DECIPHERING ARCTIC CLIMATE IN A PAST GREENHOUSE WORLD: MULTI-PROXY RECONSTRUCTIONS OF PLIOcene CLIMATE

by

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ABSTRACT

The high sensitivity of high latitudes to global climate changes is the stimulus for the study of ancient Arctic ecosystems under greenhouse conditions. With an increasing number of studies, including the most recent Intergovernmental Panel on Climate Change (IPCC) report highlighting Pliocene climate as key example for the study of Earth system sensitivity to higher levels of atmospheric CO₂, the need for accurate proxy records for this period is crucial.

In order to investigate Pliocene climate, I used stable isotopic studies of fossil molluscs, moss and wood from two fossil forest deposits in the Canadian High Arctic. Temperatures were determined for an Early Pliocene (4-5 Ma) fossil forest site located on Ellesmere Island using ‘clumped’ and stable isotopic analysis of mollusc shells and stable isotope values of fossil wood. Mollusc inferred growing season (May-Sept) temperatures derived using two independent techniques were estimated to be 11-16° C warmer than present (1950-1990) Ellesmere Island temperatures. Tree ring inferred growing season (June-July) temperatures (JJ) were 10-16° C and mean annual temperatures (MAT) were 18-20° C warmer than present (1950-1990). Mean annual and growing season (JJ) temperatures were also determined using fossil wood from a younger (2.4-2.8 Ma) late Pliocene-early Pleistocene site on Bylot Island. This deposit represents the remains of a flora that grew during an interglacial warm period during the transition to large-scale Northern Hemisphere glaciation that occurred between 2.5 to 3 million years ago. Mean annual temperatures were ~12° C and growing season temperatures were ~13° C warmer than present (1923-2010). The interglacial setting of the Bylot Island site and the warm
temperatures suggests that prior to using such sites as true analogues of future conditions we may need to consider how close the feedbacks operating then were to the feedbacks we might expect in the future. However, that temperatures so much warmer than present existed in the high Arctic during a period when levels of atmospheric CO$_2$ were at near-present levels indicates that we may be moving beyond our ability to use the Pliocene as an example of the future.
CHAPTER 1 – INTRODUCTION

1.1 Statement of the problem

The Pliocene (2.6 to 5 million years ago) represents the last period in Earth’s history when temperatures significantly warmer than the modern instrumental record persisted for an extended length of time. Modeled annual global air temperatures during the Pliocene are ~3-5°C and Arctic temperatures are ~7-10°C warmer than pre-industrial temperatures (Sloan et al., 1996; Haywood et al., 2009). Modeled mean annual Pliocene temperatures are in line with projections of future warming for the end of this century, so an improved understanding of climate during this period can provide an important perspective on a possible future world.

Abundant well-preserved Pliocene fossil forest sites exist throughout the Arctic (Matthews and Ovenden, 1990; Elias and Matthews, 2002) providing an important set of potential paleoclimatic proxies with which to investigate Pliocene Arctic climate (Fig 1).
Figure 1: Map showing the location of all Plio-Pleistocene fossil forest deposits in the North American Arctic (Fyles, 1989; Matthews and Ovenden, 1990; Funder et al., 2001; Elias and Matthews, 2002; Bennike et al., 2002; Mathews et al., 2003; this dissertation). Red dots indicate sites studied in this dissertation. Dashed lines indicate provincial and international boarders. Green line indicates the modern northern tree-line.

The need to investigate the Pliocene was highlighted in the most recent report of the Intergovernmental Panel on Climate Change (AR4), which stated that the Pliocene is an accessible example of a world that is similar in many ways to what models estimate the Earth of the late 21st century could be (Jansen et al., 2007). Investigations of Pliocene climate have also become increasingly important to test the Earth system sensitivity to changes in levels of atmospheric CO$_2$ (Lunt et al., 2010; Ballantyne et al., 2010; Pagani et al., 2010).Earth system sensitivity differs from climate sensitivity—the mean global temperature response to a doubling of atmospheric CO$_2$—by including the effects of long-
term feedbacks in addition to the response solely to atmospheric CO$_2$ (Pagani et al., 2010).

At present few records exist with which to determine the range of inter-annual variability under overall warmer conditions, such as existed during the Pliocene. The annually resolved nature of the tree-ring records presented here can also provide information on climate variability. For example in the modern climate system there are several well known oceanic and atmospheric modes of climate variability, such as the El Niño/Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO) and the Northern Annular Mode/Arctic Oscillation (NAM/AO) and the North Atlantic Oscillation that exert a strong influence on regional and global climate. Several studies have begun to investigate whether these familiar modes of climate variability were extant during the Pliocene (Wara et al., 2005; Fedorov et al., 2006; Bonham et al., 2009; Hill et al., in review). These modeling studies suggest that the mechanisms do exist to support these modes of climate variability during the Pliocene. Empirical evidence, however, for climatic oscillations operating during this time remains sparse (Bonham et al., 2009; Hill et al., in review).

The primary research objectives of this dissertation were (1) to obtain estimates of Pliocene Arctic temperatures for a well studied early Pliocene (4-5 million year old) site on Ellesmere Island using both stable and ‘clumped’ isotopes of mollusc carbonate and stable isotopes of fossil wood cellulose and to quantify temperatures in the Arctic during the Pliocene; (2) to obtain estimates of meteoric water $\delta^{18}O$ values and humidity for an early Pliocene site on Ellesmere Island; (3) to apply the methods established for
estimating temperatures from fossil wood to a Plio-Pleistocene (2 to 3 million year old) fossil forest site on Bylot Island; (4) to determine the range of inter-annual variability in Pliocene temperature estimates using tree-ring derived reconstructions of Pliocene temperatures; and finally (5) to use the annually resolved tree-ring records to attempt to determine if large-scale modes of climate variability were in operation during the Pliocene.

1.2 Pliocene Climate

Extreme seasonality and sub-zero temperatures for much of the year characterize modern Arctic climate, resulting in a prominent cryosphere with widespread land and sea ice, as well as permafrost. However, a well-developed cryosphere in the Arctic is a relatively recent development. Paleoclimate records indicate that major ice sheets did not arise until about 3 - 3.5 million years ago (Spielhagen et al., 1997; Haywood et al., 2002; Moran et al., 2006; Lunt et al., 2008a; Mattheissen et al., 2009). Throughout much of the past 100 million years the Arctic was relatively warm and largely ice-free. During the Eocene (55-34 million years ago) the Earth experienced extreme greenhouse conditions, producing global mean annual temperatures approximately 12°C warmer than present (Zachos et al., 2001). During the Eocene, high-latitude winter temperatures were as much as 20°C warmer than present (Greenwood and Wing, 1995) and Greenland was mostly ice free (Edgar et al., 2007). After the Eocene Thermal Maximum (55 million years ago) Cenozoic climate gradually cooled and entered what has been termed an ‘icehouse’ phase of climate (Zachos et al., 2001).
1.2.1 Causes of Pliocene Cooling

Over the past 4 million years, Earth’s climate has undergone a transition from warmer to cooler climates, increasing approximately 3° C in global mean temperature over this interval (Haywood and Valdes, 2004; Ravelo et al., 2004). The potential cause of Late Cenozoic cooling is a hotly debated topic (e.g. Raymo 1994; Rea et al., 1998; Maslin et al., 1998; Philander and Fedorov, 2003; Ravelo et al., 2004; Bartoli et al., 2005; Fedorov et al., 2006; Molnar and Cane, 2007; Lunt et al., 2008b).

Among the many hypotheses that have been advanced to explain Pliocene cooling is the ‘Panama hypothesis’, which suggests that the tectonically driven closure of the Panama seaway between 13 and 2.5 million years ago drove cooling (Haug and Tiedemann, 1998; Lunt et al., 2008b). The ‘Panama hypothesis’ suggests that the closure of the Isthmus of Panama led to an increase in the salinity contrast between the Pacific and Atlantic Oceans, increased oceanic northward heat transport and warmer surface waters in the North Atlantic. The changes in salinity causes increased rates of North Atlantic Deep Water (NADW) formation associated with strengthened flow of the Gulf Stream. Increases in the temperature of the North Atlantic would have led to increased atmospheric moisture in the Arctic and increased snowfall on Greenland. The coincident timing of changes in salinity in the Caribbean basin and the East Pacific, and increased ice rafted debris (IRD) near Iceland supports this hypothesis (Bartoli et al., 2005). Two recent modeling studies have sought to test this hypothesis with one study suggesting that perennial snow cover decreased at northern high latitudes when the Isthmus closed and that closure of the Isthmus of Panama did not intensify orbitally forced glaciation.
(Klocker et al., 2005). This study used an intermediate-complexity coupled atmosphere-ocean model (ECBILT-CLIO) to look at an open and closed Isthmus of Panama under different orbital configurations. The second modeling study used output for the Pliocene from a fully coupled, fully dynamic ocean-atmosphere general circulation model (HADCM3) to drive an ice sheet model (GLIMMER) of Greenland and found that although the closure of the Isthmus of Panama did intensify Atlantic thermohaline circulation and enhance precipitation over Greenland the ultimate volume difference in ice sheets between the open and closed configurations was not significant (Lunt et al., 2008b). The largest hurdle yet to be overcome by this hypothesis is the question of timing of the closure of the Isthmus of Panama (Lunt et al., 2008a; 2008b).

The ‘ENSO hypothesis’ holds that a ‘permanent El Niño-like state’ existed in the Early Pliocene and terminated in the late Pliocene (Philander and Fedorov, 2003; Wara et al., 2005; Haywood et al., 2005; Shukla et al., 2009). The transmission of El Niño-like anomalies from the tropics propagated north through large-scale planetary waves that warmed the high latitudes of the Northern Hemisphere (Philander and Fedorov, 2003). Thus a ‘permanent El Niño-like state’, or El Padre (Shukla et al., 2009), would have acted to prevent the onset of glaciation and the loss of this state would have acted as a positive forcing on Northern Hemisphere glaciation (Philander and Fedorov, 2003). Evidence for the ‘ENSO hypothesis’ includes an increasing zonal gradient in sea surface temperature during this time period (Wara et al., 2005; Haywood et al., 2005) in the eastern and western Equatorial Pacific.
Recent studies have investigated whether this ‘permanent El Niño-like state’ is in fact permanent (Wunsch, 2009; Bonham et al., 2009). Wunsch (2009) postulated that the El Padre permanent El Niño state observed in the proxy record could instead be explained either by a higher frequency ENSO cycle or by stronger El Niño events. Proxy records don’t have the temporal resolution to distinguish a permanent El Niño from either more frequent or more intense El Niño events and so could appear to be a ‘permanent’ El Niño. Bonham et al. (2009) demonstrate that a climate model can generate the same effects as a ‘permanent El Niño’ by simply warming the Earth. They suggest that rather than being driven by El Niño a deeper thermocline in the tropical Pacific can instead be obtained via a reduction in heat loss at high latitudes and resultant reduction in poleward heat transport by the ocean, an idea that was advanced by Huang and Pedlosky (2000) and by Fedorov et al. (2006).

The ‘uplift hypothesis’ suggests that the Cenozoic uplift of the Rocky Mountains and Himalayas caused larger Rossby wave amplitude and jet-stream deflection, bringing cooler air masses and increased moisture and snowfall to the Greenland ice sheet (Ruddiman and Kutzbach, 1989; Raymo and Ruddiman, 1992). However, recent studies show that there has been little change in the height of the Rockies since well before the Pliocene (McMillan et al., 2006; Moucha et al., 2008) suggesting that this hypothesis may need revising.

The ‘snow gun hypothesis’ is based on changes in the temperature of the North Pacific. $\delta^{18}O$ records from the North Pacific ODP site 882 show a warming of ~4° C in sea surface temperatures ~80 ka before the expansion of the ice sheets, perhaps related to
changes in ocean circulation (Maslin et al., 1996). Maslin et al. (1996) suggest that the warm water in the North Pacific would have provided a source of moisture for the enhanced precipitation necessary in the Arctic to form ice sheets in the Arctic and in Greenland.

Finally, decreasing levels of atmospheric CO\(_2\) could have resulted in cooler melt-season temperatures and decreased ablation, resulting in the net accumulation of snow necessary for glaciation of Greenland. Decreasing atmospheric CO\(_2\) has been suggested as the primary mechanism for intensification of Pliocene glaciation (Maslin et al., 1998; Lunt et al., 2008a). Unfortunately records of atmospheric CO\(_2\) for the Pliocene are sparse. However, recent work spurred by an interest in understanding the sensitivity of Pliocene climate to CO\(_2\) forcing as a means of understanding the possible future response of climate to changing CO\(_2\) has resulted in several new studies of levels of atmospheric CO\(_2\) during this important interval. Estimates of atmospheric CO\(_2\) for the warm part of the Pliocene (3.5-5 million years ago) are between 350 and 450 ppm (Pagani et al., 2010; Seki et al. 2010; Tripati et al., 2010), near 2010 levels of 385 ppm (Fig 2). All three studies found that a decline in atmospheric CO\(_2\) began around 3.2 to 3.5 million years ago with atmospheric CO\(_2\) levels falling to between 250 and 300 ppm by ~2.8 million years ago (Pagani et al., 2010; Seki et al., 2010; Tripati et al., 2010). This timing coincides with the beginning of large-scale Northern Hemisphere glaciation when rapid expansion of the Greenland, Scandinavian and Canadian ice sheets occurred beginning 2.75 million years ago (Kleiven et al., 2002). Also coincident with decreasing atmospheric CO\(_2\) was a rapid cooling of the North Pacific indicated by an abrupt drop of ~7.5 °C in sea-surface
temperatures at around 2.7 Ma (Maslin et al., 1996). There was also an increase in ice-rafted debris in the North Atlantic (Kleiven et al., 2002) and increases in benthic foraminiferal $\delta^{18}$O values indicative of cooling in the deep ocean (Tiedemann et al., 1994; Shackleton et al., 1995; Lisiecki and Raymo, 2005). It is also interesting to note that CO$_2$ levels of between 250 and 300 ppm are around the level of ~280 ppm suggested by DeConto et al. (2008) as the threshold of atmospheric CO$_2$ below which development of large-scale Northern Hemisphere glaciation is possible. Using Pliocene GCM output data in an off-line Greenland ice sheet model Lunt et al. (2008a) demonstrated that of all the hypotheses put forward to explain Pliocene glaciation, only decreasing CO$_2$ produced significant changes in the amount of ice present on Greenland. Pagani et al. (2010) reconstructed CO$_2$ records for the Pliocene (Fig 2) and assessed earth system sensitivity based on their results of Pliocene atmospheric CO$_2$ levels of 365 to 415ppm.

Based on these atmospheric CO$_2$ levels and estimated Early Pliocene mean global temperatures ~4°C warmer than pre-industrial would mean an Earth system sensitivity of 9.6 ± 1.4°C per CO$_2$ doubling (Pagani et al. 2010). Assuming that the conclusions of Lunt et al. (2008a) are correct and CO$_2$ levels exert the strongest control on Pliocene temperatures this would mean that Earth system sensitivities to CO$_2$ in the warmer Pliocene world are substantially higher than the range of 2 to 4.5 °C per CO$_2$ doubling climate sensitivity discussed in the IPCC (Solomon et al., 2007; Pagani et al., 2010). This suggests that key differences between the Pliocene and the present (lack of Greenland ice, open Isthmus of Panama, reduced pole-equator temperature gradient, ice-free Arctic Ocean) act to make the Pliocene climate system more sensitive to changes in CO$_2$. The
causes of the Earth system sensitivity in the Pliocene need to be understood as they may have direct relevance to our estimates of future climate sensitivity, which would need revision if, for example, the future world has no Greenland Ice Sheet.

The entire interval of Pliocene warmth occurred when atmosphere levels of CO$_2$ were likely between 365 and 450 ppm (Pagani et al., 2010; Tripati et al., 2010; Seki et al., 2010) (Fig 3). With current atmospheric CO$_2$ at 392 ppm and climbing, Earth’s climate is rapidly moving beyond the point where the Pliocene can be used to assess the sensitivity of climate to future levels of CO$_2$. However, the Pliocene still represents the most accessible example of a warmer world from which to examine the effects that warmer temperatures may have in redistributing ecosystems and changing climate feedbacks, so understanding Pliocene climate is still useful to our understanding of the future. Accurate proxy estimates of temperatures from the Pliocene, such as those presented in this dissertation can help set boundary conditions for Pliocene models and thus help to understand those climate feedbacks that may be linked to an overall warmer world. Understanding climate feedbacks is key to helping make better forecasts of future climate.
1.2.2 Mechanisms of Pliocene cooling

The question that must then be asked is what drove the decrease in CO₂ during the Pliocene, but perhaps clues lie in processes linked to the other hypotheses for Pliocene cooling. Closure of the Isthmus of Panama would have affected deep thermohaline circulation and thus contributed to a substantial reorganization of the ocean-atmosphere system. Deep ocean circulation in the North Atlantic is an important sink for CO₂, and thus this could have contributed to the draw down of atmospheric CO₂. Changes in salinity between the Pacific and Atlantic as well as increases in Arctic precipitation could have reorganized circulation in the Arctic Ocean. In the Early Pliocene the flow through
the Bering Strait was from Arctic to the Pacific (Marincovich, 2000; Lindstrom, 2001) and Arctic Ocean surface waters were dominated by the influx of Atlantic water. Today relatively fresh North Pacific water flows through the Bering Strait into the Arctic Ocean and subsequently into the North Atlantic (Woodgate and Aagaard, 2005). Strengthened flow of the Gulf Stream resulting from closure of the Isthmus of Panama would have resulted in denser saltier North Atlantic water and increased rates of North Atlantic Deep Water (NADW) formation (Haug et al. 2001). Dense North Atlantic water flowing into the Arctic Ocean would have become capped by less dense relatively fresh water flowing through the Bering Strait from the North Pacific and increased freshwater influx to the Arctic Ocean from run-off. The less salty North Pacific water flowing through the Bering Strait would have formed a cap over the warmer saltier Atlantic water and resulted in the stratification of the Arctic Ocean. This would have prevented heat flux from the warmer Atlantic-derived water into the Arctic and most likely would have led to a cooling of Arctic air temperatures. Fresher water in the Arctic Ocean also could have made it easier for sea ice to form further stratifying the ocean. Arctic Ocean cooling could have further enhanced NADW formation. A recent paper by Hu et al. (2010) suggest that changes in flow through the Bering Strait as a result of sea level fluctuations during the past 116,000 years had a profound effect on North Atlantic meridional overturning circulation (MOC). Hu et al. (2010) suggest that when the Bering Strait is closed and there is no inflow of Pacific water into the Arctic Ocean, the North Atlantic is less stratified and MOC is more vigorous resulting in increased temperatures in the Northern Hemisphere. The effect is reversed if the Bering Strait is open. It is thus possible that changes in the Atlantic-Pacific
salinity contrast could resemble an open or closed Bering Strait. Waddell et al. (2009) have suggested that storage of CO$_2$ in the Southern Ocean was enhanced around this time as well, related to the continued cooling of Antarctica and its surrounding waters.

The closure of the Isthmus of Panama could have rearranged circulation patterns in the tropical Pacific including enhancing upwelling in the eastern Pacific, as suggested by (Ravelo et al., 2004). This could have reorganized ENSO dynamics in the tropical Pacific perhaps by strengthening easterlies owing to a more pronounced zonal gradient across the Pacific. This could perhaps have resulted either in a less El Niño-like state, or if Wunsch (2009) and Bonham et al. (2009) are correct, less frequent less intense El Niño events. Enhanced upwelling in the Pacific could also have contributed to an enhanced biological pump to draw CO$_2$ out of the atmosphere (Marlow et al., 2000). Etourneau et al. (2009) identified that enhanced zonal wind systems developed in the South Atlantic at this time, strengthening upwelling related to the Benguela current. Enhanced zonal wind systems in the South Atlantic may also indicate enhanced zonal wind throughout the tropics suggesting that it may be reasonable to assume that easterlies in the tropical Pacific may have also strengthened at this time.

It seems that somehow the closure of the Isthmus of Panama coincided with a major reorganization of the ocean-atmosphere system transitioning from a relatively sluggish warm Pliocene ocean to the colder more energetic Pleistocene ocean. One key question is how such changes could relate to future warming? If, as has been suggested, all of these changes are linked to the closure of an oceanic gateway, such as the Isthmus of Panama, how relevant is Pliocene warmth to questions of future warming? The answer
lies in the apparent sensitivity of the Pliocene climate system to changes in levels of atmospheric CO$_2$ regardless of the mechanism that caused the change in CO$_2$ over the Pliocene (Lunt et al., 2010; Pagani et al., 2010).

1.2.3 Paleoproxy evidence of Pliocene climate and Milankovitch forcing

A large number of the well studied fossil forest sites in the Arctic date to the late Pliocene-early Pleistocene (between 2 and 3 million years) (Matthews and Ovenden, 1990; Funder et al., 2001; Elias and Matthews, 2002) during the transition from mid-Pliocene warmth through the onset of large-scale Northern Hemisphere glaciation (Mattheissen et al., 2009).

Elias and Matthews (2002) undertook temperature reconstructions from 11 Plio-Pleistocene fossil sites in North America ranging in age from 5.7 to 1.7 million years old. They used a mutual climate range (MCR) method using beetle fauna to reconstruct maximum summer temperatures ($T_{\text{max}}$). They found that for all assemblages dating between 5.7 and 2 million years old, estimates of $T_{\text{max}}$ ranged between 12.4 and 13.8$^\circ$ C regardless of age or location. If the temperature reconstructions of Elias and Matthews (2002) are presented in conjunction with Pliocene temperature from the high Arctic presented in this dissertation and in other studies we find similar results where temperatures estimated for the late Pliocene-early Pleistocene (2 to 3 Ma) are not substantially different from temperature estimates for the early Pliocene (4 to 5 Ma) (Fig 3).
Lisiecki and Raymo (2005), among others, have clearly shown that glacial-interglacial cycles driven by Milankovitch forcing were an important aspect of Pliocene climate. The amplitude of variability in the benthic δ¹⁸O stack between interglacial and glacial periods increased significantly at ~2.7 Ma during the onset of major Northern Hemisphere glaciation coinciding with an increase in the sensitivity of climate to precession (41 ka cycle) forcing (Lisiecki and Raymo, 2007) (Fig 2). The Lisiecki and Raymo (2005) benthic δ¹⁸O stack reveals that many of the interglacial periods indicated by lighter values of δ¹⁸O have δ¹⁸O values not much different (~0.5-1 ‰ heavier) from average δ¹⁸O values for the early Pliocene prior to the onset of Northern Hemisphere...
glaciation (Fig 3). Benthic δ¹⁸O is a proxy of the climate response of high-latitudes because it measures changes in deep-water temperatures and ice volume, which are controlled by high-latitude surface temperatures (Lisiecki and Raymo, 2007). If high Arctic Pliocene terrestrial deposits are considered in the context of the Lisiecki and Raymo (2005) stack (Fig 3) we note that the age control on the Pliocene terrestrial deposits is not well enough resolved to determine whether these represent glacial, interglacial or average conditions. Both the Bylot Island and Kap København sites show stratigraphic evidence of being interglacial deposits (Funder et al., 2001; this dissertation). This suggests that perhaps the lack of a large temperature difference between Plio-Pleistocene high Arctic terrestrial sites is because all of these deposits represent interglacial rather than average climates. This would also suggest that climates during periods of maximum insolation in the high Arctic did not change significantly over the Plio-Pleistocene transition, at least prior to 2 Ma, even with the onset of large scale Northern Hemisphere glaciation and overall Pliocene cooling. This could imply that perhaps glacial phases are more sensitive changes in Arctic feedbacks than interglacial phases and that ice volume may be the critical feedback that accounts for the high estimates of Pliocene Earth system sensitivity found by Pagani et al. (2010). This conclusion would agree with the modeling results of Haywood and Valdes (2004) who proposed that the major contributing mechanism to Pliocene warmth was the reduced extent of high-latitude ice sheets.
1.2.4 Modeling Pliocene climate

In order to understand the dynamics of Pliocene climate it is important to have a more global understanding. The mid-Pliocene warm period represented by PRISM (USGS Pliocene Research, Interpretation and Synoptic Mapping) has become an important interval for paleoclimate modeling, in part to determine whether General Circulation Models (GCMs) can successfully reconstruct climatic conditions significantly different from present-day using boundary conditions prescribed by available proxy data (Haywood et al., 2009). Aside from testing how GCMs respond to various climate forcings (e.g. Lunt et al., 2008a; DeConto et al. 2008) Modeling studies of the Pliocene have helped to shed light on the El Padre phenomenon (The permanent El Niño-like state) (Bonham et al. 2009) as well as being used to explore the sensitivity of the Greenland ice sheet to different climate forcings (Lunt et al. 2008b). Climate model reconstructions of this warm period are thus critical to both understanding past warm intervals and to refine the responses of climate models by improving their ability to recreate known warm climates. In order to prescribe boundary conditions for the Pliocene most models use the PRISM reconstruction. The twofold goals of the PRISM project are to reconstruct global conditions during the 3.3 to 3.0 Ma time interval and to develop a series of global scale, quantatative data sets for uses in modelling experiments of the mid-Pliocene (Dowsett et al., 1999). PRISM has undergone several iterations over its life span with PRISM1 incorporating 64 marine and 74 terrestrial sites representing sea surface temperature (SST), annual vegetation, continental ice, sea-ice, sea level and topography on a 2° X 2° grid (Dowsett et al., 1996).
PRISM2 advanced on this by including additional sites in the marine portion of the reconstruction, using a new SST (sea-surface temperature) interpolation set and a reduced sea level rise (Dowsett et al., 1999). Most importantly the PRISM2 data set provided the initial conditions for a series of fully coupled ocean-atmosphere model simulations of Pliocene climate (Haywood and Valdes, 2004; Haywood et al., 2007; Lunt et al., 2008b; Haywood et al., 2009a; Hill et al. in review). The latest iteration of PRISM (PRISM3D) further improves on PRISM2 by incorporating multiproxy re-examinations of SSTs, including both maximum and minimum peak SSTs to provide a climatological error bar. PRISM3D also includes revised land-ice and vegetation schemes based on independent vegetation and ice models (Hill et al., 2007; Salzmann et al., 2008) and adds deep-ocean temperature reconstructions and uses numerical models to augment fossil data (Dowsett et al., 2010).

Perhaps the most significant change, in terms of atmospheric patterns, between PRISM2 and PRISM3D is the topography used in each. PRISM2 relied used paleotopography estimates based on the work of Thompson and Flemming (1996). Based on available data Thompson and Flemming (1996) placed the maximum elevation of the Andes at ~2000 m, and the western North American cordillera at a maximum of 1500 m, both about half their modern elevations. Recent work reassessed the paleotopography of both the Andes and the North American cordillera, finding that both mountain ranges had near modern topography in the Pliocene (McMillian et al., 2006; Garzione et al., 2006; Markwick, 2007) PRISM3D now uses near modern topography as its base. PRISM3D is
to provide initial values for the next suite of Pliocene models that are being run as part of the PlioMIP model intercomparison project (Haywood et al., 2009b).

All iterations of PRISM rely on a global data set of proxy data to form the basis of their model. However, the PRISM data sets have a heavy marine bias and lack terrestrial records. Additionally there are very few high-latitude data sets that contribute to PRISM. Thus data sets such as are presented here can add crucial information to the PRISM project. PRISM4, which is currently just in the initial stages of development, plans to incorporate records with annual resolution to enhance the applicability of the PRISM work (Dowssett et al., 2010). Annually resolved studies, such as are presented here will be valuable additions to the PRISM4 reconstruction.

1.3 The Northern Annular Mode in a warmer world

The leading mode of atmospheric variability in the extratropical Northern Hemisphere is the Northern Annular Mode (NAM) (Thompson and Wallace, 2000). The NAM has also been referred to as the Arctic Oscillation (AO) (Thompson and Wallace, 1998) and the North Atlantic Oscillation (NAO) (Visbeck et al., 2001; Hurell and Deser, 2010), although there is some debate as to whether the NAO is independent of the NAM (Hurell and Deser, 2010) the two processes are highly correlated. The fate of NAM in a warming world has received much attention (e.g. Sigmond and Scinocca, 2009 and references therein). Because variability in the NAM accounts for much of the variability in climate patterns in the Northern Hemisphere (Karpechko, 2010), understanding what is likely to happen to the NAM in the future is key to discussions of climate change. In the
same sense figuring out how the NAM operated in the warm Pliocene may be key to better understanding Pliocene climate.

Early model projections about the response of the eddy-driven jet (NAM) to climate change revealed a jet stream that shifted poleward and became stronger, evidence of a more positive NAM (Fyfe et al., 1999). Other models show little change over the present NAM (Miller et al., 2006). Recently this issue has received more attention with many papers suggesting that NAM will become more negative (Deser et al., 2010; Woolings et al., 2010; Morgenstern et al., 2010; Barnes et al., 2010; Kidston and 2010).

Deser et al. (2010) examined what effect simply removing sea ice would have on the NAO. Their model leaves atmospheric levels CO$_2$ constant and only change sea ice. This may be more similar to conditions during the Pliocene when atmospheric CO$_2$ was near present. Deser et al. (2010) found that removing sea ice produced a change in the position of the jet stream that resembled the negative phase of the NAO. This result was consistent with the response observed in Magnusdottir et al. (2004) and Seierstad and Bader (2009), who noted a consistent negative NAM circulation response with reduced Arctic sea ice.

Kidston and Gerber (2010) in an investigation of the Southern Annular Mode (SAM) noted that the more equatorward a model puts the jet stream the further poleward it will shift in response to warming. This is important because most GCMs are biased towards a more equatorward jet stream. Kidston and Gerber (2010) speculate that this is because the more equatorward jet stream has more persistence and more autocorrelation and so is more stable than a jet that is poleward. Barnes and Hartmann (2010) take this a step further and note that if average North Atlantic jet stream is shifted poleward, then the
persistence of the poleward-shifted phase (positive NAM) is reduced. This is consistent with the findings of Woolings et al. (2010). The lack of persistence in the poleward shifted phase, as noted by both Barnes et al. (2010) and Kidston and Gerber (2010), is related to lower wind speeds resulting in a less stable jet. In terms of frequency of the NAM, Morgenstern et al. (2010) produced modeled NAM indices showing the negative shift in the model-produced NAM index, however, it is interesting to note that although the trend in the NAM is towards generally more negative values there is still interannual variability in the NAM. This is consistent with the findings of Barnes and Hartmann (2010) who noted that although persistence in the poleward shifted phase (positive NAM) was reduced, persistence in the equatorward shifted phase (negative NAM) remained unchanged.

Karpechko (2010) pointed out that there is still considerable uncertainty in model projections of the future NAM with differences between models accounting for up to 40% of the variance in projections of surface temperature and precipitation in the Northern Hemisphere. This suggests that understanding how the NAM changes in a warmer world is crucial to understanding future climate change. Perhaps studies of the Pliocene NAM such as that undertaken by Hill et al. (in review) are an important step in that process.
1.4 Technical Background

1.4.1 Isotopic studies of freshwater molluscs as a paleothermometer

Isotopic records of oxygen ($\delta^{18}$O) from calcite and aragonite have long been recognized as an important paleotemperature proxy (McCrea, 1950; Epstein et al., 1953; O’Neil et al., 1969). There is a strong temperature dependence on the fractionation of oxygen isotopes during the formation of calcite. In order to understand the fractionation process, however, the isotopic value of the initial water must be known (Grossman and Ku, 1986). With carbonate derived from a marine setting, this determination is fairly straightforward because the isotopic value of the ocean is generally inferred to be ~ 0 ‰ and varies according the continental ice volume (Lear et al. 2001). For freshwater carbonate this estimate is more difficult to make and several approaches have been made in order to use $\delta^{18}$O values of freshwater organisms as paleotemperature proxies. Some studies have used the inferred minimum growth temperature of the organism and equated that with the highest $\delta^{18}$O value measured, as carbonate $\delta^{18}$O is more enriched at colder temperatures, and then inferred other temperatures using that baseline (Patterson et al., 1993). Other studies have used proxies to reconstruct the source water $\delta^{18}$O value and then used the reconstructed source water $\delta^{18}$O to infer temperatures from freshwater molluscs (Fricke and Wing, 2004). Provided there is some means of estimating source water value, then the same well-constrained calcite-temperature relationship established for the marine system can be applied.
A recently developed approach is carbonate ‘clumped isotope’ thermometry. ‘Clumped isotopes’ are based on the proportion of $^{13}\text{C}$ and $^{18}\text{O}$ bound to each other within the mineral matrix. The basis for clumping these heavy isotopes into bonds with each other is the thermodynamically controlled exchange of stable isotopes amongst isotopologues of calcium carbonate, or carbonate ions, in the solution from which carbonate is precipitated (Ghosh et al., 2006; Schauble et al., 2006). This means that a progressively more random distribution of heavy isotopes amongst all possible isotopologues is favored with increasing temperatures (Schauble et al., 2006). The abundances of heavy, mass-47, isotopologues of CO$_2$ are reported using the notation $\Delta_{47}$ and is defined as the difference in per mil between the measured 47/44 ratio of the sample and the 47/44 ratio expected for that sample if its stable carbon and oxygen isotopes were randomly distributed amongst all isotopologues (Eiler and Schauble, 2004). Because this relationship relies on the temperature dependence of heavy isotopologues against the random distribution of all other isotopologues, ‘clumped isotope’ thermometry does not require knowledge of source water $\delta^{18}\text{O}$ in order to determine temperatures. Two existing studies have calibrated the carbonate ‘clumped isotope’ thermometer in molluscs (Came et al., 2007; Huntington et al., 2009) and found it to be a reliable recorder of temperature with errors of between 1.2 and 2.4$^\circ$ C.

1.4.2 Stable isotope dendrochronology

Isotopic records of oxygen ($\delta^{18}\text{O}$), carbon ($\delta^{13}\text{C}$), and hydrogen ($\delta\text{D}$) from tree rings show a strong correspondence with instrumental climate records, such as, temperature (Burk and Stuiver, 1981; Feng and Epstein, 1995; Sauer et al., 2008),
precipitation (Epstein et al., 1977; Saurer et al., 1997; Robertson et al., 2001; Liu et al.,
2004) and humidity (Buhay and Edwards, 1995; Hemming et al., 1998; Edwards et al.,
2000; Wright and Leavitt, 2006). Changes in carbon isotopic ratios have also been related
to changing moisture conditions, such as drought (Leavitt and Long, 1989) and soil
moisture status (Buhay et al., 1996; Leavitt et al., 2002). In high-latitude forests δ¹⁸O has
been used to reconstruct mean annual temperature and the North Atlantic Oscillation
(Sidorova et al. 2009; Knorre et al., 2010; Sidorova et al. 2010). Carbon isotopic ratios in
high-latitude forests have been used to reconstruct growing season temperature (Porter et
al., 2009) and soil moisture status (Barber et al., 2000; Sidorova et al., 2009). Thus,
studies of isotopic records from tree rings have increasingly become a valuable tool for
dendroclimatology (Gagen et al., 2010). In studies of Quaternary and pre-Quaternary
wood, which lack a modern calibration period, climatic inferences from isotopic values
have started to become commonplace (Jahren and Sternberg, 2003; Ballantyne et al.,
2006; 2010; Richter et al., 2008).

Water absorbed by roots is translocated to distal branches without fractionation of
δ¹⁸O and δD until it reaches the leaves, where evaporative enrichment generates an
increase in δ¹⁸O and δD values of leaf water (Roden et al., 2000). Any oxygen derived
from CO₂ has already undergone exchange and equilibration with atmospheric H₂O and
so is also dependent on the oxygen isotope value of water (Roden et al., 2000). Because
the primary source of water for trees is soil water, the primary source of oxygen and
hydrogen isotopes is also soil water and therefore much of the isotopic signal should
reflect the isotopic signature of precipitation, modified by evaporative enrichment in both
the soil and the leaves. The depth at which the roots access the water can also affect the isotope value because of differing isotopic values of groundwater (e.g. whether the water is old or new, or is it from winter, spring or summer precipitation (McCarroll and Loader, 2004)).

Once a tree takes up water, no fractionation occurs until the water reaches the leaves, where transpiration preferentially removes a higher proportion of the common isotopes ($^{16}$O and $^1$H) from the tree water reservoir, increasing $\delta^{18}$O and $\delta$D ($^2$H) values (evaporative enrichment) of residual leaf water. When tree-ring cellulose is formed from glucose, there is some exchange between glucose oxygen and xylem water in the trunk. The possible exchange of glucose oxygen with xylem water may be as high as 40% of the oxygen exchanging during the synthesis of cellulose from glucose via triose-phosphate (Roden et al., 2000), dampening the signal imparted from evaporation in the leaves and making it more likely that the isotopic value of cellulose reflects primarily the isotopic value of meteoric water. Evaporation increases the isotopic ratio of leaf water, and therefore the isotopic ratio of glucose is enriched relative to the isotopic ratio of the original xylem water, the cellulose isotope ratio will also be higher. Thus, the dominant environmental signals recorded in the water isotopic ratios ($\delta^{18}$O and $\delta^2$H) are most likely the isotopic ratio of the source water and the summer humidity.

Water isotope ($^{18}$O/$^{16}$O and $^2$H/$^1$H) ratios of tree-ring cellulose depend mainly on the isotopic composition of the water used during cellulose synthesis, albeit with a humidity signal superimposed (Edwards and Fritz, 1986; Reynolds-Henne et al., 2007). The basis for the climate dependence of water isotope ratios in tree rings is, thus, the
temperature-dependent fractionation processes that occur during evaporation and condensation. A significant number of stable isotope dendroclimatology studies have shown positive correlations between $\delta^{18}O$ of tree rings and mean annual temperature (McCarroll and Loader, 2004; Treydte et al., 2007; Saurer et al., 2008; Sidorova et al. 2009).

Trees obtain carbon from CO$_2$ in the atmosphere. CO$_2$ enters the tree via the stomata, which can be constricted by guard cells to control water loss. Thus, carbon gain and water loss are intimately related. Plants preferentially use the lighter $^{12}$C as opposed to $^{13}$C because the bonds are easier to break. The amount of CO$_2$ in leaves can also depend on the rate of photosynthesis. When stomata are open, and when the rate of photosynthesis is low, the tree has access to more CO$_2$ and $\delta^{13}$C ratios are low. The opposite is true if the stomata are closed or the rate of photosynthesis is high (McCarroll and Loader, 2004). Stomatal conductance can vary in response to environmental controls such as, relative humidity (RH), temperature, and soil moisture status. For example when plants decrease stomatal conductance to limit water loss during warm or dry conditions this results in an increase in fixation of $\delta^{13}$C (CO$_2$) by the plant (Francey and Farquhar, 1982). Therefore, warm or dry periods should be characterized by increased $\delta^{13}$C values. Because the climate signal in $\delta^{13}$C is dominated by both stomatal conductance and photosynthetic rate, the same $\delta^{13}$C response can occur because of different climatic forcings in different locations (Saurer and Siegwolf, 2007). High $\delta^{13}$C values could equally be found to relate to a sunny growing season, at a cool moist site or to a dry summer a more arid site (Gagen et al., 2010). In both cases the climate variable in
question can also be highly correlated with temperature so correlation of the $\delta^{13}$C values with summer temperature would be expected (Gagen et al., 2010). Lipp et al. (1991) used $\delta^{13}$C values to reconstruct both relative humidity and temperature. A study that Edwards et al. (2000) expanded upon by combining both the relationships established by Lipp et al. (1991) to present $\delta^{13}$C values in RH-temperature space. At high latitudes both Gagen et al. (2007) and Porter et al. (2009) found that $\delta^{13}$C records could be used to reconstruct summer temperature. As high-latitude sites tend to be cool, moist sites, the temperature response at high-latitude sites seems to be primarily related to the dual control of temperature and summer irradiance on photosynthetic rate (McCarroll and Pawellek, 2001).

1.4.3 Stable isotope paleodendroclimatolgy

One of the most intriguing finds in recent years has been the abundant well preserved fossil forest sites dating from the Pliocene. Pliocene fossil forests have been described from Alaska (Matthews and Ovenden, 1990; Matthews et al., 2003), Siberia (Bondarenko, 2007), Greenland (Funder et al., 2001), Antarctica (van Bergen and Poole, 2002) and the Canadian Arctic Archipelago (Matthews and Ovenden, 1990; Ballantyne et al., 2006; Richter et al., 2008; Ballantyne et al., 2010; Barker et al. 2010; this dissertation) (Fig 1).

Isotopes in fossil wood have enormous potential for answering questions about past climates (Gagen et al., 2010). Apart from a short juvenile phase in carbon isotope
series (McCarroll and Loader, 2004; Leavitt, 2010) there are no long term age-related trends in isotopic series derived from tree rings (Gagen et al., 2010). This lack of age-related trends means that unlike ring-width series, isotopic series from fossil wood do not need to be detrended and indexed, thereby preserving their utility to reconstruct climatic variables for samples lacking a calibration period. Previous studies have also demonstrated that oxygen isotopic series derived from tree rings show much less inter-tree variability than ring-width series (Gagen et al., 2004; Treydte et al., 2006; Roden, 2008; Leavitt, 2010). Studies by Kirdyanov et al. (2008) have indicated that the expressed population signal (EPS) (Eq 1) derived from correlations among carbon isotope time series of four Siberian larch trees is 0.9 for a given site, which indicates the common site signal is strongly captured by only four trees. This was confirmed in subsequent studies of Siberian larch that used between four and five trees to obtain a reliable climate signal (Sidorova et al. 2008; Knorre et al., 2010).

\[
\text{EPS} = \left( n \times r_{\text{mean}} \right) ÷ \left[ (n \times r_{\text{mean}}) + (1 - r_{\text{mean}}) \right]
\]

(1)

Where \( n \) is the number of trees at the site with separate isotope series and \( r_{\text{mean}} \) is the mean correlation coefficient among all pairs of isotope series. An EPS value of 0.85 is generally considered the benchmark for identifying the number of trees necessary to capture 85% of the variance at a site (Leavitt, 2010). Several studies have evaluated the minimum number of trees necessary to capture a representative isotope signal finding that in general isotopic analysis of relatively few trees can capture a representative site signal.
(Leavitt, 2010). This means that fossil wood deposits with perhaps only a few specimens present are still likely to contain a common site signal.

In recent years, studies of isotopic values in well-preserved examples of ancient wood have begun to answer questions about past climates, Pleistocene-Holocene wood has been investigated in several studies (e.g. Friedrich et al., 1999; Mayr et al., 2003; Leavitt et al., 2006). Jahren and Sternberg (2003; 2008) investigated $\delta^{13}C$, $\delta^{18}O$ and $\deltaD$ ratios of Eocene-aged (40-45 million year old) wood to reconstruct the paleotemperature and paleohumidity of a fossil forest in the Canadian Arctic generating important climatic information for the middle Eocene warm period. Jahren and Sternberg (2008) measured $\delta^{13}C$ at high resolution by sampling intra-annually allowing for an assessment of leaf senescence. Jahren and Sternberg (2003) used $\delta^{18}O$ and $\deltaD$ ratios of fossil wood cellulose to reconstruct an Eocene summer paleohumidity of 67%. The methods they employed are used in this dissertation to estimate the paleohumidity of a Pliocene forest site on Ellesmere Island in appendix B.

The first study to use isotopic ratios of tree-ring cellulose to investigate Pliocene Arctic climate was a study by Ballantyne et al. (2006). They used $\delta^{18}O$ ratios of fossil wood from the Beaver Pond locality in conjunction with ring widths to reconstruct mean annual temperatures for the site. This study was limited, however, in the methods used to build their transfer function and to reconstruct temperatures. Ballantyne et al. (2006) built their transfer function using two living trees, one from the Dawson City, Yukon (a northern Boreal site) and one from Ottawa, Ontario (a southern Boreal site) and ten rings from each tree compared to climate data. For their temperature reconstruction only 20
rings from the Pliocene log were measured and used to infer temperature. In a subsequent study, Ballantyne et al. (2010) validated the results of the previous study by comparing mean annual temperatures reconstructed from isotopes of fossil wood with other proxy-based temperature reconstructions involving vegetation co-existence and bacterial tetraethers.

Richter et al. (2008a) did a comprehensive study to examine the potential of $\delta^{18}$O from wood of Eocene through Holocene age to reconstruct temperature and $\delta^{18}$O of meteoric water. They found that the latitudinal gradient of $\delta^{18}$O in precipitation has not changed radically in the past 45 million years. Richter et al. (2008a) based their temperature reconstructions on a continental network of wood $\delta^{18}$O established in another paper (Richter et al., 2008b). The drawback to this approach is that it eliminates the ability to resolve fine-scale regional differences in the relationship between $\delta^{18}$O and temperature.

1.5 Research Goals and Approaches

In this dissertation I have undertaken a variety of approaches to investigate past climate in the Pliocene Canadian Arctic. The goal has been to better understand climate during past warm periods. The first study uses two novel proxy approaches, stable and ‘clumped’ isotope value of freshwater molluscs and moss to characterize growing season temperatures and meteoric water values for an early Pliocene (4-5 Ma) high-Arctic forest deposit on Ellesmere Island, Canada. The second study uses stable oxygen, carbon and hydrogen isotopes of fossil wood samples from the same forest deposit on Ellesmere
Island to reconstruct annually resolved records of growing season and mean annual temperature. This study also uses the fossil isotopic values to reconstruct the isotopic value of precipitation, and to infer relative humidity. The third study takes the methods developed in the second study and applies them to a younger (2.4 to 2.8 Ma) forest deposit on Bylot Island, Canada. Each appendix is presented as an independent study; yet in summary they provide a valuable contribution to understanding climate during past warm periods and help provide clues as to what we may expect under future warming.
CHAPTER 2 – THE PRESENT STUDY

This dissertation employs several methods to gain an understanding of Pliocene Arctic climate. Each of these methods provides independent means of estimating temperature with two studies focused on the early Pliocene (4-5 Ma) warm period and the third study focused on the Plio-Pleistocene transition.

2.1 Estimates of Arctic land surface temperatures during the early Pliocene from two novel proxies

In this study two independent proxies for terrestrial growing-season temperatures are used to reconstruct temperature and the oxygen isotopic compositions of meteoric water for the Early Pliocene Arctic. Samples come from the Beaver Pond site, a well-studied fossil locality at Strathcona Fiord, Ellesmere Island dated biostratigraphically to between 4 and 5 Ma. The first of these proxies involved using $\delta^{18}O$ values of cellulose from well-preserved aquatic mosses to constrain the $\delta^{18}O$ values of the water in which the molluscs grew. The $\delta^{18}O$ values of aragonitic freshwater molluscs preserved within the same peat as the moss were then measured and the Grossman and Ku (1986) equation, modified by Dettman et al. (1999) was subsequently used in order to calculate mollusc growth temperatures, inferred to represent May to September. This approach yielded $\delta^{18}O$ values of meteoric water of $-20.7 \pm 0.3\%e$ and an average growing-season temperature of $14.2 \pm 1.3^\circ C$. Temperatures and meteoric water values were also independently derived via ‘clumped isotope’ thermometry of additional mollusc shells from the same site by my co-authors Aradhna Tripati, Rob Eagle and John Eiler. This approach yielded $\delta^{18}O$ values
of meteoric water of $-22.1 \pm 0.7\%$ and an average growing-season temperature of $10.2 \pm 1.4^\circ C$. Meteoric water values are 3-5 $\%$ more enriched than present Ellesmere Island precipitation values from Strathcona Fiord and Eureka indicating that the latitudinal gradient of isotopes in precipitation may have been different in the early Pliocene. Temperatures calculated via both methods are 11-16$^\circ C$ warmer than present Ellesmere Island temperatures (May-Sept temperature = $-1.6 \pm 1.3^\circ C$ n = 50 yrs). Complete results are presented in Appendix A and in Csank, A.Z., Tripati, A.K., Patterson, W.P., Eagle, R.A., Rybczynski, N., Ballantyne, A.P., Eiler, J.M. 2011. Estimates of Arctic land surface temperatures during the early Pliocene from two novel proxies. Earth and Planetary Science Letters 304, 291-299.

2.2 Climate variability in the Early Pliocene Arctic: Annually resolved evidence from stable isotope values of sub-fossil wood, Ellesmere Island, Canada

In this study, $\delta^{18}O$, $\delta D$ and $\delta^{13}C$ composition was measured on wood preserved in early Pliocene (4-5 Ma) deposits at Strathcona Fiord, Ellesmere Island, Canada (ca. 78$^\circ N$). Four well-preserved tree trunks, identified via wood anatomy as *Larix* (larch), from this high Arctic site were measured in order to provide an annually resolved, well replicated record of Pliocene climate, although, $\delta^{13}C$ and $\delta D$ values were only measured from a single specimen (US357). Isotopic values from US357 were measured as part of my MSc thesis (Csank, 2006) and these isotope results have been reanalyzed and reinterpreted here. Although we focus primarily on the $\delta^{18}O$ values, in conjunction with ring-width measurements, to derive annually resolved mean annual and June-July
temperature and δ\textsuperscript{18}O values of precipitation from the site; we also use δ\textsuperscript{13}C to reconstruct June through August temperature and relative humidity and δD to reconstruct relative humidity and δD values of precipitation. Previous studies of fossil wood have focused on δ\textsuperscript{18}O for temperature reconstructions because it has a well known mechanistic relationship to temperature (Jahren and Sternberg, 2003; Ballantyne et al., 2006; 2010; Richter et al., 2008). Previous studies may also have focused on δ\textsuperscript{18}O because it is unclear to what extent carbon and hydrogen isotopes could be influenced by polar light conditions (Yang et al., 2009) and higher levels atmospheric CO\textsubscript{2} (Pagani et al., 2010; Tripati et al., 2010). Using transfer functions for δ\textsuperscript{18}O and δ\textsuperscript{13}C we determined Pliocene growing season (JJ) temperatures of 12.3-21.1 ± 2 °C, 11-13 °C warmer than the average 1950-1990 Ellesmere Island JJ temperature of 4.0 ± 1.2 °C. Pliocene mean annual temperatures (MAT) of –1.4 to 1.2 ± 1.9 °C, 17-20 °C warmer than the average 1950-1990 Ellesmere Island MAT of -19.7 ± 1.3 °C were determined using transfer functions and a mechanistic model approach using δ\textsuperscript{18}O values. The range presented represents the maximum and minimum average values of time series from all four Pliocene specimens. Interannual variability in temperatures is on the order of 5 °C, equivalent to the modern interannual variability in MAT for Eureka, Ellesmere Island (4.3 °C). Estimated isotope values of precipitation of –21 ‰ (δ\textsuperscript{18}O) and –150 ‰ (δD) were calculated from the isotopic values of wood cellulose. Relative humidity was estimated by both δ\textsuperscript{13}C and δD to be between 60 and 80%. Spectral analysis of isotopes and ring widths indicates significant periodicities around 20, 10-12, 5.5, 4-5, 3-3.5 and 2.3-2.8 years suggesting that decadal and sub-decadal modes of variability, similar in period to modern modes,
may have existed in the early Pliocene Arctic. Results are presented in Appendix B as a manuscript in review in the journal *Palaeogeography, Palaeoclimatology, Palaeoecology*.

2.3 Annually resolved temperature reconstructions for a late Pliocene-early Pleistocene polar forest on Bylot Island

This study involved paleotemperature reconstructions for a late Pliocene-early Pleistocene fossil forest from Bylot Island, Canada believed to be between 2.4 and 2.8 million years old. Isotopic analysis of $\delta^{18}$O from tree rings are used to reconstruct mean annual temperatures and precipitation $\delta^{18}$O values. This study also represents the first instance of a crossdated series of Pliocene-age wood used for paleoclimate reconstructions. Temperature estimates were derived from averaged $\delta^{18}$O values of tree rings from four specimens selected from the crossdated series of ten specimens.

Estimates of MAT average $-2.9 \pm 3.9\, ^\circ\text{C}$ or $11.9 \pm 4.5\, ^\circ\text{C}$ warmer than present day Bylot Island ($-14.8 \pm 2.2\, ^\circ\text{C}$). Growing season (June-July) temperatures are inferred to be $13.5 \pm 1.1\, ^\circ\text{C}$ or $12.6 \pm 1.6\, ^\circ\text{C}$ warmer than present day ($4.2 \pm 1.2\, ^\circ\text{C}$). Meteoric water $\delta^{18}$O values average $-16.2\, \%$ or 2-6\, $\%$ heavier than present values. Stratigraphically, wood deposits on Bylot Island sit above diamictite and the Bylot Island section is thought to be correlative with the Kap København deposits (Funder et al., 2001) on Greenland. Both of these deposits are inferred to represent interglacial deposits within the Plio-Pleistocene. That temperature calculated here and temperatures from Kap København both match temperatures for the mid-Pliocene warm period suggesting that temperatures
recorded during interglacial times may not be indicative of overall Pliocene cooling. Complete results are presented in Appendix C as a manuscript to be submitted to the Journal *Palaeogeography, Palaeoclimatology, Palaeoecology*.

2.4 Summary & Future Work

Temperatures calculated here using both molluscs and tree rings indicate growing season temperatures were on the order of 10 to 16 °C warmer than present (1950-1990) Ellesmere Island temperatures ($\Delta T_p$) during the Early Pliocene (4-5 Ma). Growing season temperatures calculated using tree ring records from a younger site (2.4-2.8 Ma) on Bylot Island are estimated to be 8 to 13 °C warmer than present (1923-2000) ($\Delta T_p$) temperatures. MAT determined for the early Pliocene Beaver Pond site are estimated to be 18-20 °C warmer than present (1950-1990) Ellesmere Island temperatures ($\Delta$MAT). $\Delta$MAT from the younger Bylot Island site is estimated to be 8 to 15 °C warmer than present (1923-2000) temperatures.

The difference in $\Delta T_p$ between the early Pliocene and late Pliocene sites illustrated here is only 2 to 4 °C, an interesting result if we consider that the onset of large-scale Northern Hemisphere glaciation began at around 2.75 Ma (Fig. 2). However, the Bylot Island site likely represents interglacial conditions suggesting that this difference most likely represents only the difference between the generally warmer early Pliocene and a warm interglacial during the generally cooler late Pliocene/early Pleistocene.

Although the studies presented here have addressed questions related to temperature reconstructions of the Arctic during the Pliocene, there are still unanswered questions to
be addressed in future work. The study conducted in Appendix B was limited to 4 samples that were not systematically collected from the rich deposit at the Beaver Pond site. The Beaver Pond site contains an abundance of fossil wood in many different stratigraphic layers, which presents the possibility of improving this study by collecting specimens from the same stratigraphic layer allowing for the possibility of crossdating, as was done in Appendix C. The abundant material also presents the possibility of reconstructing multiple different intervals in the early Pliocene at annual resolution by sampling wood from multiple stratigraphic layers up the entire 90 m Beaver Pond section. This could build a more detailed picture of the evolution climate through the Beaver Pond site. Crossdated records from the Beaver Pond site would allow further investigation of the interannual periodicities seen in the tree-ring width and tree-ring isotope records from this site that could allow for more concrete conclusions about the mechanism behind the interannual variability seen at this site.

Ongoing work in the field of tree ring isotopes continues to reveal the ways in which isotopes of carbon, oxygen and hydrogen in tree rings record environmental variables. It would be useful to work more on investigating the environmental relationships of tree rings and tree ring isotopes in modern analogue environments so as to develop transfer functions that could utilize all isotopic systems to reconstruct environmental parameters from Pliocene wood samples with greater confidence.

In Appendix A, a comparison between water δ^{18}O determined from aquatic moss and water δ^{18}O determined by ‘clumped isotope’ analysis of freshwater mollusc shells revealed that the results of these two techniques are within error of each other. This
presents the possibility to use freshwater mollusc shells from Pliocene deposits globally to reconstruct the latitudinal gradient of meteoric water $\delta^{18}O$ for the Pliocene. This would allow for greater accuracy in proxies, like the tree-ring $\delta^{18}O$ proxy, that are influenced by the isotopic value of meteoric water in addition to climate.

With the abundant Plio-Pleistocene aged forests located throughout the Arctic (Fig 1), it would be possible to take the techniques developed in this dissertation and use them to investigate climate and climate variability throughout the Arctic. Sites at Meighen Island (Canada), Kap København (Greenland), Prince Patrick Island (Canada), Ellef Ringnes Island (Canada), Ch’jees Bluff (Canada) and Circle (Alaska) have all been dated to between 3 million and 2.6 million years. Climate reconstructions from these sites would allow for a more whole-Arctic approach to Pliocene climate reconstruction. Additionally the greater spatial distribution of these sites would allow for a more comprehensive picture of climate that could contribute information on interannual climate variability to model reconstructions of the Pliocene, like PRISM4.

These high Arctic forests sites represent a rich source of information for both the paleoclimatic and paleontologic communities and their full potential has yet to be exploited. It is my hope that in future years I will have the opportunity to continue my studies by investigating forest sites throughout the Arctic.
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APPENDIX A

ESTIMATES OF ARCTIC LAND SURFACE TEMPERATURES DURING THE EARLY PLEISTOCENE FROM TWO NOVEL PROXIES

Estimates of Arctic land surface temperatures during the early Pliocene from two novel proxies

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A.1 ABSTRACT

During the Pliocene (2.6 to 5 million years ago) atmospheric CO₂ levels have been estimated as similar to or slightly above present levels (Tripati et al., 2009; Pagani et al., 2010), and yet Earth’s climate was considerably different. Recent evidence suggests that although global temperatures were 2-3°C warmer than pre-industrial during
the Pliocene, Arctic warming may have been amplified. Thus precise temperature records of this interval are required to assess the sensitivity of Earth’s climate to persistent levels of CO$_2$ between 365 and 415ppm. We present records of two independent proxies for terrestrial growing-season temperatures at the Early Pliocene Beaver Pond site on Ellesmere Island. $\delta^{18}O$ values of cellulose from well-preserved peat constrain the $\delta^{18}O$ values of meteoric water to $-20.7 \pm 0.3\%$, which we combined with $\delta^{18}O$ values of aragonitic freshwater molluscs found within the peat in order to calculate mollusc growth temperatures. This approach results in an average growing-season temperature of $14.2 \pm 1.3^\circ$ C. Temperatures were independently derived by applying carbonate ‘clumped isotope’ thermometry to mollusc shells from the same site, indicating an average growing-season temperature of $10.2 \pm 1.4^\circ$ C. A one-way ANOVA indicates that the differences between the two techniques are not significant as the difference in mean temperatures between both methods is no different than the difference between individual shells using a single technique. Both techniques indicate temperatures $\sim 11-16^\circ$ C warmer than present (May-Sept temperature = $-1.6 \pm 1.3^\circ$C) and represent the first thermodynamic proxy results for Early Pliocene Ellesmere Island.

A.2 INTRODUCTION

An increasing number of studies, including the most recent Intergovernmental Panel on Climate Change (IPCC) report, have highlighted the early Pliocene climate as an example of climate conditions that could result from elevated atmospheric CO$_2$ driven by anthropogenic emissions (Jansen et al., 2007). Models of Pliocene climate suggest
temperatures ~2-3°C warmer globally, and 7-15°C warmer in the Arctic relative to today (Sloan et al., 1996; Haywood et al., 2009), which is within the range projected for the end of this century (Jansen et al., 2007). The Pliocene configuration of the continents was more similar to today than other periods of climatic warmth, and estimated Pliocene CO₂ levels of ~365-415ppm were similar to or slightly above present levels (Tripati et al., 2009; Pagani et al., 2010). Therefore, the Pliocene is, in many respects, an accessible example of near-future temperatures (Jansen et al., 2007; Haywood et al., 2009). However, accurate and detailed constraints on polar temperatures in the Early Pliocene are needed to test Pliocene model predictions, but have been limited to-date. New reconstructions of Pliocene Arctic temperatures are particularly needed, as such data would constrain the extent to which global warming is amplified at high latitudes and help assess the Earth System Sensitivity of global and Arctic climate to changing levels of CO₂ (Lunt et al., 2010).

Fossil forest sites in the Canadian Arctic, which contain well-preserved plants, vertebrates and invertebrates, are amenable to the use of two novel proxies to constrain Pliocene polar temperature. The Beaver Pond locality (78° 33’ N; 82° 25’ W) on Ellesmere Island, Canada, so named because of the abundance of beaver-gnawed sticks found at the site (Rybczynski, 2008), is particularly well characterized and is unique in having abundant well-preserved remains of both plants and animals. The study site is located in Central Ellesmere Island near the head of Strathcona Fiord (Fig A1). Stratigraphically the site lies within the Beaufort Formation, a predominantly fluvial unit that forms part of the Sverdrup Basin (Fyles, 1989).
The Beaver Pond section is composed of cross-bedded sand layers and arkosic gravels intercalated with fossiliferous peat, suggesting a fluvial depositional setting. Sand layers are cross-bedded in the coarser layers and planar bedded in the finer layers; both layers contain abundant plant macrofossils. Sand layers containing discontinuous lenses of peat indicate peat-forming units likely occurred within a braided river system. The gravel units probably represent gravel bars at the river bends or between channels. Peat deposits likely formed between channels and in the river flood plain. The Beaver Pond deposit itself is a thick peat layer that likely represents an oxbow lake deposit formed when a beaver dam was built across one of the channels. The entire sequence is capped by glacial deposits of Pleistocene age.

Flora includes macrofossils of bryophytes and vascular plants. The bryophytes consist mostly of extant species (Matthews and Ovenden, 1990), whereas the vascular plants contain some extinct forms including a larch (*Larix groenlandii*) (Tedford and
Harington, 2003; Ballantyne et al., 2006). Other vascular plants include spruce (*Picea*), pine (*Pinus*), cedar (*Thuja*), alder (*Alnus*), and birch (*Betula*) (Matthews and Ovenden, 1990; Ballantyne et al., 2010). The rich mammal fauna, as diverse as that of the modern Arctic tree-line, consists of beaver (*Dipoides*), rabbit (*Hypolagus*), a small canine (*Eucyon*), shrew (*Arctisorex*), bear (*Ursus abstrusus*), wolverine (*Plesiogulo*), badger (*Arctomeles*) and a hipparionine horse (*Plesiohipparion*). Biostratigraphic age control based on first appearances of these species provides an Early Pliocene (4-5 million years old) age for the site (Tedford and Harington, 2003). The age is primarily based on the species *Ursus abstrusus* that in North America dates from 3.5 to 5 million years ago and *Plesiohipparion* that dates between 4 and 5 million years ago (Tedford and Harington, 2003).

Although it is unusual to see carbonates preserved in peat deposits, the overall fluvial depositional setting offers some insight into why this is possible at this site. Ovenden (1993) states that the assemblages of bryophytes found at the site are indicative of species that existed in neutral to slightly acidic treed fens. Fen environments are fed by ground water and/or streams and thus are much higher in pH than ombrotrophic bogs (Vitt et al. 1993). Additionally, at least one of the species used in our study (*Gyraulus albus*) is tolerant of slightly acidic low carbonate environments (Boycott, 1936; Dussart, 1979; White et al., 1999). The molluscs likely lived among the aquatic vegetation of the fen. Hydrologically a fen environment fed by groundwater is not as susceptible to large seasonal fluctuations in the $\delta^{18}$O values of the source water (White et al., 1999; Zanazzi
and Mora, 2005). Turner et al. (2010) note that groundwater influenced channel fens and oxbow lakes in the Yukon tend to show slightly depleted values weighted towards winter snow melt, however, these groundwater fed systems exhibit little to no evaporative enrichment in δ¹⁸O values during the summer months leading to fairly consistent annual δ¹⁸O values.

Previous attempts to estimate early Pliocene temperatures in this region used non-thermodynamically based proxies. Estimates of summer temperatures based on beetle assemblages are 10 ± 2º C warmer than present (Elias and Matthews, 2002). Transfer functions based on δ¹⁸O values of wood cellulose have been interpreted to reflect mean annual temperatures ~14-19º C warmer than present (Ballantyne et al., 2006; Ballantyne et al., 2010). Estimates based on plant macrofossils and molecular techniques indicate mean annual temperatures ~19º C warmer than present (Ballantyne et al. 2010). These temperature calculations rely upon empirical transfer functions, whereas, our study relies upon the thermodynamically controlled isotopic distributions within carbonate minerals and between those minerals and co-existing water.

A.3 METHODS

We use two approaches, using carbonate stable isotope values, to provide both independent estimates of temperature and baseline δ¹⁸O values for Arctic meteoric water during the Pliocene. First, we use δ¹⁸O values of cellulose of aquatic moss derived from peat as a proxy for the δ¹⁸O value of meteoric water (Sauer et al., 2001; Ménot-Combes et al., 2002; Zanazzi and Mora, 2005). These reconstructed lake water δ¹⁸O values are
compared with δ\(^{18}\)O values of well-preserved freshwater molluscs found within the peat to calculate paleotemperatures using the temperature-dependent partitioning of \(^{18}\)O/\(^{16}\)O ratios between calcium carbonate and water (Grossman and Ku, 1986; Dettman et al., 1999). A second technique, carbonate ‘clumped isotope’ thermometry, examines the proportion of \(^{13}\)C and \(^{18}\)O that are bound to each other within the carbonate mineral lattice. The basis for the clumping of these heavy isotopes into bonds with each other is thought to be the thermodynamically controlled exchange of stable isotopes amongst isotopologues of calcium carbonate, or carbonate ions in solutions from which carbonate minerals precipitate (Ghosh et al., 2006; Schauble et al., 2006; Eiler, 2007; Eagle et al., 2010; Tripati et al., 2010). The equilibrium constant for the reaction: \(\text{Ca}^{12}\text{C}^{18}\text{O}^{16}\text{O}_2 + \text{Ca}^{13}\text{C}^{16}\text{O}_3 \leftrightarrow \text{Ca}^{13}\text{C}^{18}\text{O}^{16}\text{O}_2 + \text{Ca}^{12}\text{C}^{16}\text{O}_3\) (Reaction 1; or its equivalent for dissolved inorganic carbon species) forms the theoretical basis for carbonate clumped isotope thermometry, with the doubly substituted species of CO\(_2\) (or heavy isotope, mass-47, “clump”) produced during acid digestion slightly more stable than the other isotopologues. Thus, a progressively more random distribution of heavy isotopes amongst all possible isotopologues is preferentially favoured with increasing temperatures (Schauble et al., 2006). Abundances of mass-47 isotopologues are reported using Δ\(_{47}\) (Eiler and Schauble, 2004). The Δ\(_{47}\) value is defined as the difference in per mil between the measured 47/44 ratio of the sample and the 47/44 ratio expected for that sample if its stable carbon and oxygen isotopes were randomly distributed among all isotopologues, referred to as stochastic distribution.

The Δ\(_{47}\)-temperature relationship for most modern biogenic carbonates measured to date (Came et al., 2007; Eagle et al., 2010; Ghosh et al., 2006; Huntington et al., 2009; Eagle et al., 2010; Tripati et al., 2010) generally adheres to the inorganic calcite
calibration reported by Ghosh et al. (2006). Two studies have calibrated the carbonate clumped isotope thermometer in aragonitic molluscs (Came et al., 2007; Huntington et al., 2009) and found evidence for a Δ47-temperature calibration that is similar to inorganic calcite. As a result, we apply the inorganic calibration of Ghosh et al. (2006) to aragonitic molluscs from the Beaver Pond site to estimate temperature. However, as this is a new proxy, the full scope of potential kinetic isotope effects on carbonates has not yet been resolved. We note that there may be some evidence of kinetic isotope effects in certain biogenic carbonates (e.g. surface-dwelling coral, fish otoliths), although the growth temperatures of these calibration materials were poorly constrained (Ghosh et al., 2006; Ghosh et al., 2007; Tripati et al., 2010).

A.3.1 Methods: stable isotope analysis

Mollusc shells were separated from the peat and washed in an ultrasonic bath containing deionized water before being air-dried and examined to ensure no matrix was contained within the shells. Initially three gastropod shells (Gyraulus albus) were crushed with a mortar and pestle and analyzed as whole samples. Subsequent samples of mollusc shells, four gastropods (two Gyraulus albus and two Lymnae sp.) and two bivalve shells (Pisidium sp.) were sub-sampled using a Dremel dental drill fixed in place under a binocular microscope. Carbonate was sampled along growth banding for three of the specimens, for seasonal analysis, while bulk analyses were conducted on the remaining two specimens (Table A1). Samples were analyzed for δ18O values using a Kiel-III carbonate preparation device directly coupled to a Thermo Finnigan MAT 253 mass spectrometer in dual inlet mode, with an analytical precision of 0.1‰. δ18Oaragonite was calibrated and corrected to VPDB using the standard NBS-19. Initial (bulk) analyses
were carried out at the Saskatchewan Isotope Laboratory (SIL), University of Saskatchewan and additional samples were analyzed using an identical set up located at the Environmental Isotope Laboratory at the University of Arizona.

Moss fragments from the aquatic species (*Scorpidium scorpioides*) were isolated from the peat using a binocular microscope and tweezers to select only individuals of the desired species and washed in deionised waster. The moss stems were subsequently processed to α-cellulose using the sodium chlorite bleaching method outlined by Leavitt and Danzer (1993). Moss fragments were placed in heat-sealed mesh pouches immersed in a solution consisting of 1 L deionized water, 15 g of sodium chlorite and 10 mL of acetic acid. The solution with samples was heated to 70º C and left over night with another 10 g of sodium chlorite added every 4 hours during the subsequent day. Samples were left in solution at 70º C until lignin was completely removed as indicated by the samples turning completely white. Deionized water heated to 70º C was used to rinse the samples followed by a rinse in 20º C deionized water. Ten mL of 10% sodium hydroxide was added to the samples that were subsequently heated to 80º C for 2 hours, samples were removed from solution and rinsed with deionized water. This step ensures removal of all sugars and hemicellulose from the samples. The samples were rinsed in deionized water one final time and placed in a vacuum oven at 40º C to dry for 12 hours. $\delta^{18}O_{\text{cellulose}}$ values of moss were determined at the SIL using a Thermo Finnigan TC/EA coupled to a Thermo Finnigan Delta Plus XL mass spectrometer in continuous flow mode via a Conflo III interface with an analytical precision of 0.2‰, and standard deviation on repeat measurements of 0.3‰. $\delta^{18}O_{\text{cellulose}}$ values were calibrated and corrected to VSMOW using the IAEA 601 and 602 benzoic acid standards as well as an internal cellulose standard.
Table A1: Summary table of all stable isotope measurements. Uncertainties in the temperature calculation are based on the standard errors of both the stable isotope determination of calcite and the standard error of the meteoric water value inferred from moss cellulose. Average of all molluscs is determined as the average calculated from each individual mollusc.

Nine precipitation samples were collected from Strathcona Fiord, Ellesmere Island Canada in July, 2004. Samples were collected using a bucket and then transferred...
to 50 mL nalgene bottles. Bottles were filled to the top to eliminate headspace and the
tops were sealed tightly and then wrapped with electrical tape. δ^{18}O values of moss were
determined at the SIL using a Thermo Finnigan TC/EA coupled to a Thermo Finnigan
Delta Plus XL mass spectrometer in continuous flow mode via a Conflo III interface with
an analytical precision of 0.2‰, and standard deviation on repeat measurements of 0.3‰.
δ^{18}O values were calibrated and corrected to VSMOW using Standard Light Antarctic
Precipitation (SLAP) and VSMOW.

A3.2 Methods: ‘clumped isotope’ measurements

CO_{2} analyte was obtained from carbonate samples by reacting 10mg carbonate
samples in H_{3}PO_{4} on a custom-built automated online vacuum system described in
Passey et al. (2010). Reactions were carried out at 90°C for 20 minutes, and CO_{2} was
immediately trapped at liquid nitrogen temperatures as it evolved. Each sample was then
cryogenically purified in dry ice/ethanol and liquid nitrogen traps, and purified on a gas
chromatograph to remove hydrocarbon contaminants according to the scheme outlined in
Ghosh et al. (2006) and modified as detailed in Passey et al. (2010). Sample gases were
analyzed at the California Institute of Technology for masses 44 through 49 (inclusive)
on a Thermo-Finnigan MAT 253 gas source isotope ratio mass spectrometer in dual inlet
mode for 8 acquisitions of 10 cycles, each with an integration time of 8 seconds and a
total analysis time of approximately 3 hours (including pressure balance, background
measurement and peak-centering routines). Measurements were made with a stable 16-
volt signal at mass 44, with peak centering, background measurement and pressure-
balancing before each acquisition (Huntington et al., 2009; Passey et al., 2010).
Data for this study were collected between June 12 and 14, 2009. On every day of machine use, gases heated to 1000°C were analyzed to define the stochastic distribution of isotopologues as described in Eiler and Schauble (2004). Heating of gases to 1000 °C is necessary to drive the sample gases to the stochastic distribution. Carbonate standards run during this time period consisted of an analysis of two separate extractions of a marble, yielding an average $\Delta_{47}$ value of 0.345 ± 0.003‰, and two extractions of a vein calcite 102-GC-AZ01 ($\Delta_{47} = 0.669 \pm 0.002$‰). The long-term mean $\Delta_{47}$ values for 17 analyses of the marble standard analyzed at 90°C on the same system is 0.352‰ ± 0.019‰, (Ghosh et al., 2006; Passey et al., 2010).

The error in $\Delta_{47}$ of CO$_2$ extracted once (n=1) from a carbonate sample ranged from 0.006 to 0.013‰ (one standard error), (Table A2). This uncertainty is calculated as one standard error of the mean $\Delta_{47}$ value, determined from eight acquisitions of the same gas. The external precision for carbonate sample Ell 1a (*Lymnaeae*), from which measurements were taken from four separate extractions, was 0.006‰ (Table 2). This uncertainty is calculated as one standard error of the mean $\Delta_{47}$ value determined from multiple CO$_2$ samples extracted from separate aliquots of the same material, and analyzed for eight acquisitions each.
Table A2: Individual clumped isotope measurements

<table>
<thead>
<tr>
<th>Species*</th>
<th>$\delta^{18}$O aragonite</th>
<th>$\delta^{18}$O water</th>
<th>$\Delta_{av}$ Per mil</th>
<th>#Temp. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lymnae sp.</td>
<td>8.43</td>
<td>-22.1</td>
<td>0.695 ± 0.006</td>
<td>14.7 ± 1.3</td>
</tr>
<tr>
<td>Lymnae sp.</td>
<td>8.41</td>
<td>-23.1</td>
<td>0.716 ± 0.008</td>
<td>10.5 ± 1.5</td>
</tr>
<tr>
<td>Lymnae sp.</td>
<td>8.38</td>
<td>-22.4</td>
<td>0.701 ± 0.009</td>
<td>13.3 ± 1.8</td>
</tr>
<tr>
<td>Lymnae sp.</td>
<td>8.47</td>
<td>-23.7</td>
<td>0.730 ± 0.008</td>
<td>7.8 ± 1.5</td>
</tr>
<tr>
<td>Gyraulus albus</td>
<td>9.47</td>
<td>-22.2</td>
<td>0.719 ± 0.011</td>
<td>9.8 ± 2.2</td>
</tr>
<tr>
<td>Pisidium sp.</td>
<td>11.05</td>
<td>-21.8</td>
<td>0.744 ± 0.013</td>
<td>5.3 ± 2.3</td>
</tr>
<tr>
<td>§Average all molluscs</td>
<td>9.03 ± 1.1</td>
<td>-22.6 ± 0.7</td>
<td>0.717 ± 0.007</td>
<td>10.2 ± 1.4</td>
</tr>
</tbody>
</table>

* Each measurement is from a different mollusc.
†± Uncertainty in $\Delta_{av}$ values represents the internal precision of each measurement (one standard error).
§The ± values in the average row refer to the external precision of $\Delta_{av}$ measurements to one standard error.
#Uncertainty in $\Delta_{av}$ temperature values represents the error in temperature in °C propagated from the internal or external precision of the $\Delta_{av}$ measurement (one standard error).

Table A2: Summary of all ‘clumped isotope’ measurements. Samples were all measured on whole shells with no sub-sampled members. All errors are reported as standard error.

A3.3 Methods: Moss isotopes

Variations in the $\delta^{18}$O of aquatic mosses reflect the composition of meteoric waters in which they grow. When oxygen is fixed into cellulose, the isotopic ratio of cellulose is controlled by the isotopic value of the environmental water in the cell enriched by a biologic fractionation factor of 27-29‰ (Epstein et al., 1977; DeNiro and Epstein, 1981; Yakir and DeNiro, 1990). Thus, $\delta^{18}$O of cellulose can be considered a record of cellular water at the site of fixation. However, in terrestrial plants, leaf water $\delta^{18}$O becomes enriched by evaporation of cellular water from the leaf (Roden et al., 2000). Aquatic plants do not undergo evaporative enrichment on their own and instead form cellulose in equilibrium with surrounding water (DeNiro and Epstein, 1981; Sauer
et al., 2001; Ménot-Combes et al., 2002; Zanazzi and Mora, 2005). Mosses lack both a vascular system and stomata, thus fractionation occurs as a function of diffusion and biological fractionation during cellulose synthesis. Therefore, cellulose $\delta^{18}O$ in aquatic plants should reflect the $\delta^{18}O$ value of meteoric water, after removing the effects of fractionation factors associated with cellulose synthesis and diffusion. There is a linear relationship between source water $\delta^{18}O$ and cellulose $\delta^{18}O$ (Equation 1) (Sauer et al., 2001). The intercept of this equation is indistinguishable from the reported theoretical fractionation during cellulose synthesis of 27-29‰ (Yakir and DeNiro, 1990).

\[
\delta^{18}O_{\text{cellulose}} = 0.882\delta^{18}O_{\text{water}} + 28.3\text{‰} \quad (1)
\]

In this study, we evaluated moss isotopic values from previously published field studies and examined the overall relationship between $\delta^{18}O$ of cellulose and $\delta^{18}O$ of water (Fig A2). The relationship established by Sauer et al. (2001) (Eq 1) holds true even when moss $\delta^{18}O$ values from other studies are incorporated into the regression, with Equation 1 plotting within the confidence limits of a regression line that incorporates two additional moss isotope studies (Ménot-Combes et al., 2002; Zanazzi and Mora, 2005).
Figure A2. Illustrates the relationship of moss $\delta^{18}O_{\text{cellulose}}$ to $\delta^{18}O_{\text{water}}$ for three studies (Sauer et al., 2001; Ménot-Combes et al., 2002; Zanazzi and Mora, 2005) of moss isotopes. Grey dashed line indicates the line determined by the theoretical relationship established by Sauer et al. (2001) (Eq 1). Note that the relationship determined by Sauer et al. (2001) is indistinguishable from the relationship determined by adding additional moss isotope data. Scatter in the plot represents a standard error (SE) of 0.15 ‰.

A3.4 Methods: Mollusc Stable Isotopes

Eight mollusc shells were extracted from within the peat ensuring that the cellulose and the mollusc samples were contemporaneous. Four of the mollusc shells were of the species *Gyraulus albus*, two were *Lymnaeae* sp., and two were small *Pisidium* sp. clams. Three of the mollusc shells were sampled along growth bands to attempt to resolve a seasonal signal. Temperature calculations were carried out using the Grossman and Ku (1986) equation modified for use in freshwater molluscs (Dettman et al., 1999) (Equation 2) Where $T$ is degrees in Kelvin and $\alpha$ is the fractionation between water and aragonite described by Equation 3, such that,

$$1000 \ln (\alpha) = 2.559 \left(10^6 T^{-2}\right) + 0.715 \quad (2)$$
\[ \alpha_{\text{aragonite-water}} = \frac{1000 + \delta^{18}O_{\text{aragonite (VSMOW)}}}{1000 + \delta^{18}O_{\text{water (VSMOW)}}}. \] (3)

\( \delta^{18}O_{\text{aragonite}} \) is the isotopic ratio in VPDB measured from the gastropod shells converted to VSMOW using the \( \alpha_{\text{SMOW-PDB}} \) of 1.03091 (Gonfiantini et al., 1995). \( \delta^{18}O_{\text{water}} \) is the isotopic ratio derived from the \( \delta^{18}O_{\text{cellulose}} \) values of aquatic moss and calculated using (Eq. 1). The uncertainty in the source water estimates is combined with the uncertainty in equation 2 in order to determine the uncertainty of the calculated temperature. White et al. (1999) conducted a combined laboratory and field study on the temperature relationship of biogenic aragonite from the freshwater gastropod lymnae. Based on their results they posited that there is no vital effect that is specific either to this freshwater, or to other, mollusc genera.

A.4 RESULTS & DISCUSSION

A.4.1 Inferred \( \delta^{18}O \) of meteoric water

The \( \delta^{18}O \) values of cellulose obtained from the Pliocene moss averaged 10.1 ± 0.24‰ VSMOW (n=8), implying a water \( \delta^{18}O \) value of -20.7 ± 0.3‰ (Eq. 1). Water \( \delta^{18}O \) values calculated using clumped isotope-derived temperatures combined with mollusc \( \delta^{18}O \) support a slightly lower (though similar) value of -22.6 ± 0.7‰ VSMOW (n=7). Water \( \delta^{18}O \) values determined from moss cellulose likely represent values for the moss
growing season, and are thus possibly biased more towards June and July than the isotopic values determined from ‘clumped isotopes’ that would reflect water δ¹⁸O values for the period of mollusc calcification. Moss-derived water δ¹⁸O values contain more of a summer signal and are ~ 3-5‰ enriched compared to modern Arctic summer precipitation on Ellesmere Island measured in July 2004 (δ¹⁸O = -25.4‰ ± 0.5‰ VSMOW; n=9). Estimates of Pliocene water δ¹⁸O are higher than values reported by GNIP (the global network for isotopes in precipitation) for summer precipitation on Ellesmere Island (-24 to -28‰) (IAEA/WMO, 2006), and much higher than annually averaged precipitation for Ellesmere Island (approx. -30.8 ± 6.7‰) (IAEA/WMO, 2006). Both proxies yield Pliocene values that are similar to modern annual GNIP water values from Alaska and the Yukon (-20.8 ± 2.3 ‰; Barrow, Mayo and Whitehorse stations) (IAEA/WMO, 2006). Values of meteoric water calculated in this study are similar to values of meteoric water calculated for the Paleocene and Eocene at the same location (Tripati et al., 2001; Richter et al., 2008; Jahren et al., 2009). Meteoric water values for the Eocene derived from mollusc shells are enriched relative to modern and have δ¹⁸O values that range from -16 to -23 ‰ (Tripati et al., 2001; Jahren et al., 2009; Eberle et al., 2010). Richter et al. (2008) estimated modern values of annual precipitation isotopes for Axel Heiberg Island not much different from their estimated Eocene values of – 21.9‰ using the Bowen and Wilkinson (2002) online precipitation isotope calculator. However, these inferred values are high when compared to both the IAEA and our measurements of modern precipitation. Based on this they argued that the gradient of values of δ¹⁸O of meteoric water remained unchanged over much of the Cenozoic
(Richter et al., 2008), however, this is unlikely given the results presented here which show a 3-5‰ enrichment in $\delta^{18}O$ of Pliocene Arctic water. Jahren et al. (2009) also demonstrated a difference in $\delta D$ of Eocene meteoric water and modern of $\sim 40$‰ and suggests that the latitudinal gradient in precipitation isotopes was likely different. Recent modeling studies of Eocene isotopes in precipitation have indicated that when the latitudinal temperature gradient is reduced there is a corresponding change in the latitudinal gradient of isotopes in precipitation (Speelman et al., 2010). Thus the difference in isotopic ratios between the present and the Pliocene instead likely reflects either local, regional or global changes in the hydrologic cycle, which in turn could record differences in the temperature of precipitation, in Rayleigh fractionation associated with distillative condensation during vapour transport, and possibly in the season of precipitation (i.e., the ratio of summer to winter precipitation). Furthermore if we consider that a reduction in ice volume during the Pliocene caused $\delta^{18}O$ values of meteoric water to be globally depleted by $\sim 1$‰ (Lear et al., 2000) and that the latitudinal temperature gradient was reduced (Ballantyne et al., 2010), then our estimates of a 3-5‰ enrichment in $\delta^{18}O$ of Arctic water suggests that the latitudinal gradient in $\delta^{18}O$ of meteoric water was also diminished in the Pliocene. Studies of other Pliocene sites from a range of latitudes and the use of isotope-enabled models would provide further insight into these changes.
A.4.2 Temperature estimates from independent proxies

Temperatures calculated based on the difference in $\delta^{18}$O of meteoric water, inferred from moss cellulose, and co-existing mollusc carbonate average 14.2 ± 1.3° C (Fig. A3). Temperatures calculated based on ‘clumped isotope’ thermometry average 10.2 ± 1.4° C (Fig. A3). We infer these temperatures to equal mean mollusc growing season temperatures. Figure A3 presents results obtained from all gastropod shells measured in this study. Sub-sampled shells were sampled at intervals along their growth transect and thus have a spread of values representing seasonal variations (either in $\delta^{18}$O of meteoric water or temperature). Bulk values represent whole shells that were powdered and measured. Temperatures were calculated from $\delta^{18}$O values of both sub-sampled and bulk mollusc samples using the Grossman and Ku (1986) equation and the $\delta^{18}$O value of meteoric water, inferred from moss cellulose. Temperatures are independently derived using the stochastic ‘clumped isotope’ technique and indicated as $\Delta_{47}$ (Fig. A3).
Figure A3. Paleotemperature calculations from Pliocene molluscs, *Pisidium sp.*, *Gyraulus albus* and *Lymnae sp.* $\delta^{18}$O (sub-sampled) represents multiple measurements from the same individuals with the variability in values being the result of sampling along growth whorls. $\delta^{18}$O (bulk) represents bulk aragonite values from single shells. $\Delta_{47}$ (‘clumped isotope’) represents ‘clumped isotope’ thermometry of mollusc shells. The average temperature of 12.7° is plotted along with 1σ error bars.

To assess the differences between methods we conducted a t-test grouping the data based on the method used and found that there is no significant differences between the stable isotope and ‘clumped isotope’ results (t=2.02; p = 0.07). We also conducted a one-way ANOVA, grouping the sub-sampled stable isotope results as one group, the bulk results as another and the ‘clumped isotope’ results as a third group, again finding no significant difference (F=1.26; p=0.30). If we run the ANOVA analysis using each individual mollusc as a group then we do find significant differences, however, the difference between the ‘clumped isotope’ individuals and the stable isotope individuals is no greater than the difference between two individuals in the sub-sampled isotope group.
There is no difference between the two techniques for the species *Gyraulus albus*, however, there are differences between the results obtained from both *Lymnaceae sp.* and *Pisidium*. Although both *Gyraulus albus* and *Lymnaceae sp.* grow throughout the year, both species experience much slower growth during the winter months and rapid growth between April and August (Dussart, 1979; White et al., 1999). However, these studies used species collected in the British Isles. Pip (1992) noted that in Manitoba, Canada both species of mollusc generally hibernate for the entire period of ice cover and exhibit most growth during the months of May to September, which is consistent with a study by Caquet (1993) showing that *Lymnaea* exhibits almost no growth between September and May. Additionally as this would have been high enough latitude to experience complete darkness during the winter months, it is unlikely that either species grew during the winter. Both species generally only live one to two years (Dussart, 1979; Pip 1992). Pip (1992) reports that these molluscs are born between June and September and then overwinter as sub-adults before growing to full size. That there is no difference between temperatures calculated from $\delta^{18}O$ and temperatures determined from $\Delta_{47}$ from *Gyraulus albus* could suggest that *Gyraulus albus* grows primarily during the same period as the moss whereas the greater difference between $\delta^{18}O$ and $\Delta_{47}$ temperature results determined from *Lymnaceae* and *Pisidium* could suggest that these two species exhibit a longer growing season than the moss (Fig A3). Thus the differences between the two techniques could simply be an artifact of the fact that the moss $\delta^{18}O$ value represents average water $\delta^{18}O$ values of the moss growing season (likely summer) and not the average $\delta^{18}O$ values of the mollusc growing season. Although Zanazzi and Mora (2005)
indicate that water in fen environments has little seasonal variation in isotopic values. Turner et al. (2010) suggest that there is a tendency for more depleted values during the 2 months of snow melt in the spring and if some species of molluscs start growing early in the season their $\delta^{18}$O values may biased towards more depleted winter precipitation. To test this we can use the meteoric water $\delta^{18}$O value determined from the ‘clumped isotope’ results ($-22.6 \pm 0.7\%e$), which should represent the average water value of the period of mollusc calcification, instead of the moss-inferred water $\delta^{18}$O value to calculate temperatures from the Lymnaceae and Pisidium. If we do this, the average Lymnaceae temperature of 11.8 ± 0.2° C is much closer to the clumped isotope value of 9.8 ± 2.2° C suggesting that it may be that moss and Lymnaceae growth periods do not overlap.

Three mollusc specimens were sampled at up to eight points along the growth whorl of the gastropod in two of these three mollusc samples with calculated temperatures ranging between 12.7 and 21.7° C (7.5-16.4° C if the $\delta^{18}$O water value inferred from ‘clumped isotopes’ is used) suggesting spring-summer temperatures are predominantly recorded (Fig A4). The third sub-sampled mollusc has temperatures ranging between 5.4 and 10.8° C (-2.9-2.8° C if the $\delta^{18}$O water value inferred from ‘clumped isotopes’ is used). Because all molluscs were sampled from near the centre of the shell this could represent an individual that was born late in the year and grew mostly in the fall and spring as Gyraulus albus can have two generations, one born in the spring and another born in the fall (Dussart, 1979). ‘Clumped isotope’ results and ‘bulk’ sample results likely do contain more of a winter signal than the sub-sampled gastropods because there would be some influence from the minor amount of carbonate formed during the
winter on the temperatures calculated using calcite from the entire shell. These values are 14.3 ± 4.5° C (average of all results) warmer than the modern (May-Sep) temperatures for Eureka, Ellesmere Island, of −1.6 ± 1.3° C (Environment Canada, 2008). May to September temperatures were used because these months represent a standard growing season and cover the period when molluscs have their highest growth rates (Dussart, 1979; Pip, 1992).

Figure A4: Paleotemperature calculations from three Pliocene molluscs that were sampled at intervals along a growth whorl indicating seasonal variability in single individuals. Temperatures were calculated using equation 2 and the meteoric water value determined from moss stable isotopes with grey shading indicating a range of temperatures inherent in the standard deviation of equation 1.

Without detailed information on the growing season of Pliocene molluscs or information on the seasonal variation of water δ¹⁸O, we are unable to use these data to determine seasonal variation in temperature. If low variability in water δ¹⁸O during the period of calcification is assumed, then the intra-shell variability in δ¹⁸O combined with
our inferred water $\delta^{18}$O values (-20.7 ‰) can be used in order to estimate maximum and minimum temperatures during the season of calcification. Maximum and minimum growing season temperatures of 21.7° C and 5.4° C are calculated.

Our estimates of mean calcification temperature derived from both techniques of 10.2 and 14.2° C (Fig A3) are generally similar to warm-month mean temperatures (WMMT) based on transfer functions derived from fossil beetle assemblages, also collected from the Beaver Pond locality (12.4° C) (Elias and Matthews, 2002) (Table A3). However, the temperatures determined in our study are assumed to represent an average of the entire calcification period, hereafter referred to as growing season (and likely not solely reflecting summer). These data imply that summer temperatures may actually have been warmer than what has previously been reported (Elias and Matthews, 2002). $\delta^{18}$O values of cellulose from wood collected at the Beaver Pond site on Ellesmere Island (Ballantyne et al., 2006) was used to reconstruct mean annual temperature (MAT). The tree-ring isotope-based temperature obtained from this earlier study was -5.5 ± 1.9° C, or 14.2 ± 2.5° C warmer than present (Ballantyne et al., 2006). This difference (Pliocene minus modern) is equivalent to the temperature difference calculated in our study using both proxies (14.3 ± 4.5° C) and that of the beetle study of Elias and Matthews (2002) (10 ± 2° C). These temperatures are also equivalent to WMMT calculated by Ballantyne et al. (2010).

In order to more closely compare to both the Elias and Matthews (2002) study, which is based on estimates of WMMT we can look at $T_{\text{max}}$ (maximum calculated temperature) determined both from the stable isotopes of freshwater molluscs and moss
(21.7° C) and ‘clumped isotopes’ (14.7° C) as analogous to temperatures recorded during calcite formation during the warmest months, or WMMT. A difference from modern ΔWMMT of 14.1 ± 0.7° C and 7.1 ± 1.3° C can be inferred indicating that our estimates of WMMT are within the range of values determined by Elias and Matthews (2002) and Ballantyne et al. (2010). Elias and Matthews (2002) suggested that greater warming is evident during the winter months, meaning that differences from modern (ΔT) calculated for either growing season or mean annual temperatures are likely to appear greater than estimates of ΔWMMT. Ballantyne et al. (2006) confirmed this when they determined that their tree-ring isotope based reconstruction captured more of a winter signal than the beetle records of Elias and Matthews (2002) resulting in a higher estimate of ΔT.

Ballantyne et al. (2010) conducted a proxy comparison study of temperature estimates for the Beaver Pond locality updating and expanding upon the Ballantyne et al., (2006) study. Using three independent approaches, a study of the coexisting vegetation at the site and its relationship to climate, and stable isotope values of tree-ring cellulose and bacterial tetraethers, Ballantyne et al. (2010) inferred an MAT of −0.6 to −0.4° C, or ~19° C warmer than present. These inferred temperatures from three independent proxies are considerably warmer than the temperatures calculated by previous studies and in our own study (Table A3), however, as our study is representative of the period of calcification and is biased towards the period of maximum mollusc growth, we are not able to accurately compare our results to MAT. This large difference between the seasonally biased studies presented here and in Elias and Matthews (2002) and the mean annual temperature studies presented in Ballantyne et al. (2010) suggests that there was perhaps
much greater warming, relative to modern, in Pliocene winter temperatures at high latitudes than in temperatures during the warmer months. The cause of greater winter warming could be related to a relatively ice-free Arctic Ocean during the Pliocene resulting in reduced albedo and allowing for maritime modulation of winter temperatures or to increased cloud cover during the winter months. Seasonal ice extent in the Arctic Ocean during the Pliocene is largely unknown (Zachos et al., 2008) and although a recent study by Abbot and Tziperman (2008) indicated that deep convective clouds can produce significant winter warming under ice-free conditions, the climatic effects of both of these phenomena on Pliocene winter temperatures must be further investigated.

Table A3: Comparison of results obtained in this study to results from previous non-thermodynamic temperature estimates for the Pliocene Beaver Pond locality. Note that our estimates confirm estimates from prior work. All errors are reported as standard error.

<table>
<thead>
<tr>
<th>Pliocene growing season temperature</th>
<th>Difference from modern</th>
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<tr>
<td>T (°C)</td>
<td>ΔT (°C)</td>
</tr>
<tr>
<td>Freshwater mollusc shell δ¹⁸O</td>
<td>14.2 ± 1.3</td>
</tr>
<tr>
<td>Freshwater mollusc shell ΔT (‘clumped isotopes’)</td>
<td>10.2 ± 1.4</td>
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<tr>
<td>Average this study</td>
<td>12.7 ± 1.9</td>
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<table>
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<th>Pliocene warm month mean temperature</th>
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<tr>
<td>Elias and Matthews (2002)</td>
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<td>Ballantyne et al. (2010)</td>
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<tr>
<th>Pliocene MAT</th>
</tr>
</thead>
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<tr>
<td>Ballantyne et al. (2006)</td>
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<tr>
<td>Ballantyne et al. (2010)</td>
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The results presented are averages of all measurements for each technique.

*Temperatures calculated from freshwater mollusc δ¹⁸O measurements uses the moss cellulose inferred water δ¹⁸O values for temperature determination using equation 2.
A.5 SUMMARY & CONCLUSIONS

The data we present provide constraints on the magnitude of Arctic Pliocene warmth, and on changes in Arctic hydrology associated with a warmer climate. Our temperature estimates (Table A3) indicate that the Pliocene climate of Ellesmere Island was similar to conditions experienced by modern boreal forests located ~15-20° further south. This is consistent with our surface water estimates (-20.7 to -22.6‰) that are also similar to δ¹⁸O of water values from latitudes ~15-20° south of Strathcona Fiord. With mollusc growing season temperatures (inferred as May-Sept) on the order of 10.2-14.2° C at paleolatitudes of 78° N, the Arctic Ocean was likely seasonally ice free (Haywood et al. 2009). Much greater estimates of ΔT are determined using proxies of MAT than proxies biased towards warmer months. This indicates that the difference between Pliocene and modern winter temperatures in the Arctic was greater than the difference between Pliocene and modern temperatures from warmer months. The Pliocene is a possible example of the temperatures that we may experience in the next century. Models of Pliocene climate are being constructed with an aim to improve model predictions of future climate by assessing the sensitivity of climate to forcings, such as, increased atmospheric CO₂ (Jansen et al., 2007; Haywood et al. 2009). Our study provides the first thermodynamically-based estimates of terrestrial Arctic surface-temperature and water δ¹⁸O for the Pliocene, and the temperatures we estimate for Ellesmere Island provide some constraints on the magnitude of warming that might eventually occur in the Arctic if atmospheric CO₂ levels were stabilized at present levels.
A.6 ACKNOWLEDGMENTS

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A.7 AUTHOR CONTRIBUTIONS

A.Z.C, A.T. and W.P.P conceived the study. N.R. and A.P.B. collected and provided samples and field notes. A.Z.C. conducted the stable isotope analyses on cellulose and on microsampled molluscs. A.T. and R.A.E. conducted the clumped isotope analyses, and these analyses were performed in the laboratory of J.E. W.P.P supervised the bulk stable isotope analyses. A.Z.C wrote the manuscript with contributions from A.T. and R.A.E. All authors commented on the manuscript throughout its development.
A.8 REFERENCES


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APPENDIX B

CLIMATE VARIABILITY IN THE EARLY PLIOCENE ARCTIC: ANNUALLY RESOLVED EVIDENCE FROM STABLE ISOTOPE VALUES OF SUB-FOSSIL WOOD, ELLESMERE ISLAND, CANADA

Manuscript in review at Palaeogeography, Palaeoclimatology, Palaeoecology
Climate variability in the Early Pliocene Arctic: Annually resolved evidence from stable isotope values of sub-fossil wood, Ellesmere Island, Canada

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B.1 ABSTRACT

Stable isotope climate proxies (δ¹⁸O, δD and δ¹³C values) enhance traditional ring-width data, although poor preservation of ancient wood has tended to limit development of stable isotope proxy records to the Holocene and the Late Pleistocene. Here we apply stable isotope techniques to wood that represents the remains of Mixed-Coniferous Boreal Vegetation preserved in Early Pliocene (4-5 Ma) deposits at Strathcona Fiord, Ellesmere Island, Canada (ca. 78°N). Four well-preserved tree trunks, identified through wood anatomical characteristics as Larix (larch), from this high Arctic site provide annually resolved sequences of up to 250 years from which we developed a
high-resolution, continuous secular isotope record of Pliocene climate. Stable oxygen isotope values, in conjunction with ring-width measurements were used to derive annually resolved temperature records for this site. Our ring-width and isotope-based reconstructions provide an annually resolved record, up to 250 years, of temperature and indicate average growing season (JJ) temperatures (15.8 ± 5.0 °C) 11.8 ± 5.1 °C, and mean annual temperatures (MAT) (−1.4 ± 4.0 °C) 18.3 ± 4.1 °C warmer than present. Estimated isotope values of precipitation of −16.3 ± 2 ‰ (δ\(^{18}\)O) and −150.1 ± 8.9 ‰ (δD) were calculated from the isotopic values of wood cellulose. Relative humidity estimated from both δ\(^{13}\)C and δD records ranged from 60-80%. Spectral analysis of isotopic and ring-width values exhibit periodicities around 20, 10-12, 5.5, 4-5, 3-3.5 and 2.3-2.8 years, indicating that decadal and sub-decadal modes of variability similar in period to modern modes such as the NAO and the AO existed in the early Pliocene Arctic.

B.2 INTRODUCTION

In recent decades, interest in the response of high-latitudes to global climate change has renewed interest in the study of ancient ecosystems under greenhouse conditions. Well-preserved wood from Ellesmere Island affords the opportunity to conduct the first high-resolution study of Early Pliocene Arctic temperature variability. This is particularly important as the Early Pliocene is considered an important example of future climate conditions that could result should current warming trends continue (Haywood and Valdes, 2004; Jansen et al., 2007). Early Pliocene atmospheric CO\(_2\) levels
are estimated at 365-415 ppm, similar to present day levels (Tripati et al., 2009; Pagani et al., 2010), however temperatures are estimated to have been 2-3 °C warmer and Arctic temperatures 7-15 °C warmer than pre-industrial temperatures (Haywood et al., 2009; Ballantyne et al., 2010). With Pliocene temperatures considerably warmer and levels of atmospheric CO$_2$ similar to present, detailed constraints on polar temperatures are necessary to constrain the extent to which warming was amplified at high latitudes and thus provide important information on the sensitivity of high latitudes to changing levels of CO$_2$. Proxy records of temperature can be used to constrain model predictions of climate when different boundary conditions are changed allowing an assessment of which conditions exert the most control on Pliocene climate (Lunt et al. 2010).

A large number of fossil forest sites dating to the Pliocene exist throughout the Canadian Arctic (Matthews and Ovenden, 1990). Here we report year-to-year temperature variations in *Larix* (larch) trees recovered from discontinuous peat layers within fluvial sands near Strathcona Fiord, Ellesmere Island, Canadian Arctic Archipelago, 78° 29.271’N 82° 37.973’W (Fig B1). The Pliocene fossil deposits from Strathcona Fiord have been well studied. The Beaver Pond locality, named for the abundant beaver-gnawed sticks found at the site (Rybczynski, 2008), is located at the head of Strathcona Fiord. Flora includes macrofossils of bryophytes and vascular plants including many extant species of moss, spruce, pine, cedar, alder and birch (Matthews and Ovenden, 1990; Ballantyne et al., 2010). Several fossil forest deposits exist in the Strathcona Fiord area and all are inferred to date from the same period. Dating of these deposits is based on the rich mammalian fauna from the Beaver Pond locality and places
the Strathcona forests in the Early Pliocene (4-5 Ma) (Tedford and Harington, 2003; Ballantyne et al. 2006). Although work is ongoing to further constrain the age of these deposits. Fossil wood from this site has been identified through wood anatomy as belonging primarily to the genus *Larix* (Ballantyne et al., 2006; this study). Trees at this site show the narrow rings, stunted development, and curtailed vertical growth characteristic of boreal vegetation.

Figure B1. Map showing the location of Strathcona Fiord in central Ellesmere Island.

The well-preserved wood found at the site makes it possible to use dendrochronologic techniques to obtain annually resolved climate information from the site. Dendroclimatology has proven useful in extending our knowledge of climate to periods prior to instrumental records. Use of stable isotopic techniques in combination with traditional ring-width data, has expanded this utility to periods prior to the late
Holocene (Jahren and Sternberg, 2003; Ballantyne et al., 2006; Panyushkina et al., 2008; Richter et al., 2008a). Previous studies have had success in using stable isotopes and ring widths to estimate temperatures for the Beaver Pond locality (Ballantyne et al., 2006; 2010), however, both studies only used ca. 20 rings from a single tree for their reconstruction. Here we expand on these studies by presenting annually resolved reconstructions of temperature from four trees, with a minimum of 100 years of data from each tree, providing information on between-tree variability and interannual variability in climate during the Early Pliocene.

B.3 METHODS

Fossil wood was sampled by first slicing a transverse disk using a Ryoba 330 fine toothed saw. The wood was sliced into rectangular blocks along one radius from each disk and ring widths were measured to a precision of 0.01mm. Following ring-width measurement individual rings were sampled using a scalpel to cut thin shavings of ca. 50 µm in width. Four specimens were used in this study. Specimen US357 was collected in 1992 at 78° 29’ N, 82° 37’ W and is archived in the Department of Geological Sciences at the University of Saskatchewan. Specimens NUPB 99 and NUPB 100 were collected in 2004 at 78° 26’ N, 82° 38’ W and are archived in the Department of Geological Sciences at the University of Saskatchewan. Specimen NUPB 341 was collected in 2006 by the Canadian Museum of Nature at 78° 33’ N 82° 25’ W and is archived at the Canadian
Museum of Nature. Oxygen isotope values were measured on all four specimens, whereas carbon and hydrogen isotopic values were determined only on specimen US357.

α-cellulose can provide proxy climate information from three stable isotope systems and is the easiest to isolate from other wood components that differ in isotope values (Wilson and Grinsted, 1977). The method outlined by Loader et al. (1997) was used to isolate α-cellulose from bulk wood for $\delta^{18}$O and $\delta^{13}$C analyses. Because of the exchangeability of cellulose hydrogen with other sources of hydrogen, such as atmospheric moisture or chemicals during processing, it is necessary to eliminate exchangeable components via cellulose nitration with fuming nitric acid by the acetic anhydride method prior to $\delta$D analysis (Sternberg, 1989). Specimen US357 was analyzed for $\delta^{13}$C, $\delta^{18}$O and $\delta$D values at the Saskatchewan Isotope Laboratory (SIL), University of Saskatchewan. $\delta$D and $\delta^{18}$O analyses were conducted in the SIL using a Thermo Finnigan TC/EA coupled via a Conflo III interface to a Thermo Finnigan Delta Plus XL mass spectrometer in continuous flow mode. $\delta^{13}$C values were determined using a Thermo Finnigan Flash EA via oxidation directly coupled to a Thermo Finnigan Delta Plus XL mass spectrometer interfaced by a Conflo III. Analytical errors based on repeated analysis of lab standards were, 0.2 ‰ for oxygen, 0.1 ‰ for carbon and 1.5 ‰ for hydrogen. Duplicate samples were run for each ring with a precision of 0.5 ‰ for oxygen, 0.3 ‰ for carbon and 1.5 ‰ for hydrogen. $\delta^{18}$O analyses of specimens NUPB 99, NUPB 100 and NUPB 341 were conducted at the University of Arizona and utilized a Costech HTG EA, modified for oxygen isotope analysis of cellulose (Evans, 2008), directly coupled via a ConFlo III to a Finnigan DeltaPlus XP mass-spectrometer.
Cellulose (0.30-0.35 mg) was weighed into silver capsules, pyrolyzed over glassy carbon and reported as values relative to VSMOW. Analytical precision for repeat analysis of $\delta^{18}O$ ratios of an internal sigma cellulose standard is typically 0.3‰.

**B.4 INTERPRETATION OF STABLE ISOTOPE VALUES**

**B.4.1 Mechanistic models**

In an effort to understand the relationship between tree-ring isotopes and their environment, Roden et al. (2000) developed mechanistic models containing factors (i.e. humidity, temperature, and physiology) that influence isotope ratios. Trees were grown under controlled laboratory conditions and were subsequently analyzed for $\delta^{18}O$ values in cellulose and $\delta D$ values in cellulose nitrate. Their premise was that oxygen and hydrogen isotopes incorporated into cellulose from leaf water were subject to fractionation at several points during the transition from water to cellulose, which is modeled in equations 1 and 2, where $f_o$ and $f_H$ are the fractions of carbon-bound oxygen and hydrogen, respectively; $\delta^{18}O_w$ and $\delta D_w$ are oxygen and hydrogen isotope ratios of xylem water; $\epsilon_o$ is the biological fractionation factor for oxygen during conversion from sugar to cellulose; $\epsilon_{HH}$ is the biological fractionation factor for hydrogen during conversion from sugar to cellulose in a heterotrophic reaction; whereas $\epsilon_{HA}$ is the biological fractionation factor for hydrogen in sugar converting to cellulose in an autotrophic reaction; and $\delta D_w$ and $\delta^{18}O_w$ are the oxygen and hydrogen isotope ratios of the leaf water calculated using the equation developed by Flanagan and Ehleringer (1991).

$$\delta^{18}O_{\text{cellulose}} = f_o * (\delta^{18}O_w + \epsilon_o) + (1 - f_o) * (\delta^{18}O_w + \epsilon_o) \quad (1)$$
In a field test of riparian vegetation, their model was able to explain almost all of the variance in both $\delta D$ and $\delta^{18}O$ values (Roden and Ehleringer, 2000). Subsequently Waterhouse et al. (2002) conducted a test of the Roden et al. (2000) model using oak ($Quercus$) tree-ring isotopes in central England and discovered that although the model explained all of the variance in $\delta^{18}O$ values it was not able to explain the variance in $\delta D$ values. The Roden et al. (2000) model relies on knowledge of modern plant physiology, leaf water and xylem water isotope ratios rendering it under-constrained for most paleoclimate reconstructions. Additionally temperature is not a direct term in the Roden et al. (2000) model, thus a secondary calculation involving the water isotope temperature relationship must be utilized (Dansgaard, 1964; Rozanski, 1993).

A model devised by Anderson et al. (2002) simplifies the Roden et al. (2000) model by incorporating the terms of the leaf water model, $\delta^{18}O_{wl}$ in equation 1, into the overall model (equation 3).

$$\delta^{18}O_{sw} \approx \delta^{18}O_{cellulose} - (1 - f) (1 - h) (\epsilon_e + \epsilon_k) - \epsilon_{biochem}$$

In this equation $\delta^{18}O_{sw}$ is the oxygen isotope value of source water; $f$ is the fraction of leaf water not subject to evaporation, accounting for the change in the isotope value of photosynthate caused by stem water interaction; $\epsilon_e$ is the liquid-vapor equilibrium fractionation factor (Manjoube, 1971); $\epsilon_k$ is the liquid-vapor kinetic fractionation, dependent on airflow dynamics at the leaf boundary layer — for conifers it has been inferred to be 28‰; $h$ is relative humidity and $\epsilon_{biochem}$ is the biologic fractionation factor.
for sugar converting to cellulose calculated as $27 \pm 3\%$. Anderson et al. (2002) also found that a variable $f$ value, dependent on humidity, was much more valuable in the model than a fixed value for $f$. Use of the Anderson et al. (2002) model requires knowledge of humidity, temperature (to calculate $\epsilon$), and/or source water value, thus use of mechanistic modeling equations as a method of calculating temperature remains somewhat problematic (Waterhouse et al., 2002; Anderson et al., 2002). In this study we have primarily used the mechanistic modeling equations to reconstruct the isotopic values of xylem water.

B.4.2 Transfer functions

Previous studies have noted a correspondence between $\delta^{18}O$ and $\delta^{13}C$ values of tree rings and instrumental meteorological records of temperature (Epstein and Yapp, 1976; Gray and Thompson, 1976; Wilson and Grinsted, 1977; Burk and Stuiver, 1981; Anderson et al., 1998; Ballantyne et al. 2006; Richter et al., 2008b; Saurer et al., 2008), precipitation (Leavitt and Long, 1989; Ehleringer et al., 1993; Buhay and Edwards, 1995; Robertson et al., 2001), humidity (Lipp et al., 1991; Edwards et al., 2000; Wright and Leavitt, 2006) and soil moisture (Buhay et al., 1996; McCarroll and Loader, 2004).

Ballantyne et al. (2006) developed a transfer function for reconstructing temperatures from $\delta^{18}O$ values of fossil wood in combination with ring widths. However, their transfer function was based upon only two trees, one from the Yukon and one from Ottawa, and a series of ten rings from each tree (Ballantyne et al., 2006).
Richter et al. (2008b) developed a transfer function by establishing a continental scale isotope-climate relationship with tree-ring δ¹⁸O values from around North America. However, we posit that it is unreasonable to expect the isotope climate relationship to be the same on a continental scale, and instead it is better to develop transfer functions from analogous modern sites to more closely approximate the conditions experienced by the fossil forest. This study also utilized bulk wood extractions rather than individual rings, which makes this transfer function good for a broad brush approach but limited for interannual reconstructions.

In order to improve upon these transfer functions we have analyzed 45 rings from five *Larix* trees from a site near the settlement of Chokurdak in Yakutia northern Siberia (Hughes et al. 1999). We chose Siberia because it is a modern high-latitude boreal mixed-coniferous forest dominated by *Larix*. Modern tree-ring samples were obtained from the archival collection at the Laboratory of Tree Ring Research, University of Arizona and consisted of *Larix cajenderi* (Mayr.). These samples were part of a collection used in the Hughes et al. (1999) study. Material used for our study consisted of five cross-sections crossdated as part of the Hughes et al. (1999) study. Ring width was measured from the years 1950 to 1994 with a range from 0.20 to 0.36 mm. There were no missing rings in the samples we selected and all but one of the specimens we used were part of the chronology developed in Hughes et al. (1999) and the ring-width indices (1950-1994) of the five specimens used in our study correlated well with the Hughes et al. (1999) chronology ($R^2 = 0.68$). Oxygen isotopes were measured for the period 1950 to 1994 of five specimens and showed a correlation between trees of $R^2 = 0.49$ and an expressed
population signal (EPS) of 0.81. The δ¹⁸O series of all five specimens were averaged to
form a site δ¹⁸O chronology. Our calculations of regression between the site δ¹⁸O
chronology, ring-width chronology and the mean surface temperature at Chokudak used
the period 1950 to 1972 with the period 1973 to 1994 reserved for calibration.

Using both the δ¹⁸O and ring-width chronologies we investigated the climate
relationships between ring-width, δ¹⁸O and climate parameters like temperature and
precipitation. We obtained strong correlations between the composite ring-width
chronology of our five samples and June-July (JJ) temperature for the period 1950-1994
(R² = 0.60), consistent with the results of Hughes et al. (1999). For δ¹⁸O values we
obtained the strongest correlation with mean annual temperatures (MAT) (R² = 0.27),
however there was also a weak correlation with JJ temperature and with the previous
year’s winter temperature, particularly February-March consistent with results observed
by Sidorova et al. (2009). Although ring-width indices had the strongest correlation with
climate parameters at the site, the transfer function based exclusively on them is, by
design, of local to regional applicability. Thus, in order to make use of the strong climate
signal identified in the ring-width indices we have developed a transfer function that
looks at the temperature variability by regressing the June-July temperature anomalies
against the ring-width index. The R² for the regression against the calibration period
(1950-1972) was 0.64 (MS = 25.8, F = 32.8, p = <0.001). The R² for the verification
period (1973-1994) was 0.50 (RE = 0.26; CE = 0.22). This transfer function is presented
in Equation 4;
\[ T_{JJ_{\text{anom}}} (\degree C) = -2.9344 + 2.2333 \times RW \]  

(4)

where \( T_{JJ_{\text{anom}}} \) is the temperature anomaly in \( \degree C \), RW is the ring-width index. The ring-width index was developed by first standardizing the ring-width measurements of each of our five samples by fitting a negative exponential or straight line to the ring-widths and subtracting the trend (Cook and Kariukstis, 1990) then averaging the standardized indices to form a composite index of all five samples. The error derived from the fit of the regression model is \( \pm 0.5 \degree C \).

Figure B2 presents reconstructed growing season (June-July, JJ) temperature anomalies of Siberian larch using our ring-width transfer function Equation 4.

![Figure B2](image)

Figure B2: Reconstruction of June-July temperature anomalies for Chokurdak using indexed ring-width measurements from 5 cross-dated specimens of Siberian Larch for the period 1950-1994. Calibration period is 1950-1972 and verification period is 1973-1994. Dashed line (\( J_{\text{anom}} \)) are the temperature anomalies from the instrumental station at Chokurdak determined by subtracting the mean temperature from each year. Solid line (Recon \( J_{\text{anom}} \)) are the reconstructed temperature anomalies calculated using Equation 4. Gray band is the 1\( \sigma \) standard error of the regression.
The ring-width index based transfer function (Eq. 4) provides a means of estimating the interannual variability in temperature when using tree-ring material from the region where the samples were collected under late Holocene conditions. Because the ring-width δ\textsuperscript{18}O records obtained from Chokurdak are also from a single location we chose to use an approach similar to the continental scale reconstruction of Richter et al. (2008a) and obtained other tree ring derived δ\textsuperscript{18}O records from boreal sites representing a range of different temperatures, growing season lengths and precipitation records. Localities from which we obtained data include; western Siberia (68.1° N, 60.0° E) (Holtzkamper et al., 2009), central Siberia (54.24° N, 89.57° E) (Knorre et al. 2010), northern Finland (68.56° N, 28.19° E) (Hilasvouri et al. 2009), Northern Norway (68.50° N, 15.39° E) (G. Young, pers. comm.) and the Mackenzie Delta (68.24° N, 133.48° W) (Porter et al., 2009). We also included a second isotope study from Yakutia (70° N, 148° E) involving different trees from the same site (Sidorova et al. 2008). We used the period 1950-2000 from these studies.

Using the δ\textsuperscript{18}O values of cellulose and average JJ temperature for 25 years of the 50 years we had available for each series we regressed JJ temperature against the δ\textsuperscript{18}O values of cellulose. R\textsuperscript{2} for the calibration data was 0.60 (MS = 715.6; F = 162.8; p = <0.001). R\textsuperscript{2} for the verification data was 0.54 (RE = -0.01; CE = -0.06). That the reduction of error (RE) coefficient of efficiency (CE) are negative suggests that the values reconstructed are not significantly different from the mean of the calibration period. This indicates that although the δ\textsuperscript{18}O may be capturing the mean JJ temperature it
is not providing any information on interannual variability. We are, however, using a small sample of years for both our calibration and verification periods (25 years) thus it is possible that this statistic could improve with a longer data set. The transfer function is presented in Equation 5;

\[
T_{jj}(^\circ\text{C}) = 1.9819 \times (\delta^{18}\text{O}_{\text{cellulose}} + (-17 - \delta^{18}\text{O}_{\text{sw}})) - 30.214
\]  

(5)

where \(T_{jj}\) is the temperature in \(\circ\text{C}\), \(\delta^{18}\text{O}_{\text{cellulose}}\) is the \(\delta^{18}\text{O}\) value of cellulose in VSMOW, the constant -17 is the average source water value for all sites used determined using the Online Isotopes in Precipitation Calculator (OIPC v 2.2) (Bowen and Revenaugh, 2003; Bowen, 2010), and \(\delta^{18}\text{O}_{\text{sw}}\) is the source water value for the location of the sample. The term \((-17 - \delta^{18}\text{O}_{\text{sw}})\) was added to account for the difference in source water values between the sites. The error is \(\pm 3.2 \circ\text{C}\).

In order to reconstruct an annually resolved record of JJ temperature we have combined the two functions in Equation 6;

\[
T_{jj}(^\circ\text{C}) = 1.9819 \times ((\delta^{18}\text{O}_{\text{cellulose}} + (-17 - \delta^{18}\text{O}_{\text{sw}})) - 33.1484 + 2.2333 \times \text{RW}
\]  

(6)

where \(T_{jj}\) is the annually resolved temperature in \(\circ\text{C}\), \(\delta^{18}\text{O}_{\text{cellulose}}\) is the isotopic value of tree ring cellulose, \(\delta^{18}\text{O}_{\text{sw}}\) is the source water value local to the site where the tree-ring samples were collected, and RW is the ring-width index. The error for this function determined by taking the square root of the sum of the squared errors for Equations 4 and
5 is $\pm$ 3.2° C. $R^2$ for this combined transfer function for the validation period (1973-1994) from the Yakutia location is 0.60 (RE = 0.27; CE = 0.27).

To ensure that this transfer function was applicable to sites other than north eastern Siberia we obtained ring-width and $\delta^{18}O$ data from the literature for other boreal forest sites including a second collection from the Yakutia site (Sidorova et al. 2008) and localities in western Siberia (Holtzkamper et al., 2009), central Siberia (Knorre et al. 2010), northern Finland (Hilasvouri et al. 2009) and the MacKenzie Delta (Porter et al., 2009). Climate data for these additional sites was obtained from the Global Historical Climate Network (GHCN) (http://www.ncdc.noaa.gov/ghcnm/) and where possible used the same station as was used in the original study: Khoseda Khard, Russia (67.05°N, 59.23°E) (Holtzkamper et al., 2009), Shira, Russia (54.30° N, 90.12° E) (Knorre et al., 2010), Sodankyla, Finland (67.37°N, 26.65°E) (Hilasvouri et al., 2009), Andenes, Norway (69.30° N, 16.2° E) (G. Young, pers. comm.), Inuvik, Canada (68.3° N, 133.48° W) and Chokurdak, Russia (70.62° N, 147.88° E). For each additional site the period 1950 to 2000 was used. We found that using Equation 6 we were able to reconstruct JJ temperature for the second Yakutia site with an $R^2$ of 0.46, for western Siberia with an $R^2$ of 0.21, for central Siberia with an $R^2$ of 0.23 for northern Finland with an $R^2$ of 0.47 and for the MacKenzie Delta with an $R^2$ of 0.26. This provides a measure of the applicability of our transfer function to sites outside of Yakutia.

Studies have shown that $\delta^{13}C$ from tree rings of boreal forest trees can also provide a proxy record of growing season temperatures (Hemming et al., 1998; Gagen et al., 2004; Gagen et al., 2007; Kirdyanov et al., 2008; Sidorova et al. 2008; 2009; Porter et
We also obtained data from other modern boreal forests, with sites from the Mackenzie Delta in Canada (Porter et al. 2009) and Northern Norway (G. Young, pers comm). $R^2$ for June-August temperatures regressed against tree ring $\delta^{13}C$ was 0.67 ($p = <0.001$) for the Mackenzie Delta (Porter et al., 2009) and 0.50 for Northern Norway. Using the $\delta^{13}C$ values from both of these sites we produced a transfer function for the relationship between $\delta^{13}C$ values and temperature, Equation 5;

$$T_{\text{jja}}(°C) = 1.85 \times \delta^{13}C_{\text{cellulose}} + 56.4$$ (5)

where $T_{\text{jja}}$ is the June-August temperature in °C and $\delta^{13}C_{\text{cellulose}}$ is the $\delta^{13}C$ cellulose value of the tree rings.

B.4.3 Spectral Analysis

The Multi Taper Method (MTM) of spectral analysis provides a means for spectral estimation of ‘noisy’ time series (Thomson, 1982; Percival and Walden, 1993) and has been widely applied to problems in geophysical signal analysis, including analyses of atmospheric and oceanic data and paleoclimatic data. MTM offers the appeal of being nonparametric and attempts to reduce the variance of spectral estimates by using a small set of tapers rather than the spectral window used by other methods. We used the MTM in our study as a nonparametric method for investigating quasi-periodic signatures in our time series.
B.5 RESULTS

B.5.1 Tree-ring width

Growth ring widths from the four specimens (min = 0.05 mm, median = 0.38 mm, max = 2.02 mm, n=933) are comparable to those of modern *Larix* growing at ~72° N in Siberia (Creber and Francis, 1999; Hughes et al., 1999) (Table B1). We were not able to crossdate specimens, which means that we cannot rule out the possibility of missing rings being present in some samples. However, missing rings are rare in boreal trees. It is hoped that in the future with more careful sample selection it may be possible to crossdate trees from this site. A reduction in ring width at year 80 in specimen US357 is most likely the result of an injury the tree sustained at this time, from which the tree was not able to fully recover. Specimen NUPB 99 was sampled starting at ring 20, and NUPB 100 was sampled starting at ring 100. Sample NUPB 341 was sampled from pith.

<table>
<thead>
<tr>
<th>Age (yrs)</th>
<th>US357</th>
<th>NUPB 341</th>
<th>NUPB 99</th>
<th>NUPB 100</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>*203</td>
<td>250</td>
<td>234</td>
<td>246</td>
</tr>
<tr>
<td>Avg. ring width (mm)</td>
<td>0.39</td>
<td>0.40</td>
<td>0.28</td>
<td>0.49</td>
</tr>
<tr>
<td>Range (mm)</td>
<td>0.05-1.49</td>
<td>0.14-0.85</td>
<td>0.05-0.96</td>
<td>0.06-2.02</td>
</tr>
<tr>
<td>SD</td>
<td>0.31</td>
<td>0.15</td>
<td>0.18</td>
<td>0.32</td>
</tr>
<tr>
<td>MS</td>
<td>0.40</td>
<td>0.33</td>
<td>0.28</td>
<td>0.33</td>
</tr>
</tbody>
</table>

*Pith was not present thus US357 is counted from the innermost ring

Table B1. Table of the four samples measured for oxygen isotope analysis. Age is determined from the number of rings. Range is the range of ring widths, SD is the standard deviation of the ring widths and MS is the mean sensitivity.
In Boreal forests, temperature is considered to be the dominant factor affecting tree growth (Briffa, 2000; Hughes et al., 1999). Thus, narrow rings would represent cold years whereas wide rings likely formed during warm years. Temperature interpretation from ring widths can be confounded by other factors, such as availability of water and nutrients, irradiance, insect predation, and the length of the growing season (Creber, 1977). Additionally, temperature reconstructions from ring widths are based on transfer functions applied to indices, thus a tree-ring width-based reconstruction using a modern analogue site would reconstruct the exact temperatures of that analogue site and not that of the Pliocene. A multi-proxy study (ring width, $\delta^{18}$O, $\delta^{13}$C) can be used to characterize different environmental factors affecting the tree for example $\delta^{18}$O is influenced by annual temperature, $\delta^{13}$C and ring width are influenced by summer temperature. Additionally as isotopic proxies rely on the raw value and not indices, this means that transfer functions based on modern analogues can indeed be used for temperature reconstructions (Ballantyne et al. 2006; Richter et al., 2008b; Porter et al., 2009). Another advantage posed by isotopic data is that some factors that affect ring width have no effect on the cellulose isotopes. Insect population dynamics, for example, will affect tree growth and thus ring widths. However, studies suggest that insect predation has no effect on isotope values, although it affects ring width (Kress et al., 2009). It is reasonable to assume, therefore, that any signals present both in isotopic systems as well as the tree-ring record likely result from external parameters, such as climatic changes.

B.5.2 Stable isotopes
Two-hundred and three consecutive years of $\delta^{13}C$, $\delta^{18}O$ and $\delta D$ isotope values were derived from specimen US357 (Fig. B3). $\delta^{13}C$ values range from $-23.9 \%_o$ to $-20.2 \%_o$, VPDB; $\delta^{18}O$ values range from $15.4 \%_o$ to $24.5 \%_o$ VSMOW; and $\delta D$ values range from $-204 \%_o$ to $-157 \%_o$ VSMOW. Beginning at year 80 we observe a general decrease in $\delta D$ values, from $-170$ to $-204 \%_o$ VSMOW. We note that the decrease in $\delta D$ values coincides with the decrease in ring widths. Ring widths, however, remain narrow while $\delta D$ values increase after year 98 to values averaging ca. $-180 \%_o$ VSMOW. Results for specimen NUPB 341 reveal $\delta^{18}O$ values ranging from $17.1 \%_o$ to $23.4 \%_o$ VSMOW (Fig. B4). We note a decreasing trend in $\delta^{18}O$ until year 50 that corresponds with an increasing trend in ring width, which we relate to a juvenile effect possibly related to root depth. One-hundred rings were sampled from NUPB 99 and NUPB 100: we used rings 100-200 from NUPB 99 and rings 20-120 from NUPB 100. The $\delta^{18}O$ cellulose values range from $19.6 \%_o$ to $25.0 \%_o$ VSMOW for NUPB 99 and $18.7 \%_o$ to $23.9 \%_o$ VSMOW for NUPB 100 (Fig B5).
Figure B3. $\delta^{13}$C, $\delta^{18}$O, $\delta$D and ring-width for 203 years from specimen US357.

Figure B4. $\delta^{18}$O and ring-width measurements for 250 rings from specimen NUPB 341.
Figure B5. $\delta^{18}$O and ring-width measurements for a 100-year series from specimens NUPB 99 (bottom) and NUPB 100 (top). NUPB 99 starts at ring 100 and goes to ring 200 and NUPB 100 starts at ring 20 and goes to ring 120.

B.6 DISCUSSION
B.6.1 Temperature estimates

In order to use mechanistic models to calculate temperature, the meteoric water value is required. For this study we used the meteoric water value of -20.7 ‰ derived from $\delta^{18}$O values of aquatic moss fragments isolated from peat (Ballantyne et al., 2010; Csank et al., 2011). This lake/fen water value should be an amalgam of annual meteoric water isotopic values such that we can use their calculated water $\delta^{18}$O value here to calculate temperatures from the isotopic values of wood. It is, however, possible that the
peat-inferred water values are lower than average annual precipitation values because peat would more closely represent ground water, which contains a larger winter precipitation signal (Csank et al., 2011). The Roden et al. (2000) equation does not have a term for temperature, thus temperatures were calculated by first determining the δ¹⁸O value of xylem water using the Roden et al. (2000) (Eq. 1), then the water δ¹⁸O-temperature relationship originally established by Dansgaard (1964) and updated by (Rozanski, 1993) was used to calculate the inferred temperature. This method yields mean annual temperature (MAT) estimates from -6.2 to -0.4°C with an average of -3.0 ± 2.4°C (Fig 6). The water δ¹⁸O-temperature relationship was also used to calculate temperatures using the Anderson et al. (2002) model. MAT ranged from -1.2 to +1.4°C with an average of 0.3 ± 1.1°C (Fig B6). These interpretations assume that the relationship between δ¹⁸O of meteoric water and temperature in the present has remained unchanged since the Pliocene.
Figure B6. Mean annual temperature reconstructions from 4 Pliocene trees based on $\delta^{18}$O data series. Reconstructions shown include the Roden et al. (2000) (dark grey) and Anderson et al. (2002) (light grey) mechanistic models presented with the shaded area indicating the range of values obtained if humidity is varied from 60 to 80%. Darkest shading indicates the overlap of both reconstructions.

The interannual temperature variability in a single tree derived from the Roden et al. (2000) mechanistic models showed a wide range from $-15.9$ to $+8.1^\circ$ C with a standard deviation (SD) of $4.5^\circ$ C. This range of variability may have resulted because terms such as humidity, isotopic value of leaf water and vapour pressure deficit were held constant for the entire series, although in nature these values would also vary from year to year. The interannual variability in a single tree using the Anderson et al. (2002) model was much less than that of the Roden et al. (2000) model, ranging from $-7.5$ to $+4.9$ $^\circ$ C.
likely because the Anderson et al. (2002) model is a simplified version of the Roden et al. (2000) model and so includes fewer unknowns. Much work is still needed in order to establish a reliable mechanistic model for tree-ring isotope values that is applicable to a paleoclimatic setting, something we hope to address in ongoing studies.

Two transfer functions developed for ancient wood specimens by Ballantyne et al. (2006) and Richter et al. (2008b) have been applied to the isotopic ratios and ring-width data determined from our four specimens. MAT calculated using the Ballantyne et al. (2006) transfer function range from –10.4 to –2.9°C with an average of -6.2 ± 3.1°C. Interestingly, this is similar to the MAT of -5.5 ± 1.9°C determined by Ballantyne et al. (2006) for Pliocene wood δ¹⁸O from the Beaver Pond locality; however, it differs from the MAT of -0.5 ± 1.9°C determined by Ballantyne et al. (2010). Temperatures calculated using the Richter et al. (2008b) transfer function range from –13.0 to –9.9°C with an average of -11.2 ± 2°C., although Richter et al. (2008b) state that without knowledge of humidity, their transfer function can be off by as much as 16°C. A transfer function relating MAT to ring width alone was developed for Siberian trees by Naurzbaev et al. (2004) we have also used this ring-width-based transfer function to provide a measure of average MAT independent of δ¹⁸O. Temperatures determined using the Naurzbaev et al. (2004) function indicate an average MAT of –8.5 ± 1.5°C. Our mechanistic model derived MAT of -1.9 ± 2.2°C is much warmer than temperatures calculated from our four trees using these transfer functions. However, our temperature estimates are in line with MATs calculated by Ballantyne et al. (2010) for the Beaver
Pond locality using a vegetation coexistence approach (MAT = -0.4 ± 4.1 °C), δ¹⁸O of tree rings (-0.5 ± 1.9 °C) and bacterial tetraethers (-0.6 ± 5.0 °C).

Average temperatures calculated using our growing season (June-July) temperature transfer function (Eq 6) range from 8.1 to 23.2 °C with an average of 15.8 ± 5.0 °C (Fig B7). These temperatures are 11 to 13 °C warmer than modern growing season (JJ) temperatures on Ellesmere Island (4.0 ± 1.2°C).

Using a transfer function for δ¹³C developed from records provided by G. Young (pers comm.) and obtained from Porter et al. (2009) (Eq 7) we determined growing season temperatures from the δ¹³C record of specimen US357 ranging from 12.4 to 18.7°C with an average of 15.2 ± 1.3 °C. Previous reconstructions of growing season temperatures from Strathcona Fiord include warm month mean temperatures (WMMT) of 12.4°C determined from beetle fauna (Elias and Matthews, 2002) and 14.4 ± 2°C determined from a vegetation coexistence approach (Ballantyne et al., 2010). Both WMMT values fall within the range of annually resolved values determined in this study with the beetle fauna reconstruction appearing to be biased towards the cold end of the reconstructions. Mollusc temperatures determined by Csank et al. (2011) from both ‘clumped isotope’ and stable isotope geochemistry average 12.7 ± 1.9°C, also within the range of temperatures determined here. Mollusc temperatures are, however, inferred to represent the entire period of calcification of the mollusc and so could be influenced by temperatures in the spring and fall as well as JJ temperatures.
Figure B7. Growing season (JJ) temperature reconstructions from 4 Pliocene trees. In black is the reconstruction using equation 6 with the dark grey shading indicating the error. The lighter grey line in figure (A) is the growing season (JJA) temperature reconstructed using the δ\textsuperscript{13}C record in equation 7.

Because of the lack of crossdating in our samples, it is possible that each of our four trees represents climate from a different time interval. In order to assess whether our four specimens are recording similar climates, we have examined the standard deviation in mean temperatures calculated from each tree and compared these to the interannual within-tree standard deviation and found that the average within-tree variability of ca. 10° C is greater than the maximum temperature difference between trees of 4° C, suggesting that the within-tree variability is greater than the between-tree variability. A summary of our temperature reconstructions is presented in table B2.
If we wish to compare our JJ temperature estimates to our estimates of MAT and to better compare our results to other studies, we have determined the difference in temperatures between the Pliocene and modern temperatures (ΔT) on Ellesmere Island. Modern temperatures were obtained for Eureka, Nunavut, located in central Ellesmere Island from Environment Canada (Environment Canada, 2008). The temperature record at Eureka extends from 1950 to 2010. The modern MAT at Eureka is -19.7 ± 1.3º C and modern JJ temperatures are 4.0 ± 1.2º C. Our results indicate that MAT was 18.3 ± 4.1º C warmer than present and JJ temperatures were 11.8 ± 5.1º C warmer than present (Fig. B8). The higher ΔT of MAT is consistent with winter amplification of warming. Amplification of Arctic warming particularly in winter is attributed to the loss of sea ice and surface albedo change (e.g. Holland and Bitz, 2003). Thus the greater winter warming on Ellesmere Island could be related to reduced sea ice allowing for a winter climate more moderated by the influence of the ice free Arctic Ocean. The dark, low albedo ocean readily absorbs solar energy during the summer, increasing the summer heat content of the ocean. Ice formation in the winter is then delayed, or non-existent, which then allows for increased winter heat transfer from the ocean to the atmosphere (e.g., Hall, 2004; Serreze et al., 2009; Polyak et al., 2010). When we compare our results in terms of ΔT with previous studies our results for ΔT(MAT) (18.3 ± 4.1º C) are consistent with the results of Ballantyne et al. (2010) (19.3 ± 0.4º C) but warmer than the ΔT(MAT) determined by Ballantyne et al. (2006) (14.2 ± 1.9º C). Our estimates of ΔT(JJ) (11.8 ± 5.1º C) are consistent with the results of Ballantyne et al. (2010) (10.8 ± 2.0º C), Elias and Matthews (2002) (ca. 10º C) and Csank et al. (2011) (14.3 ± 2.2º C).
This suggests that our finding of a $\Delta T$(MAT) (Pliocene-Modern) greater than the $\Delta T$ (Pliocene-modern) of JJ temperatures is consistent among multiple proxies (Fig. B8).

One of the arguments for why proxy data are important is as a means to validate model reconstructions of the past. A recent model comparison study of two mid-Pliocene models by Haywood et al. (2009) compared mid-Pliocene model results from the HADAM3 and GCMAM3 general circulation models. Results obtained for the high Arctic from both these models indicate Pliocene $\Delta T$-MATs of ca. 12º C (HADAM3) ca. 9º C (GCMAM3) and $\Delta T$-JJA temperatures of ca. 8º C (HADAM3) ca. 12º C (GCMAM3) (Haywood et al., 2009). $\Delta T$-JJA temperatures in both models are consistent with $\Delta T$-JJ temperatures determined in our study and others (Fig. B8). It is interesting to note, however, that $\Delta T$-MATs calculated in ours and other studies appear to be
Table B2. Temperature estimates from all trees. All temperatures are based on an average of all years for a given tree. *Calculated using the Roden et al. (2000) equation to find source water and the Dansgaard relationship (Dansgaard, 1964) to determine temperature. § Difference from modern temperature based on an average growing season temperature (JJ) for Eureka, Ellesmere Island of 4.0° C and a MAT of −19.7° C.

<table>
<thead>
<tr>
<th></th>
<th>growing season (JJ) temperature (°C)</th>
<th>mean annual temperature (°C)</th>
<th>Anderson et al. (2002)</th>
<th>Roden et al. (2000)*</th>
<th>Ballantyne et al. (2006)</th>
<th>Richter et al. (2008)</th>
</tr>
</thead>
<tbody>
<tr>
<td>US357</td>
<td>15.4</td>
<td>15.2</td>
<td>-1.2</td>
<td>-6.3</td>
<td>-10.3</td>
<td>-13.0</td>
</tr>
<tr>
<td>NUPB 341</td>
<td>15.3</td>
<td>15.2</td>
<td>0.3</td>
<td>-3.1</td>
<td>-6.0</td>
<td>-11.1</td>
</tr>
<tr>
<td>NUPB 99</td>
<td>15.8</td>
<td>15.2</td>
<td>0.7</td>
<td>-2.3</td>
<td>-5.3</td>
<td>-10.7</td>
</tr>
<tr>
<td>NUPB 100</td>
<td>17.0</td>
<td>15.2</td>
<td>1.4</td>
<td>-0.5</td>
<td>-2.9</td>
<td>-9.9</td>
</tr>
<tr>
<td>Averages</td>
<td>15.4 ± 1.2</td>
<td>15.8 ± 5.0</td>
<td>0.3 ± 1.1</td>
<td>-3.0 ± 2.4</td>
<td>-6.2 ± 3.1</td>
<td>-11.2 ± 1.3</td>
</tr>
<tr>
<td>Difference from modern ΔT (°C)§</td>
<td>11.4 ± 1.8</td>
<td>11.8 ± 5.1</td>
<td>20.0 ± 1.7</td>
<td>16.7 ± 2.7</td>
<td>13.5 ± 3.3</td>
<td>8.5 ± 1.8</td>
</tr>
</tbody>
</table>

significantly higher than ΔT-MATs determined from both models. This difference could be partially related to the fact that the ΔT of the models was calculated using a difference from pre-industrial temperatures for the entire Arctic region, whereas our ΔT–MAT is based on 1950-2010 instrumental data from a single station and so could be representing a difference between regional and global climate. However it is also possible that models are not adequately capturing the degree of winter warming present at high latitudes during the Pliocene. Enhanced winter warming could be related to sea-ice and ocean circulation feedbacks that aren’t adequately represented in the current Pliocene models.
Figure B8. Pliocene temperatures in terms of difference from modern ($\Delta T$). Black symbols represent MAT and white symbols represent JJ or growing season temperatures. (○) This study, (▲) Ballantyne et al. (2006), (□) Ballantyne et al. (2010), (▵) Elias and Matthews, (2003) and (◇) Csank et al. (2011). Grey shading indicates the range of values covered by our annually resolved tree-ring data. Error bars represent the 1 σ error.

B.6.2 Meteoric water

Using the modeling equations of Roden et al. (2000) and Anderson et al. (2002), we were able to obtain some estimates of temperature and source water using the isotope values derived from the Pliocene wood. For the source water $\delta^{18}O$ calculations, the necessary annually resolved paleotemperatures were determined using the annually resolved temperatures derived from $\delta^{13}C$. However, as $\delta^{13}C$ data were available only for specimen US357 the average gastropod inferred temperature of 12.7 °C from Csank et al. (2011) was used for all other specimens. Relative humidity was estimated as 74% using modern relative humidity data for Northern Canada, Norway, Siberia and Alaska; however, we varied humidity from 60% to 80% to present a range of inferred values. Source water for each year was then calculated using these parameters and the Anderson
et al. (2002) equation. Using the Roden et al. (2000) equation with calculated values for leaf water and the same $f$ and $h$ values used for the Anderson et al. (2002) equation calculations, we were able to calculate annual source water values from the tree.

The average of annual $\delta^{18}O_{sw}$ values calculated using the Roden et al. (2000) equation was ranged from $-25.1 \, ^{\circ}\text{C}$ to $-8.7 \, ^{\circ}\text{C}$ with an average of $-16.3 \, ^{\circ}\text{C}$. Source water values calculated using the Anderson et al. (2002) equation range from $-22.3 \, ^{\circ}\text{C}$ to $-10.9 \, ^{\circ}\text{C}$ with an average of $-16.3 \, ^{\circ}\text{C}$. These values are slightly more enriched than the source water $\delta^{18}O$ of $-20.7 \, ^{\circ}\text{C}$ to $-22.1 \, ^{\circ}\text{C}$ determined by Csank et al. (2011). However, Csank et al. (2011) were measuring aquatic moss and aquatic molluscs, both representative of fen water values, which may contain more of a winter precipitation signal than the inferred annual precipitation signal reflected in the $\delta^{18}O$ ratios of tree rings. Using the relationship of $\delta D$ measured in cellulose nitrate to meteoric water determined by Tang et al. (2000) we can work out what the Pliocene $\delta D$ value of meteoric water was using the $\delta D$ values measured on specimen US357. $\delta D$ values calculated in this way range from $-116.5 \, ^{\circ}\text{C}$ to $-175.2 \, ^{\circ}\text{C}$ with an average of $-150.1 \, ^{\circ}\text{C}$. If we compare the average source water $\delta^{18}O$ values calculated here ($\delta^{18}O = -16.3 \, ^{\circ}\text{C}$) to modern Canadian precipitation isotope values, the Pliocene $\delta^{18}O$ values are significantly higher than modern Ellesmere Island summer precipitation (average $\delta^{18}O$ values of ca. $-28 \, ^{\circ}\text{C}$), but are close to those recorded at the present tree-line ($-14$ to $-22 \, ^{\circ}\text{C}$) in areas such as Whitehorse, YT or Yellowknife, NWT (IAEA/WMO, 2006).
B.6.3 Relative Humidity

Several studies have noted a correlation between relative humidity and isotopic ratios of cellulose. Studies that have attempted to reconstruct relative humidity have focused primarily on $\delta^{13}$C and $\delta$D ratios of cellulose (Lipp et al., 1991; Edwards et al., 2000; Jahren and Sternberg, 2003; Porter et al., 2009). We have determined both $\delta^{13}$C and $\delta$D from specimen US357. Using the $\delta^{13}$C values in conjunction with growing-season temperatures inferred from $\delta^{18}$O and ring widths using equation 6 we are able to use the equation for a relative humidity-temperature response surface derived by Edwards et al. (2000) (equation 8) to infer growing season relative humidity;

$$ RH = \frac{([\delta^{13}\text{C} + 6] - (-0.15)T)}{-0.17} $$

where RH is the relative humidity, $\delta^{13}$C is the carbon isotope ratio of cellulose and T is temperature in °C. Relative humidity calculated from this equation averages 83% and ranges from 70% to 90%, which is within the range of average boreal forest humidity.

Jahren and Sternberg (2003) established a humidity estimate for an Eocene Arctic forest using $\delta$D ratios measured from wood cellulose. Their relationship was based in part on the Roden et al. (2000) equation for $\delta$D (Eq 2) and is presented in equation 9:

$$ \delta D_{CN} = (8.13 \times \delta^{18}O_{sw}) + 10.18 + (f_h \times \varepsilon_{HH}) + (1-f_h) \times [(\varepsilon_e + \varepsilon_k) \times (1-h) + \varepsilon_{HA}] $$

(9)
where $\delta D_{CN}$ is the isotopic ratio of nitrated cellulose; $\delta^{18}O_{sw}$ is the oxygen isotopic ratio of environmental water, $f_f$ is the fraction of exchange between xylem water and glucose (35%), $\epsilon_{\text{HH}}$ is the biological fractionation factor for hydrogen during conversion from sugar to cellulose in a heterotrophic reaction (158 ‰), $\epsilon_e$ is the equilibrium fractionation for liquid-vapour, $\epsilon_k$ is the kinetic fractionation factor for hydrogen (25.1 ‰), $h$ is humidity, and $\epsilon_{\text{HA}}$ is the biological fractionation factor for hydrogen during conversion from sugar to cellulose in an autotrophic reaction (−171 ‰).

We derived relative humidity using the Jahren and Sternberg (2003) relationship, with $\delta^{18}O_{sw}$ estimated using both the Roden et al. (2000) and Anderson et al. (2002) mechanistic models. Relative humidity averaged 60% and ranged from 12% to 102%. Although the average humidity of 60% is within the expected range for a boreal forest, the extreme range of humidity determined by the Jahren and Sternberg (2003) relationship suggests that perhaps $\delta D_{CN}$ is recording environmental variables other than simply temperature. This is consistent with results obtained by Tang et al. (2000).

B.6.4 Spectral analysis

B.6.4.1 Results

MTM spectral analysis of tree-ring and isotope series of all four trees reveals periodicity in the ring-width series in the 20-year, 12-year, 10-year, 5.5-year, 3.3-year, 3-year, 2.4-year and 2.25-year frequency bands (Fig. B9). All of these spectral peaks are above the 95% confidence limit for a red-noise spectra of an AR(1) (auto-regressive) model. Greatest power is in the 20-year and the 2.4-year periods. Periodicity in the $\delta^{18}O$
series is in the 20-year, 12-year, 6.25-year, 5.5-year, 4.7-year, 4.2-year, 3.6-year, 2.5-year and 2.1-year frequency bands (Fig 9). All of these spectral peaks are above the 95% confidence limit. Greatest power is in the 12-year and 6.25-year periods. Periods common to both the ring-width and δ¹⁸O series are a 20-year, 12-year and 5.5-year period.

Figure B9. MTM power spectrum of four Pliocene trees and the NAO and AO indices. NAO index data provided by the Climate Analysis Section, NCAR, Boulder, USA (Hurrell, 1995) ‘δ¹⁸O’ and ‘ring width’ represents the stacked spectral analyses of all four trees for δ¹⁸O and ring-width, respectively. Dashed lines represent the 90, 95 and 99% confidence limits using an AR(1) model. Light grey bars represent significant peaks in the δ¹⁸O spectra, medium grey bars represent significant peaks in the ring-width data and the dark grey bars represent peaks common in both spectra. Numbers above the chart is the periodicity in years.
B.6.4.2 Causes of decadal and multi-decadal variability

In addition to the seasonal cycle, several decadal or quasi-decadal modes of climate variability influence Arctic climate. Those that have the largest impact on the eastern Arctic are the North Atlantic Oscillation (NAO) (Hurrell, 1995) and its Arctic component the Arctic Oscillation (AO) (Thompson and Wallace, 1998), known collectively as the Northern Annular Mode (NAM) (Thompson et al., 2000).

The NAM is one of the most prominent climate modes of the Northern Hemisphere. The NAO is characterized by a fluctuation of atmospheric mass between the Arctic, measured at Iceland, and the subtropical Atlantic, measured in the Azores. When the Icelandic Low has lower than normal sea level pressures and the Azores High, higher than normal pressures, the NAO is positive. A positive NAO generates strong northerly winds over Greenland and northeastern Canada forcing cold air southward over the northwest Atlantic, whereas a negative NAO results in warmer conditions to the North (Hurrell et al., 2003). The NAO is most obvious during the winter months but can be identified at any time of year. In the framework of the NAM the NAO is simply the North Atlantic component of a larger more fundamental mode (Thompson et al., 2000). When the NAM is positive, surface pressures are low over the Arctic and high at mid latitudes, which results in strengthening of the Polar Westerlies and an enhancement of the cyclonic activity at high latitudes (Thompson and Wallace, 1998). A positive NAM results in a poleward shift of the circumpolar vortex. This results in a shift in storm tracks to the north, bringing more precipitation to many regions in the Arctic (Zhao et al., 2006). The NAM is defined as the surface signature of changes in the strength of the polar
vortex as detected in a linear principal component measurement of sea-level pressure in the Northern Hemisphere north of 20° N (Thompson and Wallace, 1998). Both the AO and the NAO have strong quasi-decadal peaks at between 8-12 year periods. Because the AO and NAO are highly correlated with each other they are often linked in spectral analyses. A positive NAM results in colder than average temperatures on Ellesmere Island whereas a negative NAM results in warmer than average temperatures. Mean annual temperatures from Eureka, Ellesmere Island (1950-2008) correlate negatively with both the NAO and AO index. June-August temperatures, however, do not correlate. Thus we would expect our δ¹⁸O record to be more likely to show a correspondence with the NAO/AO than the ring-width series. In fact Welker et al. (2005) noted a correlation between time series of δ¹⁸O derived from the Arctic shrub Cassiope tetragona from Ellesmere Island and the NAO.

Examination of the NAO and AO indices using MTM spectral analysis reveals a strong ca. 8-year period in the NAO index, as well as faint 8-year period in the AO index with other significant periods at 2.7, 2.4 and 2.25-years. Both the AO and the NAO index do present spectral peaks in the 2.5-3-year, 3.5-4-year, 4.2-4.7-year and 5.5-year periods, however, these peaks fall below the 90% confidence limit (Fig. B9). In their large-scale Northern Hemisphere proxy reconstruction of the NAO, Cook et al. (2002) found that when a longer term reconstruction was used the spectra of the NAO appeared to change with a ca. 4-year period appearing to have the highest spectral power in their reconstruction. Other significant periods in the Cook et al. (2002) reconstruction are a 2.25-year, 2.4-year 2.7-year and ca. 8-year period. That spectral analysis of our trees has
spectral frequency in common with the AO/NAO could suggest that these modern modes of climate variability were operating in some form during the Pliocene. However, our spectra do not exactly match the spectra of the AO or the NAO index, thus this does not indicate unequivocally that the AO/NAO was extant 4-5 million years ago. Instead it does suggest that decadal scale variability with periods similar to modern climate modes existed in the Early Pliocene. However, given that oceanographic and climatic conditions were proposed to have been substantially different from the modern (Haywood et al., 2000; Lear et al., 2003; Haug et al., 2005), it is possible that these modes could have been operating differently during the Pliocene. Recent modeling studies of what is likely to happen to the NAM with future warming suggest that the NAM will become generally more negative (Deser et al., 2010; Woolings et al., 2010; Morgenstern et al., 2010; Barnes et al., 2010; Kidston and Gerber 2010). Deser et al. (2010) modeled the response of the NAM to an ice-free Arctic Ocean and found that a lack of sea ice produced a change in the position of the jet stream that resembled the negative phase of the NAO, which is consistent with the response observed by Seierstad and Bader (2009), who also noted a consistent negative NAM circulation response with reduced Arctic sea ice. Barnes and Hartmann (2010) note that if the mean state of the North Atlantic jet stream is shifted poleward the persistence of the poleward shifted phase (positive NAM) is reduced, suggesting that perhaps the periodicity of the NAM could change. How this would affect the operation of a Pliocene NAM is the question of ongoing work (Hill et al. in review). Additionally given the size of the Arctic Ocean basin and the fact that it was largely ice free in the Early Pliocene, decadal variability in the Pliocene could even be
related to a hitherto unrecognized sea surface temperature cycle operating in the Arctic Ocean. If this is indicative of the AO/NAO persistence of such climate modes through very different climate regimes, suggests that they may be substantially more robust than is currently assumed or there may be some other driver of these modern climate modes that is unaffected by changes in climatic boundary conditions. Future studies will expand upon the promising results obtained in this study through analysis of additional samples from this site and samples from other Pliocene Arctic forest sites, to identify additional decadal-scale climate cycles during the Pliocene (Wara et al., 2005; Fedorov et al., 2006; Bonham et al., 2009).

B.7 CONCLUSIONS

Our temperature reconstruction reveals growing-season (JJ) temperatures $15.8 \pm 5.0 \, ^\circ C$, with a range of interannual variability of ca. $2 \, ^\circ C$. Mean annual temperatures reconstructed in this study average $-1.4 \pm 4.0 \, ^\circ C$ with an interannual range of variability of ca. $2.6 \, ^\circ C$. Growing season (JJ) temperatures are $11.8 \pm 5.1 \, ^\circ C$ warmer than present and mean annual temperatures are $18.3 \pm 4.1 \, ^\circ C$ higher than modern temperatures on Ellesmere Island, in line with other temperature reconstructions (Ellias and Matthews 2002; Ballantyne et al., 2010; Csank et al., 2011). Although growing-season temperatures are comparable to those determined from modeling studies of the Pliocene, mean annual temperatures are substantially higher. This suggests that perhaps models are not adequately capturing winter warming in the Pliocene at high latitudes. What our study
adds to previous studies of temperature is a detailed interannual reconstruction of temperatures during the Pliocene. Our annually resolved study also suggests that although temperature reconstructions for the Pliocene from Strathcona Fiord vary, all previous estimates fall within the range of natural variability represented by our annually resolved record. Spectral analysis provides evidence of sub-decadal and multi-decadal periodicities similar to well-known modern modes of climate variability in the Arctic region, such as the AO/NAO. Although it is not possible from this study to say conclusively that an AO/NAO was operating in the Pliocene with the same periodicities it has now, it is clear that there was indeed some decadal and inter-decadal climate variability in our records. Future work and better age control will help us to further investigate the mechanisms of this variability.

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B.8 REFERENCES


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APPENDIX C

ANNUALLY RESOLVED TEMPERATURE RECONSTRUCTIONS FROM A LATE PLIOCENE-EARLY PLEISTOCENE POLAR FOREST ON BYLOT ISLAND

Manuscript for submission to *Palaeogeography, Palaeoclimatology, Palaeoecology*
Annually resolved temperature reconstructions from a late Pliocene-early Pleistocene polar forest on Bylot Island

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C.1 ABSTRACT

Here we use $\delta^{18}O$ ratios measured in tree rings of crossdated sub-fossil wood to reconstruct an annually resolved record of temperature and $\delta^{18}O$ of meteoric water for an interglacial late Pliocene-early Pleistocene fossil forest found on Bylot Island, Nunavut, Canada. Our record represents the first crossdated record of Pliocene wood. Mean annual temperatures determined in this study average $\pm 2.9 \pm 3.9$ °C or $11.9 \pm 4.5$ °C warmer than present-day Bylot Island ($-14.8 \pm 2.2$ °C). Meteoric water $\delta^{18}O$ values average $-16.2$‰, $\sim 2-6$ ‰ more enriched than present values of precipitation $\delta^{18}O$. Our temperatures are comparable to mid-Pliocene modeled temperatures for the Arctic (3-5 °C warmer than present), suggesting that interglacial warm periods in the late Pliocene-early Pleistocene may have been as warm as the mid-Pliocene warm period. That both the Bylot Island...
forest deposit and the Kap København deposit represent the remains of northern tree-line vegetation that lived during warm interglacial periods within the overall cool Plio-Pleistocene suggests that forest deposits in the Arctic capture a snapshot of interglacial conditions during the Plio-Pleistocene rather than the average Pliocene climate and may not be suitable records to study Pliocene cooling.

C.2 INTRODUCTION

During the Pliocene and early Pleistocene, polar forests in North America reached north as far as the shores of the Arctic Ocean. Mummified sub-fossil wood remains found at sites throughout the Arctic afford the opportunity to study the Arctic climate system in a generally warmer world (Ballantyne et al., 2006; 2010). Modeled global temperatures are estimated to have been ~3 to 5º C warmer than present and Arctic temperatures on the order of 7-10º C warmer (Sloan et al., 1996; Haywood et al., 2009). The mid-Pliocene (3-3.5 Ma) is considered an accessible example of a future warmer world (Jansen et al., 2007). However, with present CO$_2$ levels fast approaching and soon surpassing Pliocene CO$_2$ levels of 400 ± 25 ppm (Pagani et al., 2010), the Pliocene will shortly cease to be a good example of near future climate. The extreme warmth of the Pliocene under near-present levels of CO$_2$ suggests that it is important to understand the climate system that operated during this time to better understand the sensitivity of Earth’s climate system to changing levels of atmospheric CO$_2$ (Pagani et al., 2010; Lunt et al., 2010).

One of the most intriguing facets of past warm phases was the presence of forests at both poles. Rich Pliocene fossil floras have been described from Alaska (Matthews and
Ovenden, 1990), Antarctica (Francis and Hill, 1996; van Bergen and Poole, 2002), Siberia (Bondarenko, 2007), Greenland (Funder et al., 2001) and Arctic Canada (Matthews and Ovenden, 1990; Fyles et al., 1990; Ballantyne et al., 2006; 2010), all located at or above the Pliocene polar circle.

The evolution of the polar forests can been seen in the shifts in vegetation that occurred between the earliest known Pliocene site on Ellesmere Island to the youngest sites on Meighen Island and Greenland. Flora from the well known Beaver Pond locality near Strathcona Fiord on Ellesmere Island dates to between 4 and 5 million years (Tedford and Harington, 2003) and consisted of Betula (birch), Larix (larch), Picea (spruce), Thuja (cedar), Alnus (Alder) as well as abundant mosses and ferns that are all characteristic of modern boreal flora (Matthews and Ovenden, 1990; Ballantyne et al., 2010). Deposits from a site on Meighen Island in the Canadian Arctic, date to about 3 Ma and show a mixture of tundra and boreal vegetation (Matthews and Ovenden, 1990). Palynological records from the Hvitland beds on Ellesmere Island, dated to around 2.5 Ma, show exclusively tundra vegetation (Fyles et al., 1998) indicating that by this time the polar forests had disappeared.

In paleoclimatic studies of the recent past, much research has gone into examining the range of natural variability in an effort to understand how natural variability is impacted by current warming. However, most paleotemperature records of the Pliocene simply provide temperature estimates for a single point representing several thousands to millions of years and provide no information on what the range of inter-annual variability was. In this paper we present an annually resolved temperature record for ~100 years
during the late Pliocene-early Pleistocene providing a record of inter-annual variability under overall warmer conditions in the Arctic.

Dendroclimatology is a useful tool for providing information about climate in periods prior to instrumental records. Stable isotope analysis of tree rings has extended this tool to periods lacking a modern calibration period as far back as the Eocene (Jahren and Sternberg, 2003; Jahren et al., 2008) and has recently been applied to studies of Pliocene polar vegetation (Ballantyne et al. 2006; 2010).

Here we report annually resolved temperature variations derived from tree-ring δ¹⁸O measurements from crossdated *Larix* (larch) trees recovered from Bylot Island, Canadian Arctic Archipelago, 73° N 80° W (Fig C1). Our reconstruction also represents the first crossdated sequence of Pliocene polar vegetation. Specimens were recovered from within a single stratigraphic layer from fluvial sedimentary deposits. Stratigraphically the wood layer lies within organic-rich alluvial sediments with discontinuous peat layers. This unit directly overlies a till layer inferred to relate to a local glacial event (Piraux, 2004). Dating of the site has been coarsely established by paleomagnetic analysis that indicates a succession of a normal polarity followed by a reversed polarity. This section has been inferred to either represent the period between 1.7 and 1.9 Ma, the short normal polarity period at 2.14 to 2.18 Ma or the period around the Gauss/Matuyama boundary between 2.4 and 2.8 Ma (Piraux, 2004; Zimmermann et al., 2010).
Although it is possible that the forests existed during an interglacial period around 1.9 Ma or the interglacial marine oxygen isotope stage 81 around 2.1 Ma, the absence of tree species in pollen records from the Arctic after 2.2 Ma (Fyles et al. 1998; Harris, 2005) makes it more likely that the site dates to between 2.4 and 2.8 Ma (Fig C2). These dates indicate that the Bylot Island forest is similar in age to the Kap København forest deposits from Greenland, dated to between 2.4 and 2.5 Ma (Bennike and Böcher, 1990; Funder et al., 2001). This also correlated quite closely with several marine transgressions in Alaska (Brigham-Grette and Carter, 1992) suggesting that these forest deposits may represent forests that grew at northern tree-line during interglacial periods within the Plio-Pleistocene. Ongoing work will better resolve the age and stratigraphic relationships.
of the deposit. Bylot Island trees exhibit narrow rings and small diameters associated with tree-line vegetation, with the oldest specimens aged 175 years.

Figure C2. The Lisiecki and Raymo (2005) benthic $\delta^{18}$O stack for the period between 1.8 and 3 million years. Normal and reversed polarity is indicated with normal polarity intervals indicated in black. Several of the MIS interglacial periods are indicated. Grey bars indicate possible ages of the Bylot Island deposits. Also indicated are other Pliocene deposits from the high Arctic with similar ages, including the Kap København (KK), Hvitland Beds (HB), Ile de France (IF) and Meighen Island (MI) deposits.

C.3. METHODS

Ring widths of ten specimens were measured to ±0.01 mm using a Bannister Bench measuring system (Robinson and Evans, 1980). Specimens were carefully crossdated, and crossdating was checked using COFECHA (Holmes, 1983). COFECHA checks for dating and measurement errors by log transforming and de-trending the individual series then removing autoregressive (AR) processes and cross-comparing lagged segments of each series against the chronology of all series (Holmes, 1983). This is done to ensure that the crossdating had no errors (i.e. unidentified missing rings, mis-
matched series). Four of the ten crossdated specimens were subsequently sampled for
stable-oxygen isotope analysis (Table C1).

Table C1. Ring statistics of the four samples chosen for oxygen isotope analysis. Age is determined from
the number of rings. Range is the range of ring widths, SD is the standard deviation of the ring widths and
MS is the mean sensitivity.

<table>
<thead>
<tr>
<th>Age (yrs)</th>
<th>BY 25</th>
<th>BY 24</th>
<th>BY 27</th>
<th>BY 33</th>
</tr>
</thead>
<tbody>
<tr>
<td>82</td>
<td>76</td>
<td>90</td>
<td>109</td>
<td></td>
</tr>
<tr>
<td>Avg. ring width (mm)</td>
<td>0.43</td>
<td>0.55</td>
<td>0.69</td>
<td>0.33</td>
</tr>
<tr>
<td>Range (mm)</td>
<td>0.07-1.49</td>
<td>0.14-1.21</td>
<td>0.21-1.23</td>
<td>0.12-0.81</td>
</tr>
<tr>
<td>SD</td>
<td>0.30</td>
<td>0.26</td>
<td>0.23</td>
<td>0.13</td>
</tr>
<tr>
<td>MS</td>
<td>0.30</td>
<td>0.25</td>
<td>0.27</td>
<td>0.26</td>
</tr>
</tbody>
</table>

Tree rings were annually sampled using micro-milling technology (Dettman and
Lohman, 1995; Dodd et al., 2008) along two radii to account for variation in isotope
composition around the circumference of a given ring (Leavitt, 2010). Sampled rings
were then processed to α-cellulose. Alpha-cellulose is often preferred because different
wood components differ in their isotopic values (Wilson and Grinsted, 1977, McCarroll
et al., 2004), and original isotope composition is preserved in the α-cellulose component
that can be easily isolated. A modified Leavitt-Danzer method, employing sodium
chlorite to remove lignin and sodium hydroxide to isolate α-cellulose, was used to
prepare samples for δ¹⁸O analysis (Leavitt and Danzer, 1993). Stable-oxygen isotope
measurements were conducted at the University of Arizona utilizing a Costech HTG EA
modified for oxygen isotope analysis of cellulose (Evans, 2008) and directly coupled via
a ConFlo III to a Finnigan DeltaPlus XP mass-spectrometer. A mass of 0.30-0.35 mg of cellulose was weighed into silver capsules, pyrolyzed over glassy carbon and reported as δ^{18}O with respect to Vienna-Standard Mean Oceanic Water (VSMOW). Analytical precision for repeat analysis of δ^{18}O ratios of an internal sigma cellulose standard is typically 0.3‰.

Oxygen-isotope ratios of tree-ring cellulose have been used to reconstruct temperature both in the Holocene (e.g. McCarroll and Loader, 2004; Sidorova et al., 2008; Sauer et al., 2008) and in periods in Earth’s geologic past, such as the Eocene (e.g. Jahren and Sternberg, 2003; 2008), Miocene (Richter et al., 2008a) and the Pliocene (Ballantyne et al., 2006, 2010; Richter et al., 2008a).

The oxygen isotope ratio of tree-ring cellulose is controlled by the isotopic composition of source water (rain, soil water, ground water), evaporative enrichment in the leaves and biochemical fractionations during cellulose synthesis (Roden et al., 2000; McCarroll and Loader, 2004). Water is taken up through the roots without fractionation, but fractionation occurs through evaporation from the leaves during transpiration when leaf water becomes ^{18}O enriched relative to source water. Leaf water is synthesized into sucrose without fractionation and is subsequently modified in the trunk through exchange with xylem water (source water) prior to formation of cellulose (Roden et al., 2000). These fractionations are consistent for trees of the same species and from year-to-year and thus the year-to-year variability in cellulose δ^{18}O reflects the variability in source water. The primary source of water for trees is soil water; thus, the primary source of oxygen isotopes in cellulose is also soil water. The δ^{18}O of water in the upper portion of
the soil column is related to the isotopic composition of meteoric precipitation (Tang and Feng, 2001; Anderson et al., 2002), and therefore much of the isotopic signal in cellulose δ^{18}O should reflect the isotopic signature of precipitation.

The stable isotopic composition of precipitation integrates a variety of factors including temperature and precipitation amount at the site of deposition. The fundamental control on the isotopic composition of precipitation is temperature. With increasing temperature, precipitation becomes enriched in the heavier isotope ^{18}O, in a linear relationship that was first described by Dansgaard, (1964). In addition to temperature the isotopic value of precipitation becomes depleted with increasing elevation and increasing distance from the equator (both of which also correspond to lower temperature). The relationship of precipitation isotopes to temperature is nicely summarized in Kohn and Welker (2005).

In order to understand the relationship of oxygen isotopes in tree rings to source water Roden et al. (2000) developed mechanistic models based on laboratory studies of various tree species. Their premise was that oxygen isotopes incorporated into cellulose from water are subject to fractionation at several points and can be modeled by Equation 1:

\[
\delta^{18}O_{\text{cellulose}} = f_o \times (\delta^{18}O_{\text{wx}} + \varepsilon_o) + (1 - f_o) \times (\delta^{18}O_{\text{wl}} + \varepsilon_o)
\]  

(1)

where \( f_o \) is the fraction of carbon-bound oxygen that undergoes exchange with xylem water (≈ 35 %); \( \delta^{18}O_{\text{wx}} \) is the oxygen isotope ration of xylem water; \( \varepsilon_o \) is the biological fractionation factor for oxygen during conversion from sugar to cellulose (27 ‰); \( \delta^{18}O_{\text{wl}} \)
is the oxygen isotope ratio of leaf water as determined by the Flanagan et al. (1991) model presented in equation 2:

$$\delta^{18}O_{wl} = [((\alpha_k (e_i - e_a/e_i) + R_x (e_a/e_i)) / 0.0020052) - 1] \times 1000$$  \hspace{1cm} (2)$$

where $\alpha$ is the liquid-vapor fractionation factor derived from Manjoube (1971); $\alpha_k$ is the kinetic diffusion factor of water across the leaf membrane (1.0285); $R_x$ is the molar ratio of $^{18}$O to $^{16}$O in xylem (stem) water; $e_i$ is the partial pressure of water vapor in a leaf; $e_a$ is the partial pressure of water vapor in the atmosphere; $R_a$ is the molar ratio of $^{18}$O to $^{16}$O in water vapor. A simplified model was derived by Anderson et al. (2002), which combines the terms from the Roden et al. (2000) wood cellulose model and the Flanagan et al. (1991) leaf water model into a simplified equation presented in equation 3:

$$\delta^{18}O_{sw} \approx \delta^{18}O_{cellulose} - (1 - f) \times (1 - h) \times (\varepsilon_e + \varepsilon_k) - \varepsilon_{biochem}$$  \hspace{1cm} (3)$$

where $\delta^{18}O_{sw}$ is the isotopic value of source water; $\delta^{18}O_{cellulose}$ is the isotopic value of cellulose; $f$ is the fraction of leaf water not subject to evaporation and accounting for the change in isotope value of photosynthate by stem water interaction; $\varepsilon_e$ is the liquid-vapor equilibrium fractionation factor; $\varepsilon_k$ is the liquid-vapor kinetic fractionation dependent on airflow at the leaf boundary layer (28 %); $h$ is the relative humidity and $\varepsilon_{biochem}$ is the biologic fractionation factor for sugar converting to cellulose (27 %).
(2002) found that a variable value of $f$ dependent on humidity was much more useful in the model than a fixed value for $f$.

Use of both of these models requires knowledge of humidity, vapour pressure deficit (VPD) and/or the isotopic value of the source water, thus the use of mechanistic modeling equations in a paleoclimatic context remains problematic, but if some careful assumptions are made, mechanistic models can be a useful method of determining past temperatures from tree-ring cellulose (Waterhouse et al. 2002). We have utilized both the Roden et al. (2000) (Eq 1 and 2) and the simplified Anderson et al. (2002) (Eq 3) equations to estimate source water values. However, to do this certain assumptions are necessary. For humidity (Eq 1, 2 and 3) we determined an average humidity of ~74% from instrumental data from Alaska, Northern Canada, Scandinavia and Russia and then determined sensitivity to humidity by allowing it to vary between 60% and 80%, so a range of possible values is presented. In order to determine partial pressures of water vapor ($e_i$ and $e_a$ in Eq 2) we used the saturation vapour pressure of water at 20° C for $e_a$, 20° C being the inferred temperature of a leaf during photosynthesis (Helliker and Richter, 2008). We then solved for $e_i$ by using the average stomatal conductance values for larch.

Once we determined the isotopic value of source water we were able to use the relationship between $\delta^{18}O$ of precipitation and temperature to solve for temperature (Dansgaard, 1964; Rozanski, 1993; Kohn and Welker, 2005). This of course assumes that the relationship between precipitation $\delta^{18}O$ and temperature is unchanged since the Pliocene.
In order to independently reconstruct source water values for use in the δ\(^{18}\)O transfer functions (Ballantyne et al. 2006; 2010; Richter et al., 2008b; Csank et al. (in review) we chose to use a method similar to that used by both Ballantyne et al. (2006) and Richter et al. (2008b). Both Ballantyne et al. (2006) and Richter et al. (2008b) examined the relationship between the δ\(^{18}\)O values of annual precipitation (ppt) with the δ\(^{18}\)O values of cellulose that they analyzed from various sites around the world. We decided to go a step further and include data from the large body of published δ\(^{18}\)O cellulose values from around the world rather than relying on a more limited data set. We used precipitation isotopic composition for sites reporting a measured value, but many sites in the literature used stations from the Global Network of Isotopes in Precipitation (GNIP) that were at some distance from the site at which the tree-ring samples were collected. Thus, rather than use nearby stations we elected to use the Online Isotopes in Precipitation Calculator (OIPC v 2.2) (Bowen and Revenaugh, 2003; Bowen, 2010) using the site locality information (latitude, longitude and elevation) for each δ\(^{18}\)O cellulose site to parameterize the model. The resulting regression for the relationship between δ\(^{18}\)O cellulose and δ\(^{18}\)O in precipitation is strongly correlated (R\(^{2}\) = 0.70, F = 77.4 p < 0.001) (Fig C3). The resulting regression equation (Eq 4) was used to determine the δ\(^{18}\)O value of precipitation (ppt) for Bylot Island. It should be noted that both the regression line of Richter et al. (2008a) and Ballantyne et al. (2006) fall within two standard error of our Equation 4.

\[
\delta^{18}O_{cellulose} = 0.6109 \times \delta^{18}O_{ppt} + 33.2054
\] (4)
Regression relationship between δ¹⁸O in precipitation (ppt), based on the OIPC v2.2 (Bowen and Revenaugh, 2003), and δ¹⁸O in tree-ring cellulose (the average of all rings from a site). Grey line is the regression relationship determined by Richter et al. (2008a). Curved dashed grey line is the regression relationship determined by Ballantyne et al. (2006). Black line is the relationship for all points. All data were obtained from a review of current literature (Epstein et al., 1977; Buhay and Edwards, 1995; Sauer et al., 1998; Anderson et al., 1998; Roden and Ehleringer, 2000; Sauer et al., 2000; Robertson et al., 2001; Waterhouse et al., 2002; Evans and Schrag, 2004; Shu et al., 2005; Welker et al., 2005; Ballantyne et al., 2006; Wright and Leavitt, 2006; Holzkämper et al., 2008; Norström et al., 2008; Richter et al., 2008b; Sidorova et al., 2008; Voltas et al., 2008; Battaglia et al., 2009; Hilasvouri et al., 2009; Horacek et al., 2009; Liu et al., 2009; Porter et al., 2009; Reynolds-Henne et al., 2009; Sidorova et al., 2009; Bale et al., 2010; Knorre et al., 2010).
Previous studies that have reconstructed past climate from oxygen isotope values of fossil wood have relied on modern analogs to determine transfer functions for isotope climate relationships (Jahren and Sternberg, 2003; Ballantyne et al., 2006; Richter et al., 2008a). Jahren and Sternberg (2003) developed a transfer function based on samples of *Metasequoia* targeted to be analogous to the Eocene fossil forest site that they were trying to reconstruct. Richter et al. (2008b) developed a continental-scale transfer function from several sites across North America (Eq 5):

\[
\delta^{18}O_{\text{cellulose}} = -0.01 T^2 + 0.49 T + 26.84 \tag{5}
\]

where \(\delta^{18}O_{\text{cellulose}}\) is the isotopic value of tree-ring cellulose in VSMOW and \(T\) is the mean annual temperature (MAT) in °C. Ballantyne et al. (2006) developed their transfer function to be analogous to a Pliocene forest site on Ellesmere Island that is similar to our site on Bylot Island, based upon two modern larch trees, one from the Yukon and one from Ottawa (Eq 6):

\[
T = 17.5 + 0.98 \times \delta^{18}O_{\text{ppt}} + -2.71 \times \text{RW} \tag{6}
\]

where \(T\) is the MAT in °C, \(\text{RW}\) is the ring width in millimeters and \(\delta^{18}O_{\text{ppt}}\) is the isotopic value of precipitation determined by the Ballantyne et al. (2006) regression in Figure 2 and described by Equation 7:
\[ \delta^{18}O_{\text{ppt}} = 312.75 \times e^{(-0.13 \times \delta^{18}O_{\text{cellulose}})} \]  \hspace{1cm} (7)

where \( \delta^{18}O_{\text{ppt}} \) is the inferred isotopic value of precipitation in VSMOW and \( \delta^{18}O_{\text{cellulose}} \) is the isotopic value of tree-ring cellulose in VSMOW.

Ballantyne et al. (2010) then updated this relationship by using source water values calculated from aquatic moss rather than source water values themselves reconstructed by a transfer function. However, moss samples are not available at our site to directly infer source water composition, so the relationship established in Equation 4 is used (Fig C3).

Csank et al. (in review) developed transfer functions for growing season (June-July) temperature from five specimens of Siberian larch using tree rings representing the years 1950-1994 (Eq 8). The Csank et al. (in review) transfer function, like the Ballantyne et al. (2006) function for mean annual temperature, also requires preexisting knowledge of source water:

\[ T_{jj} \,(^\circ C) = 1.9819 \times ((\delta^{18}O_{\text{cellulose}} + (-17 - \delta^{18}O_{\text{sw}})) - 33.1484 + 2.2333 \times \text{RW} \]  \hspace{1cm} (8)

where \( T_{jj} \) is the June-July temperature in \(^\circ C\), \( \delta^{18}O_{\text{cellulose}} \) is the isotopic value of tree-ring cellulose, \( \delta^{18}O_{\text{sw}} \) is the average source water value RW is the ring width index.
C.4. RESULTS AND DISCUSSION

Growth ring widths average 0.44 mm (min = 0.07 mm, max = 1.49 mm, SD = 0.23, MS = 0.31, n=760). Figure 4 shows the complete ring-width chronology of all ten crossdated specimens. All specimens were de-trended to remove growth trends and the resultant indices were averaged.

Figure C4. Ring-width chronology for ten specimens from Bylot Island. The Ring number scale axis from pith of the oldest specimen. Grey line indicates the number of trees in each part of the floating chronology.

In modern tree-ring studies from boreal forests, summer temperature is inferred to be the dominant factor affecting tree growth (Hughes et al., 1999). Temperature reconstructions derived from ring widths are based on transfer functions applied to indices, thus a tree-ring index-based reconstruction using a modern analogue site would not be useful for reconstructing the temperatures of Plio-Pleistocene Bylot Island (Gagen et al., 2007). The Ballantyne et al. (2006) transfer function incorporates both ring-width data and oxygen isotope measurements. Because oxygen isotope measurements obtained
from tree rings are not indexed, transfer functions based on modern analogues can be used for temperature reconstructions of the Pliocene (Ballantyne et al., 2006; Richter et al., 2008; Ballantyne et al., 2010). Figure C5 presents the oxygen isotope record averaged over 4 trees. Average $\delta^{18}O$ ratio is 22.1 ‰ (min =20.1, max = 24.6, SD = 0.8, n=328) (Table C2).

Figure C5. Averaged $\delta^{18}O$ chronology for four specimens from Bylot Island. The Ring number axis counts from pith of the oldest specimen. Grey line indicates the number of trees in each part of the floating chronology with four being the maximum and one the minimum.

Table C2. Statistics on the four samples chosen for oxygen isotope analysis. Age is determined from the number of rings. Range is the range of $\delta^{18}O$ values, SD is the standard deviation of the $\delta^{18}O$ and MS is the mean sensitivity.

<table>
<thead>
<tr>
<th></th>
<th>BY 25</th>
<th>BY 24</th>
<th>BY 27</th>
<th>BY 33</th>
</tr>
</thead>
<tbody>
<tr>
<td>Age (yrs)</td>
<td>82</td>
<td>76</td>
<td>90</td>
<td>109</td>
</tr>
<tr>
<td>Avg. $\delta^{18}O$ (%)</td>
<td>22.0</td>
<td>22.2</td>
<td>22.0</td>
<td>21.9</td>
</tr>
<tr>
<td>Range (%)</td>
<td>18.5-24.7</td>
<td>20.0-24.6</td>
<td>19.2-24.6</td>
<td>20.4-23.2</td>
</tr>
<tr>
<td>SD</td>
<td>1.4</td>
<td>0.9</td>
<td>1.2</td>
<td>0.7</td>
</tr>
<tr>
<td>MS</td>
<td>2.0</td>
<td>1.0</td>
<td>2.1</td>
<td>0.4</td>
</tr>
</tbody>
</table>
Temperature reconstructions for the site were carried out using the mechanistic modeling equations of Roden et al. (2000) (Eq 1, 2) and Anderson et al. (2002) (Eq 3) and the Ballantyne et al. (2006) (Eq 6), Richter et al. (2008) (Eq 5) and Csank et al. (in review) (Eq 8) transfer functions. Stable isotope values of precipitation were determined using the relationship described in Equation 4 for use in the transfer functions (Eq 6, 8). Errors are propagated through each step with the standard errors associated with Equation 4 incorporated into the errors associated with each transfer function and mechanistic model.

C.4.1 Meteoric water

The average stable isotope value of meteoric water determined from the $\delta^{18}$O$_{\text{cellulose}}$ value of tree rings using Equation 4 was $-18.2 \pm 1.6 \text{‰ VSMOW}$ and ranged from $-21.5 \pm 1.0$ to $-14.1 \pm 1.0 \text{‰}$ (Fig C6).

Figure C6. The $\delta^{18}$O$_{\text{ppt}}$ value of meteoric water as calculated by Equation 4 (○), the Roden et al. (2000) mechanistic model (■) and the Anderson et al. (2002) mechanistic model (▲). Error bars indicate the range of values represented by the annually resolved records. Grey box is the mean with standard deviation of all three techniques.
The average source water $\delta^{18}O_{sw}$ values calculated here ($\delta^{18}O = -15.5 \pm 2.9 \%e$, average all techniques) are very similar to modern values for Bylot Island precipitation. Values of $\delta^{18}O_{ppt}$ estimated from the OIPC average $-20.8 \%e$ but range from summer (June-Aug) values of $-15.3 \%e$ to winter values of $-25.2 \%e$. Average $\delta^{18}O_{ppt}$ values measured at the closest Global Network of Isotopes in Precipitation (GNIP) stations located at Resolute Bay, Nunavut are $-20.0 \%e$ and at Hall Beach, Nunavut are $-17.5 \%e$ (IAEA/WMO, 2006). These are well within the range of values of $-14.1$ to $21.5 \%e$ calculated using Equation 4.

Interannual variability in the Plio-Pleistocene isotope series could reflect either local, regional or global changes in the hydrologic cycle, which in turn could reflect differences in the temperature of precipitation, in vapour transport-related Rayleigh fractionation, and possibly in the season of precipitation (i.e., the ratio of summer to winter precipitation).

Recent isotope-enabled modeling studies of Eocene isotopes in precipitation have indicated that when the latitudinal temperature gradient is reduced, there is a corresponding reduction in the latitudinal gradient of isotopes in precipitation (Speelman et al., 2010). Considering that the latitudinal temperature gradient was reduced during the Pliocene (Ballantyne et al., 2010) and that ice volume was also reduced during the Pliocene (Lear et al., 2000), the lack of significant differences between our estimates of $\delta^{18}O$ of Pliocene Arctic meteoric water and that of modern suggests that factors other than the latitudinal temperature gradient could be driving latitudinal isotopic variability.
C.4.2 Temperature

Temperatures calculated using all methods resulted in an estimate of MAT of $-2.9 \pm 3.9^\circ C$ (Fig C7). Average MAT calculated using the Roden et al. (2000) model was $-3.4 \pm 7.1^\circ C$ with range of $-11.6$ to $6.9^\circ C$ over 106 years. Values calculated using the Anderson et al. (2002) equation was $1.7 \pm 5.1^\circ C$ ($-1.2$ to $5.3^\circ C$ over 106 years). Errors for both mechanistic models are based on multiple runs varying the humidity values. Temperatures determined using the Ballantyne et al. (2006) transfer function was $-2.8 \pm 1.7^\circ C$ with a range of $-7.9$ to $2.3^\circ C$ over 106 years. However, as the Ballantyne et al. (2006) transfer function uses an inferred value for precipitation determined from a method similar to the one we employed in Equation 4, we adjusted the transfer function by using values of precipitation isotopes determined by Equation 4 rather than those determined from Equation 7. This resulted in a warmer average temperature calculation of $-1.1 \pm 1.6^\circ C$ and a much smaller range of temperatures $-3.1$ to $1.7^\circ C$; this $5^\circ C$ range in MAT is much closer to the realistic range of MAT in a boreal forest ($5.5^\circ C$) than the 10 degree range determined using the unmodified Ballantyne et al. (2006) transfer function. Temperatures inferred from the Richter et al. (2008b) transfer function averaged $-9.1^\circ C$ ($-12.2$ to $-4.4^\circ C$). All MAT reconstructions were done for the entire floating chronology providing an annually resolved record of MAT (Fig C8).
Figure C7. MAT reconstructions as calculated by the Roden et al. (2000) mechanistic model (●), the Anderson et al. (2002) mechanistic model (○), the unmodified Ballantyne et al. (2006) transfer function (■), the Ballantyne et al. (2006) model with source waters inferred from Equation 4 (□), the Richter et al. (2008b) transfer function (▲). Error bars represent the range of values represented by the annually resolved records. Grey box is the mean temperature with the standard deviation inferred from all techniques.
Figure C8. MAT estimates for our Bylot Island chronology. Green shaded area represents the range of temperatures calculated by the Roden et al. (2000) mechanistic model with humidity ranging from 60-80%. Blue shaded area is the range of temperatures calculated by the Anderson et al. (2002) mechanistic model based on varying humidity. Green-blue area is where the two mechanistic models overlap. Also shown are temperature estimates using the Ballantyne et al. (2006) transfer function (red) and the Richter et al. (2008) transfer function (brown). The darker red line represents temperature estimates determined using the Ballantyne et al. (2006) transfer function but with source water values computed using our Equation 4.

Using the transfer function developed by Csank et al. (in review) (Eq 8) we were able to calculate growing season (June-July) temperatures of 13.5 ± 1.1 °C, with a range of 11.2 to 16.8 °C. (Fig C9).
C9. JJ temperature estimates for our Bylot Island chronology. Shaded area indicates the standard deviation of ± 1.6°C determined by Csank et al. (in review) as the error of the transfer function.

Our estimates of MAT of -2.9 ± 3.9 °C are approximately 11.9 ± 4.5°C warmer than present Bylot Island MAT (-14.8 ± 2.2 °C). Our estimates of JJ temperature of 13.5 ± 1.1°C are approximately 12.6 ± 1.6°C warmer than present Bylot Island JJ temperature (4.2 ± 1.2°C). Cellulose δ18O values from wood collected from the early Pliocene (4-5 Ma) Beaver Pond site on Ellesmere Island was used to reconstruct mean annual temperature (MAT) (Ballantyne et al. 2006; 2010) and JJ temperature (Csank et al. in review). The tree-ring isotope-based temperature obtained by Ballantyne et al. (2006) was -5.5 ± 1.9°C, or 14.2 ± 2.5°C warmer than present. Ballantyne et al. (2010) conducted a proxy comparison study of temperature estimates for the Beaver Pond locality updating and expanding upon the Ballantyne et al. (2006) study. Using three independent proxies, a study of the coexisting vegetation at the site and its relationship to climate, stable isotope values of tree-ring cellulose and bacterial tetraethers, Ballantyne et
al. (2010) inferred MAT of $-0.6$ to $-0.4^\circ$ C, or $\sim 19^\circ$ C warmer than present. These inferred temperatures are considerably warmer than the temperatures calculated by previous studies and much warmer than the temperatures that we have determined for Bylot Island. However, our study is from a site that is much younger, by a minimum of 1 million years and therefore would have experienced a much cooler climate.

Warm month mean temperature (WMMT) determined for the Beaver Pond locality on Ellesmere Island by Ballantyne et al. (2010) was $14.4 \pm 2$° C. Csank et al. (2011) determined growing season temperatures for the Beaver Pond locality of $12.7 \pm 1.9$° C, however these were based on mollusc shells and so likely represent a longer growing season than June-July. Csank et al. (in review) using the transfer function in Equation 8 reconstructed JJ temperatures for the Beaver Pond locality of $15.2 \pm 1.3$°C.

Although the Beaver Pond locality is older, the temperatures calculated there are not much warmer than the temperatures determined from our Bylot Island site. Elias and Matthews (2002) reconstructed temperatures from 11 Plio-Pleistocene fossil sites in the North American Arctic ranging in age from 5.7 to 1.7 million years old. They used beetle fauna from each site in a mutual climate range (MCR) method to reconstruct maximum summer temperatures ($T_{\text{max}}$). Elias and Matthews (2002) determined that for all assemblages, estimates of $T_{\text{max}}$ ranged between $12.4$ and $13.8$° C regardless of age or location. This appears to be constant with our result of a JJ temperature for Bylot Island of $13.5 \pm 1.1$ °C. If we examine the recent temperature reconstructions from the high Arctic in conjunction with the temperature reconstructions presented here we obtain a similar result to Elias and Matthews (2002). Temperatures estimated for the late
Pliocene-early Pleistocene (2 to 3 Ma) are not substantially different from temperature estimates for the early Pliocene (4 to 5 Ma).

The onset of Northern Hemisphere glaciations began with stepwise cooling starting at 3.5 Ma and culminating with the onset of widespread Northern Hemisphere glaciation between 2.7 and 2.4 Ma, with first appearance of a Greenland ice sheet at ~3.3 Ma and expansion of the Greenland, Scandinavian and Arctic ice sheets from 2.75 Ma on (Kleiven et al., 2002; Matthiessen et al., 2009). Thus, Arctic climate had already entered a cooler period by the time the Bylot trees were growing. Our site is closest in age and similar stratigraphically to the Kap København deposits in Northeast Greenland (Funder et al., 2001). The Kap København deposits represent a forest of predominantly boreal vegetation consisting of larch, pine, spruce, and abundant mosses, which are inferred to have existed during an interglacial period between MIS 103 and 95 (2.4 to 2.7 Ma). The Kap København forest deposits are located stratigraphically above a till unit that represents a local glaciation. Temperatures inferred for the Kap København forest deposits based on an insect fauna and vegetation coexistence approach around -4° C MAT (10-15° C WMMT), are very similar to the temperatures we infer for Bylot Island (-2.9° C MAT; 13.5° C JJ temperature). By contrast the Hvitland Beds located in northern Ellesmere Island and dated to around 2.5 Ma (Fyles et al. 1998) represent the remains of marine units and contain pollen that indicates exclusively tundra vegetation with wood and tree pollen entirely absent. The Hvitland deposits were inferred by Fyles et al. (1998) to predate the Kap København deposits and so have inferred the Hvitland Beds to represent a cool stage of climate, likely one of the glacial stages between MIS
100 and MIS 96. That both our Bylot Island deposit and the København deposits represent the remains of forests that grew during interglacial phases within the Pliocene begs the question of whether other Pliocene terrestrial deposits also represent generally warmer phases and whether temperatures inferred from these deposits might be biased warm compared to the overall cooling trend of the Pliocene. Better age control of terrestrial units thus becomes necessary if we are to put these deposits into a global context.

The mid-Pliocene (3-3.5 million years ago) has been the subject of extensive modeling studies (Chandler et al., 1994; Sloan et al., 1996; Dowset et al., 1999; Haywood et al., 2004; 2009). The modeling communities are interested in the Pliocene as an accessible example of a future warmer world. Model results for the Arctic around Bylot Island indicate temperatures between 10 and 14° C warmer than present utilizing both the GCMAM3 and HadAM3 models (Haywood et al., 2009), both of which use Pliocene boundary conditions determined from PRISM2 (Dowset et al., 1999). Mean annual temperatures inferred for Bylot Island from model output of the HadAM3 Pliocene model are around -4° C, comparable to MAT inferred from our tree-ring isotope data (Haywood et al., 2009). However, because our forest unit most likely represents a snapshot taken of an interglacial period within the Plio-Pleistocene, it may not be reasonable to compare our data with inferences from a climate model unless the model is specifically set up to model interglacial climate. Because these models were developed to model the mid-Pliocene climate optimum (3-3.5 Ma) it is interesting that our Bylot temperatures from 2.4 to 2.8 Ma match with modeled mid-Pliocene temperatures. This suggests that
temperatures during warm marine oxygen isotope stages within the Plio-Pleistocene, such as that represented by the Bylot site, were comparable to temperatures during the mid-Pliocene before the onset of large-scale Northern Hemisphere glaciation. Perhaps a challenge for future modeling studies of the Pliocene is to develop models with orbital parameters tuned to glacial and interglacial phases.

C.5. CONCLUSIONS

The data we present provide important constraints on the magnitude of Arctic amplification of Pliocene warmth, and on changes in Arctic hydrology associated with a warmer climate. Our temperature estimates indicate that Bylot Island likely represented boreal tree-line vegetation during the Plio-Pleistocene. The Bylot Island forest deposits date to between 2.4 and 2.7 Ma, placing them during the onset of Northern Hemisphere glaciation. Our mean annual temperatures on the order of $-2.9 \pm 3.9^\circ$ C and growing season (June-July) temperatures on the order of $13.5 \pm 1.1^\circ$ C suggest that the Bylot forests likely represent a warm interval possibly corresponding to an interglacial MIS between 103 and 95. The Bylot forests directly overlie diamictite representing local glaciation of Bylot Island, suggesting that Pliocene cooling was likely a step-wise event punctuated by numerous glacial advances and retreats prior to full-scale Northern Hemisphere glaciation and loss of all forest deposits from the high Arctic. The Pliocene is increasingly used as an analogue for the conditions that we may experience in the next century. Models of Pliocene climate are being constructed with an aim to improve model predictions of future climate (Jansen et al., 2007; Haywood et al., 2009). However if most
terrestrial deposits from the Arctic are biased towards interglacial warm phases and most marine deposits, as suggested by Matthiessen et al. (2009), are biased towards glacial cool phases when sedimentation rates were higher, this could indicate a disconnect in our ability to provide accurate boundary conditions to form the basis of modeling studies. More work is needed to better constrain the dating of terrestrial Arctic deposits in order to allow for a more accurate comparison with marine records to allow us to address this discrepancy. Our study provides estimates of the inter-annual variability of Arctic temperatures and water δ18O for an interglacial phase in the Plio-Pleistocene, and the temperatures we estimate for Bylot Island provide some constraints on the magnitude of warming that might eventually occur in the Arctic if atmospheric CO2 levels were stabilized at near present levels.

C.6. REFERENCES


Harris, S.A., 2005. Thermal history of the Arctic Ocean environs adjacent to North America during the last 3.5 Ma and a possible mechanism for the cause of the cold events (major glaciations and permafrost events). Progress in Physical Geography 29, 218-238.


