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# Arctic hydroclimate variability during the last 2000 years: current understanding and research challenges

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**Abstract.** Reanalysis data show an increasing trend in Arctic precipitation over the 20th century, but changes are not homogenous across seasons or space. The observed hydroclimate changes are expected to continue and possibly accelerate in the coming century, not only affecting pan-Arctic natural ecosystems and human activities, but also lower latitudes through the atmospheric and ocean circulations. However, a lack of spatiotemporal observational data makes reliable quantification of Arctic hydroclimate change difficult, especially in a long-term context. To understand Arctic hydroclimate and its variability prior to the instrumental record, climate proxy records are needed. The purpose of this review is to summarise the current understanding of Arctic hydroclimate during the past 2000 years. First, the paper reviews the main natural archives and proxies used to infer past hydroclimate variations in this remote region and outlines the difficulty of disentangling the moisture from the temperature signal in these records. Second, a comparison of two sets of hydroclimate records covering the Common Era from two data-rich regions, North America and Fennoscandia, reveals inter- and intra-regional differences. Third, building on earlier work, this paper shows the potential for providing a high-resolution hydroclimate reconstruction for the Arctic and a comparison with last-millennium simulations from fully coupled climate models. In general, hydroclimate proxies and simulations indicate that the Medieval Climate Anomaly tends to have been wetter than the Little Ice Age (LIA), but there are large regional differences. However, the regional coverage of the proxy data is inadequate, with distinct data gaps in most of Eurasia and parts of North America, making robust assessments for the whole Arctic impossible at present. To fully assess pan-Arctic hydroclimate variability for the last 2 millennia, additional proxy records are required.

## 1 Introduction

Global climate is changing rapidly, largely due to increased anthropogenic greenhouse gas emissions (IPCC, 2013). However, distinct regional differences in the magnitude of observed warming in recent decades are apparent; for example, the Arctic has warmed at more than twice the rate of the global average (Cohen et al., 2014). This *Arctic amplification* (Serreze et al., 2009) is due to complex feedback processes within the atmosphere–cryosphere–ocean system, including surface albedo and heat exchange with the ocean (Johannessen et al., 2003; Hind et al., 2016) and, most importantly, substantial losses of sea ice extent and late-spring snow cover (Overland, 2014).

Temperature increases have resulted in an intensified hydrological cycle (Huntington, 2006). Increasing precipitation in the Arctic has been linked to higher local evaporation and reduced sea ice coverage (Bintanja and Selten, 2014; Kopec et al., 2016), but also enhanced northward transport of moisture (Zhang et al., 2013). According to most climate models (see Sect. 2), precipitation will continue to increase in the coming century, with the largest changes occurring over the Arctic Ocean (Bintanja and Selten, 2014). However, there are still large uncertainties regarding hydroclimate variability and changes in the hydrological cycle in the Arctic due to incomplete or fragmentary data (Serreze et al., 2000; Screen and Simmonds, 2012). This makes it difficult to detect changes in and understand the mechanisms controlling hydroclimate variability on different timescales.

There are large spatial differences in the meteorological station distribution across the Arctic, and except for Fennoscandia and westernmost Russia, only a few observational records reach more than 75 years back in time (Bekryaev et al., 2010), making it difficult to provide a spatiotemporal understanding of hydroclimate variability. Going beyond the observational records, climate proxies are needed. Most reconstructions of past climate for the whole Arctic during the Common Era (CE) have focused on temperature (Overpeck et al., 1997; Kaufman et al., 2009; Shi et al., 2012; Hanhijärvi et al., 2013; McKay and Kaufman, 2014). However, a number of proxies recorded in natural archives, such as ice cores, lake and peat sediments, and tree rings, can provide information on hydroclimate variations in the Arctic. They provide information with different temporal and seasonal resolution. In a recent study of hydroclimate variability across the Northern Hemisphere during the last 1200 years, Ljunqvist et al. (2016) found a tendency for generally wetter conditions during the 9th–11th centuries corresponding to the Medieval Climate Anomaly (MCA, ca. 900–1150 CE), whereas the 12th–19th centuries, including the Little Ice Age (LIA, ca. 1400–1850 CE), showed more widespread dry conditions. However, for the Arctic, only 18 records with heterogeneous spatial distribution were included. Nevertheless, ongoing efforts to collect new data have resulted in a growing network that will increase our understanding of Arctic hydroclimate variability.

The aim of this review is to summarise the current understanding of Arctic hydroclimate, focusing on the last 2 millennia. The paper uses the PAGES 2k definition of the Arctic, i.e. the region north of 60° N. Section 2 briefly presents the current state and a future outlook of Arctic hydroclimate and impacts from observations and climate model simulations from the Coupled Model Intercomparison Project Phase 5 (CMIP5; Taylor et al., 2012). Section 3 reviews the various archives and proxies used to derive information on past hydroclimate variability. Section 4 presents multi-proxy comparisons of hydroclimate variability in Arctic Canada and Fennoscandia, two regions with denser networks of sites. In Sect. 5, a new compilation of Arctic hydroclimate data, which illustrates the potential to derive higher temporal resolution than that of Ljungqvist et al. (2016), is compared to model simulations from the Paleoclimate Modelling Intercomparison Project Phase III (PMIP3; Braconnot et al., 2012). The current understanding of Arctic hydroclimate during the last 2 millennia is summarised in Sect. 6, and some recommendations for future work are given in Sect. 7.

## 2 Current and future Arctic hydroclimate and its impacts

### 2.1 Observations and models

Precipitation data derived from the gridded ERA-20C dataset (Poli et al., 2013) averaged over the whole Arctic ($\geq 60°$ N) show a positive trend over the last century, with a notable increase in the last few decades (Fig. 1), which is in line with previous findings (Serreze et al., 2000; Min et al., 2008). A similar trend is seen in an ensemble of 12 historical CMIP5 simulations (1900–2005; see Table S1 in the Supplement for information on the included models). The spatial pattern of precipitation change is heterogeneous across the region (Fig. 2a and b), with the largest increases in annual precipitation over the North Atlantic and Barents Sea and, to a lesser degree, over eastern Asia, western North America, and

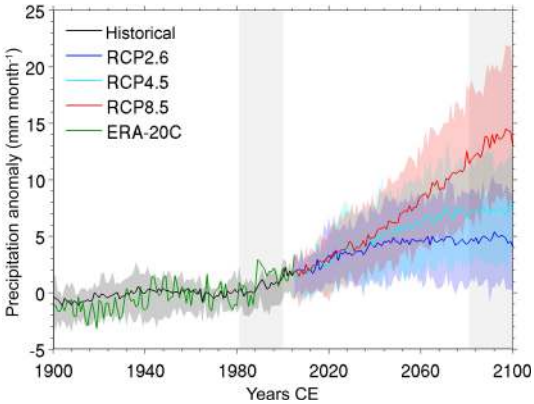


**Figure 1.** Annual precipitation anomaly (relative to the period 1961–1990) of the Arctic ($> = 60°$ N) derived from ensemble mean of historical (1900–2005) and RCP (2006–2100) simulations using 12 CMIP5 climate models (Taylor et al., 2012). The green line shows the annual precipitation anomaly derived from the ERA-20C reanalysis dataset (Poli et al., 2013). Solid lines represent multi-model ensemble means, while shading around the solid lines represents the interquartile ensemble spreads (25th and 75th quartiles). The vertical light grey shading marks the time period of 1981–2000 (left) and 2081–2100 (right).

the North Pacific (Fig. 2a). Annual precipitation decreased in western Eurasia and locally over eastern Greenland and Svalbard. The CMIP5 models show a similar pattern, although the increase is much lower and more spatially homogenous (Fig. 2b), with slightly more prominent increases over parts of the North Atlantic and Barents Sea and decreases in areas south-west and south-east of Greenland.

For future hydroclimate projections, 36 CMIP5 simulations (12 for each of the Representative Concentration Pathways: RCP 2.6, RCP 4.5, and RCP 8.5 scenarios) for the period 2006–2100 were used. Multi-model ensemble means represent robust projections of the temporal variation, spatial patterns, and seasonal cycle of the historical and future annual precipitation variability over the Arctic region. Simulations with all RCPs indicate an increase in mean annual precipitation in the coming century, ranging from 4 mm (RCP 2.6 ensemble mean) to 14 mm (RCP 8.5 ensemble mean; Fig. 1). To obtain the spatial pattern of annual precipitation changes, the spatial pattern changes were first calculated based on individual model simulations and then re-gridded to the same spatial resolution as the GFDL-CM3 model (i.e. 144 longitudinal grid cells $\times$ 90 latitudinal grid cells). The re-gridded spatial pattern changes based on the individual model simulations were then averaged to create a multi-model ensemble mean.

The most prominent increases in annual precipitation will occur over the Barents Sea, western Scandinavia, eastern Eurasia, and western North America, with decreased precipitation over the central parts of the North Atlantic (Fig. 2c to e). Moreover, the model simulations suggest an intensified precipitation cycle with increases in all months, but more prominently outside late spring–early summer (Fig. 2f and g).

## 2.2 Impacts of Arctic hydroclimate change

### 2.2.1 Impacts on Arctic environments

Changes in hydroclimate will have impacts on Arctic terrestrial and marine environments, including the cryosphere and the Arctic Ocean (ACIA, 2005). Observational studies show evidence of increased precipitation and river discharge in the Arctic, and hence a freshening of the Arctic Ocean, over the last decades (e.g. Peterson et al., 2006; Min et al., 2008). The freshening will have impacts on ocean convection in the subarctic seas, influencing the thermohaline circulation (THC, see below; Min et al., 2008). Increased ocean freshening will also have implications for marine flora and fauna distribution due to altered light and nutrient conditions (Carmack et al., 2016). Planktonic primary producers are likely to be affected, some positively and some negatively, and these impacts may cascade up the food web and alter the whole marine ecosystem structure (Li et al., 2009), thereby affecting marine biodiversity. Overall, changes in landscape and biophysical properties, biogeochemical cycling, and chemical transport associated with warmer and wetter conditions will influence ecosystem productivity (e.g. Wrona et al., 2016). Impacts on ecosystems will also affect the Arctic's indigenous populations, e.g. through increased risks to infrastructure and water resource planning (Bring et al., 2016), health (Geer et al., 2008), and subsistence-based livelihoods (Ford et al., 2014). As an example of the latter, increased occurrences of rain events during the cold season, causing the formation of ground ice and preventing winter grazing, will have negative impacts on herbivores, such as reindeer (Stien et al., 2012).

### 2.2.2 Remote impacts

In general, snow cover in the pan-Arctic region has decreased over the last several decades (Screen and Simmonds, 2012; Shi et al., 2013), although for snow on sea ice knowledge is limited due to the effects of regional variability and a lack of direct observations. This has been attributed to elevated temperatures and an increasing fraction of rain relative to snow, but also to the effects of increasing evaporation from the ocean due to the receding sea ice pack (Bintanja and Selten, 2014). In addition to the local effects described above, changes in snow cover, especially during autumn–winter, affect the atmospheric circulation, and this can have remote impacts on the hydroclimate of lower latitudes. For example, Cohen et al. (2012) suggested that a warmer and wetter Arctic atmosphere during autumn, caused by decreasing sea ice coverage, regionally favours increased snow cover in the same season, which dynamically forces negative Arctic Oscillation (AO) conditions in the subsequent winter. The

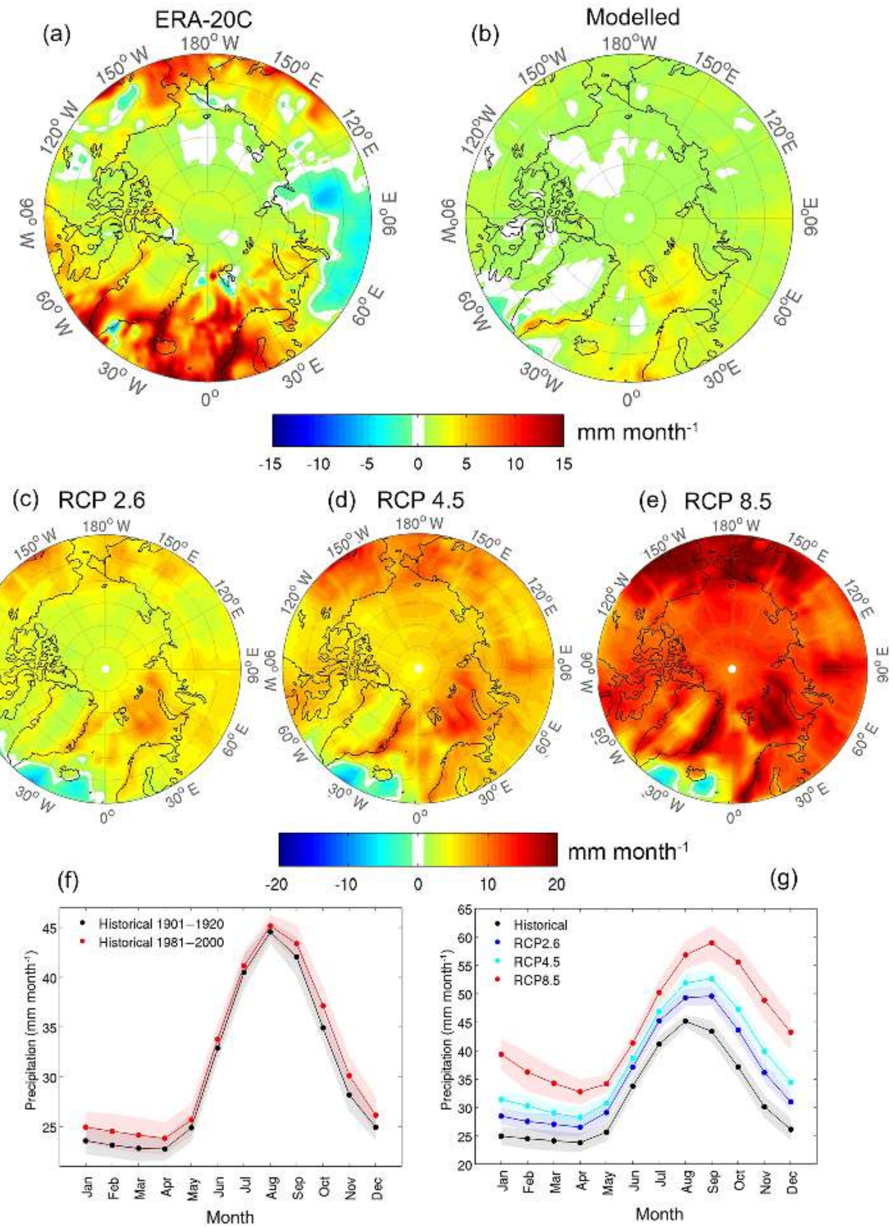


**Figure 2.** Simulated and observed Arctic precipitation. Upper panels: **(a)** observed and **(b)** modelled changes in annual precipitation over the Arctic region (> = 60° N) relative to the reference period 1901–1920 and averaged over the 1981–2000 period. The observed pattern is obtained from the ERA-20C reanalysis dataset (Poli et al., 2013). The modelled pattern is derived from ensemble mean of historical (1900–2005) simulations performed by 12 CMIP5 climate models (Taylor et al., 2012). Middle panels: multi-model (12 CMIP5 models) average changes in annual precipitation relative to the reference period 1981–2000 averaged over the period 2081–2100 under RCP 2.6 **(c)**, RCP 4.5 **(d)**, and RCP 8.5 **(e)** forcing scenarios. Lower panels: multi-model (12 CMIP5 models) average of seasonal cycle of precipitation over the Arctic (> = 60° N) for the periods of 1901–1920 (black) and 1981–2000 (red) **(f)**, and for 1981–2000 (black) and 2081–2100 (other colours) **(g)**. Solid lines represent multi-model ensemble means, while shading around the solid lines represents uncertainties expressed as ±2 standard deviations of the mean monthly precipitation over a 20-year period.

negative phase of the AO is associated with a more meridional flow of the jet stream, which allows cold Arctic air to penetrate into lower latitudes, occasionally yielding extreme weather events (Overland et al., 2016).

A distinct decline in sea ice extent and thickness has been observed in the past decades (Stroeve et al., 2012; Kwok and Cunningham, 2015). The melting of Arctic sea ice has local influences, but recent research suggests that it may also have remote impacts on mid-latitudes by perturbing local energy fluxes at the surface and modifying the atmospheric and oceanic circulation (e.g. Budikova, 2009; Francis et al., 2009). Variations in Arctic sea ice extent influences the North Atlantic Oscillation (NAO; Pedersen et al., 2016), which has a strong influence on precipitation in the North Atlantic region (Hurrell, 1995; Folland et al., 2009). Wu et al. (2013) suggested that winter Arctic sea ice concentration may be a precursor for summer rainfall anomalies over northern Eurasia and Guo et al. (2014) noted a link between spring Arctic sea ice conditions and the summer monsoon circulation over eastern Asia. On the other hand, it is also likely that lower-latitude phenomena influence Arctic sea ice conditions. For example, wintertime sea ice loss has been linked to different phases of the Pacific Decadal Oscillation (PDO; Screen and Francis, 2016).

Enhanced precipitation and melting of the cryosphere increases the run-off from the pan-Arctic land areas and lowers the salinity of the Arctic Ocean, and this will likely have significant impacts at a local and potentially global scale (Serreze and Barry, 2011; Rhein et al., 2013; Carmack et al., 2016). Since the density of the water in the Arctic Ocean determines the location of the thermocline and haloclines, changes in salinity may influence the distribution patterns of organisms and biogeochemical properties (Aagard and Carmack, 1989; Carmack et al., 2016). Moreover, salinity regulates the density of the water in the Arctic Ocean, and through outflow of Arctic water into the North Atlantic it can impact regions at lower latitudes, e.g. by affecting deep water formation in the Greenland–Norwegian and Labrador seas and thus the strength of the THC (Aagard and Carmack, 1989; Rahmstorf, 1995; Slater et al., 2007). A disruption of the THC could have global impacts (Vellinga and Wood, 2002). Density also determines the location of the thermoclines and haloclines so that salinity shifts greatly the influence distribution of organisms (Aagard and Carmack, 1989; Carmack et al., 2016).

## 3 Hydroclimate archives and proxies in the Arctic

While most archives and proxies that are widely used elsewhere to infer past climate variability can be found in the Arctic, their application require specific treatment and interpretation. The following section describes and discusses the characteristics and the limitations of these in the Arctic environment.

### 3.1 Lake sediments: varves and biomarkers

#### 3.1.1 Arctic lakes

Most lakes in the Arctic were formed just after local retreats of ice sheets, glaciers, and ice caps after the last glaciation; hence, their ages and the potential lengths of the records they contain range from the entire Holocene in Beringia and Scandinavia to only a few hundred years in Greenland or Iceland. The last 2000 years has, in general, been characterised by minor glacier fluctuations prior to the general melt of the recent decades (Solomina et al., 2016). This relative stability in surface area over the last 2 millennia makes lakes excellent recorders of hydroclimate variability for this period. What makes the lakes different in the Arctic is the very strong seasonality that is reflected in a long to very long ice cover period. A long ice cover season substantially reduces the input of particles from the watershed to the lake, frequently to an unmeasurable quantity. Therefore, what is recorded in Arctic lacustrine sediments is strongly biased towards the ice-free periods, i.e. spring snowmelt, short summer, and early autumn. Another characteristic of the Arctic is physical weathering related to gelifaction and sparse vegetation cover, making large quantities of easily eroded minerogenic matter available to be transported into lakes (Zolitschka et al., 2015).

Lake systems in the Arctic differ depending on the presence or absence of glaciers in their watershed. Snow-fed watersheds experience maximum discharge during snowmelt in spring. They become depleted in water once the snow cover has melted, reducing sediment transport in the latter part of the ice-free season. On the other hand, glacierised watersheds are not water limited; i.e. the water supply to the lake tributary can last the entire summer and autumn until temperatures decrease below zero. Discharge, and therefore sediment transport, is usually driven by temperature at the elevation of the glacier and is usually at a maximum during summer. In addition, lake systems in glacierised watersheds may on rare occasions be subjected to catastrophic floods (called Jökulhlaups), which are due to collapsing ice dams retaining vast amounts of water in intra- or supra-glacial lakes, and resulting in high sediment fluxes to the lakes.

Many watersheds in the Arctic are also affected by the presence of permafrost. During summer, the permafrost thaws in its upper part (active layer), leaving sediment easily mobilised by small amounts of rainfall. This increases the risk of slope detachments and can result in debris flows or very high sediment yields in lake tributaries (Lewis et al., 2005). The presence of permafrost also makes the dating of lacustrine sediments difficult because organic matter can be stored in the soils for a long period prior to being transported to the lakes (Abbott and Stafford Jr., 1996).

In the high Arctic, sources of organic matter in lake sediments are both allochthonous and autochthonous, i.e. produced in the watershed or the lake. The relative contribution of these sources may in part be controlled by climate (Outridge et al., 2017), although the allochthonous organic matter remains dominant, and the total amount preserved remains low (Abbott and Stafford Jr., 1996; Gälman et al., 2008). Conversely, lakes located in the southernmost part of the Arctic, such as in the boreal forests of Scandinavia or North America, experience a season with higher primary productivity. Their total organic carbon content can be relatively high

(Gälman et al., 2008) when anoxic conditions at the bottom of the water column prevail, slowing down its degradation.

#### 3.1.2 Extracting hydroclimatic information from Arctic lakes

Most of the proxies used elsewhere in the world for the purpose of reconstructing past hydroclimate can also be analysed in Arctic lakes. Extensive experience has enabled their use in the Arctic in spite of the harsh nature of the environment.

*Pollen*. Pollen can successfully be used to reconstruct precipitation because the response of plants to moisture changes is direct and well studied. Although a substantial proportion of the pollen in the high Arctic arrives from forested regions to the south, pollen assemblages can still be used to reconstruct the local conditions (e.g. Gajewski, 2002, 2006, 2015b). The sediments may be contaminated by older pollen stored in the soils or, in some cases, from Tertiary deposits in the watershed (Gajewski et al., 1995). Nevertheless, annual precipitation has been reconstructed, along with temperature (Gajewski, 2015a), using pollen assemblages and are presented in Sect. 4.2.

*Chironomids*. In the Arctic, chironomids are primarily affected by lake depths, temperature and water chemistry (Gajewski et al., 2005). Provided that changes in precipitation regime affect the depth of a lake and the pH or nutrient supply, chironomids can be used (Medeiros et al., 2015). However, most work has emphasised the reconstruction of temperature, and there would need to be relatively large changes in depth to have a noticeable effect on the chironomid community (e.g. Barley et al., 2006; Fortin et al., 2015).

*Diatoms*. These can presumably be used to reconstruct past moisture through various indirect methods. The primary control on diatoms is pH (Finkelstein et al., 2014), and to the extent this is affected by lake level variations, it could be used as an indirect proxy. Lake level changes affecting the relative area of deep and shallow water can be registered by diatoms; these have been used in other regions but not in the Arctic. A study of stable oxygen isotopes in diatom frustules allowed for a palaeohydrological reconstruction from Baffin Island, Canada (Chapligin et al., 2016).

*Hydrogen isotopes and biomarkers*. The source of environmental water in terrestrial systems is precipitation. Precipitation $\delta$D values are influenced by the location, temperature, and relative humidity of the primary evaporation source of the moisture, the air mass trajectory, and the temperature at condensation (Dansgaard, 1964; Boyle, 1997; Pierrehumbert, 1999; Masson-Delmotte et al., 2008; Frankenberg et al., 2009; Theakstone, 2011; Sjolte et al., 2014). Evaporative enrichment can cause environmental water, including lake water, soil moisture, and leaf water, to become D-enriched relative to the original precipitation. Thus, lake-sediment-based lipid $\delta$D records can provide important insights into the variability of both precipitation $\delta$D values and evaporative enrichment and thus ultimately into local hydrological changes. There are only a handful of published studies using $\delta$D values of leaf waxes and algal lipids to reconstruct past hydrological changes in the Arctic (Thomas et al., 2012, 2016; Balascio et al., 2013, 2017; Moossen et al., 2015; Keisling et al., 2017). The palaeohydrological interpretations of these $\delta$D records differ among the studies, reflecting the fact that different lake catchments respond differently to hydrologic changes (and over different timescales), but also highlights our incomplete understanding of the biological and environmental factors that influence hydrogen isotope variability in lipids. Palaeohydrologic interpretations are better constrained when lipids with $\delta$D values representing lake water (e.g. those derived from algae and macrophytes) are considered together with those representing leaf water (e.g. long-chain $n$-alkanes and long-chain $n$-alkanoic acids; Balascio et al., 2013, 2017; Rach et al., 2014; Muschitiello et al., 2015; Thomas et al., 2016). Together, $\delta$D values of these compounds can be used to quantify isotopic differences between lake water and leaf water, which can reveal changes in the duration of summer ice cover (Balascio et al., 2013), seasonality of precipitation (Thomas et al., 2016), or the vegetation type contributing lipids to the lake sediments (Balascio et al., 2017). Rach et al. (2017) propose an approach using paired terrestrial and aquatic lipid $\delta$D values and plant physiological models to quantitatively reconstruct relative humidity changes through time. This approach may prove effective in some Arctic settings.

Several physical and geochemical proxies have been used to infer past hydrology. *Mass accumulation rate* (MAR) is a measure of the amount of sediment accumulated at the bottom of the lake (e.g. Weltje, 2012). It is usually directly linked to the lake tributary discharge in lakes with low primary productivity (Petterson et al., 1999). Obtaining MAR requires an accurate age model and measurements of density. *Density*, *magnetic susceptibility*, and *elemental composition* are all indicators of the detrital input, which is again linked to the lake tributary discharge (Petterson et al., 1999; Dearing et al., 2001; Cuven et al., 2010). The *grain size* of the terrigenous fraction is an indicator of the competence of the flow (maximum discharge), its duration, and physical processes occurring in the lake water column (Lapointe et al., 2012). Together, these physical and geochemical proxies are rarely used in Arctic sedimentary sequences with massive structure because of the complexity of their interpretation; however, they have proven to be useful tools in annually laminated sediments.

#### 3.1.3 Varved sediments

Varved sediments are difficult to find and probably rarely deposited (Zolitschka et al., 2015). However, several lakes with varved sediments have been found in the Arctic, probably because the very strong seasonal contrast in sediment supply favours the formation of varves. Lakes containing varves tend to be deep enough to prevent bioturbation and are usually found in watersheds with high sediment yield. As such, many of the varved records cannot be directly compared to lakes used in diatom and pollen studies because the latter are usually studied in smaller systems. The advantages of varved sediments are that they contain their own chronology, that annual fluxes can be measured through the measurement of density, and that their properties can be calibrated against instrumental records (Hardy et al., 1996). In the Arctic, two types of varves exist: clastic varves and mixed clastic–biogenic varves, as discussed in Zolitshcka et al. (2015).

### Clastic varves

Clastic varves result from the complex interactions between sediment availability (geomorphological control), seasonal run-off variations carrying suspended sediment (hydroclimatic control), the thermal density structure of the lake water column, and the bathymetry of the lake (limnological control). These varves are typically composed of a coarse-grained lower lamina that grades into a fine-grained upper lamina (e.g. Lake DV09; Courtney-Mustaphi and Gajewski, 2013; Lake C2; Zolitschka, 1996). Additional coarse-grained laminae can be deposited and can be related to multiple pulses of snowmelt or rain events (Ringberg and Erlström, 1999; Cockburn and Lamoureux, 2008). The finest clay fraction remains in the water column and is only deposited under quiet conditions during the following winter (Francus et al., 2008). Therefore, the presence of a distinct clay cap is the main criteria for identifying a year of sedimentation (Zolitschka et al., 2015).

Several individual parameters can be measured from each varve sequence: total thickness, sublamina thickness, density, mass accumulation rate, total and sublamina grain size, elemental composition, and magnetic susceptibility. Linking these properties with hydroclimate conditions requires monitoring the processes occurring in the watershed and the lake, with each system being different. Disentangling the respective effect of the temperature from moisture is a challenge due in part to the difficulty in obtaining data for calibration in the Arctic. When comparing varve properties to observational climate data, they often contain signals of both temperature and precipitation (e.g. Table 2 of Cuven et al., 2011; Lamoureux and Gilbert, 2004), although the temperature signal has been reported more often in the literature. However, this may be because more robust measurements of instrumental temperature are available compared to precipitation (especially snow) and precipitation patterns tends to be more variable over a region, making correlation with sediment properties more difficult.

Despite these difficulties, several authors reported correlations of varve sequence data with hydroclimate. In general, the hydroclimate is revealed in the measurement of a specific part of the sedimentary cycle, and not by a parameter that integrates the whole year of sedimentation such as the total varve thickness. For instance, Lapointe et al. (2012) showed a correlation ($r = 0.85$, $p = 0.0001$) between the largest rainfall events and the coarsest grain-size fraction of each varve. Lamoureux et al. (2006) found a correlation between varve thickness of Sanagak Lake, Boothia Peninsula and snow-water equivalent in the watershed, but they were unable to calibrate the series due to lack of calibration data. Francus et al. (2002) found a correlation ($r = 0.53$, $p < 0.05$) between snowmelt intensity and the median grain size. Lamoureux (2000) found an association of sediment yield estimates of Nicolay Lake, Cornwallis Island and rainfall events.

### Mixed (clastic–biogenic) varves

In less harsh environments, such as in central Scandinavia, the vegetation of the catchment area and soils is more developed, allowing decaying organic matter to be incorporated into the lacustrine system. At the same time, the primary productivity in the water column during the warmer seasons is large enough to be recorded in the sedimentary archive. This results in the accumulation of a mixed type known as clastic–biogenic (or clastic–organic) varves. These typically contain a characteristic minerogenic lamina, usually showing graded bedding that is directly related to the duration and strength of the spring flood (e.g. Ojala et al., 2000; Snowball et al., 2002; Tiljander et al., 2003) and a biogenic lamina that can be composed of autochthonous organic matter (e.g. diatom frustules) and/or allochthonous organic debris.

Proxies measured with annual resolution on these mixed varves include (1) total varve thickness, (2) growing season lamina (GSL) thickness, (3) winter lamina (WL) thickness (Saarni et al., 2015), and (4) relative X-ray densitometry (Ojala and Francus, 2002). Correlations with climate parameters vary from site to site and sometimes through time at a single site (Saarni et al., 2015). Only a small number of lacustrine sequences, all of them from Scandinavia, have been successfully correlated with precipitation or moisture. At Lake Nautajärvi annual and winter precipitation was reconstructed using relative X-ray densitometry (Ojala and Alenius, 2005), whereas at Lake Kallio-Kourujärvi, the growing season lamina was linked to annual precipitation (Saarni et al., 2015). Rydberg and Martinez-Cortizas (2014) showed that high accumulation of snow resulted in high mineral matter content, and Wohlfarth et al. (1998) found a significant correlation between early spring–summer precipitation and total varve thickness in north-central Sweden.

As with clastic varves, it is quite difficult to separate the temperature from the moisture signal. Ojala and Alenius (2005) showed that direct annual and seasonal comparisons between raw varve data and instrumental measurements are complicated. Itkonen and Salonen (1994) showed that total varve thickness of three Finnish lakes were correlated with both temperature and precipitation, the correlation being weaker for precipitation. Nevertheless, sediment trap studies clearly but qualitatively showed the sensitivity of such systems to varying hydroclimate conditions (Ojala et al., 2013; Rydberg and Martinez-Cortizas, 2014).

## 3.2 Peat deposits

### 3.2.1 Peatland processes and peat archive

Peatlands are wetland ecosystems that preserve their developmental history over millennia. Peat deposits are products of the balance between plant production and organic matter decomposition (Clymo, 1984), and both processes are affected by climate. As a result, peat accumulation is inherently influenced by autogenic–ecological and allogenic–climatic factors, as well as their interactions (Belyea and Baird, 2006). Many peat-based proxies (see below) have been used to reconstruct peatland hydrology and water table dynamics, likely connected with regional hydroclimate. This ability of wetland communities to record hydrological change results largely from their occurrence in environments where a single extremely variable habitat factor, i.e. water supply, is predominant (Tallis, 1983). However, empirical and modelling studies show the importance of autogenic process and eco-hydrological feedbacks (e.g. Tuittila et al., 2007; Swindles et al., 2012; Loisel and Yu, 2013; Väliranta et al., 2016). Clearly, consideration of biological processes and ecological feedbacks is needed when using these systems for climate reconstructions.

Peatland plants shape their own habitat since they form their own growth substrate: peat. Hence, peatlands are capable of recording in their deposits the effects of past vegetational and ecological changes. Within the peat lies a repository of botanical, zoological, environmental, and biogeochemical information, which is important for understanding past climatic conditions. These palaeorecords are used to estimate the rates of peat formation or degradation, past vegetation, climatic conditions, and depositional environments (Moore and Shearer, 1997; Blackford, 2000). Analysis of peat deposits has undergone major developments during the last several decades regarding coring techniques, peat sampling and analysis, geochronology, identification of plant remains and other microfossils, and quantitative multivariate techniques (e.g. Barber et al., 1994, 1998; Väliranta et al., 2007; Charman et al., 2009; Chambers et al., 2011; Mathijssen et al., 2016, 2017).

Stratigraphic studies in peatlands have shown a hydroseral succession in which wet swamp and fen communities gradually develop into dry bog communities (Tallis, 1983; Korhola, 1992; Väliranta et al., 2016). These changes are largely autogenic, connected to the growth of wetland communities, and caused by climatic variability or artificial drainage. Hilbert et al. (2000) developed a model of peatland growth that explicitly incorporates hydrology and feedbacks between moisture storage and peatland production and decomposition. They suggest that drier ombrotrophic peatlands (most bogs) will adjust relatively quickly to perturbations in moisture storage, while wetter ombrotrophic peatlands (mineral-rich fens) are relatively unstable and can withstand only very small changes in water tables (Mathijssen et al., 2014). Climate change will affect the hydrology of individual peatland ecosystems mainly through changes in precipitation and temperature. As the hydrology of the surface layer of a bog is dependent on atmospheric inputs (Ingram, 1983), changes in the ratio of precipitation to evapotranspiration may be expected to be the main factor driving ecosystem change. In particular, ombrotrophic peatlands are regarded as directly coupled to the atmosphere through precipitation and hydrology change (Barber et al., 1994) such that their water levels and dominant plants will reflect the prevailing climate. More specifically, their water table variability has been shown to be highly correlated with the total summer seasonal moisture deficit (precipitation–evapotranspiration; Charman, 2007).

Modern investigations of past climate are performed with an emphasis on obtaining the highest possible time resolution for a given archive. Radiocarbon dating is one of the main methods used to establish peat chronologies. The best materials for ensuring accurate dates are aboveground remains of plants that assimilated atmospheric $CO_2$, e.g. short-lived plant macrofossils and pollen whose $^{14}C$ age is consequently not affected by an old carbon effect. Suitable materials for sample selection are *Sphagnum* mosses (branches, stems, and leaves) or, if not present, aboveground leaves and stems of dwarf shrubs (e.g. Nilsson et al., 2001; see however Väliranta et al., 2014). Age–depth models that are considered ecologically plausible and that take into account likely modes of peat accumulation include (1) linear accumulation, (2) concave curves (caused by continuing decomposition of fossil matter in the peat deposit; Yu et al., 2001), (3) convex curves (with deposits slowing their accumulation when approaching a height limit; Belyea and Baird, 2006), and (4) Bayesian models that can include prior information on stratigraphy, accumulation rate and variability, and/or detect outlying dates (reviewed by Parnell et al., 2011). The robustness of age models can be significantly improved and the uncertainties reduced by using multiple dating methods on a single core. Most commonly, the uppermost layer can be dated using atmospheric fall-out radionuclides (e.g. $^{210}Pb$; see Le Roux and Marshall, 2011) and spheroidal carbonaceous particle (SCP) profiles (Yang et al., 2001), while tephrostratigraphy can potentially be applied throughout the core (Swindles et al., 2010). With suitable statistical treatment, all results can be combined into one reliable chronology which provides the backbone for interpretations of palaeoclimatic and palaeoenvironmental change data.

### 3.2.2 Peat-based hydroclimate proxies

Peatland formation can initiate via three processes: primary peatland formation, terrestrialisation, or paludification (Rydin and Jeglum, 2006). In primary peatland formation, peat is formed directly on wet mineral soil when the land is newly exposed due to crustal uplift or deglaciation, whereas in terrestrialisation and paludification the area colonised by peatland vegetation has experienced previous sediment deposition or soil development (e.g. Tuittila et al., 2013). Information on hydroclimatic conditions can be derived from these processes, especially when the different peat formation types show systematic and isochronic patterns over wide geographic areas. For example, in paludification, the prerequisite is that the local hydrological conditions become wetter, for instance induced by climatic change, fire, or beaver damming, resulting in waterlogged soil conditions that promote peat accumulation (Charman, 2002; Gorham et al., 2007; Rydin and Jeglum, 2006). A new conceptual model of episodic, drought-triggered terrestrialisation presents infilling as an allogenic process driven by decadal to multidecadal hydroclimatic variability (Ireland et al., 2012).

Recently, Ruppel et al. (2013) presented a comprehensive account of postglacial peatland formation histories in North America and northern Europe using a dataset of 1400 basal peat ages accompanied by below-peat sediment-type interpretations. Their data, mainly focusing on boreal–Arctic regions, indicates that peat formation processes exhibited some clear spatiotemporal patterns. Unfortunately, the overwhelming majority of the basal peat accounts originate from the deepest and often the oldest parts of peatlands, and therefore the last 2 millennia are clearly under-represented in the present data. However, existing studies illustrate the potential of using peat initiation and expansion data to account for changes in regional moisture regimes, also in more recent times. The formation of new peatland areas does not necessarily decrease when the initiation rates decrease, but new peatland areas are continuously formed via lateral expansion.

Peat bulk density (or organic matter density) in downcore profiles has been used to reflect overall peat decomposition, which in many peatland regions is controlled by surface moisture and hydroclimate conditions (e.g. Yu et al., 2003). The rationale is that well-preserved peat is loose and has low organic matter density, most likely deposited under wet conditions promoting the protection of organic matter in an anaerobic environment. Peat bulk density values are typically 0.05 to 0.2 $g\,cm^{-3}$ in high-latitude regions (Chamber et al., 2011). Peat humification is another proxy for the degree of peat decomposition which can be estimated or measured in the field or laboratory using a range of methods (Chamber et al., 2011). Humification can be used as a proxy for peatland surface wetness, as moisture is a key determinant of decomposition, and regional hydroclimate in the Arctic (e.g. Borgmark, 2005; Borgmark and Wastegård, 2008; Vorren et al., 2012).

Net carbon accumulation is the balance between production and decomposition, both of which are influenced directly or indirectly by climate (Yu et al., 2009). However, recent syntheses indicate that temperature-driven production might be more important than moisture-controlled decomposition in determining net peat accumulation (e.g. Beilman et al., 2009; Charman et al., 2013). Therefore, without constraints from other proxies, it is difficult to infer hydroclimate from peat accumulation records (Mathijssen et al., 2016, 2017). As mosses are the dominant plants in peatlands, carbon isotopes from these mosses are useful for inferring peatland moisture conditions. In wet conditions, water films around moss leaves will reduce the conductance of pores on the leaf surface to $CO_2$ uptake, reducing discrimination against $^{13}C$ and resulting in high carbon isotope values (Rice, 2000). Carbon isotopes have been shown to reflect surface moisture in peatlands (e.g. Loisel et al., 2009). In addition, Nichols et al. (2009) used compound-specific carbon and hydrogen isotopes from peatlands in the Arctic to evaluate summer surface wetness and precipitation seasonality.

Because plant macrofossils reflect changing abundances of climatically sensitive peatland vegetation, they have been used not only for reconstructing the local vegetation history of peatlands but also for inferring past peatland hydrological changes and, by extension, regional climate variability (e.g. Barber et al., 1998; Hughes et al., 2000; Swindles et al., 2007; Väliranta et al., 2007; Mauquoy et al., 2008; Mathijssen et al., 2014, 2016, 2017). Traditionally, plant-based peatland surface wetness reconstructions have been qualitative or semi-quantitative based on the identification of phases of relatively low local water tables (showing increased representation of hummock species) and phases of higher local water table depths (lawn and hollow species; Mauquoy et al., 2002; Pancost et al., 2002; Sillasoo et al., 2007). More recently, ordination techniques (e.g. PCA and DCA) have been used to create a single index of peatland surface wetness based on the total subfossil dataset for a peat profile (Barber et al., 1994; Mauquoy et al., 2004; Sillasoo et al., 2007; Zhang et al., 2017) in which it is assumed that the principal axis of variability in the dataset is linked to hydrology. The most recent progress in identification and quantification techniques of plant macrofossils (e.g. the Quadrat Leaf Count method; Mauquoy et al., 2010), together with careful calibration with modern plant community data, allows for the quantification of past peatland water table fluctuations with great accuracy. Väliranta et al. (2007) developed a transfer function by calibrating plant macrofossil records against the modern vegetation–water table relationship in order to quantitatively reconstruct peatland surface wetness trends for the late Holocene. The inferred water tables showed strong fluctuations, with an overall amplitude of ca. 40 cm. During the last 2 millennia, they found generally dry conditions until ca. 1600 cal BP (ca. 400 CE), varying water tables during the following 4 centuries, and dry conditions from ca. 1200 to 700 cal BP (ca. 800–1300 CE, covering the MCA). The subsequent centuries were again variable, while the period 500–200 cal BP (1500–1800 CE, covering the LIA) was wet and the last 2 centuries dry except for the very recent years. The comparison of water table reconstructions based on macrofossils and testate amoebae at two bogs in Estonia and Finland increased the confidence in using bog plants in quantitative hydrological reconstructions (Väliranta et al., 2011).

Testate amoebae (Protozoa: Rhizopoda) are unicellular animals with distinct environmental preferences which live in abundance on the surface of most peat bogs. These amoeboid protozoans produce morphologically distinct shells, which are commonly used as surface moisture proxies in peat-based palaeoclimate studies (Mitchell et al., 2008). Although the moisture sensitivity of these organisms has been known for a long time, work over the past several decades has demonstrated the utility of testate amoebae as quantitative peatland surface moisture indicators. Their indicator value in documenting surface moisture variation has been demonstrated by coherence in reconstructions of wet and dry fluctuations within and between peatland sites (Hendon et al., 2001; Booth et al., 2006). A protocol of their use in palaeohydrological studies is provided by Charman et al. (2000) and Booth et al. (2010). Testate amoebae have been used for tracing hydrological changes in temperate peatlands in several regions of the world, as well as in the boreal and subarctic peatlands of Canada and the US (Payne et al., 2006; Loisel and Garneau, 2010; van Bellen et al., 2011; Bunbury et al., 2012; Lamarre et al., 2012, 2013). In addition to bogs, they can also be applied in fens (Payne, 2011). Recently, Swindles et al. (2015) and Zhang et al. (2017, 2018) tested the potential of testate amoebae for peatland palaeohydrological reconstruction in permafrost peatlands based on sites in Arctic Sweden and Russia, respectively. These evaluations confirmed that water table depth and moisture content are the dominant controls on the distribution of testate amoebae in Arctic peatlands, corroborating the results from studies in mid-latitude regions. New testate-amoeba-based water table transfer functions were created with good predictive powers and the transfer functions were applied to short cores from permafrost peatlands. All records revealed a major shift in peatland hydrology, which in one case coincided with the onset of the Little Ice Age (Swindles et al., 2015). The new modern training sets will enable palaeohydrological reconstruction from permafrost peatlands in Northern Europe, thereby permitting a greatly improved understanding of the long-term hydrological dynamics of these ecosystems and the general variability in hydroclimatic conditions.

## 3.3 Tree-ring data

Distinct, precisely dateable tree rings are generally formed in areas with pronounced seasonality, which results in a single period of cambial activity (growth) and dormancy per calendar year. The width, density, and isotopic compositions of a tree ring are partly determined by local weather and climate, and the closer to the ecological limit of distribution a tree grows, the more sensitive to climate it will be. Due to the large spatial distribution of trees across extra-tropical regions, their capacity to live for many years and their potential for developing precise, annually resolved chronologies, tree-ring data have been widely used to infer late Holocene variations in a range of climate parameters on local to hemispheric scales.

### 3.3.1 Tree-ring width and density

Measurements of annual tree-ring width (TRW) are perhaps the most important data source for quantitative estimates of high- to low-frequency climate variability during the past centuries to millennia. The advantage of TRW comes from its annual resolution and a comprehensible understanding of the climatic controls on the tree-ring growth dynamics (e.g. Vaganov et al., 2006). Tree rings have the advantage of numerical calibration, verification, and the potential to capture seasonal extreme events not possible using lower-resolution, less temporally well-constrained archives. The tree-ring community has generated an expansive network of TRW chronologies covering a wide range of species and ecosystems across the globe, including the boreal–Arctic ecotone. In general, trees growing close to their latitudinal or altitudinal limit of distribution will be sensitive to warm-season temperature, while trees growing in semi-arid to arid regions are limited by precipitation and moisture (St George, 2014; St George and Ault, 2014; Hellman et al., 2016). Consequently, most but not all high-latitude TRW chronologies exhibit strong positive associations with summer temperature and only weak correlations with summer or winter rainfall. Tree-ring data from the high northern latitudes have been used in several reconstructions of Northern Hemisphere temperature (e.g. D'Arrigo and Jacoby, 1993; Jones et al., 1998; Briffa et al., 2001; Esper et al., 2002; D'Arrigo et al., 2006; Schneider et al., 2015; Stoffel et al., 2015; Wilson et al., 2016) and reconstructions targeting Arctic temperatures (Overpeck et al., 1997; Kaufman et al., 2009; Shi et al., 2012; Hanhijärvi et al., 2013; McKay and Kaufman, 2014). The few chronologies in the cool boreal and Arctic regions developed from precipitation-sensitive trees are mainly located in continental climate zones, such as western Canada, Alaska, and eastern Fennoscandia (see Fig. 1 in St George and Ault, 2014). Many TRW records are negatively correlated with summer rainfall, and most of these are found in the colder high-latitude regions. Positive correlations with prior-summer precipitation are also common across the Arctic. This carry-over effect may be caused by increased photosynthetic reserve accumulation in years with sufficient moisture supplying resources that can be used for secondary tissue growth in subsequent years. A proportion of this association likely reflects the inverse relationship between summer temperature and precipitation observed in these regions.

Although TRW is the most commonly used tree-ring proxy, at high latitudes, wood densitometric measurements, specifically maximum latewood density (MXD) and its surrogate blue intensity (BI), are commonly being viewed as superior temperature proxies compared to TRW. It would seem that the strong correlation between MXD–BI and temperature would prevent their use in hydroclimate reconstructions. However, recent studies (Cook et al., 2015; Seftigen et al., 2015a, b) have indirectly used high-latitude temperature-sensitive tree-ring data to reconstruct soil moisture availability by considering the inverse relationship between available soil moisture and clear skies, higher temperatures, increased evaporation, and reduced rainfall. Thus, the negative correlation between the high-latitude tree-ring data and drought metrics, such as the self-calibrating Palmer Drought Severity Index (scPDSI; van der Schrier et al., 2006a, b) and the Standardised Precipitation Evapotranspiration Index (SPEI; Vicente-Serrano et al., 2010), can be used to generate reconstructions that are comparable to those from arid and semi-arid regions where tree growth is strongly limited by rainfall. Many high-latitude MXD data that are mainly influenced by temperature also exhibit a negative, albeit weak, statistical association with summer precipitation (Briffa et al., 2002). This mixed response explains why such data have successfully been used in reconstructions of drought indices that integrate both temperature and precipitation.

### 3.3.2 Stable isotopes in tree rings

The isotopic ratios of wood, lignin, and tree-ring cellulose are influenced by a different and more limited range of environmental and physiological controls than TRW and MXD. For this reason, stable isotopes in tree rings provide additional palaeoclimate information to support and enhance the information attainable from the physical proxies (McCarroll and Loader, 2004; Gessler et al., 2014). Similar to TRW and MXD, the strength and relative expression of these climatic controls will vary geographically and to a degree with local edaphic conditions and tree species. In simple terms, carbon isotopic variability reflects changes in the balance between the conductance of carbon dioxide ($CO_2$) from the atmosphere to the site of photosynthesis and assimilation rate, which are influenced by moisture stress and photosynthetically active radiation (PAR), respectively. Temperature and nutrient availability may also contribute to this signal through an influence upon the rate of chemical reaction and production of photosynthetic enzymes (Farquhar et al., 1982; Scheidegger et al., 2000; Hari and Nöjd, 2009). Oxygen and hydrogen isotopes are more closely related to the isotopic composition of the water used by the tree during photosynthesis, which may reflect a combination of moisture sources subsequently modified by the evaporative enrichment of leaf water (vapour pressure deficit and relative humidity) and plant physiological processes (Barbour et al., 2001; Danis et al., 2006; Treydte et al., 2014; Roden et al., 2000).

Since the earliest isotopic dendroclimatology studies conducted in the Arctic (Sonninen and Jungner, 1995; McCarroll and Pawellek, 1998; Waterhouse et al., 2000), several studies have made significant contributions to palaeohydrology (e.g. Waterhouse et al., 2000; Holzkämper et al., 2008, 2012; Sidorova et al., 2008, 2009; Porter et al., 2009). The combination of long-lived trees, robust dendrochronologies and excellent sample preservation both on land and in lakes have facilitated the development of several multi-centennial to millennial length isotopic records (Boettger et al., 2003; Kremenetsky et al., 2004; Sidorova et al., 2008; Young et al., 2010; Gagen et al., 2011; Porter et al., 2014; Loader et al., 2013). However, because moisture is rarely the dominant tree-growth-limiting factor across much of the Arctic region, there is a limitation to the hydroclimate information that can be reconstructed using the isotopic approach. Using a multi-parameter approach, several studies (Loader et al., 2013; Young et al., 2010, 2012; Gagen et al., 2011) provided sunshine–cloud estimates and were able to demonstrate large-scale shifts in the dominance of Arctic and maritime air masses over the northern Fennoscandian region during the LIA and MCA. Such multi-parameter studies are potentially very powerful as they help to develop testable hypotheses relating to the future response of the Arctic atmosphere and provide a foundation for developing a circumpolar isotope network to track changes in atmospheric circulation and its relationship to climate throughout the Common Era.

Reconstructions based upon oxygen and hydrogen isotopes have yet to reveal the same clear and stable correlations with instrumental data observed for carbon, but are likely to relate most closely to local and regional hydroclimate through their close link with stable isotopes in precipitation (Roden et al., 2000). The relative contributions of the isotopic signal from snowmelt and growing season precipitation used to form the tree rings is an area requiring investigation. Links between $\delta^{18}O$ and both moisture and temperature have been identified (Sidorova et al., 2009; Knorre et al., 2010). Further south, Hilasvuori and Berninger (2010) linked oxygen (and carbon isotopes) most strongly to cloud cover, with precipitation, relative humidity, and temperature exhibiting lesser correlations. Hydrogen isotopes did not correlate as strongly as oxygen or carbon, with the strongest statistically significant relationships being with precipitation. Seftigen et al. (2011) linked $\delta^{18}O$ to precipitation, but noted that this relationship was unstable through time, possibly due to changes in the atmospheric circulation. If the same close relationship observed between water isotope composition and tree-ring cellulose in mid-latitude regions (Danis et al., 2006; Labuhn et al., 2014; Treydte et al., 2014; Young et al., 2015) is confirmed in the Arctic, then the potential exists for developing long records of the isotopic composition of precipitation suitable for large-scale mapping of isotope climate (Hemming et al., 2007; Saurer et al., 2012; Young et al., 2015). Reconstructing the stable isotopic composition of precipitation will likely provide a more useful, more direct link to the global hydrological cycle “isotope climate” (Birks and Edwards, 2009; Bowen, 2010) than a statistical calibration of water isotopes developed against a measured (indirect) meteorological variable which may vary in the degree of its control across space or over time.

### 3.3.3 Tree-ring-based hydroclimate reconstructions

Tree-ring data have been used to locally estimate a variety of hydroclimate variables, such as precipitation, drought, streamflow, cloud cover, and snowpack (e.g. Stahle and Cleaveland, 1988; Waterhouse et al., 2000; Pederson et al., 2001; Meko et al., 2001; Woodhouse, 2003; Gray et al., 2003; Young et al., 2010; Gagen et al., 2011). Moreover, networks of tree-ring chronologies have been used to make spatial (or field) hydroclimate reconstructions (Nicault et al., 2007; Cook et al., 1999, 2004, 2010; Fang et al., 2011; Touchan et al., 2011; Hua et al., 2013). However, these studies have almost exclusively utilised tree-ring data from lower latitudes outside the Arctic region.

Within the Arctic, trees are naturally constrained to exist below the latitudinal treeline, extending as far as ca. 73° N in parts of central Siberia, and as noted above, the majority of the tree-ring data in the region come from temperature-sensitive trees. Still, with careful site selection, it is possible to find trees that are sensitive to moisture variability, and a few studies have inferred past precipitation variability using ring-width data. Indeed, a handful of reconstructions of local hydroclimate from the Arctic have been published. These have mainly focused on late spring–early summer precipitation. The longest record, and presently the most widely used high-resolution hydroclimate proxy for high-latitude Fennoscandia, comes from south-eastern Finland where Scots pine TRW was used to reconstruct annual May–June precipitation over the last millennium (Helama and Lindholm, 2003). Later an updated chronology from the same region was used to highlight the distinct and persistent "mega-drought" from the early 9th century to the 13th century CE (Helama et al., 2009). The same parameter was also reconstructed from Scots pine for east-central Sweden back to 1560 CE by Jönsson and Nilsson (2009). In North America, Pisaric et al. (2009) reconstructed the June precipitation of the North-west Territories, Canada using a TRW network. The only reconstruction of hydroclimate outside the growing season was presented by Linderholm and Chen (2005), who developed a 400-year-long winter (September–April) precipitation reconstruction with 5-year resolution based on Scots pine TRW data from west-central Scandinavia.

Focusing on hydroclimate field reconstructions, one of the earliest works to use tree rings to reconstruct past moisture variability in a high-latitude region was the North American Drought Atlas (NADA; Fig. 3a). The atlas was first released in 2004 (Cook et al., 2004), then covering the continental US and later updated (Cook et al., 2007) with an expanded tree-ring network to include parts of the Canadian Arctic. Although significant portions of the latter region are at present under-represented in NADA, the tree-ring coverage still provides valuable hydroclimate reconstructions for a number of regions. The summer PDSI reconstruction data for the Arctic part of NADA extend back to the 1000 CE, indicating slightly drier conditions during most of the MCA, except for a wet period in the 12th century, and a highly variable LIA albeit with a tendency for progressively wetter summers before the early 19th century (Fig. 3c). Two efforts have used extensive tree-ring data networks to infer past drought–pluvial variability for Fennoscandia (Seftigen et al., 2015a, b) and Europe (The Old World Drought Atlas, OWDA; Cook et al., 2015, Fig. 3a). These atlases, in which tree-ring data were used to create gridded (field) reconstructions of the SPEI (Seftigen et al., 2015b) and the scPDSI (Cook et al., 2015), included regions north of 60° N. These millennium-long reconstructions allow for detailed investigations of the MCA and the LIA. The MCA in continental Europe and southern Scandinavia was significantly drier that the LIA and the post-industrial period (1850–present, and the reconstruction suggests that the Arctic regions in Europe experienced a severe drought during this period; Fig. 3b), which is in agreement with the findings of Helama et al. (2009). Interestingly, the timing of the MCA drought seems to temporally coincide with multi-centennial droughts previously reported for large areas of North America (Cook et al., 2007), specifically in California and Nevada. This suggests a common forcing across the North Atlantic, likely related to the North Atlantic Oscillation (NAO) and/or Atlantic Ocean sea surface temperatures. However, the restricted temporal coverage of the high-latitude part of NADA does not provide an opportunity to compare hydroclimatic variability across the Arctic region during the MCA. Large-amplitude hydroclimatic variability is not only restricted to the MCA, as periods of dryness are recorded in the first half of the 15th century CE and in the 1750s–1850s and may not have been restricted to the Arctic (Cole and Marsh, 2006).

Another possible option to derive hydroclimate information from north of the treeline in the Arctic is the utilisation of annual growth rings from shrubs. For example, Zalatan and Gajewski (2006) presented a short *Salix alaxensis* growth-ring series from north-western Victoria Island in the Canadian Arctic. The width of the shrub rings was found to be correlated with winter precipitation. Although the reported record was too short to be useful for palaeoclimate studies, it may be possible to obtain longer series by using larger specimens (some are tree-sized in this area; Edlund and Egginton, 1984) or cross-dating dead and buried wood.

### 3.3.4 Pine regeneration patterns as indicators of hydrological shifts

In the high northern latitudes, tree remains can be preserved for several millennia buried in lakes or peat, which becomes so-called subfossil wood, and subfossils extracted from lakes have been used to reconstruct temperatures for large parts of the Holocene in Fennoscandia (see Linderholm et al., 2010, for a review). More or less well-preserved trees can also be

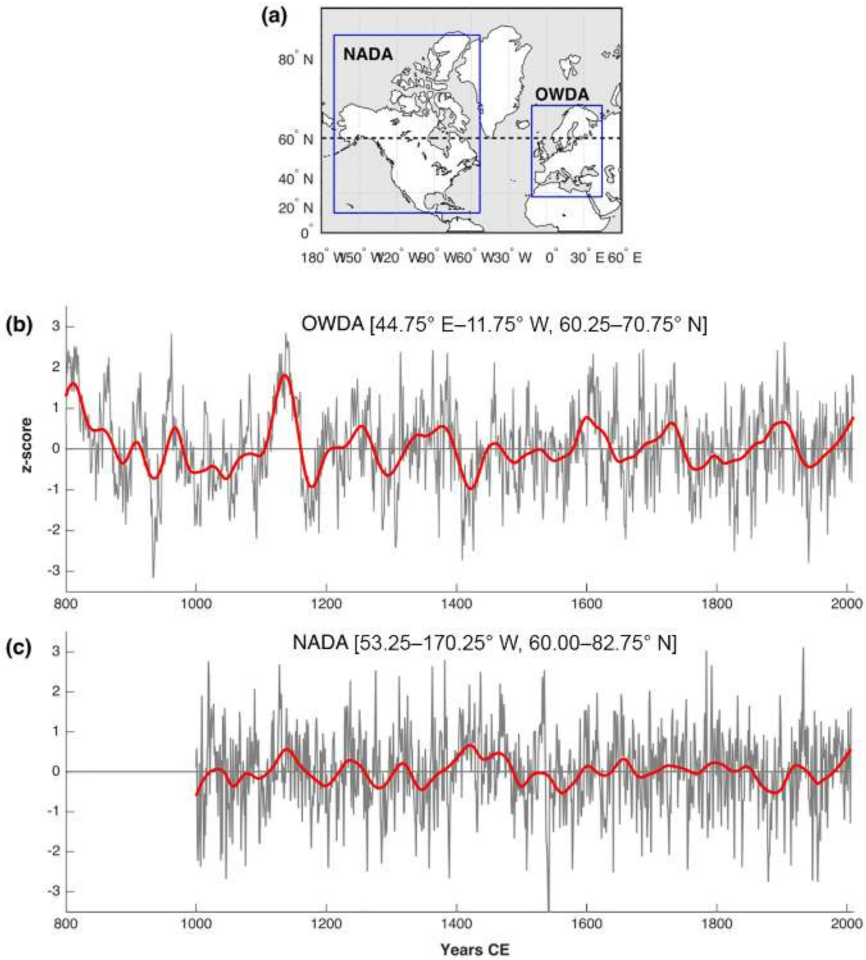


**Figure 3.** Drought atlas reconstructions over the Arctic for North America (NADA; Cook et al., 2004, https://www.ncdc.noaa.gov/paleo/study/6319, last access: 16 January 2017) and Europe (OWDA; Cook et al., 2015, https://www.ncdc.noaa.gov/paleo/study/19419, last access: 4 November 2015). **(a)** The full spatial domains of the two atlases and a regional average over latitudes > 60° N in **(b)** Europe and **(c)** North America transformed into $z$ scores and filtered with a 100-year loess (red lines).

found in dark layers of well-humified peat, an indicator of dry conditions having allowed trees to grow and to colonise the area (Gunnarson, 2008).

In west-central Sweden, more than 1000 subfossil and peatland Scots pine (*Pinus sylvestris* L.) samples have been collected since the late 1990s. Most samples come from different lakes at varying altitudes, and the temporal distributions of the dated samples show wave-like patterns of regeneration with clearly distinguishable mortality and germination phases. Such generation pulses have been related to climatic conditions favourable for seed production and successful germination, i.e. warm and dry periods (Zackrisson et al., 1995). However, Gunnarson (2008) suggested that the temporal variations of pine samples from both bogs and lakes (Fig. 4) reflect fluctuations in peatland groundwater tables and lake levels caused by regional changes in hydroclimate. It is likely that these variations have been governed by changes in precipitation rather than changes in temperature. In south-eastern Finland evidence of depositional histories of subfossil pines from lakes, where most trees have grown adjacent to or on lakeshores (so-called riparian trees), and peatland pines were combined by Helama et al. (2017a). Divergent depositional histories (i.e. replication curves) were obtained for the two environments during the Common Era. High accumulation of peatland pines during the MCA indicates dry surface conditions beneficial for pine colonisation (Torbenson et al., 2015; Edvardsson et al., 2016). This phase overlapped with a phase of low accumulation of riparian pine trees. In contrast, the accumulation of riparian pines increased towards the LIA, culminating around 1300 CE and suggesting a rising lake water level contributing to tree mortality and increased preservation potential of trees in lakes. Again, this phase overlapped with a phase of strongly declined accumulation of peatland pine trees. These results were supported by taphonomic interpretation (Gastaldo, 1988) of the depositional histories, espe-

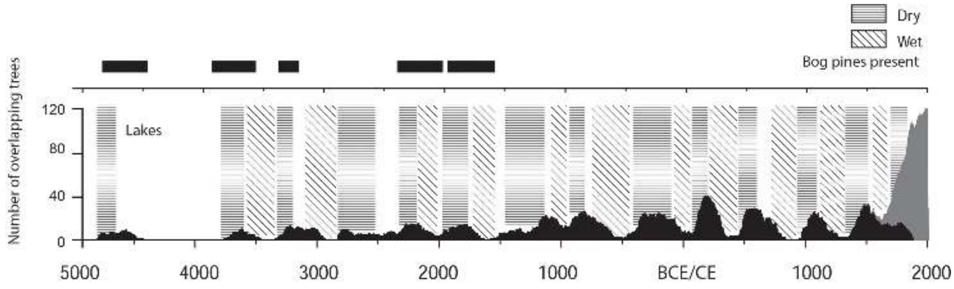


**Figure 4.** Changes in subfossil Scots pine (*Pinus sylvestris* L.) sample numbers over time (black) from lakes in the central Scandinavian Mountains (Gunnarson, 2008). The grey shaded area at the end represents living trees. Interpreted wet and dry periods shown in grey breaks and the presence of Scots pines growing on a nearby peat bog indicate drier conditions (figure adapted from Gunnarson, 2008). Data available at http://bolin.su.se/data/Gunnarson-2017 (last access: 22 June 2017).

cially their dissimilarities, and by comparisons with palaeolimnological reconstructions of water level fluctuations during the MCA and LIA (Luoto, 2009; Nevalainen et al., 2011; Nevalainen and Luoto, 2012). Similar to the study conducted in west-central Sweden (Gunnarson, 2008), the depositional histories in south-eastern Finland were found to reflect past hydroclimatic variations. Likely, the replication in pine chronologies from near the northern edge of the species range reflects summer temperature conditions, especially in subarctic sites (Helama et al., 2005, 2010). Further south, tree accumulation in different sediments seems to be more strongly influenced by recruitment and preservation potentials which, in turn are driven by local hydroclimatic conditions.

## 3.4 Glaciers

### 3.4.1 Glaciers as direct and indirect climate indicators

Glaciers respond to climate changes through variations in length, area, and volume (Oerlemans, 1994, 2001). In the Arctic and subarctic, observations and indirect evidence of glacier fluctuations have been widely used as sources of information about past climates (Solomina et al., 2016, and references therein). Changes in glacier length through advances or retreats are indirect, lagged responses to climate change, while glacier mass balance variations, as indicated by changes in ice thickness and volume, are direct responses to the annual weather conditions (Haeberli and Hoelzle, 1995). Direct measurements of glacier variability across the world, derived from annual mass balance measurements using glaciological or geodetic methods, are generally limited to the last half century (Zemp et al., 2009). In addition, annual mass balance records have been extended for several centuries using meteorological and proxy data such as historical records and tree-ring data (e.g. Lewis and Smith, 2004; Watson and Luckman, 2004; Nordli et al., 2005; Linderholm and Jansson, 2007). However, to yield information about glacier variability beyond direct observations, indirect indicators are mainly used.

There are two types of indirect glacier records: classical discontinuous series usually based on moraines delimiting the former glacier positions and continuous records from lakes (Solomina et al., 2016). Geomorphological evidence of glacier advances, such as terminal moraines or proglacial lacustrine sediments, give relative dates of glacier fluctuations, usually with some uncertainty. Lichenometry, a method through which lichen dimensions are used to infer the timing of colonisation, can provide rough estimates of moraine formation (Bickerton and Matthews, 1992; Armstrong, 2004). If the moraines contain organic material, they can be dated by $^{14}C$ (Karlén and Denton, 1976) or dendrochronological methods (Luckman, 1993; Carter et al., 1999). Cosmogenic isotopes (e.g. $^{10}Be$) can be used to directly identify the age of moraine deposition (Gosse and Phillips, 2001; Granger et al., 2013). Continuous records derived from lake sediment properties represent both the advance and retreat phases of glacier variations (Dahl and Nesje, 1994; Matthews et al., 2005; Bakke et al., 2008). As soon as the meltwater signal in proglacial lake sediments covaries with the distance between the glacier and the lake, it can serve as an indicator of glacier extent and the corresponding equilibrium line altitude (ELA), which is the altitude where accumulation equals ablation (Dahl and Nesje, 1994). Reconstructions of the ELA are based on multi-proxy sediment analysis (e.g. loss on ignition, bulk density, magnetic susceptibility grain-size distribution, and AMS dating control).

Glacier mass balance measurements demonstrate that for most regions summer temperature is the dominant control on annual mass balance (Koerner, 2005; Björnsson et al., 2013). Some exceptions have been noted; glacier advances in coastal areas of Scandinavia, SE Alaska, Kamchatka, and New Zealand in the late 20th century were forced primarily by high winter precipitation (e.g. Lemke et al., 2007). This means that in order to derive precipitation information from records of glacier variations, the data should be complemented by independent temperature reconstructions. Thus, if the advance of a glacier corresponds to inferred warm summers (which would lead to increased ablation), it is likely that the advance was due to increased precipitation during winter (and vice versa). Various summer temperature proxies have been used to interpret past glacier fluctuations: macrofossils at the upper treeline (Dahl and Nesje, 1996), pollen (Bakke et al., 2008), chironomids (Axford et al., 2009), tree-ring data (Anchukaitis et al., 2013), sedimentary chlorophyll content (Boldt et al., 2015), melt features (Henderson, 2002), borehole temperatures (Wagner and Melles, 2002), and oxygen isotopes from ice cores (Kirkbride and Dugmore, 2006). Several sources of uncertainty should be taken into account when this approach is applied, such as the lag between glacier advances and corresponding climatic forcing, which may last for decades even for moderate-sized glaciers (Oerlemans, 2001), and the dating uncertainties for both geomorphic and stratigraphic data (for details see Nesje, 2009; Solomina et al., 2015, 2016).

#### 3.4.2 Hydroclimate signals inferred from glacier fluctuations

Numerous detailed reconstructions of winter precipitation during the Holocene are available from Norway, where the mass balance of many maritime glaciers depends largely on accumulation rather than temperature changes (Nesje, 2009). Dahl and Nesje (1996) calculated winter precipitation at Hardangerjøkulen in south-central Norway using proglacial sediments and treeline altitude variations over the Holocene. They found that winter precipitation during the period from 1250 to 600 cal BP (ca. 750 to 1400 CE) were similar to today's values (reference period 1961–1990). It then increased to more than 120 % compared to the modern values until the LIA maximum (1750 CE) before being reduced again with up to 90 %. These results conflict with those obtained from Bjørnbreen in central Norway, where a comparison of the ELA with reconstructed July temperature showed that the highest values of winter precipitation during the past 2 millennia occurred in the MCA at around 1000 CE (Matthews et al., 2005). The explanation for this disagreement could to some extent be related to the high spatial variability of winter precipitation in Norway. To explore the spatial precipitation patterns in Norway during the Holocene, Bakke et al. (2008) used data from two proglacial sites at Folgefonna (southern Norway) and Lenangsbreen (northern Norway) together with a pollen-based July temperature reconstruction. They found that the differences in the distribution of precipitation were related to the changes in the position of the westerlies. The southernmost position of the westerlies, leading to a smaller S–N precipitation distribution gradient and large positive precipitation anomalies during the last 2 kyr in western Norway, occurred around 800 and 1600 CE. The suggested link between the atmospheric circulation (NAO) and precipitation–glacier fluctuations (Nesje, 2009) is supported by the advance of Nigardsbreen in southern Norway between 1710 and 1735 CE, which was attributed mainly to increased winter precipitation linked with a period of the positive mode of the NAO (Nesje and Dahl, 2003).

Early studies of glacier advances during the LIA on Svalbard interpreted them to be responses to low temperatures (Svendsen and Mangerud, 1997; Humlum et al., 2005). However, some recent studies attribute a number of advances, at least those that occurred in the 19th and early 20th centuries, to increased precipitation associated with a positive phase of the NAO (Reusche et al., 2014; D'Andrea et al., 2012). In western Svalbard, Røthe et al. (2015) suggested that open water associated with a loss of sea ice was the source of increased precipitation leading to the advance of the Karlbreen Glacier from ca. 1700 to 1500 cal BP (ca. 300 to 500 CE). The large LIA glacier advances in coastal areas on Iceland could also reflect increased precipitation (Kirkbride and Dugmore, 2006). Based on geomorphological evidence and $^{14}$C dating, Lubinsky et al. (1999) identified glacier advances during the ∼ 10th and 12th centuries, 1400 and 1600 CE, and in the early 20th century in Franz Josef Land. The last advance occurred despite warm summers, as recorded from melt features in the Windy Dome ice core, due to anomalously high snow accumulation (Henderson, 2002). A glacier advance at ca. 1400 CE was also noted in Novaya Zemlya (Polyak et al., 2004). This was a time of increased winter precipitation as interpreted from the GISP2 ice core record (Zeeberg and Forman, 2001). Wagner and Melles (2002) suggested that the Holocene fluctuations of the Ymer Ø ice cap in east Greenland depended mainly on precipitation since the inferred fluctuations disagreed with the Greenland borehole temperature. The advance of glaciers in the Miki and IC Jacobsen fjords in eastern Greenland, which have been dated with lichenometry to around 900–950 CE, corresponds to the MCA and could also reflect the response of glaciers to increased precipitation (Geirsdóttir et al., 2000).

In Alaska, most glacier advances have been related to cool summers (Anchukaitis et al., 2013; Wiles et al., 2014). However, the MCA advance of the Sheridan Glacier (Zander et al., 2013), also observed for glaciers in Alaska and western Canada (Menounos et al., 2009; Koch and Clague, 2011), can be attributed to increased precipitation due to extended La Niña-like conditions (Koch and Clague, 2011). The advance of the Sheridan Glacier in the 1600s CE coincides with warming summers as recorded by tree rings (Anchukaitis et al., 2013) and a peak in sedimentary chlorophyll (Boldt et al., 2015); it is thus probably also a sign of increased winter precipitation. Using lichenometry-dated moraines and the density of the sediment in Kurupa Lake (the Brooks Range in Alaska), Boldt (2013) produced a continuous reconstruction of ELA variations for several glaciers in the region. By regressing ΔELA against average Arctic-wide summer temperatures from Kaufman et al. (2009) and using the residuals as a proxy for winter accumulation, he identified periods of increased (150–550, 650–1000, and 1500–1650 CE) and reduced (600, 1050–1450, and 1750 CE) accumulation. In the Chugach Mountains in south-central Alaska, McKay and Kaufman (2009) used the differences between inferred summer temperature and evidence for glacier advances and retreats to suggest a period of increased winter precipitation from 1300 to 1500 CE and reduced winter precipitation from 1800 to 1900 CE, changes which were likely associated with variability in the strength of the Aleutian Low.

### 3.4.3 Hydroclimate from ice cores

Ice cores provide information of past climates through analysis of the annual layers deposited in glaciers. Several ice core parameters are used as palaeoclimate proxies, such as isotopic composition (mainly temperature), dust (e.g. storminess, aridity), air bubbles (atmospheric composition), and acidity (volcanic eruptions; Rozanski et al., 1997). Ice core data have been used to infer past hydroclimate variability, mainly at lower latitudes such as Tibet (e.g. Thompson et al., 2000; Yao et al., 2008) and the Andes in South America (e.g. Thompson et al., 1985). If annual layers can be identified and dated in ice cores, annual accumulation may be interpreted as records of past precipitation rates (Paterson and Waddington, 1984); however, the accuracy of the reconstruction is affected by processes such as redistribution by wind, melting, and dating–measuring errors (e.g. Mosley-Thompson et al., 2001). The cosmogenic isotope $^{10}$Be can provide estimates of palaeoaccumulation rates (Yiou et al., 1997). Paterson and Waddington (1984) analysed ice core accumulation rates from Camp Century (Greenland) and the Devon Island ice cap (Arctic Canada) and concluded that precipitation rates had shown only minor fluctuations during the last 2 kyr. Direct studies of ice core accumulation rates on Greenland have provided a better understanding of spatiotemporal mass balance variability (e.g. Mosley-Thompson et al., 2001; Box et al., 2013), estimated precipitation trends (Mernild et al., 2015), and assessed links between precipitation–accumulation and large-scale modes such as the NAO (e.g. Appenzeller et al., 1998). Such long-term accumulation records may be useful for hydroclimate estimations. Box et al. (2013) provided a reconstruction of annual Greenland ice sheet snow accumulation showing increasing accumulation over the last 410 years. Ljungqvist et al. (2016) used “lamina thickness” from five sites across the Greenland ice sheet as proxies for annual precipitation in their hemispheric hydroclimate reconstruction (Supplementary Table S1 in their paper).

## 4 Regional comparisons

Here we present comparisons of hydroclimate records from two of the most densely sampled regions in the Arctic.

### 4.1 Canadian Arctic

In the Canadian Arctic Archipelago, available quantitative palaeoclimate reconstructions fall into three classes of data: (a) relatively low temporal resolution (100-year) regional precipitation reconstructions from pollen assemblages for the boreal zone, (b) individual site-based reconstructions of lake level or precipitation sampled at variable but relatively low temporal resolution based on various proxies, and (c) annually resolved reconstructions typically based on varves or tree rings. The most extensively used palaeoclimate proxies in this region are pollen records from lake sediment cores. Typically annual precipitation and temperature have been the targets for reconstructions (Gajewski, 2015a). An extensive modern database of pollen data (Whitmore et al., 2005) enables quantitative reconstructions, and a number of reconstructions are available from the Canadian Arctic. A recent review of Holocene climate variations in the Canadian Arctic also indicated a number of other proxies in use (Briner et al., 2016) based on isotope or other physical or chemical measures. Presently, there are few published hydroclimate reconstructions using other proxies, although they have the potential to produce records with high temporal resolution. However, networks of these are not yet available. Most of the records are considered as temperature records, even if some have been related to moisture.

Viau et al. (2008) and Viau and Gajewski (2009) presented regional reconstructions of annual precipitation using all available pollen records from the boreal zone of Canada and Alaska (Figs. 5–6, Table 1). At the scale of this study, spatial patterns in the precipitation reconstructions are not clear. In western Canada and Alaska, there was an increase in precipitation during the past 2000 years, whereas a long-term decrease was seen towards the east. There was no clear difference between the MCA and LIA (Fig. 5). From the Canadian Arctic, only four low-resolution reconstructions of annual precipitation are available and these are based on pollen records (Peros and Gajewski, 2008, 2009; Peros et al., 2010). They all show a comparable signal, with lower precipitation during the MCA and slightly higher moisture during the LIA, and during the period from 400 BCE to 600 CE, which is slightly earlier at site KR02 from Victoria Island (Fig. 6).

### 4.2 Fennoscandia

In Fennoscandia, palaeolimnological studies have produced records indicative of past regional hydroclimatic variability. Such records are based on microfossil, macrofossil, and megafossil assemblages in addition to lithological data. Here we use 16 palaeolimnological records from the Arctic region (Table 2, Fig. 7) to illustrate hydroclimatic shifts and variations in Fennoscandia over the Common Era. The records are derived from depositional histories of subfossil trees (Gunnarson et al., 2003; Gunnarson, 2008; Helama et al., 2017a), estimates of peat humification (Gunnarson et al., 2003; Andersson and Schoning, 2010), sediment grain size (Si–Ti; Berntsson et al., 2015), varve thickness (Saarni et al., 2015), varve minerogenic lamina (“light sum”; Saarni et al., 2016), plant macrofossils (Väliranta et al., 2007), and chironomids

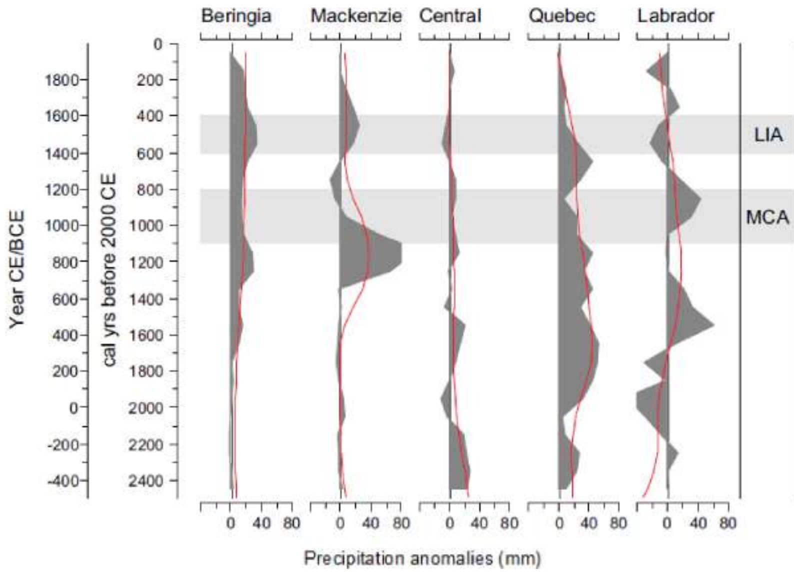


**Figure 5.** Regional reconstructions of annual precipitation from the boreal zone of North America. The average of all pollen records from the different regions are shown. Grey silhouettes are precipitation anomalies in millimetres; red lines are a loess fit (with a span of 0.2) to the data. The Beringia record is described in Viau et al. (2008) and the others in Viau and Gajewski (2009). Some large anomalies are truncated; see Table 1 for data availability.

**Table 1.** Regional (Fig. 5) and site-specific (Fig. 6) precipitation reconstructions over the Common Era from North America.

| Site | Proxy | Reference | Data availability (last access: 20 March 2018) |
|---|---|---|---|
| Beringia | Pollen | Viau et al. (2008) | https://www.ncdc.noaa.gov/paleo/study/8689 or http://www.lpc.uottawa.ca/data/reconstructions/index.html |
| Mackenzie, Central, Quebec, Labrador | Pollen | Viau and Gajewski (2009) | https://www.ncdc.noaa.gov/paleo/study/8690 or http://www.lpc.uottawa.ca/data/reconstructions/index.html |
| BC01 (Melville Island) | Pollen | Peros et al. (2010) | http://www.lpc.uottawa.ca/data/reconstructions/index.html |
| MB01 (western Victoria Island) | Pollen | Peros and Gajewski (2009) | https://www.ncdc.noaa.gov/paleo/study/6200 or http://www.lpc.uottawa.ca/data/reconstructions/index.html |
| KR02 (western Victoria Island) | Pollen | Peros and Gajewski (2008) | http://www.lpc.uottawa.ca/data/reconstructions/index.html |
| SL06 (Boothia Peninsula) | Pollen | Peros and Gajewski (2009) | https://www.ncdc.noaa.gov/paleo/study/6200 or http://www.lpc.uottawa.ca/data/reconstructions/index.html |

or cladoceran assemblages (Luoto, 2009; Luoto and Helama, 2010; Nevalainen et al., 2011, 2013; Nevalainen and Luoto, 2012; Luoto and Nevalainen, 2015; Berntsson et al., 2015). These records originate from Sweden and Finland and represent inland areas east of the Scandinavian Mountains.

Visual inspection of Fig. 7 does not indicate any strong agreement among the records, and correlations between the smoothed series (green lines in Fig. 7) are as low as 0.08. In fact, one might expect to find such disparity considering the peculiarities in local climate and range of proxy types, with an additional issue arising from dating uncertainties. However, dating issues may not constitute a critical factor for the observed low correlations among the records; comparing the depositional histories of subfossil trees from lake archives in Sweden (SWE01; Gunnarson et al., 2003; Gunnarson, 2008) and Finland (FIN12; Helama et al., 2017a), dated by means of dendrochronology and thus without dating uncertainties, results in a correlation of −0.20. The highest correlation between any pair of sites (0.83) is obtained between the two cladoceran-based lake water depth reconstructions from southern Finland (FIN08 and FIN14; Nevalainen et al., 2011, 2013). The highest inter-proxy correlation of 0.72 was found between the cladoceran-based lake water depth reconstruction (FIN07; Nevalainen and Luoto, 2012) and FIN12, with both multi-proxy records coming from southern Finland (Helama et al., 2017a). Among the Swedish data, the highest correlation of 0.42 was obtained between the peat humification index (SWE03; Gunnarson et al., 2003) and a chironomid-based record of catchment erosion (SWE05; Berntsson et al., 2015). The highest inter-country correla-

**Table 2.** Proxy records from Sweden (SWE) and Finland (FIN) indicative of hydroclimatic variations over the Common Era.

| Site | Code | Lat and long | Proxy and indication | Resolution (yr) | Reference | Data availability |
|---|---|---|---|---|---|---|
| Håckren | SWE 01 | 63.17 13.50 | Tree accumulation and lake level | 1 | Gunnarson et al. (2003); Gunnarson (2008) | http://bolin.su.se/data/ Gunnarson-2017[a] |
| Backsjömyren | SWE 02 | 62.68 14.53 | Peat humification and peatland water table | 40 | Andersson and Schoning (2010) | https://bolin.su.se/data/ Linderholm-2018[b] |
| Stömyren | SWE 03 | 60.38 15.27 | Peat humification and peatland water table | 40 | Gunnarson et al. (2003) | https://www1.ncdc.noaa.gov/pub/data/ paleo/reconstructions/hydroclimate/ ljungqvist2016/hydro_proxies/ Stomyren.txt[a] |
| Vuoksjávrátje | SWE 04 | 66.25 15.72 | Si–Ti (coarse grain size) and flooding | 1.5 | Berntsson et al. (2015) | https://www.ncdc.noaa.gov/paleo/ study/22253[a] |
| Vuoksjávrátje | SWE 05 | 66.25 15.72 | Chironomids and catchment erosion | 80 | Berntsson et al. (2015) | https://www.ncdc.noaa.gov/paleo/ study/22253[a] |
| Kontolanrahka | FIN 06 | 60.78 22.78 | Plant macrofossils and peatland water table | 10 | Väliranta et al. (2007) | https://bolin.su.se/data/ Linderholm-2018[b] |
| Iso Lehmälampi | FIN 07 | 60.33 24.60 | Cladocera and water depth (intra-lake) | 80 | Nevalainen and Luoto (2012) | https://www.ncdc.noaa.gov/paleo/ study/22212[a] |
| Iso Lehmälampi | FIN 08 | 60.33 24.60 | Cladocera and water depth (multi-lake) | 80 | Nevalainen et al. (2011) | https://www.ncdc.noaa.gov/paleo/ study/22212[a] |
| Iso Lehmälampi | FIN 09 | 60.33 24.60 | Chironomids and lake water depth | 80 | Luoto (2009) | https://www.ncdc.noaa.gov/paleo/ study/22212[a] |
| Kalliojärvi | FIN 10 | 63.22 25.37 | Varve light sum and spring floods | 1 | Saarni et al. (2016) | https://www.ncdc.noaa.gov/paleo/ study/22194[a] |
| Kallio-Kourujärvi | FIN 11 | 62.57 27.00 | Varve thickness and precipitation | 1 | Saarni et al. (2015) | https://www.ncdc.noaa.gov/paleo/ study/22193[a] |
| SE Finland | FIN 12 | 61.95 28.97 | Tree accumulation and lake water depth | 1 | Helama et al. (2017a) | https://bolin.su.se/data/ Linderholm-2018[b] |
| SE Finland | FIN 13 | 61.80 29.75 | Tree accumulation and peatland water table | 1 | Helama et al. (2017a) | https://bolin.su.se/data/ Linderholm-2018[b] |
| Pieni-Kauro | FIN 14 | 64.28 30.12 | Cladocera and lake water depth | 40 | Nevalainen et al. (2013) | https://www.ncdc.noaa.gov/paleo/ study/22213[a] |
| Pieni-Kauro | FIN 15 | 64.28 30.12 | Chironomids and streamflow | 40 | Luoto and Helama (2010) | https://www.ncdc.noaa.gov/paleo/ study/22213[a] |
| Kylmänlampi | FIN 16 | 64.30 30.25 | Chironomids and lake water depth | 100 | Luoto and Nevalainen (2015) | https://www.ncdc.noaa.gov/ paleo-search/study/22211[a] |

[a] Last access: 28 February 2018. [b] Last access: 4 April 2018.

tion of 0.54 was found between SWE05 and lake water depth (FIN14; Nevalainen et al., 2013).

The difference between the means of the proxy values between the two periods, the LIA and the MCA, was computed for the Fennoscandian records; 9 out of 16 records indicate wetter conditions towards the LIA, and 8 of these records are located in Finland. Four out of seven proxies that indicate drier LIA conditions originate from Sweden. Although these findings imply a more pronounced change towards wetter conditions in the eastern part of the region, it is also possible that part of these differences arises from the varying sensitivity of the proxies to different seasons. While most of the studied proxy records likely represent hydroclimatic variations during summer, at least four of the records indicate a relatively drier LIA (Luoto and Helama, 2010; Berntsson et al., 2015; Saarni et al., 2016) may actually reflect climatic and environmental factors attributable to boreal winter–spring phenomena, such as flooding, erosion, or streamflow. In boreal settings, a peak in run-off is generally attained during the spring season. The strength of this peak is strongly related to snowmelt and, in fact, the respective proxy data may be largely responding to antecedent snow conditions and thus winter precipitation. This has previously been described for eastern Finland, where a collection of proxy records reflecting either winter–spring or summer variability was found to exhibit contrasting hydroclimatic trends in respective variables through the MCA and LIA (Luoto and Helama, 2010). Therefore, the observed division of proxy records according to their indications of climate becoming either wetter or drier through the MCA–LIA transition may reflect, at least partly, their response to precipitation in either winter–spring or summer.

The issue of seasonal responses may be particularly interesting in the context of the long-term development of the

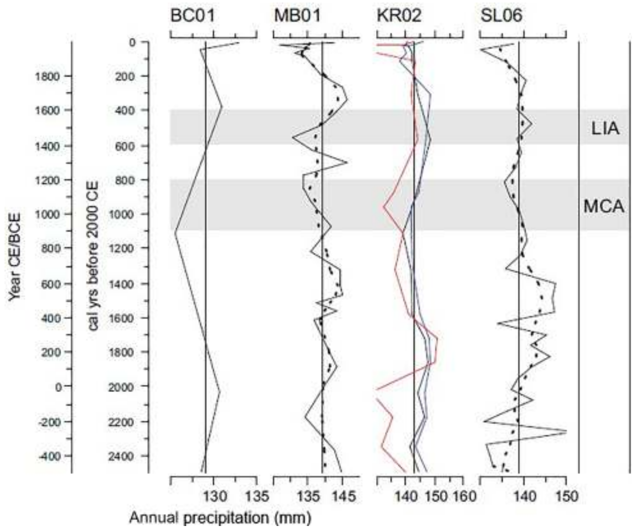


**Figure 6.** Annual precipitation reconstructions based on pollen assemblages from four lake sites in the Canadian Arctic: BC01, Melville Island (Peros et al., 2010); MB01, western Victoria Island (Peros and Gajewski, 2009); KR02, western Victoria Island (Peros and Gajewski, 2008); SL06, Boothia Peninsula (Peros and Gajewski, 2009). Dotted lines are loess lines fit to the data. For lake KR02: red represents the modern analogue technique, blue is WAPLS, and black is PLS. See Table 1 for information on data availability.

NAO. The Fennoscandian study sites are situated in a region where the positive NAO phase is attributable to increases in precipitation and thus enhanced snowfall during winter (Hurrell, 1995), but with decreased precipitation during much of the summer season (Folland et al., 2009). The different seasonal responses may at least partly explain the deviating patterns of hydroclimatic trends through the MCA and LIA among the proxies if the same climatic forcing (i.e. NAO) is anticipated to result in contrasting trends in respective records according to their target season sensitivity. These results are in line with a predominantly positive NAO phase during the MCA associated with generally wet winters but dry summers (Trouet et al., 2009), while a negative NAO phase during the LIA has been linked with dry winters and wet summers (Luoto and Helama, 2010; Luoto et al., 2013; Luoto and Nevalainen, 2017). While the view of a prolonged positive phase during the MCA has been challenged by recent proxy observations (Ortega et al., 2015), additional support for a generally positive NAO phase overlapping the MCA has also been presented (Wassenburg et al., 2013; Baker et al., 2015). Still, it is notable that not all of the analysed proxies indicated a distinct change from the MCA to the LIA. Moreover, the records are characterised by low resolution, and high autocorrelation makes it difficult to perform any statistical tests for this change, so the results should be regarded cautiously.

Compared to the hydroclimate fluctuations during the MCA and the LIA, a notable feature that characterises several of the Fennoscandian proxies (SWE03, SWE05, FIN07, FIN08, and FIN14) during the first millennium CE is a dry pre-MCA period of multi-centennial duration (Fig. 7). The timing of this phase appears to overlap with that of Dark Ages Cold Period (DACP, ca. 300–800 CE; Ljungqvist, 2010; Helama et al., 2017b, c). Apart from climatic changes related to temperature fluctuations, the DACP was likely a period of marked variable climate conditions. A review of the palaeoclimate during the DACP showed that both wet and dry conditions have been noted in north-west Europe (Helama et al., 2017b). Using peat humification records Blackford and Chambers (1991) showed indications of wet conditions for the British Isles around 550 CE. Likewise, wet rather than dry spring–summer conditions during the DACP have been noted around north-west Europe (Helama et al., 2017b). Thus, despite an indication of dry conditions during the DACP in some of the Fennoscandian records, the findings imply a general lack of agreement between the available proxy indicators.

Finally, there is no general tendency for any anomalous 20th century conditions among the records. While some of the series exhibit trends towards wetter conditions during the past century, other records indicate relatively drier conditions over the same period (Fig. 7). However, the value of this finding is limited by the fact that the post-1950s interval is not present in more than half of the records, but it is in agreement with the findings of Seftigen et al. (2015b). These results are contrasted, however, by the new precipitation tree-ring-based reconstruction just south of the region from Estonia, where an upward trend in the most recent summer precipitation was found since the 18th century CE (Helama et al., 2017d).

## 5 Arctic hydroclimate synthesis from proxies and PMIP3 simulations

### 5.1 A composite of Arctic hydroclimate variability during the last 1200 years

As noted in the Introduction, Ljungqvist et al. (2016) presented a reconstruction of Northern Hemisphere hydroclimate variability focusing on centennial variability in which the Arctic region was represented by 18 records. Here a new synthesis of Arctic hydroclimate variability extending back to 800 CE is presented, using both high- and low-resolution records. Note that this is not a quantitative reconstruction; it only provides a qualitative view of relative hydroclimate variability in the Arctic. The aim is to assess the potential to derive an Arctic hydroclimate record with more high-frequency information than that derived for the same region from the results of Ljungqvist et al. (2016).

The length of the analysis is restricted by the temporal coverage of the available series. In order to make a comparison with the PMIP3 simulations (see below), the analysis was focused on the last 1200 years. All records have been used in previous studies and are publicly available (see Table 3 and

**Table 3.** Hydroclimate proxy records from three previously published data compilations. Except for annually resolved series, the resolution listed is the mean. Letters in column 3 indicate (a) data used by Ljungqvist et al. (2016; data available at https://www.ncdc.noaa.gov/paleo/study/19725; last access: 27 January 2017); (b) data from Weißbach et al. (2016a; data available at https://doi.pangaea.de/10.1594/PANGAEA.849161) and (c) data from Sundqvist et al. (2014; data available at https://www.ncdc.noaa.gov/paleo/study/15444; last access: 7 February 2017). The series used in the qualitative hydroclimate reconstruction are shown in bold. All data analysed for the 800–2000 CE period can be accessed at http://dx.doi.org/10.6084/m9.figshare.5683666.v1.

| ID | Region | | Site | Lat (° N) | Long (° E) | Archive | Proxy | Oldest | Youngest | Resolution | References |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Hydro2k_01 | Greenland | c | N14 | 59.98 | −44.18 | Lake | BSi | 11 | 1480 | 33 | Anderson et al. (2004) |
| Hydro2k_02 | NW Norway | c | Fiskebølvatnet | 68.413 | 14.802 | Lake | Mass Acc. Rate | 786 | 1569 | 17 | Balascio and Bradley (2012) |
| Hydro2k_03 | SE Norway | c | Nattmålsvatn | 69.1793 | 17.3943 | Lake | MS (SI) | 6 | 849 | NA | Janbu et al. (2011) |
| **Hydro2k_04** | **N Alaska** | **a** | **Wolverine Lake** | **67.098** | **−158.914** | **Lake** | **Mass Acc. Rate** | **800** | **1926** | **31** | **Mann et al. (2002)** |
| Hydro2k_05 | N Norway | c | Rystad 1 | 68.2389 | 13.7839 | Peat | Humification | 776 | 1646 | 827.5 | Vorren et al. (2012) |
| Hydro2k_06 | W Greenland | c | SS16 | 66.91 | −50.46 | Lake | Diatom | 820.9 | 1999.6 | 26.8 | Perren et al. (2012) |
| **Hydro2k_07** | **C Sweden** | **a** | **Stömyren** | **60.2083** | **13.4667** | **Peat** | **Humification** | **794** | **1928** | **37** | **Borgmark and Wastegård (2008)** |
| Hydro2k_08 | W Greenland | c | SS1381 | 67.014 | −51.102 | Lake | Mineral flux | 795 | 1811 | 41 | Anderson et al. (2012) |
| Hydro2k_09 | W Hudson Bay | c | Unit Lake | 59.404 | −97.493 | Lake | AMR/IRM | 799 | 2010 | 71 | Camill et al. (2012) |
| Hydro2k_10 | E Finland | c | Saarikko | 62.25 | 27.67 | Lake | $\delta^{18}O$ | 788 | 1822 | 47 | Heikkilä et al. (2010) |
| Hydro2k_11 | N Norway | a | Over Gunnarsfjorden | 71.0383 | 28.1685 | Lake | Pollen | 814 | 1989 | 49 | Allen et al. (2007) |
| Hydro2k_12 | N Norway | c | Sellevollmyra | 69.1083 | 15.9417 | Peat | Humification | 798 | 1495 | 69 | Vorren et al. (2007) |
| **Hydro2k_14** | **Greenland** | **a** | **Crête** | **71.12** | **−37.32** | **Ice** | **Lamina** | **800** | **1973** | **1** | **Andersen et al. (2006)** |
| **Hydro2k_15** | **Grennland** | **a** | **Dye 3** | **65.11** | **−43.49** | **Ice** | **Lamina** | **800** | **1978** | **1** | **Andersen et al. (2006)** |
| **Hydro2k_16** | **Canada** | **a** | **East Lake** | **74.88** | **−109.53** | **Lake** | **Lamina** | **800** | **2005** | **1** | **Cuven et al. (2011)** |
| **Hydro2k_17** | **Greenland** | **a** | **GISP2** | **72.6** | **−38.5** | **Ice** | **Lamina** | **800** | **1987** | **1** | **Meese et al. (1994)** |
| **Hydro2k_18** | **Greenland** | **a** | **GRIP** | **72.35** | **−37.38** | **Ice** | **Lamina** | **800** | **1979** | **1** | **Andersen et al. (2006)** |
| **Hydro2k_19** | **Greenland** | **a** | **NGRIP** | **75.1** | **−42.32** | **Ice** | **Lamina** | **800** | **1995** | **1** | **Andersen et al. (2006)** |
| **Hydro2k_20** | **Alaska** | **a** | **Dune Lake** | **64.42** | **−149.9** | **Lake** | $\boldsymbol{\delta^{13}C}$ | **795** | **1992** | **16** | **Finney et al. (2012)** |
| **Hydro2k_21** | **Alaska** | **a** | **Ongoke Lake** | **59.25** | **−159.42** | **Lake** | **Diatom** | **498** | **2004** | **15** | **Chipman et al. (2009)** |
| **Hydro2k_22** | **Canada** | **a** | **Marcella Lake** | **60.07** | **−133.81** | **Lake** | $\boldsymbol{\delta^{18}O}$ | **798** | **2008** | **10** | **Andersen et al. (2007)** |
| **Hydro2k_23** | **N Norway** | **a** | **Nerfloen Lake** | **61.93** | **6.87** | **Lake** | **Particle size** | **786** | **1969** | **25** | **Vasskog et al. (2012)** |
| Hydro2k_24 | Greenland | c | Milcent | 70.3 | −44.55 | Ice | Acc. Rate | 1174 | 1966 | 1 | Andersen et al. (2006) |
| Hydro2k_27 | Alaska | c | Takahula Lake | 67.35 | −153.66 | Lake | $\delta^{18}O$ calcite | 753 | 2001 | 50 | Clegg and Hu (2010) |
| **Hydro2k_31** | **E Finland** | **a** | **Pieni-Kauro Lake** | **64.28** | **30.12** | **Lake** | **Chironomid** | **800** | **1990** | **46** | **Luoto et Helama (2010)** |
| **Hydro2k_32** | **Finland** | **a** | **Southern Finland** | **61.5** | **28.5** | **Trees** | **Ring width** | **800** | **1993** | **1** | **Helama et al. (2009)** |
| **Hydro2k_34** | **S Sweden** | **a** | **Fågelmossen 1** | **59.29** | **14.27** | **Peat** | **Humification** | **794** | **1914** | **15** | **Borgmark and Wastegård (2008)** |
| **Hydro2k_35** | **S Sweden** | **a** | **Fågelmossen 2** | **59.29** | **14.27** | **Peat** | **Humification** | **793** | **1967** | **12** | **Borgmark and Wastegård (2008)** |
| **Hydro2k_37** | **Finland** | **a** | **Kontolanrahka Lake** | **60.78** | **22.78** | **Peat** | **Humification** | **750** | **1913** | **30** | **Vairanta et al. (2007)** |
| Hydro2k_38 | Greenland | b | NGT B16 | 73.9 | −37.6 | Ice core | Acc. Rate | 1471 | 1992 | 1 | Weisbach et al. (2016a) |
| Hydro2k_39 | Greenland | b | NGT B17 | 75.25 | −37.62 | Ice core | Acc. Rate | 1363 | 1992 | 1 | Weisbach et al. (2016a) |
| Hydro2k_40 | Greenland | b | NGT B18 | 76.61 | −36.4 | Ice core | Acc. Rate | 874 | 1992 | 1 | Weisbach et al. (2016a) |
| Hydro2k_41 | Greenland | b | NGT B19 | 78 | −36.39 | Ice core | Acc. Rate | 753 | 1953 | 1 | Weisbach et al. (2016a) |
| Hydro2k_42 | Greenland | b | NGT B20 | 78.83 | −36.5 | Ice core | Acc. Rate | 775 | 1993 | 1 | Weisbach et al. (2016a) |
| Hydro2k_43 | Greenland | b | NGT B21 | 80 | −41.1 | Ice core | Acc. Rate | 1372 | 1993 | 1 | Weisbach et al. (2016a) |
| Hydro2k_44 | Greenland | b | NGT B22 | 79.34 | −45.91 | Ice core | Acc. Rate | 1372 | 1993 | 1 | Weisbach et al. (2016a) |
| Hydro2k_45 | Greenland | b | NGT B23 | 78 | −44 | Ice core | Acc. Rate | 1023 | 1993 | 1 | Weisbach et al. (2016a) |
| Hydro2k_46 | Greenland | b | NGT B26 | 77,25 | −49.21 | Ice core | Acc. Rate | 1505 | 1994 | 1 | Weisbach et al. (2016a) |
| Hydro2k_48 | Greenland | b | NGT B29 | 76 | −43.49 | Ice core | Acc. Rate | 1471 | 1994 | 1 | Weisbach et al. (2016a) |
| Hydro2k_49 | Greenland | b | NGT B30 | 75.01 | −42 | Ice core | Acc. Rate | 1242 | 1988 | 1 | Weisbach et al. (2016a) |

NA – not available

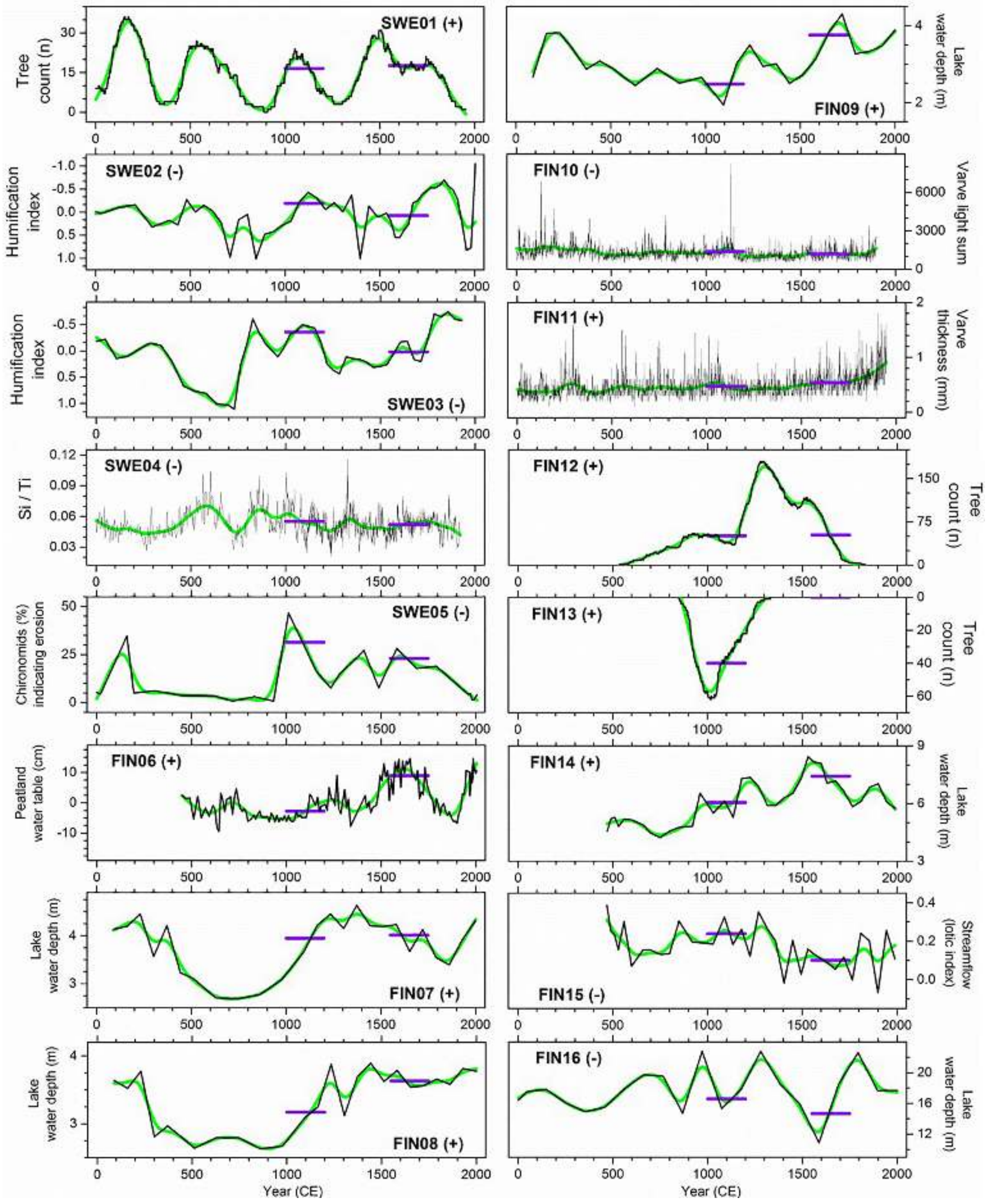


**Figure 7.** Hydroclimatic variations in Sweden (SWE) and Finland (FIN) over the Common Era (see Table 2 for details including data availability). The mean levels (violet line) during the Medieval Climate Anomaly (MCA) and Little Ice Age (LIA) were calculated from the published records (black line), those being additionally smoothed using a 200-year spline function (green line). Proxy data indicating change from the MCA towards wetter (drier) LIA conditions are noted by a plus (minus) sign. The graphs have been arranged so that wet conditions are indicated upward and dry conditions downward.

the "Data availability" section). The dataset is composed of 40 series and is based on a heterogeneous group of proxy sources: 17 records are from ice cores, 16 from lake sediments, 6 from peat, and 1 series is from tree rings (Fig. 8, Table 3). The majority of the records are located in the North Atlantic area (Fennoscandia, Greenland, and the Canadian Arctic) and Alaska.

The selection of the proxy records was based on several quality criteria (McKay and Kaufman, 2014). Specifically, all records (i) are from north of 60° N, (ii) extend back to at least 800 CE, (iii) extend into the 1900s CE in order to include the warming period of the 20th century (PAGES 2k Consortium, 2013), (iv) have an average sample resolution of less than 50 years, and (v) have at least two age control points during the defined study period. Following these criteria, 17 records were selected (see Table S2 for details). These strict selection criteria are necessary to allow for the comparison of data at centennial scales and facilitate time series analysis. The spa-

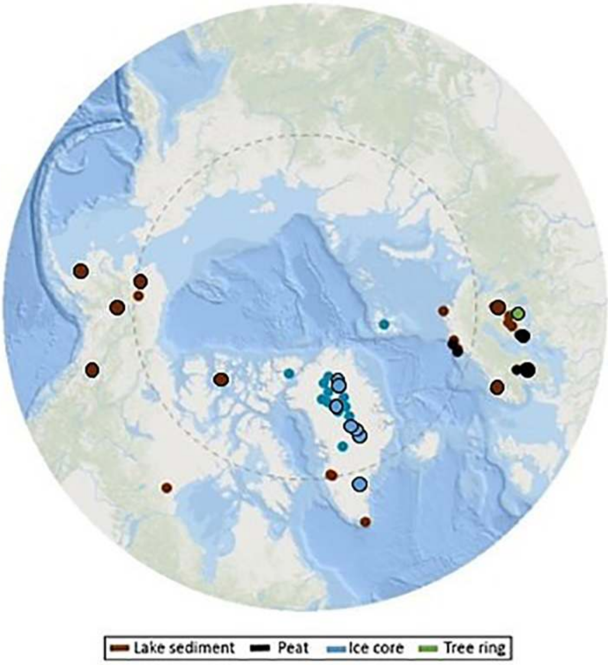


**Figure 8.** Spatial distribution of the hydroclimate proxy records available in the Arctic region. Records used for the new synthesis are highlighted by larger symbols and black borders. See Table 3 for information on the records.

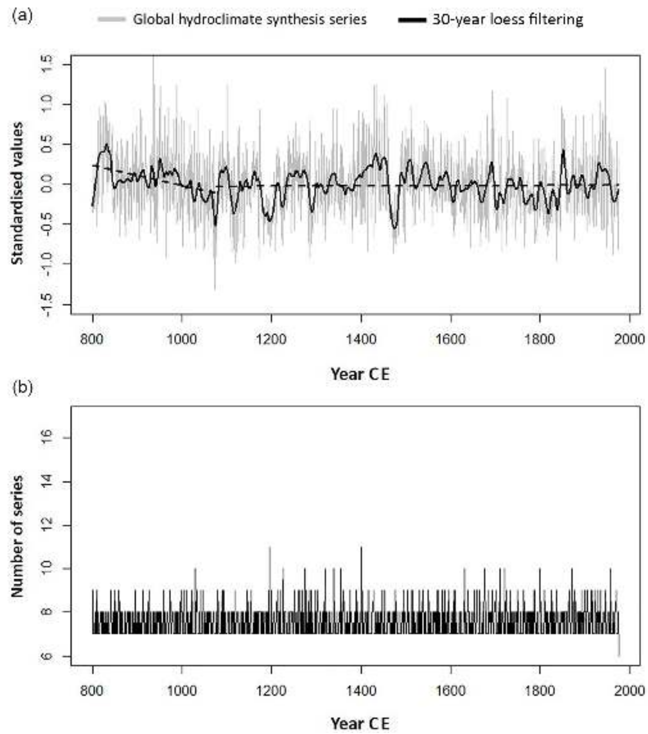


**Figure 9.** **(a)** Mean pan-Arctic hydroclimate index (grey line) based on 17 selected series (Table 3). The thick black line is a 30-year loess filter, and the dashed lines are linear trends determined by a Mann–Kendall test. Data are presented as $z$ scores. **(b)** Number of time series over time included in the synthesis.

tial coverage is mainly confined to Alaska, Arctic Canada, Greenland, and Fennoscandia, but these well-dated records, including many annually resolved records such as ice cores and varved sediments, offer the possibility to interpret hydroclimate variability in the Arctic from low to high frequencies.

To extract a common pattern from the records, we created an average signal in order to reduce the impact of random variability and enhance a possible signal (Moron et al., 2006; Hassan and Anwar, 2010). Although such a common signal obtained from several climatic proxies cannot be considered a reconstruction, it is suitable for investigating the different modes of variability present in the various records. The resulting composite reflects not only precipitation, but also a combination of processes related to the hydrological cycle (e.g. precipitation, evaporation). By calculating a standardised index of the palaeoclimatic series, we reduce the "external" variance (Zwiers, 1996; Rowell, 1998), i.e. the part of variance that is not spatially coherent. This external part of the signal can be considered as the part of the spatially independent stochastic (red or white) noise of a broad-scale climate signal.

A trend analysis was performed using the non-parametric Mann–Kendall test (Mann, 1945; Kendall, 1975), which has low sensitivity to abrupt breaks in an inhomogeneous time series. Positive values indicate that the ranks of both variables increase together, while a negative value indicates a decreasing trend. For this study, we choose the 95 % confidence level. All records were standardised to be comparable with each other.

A continuous wavelet transform (CWT) allows for the decomposition of a non-stationary time series that contains periodic or aperiodic components, noise, and progressive or abrupt changes (progressive transitions, singularities, and breaks; Debret et al., 2007; Steinhilber et al., 2012; Lapointe et al., 2017). The resulting plot of the wavelet transform, the scalogram, is a frequency contour diagram with time on the $x$ axis, frequency, wavelet scale, or equivalent Fourier period on the $y$ axis, and power on the $z$ axis. In the region of the spectrum where the zero padding decreases the power of the wavelet transform, the cone of influence, energy bands are likely to be less powerful. To determine the significance of the observed signal fluctuations, local wavelet spectra were compared to the spectra of random signals that would theoretically correspond to other realisations of the same random process. We again choose the 95 % confidence level (Torrence and Compo, 1998).

There was a significant negative trend between 800 and 1075 CE ($\tau = -0.404$, $p$ value $< 0.01$), whereas during the last 900 years, no clear trend is evident (tau $= 0.013$, $p$ value $= 0.57$; Fig. 9). A distinct decrease in the $z$ scores between 1456 and 1485 CE is also noticeable. A wavelet anal-

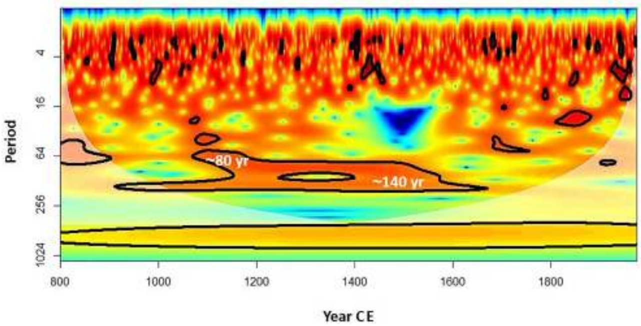


**Figure 10.** Wavelet analysis of the pan-Arctic hydroclimate record as shown in Fig. 9. Colours represent the amplitude of the signal at given time and spectral period; red equals the highest power, blue the lowest. White line corresponds to cone of influence on the wavelet coherence spectrum and global wavelet spectrum. Confidence levels of 95 % ($\alpha = 0.05$) are indicated on the wavelet spectrum with the black lines.

ysis reveals variability on multi-decadal to multi-centennial scales (Fig. 10). Because wavelet analysis is sensitive to large events that may hide the lowest frequencies recorded, the 1456–1485 CE event was extracted by wavelet filtering and the signal reconstructed by inverse Fourier transform before using CWT. An $\sim$ 80-year oscillation is present from 1050 to 1500 CE, while a $\sim$ 140 year oscillation is present from ca. 900 to about 1650 CE.

To determine the influence of the various regions on the variability recorded in our Arctic mean record, the pan-Arctic record was compared with data from the North Atlantic region (12 series) and Alaska (five series; Fig. 11). Visual comparison and correlation analysis between the Arctic mean record and each regional mean record indicate a stronger influence of the North Atlantic ($r^2 = 0.93$, $p$ value $< 0.01$) compared to Alaska ($r^2 = 0.35$, $p$ value $< 0.01$). This, however, should not be over-interpreted as 12 of the 17 records included in the pan-Arctic record are from the North Atlantic. Increasing the spatial coverage of hydroclimate proxies in Eurasia and North America will allow for a better understanding of overall hydroclimate variability in the Arctic.

## 5.2 Comparing pan-Arctic hydroclimate from proxies with PMIP3 simulations

Palaeoclimate models provide another means to investigate temporal and spatial hydroclimate variability in the Arctic during the last millennium. As a part of the third phase of the Palaeoclimate Modeling Intercomparison Project (PMIP3: Braconnot et al., 2012), last-millennium climate simulations were performed using a set of atmosphere–ocean general circulation models (AOGCMs) with the same experimental protocol (Schmidt et al., 2011). These simulations cover the period of 850–1850 CE and can be used to investigate climate responses to changes in external forcings such as solar irradiance and volcanic eruptions. Some of the models were also used to simulate climate variability for the period 1850–2005 and these are referred to as "historical simulations" (Taylor et al., 2012). In this section, 12 simulations (including six last-millennium simulations and six associated historical simulations) performed using six atmosphere–ocean general circulation models (Table S3) were used. The models used were HadCM3 (Schurer et al., 2013), IPSL-CM5A-LR (Dufresne et al., 2013), MPI-ESM-P (Jungclaus et al., 2014), CCSM4 (Landrum et al., 2013), CSIRO-Mk3L-1-2 (Phipps et al., 2011), and MRI-CGCM3 (Yukimoto et al., 2012). Simulated Arctic precipitation was then compared to the reconstructed Arctic hydroclimate by Ljungqvist et al. (2016, henceforth referred to as L16) and the new synthesis presented above. Both hydroclimate reconstructions and simulated total annual precipitation were transformed into $z$-score series because the reconstructions represent hydroclimate indices that are not necessarily comparable with annual total precipitation. Because L16 has centennial resolution, data from the simulations and the new synthesis were filtered using a Gaussian filter to preserve centennial-scale variability.

In the L16 reconstruction, it was wetter in northern than in southern Fennoscandia during the MCA compared to the LIA (Fig. 12a). Greenland shows an opposite pattern, indicating an increase in total annual precipitation during the MCA. This multiple proxy reconstruction has a limited spatial coverage in the Arctic, so hydroclimate variability can only be shown for Fennoscandia and part of Greenland. The six-model ensemble mean shows a different spatial pattern from that of the reconstruction (Fig. 12b), with increasing precipitation over most of Fennoscandia and Greenland. This discrepancy between the reconstruction and the model ensemble mean is not caused by anomalous outputs of any single model but a combination of all the models (Fig. S1 in the Supplement); the individual models show differences in spatial patterns compared to the reconstruction. Caution needs to be advised, as the magnitudes of the differences in proxy-derived hydroclimate between the MCA and LIA is consistently larger than in the model ensemble mean or in any individual model simulation. The discrepancy between model simulations and proxy reconstructions may imply that the changes in spatial hydroclimate patterns from the MCA to the LIA over Fennoscandia and Greenland are not related to changes in external forcings, but possibly internal variability. Another reason for the discrepancy between the reconstruction and the model simulations could be inadequate spatiotemporal availability of proxies across the Arctic, making it difficult to investigate changes in the spatial precipitation patterns between the MCA and the LIA. Therefore, proxy-based hydroclimate reconstructions covering a wider area of the Arctic are needed in order to make a comprehensive model–data comparison and to further investigate changes in spatial patterns of Arctic hydroclimate variability and their causes.

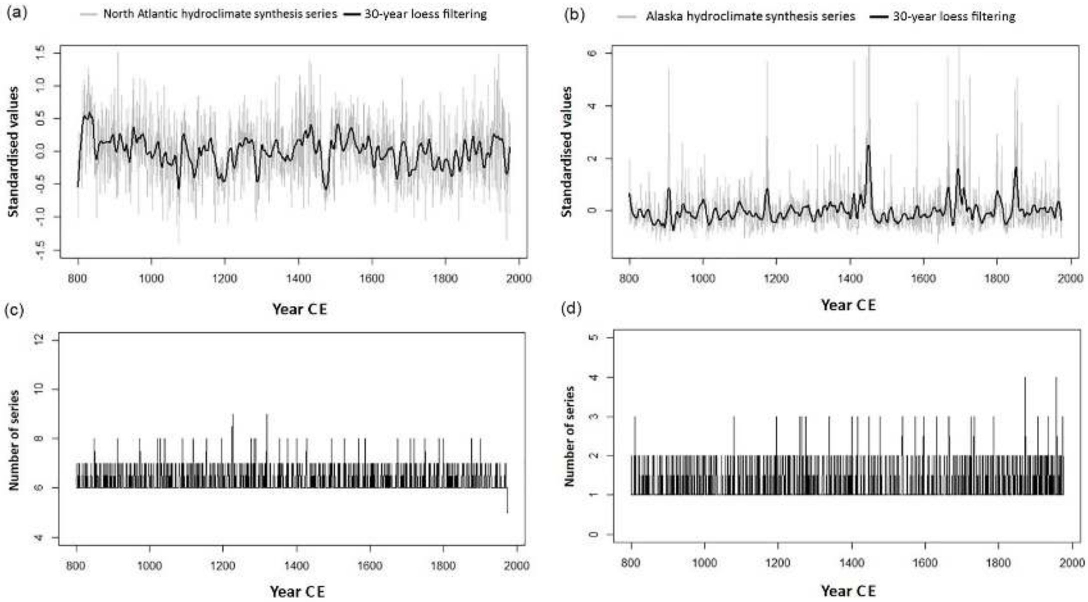


**Figure 11.** Regional hydroclimate mean series (grey) with 30-year loess filters (black) for the North Atlantic region **(a)** and Alaska **(b)**. Data are presented as $z$ scores. Panels **(c, d)** show the corresponding numbers of records through time included in the synthesis (see Table 3 for information on the records).

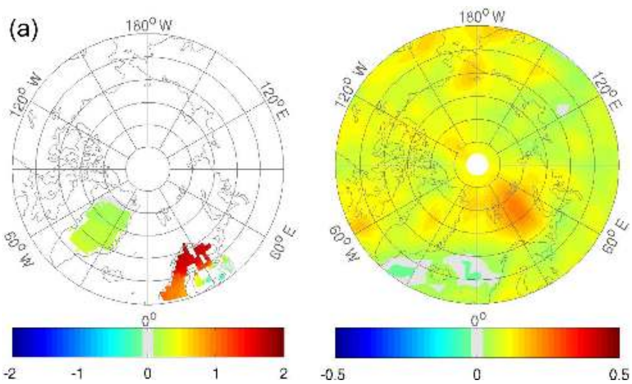


**Figure 12.** Spatial pattern of differences in annual hydroclimate between the MCA (950–1250 CE) and the LIA (1450–1850 CE) based on **(a)** hydroclimate reconstruction (Ljungqvist et al., 2016) and **(b)** ensemble mean of six last-millennium simulations. The values in **(a)** are $z$ scores of the hydroclimate index, while those in **(b)** are $z$ scores of the annual total precipitation. The $z$ scores are based on the period 850–1850 CE.

The new hydroclimate mean record shows quite coherent variability with L16 on centennial scales (Fig. 13), especially during the early MCA (ca. 900–1200) and early LIA (ca. 1400–1600). This is not surprising since they are based on many of the same proxy data. However, the new record suggests a shorter period of wet anomalies during the MCA compared to L16, and the variance of the new hydroclimate mean record is much larger after ca. 1200 CE. At a multi-centennial scale, the proxy-based reconstructions and model simulations all show drying from 800–1250 CE, increasing moisture until ∼ 1500–1600 CE, and low values from 1600–1850 CE. Compared to the model simulations, there is a discrepancy with the multi-proxy records during the later part of the MCA, where the model ensemble mean suggests a prolonged wet period lasting until 1200 CE, compared to the proxy-based records. One of the distinct features of the new hydroclimate mean record is the two distinct wet anomalies between 1400 and 1600 CE, which are more prominent than in the model simulation and where the latter is not present in L16. Overall, there is a better agreement between the model simulations and the new hydroclimate mean from the 14th century and onwards compared to L16.

## 6 Arctic hydroclimate variability in the past 2000 years

### 6.1 Current understanding

As has been shown in this review, significant efforts have been made over the last several decades to increase our understanding of hydroclimate variability in the Arctic region. However, it is also evident that the available records are insufficient to fully represent such a hydroclimatically inhomogeneous region. Moreover, there are still uncertainties regarding the temporal representation of some proxies and the interpretation of the hydroclimate information, as well as the season that is recorded by the proxy records.

Over the last 1200 years, a commonly studied period as it includes the MCA and the LIA, the proxy reconstructions do

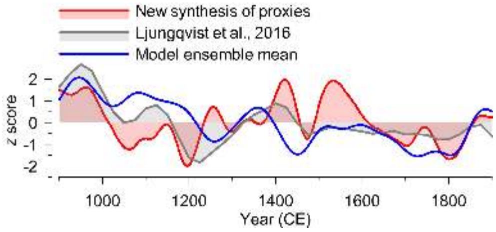


**Figure 13.** Comparison of centennial-scale annual hydroclimate variability (after application of a Gaussian filter) in the Arctic ($\geq 60°$ N) North Atlantic region from a reconstruction by Ljungqvist et al. (2016, grey), the new pan-Arctic hydroclimate proxy synthesis (red), and an ensemble mean of six last-millennium precipitation simulations (blue). See text for more information.

not provide clear evidence of systematic hydroclimate patterns across the Arctic or even regionally. In general, drier conditions during the MCA are indicated in several records in Fennoscandia (Fig. 7) and Arctic Canada (Fig. 6), but not across the North American boreal zone (Fig. 5). Similarly, the LIA seems to have been a generally wet period, as indicated by the regional comparisons and evidence of glacier advances (see Sect. 3.4), but again the picture is far from clear. The new Arctic hydroclimate mean synthesis (Fig. 13) suggests drying during the MCA, but wet conditions in the early part of the LIA and drier conditions in the latter part. This is largely in agreement with L16, although the latter shows less variability during the mainly dry LIA. At least from the LIA onwards, there is a better agreement between the model ensemble mean and the new synthesis than with L16. Both Arctic hydroclimate records derived from L16 and the composite presented here are, however, insufficient for drawing any firm conclusions for the whole region.

Hydroclimatic variations during the first millennium CE have received relatively less attention than during the MCA and LIA. Detailing the hydroclimate variability of the entire Common Era would allow for the placement of the 20th and 21st century changes in a long-term perspective. The Fennoscandian proxy series highlighted a phase of anomalous pre-MCA hydroclimate conditions during the Dark Ages Cold Period. As recently discussed (Helama et al., 2017b), this was possibly a period of noticeable climatic fluctuations, not only in temperature but also in hydroclimate. Our results highlight the need for extending proxy records to cover this climatic period.

Arctic hydroclimate proxies provide information for different target seasons, and this is likely to have an impact on any synthesis. Figure 14a shows the 20th century trends in seasonal Arctic precipitation from the ERA-20C reanalysis data (Poli et al., 2013). The trends are positive in all seasons, but most pronounced in autumn. The greatest precipitation increase occurred over the North Atlantic and Pacific oceans in all seasons (Fig. 14b) and also over the Arctic Ocean. The changes over land are less coherent in both North America

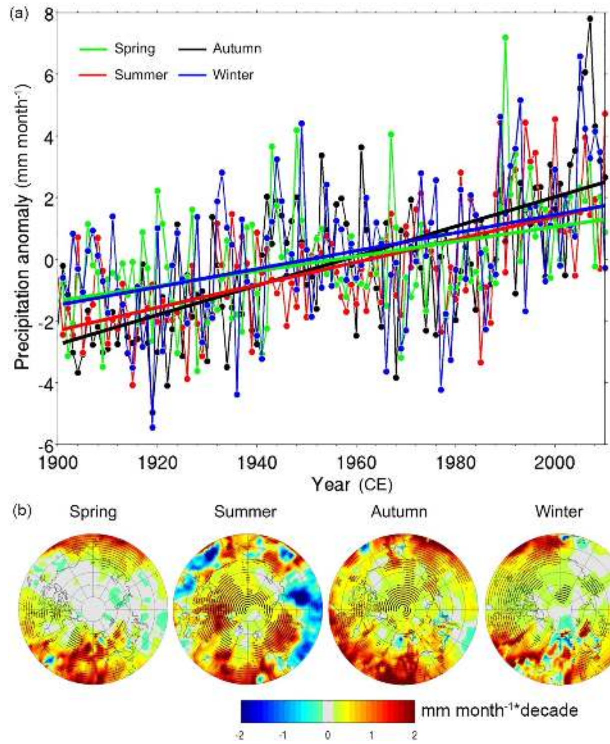


**Figure 14. (a)** Variability and linear trends of Arctic spring, summer, autumn and winter total precipitation anomalies over the period 1900–2010 CE from the ERA-20C reanalysis dataset (Poli et al., 2013). **(b)** Spatial patterns of linear trends of the Arctic spring, summer, autumn and winter total precipitation anomalies over the period 1900–2010 CE. Shading marks those grid cells where the trend is significant ($p < 0.01$).

and Eurasia, especially in summer, the target season for many proxies. Regional differences are also evident in a millennium model perspective (Fig. S2). From 900 to 1900 CE, the model ensemble mean shows slight negative trends in precipitation during spring, summer, and winter, but an obvious negative trend in (Fig. S2a). Moreover, regional differences in long-term trends are indicated both within regions and between seasons in the six studied models (Fig. S2b). The implication of this is that in order to provide an average view of hydroclimate variability for the Arctic, there must be an even distribution of high-quality, numerically calibrated, verified, and replicated climate-sensitive records. More attention should also be paid to the target season of the climate signal when developing large-scale composites to avoid the mixing of hydroclimate information across the seasons.

### 6.2 Towards a better understanding of spatiotemporal hydroclimate variability in the Arctic

Spatially explicit hydroclimate reconstructions provide excellent opportunities to study spatiotemporal variations, influences of forcings (e.g. Seager et al., 2007), and for proxy–

model comparisons. However, due to the low number of available hydroclimate proxy records from the Arctic and the imbalance in spatial coverage (Table 3, Fig. 8), it is currently impossible to prepare a field reconstruction for the whole region. As noted in Sect. 3.3, there are two tree-ring drought atlases covering parts of the Arctic (Fig. 3); however, the data representation is limited and the usage of temperature-sensitive tree-ring proxies as hydroclimate indicators needs to be properly addressed. Given the precipitation sensitivity of some high-latitude trees (St George and Ault, 2014) and more efforts in utilising isotope records from trees, it may be possible to extend any analyses of hydroclimate variability into Eurasia. Targeted regional and spatial reconstructions could be achieved for well-replicated regions, such as Fennoscandia, the Nordic Sea region, or western North America. To facilitate a compilation of Arctic hydroclimate variability, a dedicated hydroclimate proxy database needs to be developed with firm criteria for which records to include. It is encouraging that several new hydroclimate records have been made available during the process of preparing this review (see Table 2). Moreover, the new synthesis presented in Sect. 5.1 shows the potential to provide regional hydroclimate records with high temporal resolution, providing useful information on multi-decadal timescales (e.g. Fig. 13).

## 7 Recommendations for future work

Expanding the spatial coverage of hydroclimate proxy records is important, particularly for Eurasia, outside Fennoscandia, and North America. Several hydroclimate records that would add valuable information are not publicly available, so it is important to encourage palaeoclimate researchers to share and publicly archive their data.

A consistent and coherent Arctic 2k hydroclimate database should be assembled by including all necessary metadata and information on the seasonality of proxies included, following the PAGES 2k data standard, to facilitate the development of site selection criteria for a more robust and defendable synthesis of Arctic hydroclimate history.

A field reconstruction for regions with a sufficient number of hydroclimate proxy records in time and space is critical to advance our understanding of dynamical controls of Arctic hydroclimate. Presently there seem to be opportunities for a trans-Atlantic comparison, which may shed light onto observed regional hydroclimate patterns and the forcing mechanisms.

Closer collaborations between the palaeoclimate data and modelling communities are needed to address and resolve the discrepancies evident in comparisons between the existing "observational" data (reanalysis and proxies) and climate model simulations (PAGES Hydro2k Consortium, 2017).

**Data availability.** The CMIP5/PMIP3 climate data used in this paper can be obtained from https://cmip.llnl.gov/cmip5/data_getting_started.html, and the specific analyses presented here (Figs. 1–2, 12–14 and S1–S2 in the Supplement) are available at https://figshare.com/articles/Data_Linderholm_et_al_2017/5729214 (last access: 22 December 2017). The drought atlases (NADA and OWDA; Fig. 3) are accessible through https://www.ncdc.noaa.gov/paleo/study/6319 (NADA, last access: 16 January 2017) and https://www.ncdc.noaa.gov/paleo/study/19419 (OWDA, last access: 4 November 2015). For the accessability of the data presented in Figs. 5 and 6, see Table 1. The data described and partly used for the compilation of the pan-Arctic hydroclimate mean in Sect. 5 are available from the following sources (see text for references): https://www.ncdc.noaa.gov/paleo/study/15444 (last access: 7 February 2017) and https://www.ncdc.noaa.gov/paleo-search/study/19725 (last access: 27 January 2017), https://doi.org/10.1594/PANGAEA.849161 (Weißbach et al., 2016b). The syntheses presented in Figs. 9 (pan-Arctic) and 11 (North Atlantic and Alaska) are archived on Figureshare and available at https://figshare.com/articles/Global_North_Atlantic_Alaska_synthesis_record_txt/5502199 (https://doi.org/10.6084/m9.figshare.5502199).

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**Competing interests.** The authors declare that they have no conflict of interest.

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