Deglacial impact of the Scandinavian Ice Sheet on the North Atlantic climate system

Francesco Muschitiello
υστέρῳ δὲ χρόνῳ σεισμῶν ἐξαισίων καὶ κατακλυσμῶν γενομένων, μιᾶς ἡμέρας καὶ νυκτὸς χαλεπῆς ἐπελθούσης, τὸ τε παρ᾽ ύμῖν μάχην πάν ἀθρόον ἔδω κατὰ γῆς, ἢ τε Ατλαντίς νῆσος ὡσαύτως κατὰ τῆς θαλάττης δῦσα ἠφανίσθη

Plato (Timaeus)
Abstract

The long warming transition from the Last Ice Age into the present Interglacial period, the last deglaciation, holds the key to our understanding of future abrupt climate change. In the last decades, a great effort has been put into deciphering the linkage between freshwater fluxes from melting ice sheets and rapid shifts in global ocean-atmospheric circulation that characterized this puzzling climate period. In particular, the regional expressions of climate change in response to freshwater forcing are still largely unresolved.

This project aims at evaluating the environmental, hydro-climatic and oceanographic response in the Eastern North Atlantic domain to freshwater fluxes from the Scandinavian Ice Sheet during the last deglaciation (∼19,000-11,000 years ago). The results presented in this thesis involve an overview of the regional representations of climate change across rapid climatic transitions and provide the groundwork to better understand spatial and temporal propagations of past atmospheric and ocean perturbations.

Specifically, this thesis comprises i) a comparison of pollenstratigraphic records from densely 14C dated lake sediment sequences, which provides insight into the regional sensitivity of North European vegetation to freshwater forcing in the Nordic Seas around the onset of the Younger Dryas stadial (∼12,900 years ago); ii) a reconstruction of North European hydro-climate, which, together with transient climate simulations, shed light on the mechanisms and regionality of climate shortly prior to the transition into the Younger Dryas stadial; iii) studies of a ∼1250-year long glacial varve chronology, which provides an accurate timing for the sudden drainage of proglacial freshwater stored in the former ice-dammed Baltic Ice Lake into the North Atlantic Ocean; iv) a 5000-year long terrestrial-marine reconstruction of Eastern North Atlantic hydro-climate and oceanographic changes that clarifies the hitherto elusive relationship between freshwater forcing and the transient behaviour of the North Atlantic overturning circulation system. The results presented in this thesis provide new important temporal constraints on the events that punctuated the last deglaciation in Northern Europe, and give a clearer understanding of the ocean – atmosphere – ice-sheet feedbacks that were at work in the North Atlantic. This increases our understanding of how the Earth climate system functions in more extreme situations.
Svensk sammanfattning

Den långa, successivt varmare övergångsperioden, avbruten av flera kalla episoder, från den senaste istiden in i den nuvarande interglaciala värmeperioden, dvs den senaste deglaciationen/isavsmältningen, har ledrådar till vår förståelse av framtida abrupta klimatförändringar. Under de senaste årtiondena har stora ansträngningar gjorts för att dechiffra kopplingar mellan sötvattenpulser från smältande inlandsisar och snabba förändringar i den globala ocean-atmosfäriska cirkulationen, vilket var kännetecknande för denna delvis gåtfulla klimatperiod. Speciellt är de regionala klimatyttringarna av stora sötvattenflöden ett oloöst problem.

Syftet med detta projekt har varit att utvärdera den miljömässiga, hydroklimatiska och oceanografiska responsen i östra Nordatlanten på stora sötvattenflöden från den Skandinaviska inlandsisen under den senaste deglaciationen, ca 19,000 till 11,000 år före nutid. Resultaten i avhandlingen innefattar en översikt av hur det regionala klimatet påverkas vid snabba klimatiska övergångsperioder och utgör därmed ett underlag för att bättre förstå hur störningar i dåtidens atmosfär och ocean kunde spridas, både rumsligt och tidsmässigt.

Mer specifikt innefattar denna avhandling i) en jämförelse av olika pollenstratigrafiskt undersökta och noggrant 14C daterade sjösediment, vilka ger inblick i den regionala vegetationens sensitivitet för sötvattenflöden till de Nordiska haven i samband med att den kalla yngre dryas perioden inleddes för ca 12,900 år sedan; ii) en rekonstruktion av det nordvästeuropeiska hydroklimatet för ca 13,000 år sedan, vilket i kombination med transienta klimatsimuleringar klarlägger klimatets mekanismer och regionalitet strax före övergången till yngre dryas; iii) undersökningar av en ca 1250 år lång kronologi baserad på glaciala lervarv, vilken ger en exakt ålder för den plötsliga dräneringen/tappningen av sötvatten från den proglaciala Baltiska Issjön ut till Nordatlanten; iv) en 5000 år lång terrester-marin rekonstruktion av östra Nordatlantens hydroklimat och oceanografiska förändringar, vilken klargör det hittills gäckande förhållandet mellan sötvattenflöden och de transienta processerna i Nordatlantens djupvattenbildning, den s.k. termohalina cirkulationen. Resultaten i avhandlingen ger nya viktiga tidsmässiga begränsningar för de händelser som ideligen störde och avbröt utvecklingen i samband med den senaste deglaciationen i Nordeuropa. Detta ger en ökad insikt i den dåvarande Nordatlantens oceaniska, atmosfäriska och glaciala återkopplingsmekanismer, vilket ökar förståelsen för hur jordens klimatsystem fungerar under mer extrema förhållanden.
List of papers and author contributions

This thesis consists of an overview of the main aims of this PhD project, the employed methodological approach, and summaries of the key results. The appendices listed below are also included. Paper I, II and III have been published in the journals indicated and are reprinted under permission of the respective publishers. Paper IV is a manuscript.


Paper I: F.M. conceived the study, was the main contributor in terms of analyses, wrote the initial version of the paper and made the figures. B.W contributed with writing and interpretation of the results.

Paper II: F.M. conceived the study, performed isotope and geochemical analyses on lake sediment cores, interpreted the proxy data, performed statistical analysis, wrote the initial version of the paper and made the figures. F.S.R.P. analysed climate model output and contributed to the interpretation of the proxy data. J.E.W. performed the chironomid analysis. R.H.S. contributed to biomarker data evaluation. A.A.M.S. analysed the air back trajectory data. S.J.B. and N.J.W. contributed to the temperature data evaluation. A.K.C. performed the pollen analysis. B.W. led the fieldwork campaign, subsampling and identification of samples for terrestrial $^{14}$C analysis, provided insight into regional paleoenvironment and acquired financial support. All authors contributed with interpretation of the results and editing of the manuscript.

Paper III: F.M. conceived the study, performed geochemical analyses of the sediments and statistical analyses, wrote the initial version of the paper and made the figures. J.M.L. designed and performed the ice-flow model analysis. F.M and J.M.L. led the fieldwork campaign for the new sediment cores and interpreted the geochemical results. S.L.G. was the main contributor of Figure 1 and helped with the interpretation of the regional paleogeography. F.M.N. provided the ice flow model. L.B. provided the clay varve data set from Sandfjärden. A.M. contributed to varve data evaluation. B.W. provided the clay varve data sets for Östergötland, varve thickness data and IRD counts, insight into regional paleoenvironment and acquired financial support. All authors contributed with interpretation of the results and editing of the manuscript.
Paper IV: F.M. conceived the study, performed isotope and geochemical analyses on lake sediment cores, interpreted the proxy data, performed statistical analysis, wrote the initial version of the paper and made the figures. T.M.D. performed the marine multi-proxy analyses and provided the marine $^{14}$C data; M.V. contributed with the macrofossil data; R.H.S. contributed to biomarker data evaluation; S.B., S.M.D, T.L., F.S., and P.J.R. helped with interpretation of proxy data. B.W. led the fieldwork campaign for the terrestrial study site, subsampling and identification of samples for terrestrial $^{14}$C analysis, provided insight into regional paleoenvironment and acquired financial support. All authors contributed with interpretation of the results and editing of the manuscript.

The following papers are not included as a part of this thesis:


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1. Introduction

Characterising the impact of melting ice sheets on the global climate system in response to global warming requires a comprehensive understanding of the interplay between the cryosphere, oceans and atmosphere at regional scales. Specifically, as the stability of the Greenland Ice Sheet and other sources of freshwater stored over northern high-latitude continental regions are under threat (Fig. 1) (Moon et al., 2012; Shepherd et al., 2012; Hanna et al., 2013), large uncertainties are cast upon the fate of the Atlantic Meridional Overturning Circulation (AMOC) – a critical component of the Earth’s system.

In the North Atlantic Ocean – the key centre of action of the AMOC – warm and saline surface waters carried from the subtropical sector rapidly cool and sink. The process releases heat to the atmosphere with substantial impacts on hydro-climate and temperatures over large regions, and more critically over Western and Northern Europe. Shifts in atmospheric circulation patterns are thus of central concern in the debate surrounding the transient behaviour of the AMOC, as they can lead to extreme weather events over short time scales due to the movement of fronts, diversion of Rossby waves, or persistent atmospheric patterns. Therefore, future changes in ocean thermohaline properties owing to increasing meltwater discharge have the potential to drive significant shifts in regional climates with profound consequences for ecosystems and societies.

The diagnostic ability to understand and predict climate change can be aided by knowledge of past climate analogues. For instance, the transition from the Last Ice Age to the present warm Interglacial, the last deglaciation (~19,000-11,000 years ago), provides an ideal natural laboratory to decipher the physical mechanisms behind rapid climate change. The last deglaciation was a critical period of climate shifts during which every component of the Earth’s system underwent numerous abrupt and rapid large-scale changes (Fig. 2) (Denton et al., 2010; Clark et al., 2012).

![Figure 1. a, Global anomaly of mean 2 m air temperature change (T2m) estimated with the full set of GCMs from the CMIPS data base. Values are simulated with respect to 1970-1999 for experiments of the historical period (grey, 41 models), and the Representative Concentration Pathways (RCP) 4.5 (blue; 42 models) and RCP 8.5 (red; 40 models) scenarios (Moss et al., 2010). The full ensemble means are displayed as thick lines; vertical bars refer to ±1σ for the reference period 2071-2100. b, Same as (a) but for the Arctic, defined as the region north of 60° N. c, Same as a but for the Greenland Ice Sheet (GIS) defined as the region covering 60-85° N and 20-70° W and applying a land/sea mask from each GCM to confine the analysis to the land area. Mass balance terms of the GIS are also shown (Box and Colgan, 2013). Data are presented as cumulative anomalies relative to the reference period 1840-1900.](image-url)
These shifts were most prominently expressed in the North Atlantic region (Björck et al., 1996; Lowe et al., 2008; Steffensen et al., 2008), but with lower amplitudes in the Southern Hemisphere (Fig. 2) (Barker et al., 2009; Stenni et al., 2011; Shakun et al., 2012). The two longest and coldest climate reversals in the Northern Hemisphere, are commonly termed Younger Dryas (YD; ∼12,900-11,700 years ago) and Oldest Dryas (>14,700 years BP), and the informal name for the warmer interstadial prior to the YD is Bølling-Allerød (14,700-12,900 years ago), in reference to earlier pollen-stratigraphic work in Scandinavia (Iversen, 1954; Mangerud et al., 1974; Wohlfarth, 1996).

The explanation for the occurrence of multiple warm and cold intervals at the end of the Last Ice Age, when northern summer insolation was steadily increasing, has presented a major challenge for the paleoclimate community. Much research during the past decades has therefore been placed on multi-proxy analyses of terrestrial, marine and ice core archives and on correlations between the different archives to detect and quantify the impact of these dramatic climatic shifts. However, precise correlations were - and still are - difficult due to intrinsic limitations with dating techniques and the insufficient resolution of existing records (Lane et al., 2013; Blockley et al., 2014; Rasmussen et al., 2014a).

Greenland ice core records however stand out in this respect and have therefore played a pivotal role in discussing the underlying causes of abrupt climate variability. The ice-core data sets record past climatic changes in a multitude of atmospheric proxies (Steffensen et al., 2008); provide a continuous annual chronology throughout the Last Interglacial-Glacial cycle in the North Atlantic region (Rasmussen et al., 2014b); and can be synchronized to Antarctic ice cores using methane measurements and volcanic aerosol signatures (Blunier et al., 1998; Buizert et al., 2015; Sigl et al., 2015). Such synchronization allows a direct correlation and comparison of Northern and Southern Hemisphere climatic changes. This has led to the hypothesis of the Atlantic bipolar seesaw mechanisms, which attributes a large role to the AMOC in triggering abrupt global climate shifts, through an asymmetric inter-hemispheric temperature forcing (Broecker, 1998; Knutti et al., 2004; Barker et al., 2009; Stenni et al., 2011; Cvijanovic et al., 2013). Critically, this mechanism, which influenced climate in the Atlantic and neighbouring sectors, had a greater impact in northern high- to mid-latitudes during the last deglaciation (Shakun et al., 2012).

North Atlantic marine reconstructions compellingly show that the AMOC system underwent large perturbations during the last deglaciation (McManus et al., 2004; Roberts et al., 2010). However, it remains an open question as to whether sudden AMOC instabilities were a response to continental meltwater discharge from the North Atlantic seaboard (Duplessy et al., 1992; Bard, 2000; Clark et al., 2001), to changes in regional sea-
ice distribution (Bradley and England, 2008), to an intrinsic threshold behaviour of the coupled atmospheric-ice sheet system (Zhang et al., 2014), or to an intertwinment of the factors cited above. More importantly, it is largely unclear how freshwater perturbations mediated direct and indirect effects on regional shifts in climate and atmospheric circulation. Furthermore, the temporal expressions of continental climate responses relative to ocean circulation changes in the North Atlantic are still poorly resolved, thereby limiting our understanding of the true driving mechanisms and the direction of physical events associated with rapid climate change.

2. Thesis objectives and key results

One way to explore the issues broached above is to improve existing regional proxy-record chronologies to create a broad spatial network of well-constrained climate reconstructions. On the other hand, a means to directly investigate the mechanisms driving the coupled ocean-atmosphere system is to generate isotope records of precipitation from lake sediments, as the physical properties of precipitation associated with regional hydro-climate patterns are expected to respond without delay to large-scale shifts in ocean and atmospheric circulation. If analysed at adequate resolution and supported by precise chronologies, lake sediment isotope analysis has thus the potential to provide temporal climate reconstructions that deliver information on both the hydrographic and hydrological system.

In this thesis project I provide: i) a North European perspective on freshwater-driven environmental, climatic and hydrological changes during the last deglaciation; ii) an improved understanding of the ocean – atmosphere – ice-sheet feedbacks and mechanisms at work in the Nordic Seas; and iii) better chronological constraints on key hydrological and hydrographic events that occurred during the last deglaciation. Altogether, this study discloses new research directions to generate more reliable marine chronologies in the North Atlantic, thus aiding in the comparison of marine and terrestrial climatic reconstructions.

Explicitly, I have applied geochronological models to a large set of marine and lake sediment records to assign precise temporal constraints to past climatic events; generated high-resolution isotope and other geochemical data sets for two lake sediment sequences to reconstruct the deglacial hydro-climate; and assembled an extensive data set comprising all available North European quantitative paleoclimatic reconstructions spanning the last deglaciation. This overall approach was complemented with climate model simulations. The most significant results of this PhD work are:

1 - The establishment of precise geo-chronologies for eight key terrestrial and marine sedimentary records from Northern Europe and the Nordic Seas. I have reconstructed a new deglacial $^{14}$C reservoir age record for the Nordic Seas, which constitutes the best temporally resolved record of its kind and can serve as a future chronological benchmark for North Atlantic paleoclimate reconstructions.

2 - The construction of a new 1250-year long glacial varve chronology, which tracks the annual recession of the Scandinavian Ice Sheet in southern Sweden. The varve chronology was placed on an absolute time scale, and by using geochemical analyses I identified the first catastrophic drainage of the Baltic Ice Lake with an unprecedented precision (±2 years).
3 - The generation of the first North European hydro-climate reconstructions based on isotope analyses (δD) of specific molecular compounds from southern Swedish lake sediments spanning the last deglaciation.

3. Investigation area

3.1. Background

One of the central aims of this study is to investigate deglacial atmospheric circulation dynamics over Northern Europe by using lake sediment stable isotope analyses. In Northern Europe this information is scarce and mainly relies on bulk sedimentary organic matter (Ahlberg et al., 1996; O’Connell et al., 1999; Jones et al., 2002; Marshall et al., 2002; Diefendorf et al., 2006), which makes it difficult to extract climate factors from the noise of lake endogenic processes (Leng and Henderson, 2013).

In southern Sweden, some of the problems associated with stable isotope analyses on bulk sedimentary carbonates have been circumvented by using isotope measurements on specific carbonate components of lake sediments, such as mollusc shells, ostracod valves, and algae encrustations (Hammarlund and Keen, 1994; Hammarlund and Lemdahl, 1994; Hammarlund, 1999; Hammarlund et al., 1999). However, these records are fragmentary or supported by poor chronological frameworks. Therefore, to fill this gap, new well-dated isotopic records from southern Sweden based on lipid biomarker compounds were generated (Fig. 3). The records were obtained from two lake sediment sequences, Hässeldala Port (56° 16’ N; 15° 03’ E, 40 m a.s.l.) and Atteköps Mosse (56° 23’ N; 12° 51’ E, 180 m a.s.l.), located along the south-eastern and south-western coast of Sweden, respectively (Fig. 3). These sites are today small peat bogs, but contained lakes during the last deglaciation.

Stable isotope signatures on lacustrine lipid compounds have been successfully applied to reconstruct paleo-hydrological processes associated with the last deglaciation in Western Europe (Rach et al., 2014). Precisely, these isotope records have allowed reconstructing regional shifts in precipitation patterns and water vapour availability, providing knowledge on the dynamics of North Atlantic storm tracks and the physical characteristics of North Atlantic Ocean waters.

Southern Sweden is an excellent area to employ these novel isotope proxies, as here hydro-climate shifts scale in a linear fashion with upwind, near-field oceanographic metrics (Fig. 4a, b). Indeed, sea surface temperatures (SST) primarily control the amount of moisture delivered to the region by regulating the flux of moisture released from the seawater to the atmosphere (Fig. 4a). Under modern conditions, the amount of precipitation in southern Sweden is tightly linked with the prevailing westerly winds that pick up moisture from the North Sea and the Skagerrak-Kattegat basin, which constitute the main source of moisture for precipitation (Gustafsson et al., 2010). By contrast, precipitation is less abundant during an anticyclonic regime (Fig. 4b) and moist air is generally transported from the Baltic Sea only under exceptionally warm surface water conditions (Gustafsson et al., 2010).

During the last deglaciation, the region was located south of the Scandinavian Ice Sheet margin and downwind of its primary drainage route. Therefore, it is likely that meltwater discharge from the ice sheet to the North Sea resulted in lower SSTs and fresher waters. This implies a lower water-to-air moisture uptake at times of increased meltwater outflow, with relatively depleted isotope signatures of seawater as the moisture source.
became fresher. Consequently, hydro-climate records on land are likely to capture the meltwater signal as this instantaneously propagates downwind in the form of drier air reaching the coastal area and lighter stable isotopic signatures of precipitation.

In turn, terrestrial hydro-climate proxy records from southern Sweden are not only useful to reconstruct regional hydrology and precipitation patterns, but potentially ideal to better understand the regional coupling between ice-sheet, ocean and atmosphere. Furthermore, these indirect freshwater reconstructions can benefit of robust and atmospheric-based $^{14}$C chronologies obtained from terrestrial macrofossils, which circumvent the intrinsic chronological uncertainties associated with marine reconstructions.

**Figure 3.** Core locations that form part of this thesis and main sites discussed in the text. Red circled dots indicate sites where new data were generated within the framework of this thesis, i.e. Atteköps Mosse (ATK), Hässeldala Port (HÅ), marine core MD99-2284. Green dots indicate sites used for comparative analysis and/or re-evaluation of available data. 1, Sluggan Bog; 2, marine core HM79-6/4; 3, Meerfelder Maar; 4, Kvaltjern; 5, Kråkenes; 6, Kulturmyra; 7, Lake Gammelmose; 8, Lake Madtjärn. The green rectangle in south-eastern Sweden highlights the area of investigation dealt with in Paper III. The blue line denotes the approximate extension of the former Baltic Ice Lake (Björck et al., 1996) during the Late Allerød (Riede et al., 2011). The white line indicates the estimated ice margin for the Scandinavian Ice Sheet limit at 13,000 years BP (Hughes et al., 2015).

**3.2. Previous work**

The ancient lake of Hässeldala Port is located in Blekinge, southern Sweden (Fig. 3) and filled in during the Early Holocene. The site is today a small peat bog covering $\sim20$ m$^2$. Complete sediment sequences of variable depth have been retrieved and analysed using different proxies (Davies et al., 2003, 2004; Wohlfarth et al., 2006; Kylander et al., 2013; Steinthorsdottir et al., 2013; Ampel et al., 2015; Muschitiello et al., 2015a).
The basin contains a sedimentary sequence that covers the period between the Late Bølling and Early Holocene pollen zone (∼14,500-9500 years ago). The sediments have been extensively studied over the last decade using a variety of biological and geochemical proxies. The tephro-chronological framework of the site was first established by Davies et al. (2004, 2003). The pollen- and litho-stratigraphy was established by Wohlfarth et al. (2006). More recently, sediment geochemistry (Kylander et al., 2013), fossil leaf stomata (Steinthorsdottir et al., 2013), diatom (Ampel et al., 2013), and biomarker records (Muschitiello et al., 2015a) have been investigated.

The ancient lake of Atteköps Mosse is located in southwestern Sweden close to the border between the provinces of Skåne and Halland (Fig. 3). It is a small basin that filled in during the Early Holocene. The basin, which today is a peat bog covering ∼200 m², has a full deglacial and Holocene stratigraphic sequence (from ∼16,000 years ago to present) (Veres, 2001). The site has been previously investigated by Veres (2001), who established a litho-stratigraphy, analysed the sediments for loss-on-ignition, grain-size, magnetic susceptibility and 14C dating.

**Figure 4. a**, Modern moisture source distribution and transport to the main study area. Summer (JJA) correlation between specific humidity in southern Sweden (averaged across 56-57°N and 12-16°W; HadCRUH) and sea-surface temperature over the adjacent seas relative to the period 1974-2003 (contour; HadSST1). The yellow line delimits the area where the correlation is 95% CI. **b**, Summer (JJA) relationship of blocking circulation versus precipitation (CRU-TS3.23) in southern Sweden (as defined in a). Blocking circulation is here characterized as a pressure index defined by the 850 hPa atmospheric pressure difference (Trenberth’s NH) between 10°E and 40°W at 65°N. A positive blocking index indicates northward flow and negative values southward flow. Thick lines represent the 10-year moving averages. Precipitation data is expressed as an anomaly relative to the period 1974-2003 and presented on a reverse axis. Northward flow negatively correlates with precipitation anomalies in the study area ($R^2 = 0.50$). The yellow dashed line indicates the total observed freshwater (FW) storage anomaly of the Norwegian Sea relative to the observational period (Glessmer et al., 2014).

### 4. Materials, methods and applications

#### 4.1. Sampling

At Hässeldala Port, a number of cores have been collected over the years and the present isotope analyses refer to Core 5. The sediment sequences were collected in 2011 using a
Russian corer (10 cm diameter, 1 m length) with 0.5 m overlap between the successive cores. The lithology of Core 5 is shown in Table 1. Core 5 was not only sub-sampled for stable isotope analyses but also for loss-on-ignition, fossil-leaf stomatal, diatoms, and $^{14}$C dating (Steinthorsdottir et al., 2013; Ampel et al., 2015).

For the site of Atteköps Mosse, sediment sequences were collected in 2010 using a Russian corer (7.5 and 5 cm diameter, 1 m length) with 0.5 m overlap between the successive cores. The cores were first scanned for X-ray fluorescence analyses and the lithologies were then described (Table 2). Based on these, a composite stratigraphy was created, which was the basis for further sub-sampling. Sub-samples were taken for loss-on-ignition, carbon and nitrogen, chironomid, biomarker, ancient DNA, and tephra and $^{14}$C dating.

**Table 1** – Lithostratigraphic description of the Håsseldala sediment succession. Sediment units are numbered according to the reference lithostratigraphy of Wohlfarth et al. (2006).

<table>
<thead>
<tr>
<th>Unit</th>
<th>Depth (cm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>12-11</td>
<td>270.5-303.5</td>
<td>Dark brown peaty gyttja or gyttja peat</td>
</tr>
<tr>
<td>10</td>
<td>303.5-308.5</td>
<td>Dark brown gyttja</td>
</tr>
<tr>
<td>9</td>
<td>308.5-322.5</td>
<td>Brown gyttja</td>
</tr>
<tr>
<td>8a</td>
<td>322.5-332.5</td>
<td>Light brown algae gyttja clay/clayey algae gyttja</td>
</tr>
<tr>
<td>8b</td>
<td>332.5-334.5</td>
<td>Brown clayey silty algae gyttja</td>
</tr>
<tr>
<td>8c</td>
<td>334.5-338.5</td>
<td>Light brown algae gyttja clay or clayey algae gyttja</td>
</tr>
<tr>
<td>7a</td>
<td>338.5-341.5</td>
<td>Bioturbated zone; mix of the upper light brown layer and the lower dark brown gyttja</td>
</tr>
<tr>
<td>7b-6</td>
<td>341.5-348.5</td>
<td>Medium brown clayey algae gyttja; visible plant macros</td>
</tr>
<tr>
<td>5</td>
<td>348.5-358.5</td>
<td>Medium brown gyttja clay/clay gyttja</td>
</tr>
<tr>
<td>3a</td>
<td>358.5-362.5</td>
<td>Brown clayey algae gyttja</td>
</tr>
<tr>
<td>3b</td>
<td>362.5-364.5</td>
<td>Light brown-yellowish silty clay</td>
</tr>
<tr>
<td>2</td>
<td>364.5-369.5</td>
<td>Yellowish-beige silty clay</td>
</tr>
</tbody>
</table>

4.2. $^{14}$C dating

Radiocarbon ($^{14}$C) dating is the most widely used technique to infer the down-core age of lake and marine sedimentary records. $^{14}$C atoms are produced in the upper atmosphere, where cosmic rays lead to the collision of free neutrons with nitrogen atoms ($^{14}$N) causing a displacement of protons that turns $^{14}$N in $^{14}$C. $^{14}$C atoms are oxidised to carbon dioxide ($^{14}$CO$_2$), mixed in the atmosphere and oceans, and taken up by living organisms via metabolic activity. Upon death of the organism, CO$_2$ uptake ceases, whereas the radioactive $^{14}$C slowly decays into the stable element $^{14}$N (Libby, 1952).

Lake and marine sediments generally contain a certain amount of organic carbon in the form of fossil material (plant macro remains and foraminifera, respectively, among many others), which can be dated by the radiocarbon method. This method allows measuring the decay of $^{14}$C from the time of an organism’s death until present, i.e. the time of measurement (Libby, 1952).

The atmospheric $^{14}$C content varies over time due to changes in production rates and owing to carbon exchange between different reservoirs, e.g. the oceans, atmosphere, biosphere, and cryosphere. Therefore, to estimate the absolute age associated with a $^{14}$C measurement, it is necessary not only to account for the related decay factor, but also for
changes in past atmospheric $^{14}$C levels. This requires the use of a calibration curve, whereby the atmospheric $^{14}$C content can be directly equated to a calendar age (Stuiver and Kra, 1986; Stuiver and Braziunas, 1993).

The internationally ratified radiocarbon calibration curve IntCal13 (Reimer et al., 2013) is a calibration data set based on several absolutely dated records that have incorporated carbon from the atmosphere at the time of formation. For the last deglaciation, the IntCal13 curve is based on a number of independent records (Fig. 5a). After 13,900 calibrated years before 1950 AD (hereafter cal. years BP), the calibration curve is defined by precise dendrochronological measurements (Hua et al., 2009; Friedrich et al., 2004; Kromer et al., 2004). Prior to 13,900 cal. years BP the curve relies on $^{14}$C measurements from marine sediments (Bard et al., 2013; Hughen et al., 2006, 2004), corals (Durand et al., 2013; Fairbanks et al., 2005; Bard et al., 1990), speleothems (Southon et al., 2012; Hoffmann et al., 2010; Beck et al., 2001), and varved lake sediments (Bronk Ramsey et al., 2012).

**Table 2 – Lithostratigraphic description of the Atteköps mosse sediment succession.**

<table>
<thead>
<tr>
<th>Unit</th>
<th>Depth (cm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>402-462</td>
<td>Dark brown coarse detritus gyttja or peat</td>
</tr>
<tr>
<td>8</td>
<td>462-510</td>
<td>Dark brown fine detritus gyttja</td>
</tr>
<tr>
<td>7e</td>
<td>510-511.5</td>
<td>Brown to greyish silty gyttja layer</td>
</tr>
<tr>
<td>7d</td>
<td>511.5-542</td>
<td>Medium brown silty gyttja/algae gyttja</td>
</tr>
<tr>
<td>7c</td>
<td>542-549</td>
<td>Brown silty/algae gyttja</td>
</tr>
<tr>
<td>7b</td>
<td>549-582.5</td>
<td>Dark brown silty algae gyttja</td>
</tr>
<tr>
<td>7a</td>
<td>582.5-590</td>
<td>Brown silty gyttja/algae gyttja</td>
</tr>
<tr>
<td>6b</td>
<td>590-613.5</td>
<td>Brown silty gyttja</td>
</tr>
<tr>
<td>6a</td>
<td>613.5-625</td>
<td>Silty dark grey clayey gyttja</td>
</tr>
<tr>
<td>5f</td>
<td>625-627.5</td>
<td>Brownish-grey clayey silt</td>
</tr>
<tr>
<td>5e</td>
<td>627.5-629</td>
<td>Dark brown clayey silt</td>
</tr>
<tr>
<td>5d</td>
<td>629-655</td>
<td>Brownish-grey clayey silt</td>
</tr>
<tr>
<td>5c</td>
<td>655-664</td>
<td>Brownish-orange clayey silt</td>
</tr>
<tr>
<td>5b</td>
<td>664-673</td>
<td>Brown-greyish silt</td>
</tr>
<tr>
<td>5a</td>
<td>673-682</td>
<td>Dark brown silt</td>
</tr>
<tr>
<td>4</td>
<td>682-702</td>
<td>Alternating layers with dark brown mosses and greyish-brown clayey silt/silt clay</td>
</tr>
<tr>
<td>3d</td>
<td>702-704</td>
<td>Grey silt</td>
</tr>
<tr>
<td>3c</td>
<td>704-708</td>
<td>Grey coarse sand</td>
</tr>
<tr>
<td>3b</td>
<td>708-710</td>
<td>Grey sandy silt with mosses</td>
</tr>
<tr>
<td>3a</td>
<td>710-714.5</td>
<td>Grey fine sand</td>
</tr>
<tr>
<td>2</td>
<td>714.5-724</td>
<td>Alternating layers of dark brown mosses and greyish-brown clayey silt</td>
</tr>
<tr>
<td>1e</td>
<td>724-728</td>
<td>Brownish-grey clayey silt</td>
</tr>
<tr>
<td>1d</td>
<td>728-731</td>
<td>Dark grey fine sand layer</td>
</tr>
<tr>
<td>1c</td>
<td>731-740</td>
<td>Grey clayey silt</td>
</tr>
<tr>
<td>1b</td>
<td>740-744.5</td>
<td>Grey silt with thin sand layers (1 mm)</td>
</tr>
<tr>
<td>1a</td>
<td>744.5-750</td>
<td>Grey fine sand layers</td>
</tr>
</tbody>
</table>
Figure 5. Radiocarbon calibration and Bayesian age-depth modeling. a, Deglacial portion of the IntCal13 radiocarbon calibration curve; raw data composing the related database (Reimer et al., 2013) is color coded by the type of proxy and expressed with errors (±2σ). b, Example of a calibration of a radiocarbon date. The red probability distribution represents the measured 14C value of the radiocarbon date. The grey-colored area indicates the calibrated probability distribution of the radiocarbon date using the IntCal13 calibration curve (yellow). The thick black lines show the calibrated age ranges that encompass the 95% CI of the measured 14C date. The dashed and solid thin lines illustrate the intersection of the errors (±2σ) and the mean of the 14C age, and the calibration curve, respectively. c, Example of a depositional process with sediment progressively accumulating over time (yellow columns). Black dots show changes in accumulation rate and the black line reflects the “true” age-depth history of the sedimentary record. The red triangles represent radiocarbon-dated samples and the red line reflects a tentative age-depth relation based on linear interpolation between the chronological constraints. d, Construction of a probabilistic age-depth model (see text for details).

Calibration to the atmospheric 14C curve (Fig. 5a, b) is a highly suitable method to convert the 14C age of samples from terrestrial organisms to absolute age. For marine samples, however, the 14C content reflects the 14CO2 dissolved in the ocean. Oceans are depleted in their 14C content relative to that of the atmosphere. This results in an apparent 14C age difference between the ocean water and the contemporaneous atmosphere, which is also referred to as radiocarbon reservoir age (Stuiver and Braziunas, 1993). During the last deglaciation, the magnitude of the reservoir age varied over time and space primarily as a function of the strength in the rate of ocean ventilation and terrestrial freshwater discharge (Waelbroeck et al., 2001; Björck et al., 2003; Bondevik et al., 2006; Thompson et al., 2011). Given that the last deglaciation was characterized by large and rapid changes in ocean circulation (e.g. McManus et al., 2004), the timing of which can only be constrained
by means of marine radiocarbon chronologies, a regional assessment of changes in reservoir age in the North Atlantic is still a challenging endeavor.

Radiocarbon dating and calibration are not only the first step towards establishing the age-depth relationship for lake and marine sedimentary sequences. The second step to infer a reliable age-depth relationship involves probabilistic age-depth modeling. This can be based on Bayesian statistics, an approach that has been widely applied in this thesis. In the following, the method and concepts behind Bayesian age-depth modeling are discussed. An outline of the method and calculations to generate the marine $^{14}$C reservoir age record presented in this thesis is also provided.

### 4.2.1. Bayesian age-depth modeling

Bayesian age-depth modeling has become increasingly popular in the last decades to reconstruct accumulation histories of radiocarbon-dated geological records (Buck et al., 1991; Blaauw and Christen, 2005; Parnell et al., 2008; Bronk Ramsey, 2008). Bayesian statistics combine data and prior information to infer the posterior distributions, i.e. predictive probability distributions that are conditional on the observed data. Bayesian depositional models are thus constructed based upon the available age measurements and using explicit prior parameter constraints such as – for instance – positive accumulation, mean accumulation rates and mean accumulation rate variance. In a depositional context, the approach aims at mathematically finding a representative set of possible ages associated with each depth interval in a sedimentary record. The full mathematical formalism for the model elaboration can be found in Blaauw and Christen (2005, 2011) and Bronk Ramsey (2008).

The model operates via a Markov Chain Monte Carlo (MCMC) sampling method, which simulates a distribution of possible solutions (Gilks et al., 1996), with a probability that is a product of the prior distribution of each parameter and the likelihood probabilities of the observed data. Therefore, the resulting posterior distributions are a probabilistic representation of the depositional history of the sedimentary record that fully accounts for the available age measurements.

In Bayesian age models, radiocarbon dates are treated in their calibrated form (Fig. 5a, b). The calibration of a $^{14}$C date $i$ with value $R_i$ and uncertainty $\delta R_i$ is obtained via comparison to the calibration curve, which provides a continuous estimation of the $^{14}$C age over time $R(t_i)$ and the associated uncertainty $\delta R(t_i)$. The agreement between the $^{14}$C age and the calibration curve at each point in time, $t_i$, can be expressed as a likelihood, $P_i(t_i)$.

$$P_i(t_i) \propto \exp \left( - \frac{(R_i - R(t_i))^2}{2(\delta R_i^2 + \delta R(t_i)^2)} \right)$$

(1)

This provides the calendar age probability distribution of a calibrated $^{14}$C date (Fig. 5b, grey patch).

To simulate the posterior distributions with MCMC, the model is generally driven by the Metropolis-Hastings sampler algorithm (Metropolis et al., 1953; Hastings, 1970). The
algorithm is used to obtain a candidate matrix $Y$ from a probability density function referred to as proposal distribution $s(Y)$. The candidate point $Y$ is then used as the new state $t$ of the chain with a probability given by $\alpha$:

$$R(t) = \frac{P_{\text{model}}(Y_t)}{P_{\text{model}}(Y_{t-1})} \quad \text{with} \quad \alpha = \min\{1, R(t)\}. \quad (2)$$

In the event the candidate matrix $Y$ is rejected, the chain stays at the same point and the state is set to $X_t = X_{t-1}$. The candidate matrix $Y$ is accepted if

$$U(0,1) \leq \alpha \quad (3)$$

with $U(0,1)$ being a uniform random number between 0 and 1. When $\alpha = 0$ all candidate matrices are accepted, whereas when $\alpha = 1$ only candidates with probability equal to 1 are accepted. In a simplified configuration, the model probability can be defined by three main contributions – time ($T$), depth ($D$) and accumulation rate constraints ($Z$).

$$P_{\text{model}} = P_T \cdot P_D \cdot P_Z \quad (4)$$

The full mathematical specification for each of the probabilities can be found in Bronk Ramsey (2008). The model is guided by a score given in terms of the negative logarithm of the posterior probability, which provides an estimate of the model performance, while the convergence of the MCMC chain around the ‘true’ parameter values is monitored by accepted $P_{\text{model}}$ values. For each step of the MCMC chain (typically more than $10^6$ iterations in total) several age estimations for every input and interpolated depth are generated (Fig. 5d). This allows to estimate age probability distributions for every given depth interval, and for the associated confidence intervals.

### 4.2.2. Reservoir age estimation

Marine radiocarbon reservoir ages are expressed as $R(t)$ and $\Delta R(t)$ (Stuiver and Braziunas, 1993). $R(t)$ is defined as the departure of a measured marine $^{14}$C age, $^{14}$C$_{M}(t)$, from the corresponding contemporaneous atmospheric $^{14}$C age, $^{14}$C$_{ATM}(t)$, on the IntCal13 calibration curve (Reimer et al., 2013) at the calibrated age $t$ of deposition of the $^{14}$C sample material (equation 5).

$$R(t_i) = ^{14}\text{C}_M(t_i) - ^{14}\text{C}_{ATM}(t_i) \quad (5)$$

The calibrated age $t$ can be inferred from a $^{14}$C date from a terrestrial sample obtained at the same depth as the marine sample, or more accessibly using age output from a depositional model, as described above (e.g. Paper IV).
On the other hand, $\Delta R$ is defined as the departure of a measured marine $^{14}$C age, $^{14}C_M(t)$, from the corresponding contemporaneous global marine $^{14}$C age, $^{14}C_{MAR}(t)$, on the Marine13 calibration curve (Reimer et al., 2013) (equation 6).

$$\Delta R(t_i) = ^{14}C_M(t_i) - ^{14}C_{MAR}(t_i)$$  \hspace{1cm} (6)

The parameters $^{14}C_M$, $^{14}C_{ATM}$, and $^{14}C_{MAR}$ are accompanied by errors, which are normally distributed, i.e. the measured radiocarbon mean value and the associated uncertainty. Calibrated ages $t_i$ can be obtained using MCMC output from an age-depth model and probability density functions can be calculated for each $R(t)$ and $\Delta R(t)$ values (e.g. Olsen et al., 2009, 2014).

### 4.3. Synchronization of climate records

A meticulous stratigraphic alignment of proxy data from sediment cores is essential to accurately transfer information across records recovered from the same sedimentary basin (e.g. Paper II). The alignment can also be applied to records located within a confined region, thereby allowing synchronization of less well-dated records to more robustly dated reconstructions. This relies on the assumption that climate conditions were similar throughout the region and that the resolution of at least one of the records is coarser than the timing required for a climate event to propagate between the core locations (e.g. Lane et al., 2013). In particular, the latter application can be particularly helpful in refining marine chronologies (e.g. Paper IV), which are generally prone to large uncertainties associated with often-unknown regional reservoir age corrections.

The stratigraphic alignment of sediment sequences is commonly pursued through an interactive adjustment of time series, which consists of manually linking user-defined tie points (Björck et al., 2003; Austin et al., 2011). This approach is qualitative and prone to subjectivity, with limited reproducibility. It also implies assuming constant sedimentation rates between tie points, which involves errors that are difficult to assess.

However, over the last two decades a number of deterministic algorithms have been developed to automate the alignment process (e.g. Lisiecki, 2002; Lisiecki and Herbert, 2007; Malinverno, 2013; Lin et al., 2014). Such approaches involve deformation of the entirety of one proxy record onto a reference time series, thus allowing tuning multiple sedimentary records in a more flexible fashion, which accounts for uneven compaction and/or expansion of sediments over time.

The stratigraphic alignment algorithm used in this thesis (Andersson and Muschitiello, in preparation) was largely inspired by the work of Malinverno (2013). The algorithm is driven by a Markov-Hasting MCMC method (similar to that adopted for age modelling), where the algorithm jumps from the current state to the next state based on a stochastic perturbation of the current state introduced at a random position in the depth/age scale of the reference proxy record. If the stochastic perturbation improves the fit between the two proxy time series – as monitored by a probability criterion – the algorithm accepts the new state, where the acceptance criterion is described in equation (2) and (3). The goodness of the fit ($F$) of the perturbed state relative to that of the unperturbed state is computed as:
\[ F = \sqrt{\frac{\sum_{i=1}^{n} (Y_B(i)_{perturbation} - Y_A(i)_{linear})^2}{n}} \]  

(7)

where \( n \) are the number of overlapping data points and \( Y_A(i)_{linear} \) is the linear interpolation of the reference proxy record A, fitting the depth/age scale of the perturbed record B. To account for the finite differences between the reference and fitted proxy time series, an analogous fit, \( fit_{FD} \), is also computed. The fits are assumed to be normally distributed (with standard deviation \( \sigma \)) and the total probability for the new state \( i \) is expressed as:

\[ P(i) = \frac{e^{-fit(i)/2\sigma^2}}{\sqrt{2\pi\sigma^2}} \cdot \frac{e^{-fit_{FD}(i)/2\sigma^2}}{\sqrt{2\pi\sigma^2}} \]  

(8)

where \( \sigma \) is the parameter that defines the expected similarity between the time series.

The MCMC chain is typically run for more than \( 10^6 \) iterations. The algorithm is relatively fast and provides a robust method to find an optimal fit between two proxy time series also accounting for their local variability (e.g. Paper II and Paper IV). A full account of the mathematical formulation associated with the algorithm will be presented elsewhere (Andersson and Muschitiello, in preparation).

4.4. Hydrogen-isotopic composition of lipid biomarker and paleo-hydrological application

At the molecular level, the analysis of organic matter preserved in sedimentary records can provide much more detailed environmental information as compared to bulk sedimentary geochemistry. For instance, specific molecular compounds can be related to particular precursor organisms and/or group of organisms thereby providing insights into the prevailing environmental conditions within definite biological systems and ecological niches (e.g. Meyers and Ishiwatari, 1993; Didyk et al., 1978). This class of molecular compounds is also known as ‘biomarkers’.

In lake sediment studies, biomarker analysis on organic matter from photosynthesizing organisms allows for the separation of aquatic and terrestrial sedimentary components, which enables examining environmental conditions in the lake and the surrounding catchment. As water is the main source of hydrogen for photosynthesizing organisms, the hydrogen-isotopic composition (\( \delta D \)) of sedimentary lipid biomarkers has emerged as a powerful tool in the study of ancient environments and climates (Estep and Hoering, 1980; Sternberg, 1988).

Among the most routinely used lipid biomarkers are the \( n \)-alkane hydrocarbons, which are ubiquitous constituents of biological systems and excellently preserved in sedimentary records spanning a variety of geological time scales (Eglinton and Eglinton, 2008). Both the membranes of algae and aquatic plants, and the cuticular waxes of higher terrestrial plant leaves contain large amounts of \( n \)-alkanes, and specifically short-chain and long-chain \( n \)-alkanes, respectively (e.g. Ficken et al., 2000). The \( \delta D \) values of these aquatic and
terrestrial \( n \)-alkanes are generally offset from, and highly correlated with, the \( \delta D \) composition of the source water used by the precursor organisms following hydrogen-transfer reactions, i.e. intracellular water for algae and aquatic plants, and leaf water for terrestrial plants (Fig. 6) (e.g. Sessions et al., 1999; Sachse et al., 2004). This is the result of a number of environmental parameters (e.g. precipitation amount and source, temperature, relative humidity) and physiological processes involving intracellular water (e.g. leaf physiology, salinity, light intensity, biosynthetic pathway) that control the isotopic fractionation between hydrogen in the source water and in the organic lipids, also known as net or apparent fractionation (Sachse et al., 2012).

Although many of these processes are still a matter of study, the \( \delta D \) composition of lipid biomarkers from lacustrine sediments has rapidly become an established proxy to reconstruct paleo-hydrological conditions and \( \delta D \) of precipitation (e.g. Aichner et al., 2010; Rach et al., 2014). As such, \( \delta D \) values of short-chain \( n \)-alkanes (\( n-C_{17-23} \)) from aquatic algae and submerged plants collected from lake-surface sediments along climatic gradients have revealed a high correlation with lake-water \( \delta D \) values (Fig. 6) (Huang et al., 2004; Sachse et al., 2004, 2006). Analogously, \( \delta D \) values of long-chain \( n \)-alkanes (\( n-C_{27-31} \)) from leaf waxes of terrestrial plants extracted from lake-surface sediments along climatic gradients are highly correlated with precipitation \( \delta D \) values (Fig. 6a, b) (Huang et al., 2004; Sachse et al., 2004; Garcin et al., 2012).

This finding suggests a good preservation of the source water \( \delta D \) signal whereby temporal and spatial (within the catchment) integration may reduce the variability associated with individual biological sources and specific processes. However, even though both aquatic and terrestrial plants undergo isotopic fractionation, which depends on the specific biosynthetic pathway, the net or apparent fractionation of the \( \delta D \) values of terrestrial \( n \)-alkanes is strongly affected by two additional fractionation steps: soil-water evaporation and leaf-water transpiration processes (Fig. 7) (e.g. Sachse et al., 2006, 2012). These parameters, which are controlled by plant anatomy conditions, relative humidity and soil moisture availability, can be framed into a mechanistic Craig-Gordon model (Craig and Gordon, 1965) for open water bodies where leaf-water evaporative enrichment (\( \Delta D_e \)) can be expressed as:

\[
\Delta D_e = \epsilon^+ + \epsilon_k + (\Delta D_v - \epsilon_k) \frac{e_a}{e_i} \tag{9}
\]

\( \epsilon^+ \) is the temperature-dependent liquid-to-vapour equilibrium fractionation at the water-air interface, \( \epsilon_k \) is the kinetic fractionation during diffusion of vapour from the intracellular space in the leaf to the atmosphere, \( \Delta D_v \) is the isotopic enrichment/depletion of vapour in the atmosphere relative to the source water, and \( e_a/e_i \) is the leaf-to-air vapour pressure ratio, which is an expression of relative humidity, leaf temperature and air temperature (Sachse et al., 2012).

Soil-water and leaf-water evapotranspiration are poorly understood owing to the large number of biological unknowns and integration steps associated with the net or apparent fractionation, and due to the lack of experimental culture-based studies. Although an empirical understanding of all the processes behind the net or apparent fractionation in higher plants may help to pave the way for quantitative applications to paleo-hydrological reconstructions, at present, the isotopic difference between terrestrial and aquatic \( n \)-
alkanes can only be employed as a qualitative indicator for reconstructing changes in catchment evapotranspiration (Fig. 7). However, before using lipid biomarker δD values for paleo-hydrological interpretations, it is advisable to characterize the ancient environment using a multi-proxy approach. In particular, vegetation reconstructions based on pollen and macrofossil information can help to factor out or account for possible biological shifts thus allowing disentangling of climatic versus physiological effects on leaf-water δD.

**Figure 6.** Relationship between source-water δD and n-alkane δD values from lake-surface sediments. The relationship is presented for long-chain n-alkanes (n-C29) (a) and short-to-mid-chain n-alkanes (n-C17, n-C23) (b, c). The δD values of n-C23 alkanes of sediments are compared to those from their primary biological source, Potamogeton plants, from the respective lake sediments (data from Aichner et al. (2010)). δD values are given with their ±1σ uncertainty for replicate measurements (lipids and lake water), and errors are calculated from precipitation or obtained from the Atomic Energy Agency GNIP database. Error bars presented in c reflect the ±2σ uncertainty. Modified after Sachse et al. (2012).

**Figure 7.** Conceptual overview of the hydrogen-isotopic relationship between source water and sedi mentary n-alkanes of aquatic and terrestrial plants (not to scale). The red dot illustrates a hypothetical mixture of water pools within the leaf, constituting the ultimate hydrogen source for lipid biosynthesis. εbio, biosynthetic hydrogen-isotopic fractionation; εl/w, isotopic fractionation between lipids and source water. Modified from Sachse et al. (2012).
4.5. Lipid extraction and δD analysis

Lipid extraction was performed on freeze-dried sediment samples (2-8 cm³) via sonication with dichloromethane:methanol (9:1) for 20 minutes, and subsequent centrifugation. This was repeated three times and supernatants were combined at each step. Aliphatic hydrocarbon fractions were isolated from the total lipid extract using silica gel columns (5% deactivated) that were eluted with pure hexane. The saturated hydrocarbon fraction was desulphurized by elution through 10% AgNO₃-impregnated silica gel using pure hexane as eluent. Saturated hydrocarbon fractions were identified based on mass spectra from the literature and retention times. The concentration of individual compounds was estimated based on the comparison of peak areas relative to that of squalane, used as an internal standard added to the samples before lipid extraction.

Isotope ratios \( R \) (\( R = \frac{\text{D}}{\text{H}} \) with \( \text{D} \) or \( \text{D} \) for deuterium and \( \text{H} \) or \( \text{H} \) for protium) are expressed as δD values in per mil (‰), which reflect the relative deviation of \( R \) in the sample from a standard (Vienna Standard Mean Ocean Water with \( \delta \text{D} = 0 \% \)).

\[
\delta \text{D} = \frac{R_{\text{sample}} - R_{\text{standard}}}{R_{\text{standard}}} \times 1000
\]

δD values were determined using a Thermo Finnigan Delta XL mass spectrometer and all analyses were performed in triplicate. A standard mixture of \( n \)-alkanes with known δD composition (mix A4, provided by A. Schimmelmann, Indiana University, USA) was run several times daily to calibrate the measured δD value of a reference gas used to bracket all analyses. Only sample values that were characterised in the isotope-ratio mass spectrometer chromatograms by baseline separated peaks and were of high enough peak size to fall within the linearity range of the instrument were used for data interpretation.

5. Summaries of Paper I-IV

5.1. Paper I - On the timing of environmental shifts across Northern Europe at the Allerød-Younger Dryas transition

The Allerød-Younger Dryas (AL-YD) transition is the last major large-scale climate shift to severe cold conditions before the start of the present warm interglacial. Thanks to the disposal of a large number of terrestrial proxy reconstructions supported by reasonably good chronologies, the AL-YD transition constitutes an excellent workplace to explore leads and lags in response to rapid climate change.

The onset of the YD in Northern Europe is defined by a distinct shift in pollen and macrofossil records from lake sediment sequences. The pollen-stratigraphic boundary has long been used as a common regional chronological constraint (Mangerud et al., 1974; Wohlfarth, 1996; Björck et al., 1998; Lowe et al., 2008) in Northern Europe, since it was assumed that it reflects a rapid and synchronous response of the regional vegetation to the cooling related with the YD. However, at fine chronological resolution, climate records show that even though climate events may have been abrupt at a local scale, they can spread in a time-transgressive fashion over wider geographical scales (Lane et al., 2013).
Similarly, environmental changes do not necessarily respond linearly to climate shifts (Claussen et al., 2013; Rach et al., 2014).

Therefore, if differences in the vegetation response time between sites existed, these would not be fully captured in low-resolution records. Furthermore, prior to the arrival of the latest radiocarbon calibration curve IntCal13 (Reimer et al., 2013), the occurrence of a long radiocarbon age plateau around 13,000-12,800 cal. years BP (Reimer et al., 2009) had made it difficult to reasonably narrow down the age uncertainty of the pollen-stratigraphic AL-YD boundary.

The new radiocarbon calibration curve (Fig. 5a) has significantly improved the accuracy around the aforementioned radiocarbon age plateau, which is now constrained by tree-ring $^{14}$C data (Hua et al., 2009). Thus, the new calibration curve offers the opportunity to re-examine some of the most densely dated North European terrestrial chronologies.

In Paper I, the radiocarbon chronologies of four key sites spanning the AL-YD transition were revisited and compared by consistently constructing new Bayesian age-depth models for each sedimentary record. The chronologies of the four sites – Lake Kråkenes (Birks et al., 2000), Lake Madtjärn (Björck et al., 1996), Lake Gammelmose (Andresen et al., 2000), and Sluggan Bog (Lowe et al., 2004) – are underpinned by a large number of AMS $^{14}$C dates derived from terrestrial plant macrofossils and the local AL-YD pollen stratigraphic boundary is finely constrained in terms of sampling resolution. The age-depth models were produced using two different routines, OxCal (Bronk Ramsey, 2010) and Bacon (Blaauw and Christen, 2011) after calibration with the IntCal13 data set (Reimer et al., 2013). The results show a clear, geographically consistent and diachronous signal of vegetation changes over Northern Europe, with an early AL-YD transition at ∼13,100-12,900 cal. years BP in the British Isles region and Denmark, and a later AL-YD transition at ∼12,750-12,600 cal. years BP in southern Sweden and western Norway (Fig. 8). It is hypothesized that the early transition was associated with regional cooling owing to increased freshwater outflow into the Nordic Seas from the southern margin of the Scandinavian Ice Sheet. By contrast, the second phase was probably brought about by large-scale cooling caused by a widespread climate reorganization associated with a southward diversion of the North Atlantic westerly wind belt (Brauer et al., 2008; Rach et al., 2014).

A potential downside of this study is that the results rely heavily on the original definition of the pollen zone boundary at each site. For instance, new vegetation reconstructions from the site of Hasselø in Denmark suggest that here the AL-YD transition was concomitant with that recorded at the Swedish and Norwegian sites (Mortensen et al., 2015). While these new results disagree with our conclusion of an early AL-YD vegetation shift in Denmark, the age uncertainties that accompany the local pollen zone boundary at Hasselø preclude any conclusive say on this matter.

Chronological assessments of a number of independent temperature reconstructions from British sites point however at an early phase of cooling starting as early as ∼13,100 cal. years BP (Elias and Matthews, 2014). This evidence strongly argues in favour of potentially early vegetation shifts in the British Isles, thus supporting our results and interpretations.

Paper I highlights the importance of establishing robust and coherent chronologies, but also shows some of the limitations of relying on pollen data solely for climate reconstructions.
5.2. Paper II - North Atlantic hydro-climate patterns around the start of the Younger Dryas stadial

A number of studies have shown that hydrological and climate shifts spread rather uniformly across the North Atlantic domain at the onset of the YD cold period (Grafenstein et al., 1999; Birks and Ammann, 2000; Rach et al., 2014; Bartolomé et al., 2015), resulting in similar signs of changes across Europe with respect to those recorded in Greenland ice-core proxies. This has generally been attributed to changes in sea-ice coverage over the North Atlantic, which caused mid-latitude storm tracks to divert across a wide region, thus creating a tele-(connective) mechanism that linked the Greenland and European climate systems (Brauer et al., 2008).

Unsurprisingly, this conclusion has somewhat provided a justification for aligning European climate records to ice-core isotope stratigraphies (e.g. Grafenstein et al., 1999; Bakke et al., 2009; Lang et al., 2010). Such alignments have mainly been applied when available chronologies were not sufficiently reliable to allow for an accurate temporal comparison with events recorded in Greenland ice cores. This however has limited our ability to investigate whether regional atmospheric patterns across the North Atlantic were broadly consistent or not during the last deglaciation.

Paper II addresses some of these issues by studying new well-dated hydro-climate (δD on lipid biomarkers) and temperature (chironomid inferred) data from a lake sediment record from southern Sweden (Hässeldala Port – Fig. 3). The records were examined relative to hydro-climate events recorded in Greenland ice cores. The chronological comparison was also facilitated by the recent synchronization of the ¹⁴C and ice-core time scales using the common cosmogenic isotope variations in tree-ring and ice-core records (Muscheler et al., 2014). Finally, the proxy-based reconstructions were coupled to climate model simulations in order to investigate the ocean and atmosphere parameters responsible for the observed spatial hydro-climate patterns.

The proxy records indicate progressively drier and colder summer conditions in southern Sweden during the few centuries preceding the start of the YD stadial (∼13,100-12,880 cal. years BP) as opposed to wetter and/or warmer conditions observed in Greenland (Fig. 9a). As the δD records are expected to qualitatively track the rate of freshwater discharge from the southern margin of the Scandinavian Ice Sheet to the adjacent North and Norwegian Seas (see section 2) – the main moisture source of precipitation – it is suggested that this period coincides with increased Scandinavian Ice Sheet meltwater forcing in the Nordic Seas.
This hypothesis was tested by using a transient climate simulation performed with a coupled atmosphere-ocean model (Liu et al., 2009; He et al., 2013). Precisely, the regional climate response to a weak freshening (0.011 Sv) of the North and Norwegian Seas was investigated. The freshwater forcing generates a sea-level pressure (SLP) dipole across the North Atlantic, with relatively higher SLP over Northern Europe and lower SLP over Greenland (Fig. 9b). In the model, the dipole is a direct expression of increased sea-ice cover in the Norwegian and Barents Seas resulting from the freshwater input.

This physical mechanism is capable to fully explain the divergent hydro-climate and temperature shifts recorded shortly prior to the onset of the YD stadial both in southern Sweden and Greenland. Interestingly, the hydro-climate dipole in the model is bound to the presence of freshwater anomalies in the eastern sector of the Nordic Seas, which are necessary to account for the shifts observed in the reconstructions. In fact, the dipole is not simulated when freshwater is released from North American sources. Moreover, the model results compellingly support the interpretation of the δD records in terms of regional meltwater forcing.

In addition, these results imply that aligning North European records to Greenland climate signals is not a viable option if we are to understand leads-lags and spatial patterns of climate response to freshwater forcing.

In conclusion, this study highlights a previously unrecognized sensitivity of North Atlantic hydro-climate to Scandinavian Ice Sheet meltwater forcing. More importantly, it shows that Scandinavian Ice Sheet meltwater discharge to the Nordic Seas can control the timing and signs of the isotopic shifts registered in Greenland ice cores shortly prior to the YD.
stadiial. This is a potentially valid mechanism to explain the early vegetation shifts and drop in temperatures observed in the British Isles and outlined in Paper I.

5.3. Paper III - Glacial varve evidence for a catastrophic outburst of meltwater synchronous with the onset of the Younger Dryas stadial

Since the late 80’s, when Wallace Broecker and colleagues (Broecker et al., 1989) hypothesized that a catastrophic meltwater pulse from the North American continent was the main triggering mechanism for the onset of the YD, the search for the flood pathway has been one of the most controversial topics in paleoclimate sciences.

Yet, despite that the drainage or flood hypothesis is at present the classical explanation for the start of the YD, the reconstructed timing of the catastrophic meltwater outburst from the Laurentide Ice Sheet is still a matter of debate. Reconstructed ages for this event (Lowell et al., 2005; Fisher et al., 2009; Murton et al., 2010; Not and Hillaire-Marcel, 2012; Breckenridge, 2015) are systematically too young or too uncertain with respect to the hydro-climate shifts that mark the onset of the YD stadial as observed in Greenland ice-core records, where it is referred to as Greenland Stadial 1 (GS-1; 12,846 ± 69 ice years at 1σ – Rasmussen et al., 2006). This casts some doubts on the causal relationship between the drainage of proglacial lakes in North America and major shifts in atmospheric and ocean circulation at the onset of the YD stadial.

The Scandinavian Ice Sheet, on the other side of the North Atlantic Ocean, is perhaps one of the most overlooked drivers of deglacial climate change. During the Late AL (~13,000 cal. years BP), the ice front was located south of the south central Swedish lowland area (Björck, 1995; Lundqvist and Wohlfarth, 2000; Hughes et al., 2015) and the Baltic Ice Lake (Fig. 3) was dammed up. Rapid recession of the ice margin during this period at the water divide near Mt. Billingen (Fig. 3) generated a spillway system that connected the Baltic Ice Lake to the sea in the west, which resulted in an abrupt 5-10 m lowering of the Baltic Ice Lake (Björck, 1995; Björck et al., 1996).

The Baltic Ice Lake drainage has long been a contentious issue. However, new evidence now confirms that a catastrophic outflow of meltwater actually took place near Mt. Billingen (Swärd et al., 2015). However, these reconstructions lack a precise chronology that allows pinning down the exact age of this event.

Paper III constrains the timing of the Late AL drainage of the Baltic Ice Lake by re-evaluating an annually resolved glacial varve chronology from southeastern Sweden (Wohlfarth et al., 1998). The new composite 1257-year long varve chronology (~13,200-12,00 cal. years BP) is based on 57 records and provides insights into the timing of ice recession and depositional events within the Baltic Ice Lake. In addition, the chronology was placed on an absolute time scale using the Vedde Ash volcanic marker and new 14C age modelling. This allowed comparing for the first time the melting history of the Scandinavian Ice Sheet to the Greenland ice-core and radiocarbon time scales (Rasmussen et al., 2006; Reimer et al., 2013) with high accuracy and resolution.

Geochemical and sedimentological analyses of the glacial varve records indicates a rapid change in sedimentation regime and a long-lasting disappearance of ice-rafted debris in the Baltic Ice Lake, respectively, which coincided with the start of GS-1. This depositional event took place at 12,847 ± 2 years (1σ) on the ice-core time scale and at 12,876 ± 22 cal. years (1σ) on the IntCal13 time scale. The event occurred 726 ± 2 years after the deposition of the Vedde Ash as compared to 725 ± 6 years for the start of GS-1 in ice-core
records. A simplified ice-sheet model indicates that the change in sedimentation regime and especially the drop in ice-rafted debris transport can be explained by ice-margin stabilization in response to a large drop in the Baltic Ice Lake water level.

Figure 10. Radiocarbon calibrated age of the first drainage of the Baltic Ice Lake inferred from a wiggle-matching model underpinning the glacial varve chronology from Östergötland (red). The related probability is compared to the calibrated age of all available $^{14}$C dates that indirectly constrain this event. Dates from Blekinge and Arkona Basin refer to the timing of a major lowering of the Baltic Ice Lake. Dates from Hunneberg constrain the timing when the spillway at Mt. Billing became ice free. The blue bar indicates the age probability for the start of Greenland Stadial 1 (GS-1) on the IntCal13 time scale. A list of references to the respective dates is presented in Paper II.

Results from a systematic re-calibration of all available $^{14}$C dates that constrain the timing of deglaciation at the Mt. Billing outlet and the related Baltic Ice Lake water level drop (Muschitiello et al., 2015b, 2015c) argue in favour of our hypothesis (Fig. 10). A mechanism, only hinted at in Paper II, is also proposed, for which a catastrophic outflow of meltwater may have induced excess sea ice in the Norwegian and Barents Seas, which recirculated into the subpolar North Atlantic gyre. It is suggested that sea-ice recirculation in the Nordic Seas can cause the coupled atmosphere-ocean system to cross thresholds beyond which a stadial climate regime is triggered. Critically, this is a robust feature in Earth-System and General Circulation models (Fig. 11) (Drijfhout et al., 2013; Lehner et al., 2013) and thus provides a plausible physical mechanism for the inception of sustained cold climate events of the past.

5.4. Paper IV – Deglacial AMOC and meltwater forcing in the Nordic Seas: the ocean perspective

The common explanation for the inception of sustained cold stadial periods during the last deglaciation involves an abrupt weakening of the AMOC via freshwater forcing in the North Atlantic Ocean (McManus et al., 2004; Praetorius et al., 2008; Clark et al., 2012). However, paleoceanographic evidence for these abrupt ocean circulation changes and their relationship with freshwater forcing remains elusive. Moreover, the representation of freshwater sources, timing and magnitude varies between climate models (cf. Zhang et al., 2014), as does the sensitivity of the simulated AMOC to freshwater perturbations (Fig. 12) (e.g. Rahmstorf et al., 2005).

Among the most controversial issues that arise when simulating past climate change using proxy-based reconstructions, is that climate models require unrealistically large freshwater perturbations and that the longevity of the simulated stadial is entirely dependent upon the duration of the applied freshwater forcing (Ganopolski and Rahmstorf, 2001; Knutti et al., 2004). This implies that the stability of the AMOC is
systematically overestimated in climate models (Hofmann and Rahmstorf, 2009), precluding the comprehension of possible bi-stable regimes of the overturning circulation system.

**Figure 11.** Abrupt southward progression of the North Atlantic sea-ice margin in response to enhanced sea-ice production in the Barents as simulated with the EC-Earth climate model under Pre-Industrial boundary conditions. 

- **a**. Averaged annual sea-ice anomalies for model years 430-450 relative to the climatology for the years 200-400 (just prior to the cold event).
- **b**, as in (a) but for the averaged annual anomaly for the years 450-550, during the peak of the cold event. Values are expressed as a fraction of 1.
- **c**. Atlantic Meridional Overturning Circulation time series in sverdrups (Sv= 10⁶ m² s⁻¹). The black line shows the maximum overturning and the red line the overturning strength at 36°N at deeper depth (1600 m). The colored areas show the periods of integration for (a) and (b) as well as the reference period used to estimate the anomalies. Modified after Drijfhout et al. (2013).

**Figure 12.** Sensitivity of the North Atlantic thermo-haline circulation to freshwater forcing. The panel shows hysteresis curves found in coupled 3-D global ocean models. Circles indicate the present-day climate state of each model. Modified from Rahmstorf et al. (2005).
Therefore, little is known on the transient behaviour of the AMOC and its true sensitivity to freshwater forcing during the last deglaciation. Furthermore, due to large uncertainties with the marine reservoir effect, it is still unclear whether shifts to a weak-state of the AMOC are the trigger or a mere response to changes associated with other components of the climate system, in first instance sea ice and atmospheric circulation.

In Paper IV, the deglacial history of the upper limb of the AMOC – the North Atlantic inflow to the Nordic Seas – is reconstructed using SST and δ¹⁸O records from marine core MD99-2284 (Fig. 3, 13), which is located at the gateway for transport of oceanic heat flux to northern latitudes. By synchronizing the upwind SST signal to downwind hydro-climate records (δD on lipid biomarkers) from the terrestrial site of Atteköps Mosse, it was possible to provide core MD99-2284 with a precise atmospheric-based ¹⁴C chronology. Moreover, we infer meltwater discharge from the Scandinavian Ice Sheet to the adjacent Nordic Seas by reconstructing the regional evolution of the marine ¹⁴C reservoir age (Fig. 13), a proxy indicating the contribution of continental freshwater containing dissolved inorganic carbon with low ¹⁴C activity.

The reconstructions indicate a substantially unaltered strength of the North Atlantic Inflow and AMOC throughout the warm interstadial phase (∼14,700-12,900 cal. years BP). This is surprising as the marine records suggest a significant and variable outflow of freshwater from the Scandinavian Ice Sheet (Fig. 13), with a total ice-melt discharge estimated at ∼2.8 ± 0.3 m sea-level equivalent (Hughes et al., 2015).

By contrast, synchronously with the start of GS-1, the inflow of saline subtropical waters critically decreases together with a major weakening of the AMOC (Fig. 13). This implies a tight coupling between ocean and atmospheric perturbations within the North Atlantic system.

The previous study (Paper III) provides evidence for a catastrophic drainage of meltwater from the Scandinavian Ice Sheet and a plausible physical mechanism involving injection of extra sea ice into the subpolar gyre. In the light of the apparent inertia of the AMOC system to long-term freshwater fluxes, it is likely that the abrupt shift to a weak AMOC mode at the onset of GS-1 was the result of sea-ice–wind feedbacks rather than changes in buoyancy forcing. In conclusion, despite sudden meltwater pulses from ice-dammed continental lakes deliver only a fraction of freshwater to the oceans as compared to the amount that decaying ice sheets can release over millennia altogether, the associated feedbacks have a crucial impact on the AMOC mean state and can likely push the system over its tipping point. This new framework for understanding rapid climate mode shifts is a critical benchmark for designing future climate model experiments.

### 6. Terrestrial-marine proxy comparison

To evaluate the regional significance of the hydro-climate records generated within this thesis project, the δD values derived from the aquatic components (δD_{aq}) of lake sediments from Hässeldala Port (HÄ) and Atteköps Mosse (ATK) are here compared to each other and other regional marine records.

In Paper II and III it was argued that, at least prior to the onset of the Holocene, the ΔδD_{terrestrial-aquatic} values from HÄ and ATK records were mainly dependent upon the local hydrological conditions rather than upon changes in local vegetation. Across the key climatic transitions discussed in this thesis, e.g. at the onset of AL and YD, the ΔδD_{terrestrial-
Figure 13. a, Sea-surface temperatures (SST) and b, near-pycnocline ice-volume corrected seawater δ¹⁸O from marine core MD99-2284 compared to Atlantic Meridional Overturning Circulation proxy reconstruction (McManus et al., 2004). Shadings reflect the 68% CI based on both analytical and chronological errors. c, Reconstructed North Atlantic surface ocean reservoir ¹⁴C ages (ΔR). White dots refer to re-evaluated ¹⁴C data from the western coast of Norway (Bondevik et al., 2006) and red dots refer to ¹⁴C data obtained from marine core MD99-2284. The shading reflects uncertainties in the reconstruction based on dating and measurement errors (95% CI). Darker shading indicates more likely ΔR values. The mean is shown as a red line. The white arrow shows the present-day ΔR (Bondevik et al., 2006). d, Varve- (red) and radiocarbon-based (blue) age estimates (±1σ) for the drainage of the Baltic Ice Lake (Muschitiello et al., 2015b, 2015c). e, Volume evolution of the Eurasian ice sheets expressed in m sea-level equivalent (Hughes et al., 2015). SIS, Scandinavian Ice Sheet; SBKIS, Svalbard, Barents and Kara Sea Ice Sheet; BIIS, British-Irish Ice Sheet. Greenland stratigraphic events are displayed on the IntCal13 time scale. Colored bars show the two major phases of surface-water cooling in the Norwegian Sea.
aquatic records indicate hydrological shifts towards relatively drier conditions with respect to the preceding period. These hydrological shifts would have enhanced evaporative deuterium enrichment of lake-water and thus of $\delta D_{aq}$. However, the $\delta D_{aq}$ records from HÅ and ATK show large isotopic shifts towards more negative values at the start of both AL and YD, and YD, respectively. This implies that local hydrological processes operated in the opposite direction of the observed shifts in $\delta D_{aq}$ and were likely not a primary factor in controlling the $\delta D$ composition of lake-water. Therefore, the $\delta D_{aq}$ records should reflect, to a large extent, changes in the isotopic composition of the precipitation source, which are primarily related to the surface hydrography of the eastern sector of the Nordic Seas (see section 3.1).

The $\delta D_{aq}$ signals, which only overlap for $\sim2,000$ years (~14,000-12,000 cal. years BP), display a systematic offset but also some differences, with HÅ showing a larger negative shift in $\delta D_{aq}$ values at the transition into, and during the first half of the YD relative to ATK (Fig. 14).

The $\delta D_{aq}$ values at HÅ are $\sim30$-$70\%$ lower than those at ATK ($\sim60$-$100\%$ when compared with HÅ’s $\delta D$ record from C$_{21}$ alkanes). The systematic $\delta D$ offset between the two records may arise from three possible reasons or from a combination of these: the Rayleigh rainout effect (Gat, 1996); differences in the $\delta D$ composition of the moisture source contributing precipitation to each site; differences in the length of the thawing season between the eastern and western coast. In southern Sweden, the first case is a common phenomenon, whereby the isotopic ratios of precipitation decrease as the air masses move eastwards from the west coast (Jonsson et al., 2010). For instance, instrumental data show that the present-day annual difference in $\delta D$ values of precipitation across Northern Europe (e.g. from the British Isles to the Baltic countries) ranges between $-65$ and $-110\%$ (Bowen, 2003).

As to the second reason, it is plausible that a portion of the precipitation delivered to HÅ – especially in summer when the circulation was more anticyclonic (Muschitiello et al., 2015c) – was inherited from the leeward side of the Scandinavian Ice Sheet. Here, the Baltic Ice Lake constituted a moisture source characterised by heavily deuterium-depleted meltwater as opposed to the North Sea seawater on the windward side of the ice sheet. This could explain systematically lower $\delta D$ values of precipitation integrated in HÅ’s sediments with respect to ATK. However, the isotopic signal incorporated by the aquatic vegetation reflects the $\delta D$ composition of the lake-water during the growing season, i.e. late spring/early summer (Sachse et al., 2004). It was demonstrated that the early summer isotope signatures of lake-water in southern Sweden are strongly affected by the replenishment of the local aquifer by meltwater associated with winter and spring snowfall (Muschitiello et al., 2013).

It is therefore likely that $\delta D$ values recorded in HÅ’s sediments are reminiscent of the isotopic composition of the precipitation carried by the dominant westerly winds during the cold season, when the circulation was zonal. Altogether, this would mean that the $\delta D$ offset observed between ATK and HÅ during the last deglaciation is probably more related to the distillation of moisture transported across southern Sweden from the sea to the inland rather than to a signal from eastern sources.

Another possible scenario involves relatively shorter summers at HÅ relative to ATK. In fact, here the proximity to the cold Baltic Ice Lake water body may have led to a longer
Figure 14. a, Schematic upper circulation of the North Sea and location of marine coring sites JM99-1200 (Ebbesen and Hald, 2004), MD99-2284 (this thesis), HM79-6/4 (Karpuz and Jansen, 1992), and terrestrial sites Atteköps Mosse, and Hässeldala Port (this thesis). b, Comparison between terrestrial δD records of precipitation, annual sea-surface temperature (SST) and spring sea-ice reconstructions from the Norwegian Sea. The data set of marine core MD99-2284 was synchronized to Atteköp’s time scale as explained in Paper IV. The age models of core JM99-1200 (Ebbesen and Hald, 2004) and HM79-6/4 (Karpuz and Jansen, 1992) were established using the available ¹⁴C dates corrected for variations in regional reservoir age as reconstructed in Paper IV. (Caption continues on page 29)
thawing season and colder summers as compared to the western coast of Sweden. This may have postponed and lengthened the inflow of deuterium-depleted snowmelt in the lake discussed above, thereby resulting in relatively lower $\delta_{Daq}$ signature at HÅ. Such hypothesis is supported by chironomid-inferred summer temperature records, which indicate a minimum of 1-2 °C colder summers at HÅ (Muschitiello et al., 2015c) than at ATK (not shown) during ~14,000-12,000 cal. years BP. In particular, variable summer temperature differences could also provide an explanation for the observed transient discrepancies in $\delta D$ between the two sites.

Further clues can also be drawn by comparing the $\delta D_{aq}$ records to other regional marine proxy reconstructions, since we expect the hydro-climate data sets to also depend on the sea surface temperature conditions at the marine source of precipitation. In the following, three precisely dated marine sediment records from the Norwegian Sea are introduced (Fig. 14a, b): core JM99-1200 (Ebbesen and Hald, 2004), core MD99-2284 (Paper IV), and core HM79-6/4 (Karpuz and Jansen, 1992).

The chronology of MD99-2284 is based on the synchronization to ATK’s $\delta D$ records and it has been discussed in Paper IV. The radiocarbon chronology of JM99-1200 and HM79-6/4 have here been revisited by using a Bayesian age-depth model (Bacon), the Marine13 calibration curve, and by applying marine reservoir correction factors according to the reconstruction presented in Paper IV. The SST reconstructions from core JM99-1200 and MD99-2284 are based on foraminifera assemblages and indicates sub-surface conditions. By contrast, the SST reconstruction from HM79-6/4 is based on diatom assemblages that yield temperature conditions at a shallower depth (in the photic zone). The core JM99-1200 has also been studied for biomarker-based reconstructions of sea ice conditions (Cabedo-Sanz et al., 2012).

A interesting feature that arises from the comparison between the terrestrial and marine records is that the $\delta D$ record from ATK tracks the sub-surface water temperature signal reasonably well (Fig. 14b), which provided the foundations for the synchronization discussed in Paper IV. By contrast, the shifts in $\delta D$ values from HÅ sediments appear to be in better agreement with variations in seasonal sea-ice cover and surface water temperatures in the Norwegian Sea at the start and throughout the YD (Fig. 14b). Especially, the pre-YD cooling starting at ~13,000 cal. years BP, which has been identified in the HÅ isotopic records (Paper II), is evident in the marine records, whereby a shift towards near-permanent sea-ice conditions (Cabedo-Sanz et al., 2012) and colder SSTs are initiated a few centuries earlier than the start of GS-1 (Fig. 14b).

The question arises as to why hydro-climate proxies from ATK are more sensitive to sea sub-surface conditions, while records from HÅ respond to surface dynamics. During the deglaciation, ATK was located close to the coast at the head of a long fjord system, which is today known as the Skagerrak-Kattegat (Fig. 3). Modern high-latitude fjords, such as in Greenland or in Scandinavia, are highly stratified, with warm subtropical waters flowing beneath a shallow brackish or freshwater layer (Stigebrandt, 1981; Straneo et al., 2010). The surface waters are constantly replenished throughout the fjords and at the head, the deeper seawater can be entrained in the shallow fresher layer owing to the turbulence induced by the action of wind and by runoff from land (Stigebrandt, 1981). This can result in strong mixing and upwelling of sub-surface waters at the head of the fjord.

The presence of the Scandinavian Ice Sheet during the last deglaciation would have promoted the occurrence of descending katabatic winds, sweeping off sea ice, icebergs and ultimately freshwater from the northern coasts of the Skagerrak. Moreover, a stronger
anticyclonic circulation regime over the ice sheet during summer (Muschitiello et al., 2015c) may have been more conducive to steer meltwater westwards, out from the Skagerrak-Kattegat complex, into the Norwegian Sea and eventually northwards (Fig. 14a). In particular, this paleo-hydrographic flow associated with a stronger summer meridional circulation is consistent with the spatial pattern recorded with instrumental data (Fig. 4b).

Such a pattern of circulation within the fjord, with fresh outflow at the surface balanced by saltier, sub-surface inflow and vertical mixing at the head, can explain why the δD signatures of precipitation at ATK track the sub-surface water temperature. Bearing in mind that the δD signal at ATK is broadly consistent with other independent lacustrine δ^{18}O records from the south-western coast of Sweden (Hammarlund and Keen, 1994; Hammarlund and Lemdahl, 1994), this implies that the marine moisture source for ATK and the surrounding area was predominantly associated with the head of the Skagerrak-Kattegat fjord. Conversely at HÅ, further to the east, the δD signatures of precipitation probably integrated a sea surface signal over a relatively wider region.

The mechanism involving the integration of vertical physical properties of different seawater masses by precipitation in two independent terrestrial records would not just provide an additional explanation for the transient differences between δD values at ATK and HÅ. It could also provide a further explanation for the observed offset in δD values of precipitation discussed above. Since the moisture carried to ATK was inherited from relatively closer marine waters (i.e. relatively warmer and more saline), the associated isotopic signatures were more enriched in D as compared to HÅ, where the moisture originated from a broader and overall fresher source.

Local versus regional hydro-climate sensitivity between the two Swedish sites would also clarify why reconstructed summer temperatures at ATK (chironomid-inferred; not shown) do not agree with the record from HÅ (Muschitiello et al., 2015c) or other North European deglacial temperature records (Heiri et al., 2007; Elias and Matthews, 2014). In contrast, the temperature record from HÅ (Muschitiello et al., 2015c) is in good agreement with the summer temperature history in the British Isles (Elias and Matthews, 2014), thus supporting the regional significance of the climate records generated at this latter site.

Figure 14 (continued). The SST records from core JM99-1200 and MD99-2284 are based on foraminifera assemblage counts and reflect sub-surface conditions, whereas the record from core HM79-6/4 is based on diatoms and reflects surface conditions. Note the general agreement between the two sub-surface temperature records. The sea-ice reconstruction (Cabedo-Sanz et al., 2012) was obtained from the same sedimentary record of core JM99-1200 and is based on the P_{B10}P_{25} index, defined as the abundance of the biomarker brassicasterol versus IP_{25}. Greenland stratigraphic events according to the IntCal13 time scale (Muscheler et al., 2014) are also displayed. Isotope and SST records are presented with bars indicating ±1σ error associated with both analytical and chronological uncertainty. The age uncertainties of the sea-ice record are the same as for the SST reconstruction of core JM99-1200. The ages for the Vedde Ash and Saksumarvatn tephra used to construct the chronologies were based on age-modeling results from Lohne et al. (2013). The interval characterised by frequent sea-ice cover and low SSTs in the Norwegian Sea is highlighted in grey.
In conclusion, this comparative analysis brings up the complexity of interpreting δD paleo-records as a function of hydro-climate factors and highlights the need to couple empirical reconstructions with isotope-enabled climate models that can help accounting for shifts in the precipitation moisture source.

7. Current work and unpublished data

7.1. Impact of the Scandinavian Ice Sheet on regional climate using a spatially high-resolution climate model

The on-going work related to the research presented in this thesis focuses on the use of spatially high-resolution climate model simulations (using CESM1.0.5) to investigate YD summer climate over Europe (Schenk et al., in preparation). The model simulations are employed to understand the importance of northward heat transport and ocean-to-atmosphere heat flux over the North Atlantic under a weak AMOC regime, such as during the YD (McManus et al., 2004). The main motivation of this study is to shed light on the elusive regional climate impact of a cold North Atlantic Ocean during a period with high and increasing insolation forcing (Fig. 1). Moreover, the simulations, which include new data-calibrated ice-sheet model reconstructions, are being compared to an extensive European data set of proxy-based quantitative temperature records (chironomids, aquatic pollen, terrestrial plant macrofossils).

Preliminary results (Schenk et al., in preparation) suggest that, conversely to earlier coarse resolution climate simulations of the YD (e.g. Renssen et al., 2015), the competition between a cold ocean and high orbital forcing results in warmer summer conditions over Eurasia, with the exception of coastal and high elevation sites (Fig. 15a). In spite of 10 °C colder SSTs in the North Atlantic, the presence of the Scandinavian Ice Sheet significantly impacts the regional atmospheric flow preventing cold westerly winds from the Atlantic to penetrate inland, resulting in northerly flow over the Nordic Seas and increased blocking circulation over the ice sheet. In turn, this circulation pattern induces a high radiative balance of surface energy fluxes, thus explaining the summer warming in continental Europe, whereas the cold ocean has a greater influence along the coastal regions (Fig. 15a).

The plant macrofossil-based temperature reconstructions from the proxy compilation (Minna Väliranta and Maija Heikkilä – University of Helsinki) are broadly consistent with the pattern of warming observed in the simulations (Schenk et al., in preparation). However, chironomid-based reconstructions show summer cooling during the YD. The simulations suggest that this can be explained by a progressive shortening of the growing season in summer (Fig. 15b). In particular, a longer and colder spring can postpone melting of lake ice, which in turn cools lakes even during summer. Hence, this snowmelt process may have important implications for the interpretation of lake-sediment biological proxies.

7.2. Sensitivity of the Scandinavian ice Sheet to volcanic forcing

Further on-going work surrounds the comparison of the new annual glacial-varve chronology presented in Paper III with Greenland glaciochemical records at the end of the last deglaciation. The glacial varve chronology (Muschitiello et al., 2015b), which has been synchronized to the Greenland Ice Core Chronology 2005 (Rasmussen et al., 2006) via the common Vedde Ash time marker, offers for the first time the opportunity to compare...
melting rate variations of the Scandinavian Ice Sheet to ice-core volcanic records at annual resolution.

The comparison shows that years characterised by anomalous Scandinavian Ice Sheet melting coincide in time with volcanic eruptions as recorded in ice-core aerosol loading records (Fig. 16) (Zielinski et al., 1996). By using output from climate model simulations (Jungclaus et al., 2010), it is shown that explosive volcanic eruptions can generate an instantaneous climate response in the North Atlantic, which results in a substantial decrease in seasonal precipitation.

These results may suggest that ice-sheet melt anomalies identified in the varve record are potentially a result of snow-albedo feedbacks that lowered the reflectance of bare ice under reduced snow accumulation conditions, a mechanism particularly efficient in driving ice-mass loss in modern glaciers and ice sheets (Francou et al., 2003; Van Tricht et al., 2016).

This analysis may provide the first evidence for the sensitivity of continental ice masses to volcanic forcing during ice age terminations. This has important implications with respect to the tremendous amount of meltwater trapped by recessing ice sheets and its pivotal role on rapid climate change.

7.3. Unpublished data sets

In addition to the proxy records and model simulations published in Paper II and IV and presented in this report, there is a large number of unaccounted data and model output behind this project. In the following I outline some of the unpublished proxy data sets that were generated during my PhD.

Hässeldala’s sediments were thoroughly investigated for isotope and biomarker analysis. Unpublished data comprise δD records from C20 Highly Branched Isoprenoids, δ18O on cellulose, d-excess, and δ13C on a suite of n-alkanes.

Unpublished data from Atteköp’s sediments comprise X-ray fluorescence data, δD records from C20 Highly Branched Isoprenoids and chironomid-based temperature records (investigator: Tomi P. Luoto – University of Helsinki) and plant macrofossil-based temperature records (Minna Väliranta – University of Helsinki).

8. Future work

Future work will primarily focus on a set of sensitivity climate experiments using the CCSM3 model (Frederik Schenk – Stockholm University). The simulations will aim at a better understanding of physical processes behind the occurrence of an “early cooling” in Northern Europe and a hydro-climate dipole across the North Atlantic during the Late AL, as observed in Paper I and II of this thesis, respectively.

The sensitivity experiments will be designed by prescribing different amounts and rates of freshwater forcing from the Laurentide Ice Sheet and Scandinavian Ice Sheet. Additional sensitivity experiments will be run to better examine the impact of sea ice anomalies in the Norwegian Sea and Barents Sea during the same period (Francesco Pausata – Stockholm University).

Time slice experiments (12,000 versus 13,000 years BP) will also be run with the ECHAMiso model (Jesper Sjolte – Lund University) to allow for a isotope proxy-model
Figure 15.  

**a**, Modeled July surface air temperature anomaly between the Younger Drays stadial and the preceding warm Allerød interstadial (12,000 minus 13,000 model years BP).  

**b**, Same as (a) with summer temperature anomalies from proxy data overlain.  

**c**, Same as (a) for Growing Season Length. Positive (negative) values in (c) indicate a longer (shorter) growing season of land vegetation. The start (end) of the growing season is defined as the period characterized by six straight days with a temperature above (below) 5.5° C (Nemani et al., 2003). Significance levels are indicated by black stippling (95% CI). Data from Schenk et al. (in preparation).

comparison at the transition into the YD stadial.

Finally, using the state-of-the-art, high resolution (1/6°, ~18 km), coupled ocean sea-ice circulation model MITgcm (Alan Condron – Massachusetts Institutes of Technology), a set of transient simulations will be performed to resolve the circulation of the ocean and sea ice associated with the drainage of the Baltic Ice Lake at the onset of the YD stadial. These
simulations will shed light on the trajectories of meltwater at a resolution 10-15 times higher than other GCMs and help to understand the role of the Baltic Ice Lake drainage on AMOC stability.

Future proxy analyses will focus on generating new isotope records from southern Scandinavia that allow extending the terrestrial-marine synchronization discussed in Paper IV and thus the chronology of marine core MD99-2284 into the Holocene. If this attempt turns out to be successful, by dating new \(^{14}\text{C}\) sample from core MD99-2284, it will be possible to extend the North Atlantic marine reservoir age record into the Early Holocene.

**Figure 16.** Comparison between the new Late Glacial varve-clay chronology from Östergötland and Greenland ice-core volcanic records. **a,** Varve thickness standardized anomalies of the portion of the varve chronology composed of 55 overlapping varve diagrams (Muschitiello et al., 2015b). **b,** Volcanic signal recorded in GISP2 (\(\text{SO}_4^{2-}\)) and NGRIP (\(\text{H}^+\)) ice cores. Volcanic aerosol sulphates deposited at the GISP2 site are presented both as absolute values (orange) and as flux (red). The GISP2 record was synchronized to the Greenland Ice Core Chronology 2005 time scale via common volcanic markers (Rasmussen et al., 2007, 2008). **c,** Results from Monte Carlo significance tests of synchronicity between exceptionally thick varve years and volcanic events. In the left-hand panel synchronicity was tested using 1,000 permutations of the varve thickness anomalies. In the right-hand panel synchronicity was tested using 1,000 individual realizations of the varve thickness record with similar red noise spectral characteristics. The green area indicates the region above the 95\(^{\text{th}}\) percentile and the red stars indicate the estimated agreement (%) between varve anomalies and volcanic events. Exceptionally thick varve years corresponding to anomalies in atmospheric sulphate loading and the related volcanic event are also indicated. The 'b2k' convention of the Greenland Ice Core Chronology 2005 is here converted into BP (1950 years AD).
Acknowledgements

Funding for this project was provided by the Swedish Nuclear Waste Management Company (SKB). Additional funding was provided for various activities by the Bolin Centre for Climate Research and the Department of Geological Sciences, Stockholm University. I would also like to acknowledge financial support through INTIMATE cost-action.

Along with institutions, there are people. First and foremost, I would like to thank my team of advisors. Barbara, like the last deglaciation, our relationship has been a bumpy road with a lot of ups and downs. I am aware I haven’t been the easiest student to deal with and due to my temper and my over-enthusiasm I caused you, to put it mildly, quite some annoyance. Still, you always remained professional, supported me, advised me wisely, and eventually even indulged to some of my wacky ideas. You have also been my harshest critic and reviewer. I believe this has made me a more thorough researcher and certainly a better writer. Now I know that if I can sell you an idea, well, then I can sell it to anyone. I will always be grateful to you for letting me play with the mud in your sandbox, which has made these last four years one of the most fantastic experience of my life.

Thank you Rienk for passing the “art” forward to me, for the passionate training and discussions inside and outside the laboratory rooms. Isotopes rule and will always do!

A great thanks to my parents for their love and the ceaseless support throughout my eleven long years as a university student. Mom, I promise I’ll find a job now.

I would also like to thank another very special group of people: the Saarikoskis. Sure enough, you are like a second family to me. Thank you for all the love (and the awesome breakfasts).

Thanks to all my friends. I apologize I cannot name you all here one by one. However, I must mention some leading figures. Robert, I could have not possibly accomplished this PhD without you. You have been a great source of motivation for my work. Thank you for all the English synonyms, the “sandwiches”, the endless party nights and all the (mis)adventures we have been through together. Please, forgive me for the noise in Norway. Thank you James for never declining an invitation to the pub. With you I had some of the most exciting (and productive) science talks in front of a pint that I’ve ever had. Cat, Lisa, Patrik, you have been true friends and definitely a constant that have accompanied me throughout this journey. Thank you for all the fun you brought.

Among my colleagues, there are a number of awesome characters that I would like to acknowledge. In no particular order: the hard-working and relentless guys from the fourth floor, Hugsy, Reuby, Cliff, little Alex, and Barbarella; my officemates Natalia, Pedro, Francis, Liselott, and the beautiful princess Moo; and of course the silica friends Wim and Patrick.

Outside the department I have also been fortunate to make some great friends who have reminded me that life is not just about paleo-problems. Thank you Justine, Elissa, Matilda, Friedman, Nico, Jan, Riccard, Josef and Marco.

On a less human note, I must thank a number of anthropomorphic deities for the constant presence when things went wrong, as well as sushi and nachos for the necessary boost of nutrients that helped me to deliver the job.

To these folks and the rest of the departmental staff: this time would not have been the same without you. A special thank goes to Jane, Carina, Anna and Heike for the help in getting my laboratory work moving day after day. Thank you for coping with an annoying and at times cranky Italian who does not always feel comfortable in a lab coat.
I would like to thank Svante Björck for the inspiration and fruitful discussions through the last years (and also for writing my Swedish abstract!). A huge thanks to Eve for always having an open door and a good piece of advice. I’m also very grateful to Martin Jakobsson, who granted a young and unskilled limnologist with the opportunity to experience an unforgettable research cruise across the Arctic Ocean.

Last but certainly not least, Pilvi, my most treasured joy. Thank you for everything.
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