



An overview of Holocene climate reconstructions in northernmost Fennoscandia

A contribution to the *Stone Age Demographics* project

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Abstract

An overview of climate reconstructions considering summer air temperatures and effective precipitation is provided for northernmost Fennoscandia. During the earliest part of the Holocene (11,700–10,000 cal. BP), temperatures rose rapidly and were followed by mild, wet and variable conditions. An early major warming peaked around 9500 cal. BP, although many records indicate that the main Holocene warming first occurred about c. 8000 cal. BP. The sub-regional pattern of climate change suggests a defining influence of the westerlies and the North Cape Current. Non-analog climatic conditions and lags in vegetation responses to climate change may explain some of the discrepancies seen in the early Holocene between proxies. In contrast to the perceivable variable onset of the main Holocene warm period, maximum temperatures are relatively consistent between the records, indicating temperatures $1.5 \pm 0.5^\circ\text{C}$ above present. Precipitation was generally high from 10,000 cal. BP but decreased towards 8000 cal. BP when dry climatic conditions became predominant. After a stable period 8000–6000 cal. BP a gradual cooling was initiated, with a more abrupt period of change 4500–3800 cal. BP when the warm and dry climate of the mid-Holocene changed into the cool, wet and unstable climate of the late Holocene. Modern conditions were reached c. 2800 cal. BP. The Holocene Thermal Maximum may be defined several different ways: as temperatures distinctly above modern delimited to 9500–4000 cal. BP; as peak temperatures 9500–6000 cal. BP; and/or as climax vegetation in the period 8000–4000 cal. BP. Prior to 8000 cal. BP vegetation probably lagged behind the warming, whereas in the period 8000–4000 cal. BP an equilibrium between climate and vegetation was established.

Keywords: climate change, Fennoscandia, Holocene thermal maximum, Holocene thermal decline, northern Europe, temperature reconstructions

1 Introduction

Reconstructions of climate and climate divergence are a prerequisite to understanding many aspects of human livelihood in the Holocene. Previous research has often focussed on the resulting environmental change with an emphasis on vegetation developments and terrestrial species. Increasingly, there has been an interest in individual climate events, in particular the 8.2 BP cold event and the consequences for human societies. There has been less engagement with the relevance and impact of climate as a long-term factor. As part of the research project "*Stone Age Demographics*" on demographic patterns in northern Norway this report seeks to provide an overview of both the general trends as well as the divergence in the available climate data from northern Fennoscandia.

Northernmost Fennoscandia holds a key position to understanding Holocene climate change in the North Atlantic region. The region is one of the best studied areas in the world, with several high quality multi-site studies available (e.g. Seppä 1996; Barnekow 2000; Seppä & Hammarlund 2000; Seppä & Birks 2001; Bjune *et al.* 2004; Jensen & Vorren 2008; Seppä *et al.* 2009; Huntley *et al.* 2013; Kullman & Öberg 2015; Väliiranta *et al.* 2015). There has also been a keen interest in the paleo-oceanography of the North Atlantic and Barents Sea, providing well documented understanding of changes in sea surface temperatures and the strength of the major sea currents during the Holocene (e.g. Sarinthein *et al.* 2003; Andersen *et al.* 2004a; Eldevik *et al.* 2014).

Unfortunately, paleoclimatic proxy-data are notoriously noisy, and it is often difficult to discern the "true" climatic signal. There are numerous reasons for this, e.g. local variations, poor modern analogues, dating errors, vegetation response lags, statistical errors and poor correspondence between proxies and single climatic parameters. Different climate proxies will show dissimilar ecological tipping points regarding one or several climate parameters. Hence, one and the same climatic event could potentially be signaled as several temporally separate climate "events". Some investigations have tried to mitigate such uncertainties by using a large number of records, from which statically mean climatic trends can be drawn (e.g. Seppä *et al.* 2009; and for a larger region Sejrup *et al.* 2016). Even though such investigations are valuable, they heavily depend on one or a few types of proxy-data and focus on a single climatic parameter, most commonly derived from pollen-based transfer-functions. So, despite the large number of records, results would still be vulnerable to any bias following that specific approach (e.g. Paus, 2013). In addition, the "noise" may actually contain meaningful information as it represents different aspects of the past climate (Huntley 2012). A more inclusive approach to the array of available proxies will thus provide a better assessment of past climate conditions.

In the present investigation, the most relevant sites/investigations and the full spectra of climate proxy data is provided, including qualitative and/or temporally limited data not suitable for numerical or statistical analysis.

The main objectives for the present study are to provide an overview of both the general trends as well as the divergence in the available climate data from northern Fennoscandia, with focus on the Holocene Thermal Maximum (HTM), including the build-up and subsequent decline in average temperatures. In addition, potential forcing factors and atmospheric circulation that are in agreement with the observed patterns will be discussed. The overview is also meant to increase the accessibility of the data of this key region for larger scheme investigations, not least for the benefit of disciplines such as ecology and archaeology as well as paleoclimatic studies targeting more specific questions.

1.1 Regional setting

The area under investigation is limited to Fennoscandia north of c. 68°N and extending to c. 71°N at the northernmost point of the European mainland (Fig. 1). The latitudinal extension is between c. 14 to 36°E. Despite the high latitude, the region experiences a mild climate because warm Atlantic water is brought north by ocean currents. The warm waters off the west coast in combination with predominantly westerly winds give rise to large climatic variation within the investigation area, this being enhanced by the Scandes Mountains which place much of the interior and eastern parts of the investigation area in a rain shadow. The western coastal area (as far as Magerøya and the west coast of the Porsanger peninsula) experiences a maritime climate with high precipitation (>1500 mm/yr at the outer coast), cool summers and mild winters. Along the coast east of Magerøya, the climate is drier (precipitation <750 mm/yr) and a few degrees colder during both the summer and winter. The interior experiences a more continental climate with little precipitation (<500 mm/yr) and strong seasonal contrasts (Dannevig & Harstveit, 2013). The main climatic variance can thus be described by two axes: a SW–NE axis along the coast with declining precipitation and to a lesser degree declining temperature towards the northeast; and a coast–inland axis with declining precipitation and increasing seasonal contrasts, especially colder winters, towards the interior. The climatic conditions are exemplified in Table 1. The standard climate reference period 1961–1990 is used to define the modern/present climate conditions.



Figure 1. Left: Overview map with warm Atlantic sea currents marked. Right: The main investigation area (c. 68–71°N) with key-sites and areas referred to in the text marked.

Table 1. Modern climatic conditions (met.no 1961-1990 – the standard reference period for long-term climate change assessments, in this report referred to as modern/present).

Met. station	Coordinates	Climatic region	July temp.	January temp.	Annual temp.	Annual precipitation
Tromsø	69.65°N, 18.94°E	SW coast	11.8°C	-4.4°C	2.5°C	1030 mm/yr
Vardø	70.37°N, 31.10°E	NE coast	9.2°C	-5.1°C	1.3°C	560 mm/yr
Karasjok	69.47°N, 25.48°E	Inland	13.1°C	-17.1°C	-2.4°C	370 mm/yr

2 Proxy records and climate reconstructions

The most relevant investigations of Holocene climate change in northernmost Fennoscandia are considered and key results described. In addition to climate reconstructions from the main study area, marine records from the northern North Atlantic and Barents Sea are considered as sea currents are expected to have a strong influence on the terrestrial climate. In addition, records from the Greenland ice cores are included, considering their importance for defining climate events in the North Atlantic region (and beyond).

From the western part of the investigation area there are numerous pollen transfer-function based climate reconstructions available: Lake Tibetanus near Torneträsk (Hammarlund *et al.* 2002), Lake Gaupstjern and Lake Jervtjern in the Dividalen region (Jensen & Vorren 2008), as well as Lake Barheivatn (Bjune *et al.* 2004), Lake Dalmutladdo (*ibid.*), Lake Toskaljavri (Seppä & Birks 2002) and Lake Tsoulbmajavri (Seppä & Birks 2001) in the Kilpisjärvi region. Several sites on Andøya (Birks *et al.* 2014) and Jansvatnet near Hammerfest (Birks *et al.* 2012) reconstruct climate for the earliest part of the Holocene. In

the eastern part of the investigation area, there are Lake Ifjord in the NE Norwegian part of the investigation area (Seppä *et al.* 2002a), and Lake KP-2 and Lake Yarnyshnoe-3 on the Kola Peninsula (Seppä *et al.* 2008). In addition to these individually published records, July temperatures have been reconstructed from 23 pollen records in Norway and northernmost Fennoscandia (Seppä *et al.* 2009), as well as Fennoscandia as a whole based on 59 records (Sejrup *et al.* 2016). Investigations using diatom- and/or chironomid-based climate transfer-functions are also more common in the western part of the investigation area, around Torneträsk (Bigler *et al.* 2002, 2003, 2006; Larocque and Hall 2004) and Kilpisjärvi (Korhola *et al.* 2000; Seppä *et al.* 2002b).

A more direct way of inferring past summer temperatures than through transfer-functions are tree-line studies based on pollen, macrofossils and/or megafossils. In addition, indicator species, the general vegetation type, and several other types of proxies are useful in climate reconstructions, especially for determining the effective precipitation (precipitation minus evapotranspiration). As these studies vary in character and often contain qualitative information they will be described in more detail (see paragraphs below). Dendroclimatology is not considered here as the method is less suitable for reconstructing long-term climatic trends.

The marine records considered here are those most relevant for determining the strength of the Norwegian Atlantic Current and its eastern continuation as the North Cape Current. Sea surface temperatures (SST) are presented from the Vøring Plateau off the mid-Norwegian coast (core MD95-2011; Calvo *et al.* 2002; Andersen *et al.* 2004a), from the SW Barents Sea (core PSh-5159N; Chistyakova *et al.* 2010; Risebrobakken *et al.* 2010), W Barents Sea (Sarnthein *et al.* 2003) and E Barents Sea (Duplessy *et al.* 2001, 2005). Also, the conditions around Svalbard are of interest as they depend on these currents, and the influence of Atlantic water (Werner *et al.* 2016), coastal water temperatures (Mangerud & Svendsen 2018) and ice cover (Berben *et al.* 2017) are considered.

The most important paleoclimatic investigations from northernmost Fennoscandia concerning the temporal constraints and amplitudes of the warmest Holocene periods are presented in Table 2. The driest period of the Holocene and wet shifts are provided in Table 3, and cold spells and/or shorter periods with abrupt cooling are presented in Table 4. Non transfer-function based climate reconstructions from the investigation area are discussed below.

2.1 Pollen and macrofossils used in tree-line studies

In the Lake Torneträsk area in NW Sweden (see Fig. 1), the sediments of a number of lakes have been analysed. By analysing lakes at different altitudes, Barnekow (2000) concluded that pine and birch were growing at maximum altitude c. 6300–4500 cal. BP. When corrected for land uplift of 20-40 m, this corresponds to a temperature of 1.5-2°C above

present. This was also a dry period with a lowering of the lake level of 1-1.5 m (Lake Badsjön). After 4500 cal. BP, a more humid and cooler climate caused pine to retreat downslope.

In Dividalen (see Fig. 1), north of Torneträsk, pollen and macrofossils have been investigated at several sites (Jensen & Vorren, 2008). Here birch became established up to c. 300 m asl by 10,100 cal. BP, and expanded to above c. 800 m asl before 9400 cal. BP, where it remained until 3300 cal. BP. Pine became established up to 400 m asl after 8500 cal. BP and about 600 m asl after 8200 cal. BP. Alder expanded from 400 m asl to nearly 800 m asl during the period 7900–7600 cal. BP. The maximum forest distribution lasted until c. 6000 cal. BP. More pronounced lowering of the forest-line occurred c. 4600 and 3000 cal. BP.

At Lake Dalmutladdo (355 m asl) in the Kilpisjärvi area (see Fig. 1), fruits and leaves of birch indicate local stands were present at 10,200 cal. BP (Bjune *et al.* 2004). At about 7300 cal. BP, the early Holocene birch forest was replaced by pine forest, and the pine treeline became established 250-300 m higher than today and remained there until about 4000 cal. BP. In the nearby Lake Toskaljavri, the pollen signal suggests a cool, moist climate in the early Holocene that supported birch forest in the area 9600 cal. BP onwards (Seppä *et al.* 2002b). Pine immigrated 8300 cal. BP and formed the altitudinal treeline 6100–4000 cal. BP. Plant communities typical of dry, oligotrophic heaths indicate a dry climate for the period 7000–4000 cal. BP, while alpine plant communities show an inversed pattern expanding after 4000 cal. BP. Pine pollen percentage values decline at 3800 cal. BP. At Lake Tsuolbmajavri further to the east, a gradual expansion of pine started 9200 cal. BP (Seppä & Weckström 1999). It reached a maximum 7200–6000 cal. BP, and the pine tree-line retreated from the area at 4600 cal. BP. During the late Holocene, peatland development increased, indicating a cooler and moister climate.

The forest-line in these three areas (Torneträsk, Dividalen, Kilpisjärvi, see Fig. 1) reached 200-300 m above present at its altitudinal maximum, which with a lapse rate of 0.6°C/100 m would correspond to a (summer) temperature in the range 1.2–1.8 °C above present. In this area the land was c. 100 m lower at 9000 cal. BP relative to today because of later isostatic uplift (Møller & Holmeslet 2002). Using the maximum increase in forest-line (+1.8°C) and adjusting this for land-uplift (c. -0.3°C at 6000 cal. BP, see Møller & Holmeslet 2002), the maximum HTM temperature would have been about 1.5°C warmer than today.

In the northernmost part of the investigation area, Huntley *et al.* (2013) used a direct analogue method to infer July temperatures from three lake-records along the Barents Sea coast (liten Cáhppesjávri, over Gunnarsfjorden, over Kobbkrokvatnet). The estimated mean July temperatures deviated from modern temperatures by -1–2°C for the period 11,900–10,200 cal. BP, +0–1.5°C for 10,200–8500 cal. BP, +0.5–2.0°C for 8500–4300 cal. BP;

and +0–1.0°C for 4300 to present. From the western part of the Varanger Peninsula pollen records indicate the local presence of birch forests from c. 11,000 cal. BP and the establishment of pine in the lowlands as early as 10,300 cal. BP (Seppä 1996). The local presence of pine indicates that July temperatures were at least 1.2–1.5°C higher than present. Both the birch and pine tree lines started to retreat c. 5500 cal. BP. Aquatic macrofossils from northern Finland indicate a warm early- and mid-Holocene c. 11,000–6500 cal. BP, with mean July temperatures 2°C above modern (Väliranta *et al.* 2015; end of warm period as indicated by Shala *et al.* 2017).

Further east on the Kola Peninsula pollen and diatom evidence from Chuna Lake indicate a climatic optimum between 9000 and 5400/5000 cal. BP with warm and dry conditions, followed by a gradual cooling and increase in moisture (Solovieva & Jones 2002). Pollen accumulation rate (PAR) values in general and pine PAR in particular remained high until 4200 cal. BP. At the Poteryanny Zub Lake, further north on the Kola Peninsula, local pine (stomata finds) appeared 9100 cal. BP. The vegetation changed from birch woodland to pine forest at about 7800 cal. BP and reverted back to birch woodland at c. 2800 cal. BP (Gervais *et al.* 2002, labelled KP-2 in Seppä *et al.* 2008). Stomata and pollen in Lake Yarnyshnoe-3 indicate that pine (*Pinus sylvestris*) reached the north coast of the Kola Peninsula already 8900 cal. BP and retreated again c. 4500 cal. BP (Snyder *et al.* 2000). A pine tree-line near the Barents coast would represent a warming of c. 2°C above today (Gervais *et al.* 2002). In northern Russia macrofossil evidence show that the forest advanced to near the current arctic coastline between 10,000 and 8000 cal. BP (9000–7000 BP) and retreated to its present position between 4500 and 3000 cal. BP (4000–3000 BP), although the northern tree-line on the Kola Peninsula was established later than more easterly regions (MacDonald *et al.* 2000). Mean July temperatures along the northern coastline of Russia may have been 2.5°C to 7.0°C warmer than today. Given the near absence macrofossils north of the modern tree-line dated to before 9500 and after 4000 cal. BP, these dates are here used to delimit the HTM.

Table 2 Holocene thermal maximum and early Holocene warming for selected investigations. Area – see map Fig. 1; Data – type of proxy data; C. rec – climate reconstruction method. HTM age – warmest period as defined by the author(s); HTM temp °C – July air temperature or sea surface temperatures above present; Early Holocene warming – pronounced warming preceding the HTM as defined by the authors. Abbreviations: Aqu mac – Aquatic macrofossils; BM – biomarkers; Chi – Chironomids; DA – Direct analogue; Dia – Diatoms; For – Foraminifera; GR – Glacier reconstructions; IS – indicator species; IT – isotopes; Mac – macrofossils; Mar mol – Marine molluscs; Mega – Megafossils (pine); Pol – Pollen; Sed – Sediments; Sto – Stomata; TF – Transfer functions; TLR – Tree-line reconstructions; # – number of sites if more than one for investigations based on ice, sediment or peat profiles. ¹Site (Lake Ifjord) also included in Seppä 1996. ² Northern sub-set. *See also Shala et al. 2017.

Area	Data	C. rec	HTM age (cal. BP)	HTM temp °C	Early Holocene warming	Reference
NW Norway (Barh.)	Pol, Mac	TF, TLR	7000–5000	+2°C	+1.5°C (9500)	Bjune et al. 2004
NW Norway (Dalm.)	Pol, Mac	TF, TLR	7500–3500	+2°C	+0.5°C (9000)	Bjune et al. 2004
NW Norway	Pol, Mac (#4)	TF, TLR	8000–6000	+1.5–3.0°C	+0.5–1.5°C (9400)	Jensen & Vorren 2008
NW Sweden	Pol, Mac (#6)	TLR	6300–4500	+1.5–2°C	N/A	Barnekow 2000
NW Sweden	Pol	TF	8800–5000	+1.5–2°C	+1°C (9500)	Hammarlund et al. 2002
NW Sweden	Pol	TF	8000–5000	+0.5–1.0°C	0°C (9500)	Bigler et al. 2002
NW Finland	Pol	TF	8000–6500	+1.6–1.8°C	+1.0–1.4°C (9600–9000)	Seppä & Birks 2002
NW Finland	Pol	TF, TLR	8000–6000	+1.4–1.7 °C	+0°C (9500), +1°C (9300)	Seppä & Birks 2001
NE Norway	Pol (#3)	DA	8500–4300 (10200–4300)	+0.5–2°C	+0–1.5°C (10200–8500)	Huntley et al. 2013
NE Norway	Pol (#4)	TLR	10300–5500	+1.2–1.5°C	(HTM 10300–5500)	Seppä 1996
NE Norway ¹	Pol	TF	9800–6500	+1.5°C	+1°C (11200–10900)	Seppä et al. 2002a
NW Russia	Pol (#2)	TF	7500–6500	+1.5°C	±0°C (9000–3000)	Seppä et al. 2008
NW Russia	Pol, Sto	TLR	8900–4500	+2°C	Birch expand c. 11000	Snyder et al. 2000
NW Russia	Pol, Dia	TF, TLR	9000–5400/5000	+1.5–2°C	N/A	Solovieva et al. 2005
N Russia	Mac	TLR	10000–4000	+2.5–7°C	+2.5–7°C	MacDonald et al. 2000
NW Sweden	Mega	TLR	9500–9000	+2.3°C	(HTM 9500–9000)	Kullman 1999
NW Sweden	Mega	TLR	9500–8500 (9500–4500)	+3.0°C	(HTM 9500–8500)	Kullman & Öberg 2015
NW Finland	Mega	TLR	6000–4000	>+1°C	N/A	Helama et al. 2004
NW Sweden	Dia	TF	7800–2300	N/A	cool (9000–8000)	Bigler et al. 2006
NW Sweden	Dia	TF	9000–6000	+1.2°C	+0.5°C (9500)	Bigler et al. 2002
NW Sweden	Dia, Chi	TF	9500–4800	+1.5–2°C	+2°C (9500–8500)	Bigler et al., 2003
NW Finland	Chi	TF	8300–3800	+0.5°C	N/A	Seppä et al. 2002b
NW Finland	Dia	TF	7000–5000	+1.0°C	N/A	Korhola et al. 2000
N Finland	Aqu mac (#4)	IS	11000–6500	+2°C	(HTM 11000–6500)	Väliranta et al. 2015*
NW Norway	Sed	GR	8800–3800	No glacier	N/A	Bakke et al. 2005
NW Norway	Pol, Mac, Chi	TF	N/A	N/A	+1°C (from 9700)	Birks et al. 2014
NW Norway	Pol, Mac, Chi	TF	N/A	N/A	+1°C (10000–9400)	Birks et al. 2012
Norwegian Sea	Dia (#3)	TF	9500–6500	+4–5°C	+4°C (9500)	Andersen et al. 2004a
Norwegian Sea	Sed	U ^K ₄₇	8500–5500	+1.5°C	+1°C (11500–8500)	Calvo et al. 2002
SW Barents Sea	For	IS%	9700–7600	+1–2°C	+1–2°C (11200–11000)	Chistyakova et al. 2010
W Barents Sea	For	TF	10700–8800, 8200–7700	+4°C	+4°C (from 10700)	Sarnthein et al. 2003
E Barents Sea	For, δ ¹⁸ O	IS, δ ¹⁸ O	7800–6800	Warm	Warming trend from 10000	Duplessy et al. 2001
W Svalbard	For (#2)	TF	10600–5000	+6°C	Atlantic wat. 10600–8500	Werner et al. 2016
Svalbard, coastal	Mar mol	IS	10200–9200, 8200–6200	+6°C/+4°C	+6°C (from 10200)	Mangerud, Svendsen 2018
E Svalbard	BM	BM	9500–5900	Low ice conc.	N/A	Berben et al. 2017
Greenland	Ice δ ¹⁸ O	δ ¹⁸ O	10000–7000	Warm	(HTM 10000–7000)	Vinther et al. 2009
E Greenland	Chi	TF	10000–5500	+3–4°C	N/A	Axford et al. 2017
Greenland, N Canada	Various (#22) ²	Various	9500–3500 ²	+1.5–5°C	N/A	Briner et al. 2016
Norway, N Fen.scan.	Pol (#23)	TF	8000–6000 (8000–4800)	+1.5°C	N/A	Seppä et al. 2009
Fennoscandia	Pol (#59)	TF	8000–6500 (9000–4500)	+1.5–2°C	N/A	Sejrup et al. 2016

2.2 Megafossils

Megafossils from near the shore of Lake Pompe (999 m asl, named Lake Njulla in other publications) indicate local presence of pine, birch and alder 9500–9000 cal. BP, and for birch and alder also 6200–5200 cal. BP (Kullman 1999). The highest position of the pine tree-line occurred during the early Holocene and was c. 500 m above present, which suggest that summers were c. 2.3°C warmer than today (corrected for land-uplift with -0.6°C). After recession of high-mountain glaciers in the Tärna and Kebnekaise mountains, NW Sweden, the exposed macro- and megafossils of wood were collected and dated (Kullman & Öberg, 2015). During the interval 9500–8500 cal. BP pine and birch grew 600–700 m higher than today, which corresponds to summer temperatures 3.0°C above present (corrected for land uplift). Tree growth ceased around 4500 cal. BP. Both these investigations are in line with megafossil records further south in the Scandes (Kullman 2013; Paus & Haugland 2017).

Megafossil records of pine from the Enontekiö region (around Kilpisjärvi, NW Finland) show a rise in pine forest limit until 6050 cal. BP, when it reached 500 m above past sea level (Helama *et al.* 2004). There it remained until c. 4050 cal. BP when a relative abrupt decline in the altitudinal pine-forest limit occurred, although no pine germination is detected at the highest sites after 4350 cal. BP. After this the tree-line gradually declined until 2550 cal. BP. There are no indications of high altitude pine germination for the period 2750 to 1850 cal. BP. Submerged Scots pines from northern Finland also indicate that the climate was relatively dry in the period between c. 9000 and 4500 cal. BP (8000–4000 BP) showing a stable climate with relatively little interannual variability in the tree-rings (Eronen *et al.* 1999). This period was followed by a shift towards a cooler, more humid and variable climate, with most rapid changes around c. 4500 cal. BP and 2500–2000 cal. BP. The increased humidity in this region is clearly evident by increased lake levels that submerged (and preserved) pines formerly growing on dry land.

Table 3. The mid Holocene dry period and late Holocene paludification (end of dry period) as recorded in selected records. Wet shift indicates a temporal increase in effective precipitation superimposed on the general trend. Age intervals are in cal. yr BP. *Winter precipitation **Rapid decline from c. 9500 cal. BP.

Area	Data/type	Dry period	Wet shifts	Reference
NW Norway	Ombrotrophication	>7500–5100/2500	3600–2900	Vorren 2001
NW Norway	Peat humification	>7500–4500	3800	Vorren <i>et al.</i> 2012
NW Norway	Geochemistry, magnetism	>9500–4300	4800	Balascio & Bradley 2012
NW Norway	Pollen, leaf wax biomarkers	7800–5300	2800, 2500–1500	Balascio & Anderson 2016
NW Norway	Sediments/glacier*	8800–3800		Bakke <i>et al.</i> 2005
NW Sweden	Oxygen isotopes	6600–2700**		Hammarlund & Edwards 1998
NW Sweden	Macrofossils, SIRM	6300–4500		Barnekow 2000
NW Finland	Pollen/vegetation	7000–4000		Seppä <i>et al.</i> 2002b
N Finland	Cladocera, diatoms	9000/7000–6000/4500		Hyvärinen & Alhonen 1994
N Finland	Cladocera	7000–4000	(1500)	Korhola <i>et al.</i> 2005
N Finland	Submerged pines	9000–4500	2500–2000	Eronen <i>et al.</i> 1999
NE Norway	Pollen/vegetation	9000–4000		Allen <i>et al.</i> 2007

Table 4. Recorded mid- and late Holocene cold events and/or pronounced cooling organised in four temporally distinct groups: 8.2 cold event, Mid Holocene Cooling (MHC), mid- to late Holocene transition 4200–3800 cal. BP, and the 2.8 cold event. "Other" include the Dark Age Cold Period (DACP) and the Little Ice Age (LIA).

Area	Data/type	8.2 event	MHC	4.2/3.8 events	2.8 event	other	Reference
Fennoscandia	Terrestrial	8300-8000	(6000, 4800)	3800–3000		500-100	Seppä <i>et al.</i> 2009
NW Norway	Terrestrial	8200	5500	3800	2800	500	Var. pollen ¹
NW Sweden	Terrestrial	8200	4800				Kullman 2013
NW Russia	Terrestrial		5400/5000	4200	2800		Var. pollen ²
NW Finland	Lacustrine		5500–5000	4200			Seppä <i>et al.</i> 2002
NW Sweden	Lacustrine	8400	(4800)		2700		Bigler <i>et al.</i> 2003
NW Sweden	Lacustrine		5400		2900, 2700	700	Bigler <i>et al.</i> 2006
NW Finland	Lacustrine	8300	(5000)	4200	3000	7200, 400	Korhola <i>et al.</i> 2000
N Atlantic	Marine	8200	5900	4200	2800	1400	Bond <i>et al.</i> 1997
N Atlantic	Marine	8300, 7900	4700	4300	2800	6400	Andersen <i>et al.</i> 2004b
N Atlantic	Marine	8100	5800		2800		Calvo <i>et al.</i> 2002
N Atlantic	Marine		5600	3900	2700	1300, 300	Moros <i>et al.</i> 2012
Svalbard	Marine		5000	3700			Rasmussen <i>et al.</i> 2014
Svalbard	Marine	9200–8200		3600			Mangerud & Sven. 2018
Global (#81)	Cold/wet	8200	6300, 4700	(4200)	2700	1550, 550	Wanner <i>et al.</i> 2011
Global (#50)	RCC	9000–8000	6000–5000	4200–3800	3500–2500	1200–1000	Mayewski <i>et al.</i> 2004

¹) Interpreted from pollen-based climate reconstructions by Seppä & Birks (2001, 2002), Bjune *et al.* (2004) and Jensen & Vorren (2008). ²) Noted by Solovieva & Jones (2002) or Gervais *et al.* (2002).

2.3 Other climate proxies

Coastal to inland mires in Troms, NW Norway, show a transition from geogenic to ombrogenous peat in the time span 5100–2500 cal. BP, with a concentration in the interval 3600–2900 cal. BP (Vorren 2001). Around 2900 cal. BP (2750 BP) there was a major change to cooler and wetter conditions. A peat humification record from the Lofoten region, NW Norway, indicates a relatively stable warm and dry climate until 4500 cal. BP (Vorren *et al.* 2012). Especially cold and wet conditions occurred in the period 3800–3000 cal. BP, followed by a highly variable climate until 1400 cal. BP. Lake sediment properties from the same area also indicate potential regional cooling events 10,900–10,200 and 9200–8000, a stable and productive period 7800–4800 cal. BP, and a distinct shift towards a wetter and more variable climate initiated at 4300 cal. BP (Balascio & Bradley 2012). Aeolian sand influx (ASI) in a lake in the Vesterålen area, just north of Lofoten, indicate a clear increase in the intensity of west-southwestern winds after 2800 cal. BP (Nielsen *et al.* 2016). Pollen and leaf wax biomarker data from a bog NE of Tromsø indicate dry conditions 7800–5300 cal. BP, and the biomarkers reveal a distinct shift to wetter conditions after c. 2800 cal. BP (Balascio & Anderson 2016). The Lenagsbreene glaciers in Lyngen, Troms, melted away around 8800 cal. BP and were reformed c. 3800 cal BP (Bakke *et al.* 2005). A speleothem stable isotope record in northern Norway reveals an early Holocene (10,330–7700 cal. BP) with a highly variable climate, a mid-Holocene (7700–4110 cal. BP) with low variability and a late Holocene (c. 4110 cal. BP to present) with high variability and a shift between two distinct modes (Linge *et al.* 2009). Oxygen isotope records from NW Sweden (Torneträsk area) indicate a decline in precipitation from about 9500 to 6500 cal. BP and a slight increase c. 2700 cal. BP (Hammarlund & Edwards 1998; Hammarlund *et al.* 2002). Varved sediments from Lake Sarsjön, northern Sweden, reveals a distinct increase in mineral deposition at 3710 ± 86 cal. BP, probably associated with a change towards a more unstable and maritime type of climate (Snowball *et al.* 1999).

3 Interpreting the climatic records

All palaeoclimatic reconstructions are associated with some degree of uncertainty. Most organisms respond to a multitude of climatic (and other) parameters but are often used as a proxy for a single climatic parameter, which may bias the result (see Huntley *et al.* 2012). The most common method for reconstructing terrestrial climate is the pollen-based transfer-function method (e.g. Seppä *et al.* 2004). This method reconstructs past temperatures, but it has its limitations, and criticism largely follows three lines: 1) The modern, human impacted vegetation used for training sets are poor analogues for past conditions; 2) Pollen dispersal-deposition and vegetation response to climate are too complicated to be accurately captured by a few site-independent parameters; and, 3)

There might be a lag of several thousands of years between the onset of long-lasting climate ameliorations (as in the early Holocene) and the establishment of equilibrium between vegetation and climate (Paus, 2013). Transfer functions for aquatic taxa as chironomids and diatoms would theoretically be a more direct measurement of climate conditions, but they also suffer from poor modern analogues and autocorrelation with terrestrial vegetation affecting nutrient input and pH (see Bigler *et al.* 2002; Nyman *et al.* 2008).

Latitudinal or altitudinal tree-line studies using macro- or megafossils are good indicators of the HTM, as the tree-line is closely correlated to the mean July temperature. Even as such it is not without potential bias. Early Holocene mountain sites have experienced substantial isostatic uplift, which must be extrapolated from coastal areas where it can be determined. Another assumption is that the lapse rate has remained constant. In the investigation area, the modern difference in tree-line from the outer coast to the western part of the inland area is about 400 m (Moen 1999), and while birch (*Betula pubescens*) correlates with a July mean of 12°C in oceanic western Norway it drops to 10°C in eastern Norway (Odland 1996). Variance in oceanicity is thus likely to have a major impact on the tree-line as well as the vegetation composition as a whole. The increased summer insolation during the early Holocene would also give a higher increase in growing degree days in mountain areas compared to the lowlands, possibly even more enhanced by a stronger massenerhebung effect. These potential biases in climate reconstructions highlight the importance of using several types of proxies in order to minimise the risk of misinterpretation.

3.1 Structuring Holocene climate change

In order to summarize and provide a better overview of the climate change during the Holocene it has here been structured into periods with relatively uniform climatic conditions or trends. The most common, general division of the Holocene climate is into three partitions: an early cool period with increasing temperatures, a warm middle phase with maximum temperatures, and a cool late phase with declining temperatures (e.g. von Post 1946; Firbas 1949; Hafsten 1970; Wanner *et al.* 2011). On a general level the records considered here fit well within this scheme, but at closer examination problems arise as different climate characteristics are out of phase. Because of this a more neutral labelling of the Holocene into an early, middle and late period is applied. Each of these periods are then divided into an early (EP) and late phase (LP) based on the climatic characteristics. With the exception of the start of the Holocene and the well dated cold spell at 2800 cal. BP the climatic zones are set to even millennia. The climatic conditions during the different periods are described below and have been summarised in Table 5, and the schematic temperature development is presented in Fig. 2.

Early Holocene EP, 11,700–10,000 cal. BP, cold, rapid warming, -2 to +1°C

At the onset of the Holocene ca. 11,700 cal. BP the Scandinavian Ice Sheet covered most of the inner fjord area and all of the inland (Romundset *et al.* 2017). At the end of the period it had dwindled down to a fraction of its former size situated SSW of the investigation area (Stroeven *et al.* 2016). June insolation would have been at its peak 12,000–11,000 cal. BP, and July insolation 10,000–9000 cal. BP (65°N, Berger & Loutre 1991; Berger 1992). Vegetation development during the early Holocene was strongly delayed by retreating glaciers, katabatic winds, soil development and migration, showing strong local and regional variance. At the coast, this delay is especially evident prior to c. 11,000 cal. BP, while this situation continues longer in the inland closer to the ice margin.

The dry conditions of the Younger Dryas prevailed some time into the early Holocene (Birks 2015). The vegetation was characterized by shrub-tundra / heathland, although birch forest started to expand after c. 11,000 cal. BP (e.g. Seppä & Hammarlund 2000; Birks *et al.* 2012; Huntley *et al.* 2013; Sjögren & Damm 2018). Several records suggest rapidly increasing temperatures during this period, from a few degrees below present (c. -2°C) at the start of the Holocene (11,700 cal. BP) to temperatures at or slightly above present (+1°C) around 10,000 cal. BP. However, the climate is variable and there are several severe setbacks in the warming recorded (e.g. Seppä *et al.* 2002a). Notably, aquatic macrofossils in general indicate about 2°C higher temperatures than pollen-based studies in the region, and record temperatures 2°C above present already at 11,000 cal. BP (Väliranta *et al.* 2015). It is possible that terrestrial vegetation was not in equilibrium with the climate because of soil maturation and migrational lag, or that the microclimate of the aquatic taxa rather reflects the high solar insolation than air temperatures and/or a shorter but warmer growing season. Climatic reconstructions for this period are relatively sparse, and the environment is heavily influenced by the rapidly melting Scandinavian Ice Sheet. Nevertheless, the general picture is that of a variable climate with rapidly increasing temperature and a shift from the dry Younger Dryas conditions towards increased precipitation.

Early Holocene LP, 10,000–8000 cal. BP, wet, warm and variable, +0 to +2°C

Tree-birch started to spread c. 11,000 cal. BP, and around 10,000 cal. BP most of the early Holocene shrub-tundra had been replaced by oceanic birch-forests across northernmost Fennoscandia, indicating a shift towards a relatively mild and moist climate strongly influenced by westerlies (Seppä & Hammarlund 2000; Sjögren & Damm 2018). Megafossils from pine indicate that the Holocene pine forest-line peaked in altitude during the period 9500–9000 cal. BP, corresponding to summer temperatures 2–3°C above present (Kullman 1999; Kullman & Öberg 2015). An increase in sea surface temperature is evident in most marine records ranging from 10,700 to 9000 cal. BP, and a start of the HTM around 10,000 cal. BP can also be seen in records from Greenland and northern Russia. A climatic

shift seems to occur around 9000–8500 cal. BP, when the megafossils indicate a decrease in summer temperatures, and other investigations suggest that the climate became drier (Eronen *et al.* 1999; Hyvärinen & Alhonen 1994; Allen *et al.* 2007), most likely including a reduction in winter precipitation (Hammarlund & Edwards 1998; Bakke *et al.* 2005). A few other records indicate a cooling around 9000 cal. BP (Korhola *et al.* 2000; Sarnthein *et al.* 2003; Mangerud & Svendsen 2018). On the other hand, several records also indicate a warming at the same time (Bigler *et al.* 2002; Solovieva & Jones 2002; Allen *et al.* 2007). Notably, even though pine abundance did not reach its maximum (as determined by pollen and macrofossils) until later, it did expand in the period after 10,000 cal. BP and especially after 9000 cal. BP (Seppä & Hammarlund 2000; Snyder *et al.* 2000; Gervais *et al.* 2002; Sjögren & Damm 2018). Mega- and macrofossils of pine dated to c. 9500 cal. BP are spread across the region indicating a peak occurrence at this time (Kullman 1999, Kullman & Öberg 2015, Väiliranta *et al.* 2015; Virdnejávri Ua-46463, Marianne Skandfer, personal communication 2017). Overall, the period is characterized by warm, variable and wet conditions as indicated by many records, with peak temperatures around 9500 cal. BP. Although, not all records agree with this, suggesting a complex environmental context at the time which prevents firm conclusions.

Mid Holocene EP, 8000–6000 cal. BP, dry, warm and stable, +1 to 2°C

Most transfer-function based temperature reconstructions indicate that the period with highest summer temperatures started around 8000 cal. BP. Pine (*Pinus sylvestris*) and alder (*Alnus incana*) became common and replaced much of the previous birch-fern forest (Sjögren & Damm 2018), and the tree-line reached its highest altitude and latitude in some areas. Many records also indicate drier conditions and there are few indications of climatic fluctuations in the records. The overall impression of this period is that it was warm, dry and stable, and the warmest and climatically most stable millennia of the Holocene occurred 8000–7000 cal. BP (e.g. Seppä *et al.* 2009).

Mid Holocene LP, 6000–4000 cal. BP, dry and warm, +0.5 to 1.5°C

The general climate is similar to the preceding period, though deviating by showing records of increasingly colder and more unstable conditions. A major cold event seems to occur around 5500 cal. BP (see Table 4). Some records based on the extension of pine indicate the highest temperature in this period (Barnekow 2000; Helama *et al.* 2004), but the peak abundance of pine could be explained by reduced precipitation, local variations and/or migratory lags, rather than temperature alone. The shift between the warm mid Holocene and cold late Holocene could not be pinpointed down to a single century but to a transition phase 4500–3800 cal. BP, even though the dates 4500, 4200 and 3800 recur in many investigations. Here the even time point 4000 cal. BP is used as a simplified definition for this gradual or step-wise change.

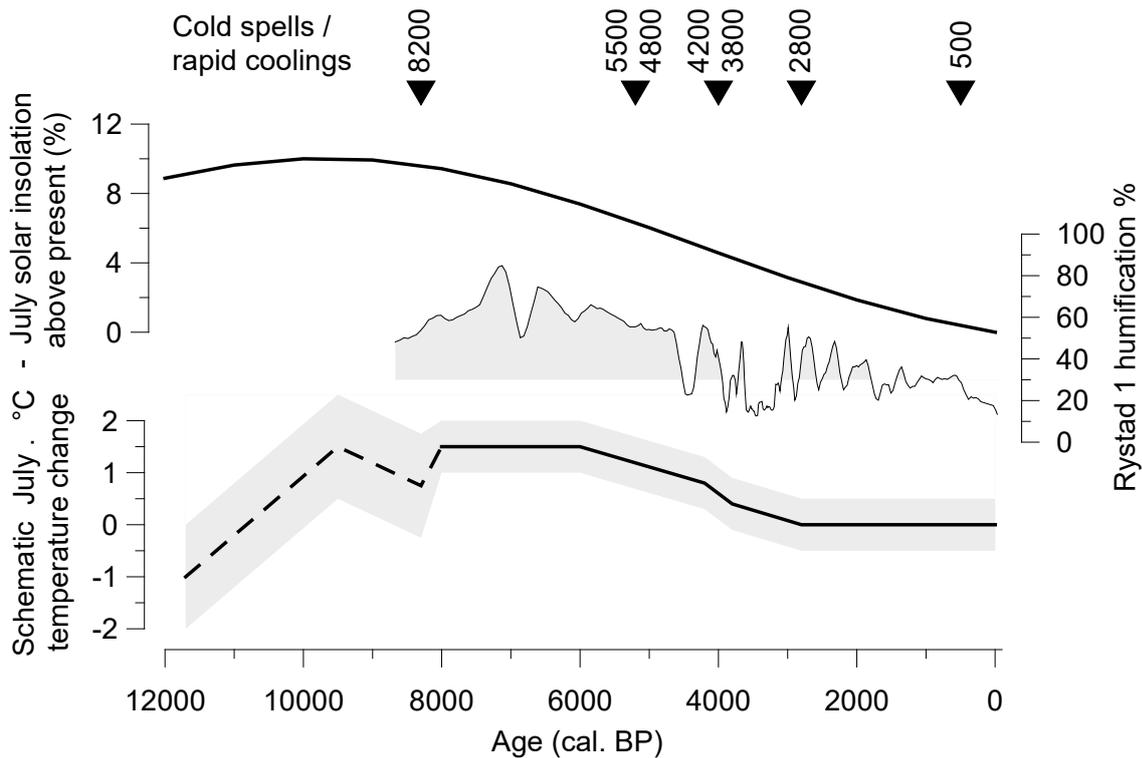


Figure 2. Schematic synthesis of climate change in northernmost Fennoscandia compared with July solar insolation at 65°N (Berger & Loutre 1991) and the late Holocene increase in effective precipitation as exemplified by the Rystad 1 humification record (Vorren *et al.* 2012). The schematic temperature curve is the general deviation of July air temperature compared to modern as deduced from the available climate records (see text and Table 2). Dashed line indicates a more uncertain development with more spatial variation, larger discrepancies between proxies and fewer continuous reconstructions. The grey fields indicate the common range within and between records. Cold spells / rapid cooling declines as reflected in the terrestrial climatic records (see Table 4).

Late Holocene EP, 4000–2800 cal. BP, wet and variable +0 to 1°C

The long-term gradual decline in summer temperatures became more abrupt around 4000 cal. BP, but the change towards much wetter conditions 4500–4000 cal. BP was more profound (see Table 3). Most likely the gradually declining solar insolation and the associated reforming of Arctic sea-ice reached a threshold that resulted in a major reconfiguration of the climatic system (see e.g. Coumou *et al.* 2018).

Late Holocene LP, 2800–0 cal. BP, wet, cool and variable, -0.5 to +0.5°C

In competition with the 8.2 event, the most pronounced cold spell occurred c. 2800 cal. BP, marking the end of the warmer climate and the start of the modern climatic conditions. Notably though, many proxy records show no indications of these cold spells which suggests that the abrupt decline in temperature was not too severe and/or of short

duration. Well known cold spells in the late Holocene are the Dark Ages Cold Period c. 1400 cal. BP (see Helama *et al.* 2017) and the Little Ice Age c. 350–100 cal. BP (Lamb 1995).

Table 5. Schematic summary of Holocene climate conditions in northernmost Fennoscandia. Atmospheric circulation indicates the predominate mode as suggested by the inferred climate conditions.

	Early Holocene		Mid Holocene		Late Holocene	
	11,700– 10,000	10,000– 8000	8000– 6000	6000– 4000	4000– 2800	2800– present
Climate conditions						
Solar insolation	Very high	Very high	High	Moderate	Low	Low
Norwegian current	Increasing	Strong	Strong	Decreasing	Weak	Weak
Climate trend/type	Warming	Variable	Stable	Cooling	Cooling	Variable
Moisture	Dry–wetter	Wet	Dry	Dry	Wet	Wet
July temperature	Cold	Warm	Warm	Warm	Moderate	Cool
Atmospheric circ.	Variable	Zonal	Meridional	Meridional	Zonal	Zonal
Δ July temp (°C)	-0.5±1.5	+1±1	+1.5±0.5	+1±0.5	+0.5±0.5	±0.5

4 Discussion

4.1 The onset of the Holocene Thermal Maximum

With few exceptions, the peak temperature of the Holocene is fairly similar between proxies, about $1.5\pm 0.5^{\circ}\text{C}$ higher than present. This is in line with estimates of HTM temperatures in northern North America (north of 60°N) that on average measure $1.6\pm 0.8^{\circ}\text{C}$ above present (Kaufman *et al.* 2004). Based on the altitudinal and latitudinal tree-line reconstructions the HTM started c. 10,000–9000 cal. BP, and the same is evident from the vegetation composition in the NE part of the investigation area. Most other climate reconstructions, especially in the western part, place the start of the HTM around 8000 cal. BP. Generally, the HTM seemed to terminate around 6000 cal. BP, even though the temperature decline appeared as gradual until the major climate shift around 4500–3800 cal. BP. The HTM in northernmost Fennoscandia is thus closely associated with the strength of the northbound sea currents and the sea surface temperatures. For the north Atlantic Anderson *et al.* (2004a) divided the Holocene in three periods based on reconstructions of sea surface temperatures (SST) from three north Atlantic high-resolution sediment cores: the Holocene Climate Optimum 9500–6500 cal. BP, the Holocene Transition Period 6500–3000 cal. BP, and the cool Late Holocene Period 3000–0

cal. BP. This is also in line with the development seen in the Greenland ice records with a HTM c. 10,000–7000 cal. BP (Vinther *et al.* 2009).

Summer solar insolation at 70°N peaked around 11,000 cal. BP (Huybers 2006; Huybers & Eisenman 2006), although the Scandinavian Ice Sheet (SIS) had a cooling effect both directly by increased albedo and katabatic winds, but also by providing large amounts of cold, fresh meltwater to the Norwegian Coastal Current. Most of the SIS ice volume was melted by 10,500 cal. BP. This reduced the outflow of meltwater to the Atlantic, although final deglaciation of the Scandinavian Ice Sheet did not occur until 9100 cal. BP (Cuzzone *et al.* 2016). The onset of the HTM thus largely coincides with the peak summer insolation and the (almost) deglaciated SIS. Atmospheric circulation models suggest that the Laurentide Ice Sheet (final deglaciation c. 6700 cal. BP; Ullman *et al.*, 2016) had a chilling effect on regions directly downwind, delaying the onset of the HTM until 9000–8000 cal. BP (e.g. Renssen *et al.* 2017). This could potentially have had a moderate (<1°C) chilling effect on the region, although it does not explain the differences in reconstructed temperatures between proxies and sub-regions. Possibly, the demise of the Laurentide Ice Sheet affected the strength and direction of the westerlies resulting in increasingly drier conditions towards 7000 cal. BP.

A lag of 500–1500 years in the terrestrial records is not unexpected as it will take some time for the vegetation to establish equilibrium with the climate. Pollen based transfer-function temperature reconstructions from Fennoscandia (see present paper; Seppä *et al.* 2009; Sejrup *et al.* 2016) generally indicate a start of the HTM around 8000 cal. BP which appears substantially delayed compared to other reconstructions. Most probably this indicates a combination of changes in other climatic conditions than July/summer temperature, as direct insolation, precipitation and seasonal contrasts, and biotic factors as soil maturation, migratory lag and species competition.

In the climate reconstructions considered here, some discrepancies between proxies become obvious. In Lake Yarnyshnoe-3 on the Kola Peninsula pine stomata (Snyder *et al.* 2000) indicate the presence of pine some 100 km north of its current distribution between c. 9000 and 4500 cal. BP, representing a summer temperature 2°C above present. In Poteryanny Zub, another lake on the Kola Peninsula, stomata indicate local presence of pine c. 9000 although no expansion is seen in the pollen record until 8000 cal. BP (Gervais *et al.* 2002). In contrast, the pollen-based transfer-function temperature reconstructions from these two lakes indicate a HTM around 7500–6500 cal. BP with temperature 1.5°C above present (Seppä *et al.*, 2008). In the western part of the investigation area, pine was growing at high altitude in the Scandes from about 9500 cal. BP (Kullman 1999; Kullman & Öberg 2015), while most pollen-based transfer-function temperature reconstructions indicate maximum temperatures no earlier than 8000 cal. BP (see Table 2). So, a lag of c. 1500 years appears between the highest-lying / northernmost tree-line establishment and

climax vegetation / pollen-based transfer-function temperature reconstructions. This lag of 1500 years is of similar magnitude as observed in mountain areas in mid-Scandinavia (Paus, 2013). Hence, tree-line studies across the region indicate a warming of c. +2°C some 1500 years prior to the bulk of pollen-based transfer-function temperature reconstructions.

Non-analogue climate (e.g. Huntley 2012) and soil maturation (e.g. Pennington 1986) may explain some of these discrepancies. Pollen-based transfer-function estimates for both terrestrial and aquatic taxa have presented a poor fit for the early Holocene temperatures (e.g. Bigler *et al.* 2002; Nyman *et al.* 2008, Seppä *et al.* 2008). Model-based quantitative vegetation reconstructions from the Dividalen area have showed that changes in July temperature could explain changes in the vegetation after 7400 cal. BP, but not prior to 7400 cal. BP, suggesting that other factors were decisive for the vegetation dynamic in the early Holocene (Sjögren *et al.* 2015). The main reason for these discrepancies is likely the non-analogue climate conditions of the early Holocene, particularly concerning solar insolation. Peak summer solar insolation at 70°N occurred c. 11,000 cal. BP and peak annual insolation 11,000–8000 cal. BP (Huybers 2006; Huybers & Eisenman 2006). At 11,000 cal. BP, annual (integrated) insolation was only 2% above modern. Insolation above 275 W/m², roughly equivalent with the insolation that today is required to reach temperatures above 0°C, was 5% higher. Although peak summer insolation, today between 475–500 W/m² and equivalent to a monthly mean temperature around 10°C at the northern coast, was 82% higher. Combined with a strong influence of the westerlies, this would have resulted in non-analogue climate conditions with warm and wet summers combined with long, cold and snow-rich winters. Possibly, the tree-line would have responded more positively to these conditions than the forest-line/density, partially explaining the discrepancies between proxies. Nevertheless, climate change will inevitably occur prior to vegetation change, resulting in a shorter or longer lag. When different proxies / records show asynchronous responses to climate change, it should be the earliest records that provide the most likely date for the onset of change (allowing for dating error and comparable quality), in this case 9500 cal. BP.

4.2 The Holocene Thermal Decline

From about 7000–6000 cal. BP there is a general decrease in July temperature which has been attributed to the gradual decline in solar insolation (Sejrup *et al.* 2016, Zhang *et al.* 2017), although the period 6000–5000 also coincided with low solar activity (Mayewski *et al.* 2004). Colder climate and contrasting patterns of hydrological change have been recognised worldwide in the period 5600–5000 cal. BP (Magny & Haas 2004), including at Spitsbergen c. 5500 cal. BP (Alsos *et al.* 2016). A striking example from Scandinavia is the sudden extinction of the pond turtle (*Emys orbicularis*) from southern Sweden around

5500 cal. BP (Sommer *et al.* 2009), where it had been present since 9800 cal. BP. Many records presented here show a decline in temperature and/or a cold spell in the period 6000–4800 cal. BP. Even though the abruptness and amplitude may be discussed, this “mid Holocene cooling” is a period with clear changes toward a colder climate constituting the early phase of the more general Holocene Thermal Decline. However, the main shift towards cooler and especially wetter climate, occurred 4500–3800 cal. BP (see Table 3 and 4). The cold and wet conditions of the late Holocene are also associated with a more variable climate (e.g. Linge *et al.* 2009; Balascio and Bradley 2012; Vorren *et al.* 2012). The high-resolution record from Lake Igelsjön further south in Sweden reveals that this transition occurred in two steps: the first between 4450 and 4350 cal. BP and the second and more pronounced between 4000 and 3800 cal. BP (Jessen *et al.* 2005). Weaker westerlies and reduced frequencies of blocking anticyclones over the Baltic area would allow recurrent penetration of NE winds bringing down cold and dry air-masses from the Arctic. This would enhance the cooling tendency in especially the NE part of the investigation area particularly after 4000 cal. BP. The pine forests and alder occurrences remained relatively intact, at least in the western part of the investigation area, until this change (Sjögren & Damm 2018), as do some estimates of the HTM. So even if July temperatures declined, the HTM can be defined as the stable climate period that corresponded to the mid-Holocene climax forest, showing the edaphical or ecological Holocene optimum (Høeg *et al.* 2018), delimited to 8000–4000 cal. BP.

After c. 4000 cal. BP, the gradual decline in temperature continued until a pronounced cold event occurred c. 2800 cal. BP, most evident in many proxies. This cold event is globally recognised and has been associated with a minima in solar activity (van Geel *et al.* 1996). After 2800 cal. BP long-term climatic conditions are similar to modern. Shorter warm and cold spells are regionally evident as the Dark Ages Cold Period c. 1400 cal. BP (see Helama *et al.* 2017) and the Little Ice Age c. 350–100 cal. BP (see Lamb 1995).

4.3 Spatial variation

During the early Holocene, the expansion of pine was time transgressive, ranging from 10,900–8900 cal. BP in the north-east to 8800–6300 cal. BP in the west and south-west (Seppä & Hammarlund, 2000). Sjögren & Damm (2018) narrowed the timing to 10,200 cal. BP in the NE and 7300 cal. BP in the W and SW. Seppä & Hammarlund argue that this may reflect gradually decreasing influence of Atlantic air masses during the early and middle Holocene. The present pollen and terrestrial macrofossil-based climate reconstructions corroborate this view. Records from NE Norway indicate a start of the HTM c. 10,000 cal. BP, similar as in N Russia. Though records from the Kola Peninsula point to a somewhat later onset, c. 9000 cal. BP. Most records from the SW and W indicate a start of the HTM from 8000 cal. BP or later. The sites closest to the western coast, Lake Barheivatn and

Lake Dalmutladdo actually record some of the latest onsets of the HTM: 7000 and 7500 cal. BP, respectively.

A key site to unravel the climate conditions is Lake Tibetanus in the Torneträsk area, NW Sweden, with a sediment record going back 10,000 yr (Berglund *et al.* 1996; Hammarlund & Edwards 1998; Hammarlund *et al.* 2002). The oxygen isotope and pollen records indicate moist, maritime conditions during the early part of the record. Maximum influence of the westerlies (high zonal index) occurred c. 9500 cal. BP, then the influence rapidly declined until 8800 cal. BP, and more gradually until 6300 cal. BP. The main vegetation changes closely follow the decline in precipitation, and pollen inferred temperature is inverse to precipitation over the early part of the record, prior to 6500 cal. BP. The driest conditions of the Holocene seem to have occurred c. 6500–6000 cal. BP (Hyvärinen & Alhonen 1994; Hammarlund *et al.* 2002), when the Norwegian Atlantic Current was strongly weakened (Andersen *et al.* 2004a). The reduced northward flow of warm surface water would by itself indicate a weakening of the westerlies, and colder water would also reduce the moisture uptake, suggesting (relatively) weaker and drier westerlies. This is concurrent with the expansion of pine in some areas (Barnekow 2000; Helama *et al.* 2004), the last major expansion of pine during the Holocene.

The early presence of pine in high altitude areas of the Scandes (Kullman 1999; Kullman & Öberg 2015) seemingly contradicts a peak influence of western air masses at the time (Hammarlund & Edwards 1998). One possible explanation is topography, i.e. that pine grows in areas largely sheltered from the westerlies, and the finds normally occur in the climatically most favourable areas (Kullman, 2013). A second explanation is that the strength of the westerlies was very variable. Interestingly, Huntley *et al.* (2013) report coincidental peaks in dwarf birch pollen (*Betula nana*) in two lakes c. 9600–9500 and 8900 cal. BP, which they interpret as cooling events caused by a weakening of the North Cape Current. As this current is largely driven by the strength of the westerlies it would indicate a reduction of western air masses and a more continental climate, favouring the establishment of pine.

The North Cape Current would have a direct influence on the NE–E coastal part of the investigation area. Interestingly, the warmest condition of the eastern Barents Sea (7800–6800 cal. BP, Duplessy *et al.* 2001, 2005) coincides with the pollen-based transfer-function derived temperature peak at the north coast of the Kola Peninsula (7500–6500 cal. BP., Seppä *et al.* 2008). The relatively late peak in temperatures of the eastern Barents Sea has been attributed to the melting of local glaciers (Duplessy *et al.* 2001, 2005), and the cold coastal water of the Kola coast might have had a restraining effect on the establishment of pine forests compared to NE Norway. The sub-regional climate variation in northernmost Fennoscandia can thus largely be explained by spatial and temporal variation in the influence of westerly air masses.

4.4 Forcing factors

The climate dynamic of the Holocene could largely be explained by two forcing factors – summer solar insolation and polar ice. In the earliest part of the Holocene the climate was strongly influenced by the remaining ice sheets. Meltwater bursts, especially in the earliest period prior to 10,000 cal. BP, would disrupt the sea currents causing temporary cold spells and a more Arctic climate. Even though summer solar insolation was high, the remaining large ice sheets in the Arctic would maintain a high latitudinal temperature gradient that would fuel a zonal atmospheric circulation with strong westerlies. After most of the inland ice sheets were gone this would result in a maritime climate, but due to the high summer solar insolation the summers would nonetheless be warm. During the middle Holocene summer solar insolation was still rather high, and with the inland ice-sheets completely gone and with presumably little sea-ice in the Arctic the altitudinal temperature gradient would be low. This would result in weak westerlies and a more meridional atmospheric circulation, i.e. drier and more continental conditions (see Yu and Harrison, 1995). During the late Holocene the weak summer solar insolation together with the reforming of glaciers and especially the expansion of Arctic sea-ice would yet again increase the latitudinal temperature gradient. The result would be a more zonal atmospheric circulation with increased influence of the westerlies. This would again give rise to a mild and wet climate, but with the low summer insolation of the late Holocene summers would be cool. If this is used as an analogue for future climate change, the current loss of Arctic sea-ice caused by global warming would weaken the latitudinal gradient, which could result in a new, major reconfiguration of the atmospheric circulation to a more meridional type (see Coumou *et al.* 2018).

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Correction notice

15 March 2021: "Track changes" marks have been removed from page 17.