al. 2013, Shala et al. 2012, 2013, 2014). These papers form the basis for the descriptions and interpretations given below. Detailed studies on a nine meter thick gyttja bed dated to MIS 5e are in progress.

A tentative correlation of the stratigraphy in the Sokli basin with the regional stratigraphy of Finnish Lapland (Korpela 1969, Hirvas 1991), and the northwest European mainland and marine stratigraphies, was made by Helmens et al. (2000). The results of a detailed OSL dating program using SAR dose protocol were presented in Alexanderson et al. (2008). Although in sequence, the quartz OSL dates have large standard errors mainly due to small sample sizes, relatively poor luminescence characteristics, and uncertainties in dose-rate determinations. Nevertheless, the ages group according to stratigraphic units and are in agreement with the earlier made land-sea comparison (Helmens et al. 2007a; Figure 3-5). The lower part of the Sokli sedimentary sequence includes two thick gyttja deposits with interglacial boreal forest pollen floras, separated by sand and gravel, which correlates to MIS 5e–5c. These sediments, which can be traced from the central to the marginal part of the basin, overly glacio-lacustrine silt and sand, and till (MIS 6). The upper part of the sequence comprises three till beds each overlain by glacio-fluvial and -lacustrine sediment. The lowermost glacial deposit is overlain by gyttja (MIS 5a) and the middle one by sands (early MIS 3), whereas the uppermost glacio-lacustrine sediments grade into Holocene gyttja and peat. The three gyttja beds and associated minerogenic sediment in the lower portion of the Sokli record are here defined as the Nuortii Interglacial (MIS 5e) and Sokli I (MIS 5c) and Sokli II (MIS 5a), and the intercalated sands/gravels (MIS 5d) and till (MIS 5b) are defined as the Savukoski 1 and 2 Stadials, respectively. The interstadial of early MIS 3 age at Sokli has been defined the Tulppio Interstadial (Helmens et al. 2007a), and the stadials preceding and following the Tulppio Interstadial are here named Savukoski 3 (MIS 4) and Savukoski 4 (late MIS 3 – MIS 2). Shrub tundra with possible tree birches, and near current climate conditions, is reconstructed for the Tulppio Interstadial. The Sokli basin was glaciated during Savukoski 2-4.



**Figure 3-5.** Simplified chronology, lithology combined with inferred depositional environments, and pollen record of the Sokli sedimentary sequence in northern Finland (Figure 1-1). Stratigraphy and marine stages are given to the right as well as temperature reconstructions based on macrofossils of indicator plant species and on chironomids. The Late Pleistocene record combines data from the central Sokli basin (depth interval 3–21 m) with evidence from the basin's margin (10–11.5 m). The Holocene record is from Lake Loitsana in the western portion of the Sokli sedimentary basin. Based on Bos et al. (2009), Engels et al. (2008, 2010, personal comm.), Helmens et al. (2000, 2007a, 2009, 2012), Shala et al. (2012, 2014) and Salonen et al. (2013); local stratigraphy is adapted from Helmens et al. (2007a). In the most recent borehole (2010) from the near-central portion of the Sokli basin, the Nuortti diatom gyttja bed (MIS 5e) and the Sokli II gyttja deposit (MIS 5a) attain total thicknesses of nine and three meter, respectively. These sediments are presently subject to detailed studies; first results on the Nuortti gyttja deposit support the vegetation development depicted in Figure 3-5 (J.S. Salonen and K.F. Helmens, unpublished data).

### 3.4.1 Environmental reconstructions for MIS 5

The pollen assemblages encountered in the Nuortti and Sokli I gyttja beds (MIS 5e and 5c, respectively) reveal a vegetation succession very similar to the one recorded in the Holocene gyttja deposit in the Sokli basin (Figure 3-5). They show the establishment of pioneer *Betula* forest followed by the spread of *Pinus* and then *Picea*. *Larix* was present both during the Nuortti and Sokli I but *Larix* pollen have not been encountered in Holocene records from Finland. The present range of larch starts some 500 km east of Sokli (Figure 1-1) where, with increasing continentality over northern Russia, *Larix sibirica* becomes an important component of the northern taiga and the vegetation near the arctic and alpine forest limits. The presence of *Larix* might point to an enhanced degree of continentality during Sokli I (MIS 5c) compared to the Holocene. This is supported by detrended correspondence analysis (DCA) ordination, in which the Sokli I pollen assemblages show a better fit with west Russian calibration samples compared to the local, Fennoscandian calibration set (Salonen et al. 2013). A high-resolution pollen study on the Nuortti interglacial bed is in progress. A decrease in *Betula* pollen is registered during early Sokli I (Figure 3-5), similar as during the early Brørup at Oerel (Figure 3-3), and might correspond to the Montaigu phase at La Grand Pile in France. Tree *Betula* started to dominate again during the late Nuortti and late Sokli I. The cooling at the end of Sokli I resulted in glaciation of the Sokli site (Savukoski 2 Stadial; MIS 5b).

The multi-proxy records reconstruct different local conditions during the Nuortti, Sokli I and Holocene interglacial intervals. A relatively large and deep lake, bordered by a limited zone of wetland only, is recorded for major part of the Nuortti. The lake history during the Holocene was complex. It included an early phase with glacio-lacustrine sedimentation followed by a lake phase with changing influxes of fluvial sediment due to adjacent wetland development and expansion; the lake was mostly fed by groundwater from a nearby esker during the late Holocene. The infilling of an oxbow lake and subsequent return to stream channel deposition is recorded during Sokli I. Due to the gradual terrestriation of the oxbow lake, Poaceae (Figure 3-5) as well as Cyperceae are over-represented in the pollen diagram, artificially reducing AP during Sokli I; ferns and the tree *Alnus glutinosa* formed important elements in the shore vegetation. Currently, the northern limit of *A. glutinosa* is situated some 400 km south of Sokli. Rich macrofossil assemblages in the Sokli I sediments, including *Najas tenuissima*, *Elatine triandra*, *Typha*, *B. tripartita*, *Glyceria fluitans*, *R. sceleratus* and *A. glutinosa*, indicate that the boreal environment at Sokli during MIS 5c experienced July temperatures several degrees higher than the present-day mean value of 13°C (Figure 3-5). Chironomid-inferred mean July air temperatures are generally somewhat lower. The difference might be due to the limited temperature gradient of the modern chironomid calibration set which at the warm end reaches to 16°C. Due to this so-called ‘edge-effect’, the true paleo-temperatures might be underestimated in the chironomid-based reconstruction.

The Nuortti and Sokli I deposits are separated by fluvial gravels and sands overlain by silty sediment deposited by a braided river system (Savukoski 1 Stadial; MIS 5d). The pollen record in the upper gravelly and overlying silty sediment, with high percentage values of Poaceae, *Artemisia* and Chenopodiaceae as well as *Betula nana*, indicates the presence of a steppe-tundra. The local presence of conifers and *Betula* trees is indicated in the macrofossil record. Mean July temperatures of at least 12–14°C are recorded by chironomids and indicator plant species (e.g. *Crassula aquatica*). The steppe-tundra vegetation, high summer temperatures and inferred braided river system indicate severe continental climate conditions. The chironomid-inferred July temperatures for Savukoski 1 are somewhat lower than those estimated by plant macrofossils. The application of a transfer function using chironomid-climate calibration data from Siberia, instead of Norway-Svalbard in Figure 3-5, has produced July temperatures more in line with the macrofossil-inferred values (Engels, personal comm.). This can be explained by the fact that chironomid taxa require higher July temperatures under more continental conditions in compensation for e.g. a shorter growing season (Self et al. 2011).

Sokli II (MIS 5a) is represented by laminated sands and silts grading into sandy gyttja and a vegetation development towards open *Betula* forest or shrub tundra close to the *Betula* tree-line. Recent coring has recovered a much thicker sandy gyttja bed (up to two meter in thickness) than the one indicated in Figure 3-5. It is probable that the bed in Figure 3-5 has been truncated during subsequent glaciation (MIS 4) and as such does not give the complete vegetation development at Sokli during MIS 5a. A rich wetland assemblage is recorded suggesting a relatively small, shallow lake. This return to lacustrine sedimentation in the central Sokli basin is most probably the result of compaction of the older part of the sedimentary sequence during the previous MIS 5b glaciation, creating a new water-logged depression.

### 3.4.2 Comparison with other records

A till-covered gyttja deposit at Tepsankumpu (Finnish Lapland; Figure 1-1) has been interpreted to represent a major part of the Eemian Interglacial (MIS 5e; Hirvas 1991, Saarnisto et al. 1999) based on litho-stratigraphic position and correlation with pollen sequences further south, including fragmented marine sequences from Ostrobothnia in western Finland (e.g. Eriksson 1993). The Tepsankumpu Interglacial (Donner et al. 1986) is characterized by an initial phase with *Betula* forest followed by the spread of *Pinus*, *Picea* and then *Larix*. All pollen zones are tree *Betula*-dominated, however, and its correlation with MIS 5e is not totally proven. *Corylus* and *Quercus* are represented near continuously by low pollen percentage values in the *Betula-Pinus* forest phase suggesting that their northern limits were further north during the interglacial temperate sub-stage than during the Holocene climatic optimum in northern Finland (Saarnisto et al. 1999). The *Corylus* record is also near-continuous throughout major part of the *Betula-Pinus-Picea* zone at Leveaniemi in northern Sweden (Figure 1-1), which represents the most complete Swedish site with sediments correlated to the Eemian (Lundqvist 1971, Robertsson 1991a, 1997, 2000). The composition of the pollen flora together with plant macrofossils and insect remains reflect a climate more favourable than during the Holocene optimum; mean annual temperature was possibly as much as 4°C higher than today in the same area (Robertsson 1997 and references therein).

### 3.4.3 Environmental reconstructions for MIS 4-2

Inter-bedded in-between two till beds (Savukoski 3 (MIS 4) and Savukoski 4 Stadials (late MIS 3 – MIS 2), the Tulppio Interstadial sediments (early MIS 3) are comprised of glacio-fluvial gravel and sand with on top a two meter thick laminated clay-silt sequence (Figure 3-5). The clay-silt unit fines up to clayey sediment and then coarses up to sands at the top. Multi-proxi evidence indicates that the clay-silt sequence, and similar sediments found underlying the Holocene gyttja deposit, is of glacio-lacustrine origin. Geomorphologic and Digital Elevation Model (DEM) data have been used to reconstruct the glacial lake evolutions showing that the coring-site is located in a previous sheltered lake embayment and that several rivers draining most probably non-glaciated terrain flowed into the glacial lake. This explains the general limited influence of the ice sheet at the coring-site, resulting in proxy records that in detail register the surrounding biotic environment and reconstruct the regional climate.

Radiocarbon dating on seeds of tree *Betula* places the last deglaciation of the Sokli area just prior to ca. 10,700 (calibrated) kyr BP, i.e. near the Late Glacial-early Holocene boundary, in agreement with the deglaciation chronology of Johansson (2007). It also shows that tree birches were present on the recently deglaciated terrain surrounding the glacial lake. Macrofossils of tree *Betula* were not encountered in the early MIS 3 glacio-lacustrine sequence but *Betula* pollen percentages here of up to 30% (and of *Pinus* to 20%) indicate that the distributional ranges of these trees were probably only few hundred kilometres south or south-east of Sokli; it is possible that tree *Betula* was present in the Sokli area in favourable spots. The Tulppio Interstadial pollen assemblages reflect a rich, low-arctic shrub tundra vegetation remarkably similar in composition to modern tundra in the continental sector of northern Fennoscandia. Mean July air temperatures in the magnitude of present-day values are reconstructed by chironomids, macrofossils of aquatic plants (*Potamogeton mucronatus*) and of Bryzoa, and diatoms. Percentages values for steppe elements such as *Artemisia* and Chenopodiaceae are low in the pollen record (Figure 3-5) and suggest that also winter temperatures were not significantly lower than today. Large abundances of pooid phytohyths in the Tulppio Interstadial as well as early Holocene glacial lake sediment reveal that the high pollen percentage values for Poaceae are probably due to an over-representation of pollen from a local wetland source. A shallowing lake with an expanding wetland zone (and increasing Cyperceae pollen percentages) is registered in the upper Tulppio sandy sediment.

An early MIS 3 age for the Tulppio Interstadial is based on OSL and <sup>14</sup>C dating (Figure 3-5). The interstadial is correlated with the prominent GIS 14 at around 53 kyr BP (Figure 4-1). The correlation takes into consideration that gradual retreat in the northeastern sector of the FIS, from its MIS 4 limit in Russia (Figure 1-1), probably started during OIS 16 in the earliest part of MIS 3 (at ca. 60 kyr BP). If the laminations in the Tulppio sediment have an annual origin, then only some 400 years of sedimentation is recorded (Helmens et al. 2000). It is possible that the Sokli area subsequently was glaciated soon again, or that ice-free conditions persisted longer into MIS 3 but evidence for this has been eroded away during the last glaciation.

### 3.4.4 Comparison with other records

The age of a 15 cm-thick organic sediment layer, found interbedded in-between till west of Sokli near Petäjäseltä (Figure 1-1), has recently been bracketed by OSL (including dates of ca. 73–58 and 32 kyr BP on sand below and above the organic bed, respectively) and  $^{14}\text{C}$  dating (ca. 35 (uncalibrated) kyr BP on wood from the organic bed) to late MIS 3 at ca. 35 kyr BP (Sarala and Eskola 2011, Väiliranta et al. 2012). The pollen assemblage with around 30% for tree *Betula* and up to 40% *Pinus* is interpreted to represent an environment at or very close to the tree line. Macrofossil evidence shows the local presence of tree *Betula* and of the boreal aquatic plant *Nuphar* indicating a minimum mean July temperature of 12°C (Väiliranta et al. 2012). Till-covered peat containing large pieces of wood from central Finnish Lapland has recently been dated by OSL and  $^{14}\text{C}$  on wood to the early part of MIS 3 (Väiliranta M, pers. comm.); the peat contained abundant macrofossils of tree *Betula*.

Extensive dating by OSL, TL and  $^{14}\text{C}$  on sediment sequences from southwestern Finnish Lapland (northern Finland) has been conducted by Mäkinen (2005). The sediments include organic beds with pollen assemblages typical for the so-called Peräpohjola Interstadial defined by Korpela (1969). The assemblages exhibit varying proportions of *Betula* and *Pinus* pollen and indicate *Betula*-dominated open vegetation. At Kauvonkangas (Figure 1-1), the local presence of *Salix* stands and *Betula nana* heaths, as well as watery, sedge-dominated mires and grasslands, are reconstructed; tree *Betula* pollen percentage values reach to 35% (Eriksson 2005). Originally, Korpela (1969) assigned a Middle Weichselian (MIS 3) age to the Peräpohjola Interstadial based on the radiocarbon chronology which consisted of mostly infinite ages. Based on later studies on the stratigraphy of Finland and Sweden and correlation with the northwest European mainland stratigraphy, an Early Weichselian age (MIS 5c or 5a) was considered as more likely (Hirvas and Nenonen 1987, Donner et al. 1986, Forsström 1988). However, the new chronology by Mäkinen (2005) shows a clear cluster of dates in the range 57–37 kyr BP and this strongly suggests an MIS 3 age for the Peräpohjola sediments. Radiocarbon datings on mammoth bones from central and southern Finland have further provided a consistent series of dates from 32 to 22.5 kyr BP indicating the existence of large ice-free areas in Finland during late MIS 3 – early MIS 2 (Ukkonen et al. 1999).

Wohlfarth (2010) recently re-assessed the quality of both published and unpublished  $^{14}\text{C}$  age measurements on pre-LGM organic deposits from Sweden. The evaluation shows that acceptable datings range between ca. 60 and 35 kyr BP for northern and central Sweden and 40–25 kyr BP for southern Sweden. Radiocarbon datings on mammoth bones from central and southern Sweden have been radiocarbon-dated to mainly ca. 44–26 kyr BP (Ukkonen et al. 2007). These chronologies for ice-free conditions in Sweden during MIS 3 are in line with those obtained for Finland mentioned above. As mentioned by Wohlfarth (2010), a conclusion for possible ice-free conditions in central and northern Sweden during parts of MIS 3, based on finite  $^{14}\text{C}$  ages in the range of ca. 40–30 kyr BP, was earlier made by Lundqvist (1967, 1978). The latter view, however, was also later revised and the till-covered sediments in Sweden were mostly correlated with the MIS 5 Brørup or Odderade Interstadials (e.g. Robertsson 1991b, Robertsson and García Ambrosiani 1992).

The sites evaluated in Wohlfarth (2010) include the stratigraphic key sites at Pilgrimstad (central Sweden) and in Norrbotten (northern Sweden). The till-covered gyttja at Pilgrimstad (Figure 1-1) contains a pollen flora reflecting an open subarctic vegetation of shrubs and herbs with *Betula* trees and possible some admixture of *Picea*; the shrubs *Juniperus* and *Betula nana* were important components in the flora together with a rich herbaceous plant cover (Robertsson 1988, 1991b). The Pilgrimstad gyttja has been correlated with the Jämtland Interstadial defined by Lundqvist (1967). A recent re-examination of the Pilgrimstad sediments by Alexanderson et al. (2009) have provided OSL ages between 52 and 36 kyr BP and  $^{14}\text{C}$  ages on twigs at ca. 44 and 39 kyr BP, in line with the  $^{14}\text{C}$  chronology accepted in Wohlfarth (2010). Furthermore, recently performed detailed OSL dating at Idre (Figure 1-1) date a deglacial phase with glacio-lacustrine sedimentation to early MIS 3 at ca. 54–41 kyr BP (Möller et al. 2012).

A distinct set of till-covered glacial landforms associated with fine-grained sediments rich in fossils is present in north-eastern Norrbotten. The local stratigraphy was studied in detail by Lagerbäck and Robertsson (1988) and was recently re-investigated by Hättestrand (2008) and Hättestrand and Robertsson (2010). In the sediment sequence recovered from a kettle hole in the esker Riipiharju (Figure 1-1), three interstadials are distinguished, i.e. Tärendö I, IIa and IIb (Hättestrand 2008). The Tärendö IIa and IIb interstadial sediments hold pollen assemblages in which tree *Betula* reaches

values to 40–50% separated by a cold and dry phase with distinct representations of *Artemisia* and Poaceae. Similar as at Pilgrimstad, *Pinus* pollen percentage values are low. Hättestrand (2008) has tentatively correlated the *Betula* phases of Tärendö II with GIS 14 and 12. This early MIS 3 age assignment is supported by the radiocarbon chronology re-examined in Wohlfarth (2010), with acceptable  $^{14}\text{C}$  ages for the sediments in NE Norrbotten falling in the time-interval ca. 50–35 kyr BP, and preliminary OSL dating at Riipiharju that suggests that Tärendö II is younger than 70 kyr BP (Alexanderson et al. 2011).

In Norway, an extensive data set of geological sections and absolute age determinations (mostly  $^{14}\text{C}$  dates) from the mountainous coast has suggested rapid phases of ice retreat (reaching to inland areas) along the western margin of the FIS during the late part of MIS 3 and MIS 2 (Olsen et al. 2001). An interstadial dated to MIS 3 along the Norwegian coast is the Ålesund Interstadial defined on the basis of cave sediments (Mangerud et al. 1981). Over 30  $^{14}\text{C}$  dates have been obtained on well-preserved bones from the Ålesund Interstadial beds, yielding ages in the range 34–28 (uncalibrated) kyr BP; the interstadial is correlated with GIS 8-7 (Mangerud et al. 2010). Sub-till, fine-grained laminated deposits at altitudes of ca. 600 m.a.s.l. in Folldal (southern, central Norway) are interpreted to have been deposited in an interstadial period defined as the Gråmobekken Interstadial (Thoresen and Bergersen 1983). Pollen in the mostly glacio-lacustrine sediment consists of 30 to over 50% of *Betula* pollen; NAP is dominated by Poaceae and includes up to some 5% of *Artemisia* pollen only. According to the authors, this assemblage indicates a scattered tree vegetation.  $^{14}\text{C}$  dating on organic matter from the minerogenic sediment gave an age of ca. 32.5 (uncalibrated) kyr BP (Thoresen and Bergersen 1983). The sediments at one of the type sections for the Gråmobekken Interstadial, i.e. at Djupdalsbekken (Figure 1-1), have recently been re-dated. The dated material consisted of macrofossil remains of *Betula nana*, *Salix*, *Empetrum*, *Silene dioica* and *Selaginella* and gave an age of ca. 31 (uncalibrated) kyr BP (Paus et al. 2011). According to Thoresen and Bergersen (1983), the Gråmobekken Interstadial sediments indicate that central parts of southern Norway were ice-free during at least part of the Middle Weichselian (MIS 3). This is now supported by the dating on terrestrial plant macrofossils by Paus et al. (2011).

In Figure 1-1B, the ice-extent for early MIS 3 corresponds to the maximum extent of ice-cover over northern Fennoscandia during an interstadial dated to ca. 53 kyr ago (Helmens and Engels 2010). This is based on comparisons of the Sokli stratigraphy with OSL-dated sediment sequences in eastern Finland (Ruunaa in Figure 1-1; Lunkka et al. 2008) and western Finland (Hitura in Figure 1-1; Salonen et al. 2008), the OSL-dated sediments at Pilgrimstad (Alexanderson et al. 2009), and the radiocarbon chronology of Olsen et al. (2001). The re-assessed radiocarbon chronology in Wohlfarth (2010) and the preliminary OSL dating result at Riipiharju (Alexanderson et al. 2011), however, suggest that also northern Sweden was ice-free during early MIS 3 and glaciers might have been restricted to the Scandinavian mountain range only. Furthermore, the recently obtained data from northern Finland, Sweden and Norway discussed above indicate that ice-free conditions with *Betula*-dominated vegetation persisted, possibly interrupted by glaciation, into the middle part of MIS 3 to ca. 35 kyr BP. This means that the early MIS 3 Tulppio Interstadial sediments in the long Sokli sequence were probably truncated during the last glaciation (Figure 4-1).

### 3.5 Lake Yamozero (Russia)

A 22 m long sediment core from Lake Yamozero on the Timan ridge in north-western Russia (lat. 65°01'N, long. 50°14'E; elevation 213 m.a.s.l.; Figure 1-1) has been analysed in detail for pollen and sediment characteristics, and has been subjected to a detailed dating program including OSL dating on quartz using SAR dose protocol and AMS  $^{14}\text{C}$  dating on mostly terrestrial plant material, by Henriksen et al. (2008). The lake basin (ca. 30 km<sup>2</sup>) occupies a closed depression above a heavily fractured part of the Timan ridge. The latter forms a low bedrock range (maximum elevation at 470 m) between the Mezen and Pechora river basins. According to Svendsen et al. (2004), the Yamozero basin is situated outside all Weichselian (MIS 5d-2) glacial limits (Figure 1-1). The lake is surrounded by mires and open boreal forest dominated by *Picea*.

Reconstruction of the climate based on the pollen sequence shows two rather similar transitions from cold, glacial to warm, interglacial conditions in the lower and uppermost (Late Glacial – Holocene) part of the core (Figure 3-6). In the middle part (dated to MIS 4 and early MIS 3), generally cold



conditions prevailed with local steppe or shrub-bush tundra vegetation; two mild intervals are recognized with dwarf shrub tundra and *Picea*. Two unconformities are identified in the Yamozero sequence, i.e. just before ca. 70 kyr BP and after ca. 40 kyr BP. The latter are interpreted as periods of subaerial exposure with wind erosion implying very dry conditions. The lake basin started to fill up again at ca. 18 kyr BP. Two conflicting age models are presented for the ‘warmer-than-present’ unit near the base of the record. The ‘conservative’ model (alternative 1) places the unit in the early part of the Eemian (MIS 5e). Alternative 2, which uses the scattered OSL dates, suggests an early Odderade age (MIS 5a). The description and interpretations given below follow Henriksen et al. (2008).

### 3.5.1 Environmental reconstructions for MIS 5

Organic silts and gyttja near the base of the Yamozero sequence are interpreted to be deposited in a lake that gradually became shallower and more productive (Figure 3-6). The increasing representations for *Betula* and *Juniperus* as well as for *Urtica* (indicating local fertile soils) show the development of shrub tundra or possibly sparse tree *Betula* vegetation. The vegetation development is sharply interrupted by a rise in steppe plant taxa (*Artemisia* and Chenopodiaceae) and in pollen of e.g. the tundra species *Dryas*. Using the first age alternative, this lower unit would be of late Saalian age (MIS 6) and the cooling may correlate with the Zeifen-Kattegat climate oscillation (e.g. Kukla 2002b), as also recorded on the Central Russian Plain (e.g. Zelikson 1995). According to age alternative 2, which is favored by Henriksen et al. (2008) because it takes into account the OSL chronology, these sediments are of MIS 5b age (Rederstall Stadial; Figure 4-1).

The lower organic unit is capped by a compact silty clay bed. The clayey sediment presumably accumulated in relatively deep water, contemporaneous with the highest shoreline at 15 m above the present lake. Maxima in pollen percentage values for shrubs (e.g. *Juniperus*) in the lower part of the clay bed are followed by peak values for *Picea* (and *Betula*) and the representation of temperate tree and shrub taxa including *Quercus*, *Ulmus* and *Corylus*. It is suggested that stands of *Ulmus* may have been present on drier and more fertile soil, indicating summer temperatures warmer than today. The clay bed is correlated with the early parts of the Odderade Interstadial (MIS 5a; alt. 2) or of the Eemian Interglacial (MIS 5e; alt. 1).

### 3.5.2 Environmental reconstructions for MIS 4-1

The middle part of the Yamozero record includes a 13 m thick sequence of laminated lacustrine silts deposited under fluctuating ‘cold’ climates of the Middle Weichselian. Linear regression of 10 OSL dates, ranging in age from 32 to 82 kyr BP, and partly in reverse order, date this part of the record to the time interval ca. 72–42 kyr BP (MIS 4 to early MIS 3). *Artemisia* and Chenopodiaceae are registered by relatively high values throughout the silts (Figure 3-6).

LOI is slightly higher (up to ca. 10%) in the lower ca. 8 m of the silt sequence where AP reaches highest values. Around 60 kyr BP, pollen of *Papaver* (indicative of unstable soils) as well as reworked microfossils are absent. *Betula* pollen percentage and concentration values, and pollen of *Juniperus* and Ericales, reach maxima. Additionally, sparse finds of *Picea* stomata occur. The vegetation during this interstadial is interpreted as dwarf shrub tundra close to *Picea* forest-tundra (alternatively, forest tundra or open *Betula* forest). Furthermore, the lowermost part of the silt sequence shows an interval with relatively mild conditions. This mild period follows peak values in *Artemisia*. The warmer phases at Yamozero may correlate with the deglacial phase around 65–60 kyr BP (Mezen transgression), with marine molluscs suggesting that shallow water temperatures in the White Sea were only slightly cooler than today (Jenssen et al. 2006).

Pollen assemblages in the upper part of the silt sequence (after ca. 55 till ca. 40 kyr BP), as well as in sandy silts dated to around 16 kyr BP (late MIS 2), mostly represent local steppe or desert vegetation. The Late Glacial part of the record is similar to the vegetation development registered in the organic silts near the base of the Lake Yamozero sequence. The cooling is dated to the Younger Dryas cooling event (ca. 12–11 kyr BP). The Holocene part of the Yamozero record is represented by gyttja and shows rapid changes in the dominance of *Betula*, *Picea* and eventually *Pinus*. Similar as to the interglacial clay bed near the base of the record, *Artemisia* and Chenopodiaceae show very low percentage values only.

## 4 Discussion

### 4.1 Limitations of proxy records and the value of multi-proxy comparisons

#### 4.1.1 Pollen records

Long pollen records from Europe have provided invaluable information on changes in vegetation and climate during the Late Pleistocene. As ice built up and ice sheets spread over northern Europe under a cooling climate, vegetation belts shifted southward over distances of up to thousands of kilometers, sensitively recorded by plant microfossils. Because of the magnitude of these displacements, however, the response of vegetation to a warming climate can be expected to have been delayed to various degrees. Vegetation development in Europe during the Late Pleistocene strongly depended on the severity of climate conditions directly preceding a warming trend as well as the magnitude and duration of the warming event. These factors determined the extent of glaciation and rate of deglaciation of a site as well as the migration of plants from glacial plant refugia.

Previously it was thought that continental ice sheets need to have very long response times under changing climates. The classic glaciation maps for northern Europe depict gradual growth in the Fennoscandian Ice Sheet (FIS) over several tens of thousands of years, starting in MIS 4 at around 65 kyr BP and culminating in the LGM at ca. 20 kyr BP (Mangerud 1991, Lundqvist 1992, Donner 1995, Kleman et al. 1997). Vegetation belts are drawn parallel to an ice-front retreating in north-western directions, towards the Scandinavian mountains, maintaining this direction also during the warmer stages MIS 5c and 5a. Vegetation response was considered slow with trees migrating from plant refugia situated south of the major mountain ranges in southern Europe, i.e. on the Balkan, Italian and Iberian Peninsulas (e.g. Willis and Niklas 2004; Figure 1-1).

Recent data, however, is revealing a dynamic ice sheet behavior including both fast build-up and decay. Extensive ice-free areas over central Fennoscandia are reconstructed for late MIS 3 prior to ca. 30 kyr BP based on <sup>14</sup>C-dated mammoth remains from Finland and Sweden (Ukkonen et al. 1999, 2007); ice subsequently expanded to its maximum LGM position on the northern Russian Plain (Figure 1-1B) within a timeframe of some 10 kyr only (Lunkka et al. 2001). Rapid ice sheet growth is supported by marine data indicating a fall in sea-level by some 30–40 m at the onset of the LGM (Lambeck et al. 2002). Extensive deglaciation in the eastern, continental sector of the FIS is recorded for early MIS 3 around ca. 50 kyr BP that allowed the establishment of *Betula* tree-line vegetation in northern Finland (Helmens et al. 2007b, Bos et al. 2009, Helmens and Engels 2010). Plant response was relatively fast. Trees probably migrated from cryptic glacial plant refugia situated to the southeast or east, on the large plains of the Baltic States or northern Russia that remained mostly unglaciated during MIS 4 (Bos et al. 2009; Figure 1-1B). The existence of tree populations in northerly small pockets of environmentally favorable conditions, in some cases close to the edge of the LGM ice-caps, is indicated by mounting evidence from pollen, plant macrofossils and macrofossil charcoal assemblages; trees in these refugia mostly involved those currently found in the boreal forest including *Pinus*, *Picea*, *Larix*, *Betula* and *Salix* (Willis and van Andel 2004, Paus et al. 2011, Väiliranta et al. 2011, Parducci et al. 2012).

#### 4.1.2 Insect and macrofossil records

The fastest response to climate warming can be expected to be recorded by rapid colonizers such as insects (e.g. Coleoptera, chironomids) or aquatic/wetland plants which do not depend on soil formation. These proxies might provide temporarily better climate estimates than terrestrial pollen assemblages particularly in the case of short-lasting climate events. It should be stressed, though, that insect records as well as aquatic/wetland indicator plant species identified in the macrofossil record often only provide estimates on July temperatures. The summer temperatures are commonly used to infer a 'warm' or 'cold' climate. A reconstruction of climate regime (or degree of continentality), however, can not be made without knowledge of winter temperatures. To provide an example, brief summers with a mean July temperature of 13°C in sub-continental cold northwestern Russia are currently accompanied by mean January temperatures exceeding –20°C and zonal vegetation is sub-

arctic forest-tundra (Figure 1-1). 13°C July temperatures in Fennoscandia, on the other hand, e.g. in northern Finland and along the western Norwegian coast, are linked with mean January temperatures of ca. -10°C and 0°C, respectively, resulting in northern boreal forest or even temperate vegetation under relatively mild maritime conditions. Valuable climate data, however, can be obtained when the insect and aquatic/wetland plant macrofossil data is combined with pollen evidence. Plant indicators for strong continental conditions include the steppic herbs *Artemisia* and Chenopodiaceae, and also Poaceae, although a main constituent of the northern Eurasian tundra, primarily attain high pollen percentage values in the steppes (Tarasov et al. 1998). The concept of continentality becomes even more important when considering that climate variability during the Late Pleistocene was probably for a major part manifested through changes in winter temperatures and precipitation (e.g. Klotz et al. 2004, Denton et al. 2005, Kelley et al. 2008, Helmens et al. 2012; this report).

It should further be noted that quantitative reconstructions of climate parameters from insects, macrofossils or pollen are based on present-day plant/insect-climate relationships. These relationships may not necessarily have been the same in the past. Problems might arise in the case of non-analogue vegetations/insect assemblages or climates. Also human-influence on present-day plant cover may influence the climate reconstruction.

Another point to be raised in the context of the present report is the findings of macrofossils in the sediment record. This to great extent depends on taphonomy with, for instance, distance to shore or inflow of running water close to the coring-site greatly influencing the amount of macrofossil remains in the lake sediment record (e.g. Hannon and Gaillard 1997, Birks 2003, Väiliranta 2006, Helmens et al. 2009, Shala et al. 2014). This means that the absence of for instance seeds of tree *Betula* does not give conclusive evidence for the absence of this tree in the surrounding terrestrial vegetation or wetland.

#### 4.1.3 The sediment record

In addition to pollen, plant macrofossils and insect remains, the present compilation focuses on the sediment record at the coring-sites. Most obvious are till beds that record former glaciations and indicate initial cool and relatively wet climate conditions. Pollen analysis on marine cores, for instance, shows that ice build-up in early MIS 5d took place during the late Eemian (Turon 1984, Sánchez-Goñi et al. 1999, Shackleton et al. 2003) in a cool but still relatively wet climate (Guiot et al. 1989, 1992). Too dry climate conditions inhibit ice sheet initiation. Lake level fluctuations depicted in the sediment record can also provide information on past changes in temperature and precipitation. However, a fall in lake level might also be the results of a local drying effect related to forest establishment (see e.g. the Oerel record under 3.2.1.; Figure 3-3), through increased evapotranspiration and decreased surface-runoff, or a combination of warming (increasing evaporation) and forestation. The latter could explain the distinct lithology changes at Horoszki Duże (Figure 3-4).

The identification of peat beds in the sedimentary record is important, as well as other evidence (e.g. macrofossils), pointing to the former presence of wetland at or nearby the coring-site. Pollen from local wetland habitats are generally over-represented in the pollen record and can obscure the regional vegetation/climate signal (e.g. Salonen et al. 2013). For this reason, sedges (Cyperaceae) that attain relatively high percentage values through major part of the Lake Yamozero record (Henriksen et al. 2008), and the grass (Poaceae) *Phragmites* at Horoszki Duże (Granoszewski 2003), have been excluded from the pollen sums in Figures 3-4 and 3-6. Botanical evidence also points to wetland expansion during various phases of lake infilling at Sokli (Figure 3-5), particularly increasing Cyperaceae pollen values (e.g. Helmens et al. 2000, 2012). Since sedges are an important tundra as well as wetland element, their pollen have not been excluded from the sum in the Sokli record; instead, the phases with wetland expansion are highlighted along the sediment column in Figure 3-5 and are taken into consideration in the interpretation of the regional vegetation and climate (Bos et al. 2009, Helmens et al. 2012, Salonen et al. 2013). Local peat accumulation is further distinctly recorded at Oerel (Behre et al. 2005). However, pollen of heath (Ericales) that attain high percentages during phases of peat deposition at Oerel (see e.g. *Sphagnum* curve in Figure 3-3) are not excluded from the sum (following Behre and Lade 1986, and Behre 1989). Peat accumulation during e.g. the Oerel and Glinde Interstadials might therefore hamper a holistic view of the regional vegetation composition in northern Germany during these early MIS 3 interstadials.

Furthermore, information on sedimentation rates is crucial for estimates on timing and duration of climate events recognized in the generally poorly dated terrestrial sequences. Analysis by Poneel (1995), for instance, reveals that running-water Coleoptera are predominant in the forested parts of the La Grande Pile sequence, in contrast to mostly standing-water beetles which characterise the cold intervals. Poneel (1995) concludes that rates of sediment accumulation were most probably not constant throughout the La Grande Pile sediment sequence. This warns us for the danger of aligning terrestrial records by assuming constant sedimentation rates, even those records obtained from relatively large basins.

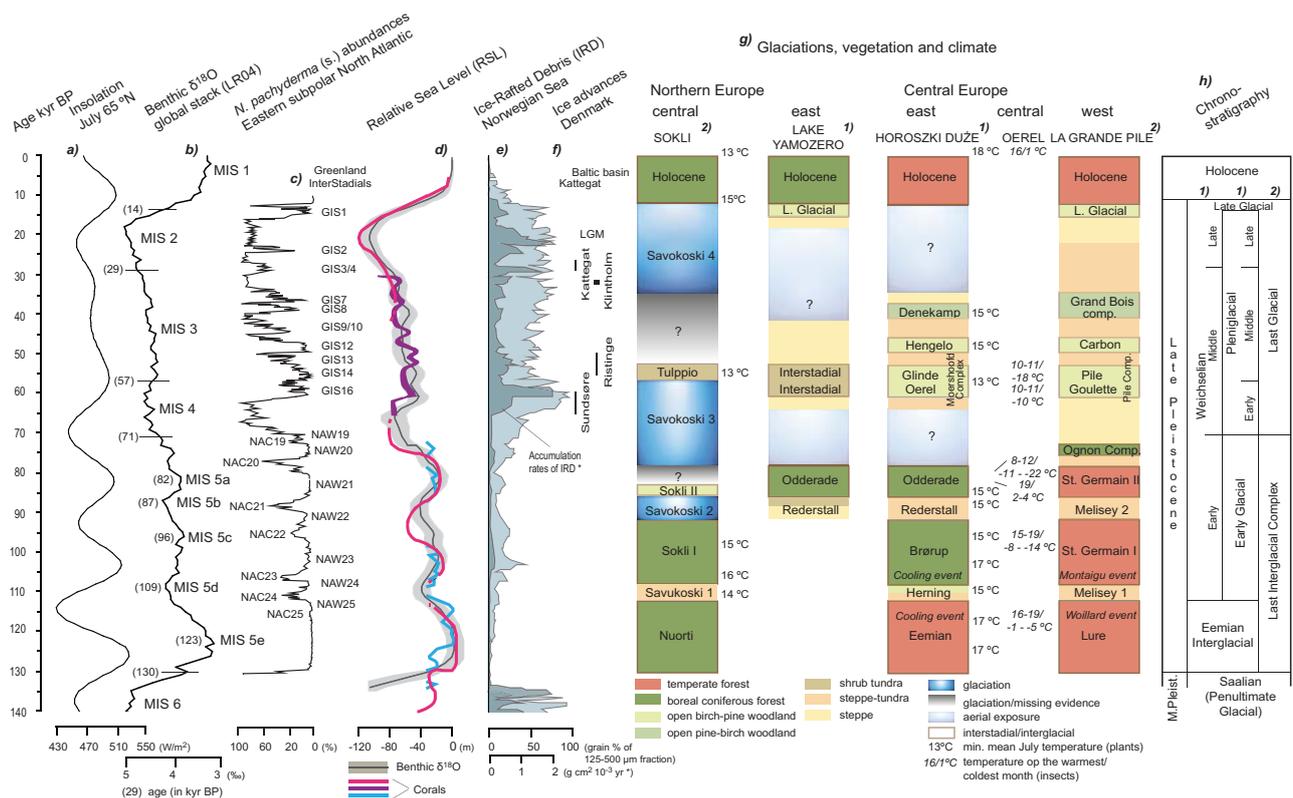
#### 4.1.4 Comparison with marine records

It is well-acknowledged that the most valuable climate reconstructions are obtained through multi-proxy comparisons. Multi-proxy studies reduce the limitations and uncertainties related to single proxies and facilitate discrimination between local (catchment-scale) and regional (climate) processes driving the proxy records. In the final part of this compilation study, the central and northern European terrestrial data will be further compared with the rapidly accumulating proxy evidence from the ocean realm including the eastern sub-polar North Atlantic Ocean (Figure 4-1). The latter include a composite record of percentage abundances of the polar planktonic foraminifera *Neogloboquadrina pachyderma* sinistral (Figure 4-1c), which is a proxy for sea-surface temperatures (SST; Bond et al. 1993, McManus et al. 1994), and a stacked ice-rafted debris (IRD) record from the Vøring Plateau in the Norwegian Sea (Figures 1-1 and 4-1e), which measures ice advances onto the continental shelf (adapted from Baumann et al. 1995). Although IRD is not a simple proxy for ice sheet size, the IRD curve in Figure 4-1e is in broad agreement with the Late Pleistocene glaciation curve from the south-west flank of the FIS (Mangerud et al. 2011). Information on fluctuations of the FIS is further given in Figure 4-1f (showing OSL-dated till sequences in Denmark according to Larsen et al. (2009a, b) and Houmark-Nielsen (2010)) and Figure 1-1B that compiles data for the western portion of the Eurasian IceSheet (including the FIS) based on Svendsen et al. (2004), Carr et al. (2006), Helmens and Engels (2010) and Mangerud et al. (2011).

Figure 4-1 also gives curves for relative sea level (RSL) changes for the Late Pleistocene (Figure 4-1d), which are inversely related to global ice volume changes, based on high resolution, benthic foraminifera oxygen isotope ratios (Waelbroeck et al. 2002) and coral-based sea-level reconstructions on Huon Peninsula, Papua New Guinea and Barbados (Chappell 2002, Cutler et al. 2003, Thompson and Goldstein 2005). The composite RSL record of Waelbroeck et al. (2002) is mostly based on one Equatorial Pacific Ocean benthic foraminifera record which seems to be largely driven by ice volume and less by variations in deep water temperatures. The curve by Cutler et al. (2003) includes fossil corals that also grew during relatively low sea-level stands; it is considered an initial step in establishing a continuous, high-resolution sea-level curve for the last climate cycle based on U/Th-dated corals. The study by Chappell (2002) includes reconstructions of sea level rises, based on shallow-water reef thickness, in an attempt to provide a preliminary millennial-scale RSL record for MIS 3. Thompson and Goldstein (2005) improved the resolution for MIS 5 using open-system age equations for relative highstands inferred from raised corals.

Figure 4-1 further shows the LR04 (Lisiecki and Raymo 2005) global benthic stack (which measures global ice volume and deep ocean temperature) with marine climate-stratigraphic stages (MIS; Figure 4-1b). North Atlantic cold (NAC) and warm water events (WAC) during MIS 5 are additionally indicated along the high-resolution SST record in Figure 4-1c (McManus et al. 1994), whereas the warm events of millennial duration during MIS 3-1 (-1) are indicated in Figure 4-1c according to their denomination in the high-resolution Greenland ice core record (Greenland InterStadials (GIS); Johnsen et al. 1992, Dansgaard et al. 1993). To the left in Figure 4-1 (a), variations in summer insolation at high northern latitudes are shown (Berger and Loutre 1991), which is a major driver in orbital-scale climate changes.

Different dating methods have been applied on different materials for the records shown in Figure 4-1, and different tuning/correlation approaches (see figure caption). Therefore, the figure does not allow detailed comparisons but is aimed to explore the general patterns of environmental change and climate evolution over the Late Pleistocene. Additionally, the figure does not take into account the various degrees of diachroneity between terrestrial stratigraphic units and marine isotope stages.



**Figure 4-1.** Compilation figure showing comparisons between marine- and terrestrial-based (central and northern European) proxy records for the last 140 kyr. Marine and ice-core chrono-stratigraphic units are given to the left and European chrono-stratigraphy (with two alternative subdivisions) is to the right. The marine and ice-core stratigraphies include Marine Isotope Stages (MIS in (b)), North Atlantic Cold (NAC) and North Atlantic Warm events (NAW in (c); McManus et al. 1994), and Greenland InterStadials (GIS in (c); Johnsen et al. 1992, Dansgaard et al. 1993). Also shown to the left are variations in July insolation at 65°N latitude (Berger and Loutre 1991). Note that the figure does not take into account possible diachroneities between terrestrial stratigraphic units and marine isotope stages; furthermore, different dating methods and tuning approaches applied to the records hamper a detailed, high-resolution time comparison.

The time-scale for the global benthic stack in (b) is based on orbital tuning and accumulation rates (Lisiecki and Raymo 2005). The composite curve in (c) combines data from North Atlantic cores V29-191 (prior to 60 kyr BP; McManus et al. 1994) and V24-11 (after 60 kyr BP; Bond et al. 1993; for locations see Figure 1-1). The time-scale of core V29-191 is tuned to the MD95-2042 chronology (Shackleton et al. 2002) as in Rousseau et al. (2006). It was slightly stretched in the time-interval 130–75 kyr BP to obtain a better match with the NorthGRIP time-scale (NGRIP Members 2004). Minor stretching was additionally required in the original time-scale of McManus et al. (1994) between 75–60 kyr BP to obtain a match with core V24-11. The original radiocarbon dates of core V24-11 are converted to calendar ages as in Lambeck et al. (2002). The curves of RSL changes in (d) are according to Waelbroeck et al. (2002; grey line with associated confidence interval), Chappell (2002; pink), Cutler et al. (2003; purple) and Thompson and Goldstein (2005; blue). The age model for the composite RSL record of Waelbroeck et al. (2002) is based on correlation with the orbitally-tuned benthic stack of Martinson et al. (1987; SPECMAP) and calibrated radiocarbon ages. The coral-based RSL reconstructions are dated by U/Th. Time control for the stacked IRD record in (e) is based on correlation of the oxygen-isotope stratigraphy of *N. pachyderma* (s) in the cores with the SPECMAP reference series of Martinson et al. (1987) and on radiocarbon dates (Baumann et al. 1995). The record was slightly stretched in (e) for the IRD maxima to broadly correlate to the isotope maxima of MIS 2, 4 and 6. Note that the minor but significant IRD maximum during MIS 5 was originally dated to MIS 5b (around ca. 90 kyr BP; Baumann et al. 1995). (f) shows OSL-dated till sequences in Denmark according to Larsen et al. (2009a, b) and Houmark-Nielsen (2010). References to the data in (g) are given in the text. Note that the vegetation types high-lighted in the columns correspond to the inferred 'climax' vegetation during the stratigraphic interval in consideration. The chrono-stratigraphies of the Lake Yamozero and Horoszki Duże records follow alternative 2 favored in Henriksen et al. (2008) and alternative A of Granoszewski (2003), respectively (see Figures 3-4 and 3-6).

## 4.2 Late Pleistocene environmental history and climate evolution of central and northern Europe

The environmental and climate data for the Late Pleistocene inferred from the central and northern European terrestrial records is summarized and compared with the marine proxy data in Figure 4-1. The terrestrial and marine data reveals an overall increasing deterioration of the climate, accompanied with increasing continentality, over the Last Glacial-Interglacial (LI-G) cycle, starting at ca. 120 kyr BP in late MIS 5e (or mid-Eemian Interglacial) and culminating at around ca. 20 kyr BP in late MIS 2 (LGM). A similar conclusion has been drawn by Caspers and Freund (2001) based on a compilation of Weichselian (MIS 5d-2) data for northern central Europe including permafrost features. Superimposed on this cooling trend, however, the present review shows that environmental and climate conditions were distinctly different during MIS 5 compared to MIS 4-2.

### 4.2.1 MIS 5 (ca. 130–70 kyr BP)

MIS 5 is characterized by three major forested intervals, roughly corresponding to MIS 5e, 5c and 5a, both in central and northern Europe. Mildest conditions occurred during the Eemian Interglacial (MIS 5e; Figure 4-1).

At La Grande Pile, in temperate west-central Europe (Figure 1-1), mixed deciduous vegetation with e.g. *Ulmus*, *Quercus*, *Corylus* and *Carpinus* developed during MIS 5e, 5c and 5a (Figures 3-1 and 4-1). Pollen and Coleoptera data shows that the early part of the Eemian (locally named Lure; MIS 5e) was the most temperate interval with winter temperatures exceeding present-day values by several degrees. The late part of the Lure, and St. Germain I and II (MIS 5c and 5a, respectively), had more continental climate regimes (Guiot et al. 1992, 1993, Cheddadi et al. 1998). *Picea* is overall well-represented in the pollen records for St. Germain I and II, pollen of the thermophilous plant taxa *Hedera* is rare, and *Larix* appears during the closing phase of both periods. Summer temperatures averaged at ca. 16–17°C (compared to current mean July temperature values up to ca. 18°C) and winter temperatures at ca. –3 to –10°C (compared to January values presently down to ca. –2°C) in the northern Alpine foreland during St. Germain I and II (Klotz et al. 2004; Figure 3-2A).

The Eemian vegetational succession shows strong uniformity across the temperate zone of central Europe (Menke and Tynni 1984, Zagwijn 1996, Turner 2002b) and Eemian forest dynamics similar as at La Grande Pile are recorded at Oerel and Horoszki Duże (Figure 1-1). In contrast to the temperate forest developments at La Grande Pile during St. Germain I and II, however, the Brørup (MIS 5c) and Odderade (MIS 5a) at Oerel (Figure 3-3) and Horoszki Duże (Figure 3-4) were characterized by boreal forest with *Betula*, *Pinus*, *Picea* and *Larix* (Figure 4-1). Indicator plant species in the Horoszki Duże sequence provide minimum mean July temperatures of 17°C, 15–17°C and 15°C for the Eemian, Brørup and Odderade, respectively (Figure 4-1; compared to a current value at ca. 18°C). Coleoptera assemblages in the Oerel sequence indicate temperatures of the warmest month at ca. 16–19°C, 15–19°C and 8–12°C during the late Eemian, Brørup and Odderade, respectively (present-day mean July value is at ca. 16°C). Temperatures of the coldest month were significantly lower than the current mean January temperature of ca. 1°C, i.e. ca. –8 to –14°C and –11 to –22°C during the Brørup and Odderade, respectively, and –1 to –5°C for the late Eemian (Figure 4-1). At Gröbern (Figure 1-1), mean July temperatures to ca. 19°C (compared with ca. 17°C at present) are reconstructed for the Eemian based on pollen and plant macrofossils, and mean July and January temperatures of ca. 17°C and around ca. –12°C, respectively, for the Brørup and Odderade (mean January temperatures are currently at ca. –2°C); winter temperatures during most of the Eemian were ca. –3°C (Kühl et al. 2007; Figure 3-2B).

The data summarized above for central Europe indicates an enhanced degree of continentality during MIS 5c and 5a compared to MIS 5e and the Holocene. It is probable that the absence of temperate forest in major part of central Europe during MIS 5c and 5a was due to factors related to increased continentality including low winter temperatures, low precipitation values, thin snow cover, frost, and a shorter growing season. As noted by Behre (1989) and Kühl et al. (2007), summer temperatures were high enough, and the forested phases of the Brørup and Odderade most probably lasted long enough, for the re-immigration of temperate trees. Casper and Freud (2001) note a difference in forest development along a W-E transect over central Europe in the Brørup, with an eastwards decrease in pollen percentage values for *Quercus*, and ascribe this differentiation in vegetation to

a combination of winter temperatures and precipitation values. Salonen et al. (2012) show that winter temperatures presently have a relatively large influence on deciduous *Quercus* as well as *Corylus*. They base this conclusion by analyzing the relative influence of different climatic parameters on the representation of plant taxa in modern pollen samples from northern Europe. According to Seppä et al. (2004), growing season length and frost sensitivity are major factors determining the current northern distribution limits of most temperate trees in northern Europe. Data from the northern Alpine lowland also show a decreased representation of thermophilous plant at more continental eastern sites during St. Germain I and II (Klotz et al. 2004). The paleo-botanical data of Granoszewski (2003), showing boreal forest with *Larix* and significant representations of Poaceae and *Artemisia*, points to relatively strong continental conditions in eastern Europe during the late Brørup and Odderade.

At Sokli in northern boreal Europe (Figure 1-1), forest development during the Nuortti (MIS 5e) and Sokli I (MIS 5c) was similar to the vegetation development in the Holocene (Figures 3-5 and 4-1). It included the establishment of pioneer *Betula* forest followed by the spread of *Pinus* and *Picea*. A somewhat more continental climate regime, with mean July temperatures exceeding the present-day value of 13°C by several degrees, is reconstructed for Sokli I compared to today (Helmens et al. 2012, Salonen et al. 2013). *Alnus glutinosa*, which current northern distribution limit is situated some 400 km south of Sokli, was an important element in the boreal forest during Sokli I. The studied Sokli II deposit (MIS 5a) has been truncated and only shows the early *Betula* zone. A detailed analysis on the Nuortti deposit (MIS 5e) in the Sokli sequence is in progress. The MIS 5 chronology of the interglacial bed near the base of the Lake Yamozero sequence in north-western Russia (Figure 1-1) is unclear. It shows a period with high pollen percentages for *Picea* and relatively high contributions for several thermophilous taxa (*Quercus*, *Ulmus* and *Corylus*; Figure 3-6); Henriksen et al. (2008) favor a MIS 5a age over a MIS 5e age assignment based on OSL dating (Figure 4-1).

The composition of the boreal forest as well as summer temperatures seem to have been remarkably similar over the northern part of the temperate zone and the boreal zone of Europe during MIS 5c and suggest relatively weak north-south gradients. The reasons for the weaker gradients over northern Europe during MIS 5c compared to present-day are most probably contemporary orbital forcing, with several feedback mechanisms amplifying the primary effects of orbitally induced insolation, as discussed by Välranta et al. (2009) and Engels et al. (2010). Earlier, stronger than present-day climate gradients over northern Europe during the Brørup (MIS 5c) and Odderade (MIS 5a) have been suggested based on correlations between long proxy records from central Europe with fragmented stratigraphic data from Fennoscandia (Behre 1989, Guiter et al. 2003).

In Mediterranean southern Europe, the Eemian (MIS 5e) forested interval has been dated in marine cores off the coast of Iberia to the time-interval 126 and 110 kyr BP (Sánchez-Goñi et al. 1999). The length of the Eemian Interglacial has been estimated to ca. 17.5 kyr based on laminated sediments at Lago Grande di Monticchio in southern Italy (Brauer et al. 2007; Figure 1-1). A duration of no more than 13 kyr for the Eemian forest phase is indicated by annually laminated diatomite sections in Germany (e.g. Turner 2002a). The Woillard event identified at La Grande Pile in the late Eemian (Figure 3-1) seems also to be recorded at Horoszki Duże (Figure 3-4); the event is correlated with NAC 25 (e.g. Rousseau et al. 2006). This would indicate that the Eemian in temperate Europe lasted at least until ca. 115 kyr BP (Figure 4-1). The latter is supported by U/Th dates of ca. 115 kyr BP on peat deposited in the northern Alpine foreland during the period of cooling towards the end of the Eemian (Preusser 2004). The duration of the Brørup (MIS 5c) and Odderade (MIS 5a) are estimated at 5–10 kyr based on annually laminated diatomites and peat accumulation rates (Behre and van der Plicht, 1992, and references therein).

Relative mild climate conditions prevailed in the eastern subpolar surface North Atlantic Ocean for most of MIS 5 with peak warmth being recorded for MIS 5e (McManus et al. 1994; Figure 4-1c). Warm season sea surface temperature (SST) estimates from planktonic foraminifera assemblages, based on the modern analog technique, reveal that interglacial warmth during MIS 5e lasted ca. 10 kyr and the climatic optimum ca. 2–3 kyr (Oppo et al. 2006). MIS 5c and MIS 5a are estimated ca. 6°C and 4°C colder, respectively, than early MIS 5e; an approximate cooling of 4°C is recorded between peak MIS 5e warmth and NAC 25 (Oppo et al. 2006).

Global sea-level peaked 5.5 to 9 meters above present sea level during MIS 5e (Dutton and Lambeck 2012). Globally averaged-sea level is estimated to ca. 20 m lower for MIS 5c and MIS 5a than at present (Figure 4-1d). Speleothem encrustations from coastal caves on the island of Mallorca in

Mediterranean Europe, however, indicate a sea-level highstand that was slightly higher than at present, and only slightly lower than the MIS 5e sea level, at 81 kyr BP during late MIS 5a (Dorale et al. 2010); the data from Mallorca is consistent with a number of RSL estimates from tectonically stable locations around the world for MIS 5a. According to Dorale et al. (2010), MIS 5a was probably at least as ice-free as the present. It should be kept in mind that all proxies have uncertainties and are estimations only. Benthic foraminifera oxygen isotope ratios carry a composite signal of global ice volume and deep ocean temperatures, whereas assumptions about the magnitude and constancy of uplift rates is a source of uncertainty in the coral-based sea level estimates from tectonically uplifting coastlines as shown in Figure 4-1d. Interestingly, mild climate conditions with both summer and winter temperatures above present-day values are recorded at Oerel during the early part of the Odderade (MIS 5a) *Pinus* phase (Figure 4-1g). The quantitative reconstructions by Klotz et al. (2004) further suggest that the optima in both mean temperatures of the warmest and coldest month were somewhat higher during St. Germain II (MIS 5a) compared to St. Germain I (MIS 5c) in the northern Alpine foreland.

The MIS 5e, 5c, 5a forested intervals in temperate and boreal Europe were separated by cold MIS 5d and 5b defined as Melisey 1 and 2 in France, the Herning and Rederstall Stadials on the north European mainland, and the Savukoski 1 and 2 Stadials in northern Finland. During the stadials, open vegetation dominated with a significant contribution of steppe elements (e.g. Poaceae, *Artemisia*, Chenopodiaceae) indicating strong continental conditions (Figure 4-1). Winter temperatures in the range of ca. -15 to -20°C are reconstructed for central Europe (Klotz et al. 2004, Kühl et al. 2007). Summer temperatures, however, remained high enough for boreal trees and their local presence is confirmed by macrofossil remains both in eastern Poland (Granoszewski 2003) and northern Finland (Helmens et al. 2012). Minimum mean July temperatures of 15°C are reconstructed based on macrofossil evidence at Oerel (Figure 3-3) and Horoszki Duże (Figure 4-1), and mean July temperatures of at least 12–14°C are recorded by indicator plant species and chironomids at Sokli (Figures 3-5 and 4-1). Overall, summer temperatures during MIS 5d and 5b differed little from those reconstructed for the relatively warm MIS 5c and 5a (Figures 3-2 and 4-1). Precipitation values are difficult to reconstruct quantitatively but the pollen-based reconstructions by Klotz et al. (2004) and Kühl et al. (2007) suggest lowered precipitation values during MIS 5d and 5b compared to MIS 5c and 5a (Figure 3-2).

The records that are reviewed in this report indicate that the older parts of the MIS 5d and 5b stadial intervals had a more severe climate than the younger parts, and that overall MIS 5b had a less favorable climate than MIS 5d. The absence of any really cold-adapted Coleoptera (Ponel 1995), and the apparent persistence of nearby plant refugia (Pons et al. 1992), suggest that MIS 5d and 5b at La Grande Pile were not very arid nor cold compared to MIS 4 and 2 (see below). The FIS was mostly mountain-centered and only during MIS 5b advanced far enough eastwards for the Sokli site in northern Finland to become glaciated (Figures 1-1B and 4-1). Note that the minor but significant peak in IRD halfway through MIS 5 in Figure 4-1e was originally dated to MIS 5b (Baumann et al. 1995; see caption of Figure 4-1). This MIS 5b glaciation recorded in the IRD sequence from the Norwegian Sea compares with the MIS 5b Bøne glaciation recognized at Fjøsanger along the southwestern Norwegian coast (Mangerud et al. 1981, 2011). Combined pollen and isotope studies on North Atlantic marine cores indicate that the early MIS 5d glaciation took place in the late Eemian (Turon 1984, Sánchez-Goñi et al. 1999, Shackleton et al. 2003) in a cool but still relatively wet climate (Guiot et al. 1989, 1992). It is probable that ice accumulation during MIS 5b also initiated under still relatively moist conditions in the late part of the warm stage correlated to MIS 5c (i.e. St. Germain I/Brørup). Humid conditions at the end of the Eemian, with large-scale solifluction, are reconstructed at Stenberget in southern Sweden (Figure 1-1; Berglund and Lagerlund 1981). However, according to Mangerud et al. (1979), the late Eemian forests at Fjøsanger indicate a climate which would exclude the growth of any large ice-masses in Fennoscandia.

The relative mild climate conditions in the subpolar surface North Atlantic Ocean during MIS 5 were punctuated by a series of cool intervals (NAC 24-19 in Figure 4-1c). The summer SST record shows greatest cooling during NAC 24 and 21 (MIS 5d and 5b, respectively) of up to ca. 2°C (Oppo et al. 2006). The marine, as well as terrestrial records, show that millennial scale climate events were superimposed upon the longer lasting orbitally induced climate fluctuations of MIS 5. The short-lasting climate oscillations in the La Grande Pile record near the end of the Eemian (Woillard event) and in the early part of St. Germain I (Montaigu event) are correlated to NAC 25 and 23, respectively (Kukla et al. 1997, McManus et al. 2002, Rousseau et al. 2006; Figure 4-1). The

Montaigu cooling event shows the total disappearance of deciduous taxa accompanied by a rise in *Pinus* at La Grand Pile (Figure 3-1), a distinct return of open plant communities with e.g. *Artemisia* at Horoszki Duże (Figure 3-4), and decreased *Betula* pollen values at less sensitivity sites in northern Germany (Figure 3-3) and northern Finland (Figure 3-5); a distinct decrease in winter temperatures and precipitation values is reconstructed in the northern Alpine foreland for the Montaigu event (Figure 3-2A). The climate amelioration which resulted in the re-appearance of boreal forest both in France (Ognon Complex in Figure 4-1g), southern Germany (Müller et al. 2003) and Switzerland (Dürnten Interstadial; Welten 1982), following St. Germain II (MIS 5a), is correlated with NAW 20 (Müller et al. 2003).

#### **4.2.2 MIS 4-2 (ca. 70–15 kyr BP)**

MIS 4-2 included two glacial maxima i.e. during MIS 4 and 2. The coldest climate conditions occurred during late MIS 2 (LGM).

However, full glacial conditions in Europe and the North Atlantic Ocean were reached also during MIS 4. During this period, the FIS covered the Nordic countries of Norway (Mangerud et al. 2011), Sweden (Anjar 2012) and Finland (Saarnisto and Lunkka 2004), and in the west probably jointed the British Ice Sheet (Carr et al. 2006; Figure 1-1B). With glacier ice advancing onto the continental shelf a broad maximum in ice-rafted debris is recorded in the Norwegian Sea (Figure 4-1e). Glaciation onto the north European mainland is registered in Denmark (Figure 4-1f); the Sundsøre Till represents ice advancing through the Kattegat to the west whereas the Ristinge advance took place through the Baltic Basin in the east (Figure 1-1).

The Sundsøre and Ristinge glaciations are dated by OSL to MIS 4 and early MIS 3, respectively (Figure 4-1f; Larsen et al. 2009a, b, Houmark-Nielsen 2010). Time control for the IRD record of Figure 4-1e is based on correlation of oxygen-isotope stratigraphies in the cores with the SPECMAP reference series (Martinson et al. 1987) and on radiocarbon dating (Baumann et al. 1995). The record is a stack of 5 different marine cores. The IRD record has been slightly stretched in Figure 4-1e for the IRD maxima to broadly correlate to the isotope maxima and RSL minima of MIS 6, 4 and 2. In the original work of Baumann et al. (1995), the IRD maxima of MIS 4 was dated at around 55 kyr BP at the transition to MIS 3. Dating is a problem and hampers detailed age comparisons. MIS 3 is at the limit of the radiocarbon dating method and comparisons of  $^{14}\text{C}$  ages are further complicated by the fact that the method is used on a large variety of materials (e.g. bulk organic matter, macrofossils, bone, shells) collected from very different settings (e.g. ocean, lake). OSL is recently providing important age estimates for terrestrial records. However, the OSL dating method is not without problems and interpretations of results have not always been straightforward in Fennoscandia (e.g. Alexanderson and Murray 2007). Inconsistent OSL results (including both over- and underestimates of true ages) may be due to sampling site factors (e.g. the difficulty to correctly assess water content) as well as the luminescence properties of a sample (e.g. incomplete bleaching, fading of the OSL signal, weak signals); furthermore, error limits only include quantifiable contributions and do not include the contribution from e.g. any systematic overestimates arising from incomplete bleaching (Alexanderson and Murray 2012). Despite the problems with dating, however, the data from northern Europe and the oceans points to major glaciation around MIS 4.

Silty lacustrine sediments accumulated in the La Grande Pile depression during MIS 4 in which peak pollen percentage values for steppe plant taxa (e.g. *Artemisia*) point to strong continental conditions (Figure 4-1g). A distinct minimum in mean annual temperatures is reconstructed for MIS 4 at La Grande Pile (Guiot et al. 1989). In eastern Europe, at Lake Yamozero in northern Russia and probably also at Horoszki Duże in eastern Poland, subaerial exposure with wind erosion due to very dry conditions caused unconformities in the sedimentary sequences (Figure 4-1).

Environmental and climate conditions were even more severe during MIS 2 compared to MIS 4. Glaciations recorded in Denmark include the Kattegat advance from Norway via the Kattegat through and the Klintholm glaciation through the Baltic basin dated by OSL around the transition from MIS 3 to MIS 2 (Figure 4-1f; Larsen et al. 2009a, b, Houmark-Nielsen 2010). Denmark was overridden by the LGM ice advance from the north at ca. 23 kyr BP, and this was followed by a retreat after 21 kyr BP (Larsen et al. 2009a). The FIS advanced well onto the north-eastern European mainland during the LGM (Figure 1-1B). A broad maximum characterises the IRD record during

MIS 2 (Baumann et al. 1995) and globally-averaged sea-level at the LGM was at an absolute minimum, i.e. at approximately 120 m under today's level (Figure 4-1). In northeast Europe, very dry conditions caused an unconformity in the sedimentary record of Lake Yamozero (Figures 3-6 and 4-1), and it is most probably also dryness that limited growth in the north-eastern portion of the Eurasian Ice Sheet (Figure 1-1B). The lake basin at Lake Yamozero started to fill up again at ca. 18 kyr BP (Henriksen et al. 2008). MIS 2 at La Grande Pile shows continental steppe-tundra vegetation followed by a dominance of steppe plant communities during the LGM (Figure 4-1g). The correspondence between the polar foraminifera *N. pachyderma* abundances and SST weakens as percentages approach 100% (Figure 4-1c) and hence may correspond to a wide range of surface temperatures (e.g. Kohfeld et al. 1996).

The relative warm and humid MIS 3 period (57–29 kyr BP) was characterized by millennial-scale climate variability (Figure 4-1c–d). In contrast, climate variability of millennial duration was reduced during MIS 4 and 2 (Capron et al. 2010 and references therein). Temperature reconstructions for the warming events in early and middle MIS 3 (GIS 16 to 8) based on the Greenland ice core record vary between 8 to 15°C of abrupt warming followed by gradual cooling (Capron et al. 2010 and references therein). These temperature inferences, however, are biased by the origin and seasonality of the precipitation (Jouzel et al. 1997). Denton et al. (2005) note a mismatch between mean annual temperatures deduced from Greenland ice cores in comparison with snowline changes in e.g. East Greenland and ascribe the huge temperature declines during MIS 3 cold events largely to the winter season (related to extensive winter sea ice across the North Atlantic); summer temperature changes were relatively modest. The LGM, however, involved considerable summertime cooling as well (Denton et al. 2005). A recent study by Kelley et al. (2008) supports the conclusion by Denton et al. (2005) that the coolings were mainly a wintertime phenomena. Their study reconstructs equilibrium line altitudes from Late Glacial to early Holocene moraines on eastern Greenland, indicating a summertime-cooling of ca. 4–6.5°C compared to today's values and compared to a mean annual temperature depression of ca. 15°C inferred from the ice core data.

The magnitude of summer SST changes in the eastern subpolar surface North Atlantic during MIS 3 (GIS 15-9) is estimated at approximately 7°C, and temperatures were ca. 2–3°C lower during the MIS 3 warm events compared to MIS 5c and 5a (Oppo et al. 2006, Dickson et al. 2008). Interstadial warmth lasted relatively long in the early part of MIS 3 before ca. 45 kyr BP, whereas the middle part of MIS 3 (before ca. 35 kyr BP) shows stadials of increasing intensity punctuated by relatively brief interstadials (Figure 4-1c); late MIS 3 appears mostly cold. GIS 14 in early MIS 3 at around 53 kyr is the most prominent interstadial. Stable, warm conditions during GIS 14 lasted for about 2500 years and were followed by another 2000-yr period of intermediate temperatures before a very short cold interval started before GIS 13 (Huber et al. 2006). In a speleotherm record from the Austrian Alps, the maximum warmth during GIS 14 is dated to ca. 54.5–52 kyr BP (Spötl et al. 2006). A series of MIS 3 sea-level rises is reconstructed from corals by Chappell (2002; Figure 4-1d). Especially the longer and pronounced GIS 16/17, 14, 12 and 8 are associated with sea-level increases estimated at 15–25 m in the sea-level record from the Red Sea (Arz et al. 2007). Note that summer insolation at high latitudes was distinctly higher during GIS 16-12 than today (Figure 4-1a).

The climate evolution during MIS 3 described above is clearly registered in the central and northern European terrestrial data. Indicator plant species and chironomids from interstadial deposits reconstruct mean July temperatures similar or only slightly lower than those inferred for MIS 5c and 5a (Figure 4-1g and Engels et al. 2008). Estimates of temperatures from diatoms depict only modest differences in July temperatures of 0.5–2°C between interstadials and stadials in the time interval 36–18 kyr BP at Les Echets in France; this despite large shifts in lake productivity, catchment runoff and vegetation (Ampel et al. 2010). The inferred steppe-tundra vegetation in central Europe during the MIS 3 stadials (Figure 4-1) points to strong continental conditions and suggests relative warm, short summers and long, very cold winters.

A depression in winter temperatures by ca. 10 to 20°C is inferred at Oerel during the early MIS 3 interstadials based on Coleoptera data (Figure 4-1g). Vegetation at Sokli during the early MIS 3 Tulppio Interstadial, however, was remarkably similar to today's conditions in northern Fennoscandia (Bos et al. 2009) and suggests only modest wintertime temperature depression. The MIS 3 warming registered at Sokli, however, did not result in boreal forest. This is most probably

due to a delayed vegetation response to the relatively brief early MIS 3 warmings, which followed the glacial maximum of MIS 4. MIS 3 interstadials lasted long enough, and warming was intense enough, for the establishment of boreal trees in both central and northern Europe. Macrofossil and stomata evidence for the local presence of boreal trees is found at Horoszki Duże (Granoszewski 2003), in the Baltic countries (Satkunas et al. 2003), in northern Finland (Väliranta et al. 2012), and at Lake Yamozero in northern Russia (Henriksen et al. 2008).

When taking into account the uncertainties in dating for the MIS 3 time interval, it can be stated that the interstadials recognized in the proxy records of central and northern Europe broadly correlate in time. The early MIS 3 Goulotte and Pile Interstadials (or Pile Interstadial Complex) at La Grande Pile probably correspond to the Oerel and Glinde Interstadials at Oerel; the two interstadials recognized at Lake Yamozero; the Tulppio Interstadial at Sokli; and the Moershoofd Interstadial Complex in The Netherlands. The interstadials probably correlate to GIS 16 and 14 (Dansgaard et al. 1993; Figure 4-1). Open *Betula-Pinus* forest, with *Picea*, *Larix* and possibly *Quercus*, is reconstructed at La Grande Pile in France (Woillard 1978, de Beaulieu and Reille 1992); small stands of tree *Betula* with *Pine* and *Larix* occurred at Horoszki Duże in eastern Poland (Granoszewski 2003); and shrub tundra with *Picea* is inferred at Lake Yamozero in northern Russia (Henriksen et al. 2008). At Sokli in northern Finland, shrub tundra with the possible presence of tree *Betula* is recorded (Bos et al. 2009). The Oerel and Glinde Interstadials in northern Germany, however, are inferred as tree-less shrub tundra (Behre and Lade 1986, Behre 1989). It is possible that the high NAP values in the Oerel pollen record are due to an over-representation of pollen from the local wetland vegetation (Figure 3-3) and mask the signal of the regional vegetation here.

The  $^{14}\text{C}$  ages for the middle and late MIS 3 Hengelo and Denekamp Interstadials roughly correspond to those for the Charbon Interstadial and Grand Bois Interstadial Complex, and the age assignments for GIS 11/12 and 8, respectively (Figure 4-1; Dansgaard et al. 1993). The Charbon and Grand Bois Complex show *Betula-Pinus* and *Pinus-Betula* open forests, respectively, in France (Woillard 1978, de Beaulieu and Reille 1992). The presence of tree *Betula* with *Larix*, and of *Pinus*, is inferred in eastern Poland for the Hengelo and Denekamp, respectively (Granoszewski 2003). In contrast to the closed forests recorded for the warm stages of MIS 5, however, the MIS 3 vegetation was open. At La Grande Pile, for instance, the interstadial vegetation of MIS 3 is interpreted as increased expansion or blossoming of woodstands or shrubs in a still open vegetal environment (de Beaulieu and Reille 1992). Steppe communities or desert vegetation are reconstructed at Lake Yamozero in northern Russia for the middle part of MIS 3 (Henriksen et al. 2008) and, combined with the presence of open *Pinus* forest in central Europe (Figure 4-1), suggest less favorable (drier) climate conditions during the middle part of MIS 3 compared to early MIS 3.

The relatively mild climate conditions in early MIS 3 caused large-scale deglaciation of the FIS (Figure 1-1B; Helmens and Engels 2010). Preliminary OSL dating (Alexanderson et al. 2011) and a re-assessment of existing  $^{14}\text{C}$  dating results (Wohlfarth 2010) have recently suggested that also northern Sweden was ice-free during early MIS 3. The latter would restrict glaciation to the Scandinavian mountains as a result of early MIS 3 warming. The existence of extensive ice-free areas over Fennoscandia during the middle part of MIS 3 until ca. 35 kyr BP is indicated by  $^{14}\text{C}$  datings on e.g. macrofossil remains of terrestrial plants and mammoth remains, OSL and TL dating, and re-assessments of existing  $^{14}\text{C}$  dating results (Ukkonen et al. 1999, 2007, Olsen et al. 2001, Mäkinen 2005, Wohlfarth 2010, Paus et al. 2011, Sarala and Eskola 2011). This indicates that the MIS 3 lake sequence in the Sokli basin most probably has been truncated during the last glaciation in late MIS 3/MIS 2 (Figure 4-1). Pollen assemblages and macrofossil remains in the MIS 3 deposits in Fennoscandia show *Betula*-dominated vegetation and the local presence of tree *Betula* (Thoresen and Bergersen 1983, Eriksson 2005, Väliranta et al. 2012).

Earlier, the existence of extensive ice-free areas over Fennoscandia during (parts) of MIS 3 have been suggested in Finland by Korpela (1969), in Sweden by Lundqvist (1967, 1978), and in Norway by Thoresen and Bergersen (1983). This idea, however, was later drastically changed and glaciation was inferred to have persisted over large parts of Fennoscandia from MIS 4 into MIS 2 (Mangerud 1991, Lundqvist 1992, Donner 1995, Kleman et al. 1997). Several factors seem to have contributed to this change: 1)  $^{14}\text{C}$  dates, often on bulk sediment and near the limit of the radiocarbon dating method, were found to be unreliable, 2) the presence of *Betula*-dominated vegetation in Fennoscandia during MIS 3 was found to be incompatible with tundra type vegetations reconstructed

on the northern European mainland, and 3) the marine oxygen isotope record was used as a proxy for global ice volume, disregarding e.g. the affect of deep ocean water temperature changes on the isotope values, and suggesting a relatively large global ice volume during MIS 3 (Figure 4-1b). The data reviewed in the present report, however, strongly supports the earlier view of a very restricted ice-cover over northern Europe during MIS 3.

It is a challenge for future research to determine the global ice volume and extent of glaciations during MIS 3. This is particularly relevant since ice sheet dynamics are identified as one of the major drivers in the millennial scale climate variability during MIS 3. MIS 3 lasted some 30 kyr and, when taking the dynamic character of ice sheets into consideration, it might have been characterized by a series of ice advances out of the Scandian mountain range. In order to detect these glacial events in the geological record, however, more long sediment records from northern Europe are needed and efforts should be put on dating of sediments. At the same time, efforts should be put on further refining the reconstructions of global ice volume based on e.g. marine data and corals and the geological record of Laurentide glaciation in northern America. The data reviewed in this report indicates that climate variability during the LI-G cycle was mainly in terms of changes in degrees of continentality including large shifts in winter temperatures and precipitation values; July temperatures, on the other hand, remained relatively stable. These unfavourable climate conditions for snow accumulation, together with the registration of peat accumulation, lake sedimentation and tree birches over large parts of Fennoscandia until ca. 35 kyr BP, suggest that glaciation in general has been restricted during MIS 3. It is possible that changes in sea-ice cover, and the related changes in continentality, were more distinctive than glaciations for the LI-G cycle.

### 4.3 Stratigraphic implications

MIS 5 was defined as an interglacial period in the original marine oxygen isotope stratigraphy (Emiliani 1955). It was subsequently subdivided into stages 5a to 5e, with MIS 5e alone correlative with the last interglacial (Eemian Interglacial) on the European continent (Shackleton 1969). In the literature there is at present no consistency in the delimitation of the last interglacial in the marine record. It is either referred to as corresponding exclusively to MIS 5e (e.g. Martinson et al. 1987, McManus et al. 2002, Cutler et al. 2003) or encompassing the whole of MIS 5 (e.g. McIntyre and Ruddiman 1972, McManus et al. 1999, Oppo et al. 2006, Martrat et al. 2007). Confusion consists also in the European terrestrial stratigraphy. The Eemian Interglacial (MIS 5e) is denoted as the last interglacial period in the NW European mainland stratigraphy, whereas the environmental records from west-central (Woillard 1978, Turon 1984) and northern Europe (Helmens et al. 2012) prescribe a Last Interglacial Complex that encompasses the entire stage 5. In Quaternary stratigraphy, there is no unambiguous definition of an interglacial (Kukla et al. 2002b).

The proxy records compiled in the present report show that environmental conditions and climate evolution were different during MIS 5 compared to MIS 4-2. MIS 5 was mostly mild with warmest or peak interglacial conditions at the very start in early MIS 5e. MIS 4-2 was mostly cold with most extreme or peak glacial conditions in the closing phase in late MIS 2. Cool phases of short duration, with mountain-centered glaciation, punctuated MIS 5; coldest and driest conditions are recorded for MIS 5b. Similarly, the warm episodes of MIS 3 to early MIS 1 age were of relative short duration only; most favourable climate conditions, accompanied by large-scale deglaciation, occurred in early MIS 3. The present compilation favors a subdivision of the last climate cycle in an early interglacial part (MIS 5) and a late glacial part (MIS 4-2). It questions the interstadial-stadial character of the Early Weichselian, or Early Glacial, in the northwest European mainland stratigraphy, and suggests the inclusion of the Early Weichselian (MIS 5d-a) together with the Eemian Interglacial (MIS 5e) into a Last Interglacial Complex (MIS 5).

The characterization used here for subdividing the last climate cycle into an early, interglacial part (MIS 5) and a late, glacial part (MIS 4-2) can also be applied to proxy records from southern Europe, i.e. from highly sensitive sites located close to important glacial tree refugial area, such as the high-resolution pollen record from Lago Grande di Monticchio in Italy (Brauer et al. 2007). Outside the present area of interest, the long pollen records from the Colombian Andes at tropical latitudes (e.g. Groot et al. 2011) and the Chinese loess/paleosol sequence (e.g. Rokosh et al. 2002, Lu et al. 2007), i.e. some of the most complete records of Quaternary climate derived from terrestrial

deposits, allow a similar subdivision of the Late Pleistocene into an overall mild MIS 5 (interglacial) and overall cold MIS 4-2 (glacial). Long and high-resolution marine proxy records from the subpolar North Atlantic (McManus et al. 1999) and the Iberian margin at mediterranean latitudes (Martrat et al. 2007), as well as the long and high-resolution pollen record from Tenagi Philippon in Greece (Tzedakis et al. 2003), show that the last four climate cycles are all characterized by an overall cold, glacial half following a relatively warm, and largely ice-free, interglacial part, each with a duration of ca. 50 kyr. The cycles are not identical due to different orbital configurations.

Apart from identifying stratigraphy implications, this compilation comes to a conclusion rather similar to the one made earlier by Guiter et al. (2003). The latter paper compares long lake sequences from France with other west European limnic deposits, fluvial systems and loess accumulations, including both qualitative and quantitative data, for the LI-G cycle. Guiter et al. (2003) state that 'the old idea of warm interglacial periods alternating with long glacial periods is not valid any more; the dominantly temperate MIS 5 lasted ca. 50 kyr, which does not differ much with the duration of the glacial (MIS 4-2) at ca. 50 kyr'. The present report adds that warm conditions, accompanied by large-scale deglaciation, additionally occurred in the Last Glacial during MIS 3. A major departure from the classic idea of extensive glaciation over Fennoscandia during a major part of the last climate cycle was recently taken by Lundqvist (2011). According to Lundqvist (2011), the extensive glaciation of Sweden during MIS 2 seems to be as unique as the absence of glaciation during MIS 5e. During the remaining ca. 80% of the last climate cycle, glaciation might have been restricted to the Scandianvian mountains and northern parts of Sweden, or the mountains only (Lundqvist 2011).

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