R.H. Smittenberg (Referee)

The paper presents a detailed investigation of a sedimentary deposit from the Pliocene period, located in the Canadian Arctic. The results give insight in the past environment and climate of this high arctic region from a geologic era with atmospheric CO2 levels comparable to that of today (as indicated by earlier studies), and is thus relevant for our understanding how the long-term climate may develop in a high CO2 world, with focus on the high Arctic region. The majority of the paper is well written, however some parts are not. My main concern lies in the section about the atmospheric CO2 reconstruction. Although I do agree that the basic concept that higher CO2 availability for plants could, in principle, lead to a stronger fractionation against 13C, and thus that the 13C content of fossil plants could possibly be used to reconstruct past levels of atm. CO2 (in essence using the same approach Pagani et al took in using the 13C content of specific algal lipids, long-chain alkenones), I find that there are some major flaws in their execution.

RE: We have addressed Dr. Smittenberg's concerns that have lead us to revisit our analysis and a major revision of our manuscript.

Below is a more detailed list of comments.

Abstract: there are many issues with the English style and exact and careful phrasing. For instance, one needs to assume that CO2 concentrations are in the atmosphere. In line 24: isotope ratios of 440 ppm? Line 30: 'furthest northern evidence' (northern- most?) .

RE: Changes have been made to the text to clarify these ambiguities.

p2line6. No newer references?

RE: A more recent paper, Francis and Skific 2015, has been added.

P2l16-18. Revise / make clear and expand what the relevance is of the 100k vs 41K orbital cycles, give references. Or leave out.

RE: This sentence has been removed.

P2l30. In the rest of the paper it does not become very clear how fire has a large impact as climate amplifier. What is a 'proximal mechanism'?

RE The term 'proximal' has been removed, and it has been clarified that this hypothesis was a motivator for the study rather than an outcome.

P3: Generally written in a very sloppy manner.

P3l1-12. Confusing piece mixing up sea ice conditions, industrial black carbon and natural (counteracting) effects. How could one have observed temperatures in the Pliocene? In other words: revise.

RE: Although this article was motivated by the question of fire's role in Arctic amplification, it does not directly address it within this paper, and so this paragraph discussing the interactions of fire on climate has been removed.

P3l16. Check writing P3l20-25: Particularly badly written. To what does 'This' exactly refer to (l21)? Dating uncertainties suggest an additional hypothesis? Can proxies be 'deposited'?

RE: This paragraph has been edited to improve clarity

P4l10/11. Entirely unclear: 'spanned the 1 m remaining of Unit II as per Mitchell et al' P4l17. 'samples of these 2006'-?

RE: These sections have been edited to improve clarity

P4l32. 'Approximately' 200.00 mg Be (and thus not 200.01)?? same for 150.00 g quartz?)

RE: This has now been changed to approximately 200 and approximately 150.

Section 2.3 and 3.2 (P5l12. Carbon isotopic discrimination) The authors start using their equation 1 (taken from Farquhar 1989), derived for C3 plants with stomata, to describe fractionation against 13C by bryophytes. This is fine, they start in the same way as Fletcher et al (2008), as astomatous plants like bryophytes do (isotopically) behave fairly similarly. However, instead of taking the well developed model used by Fletcher et al (the basic concept and many tests described in GCA vol 70, p5676; but also see Fletcher 2005 in GlobBiogeochemCycles) they try to re-invent the wheel – however a very crooked one. The substitution of eq 2 into eq 1 is fine as long as one wants to back-correct for (paleo)height once a p(a) has been estimated from any transfer function of Delta13C to pCO2. However one cannot simply substitute p(i)/p(a) (a ratio between 0-1) simply by p(a), it totally changes the equation/model and even units. Moreover, I really do not see any reason or advantage of using the natural log of Delta13C instead of Delta13 - unless one wants to focus on the height term of eq. 2. Set their overly simplified and actually wrong theoretical exercise aside, the authors then compare Delta13C with pCO2 from a range of altitudes (not taking into account also lower pO2 levels) to arrive at some empirical relation between these two. As they write in their discussion, there are many confounding factors that could have influenced the observed C isotope fractionation - indeed resulting at different slopes for each site. Choosing a simple polynomial fit through this data has no theoretical basis at all, and is highly biased by the few Andean results and the Swiss sites. The majority of their plot comes from a Polish site – however in that original article the primary cause for the 13C discrimination was thought to be temperature, not altitude, although these two factors do co-vary. In the discussion (section 4.2) the authors are reasonably cautious about their model, however in my opinion the their framework is in any case ready for the trash bin and should get removed from the paper. I really wonder why the authors have not taken the model and results of Fletcher (2005, 2006, 2008), which does have a solid theoretical base ground in isotope systematics but also plant physiology. The first thing the authors need to do is discuss their results within the framework and transfer functions from Fletcher. Once they do that , I am skeptical if their data is not too compromised many environmental factors like temperature or humidity, but this remains to be seen. Also note that the framework of Fletcher only appears to work with reasonable (un)certainty on a larger amplitude of pCO2 between 300-2500 ppm. I don't think that the

confounding factors give 'subtle differences' (p9l31). In the end, interpreting the 13C values from bryophytes from one single location appears to be a very uncertain enterprise. The authors need to provide a solid error assessment, at the moment this is highly under-developed.

RE: We agree that reconstructing past CO_2 levels, considering all the underlying assumptions and sources of error, is no trivial task. In fact, our first attempts to reconstruct past CO_2 levels using $\Delta^{13}C$ of bryophytes, were based on the original approach pioneered by White et al. (1994). However, we found that many of these previously proposed theoretical models that have been since refined in the BRYOCARB model of Fletcher et al. (Fletcher et al. 2008) had far too many physiological variables that were poorly constrained when dealing with paleoenvironments and thus had too many degrees of freedom to provide reliable estimates of past CO_2. Dr. Fletcher actually provided us with his BRYOCARB model and we also found it difficult to produce reasonable estimates of Pliocene CO_2 concentrations based on the number of tunable parameters, which is why we opted to derive our own independent empirical transfer function based on the increase in $\Delta^{13}C$ of bryophytes as a function of changes in the partial pressure of CO_2 with elevation. Our empirical approach is based on the same physical principles and underlying assumptions as the BRYOCARB model: 1.) plants that lack stomates have no mechanism to actively regulate their CO_2 gradient and thus are sensitive to the partial pressure of CO_2 in their environment. 2.) as the partial pressure of CO_2 increases, more CO_2 is driven into the moss through diffusion and a greater pool of CO_2 is available for discrimination by rubisco (Farquhar, Ehleringer and Hubick 1989). Based on the comments from Reviewer #1 and Reviewer #2, we have revisited the BRYOCARB model to better evaluate our results.

One test is to see how well the BRYOCARB model performs at predicting the observed $\Delta^{13}C$ of modern mosses across the gradient in partial pressure of atmospheric CO_2. We iteratively optimized tunable parameters in the BRYOCARB model to predict the overall mean value of observed $\Delta^{13}C$ values 19.6 ± 2.1sd. It is clear that both BRYOCARB and our empirical model show the expected increase in $\Delta^{13}C$ with pCO_2; however, the slopes of these relationships are markedly different (*Fig. C1*). There are clearly processes affecting the Moss $\Delta^{13}C$ values that are not necessarily captured by the BRYOCARB model and are contributing to the large error envelope in our empirical model. The BRYOCARB model does not appear to be sensitive enough over this range of atmospheric CO_2 partial pressures to simulate the response in the observed $\Delta^{13}C$ moss values. Furthermore the RMSE from our empirical model is 1.8 $\Delta^{13}C$, whereas the RMSE for the BRYOCARB model is 2.1 $\Delta^{13}C$.

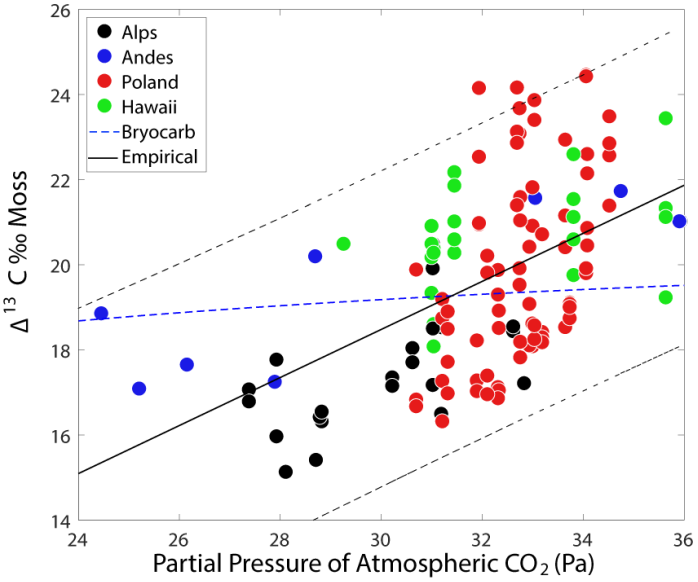

Figure C1. Predictions from BRYOCARB model (blue dashed line) and our empirical transfer function (black solid line with error envelope) compared with observed moss isotopic discrimination (points). Moss $\Delta^{13}C$ observations are from elevational transects in the Swiss Alps, Peruvian Andes, Poland, and Hawaii.

As mentioned by Dr. Smittenberg it has been previously noted that temperature may explain the variability in $\Delta^{13}C$ with elevation (Skrzypek et al. 2007). To test this explanation, we evaluated to what extent temperature could affect the isotopic discrimination across the elevation transects of observed moss $\Delta^{13}C$ and ultimately the impact on our Pliocene CO_2 estimates. Using the BRYOCARB model, with all other variables held constant, we determined that some of the variance in $\Delta^{13}C$ that is not described by partial pressure in atmospheric CO_2 (Fig. C1), may be explained by temperature (Fig. C2). In fact, assuming a moist adiabatic lapse rate 5 deg C/ km for the published data from the Peruvian Andes, we see that the data span ~2700m in elevation correspond to a 13.5 °C range in temperature. According to BRYOCARB this can result in a 1.0 ‰ enrichment in moss $\Delta^{13}C$ as temperature decreases with elevation (Fig. C2). Unfortunately, this temperature effect is in the ***opposite direction*** as the observed depletion in moss $\Delta^{13}C$ with elevation (Fig. C1), indicating that decreased temperatures at higher elevations should lead to increased discrimination. According to our mean summer temperature reconstruction from tetraethers (14.1 to 17.5 °C), temperature variations over the Pliocene could result in a 0.2 ‰ response in $\Delta^{13}C$ and thus a fairly negligible effect on our empirical estimates of Pliocene CO_2 concentrations.

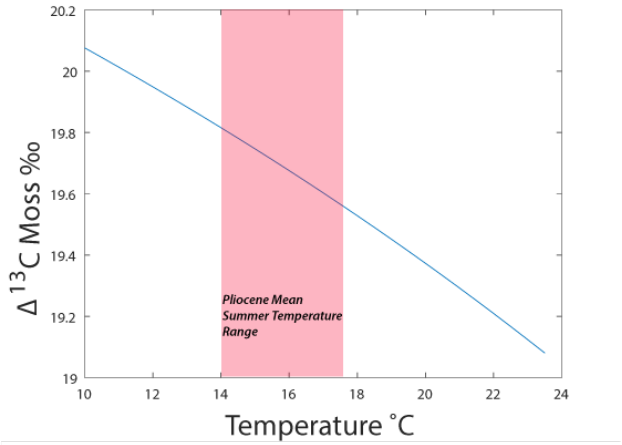

Figure C2. Changes in moss $\Delta^{13}C$ as a function of temperature. Where the entire range of temperatures represents the lapse rate of moss samples from the Peruvian Andes and the red box represents the mean summer temperature range for the Arctic derived from our tetraether measurements.

It was also suggested by Dr. Smittenberg that moss $\Delta^{13}C$ may vary as a function of decreasing partial pressure of O_2 in the atmosphere with elevation. We also tested this explanation, by revising the original BRYOCARB model so that it was formulated in terms of pO_2 instead of O_2 concentration, while holding all other variables constant. We found that $\Delta^{13}C$ does indeed change as a function of elevation; however, the relationship is ***negative*** (Fig. C3), in contrast to the positive relationship in response to pCO_2 (Fig C1). Furthermore, there is no evidence that the relatively high concentration of O_2 21% was significantly different during the Pliocene. Therefore we do not suspect that changes in pO_2 have an appreciable effect on our pCO_2 reconstructions because they do not explain the observed increase in moss $\Delta^{13}C$ with elevation and probably did not change significantly at our site during the Pliocene.

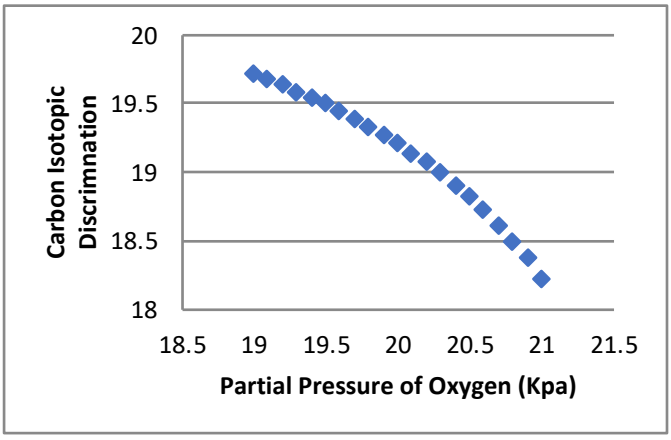

Figure C3. Changes in moss $\Delta^{13}C$ only as a function of pO_2 showing an increase in $\Delta^{13}C$ with elevation (i.e. reduced pO_2).

We have included equations 1 and 2 simply to demonstrate the physical relationship between isotopic discrimination and the partial pressure of atmospheric CO_2. Dr. Smittenberg is correct in that we cannot simply replace pi/pa with pa and substitute equation 2 into equation 1, we have revised the text to reflect that these equations simply provide the physical basis used to derive an empirical transfer function to predict atmospheric CO_2 (P6 L13-15).

We have simplified our empirical transfer function, where we now use a linear function to predict $\Delta^{13}C$ moss from the partial pressure of atmospheric CO_2 across an elevational gradient (Fig. C1). We also compare our empirical transfer function with the theoretical predictions from the BRYOCARB model.

Lastly, we do not understand what the reviewer is implying by stating that our 'data is too compromised' by environmental factors to provide reasonable estimates of past CO_2 concentrations. It is not clear whether they are referring to our measurements, or the data that we have compiled from the literature. Our measurements are no more compromised than the previous studies that have used ^{13}C of mosses to estimate concentrations of atmospheric CO_2 in the past (White et al. 1994, Fletcher et al. 2008). We agree with Smittenberg that the assumptions made in estimating past CO_2 levels based on moss ^{13}C measurements contribute considerable error; however, the assumptions regarding how the diffusion of CO_2 into the organism is regulated by the partial pressure of CO_2 in its environment is the basis for many commonly used proxies of past CO_2, including isotopic measurements of alkenones (Pagani et al. 2010).

About the assessment of the Pliocene 13CO2 value using buckbean 13C values: this appears to be a reasonable approach, although it would be good to get this estimate confirmed by measuring some more plants. Note Stuiver&Braziunas (1989, Nature328, p58 Tree cellulose 13C/12C isotope ratios and climatic change) who observe a relation between latitude and fractionation (likely cause by changes in Temp and humidity).

RE: In order to estimate past CO_2 concentrations from isotopic discrimination in mosses, we must know the isotopic composition of source CO_2 against which mosses are discriminating. In order to do this we used the original approach of White et al. (1994), such that identifiable C3 paleo-vegetation was analyzed for its isotopic composition and compared to the isotopic composition of modern plants from the same taxonomic group. Although, we agree that we only used 4 modern buck bean (*Menyanthes trifoliata*) samples from different locations across the boreal forest to assess the stomatal sensitivity (i.e. pi/pa) to different environments, we found very little variability across these values of only 0.4 ‰. While Stuiver et al. analyze carbon isotopes in coniferous trees with a global distribution, the distribution of modern buckbean is mainly restricted to the boreal forest, so we don't suspect that the large climatic gradients of temperature and humidity identified as factors affecting conifer $\Delta^{13}C$ would have the same impact on buckbean. Furthermore, our estimates of atmospheric $\delta^{13}CO_2$ during the Pliocene are statistically indistinguishable from estimates derived from planktonic foraminifera (Ravelo et al. 2004).

About the measurements of the plant 13C values: why was chosen to measure cellulose instead of bulk tissue? What is the expected difference in 13C, knowing that sugars typically have less depleted 13C values than the bulk (lipids being more de- pleted)? The description of the preparation and isotope measurement methods is very limited and should get expanded.

RE: Essentially, because these samples were embedded in peat, they were quite dirty and we wanted to ensure that they were clean and that we were performing our isotopic measurements on the same

molecules across all samples. More details have been added regarding the rational for sample preparation and hollocellulose extraction, as well as the reference providing details on the method (P5 L2).

Section 2.4 The sections about the brGDGTs as well as the one about fire, vegetation and climate are well written, and extensive and critical enough, and I have no real comments here. However, to make the jump from the local fire frequency at one location to the notion that fire could have been a global climate feedback mechanism during the Pliocene is a very large jump to conclusions (p19l12-16). It is fine to mention this possibility, but I would not use the word 'reveal' (line 12) but use a more careful wording (e.g. indicates, suggests).

RE: The wording here has been changed as suggested

P6l32. How 'well' are the brGDGTs really preserved?

RE: GDGTs are well preserved in anoxic sediments that have not experienced burial in the sub-surface (see Schouten et al., 2013, Org. Geochem. 54, 19-61 for a review). This applies for the Pliocene Arctic deposits investigated here.

P7l1 That the brGDGTs are 'thought to be sourced by a wide array of acidobacteria within the soil' is still under in- vestigation and there is still only scant evidence. For one, brGDGTs are also produced aquatically. Rephrase.

RE: The sentence has been rephrased.

P7l23. How was a concentration of 10 mg ml-1 (of brGDGTs, if one reads the text) made? Concentration of Total lipid extract or polar fraction?

RE: Based on the weight of the polar fraction. This is now clear from the modified sentence.

P7l25 mass spectrometry. UHPLC or HPLC?.

RE: This is now explained.

P7l28. From where does the transfer function error come?

RE: This sentence has been eliminated because the transfer functions are only introduced later. In the remainder of the text we do not discuss the errors in the transfer functions; we only provide the standard deviations of the estimated temperature from our sample set

P8l1 minus term missing

RE: This has been corrected.

3. Results P9l11. Not clear what the maximum probability of age of 4.5 Ma means, when earlier the most likely age is estimated at 3.9 Ma?

RE: The 3.9 MA is simply the unweighted mean of the 4 samples. The most probably age based on the convolution of all of their individual probability distribution functions is 4.5. In the next couple of sentences we explain that, because the PDF are not complex owing to reaching the radiodecay-based saturation of the isotopes, we recommend using the simple mean. This section has been reworded for clarity.

P9l19. Is an error of 104 to 105 years relevant on the geologic timescale of millions of years?

RE:104 and 105 years should have been 10^4 and 10^5 years. The superscripting is corrected at the earlier mention (Page 4), and this section has been rewritten for clarity.

Conclusions: Depending on any revision of the paper, alter or remove mention to past CO2 levels. Also be more careful in the conclusions with respect to the role of fire on climate. The paper showcases well that fire was part of the arctic climate – however this is not so different from the present day boreal realm, and the paper does not at all investigate, model or discuss this aspect. The same is true for the last sentence about present day arctic climate change, the paper does not focus at all on the present day arctic.

RE: In the revised manuscript we have provided two independent reconstructions of past CO_2 levels- one from the theoretical BRYOCARB model and the other from our empirical relationship. Both of these approaches have their assumptions and biases, so including both estimates provides a full range of potential CO_2 concentrations during the Pliocene.

The ramifications of fire for the radiative budget, and potential for inclusion in future models has been reduced in the conclusion. Extension to present-day climate-change has been removed.

Figure 3: No references given for the Polish and Hawaiian sites. Fig. 4 I would also plot the originally measured (estimated) Delta13C values,not only reconstructed pCO2 (but as stated above, I find that this aspect of the paper needs an overhaul in any case).

RE: Figure 3 has been revised and the references have been added.

Elevated atmospheric CO_2 and fire linked to arctic amplification of temperature during the Early to mid-Pliocene

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Abstract. The mid-Pliocene is a valuable time interval for understanding the mechanisms that determine equilibrium climate at current atmospheric CO_2 concentrations. One intriguing, but not fully understood, feature of the early to mid-Pliocene climate is the amplified arctic temperature response. Current models underestimate the degree of warming in the Pliocene Arctic and validation of proposed feedbacks is limited by scarce terrestrial records of climate and environment, as well as discrepancies in current CO_2 proxy reconstructions. Here we reconstruct the CO_2, summer temperature and fire regime from a sub-fossil fen-peat deposit on west-central Ellesmere Island, Canada, that has been chronologically constrained using radionuclide dating to 3.9 +1.5/-0.5 Ma. An empirical transfer function was derived and applied to carbon isotopic measurements of paleo mosses to yield an estimate of Pliocene mean atmospheric CO_2 concentrations of 410 ± 50 ppm, which are slightly lower than theoretical model predictions of 510 ppm. The estimate for average mean summer temperature is 15.4±0.8°C using specific bacterial membrane lipids, i.e. branched glycerol dialkyl glycerol tetraethers. Macro-charcoal was present in all samples from this Pliocene section with notably higher charcoal concentration in the upper part of the sequence. This change in charcoal was synchronous with a change in vegetation that saw fire promoting taxa increase in abundance. Paleovegetation reconstructions are consistent with warm summer temperatures, relatively low summer precipitation and an incidence of fire comparable to fire adapted boreal forests of North America, or potentially central Siberia.

To our knowledge, this study represents the northern-most evidence of in Earth's history and highlights the important role of forest fire in the ecology and climatic processes of the Pliocene High Arctic. The results provide evidence that terrestrial fossil localities in the Pliocene High Arctic were probably formed during warm intervals that coincided with relatively high CO_2 concentrations that supported productive biotic communities. This study indicates that interactions between paleovegetation and paleoclimate were mediated by fire in the High Arctic during the Pliocene, even though CO_2 concentrations were similar to modern.

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1 Introduction

Current rates of warming in the Arctic are almost double the rate of global warming. Since 1850, global land surface temperatures have increased by approximately 1.0°C, whereas arctic surface land temperatures have increased by 2.0°C (Jones and Moberg, 2003; Pagani et al., 2010, Francis and Skific, 2015). Such arctic amplification of temperatures has also occurred during other warm climate anomalies in Earth's past. Paleoclimate records from the Arctic indicate that the change in arctic summer temperatures during past global warm periods was 3–4 times larger than global temperature change (Miller et al., 2010). While the latest ensemble of earth system models (ESMs) provide fairly accurate predictions of the modern amplification of arctic temperatures hitherto observed (Marshall et al., 2014), they often under-predict the amplification of arctic temperatures during past warm intervals in Earth's history, including the Eocene (33.9–56 Ma; Huber, 2008; Shellito et al., 2009), and the Pliocene (2.6–5.3 Ma; Dowsett et al., 2012; Salzmann et al., 2013) epochs. These differences suggest that either the models are not simulating the full array of feedback mechanisms properly for past climates, or that the full array of fast and slow feedback mechanisms have not fully engaged for the modern Arctic. If the later, the Arctic region has yet to reach the full amplification potential demonstrated in the past.

The Pliocene is an intriguing climatic interval that may offer important insights into climate feedbacks. Atmospheric CO_2 values varied (Royer et al., 2007) decreasing from values comparable to modern (Haywood et al., 2016; Pagani et al., 2010; Stap et al., 2016), to lower levels (Raymo et al., 2006); a state transition that may revert in the future under high CO_2. Of additional importance, continental configurations were similar to present (Dowsett et al., 2016). While global mean annual temperatures (MATs) during the Pliocene were only ~ 3°C warmer than present day (Fig. 1), arctic land surface MATs may have been as much as 15 to 20°C warmer (Ballantyne et al., 2010; Csank et al., 2011a; Csank et al., 2011b; Fletcher et al., 2017). Further, arctic sea surface temperatures may have been as much as 10 to 15°C warmer than modern (Robinson, 2009), and sea-levels were approximately 25m higher than present (Dowsett et al., 2016). As such, the terrestrial environment of the Arctic was significantly different, with tree line ecosystems at much higher latitudes nearly eliminating the tundra biome (Salzmann et al., 2008).

Several mechanisms have been proposed as drivers of arctic amplification, including vastly reduced sea-ice extent (Ballantyne et al., 2013), cloud and atmospheric water vapor effects (e.g. Feng et al., 2016; Swann et al., 2010), vegetation controls on albedo (Otto-Bliesner and Upchurch Jr, 1997), and increased meridional heat transport by the oceans (Dowsett et al., 1992) though it is now considered to be of lesser influence (Hwang et al., 2011). We propose that fire in arctic ecosystems may also be an important mechanism for amplifying arctic surface temperatures during the Pliocene, and so seek to understand its characteristics through quantification from the sediment record.

Although it is generally thought that atmospheric CO_2 concentrations of ~ 400 ppm provided the dominant global radiative forcing during the mid-Pliocene, CO_2 proxies over the Pliocene do not all agree (Fig. 1). Reconstructions of Pliocene CO_2 range between 190 and 440 ppm (Martinez-Boti et al., 2015; Seki et al., 2010). While CO_2 estimates from stomata and paleosols tend to be less precise, they are within the range of boron and alkenone derived estimates (Royer, 2006; Foster et al. 2017). Thus, there is no clear consensus on CO_2 concentrations in Earth's atmosphere during the Pliocene. Dating uncertainties are an additional confounding factor complicating site to site comparisons. Although modelled direct effects of this level of CO_2 variation may be small (Feng et al., 2017), reconstructing the

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CO_2 from the same deposits from which paleoclimate and paleoecological proxies are derived, may help reconcile previous estimates and contribute to constraining climate sensitivities during the Pliocene.

To advance our understanding of arctic amplification during past warm intervals in Earth's history such as the Pliocene this investigation targets an exceptionally well-preserved arctic sedimentary sequence to simultaneously reconstruct atmospheric CO_2, summer temperature, vegetation and fire from a single site.

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2 Methods

2.1 Site description

To investigate the environment and climate of the Pliocene Arctic we focused on the Beaver pond (BP) fossil site, located at 78° 33′ N (Fig. 2) on Ellesmere Island. The stratigraphic section located at ~380 meters above sea level (MASL) today includes unconsolidated bedded sands and gravels, and rich organic layers including a thick fossil rich peat layer, up to 2.4 m thick, with sticks gnawed by an extinct beaver (*Dipoides spp.*). The assemblage of fossil plants and animals at BP has been studied extensively to gain insight into the past climate and ecology of the Canadian High Arctic (Ballantyne et al., 2006; Csank et al., 2011a; Csank et al., 2011b; Fletcher et al., 2017; Mitchell et al., 2016; Rybczynski et al., 2013; Tedford and Harington, 2003; Wang et al., 2017). Previous paleoenvironmental evidence suggests the main peat unit is a rich fen deposit with a neutral to alkaline pH, associated with open water (Mitchell et al., 2016), likely a lake edge fen or shallow lake fen, within a larch-dominated forest-tundra environment (Matthews and Fyles, 2000), not a low pH peat-bog. While the larch species identified at the site, *Larix groenlandia*, is extinct (Matthews and Fyles, 2000), many other plant remains are Pliocene examples of taxa that are extant (Fletcher et al., 2017).

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The fen-peat unit examined in this study was sampled in 2006 and 2010. The main sequence examined across the methods used in this study includes material from Unit II, the entire span of Unit III, and material from Unit IV per Mitchell et al. (2016), with a total sampled profile of 1.65 m. Unit III has been estimated to represent ~20 000 years of deposition based on modern northern fen growth rates (Mitchell et al., 2016). The atmospheric CO_2 estimates from this locality were based on 22 sample layers from the 2006 field campaign, and the charcoal was based on 31, while the temperature estimates from specific bacterial membrane lipids were taken from 22 of the sample layers collected in 2006 and an additional 12 samples collected in 2010. The same samples from the 2006 season were analyzed for each of CO_2, mean summer temperature and char count where contents of the sample allowed. Pollen was tabulated from 10 samples from the 2006 sequence, located at different stratigraphic depths.

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2.2 Geochronology

While direct dating of the peat was not possible, we were able to establish a burial age for fluvial sediments deposited approximately 4–5 m above and 30 m to the southwest of the peat. We used a method based on the ratio of isotopes produced in quartz by secondary cosmic rays. The cosmogenic nuclide burial dating approach measures the ratio of cosmogenic ^{26}Al ($t_{½}$ = 0.71 Ma) and ^{10}Be ($t_{½}$ = 1.38 Ma) in quartz sand grains that were exposed on hillslopes and alluvium prior to final deposition at BP. Once the quartz grains are completely shielded from cosmic rays, the ratio of

the pair will predictably decrease because ^{26}Al has double the radiodecay rate of ^{10}Be. In 2008, four of the medium to coarse grained quartz samples were collected from a vertical profile of planar crossbedded fluvial sands between 8.7 and 10.4 m below the overlying till surface. The samples were 5 cm thick, separated by an average of 62 cm, and should closely date the peat (the sandy braided stream beds represent on the order of ~10^4 years from the top of the peat to the highest sample). Quartz concentrates were extracted from the arkosic sediment using Frantz magnetic separation, heavy liquids, and differential leaching with HF in ultrasonic baths. When sample aliquots reached aluminum concentrations <100 ppm (ICP-OES) as a proxy of feldspar abundance, the quartz concentrate was subjected to a series of HF digestion and rinsing steps to ensure that more than 30% of the quartz had been dissolved to remove meteoric ^{10}Be. Approximately 200 mg of Be extracted from a Homestake Gold Mine beryl-based carrier was added to 150 g of each quartz concentrate (no Al carrier was needed for these samples). Such large quartz masses were digested because of the uncertainty in the abundance of the faster decaying isotope. Following repeated perchloric-acid dry-downs to remove unreacted HF, pH-controlled precipitation, column chemistry ion chromatography to extract the Be and Al ions, precipitation in ultrapure ammonia gas, and calcination at temperatures above 1000°C in a Bunsen flame for three minutes, oxides were mixed with equal amounts of niobium and silver by volume. These were packed into stainless steel targets for measurement at Lawrence Livermore National Laboratory's accelerator mass spectrometer (AMS). Uncertainty estimates for ^{26}Al/^{10}Be were calculated as 1σ by combining AMS precision with geochemistry errors in quadrature. For a complete detailed description of TCN methods see Rybczynski et al. (2013). The ages provided here are updated from Rybczynski et al. (2013) by using more recent production rate information and considering the potential for increasing exposure to deeply penetrating muons during the natural post-burial exhumation at BP.

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2.3 Atmospheric CO_2 Reconstruction

In order to reconstruct atmospheric CO_2 concentrations during the Pliocene, we derived a method based on the different sensitivity of isotopic discrimination of plant groups to their environment (Farquhar et al., 1989; Fletcher et al., 2008; White et al., 1994). Specifically, we used measurements of stable carbon isotopic discrimination in C3 vegetation to approximate the carbon isotopic signature of the atmosphere, and measurements of carbon isotopic discrimination in bryophytes to estimate the partial pressure of atmospheric CO_2, which was then converted to atmospheric CO_2 concentration. According to theory (Farquhar et al., 1989), plants discriminate ($\Delta^{13}C$) against the heavier isotope in atmospheric CO_2, such that:

$$\Delta^{13}C = a + (b - a)\frac{p_i}{p_a} \quad (1)$$

where the fractionations of atmospheric CO_2 due to diffusion (a = ~ -4.4 ‰) and carboxylation by the enzyme rubisco (b = ~ -27 ‰) are constraints. Thus, isotopic fractionation in C3 plants ($\Delta^{13}C_{C3}$) is largely a function of stomatal control of partial pressure of intercellular CO_2 – (p_i) with respect to the partial pressure of atmospheric CO_2 (p_a). However, bryophytes lack stomata and thus a mechanism for actively regulating p_i, such that isotopic fractionation ($\Delta^{13}C_{bryo}$) varies mainly as a function of partial pressure in atmospheric CO_2 (i.e. p_a). While other environmental factors,

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such as humidity, temperature, light availability, and microclimate may also play important roles in isotopic discrimination in bryophytes (Fletcher et al., 2008; Ménot and Burns, 2001; Royles et al., 2014; Skrzypek et al., 2007; Waite and Sack, 2011; White et al., 1994), the first order control on discrimination is the partial pressure of atmospheric CO_2 (Fletcher et al., 2008; White et al., 1994). Because atmospheric CO_2 is relatively well mixed in the troposphere its mean annual concentration does not differ significantly by location. However, because total atmospheric pressure decreases with atmospheric height (h), the partial pressure of atmospheric CO_2 must also decrease according to the following exponential function:

$$p_{a(h)} = p_{a(i)} e^{-h/H} \quad (2)$$

such that the partial pressure of atmospheric CO_2 at any given height in the atmosphere ($p_{a(h)}$) can be calculated based on the initial atmospheric partial pressure of atmospheric CO_2 ($p_{a(i)}$) and a reference height (H = 7600 m), where atmospheric pressure goes to 0.37 Pa (Bonan, 2015). Therefore assuming that carbon isotopic discrimination in bryophytes varies in response to the partial pressure of atmospheric CO_2 we can predict from basic physical principles an increase in $\Delta\ ^{13}C_{bryo}$ in response to an increase in $p_{a(h)}$. Furthermore, if the assumptions of this empirical relationship are valid, then this empirical relationship can in theory be used to predict the partial pressure of atmospheric CO_2 based on carbon isotopic measurements of bryophytes.

To test this prediction, we compiled data from four studies investigating carbon isotopic variability of different bryophytes, primarily mosses, along elevational transects at different locations. Based on the elevations and locations of moss samples, the atmospheric partial pressure of atmospheric CO_2 was estimated from ERA-interim reanalysis data of total atmospheric pressure (Dee et al., 2011) in conjunction with globally averaged atmospheric CO_2 concentrations (Global View-CO_2, 2013) from the years moss samples were collected. For our analysis we only included measurements of carbon isotopic variability in non-vascular mosses and all isotopic values were normalized to cellulose based on the empirical relationship reported by Ménot and Burns (2001). Carbon isotopic discrimination values for all plant material was calculated as:

$$\Delta^{13}C = (\delta^{13}C_{atm} - \delta^{13}C_{plant})/(1 + \delta^{13}C_{plant}/1000) \quad (3)$$

where $\delta^{13}C_{plant}$ represents the C isotopic composition of plant cellulose and $\delta^{13}C_{atm}$ represents the mean annual carbon isotopic composition of atmospheric CO_2 of the year when samples were collected (Global View-CO_2, 2013), or in the case of sub-fossil mosses, when the samples were growing. The response $\Delta\ ^{13}C$ to pCO_2 across elevational gradients in modern mosses was then used to calibrate the theoretical model BRYOCARB that has been developed to reconstruct past CO_2 levels based on measurements of ^{13}C in paleo bryophytes. Thus, our approach provides two independent estimates of Pliocene CO_2 concentrations – one empirically derived from our transfer function and the other predicted from the BRYOCARB model calibrated to modern mosses and constrained by our paleoclimate reconstructions.

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In order to derive estimates of atmospheric CO_2 concentrations during the Pliocene, isotopic composition of source CO_2 from the atmosphere (ie. $\delta^{13}C_{atm}$) was estimated during the Pliocene to solve for $\Delta^{13}C$ of mosses (Eq.(3)). This was accomplished by simultaneous measurements of $\delta^{13}C$ in the C3 plant buckbean (*Menyanthes trifoliata* L.) that was identified as a subfossil specimen at the BP site and was also collected from four different sites in the Canadian Boreal Forest. Although it has been demonstrated that $\Delta^{13}C$ of C3 plant material is sensitive to many factors, including mean annual precipitation and altitude (Diefendorf et al. 2010), there is less variability within biomes, so modern buckbean was sampled from within the Canadian Boreal biome that we suspect is very similar to the BP site based upon paleovegetation (Ballantyne et al. 2010, Fletcher et al. 2017). Measurements of $\delta^{13}C$ on modern buckbean were used to constrain estimates of p_i / p_a using modern estimates of $\delta^{13}C_{atm}$ from when the buckbeans were collected. This constrained modern value of p_i / p_a was then applied to our sub-fossil buckbean samples to estimate $\delta^{13}C_{atm}$ during the Pliocene. All plant and moss material were rinsed and placed in a sonicating bath with deionized water to remove any paleosoil from samples. The diagnostic material for mosses was leafy material, whereas buck bean was identified base on seeds. Therefore, to ensure that our isotopic measurements were made on similar compounds, cellulose was extracted from samples according to Leavitt and Danzer (1993). All carbon isotopic measurements were performed at University of Arizona's environmental isotope laboratory.

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2.4 Paleotemperature Reconstruction

Paleotemperature estimates were determined based on the distribution of fossilized, sedimentary membrane lipids known as branched glycerol dialkyl glycerol tetraethers (brGDGTs) that are well preserved in peat bogs, soils, and lakes (Powers et al., 2004; Weijers et al., 2007c). These unique lipids are thought to be synthesized by a wide array of Acidobacteria within the soil (Sinninghe Damsté et al., 2011; Sinninghe Damsté et al., 2014) and presumably other bacteria (Sinninghe Damsté et al., 2018) in soils and peat bogs but also in aquatic systems. Previously, it has been established that the degree of methyl branching (expressed in the methylation index of branched tetraethers; MBT) is correlated with mean annual air temperature (MAT), and the relative amount of cyclopentane moieties (expressed in the cyclization index of branched tetraethers; CBT) has been shown to correlate with both soil pH and mean annual air temperature (Weijers et al., 2007b). Because of the relationship of the distribution of these fossilized membrane lipids with these environmental parameters, it has been used for paleoclimate applications in different environments including coastal marine sediments (Bendle et al., 2010; Weijers et al., 2007a), peats (Ballantyne et al., 2010; Naafs et al., 2017), paleosols (Peterse et al., 2011; Zech et al., 2012), and lacustrine sediments (Loomis et al., 2012; Niemann et al., 2012; Pearson et al., 2011; Zink et al., 2010).

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Improved separation methods (Hopmans et al., 2016) have recently led to the separation and quantification of the 5- and 6-methyl brGDGT isomers that used to be treated as one since the 6-methyl isomers were co-eluting with the 5-methyl isomers (De Jonge et al., 2013). This has led to the definition of new indices and improved MAT calibrations based on the global soil (De Jonge et al., 2014), peat (Naafs et al., 2017), and African lake (Russell et al., 2018) datasets.

Sediment samples were freeze-dried and then ground and homogenized with a mortar and pestle. Next, using the Dionex™ accelerated solvent extraction (ASE), 0.5–1.0 g of sediment was extracted with the solvent mixture of

dichloromethane (DCM):methanol (9:1, v/v) at a temperature of 100°C and a pressure of 1500 psi (5 min each) with 60% flush and purge 60 s. The Caliper Turbovap®LV was utilized to concentrate the collected extract, which was then transferred using DCM and dried over anhydrous Na_2SO_4 before being concentrated again under a gentle stream of N_2 gas. To quantify the amount of GDGTs, 1 µg of an internal standard (C46 GDGT; Huguet et al., 2006) was added to the total lipid extract. Then, the total lipid extract was separated into three fractions using hexane:DCM (9:1, v:v) for the apolar fraction, hexane:DCM (1:1, v:v) for the ketone fraction and DCM:MeOH (1:1, v:v) for the polar fraction, using a column composed of Al_2O_3, which was activated for 2 h at 150°C. The polar fraction, which contained the GDGTs, was dried under a steady stream of N_2 gas and weighed before being then re-dissolved in hexane:isopropoanol (99:1, v:v) at a concentration of 10 mg ml^{-1} and subsequently passed through a 0.45 µm PTFE filter. Finally, the polar fractions were analyzed for GDGTs by ultra-high performance liquid chromatography – atmospheric pressure positive ion chemical ionization – mass spectrometry (UHPLC-APCI-MS) using the method described by Hopmans et al., (2016). The polar fractions of some samples were re-run on the UHPLC-APCI-MS multiple times and the average fractional abundances of the brGDGTs was determined.

For the calculation of brGDGT-based proxies, the brGDGTs are specified by the Roman numerals as indicated in Fig. S1. The 6-methyl brGDGTs are distinguished from the 5-methyl brGDGTs by a prime. The novel indices, including MBT'_{5Me} based on just the 5-methyl brGDGTs and the CBT′ that was used to calculate the pH (De Jonge et al., 2014):

$$MBT'_{5Me} = ([Ia] + [Ib] + [Ic]) / ([Ia] + [Ib] + [Ic] + [IIa] + [IIb] + [IIc] + [IIIa] + [IIIb] + [IIIc]) \quad (4)$$

$$CBT' = -^{10}\log[([Ic] + [IIa'] + [IIb'] + [IIc'] + [IIIa'] + [IIIb'] + [IIIc'])/([Ia] + [IIa] + [IIIa])] \quad (5)$$

The square brackets denote the fractional abundance of the brGDGT within the bracket relative to the total brGDGTs. Mean summer air temperature (MST) was determined using the distributions of aquatically produced brGDGTs in the lake calibration developed by Pearson et al. (2011). When this calibration is used the fractional abundances of IIa and IIa′ must be summed because these two isomers co-eluted under the chromatographic conditions used by Pearson et al. (2011):

$$MST\ (°C) = 20.9 + 98.1 \times [Ib] - 12 \times ([IIa] + [IIa']) - 20.5 \times [IIIa] \quad (6)$$

MAT and surface water pH were also calculated using a novel calibration created using sediments from East African lakes analysed with the novel chromatography method and based upon MBT'_{5Me} (Russell et al., 2018).

$$MAT = -1.2141 + 32.4223 * MBT'_{5Me} \quad (7)$$

$$Surface\ water\ pH = 8.95 + 2.65 * CBT' \quad (8)$$

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2.5 Vegetation and Fire Reconstruction

For charcoal, a total of thirty 2 cm^3 samples were taken at 5 cm intervals from depths from 300 and 301.45 MASL at the BP site, with an additional $2cm^{-3}$ sample collected at 301.65 MASL. All samples were deflocculated using sodium hexametaphosphate and passed through 500, 250 and 125 µm nested mesh sieves. The residual sample caught on each sieve was then collected in a gridded petri dish and examined using a stereomicroscope at 20-40X magnification to obtain charcoal concentration (fragments cm^{-3}). Charcoal area (mm^2 cm^{-3}) was measured for each sample using specialized imaging software from Scion Corporation. For a detailed description of methods see Brown and Power (2013).

Vegetation was reconstructed using pollen and spores (herein pollen) at selected elevations at an upper and lower elevation that corresponded with changes in charcoal. Samples were processed using standard approaches (Moore et al., 1991), whereby $1cm^3$ sediment subsamples were treated with 5% KOH to remove humic acids and break up the samples. Carbonates were dissolved using 10% HCl, whereas silicates and organics were removed by HF and acetolysis treatment, respectively. Pollen slides were made by homogenizing 35 µl of residue, measured using a single-channel pipette, with 15 µl of melted glycerin jelly. Slides were counted using a Leica DM4000 B LED compound microscope at 400–630x magnification. A reference collection and published keys (McAndrews et al., 1973; Moore et al., 1991) aided identification.

In addition to tabulating pollen and charcoal, a list of taxa derived from Beaver Pond was previously compiled in Fletcher et al. (2017). Extant species from this list were selected and their modern observations extracted from the Global Biodiversity Information Facility (GBIF.org, 2017). Observation data was grouped by 5° latitude 5° longitude grids cells, and the shared species count calculated using R (R Core Team, 2016). Modern fire frequency was mapped using the MODIS 6 Active Fire Product. The fire pixel detection count per day, within the same 5° latitude 5° longitude grids cells was counted over the ten years 2006–2015, and standardized by area of the cell. The modern climate maps were generated using data from WorldClim 1.4 (Hijmans et al., 2005). The values for the bioclimatic variables mean temperature of the warmest quarter (equivalent to mean summer air temperature; MST) and precipitation of the warmest quarter (summer precipitation) were also averaged by grid cell. The shared species count, climate values, and fire day detections were mapped to the northern polar stereographic projection in ArcMap 10.1.

3 Results

3.1 Geochronology

The burial dating results with $^{26}Al/^{10}Be$ in quartz sand at 10 m below modern depth provides four individual ages. From shallowest to deepest, the burial ages are 3.6 +1.5/-0.5 Ma, 3.9 +3.7/-0.5 Ma, 4.1 +5.8/-0.4 Ma, and 4.0 +1.5/-0.4 Ma (Table S3), with an unweighted mean age of 3.9 Ma. The convoluted probability distribution function yields a maximum probability age of 4.5 Ma. Unfortunately, the positive tails of the probability distribution functions of two of the samples exceeds the radiodecay saturation limit of the burial age. Therefore, their probability distributions do not reflect the actual age probabilities and uncertainty. Given the positive tail in the probability distribution functions, and the inability to convolve all samples, we recommend using the unweighted mean age, 3.9 Ma, with an uncertainty

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of +1.5/-0.5 Ma as indicated by the two samples with unsaturated limits. Despite the apparent upward younging of the individual burial ages, the 1σ-uncertainties overlap rendering the samples indistinguishable.

3.2 Atmospheric CO_2 Reconstruction

As expected, carbon isotopic discrimination in mosses shows a positive relationship with partial pressure of atmospheric CO_2 both in empirical observations and theoretical predictions (Fig. 3). However, a much greater change in $\Delta^{13}C_{moss}$ is observed in response to p_a than is predicted from the optimized BRYOCARB simulations. The empirical fit to the observed change in $\Delta^{13}C_{moss}$ in response to p_a is slightly better (RMSE = 1.8 ‰) than the theoretical prediction from the BRYOCARB model (RMSE = 2.1 ‰), but the slopes are quite different, with our empirical slope (0.56 ‰/p_a) an order of magnitude greater than the linear approximation of the BRYOCARB slope (0.07 ‰/p_a), suggesting that other non-linear processes and not just p_a may be affecting $\delta^{13}C_{moss}$ variability with elevation.

While there does appear to be a global relationship between p_a and $\Delta^{13}C$ of mosses, there are notable differences among sites. Moss $\Delta^{13}C$ values tended to be generally lower in the Swiss Alps (mean = 17.4 ‰) and higher in Hawaii (mean = 20.6 ‰) and the slope of the relationship between p_a and $\Delta^{13}C$ appears to vary across sites with the Andes having the smallest slope and Poland having a much greater slope. We used the BRYOCARB model to test the sensitivity of $\Delta^{13}C$ to other variables that change as a function of elevation (e.g. temperature and pO_2). According to our BRYOCARB simulations, with all other variables held constant decreased temperature with increased elevation should slow metabolic rates resulting in an increase in $\Delta^{13}C$ (Supplemental Figure S3), which directly contradicts observations (Fig. 3). Furthermore, the range of mean summer temperature estimates from the Pliocene BP site could only explain ~0.2 ‰ isotopic response in our moss samples. Similarly we evaluated the effect of just changing pO_2 in our BRYOCARB simulations and found a decrease in $\Delta^{13}C$ with increasing pO_2 that is opposite to the $\Delta^{13}C$ response of mosses to partial pressure across all elevational transects. We also evaluated model performance using a global standard atmospheric sea level pressure of 101.325 kPa, or site-specific atmospheric pressure estimates from ERA-interim reanalysis data. We found that the model using site specific atmospheric pressure estimates performed better at predicting $\Delta^{13}C_{moss}$ (RMSE = 1.096 ‰) than the model using global standard atmospheric sea level pressure (RMSE = 1.216 ‰). Therefore, it appears that partial pressure of atmospheric CO_2 is the primary physical mechanism explaining the global relationship between $\Delta^{13}C$ of mosses and elevation and that other factors, such as water availability that may be mediated by different lapse rates (Ménot and Burns, 2001; Royles et al., 2014; Skrzypek et al., 2007; Waite and Sack, 2011), may explain variability among sites. Thus, the optimal model characterizing the observed modern relationship between $\Delta^{13}C_{moss}$ and the p_a was:

$$13C\,\Delta 13C_{moss} = 0.56 \times pCO_2 + 1.55 \quad (9)$$

Based on our analysis of cellulose extracted from four different *Menyanthes* L. (i.e. buckbean) plants growing at four different locations in the modern boreal forest, we found $\Delta^{13}C$ of buckbean to be fairly constant 16 ± 0.4 ‰, yielding an estimate of p_i / p_a in modern buckbean of 0.51. Applying this modern of p_i / p_a to our $\delta^{13}C$ measurements from sub-fossil buckbean we obtained estimates of $\delta^{13}C_{atm}$ during the Pliocene of -6.23 ± 0.9 ‰. Using our empirical

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transfer function (Eq. 9) in combination with these estimates of $\delta^{13}C_{atm}$, we were able to approximate atmospheric CO_2 concentrations over the Pliocene interval captured at the BP site (Fig. 4). We estimated a mean atmospheric CO_2 concentration over this interval of 410 ± 50 ppm (mean ± transfer error and instrument error) with considerable variability between a minimum atmospheric CO_2 concentration of 296 ppm and a maximum atmospheric CO_2 concentration of 480 ppm. Predicted values of Pliocene CO_2 from the BRYOCARB model were slightly higher at 510 ppm, but the single standard deviation across all estimates was extremely high (967 ppm), suggesting that the BRYOCARB simulations are not significantly different from our empirical model estimates; however, the BRYOCARB model is too sensitive to our range of $\Delta^{13}C_{moss}$ estimates and thus not very precise.

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3.3 Paleotemperature Estimates

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3.3.1 Provenance of branched GDGTs

Previously, brGDGT derived MAT estimates (-0.6 ± 5.0 °C) from BP sediments were developed using the older chromatography methods that did not separate the 5- and 6- methyl brGDGTs, and a soil calibration (Ballantyne et al., 2010). In marine and lacustrine sediments, bacterial brGDGTs were thought to originate predominantly from continental soil erosion arriving in the sediments through terrestrial runoff, however, a number of more recent studies have indicated aquatically produced brGDGTs could be affecting the distribution of the sedimentary brGDGTs and thus the temperature estimates based upon them (Warden et al., 2016; Zell et al., 2013; Zhu et al., 2011). Since the discovery that sedimentary brGDGTs can have varying sources, different calibrations have been developed depending on the origin of the brGDGTs, i.e. soil calibration (De Jonge et al., 2014), peat calibration (Naafs et al., 2017) and aquatic calibrations (i.e. Foster et al., 2016; Pearson et al., 2011; Russell et al., 2018). Therefore, several studies have recommended that the potential sources of the sedimentary brGDGTs should be investigated before attempting to use brGDGTs for paleoclimate applications (De Jonge et al., 2015; Warden et al., 2016; Yang et al., 2013; Zell et al., 2013). In this study, we examine the distribution of brGDGTs in an attempt to determine their origin and consequently the most appropriate calibration to utilize in order to reconstruct temperatures from the BP sediments.

Branched GDGTs IIIa and IIIa′ on average had the highest fractional abundance of the brGDGTs detected in the BP sediments (see Fig. S1 for structures; Table S1). A previous study established that when plotted in a ternary diagram the fractional abundances of the tetra-, penta- and hexamethylated brGDGTs, soils lie within a distinct area (Sinninghe Damsté, 2016). To assess whether the brGDGTs in the BP deposit were predominantly derived from soils, we compared the fractional abundances of the tetra-, penta- and hexamethylated brGDGTs in the BP sediments to those from modern datasets in a ternary diagram (Fig. 6). Since the contribution of brGDGTs from either peat or aquatic production could affect the use of brGDGTs for paleoclimate application, in addition to comparing the samples to the global soil dataset (De Jonge et al., 2014), peat and lacustrine sediment samples were added into the ternary plot to help elucidate the provenance of brGDGTs in the BP sediments. According to Sinninghe Damsté (2016), it is imperative to only compare samples in a ternary diagram like this where all of the datasets were analyzed with the novel methods that separate the 5- and 6-methyl brGDGTs since the improved separation can result in an increased abundance of hexamethylated brGDGTs. Recently, samples from East African lake sediments were analyzed using these new methods (Russell et al., 2018) and so these samples were included in the ternary plot for comparison (Fig.

6). Although the lakes from the East African dataset are all from a tropical area, they vary widely in altitude and, thus, in MAT. We separated them into three categories by MAT (lakes >20°C, lakes between 10-20°C and lakes<10°C). By comparing all the samples in the ternary plot, it was evident that the BP samples plotted closest to the lacustrine sediment samples from regions in East Africa with a MAT <10°C, suggesting that the provenance of the majority of the brGDGTs from the BP sediments was not soil or peat but lacustrine aquatic production.

The average estimated surface water pH for the BP sediments (8.6±0.2) calculated using eq. (8), is within the 6 –9 range typical of lakes and rivers (Mattson, 1999). This value is near the upper limit of rich fens characterized by the presence of *S. scorpioides* (Kooijman and Westhoff, 1995; Kooijman and Paulissen, 2006) and is higher than what would be expected for peat-bog sediments that are acidic (pH 3–6; Clymo, 1964) and which constitute most of the peats studied by Naafs et al. (2017). A predominant origin from lake aquatic production is in keeping with previous interpretation of the paleoenvironment of the BP site, which was at least at times covered by water as evidenced by fresh water diatoms, fish remains and gnawed beaver sticks in the sediment (Mitchell et al., 2016).

3.3.2 Aquatic Temperature Transfer Function

Since there is evidence that the majority of the brGDGTs in the BP sediments are aquatically produced, an aquatic transfer function was used for reconstructing temperature. When we apply the African lake calibration (Eq. 7), the resulting estimated MAT for BP is 7.1 ± 1.0 °C. This value is high compared to other previously published estimates from varying proxies, which have estimated MAT in this region to be in the range of -5.5 to 0.8°C, (Ballantyne et al., 2010; Ballantyne et al., 2006; Csank et al., 2011a; Csank et al., 2011b; Fletcher et al., 2017). A concern when applying this calibration is that it is based on lakes from an equatorial region that does not experience substantial seasonality, whereas, the Pliocene Arctic BP site did experience substantial seasonality (Fletcher et al., 2017). Biological production (including brGDGT production) in BP was likely skewed towards summer and, therefore, summer temperature has a larger influence on the reconstructed MAT. Unfortunately, no global lake calibration set using individually quantified 5- and 6-methyl brGDGTs is yet available. Therefore, to calculate mean summer air temperatures (MST, Eq. 6) we applied the aquatic transfer function developed by Pearson et al. (2011) by combining the individual fractional abundances of the 5- and 6-methyl brGDGTs. The Pearson et al. (2011) calibration was based on a global suite of lake sediments including samples from the Arctic, thus covering a greater range of seasonal variability. The resulting average estimated mean summer temperature was 15.4 ± 0.8 °C, with temperatures ranging between 14.1 and 17.4 °C (Fig. 4). This is in good agreement with recent estimates based on Climate Reconstruction Analysis using Coexistence Likelihood Estimation (CRACLE; Fletcher et al., 2017) that concluded that MSTs at BP during the Pliocene were approximately 13 to 15°C.

3.4 Vegetation and Fire Reconstruction

All sediment samples from BP contained charcoal (Fig. 4), indicating the consistent prevalence of biomass burning in the High Arctic during this time period. However, counts were variable throughout the section, with the middle and lower sections (18 fragments cm^{-3}) containing less charcoal compared to the upper section upper section (710 fragments cm^{-3}). Overall, samples from BP contained on average 100.0 ± 165 fragments cm^{-3} (mean ± 1 σ), with

charcoal area averaging 12.3 ± 20.2 mm^2 cm^{-3}. The variability of charcoal within any given sample was relatively low with a 1 σ among charcoal area of approximately 2 mm^2 cm^{-3}.

The three sections analysed for pollen reveal variations in vegetation (Figs. 4 and 5). Near the bottom of the section (300.3-300.4 MASL), *Larix* (26%) and *Betula* (17%) were the dominant trees. *Alnus* (6%) and *Salix* (6%) together with ericaceous pollen (4%) were relatively high. In contrast, low numbers of *Picea* (3%), *Pinus* (3%) and fern spores were recorded. Additional wetland taxa like *Myrica* (5%) and Cyperaceae (6%) were also noted. Overall, the non-arboreal (23%) signal was well developed. Crumpled and/or ruptured inaperturate grains with surface sculpturing that varied from scabrate to verricate were noted in the assemblage (12%), but could not be definitely identified. It is possible that these grains represent *Populus*, Cupressaceae or additional Cyperaceae pollen. Between 301.15-301.25 MASL, *Larix* (38%) and *Betula* (21%) increased in abundance, followed by ferns (7%). Cyperaceae remained at similar levels (6%) whereas *Picea* and *Pinus* decreased to 2% and 1%, respectively. Unidentified inaperturate types collectively averaged 14%. *Larix* pollen (23%) remained abundant near the top of the section (301.35-301.45 MASL), whereas *Betula* (2%) decreased. *Picea* (16%) *Pinus* (6%) and ferns (23%) increased in abundance. Of the ferns, trilete spores and cf. *Botrychium* were most abundant, followed by cf. *Dryopteris*. Inaperturate unknowns (10%) were also observed. Other notables included Ericaceae (2%) and Cyperaceae (2%). While rare, Onagraceae grains were also observed (Fig. 5).

According to GBIF-based mapping exercise, the paleofloral assemblage at BP most closely resembles modern day vegetation found in northern North America, particularly on the eastern margin (e.g. New Hampshire, New Brunswick and Nova Scotia) and the western margin (Alaska, Washington, British Columbia, and Alberta; Fig. 7a), and central Fennoscandia. Of these areas, the western coast of northern North America and eastern coast of southern Sweden has the most similarity to the reconstructed BP climate in terms of MST (Fig. 7b) and summer precipitation (Fig. 7c).

While high counts of active fire days are common in the western part of the North American boreal forest, it is not as common in the eastern part of the North American boreal forest (Fig. 7d), likely due to the differences in the precipitation regime. There was also low fire counts in Fennoscandia likely due to historical severe fire suppression (Brown and Giesecke, 2014; Niklasson and Granström, 2004). Therefore, based on our reconstruction of the climate and ecology of the BP site, our results suggest that BP most closely resembled a boreal-type forest ecosystem shaped by fire, similar to those of Washington, British Columbia, Northwest Territories, Yukon and Alaska (but see Sect. 4.3).

4 DISCUSSION

4.1 Geochronology

The plant and animal fossil assemblages observed at BP suggest a depositional age between 3 and 5 Ma (Matthews Jr and Ovenden, 1990; Tedford and Harington, 2003). This biostratigraphic age was corroborated with an amino-acid racemization age (>2.4 ± 0.5 Ma) and Sr-correlation age (2.8–5.1 Ma) on shells (Brigham-Grette and Carter, 1992) in biostratigraphically correlated sediments on Meighen Island, situated 375 km to the west-north-west. The previously calculated burial age of 3.4 Ma for the BP site is a minimum age because no post-depositional production of ^{26}Al or

^{10}Be by muons was assumed. If the samples are considered to have been buried at only the current depth (ca. 10 m, see supplemental data) then the ages plot to the left and outside of the burial field, indicating that the burial depth was significantly deeper for most of the post-depositional history. The revised cosmogenic nuclide burial age is 3.9 +1.5/-0.5 Ma. It is the best interpretation of burial age data based on improved production rate systematics (e.g. Lifton et al., 2014), and more reasonable estimates of erosion rate and ice cover since the mid-Pliocene (see Table S4). As the stratigraphic position of the cosmogenic samples is very close to the BP peat layers, we interpret the age to represent the approximate time that the peat was deposited.

4.2 Pliocene atmospheric CO_2 levels

We have derived a transfer function that allows us to predict the partial pressure of atmospheric CO_2 in Earths' past based on carbon isotopic measurements in byrophytes. However, many of the studies included in our transfer function identify other mechanisms that may also influence carbon isotopic discrimination in bryophytes. Because these other mechanisms may violate the assumptions of applying this transfer function to the past or contribute error to our reconstructions of atmospheric CO_2 concentrations during the Pliocene, we discuss these mechanisms below.

It has been suggested that in the absence of stomatal regulation, that surface water may control the gradient in partial pressure (i.e. p_i / p_a) in bryophytes (White et al., 1994), due to the greater resistance to diffusion of CO_2 in water than in the atmosphere. For instance, Ménot and Burns (2001) found that most mosses growing along an elevational transect in Switzerland experienced discrimination with elevation in response to decreased partial pressure, except one species *Sphagnum cuspidatum* Ehrh. ex Hoffm., which grows almost exclusively in wet hollows. In a study of Hawaiian bryophytes Waite and Sack (2011) found consistent slopes of less isotopic discrimination with elevation in all species, however, species growing on young substrate showed significantly less isotopic discrimination. The most likely explanation is that lack of canopy cover on the older substrates lead to greater photosynthetic rates, which lead to reduced p_i. Lastly, decreased discrimination of mosses growing along an elevational transect in Poland (Skrzypek et al., 2007), was found to be highly correlated with temperature. Although temperature is the primary factor driving most metabolic reactions, it does not provide a physical mechanism explaining the relationship between elevation and isotopic discrimination in mosses. Skrzypek et al. (2007) found slightly different relationships between elevation and carbon isotopic discrimination in mosses growing on the windward versus leeward side of their elevational transects suggesting that changes in lapse rate may also play a factor. Collectively, these studies suggest that microclimatic factors may explain differences in isotopic discrimination of mosses within and among different sites possibly contributing to different intercepts for sites reported in Fig. 3, and that dry vs. moist lapse rates may also play a role in regulating the different slopes among sites. In fact, the greatest elevational range reported among sites was for the elevational transect in the Andes (320 to 3100 m), but this site did not experience the widest range in $\Delta^{13}C_{moss}$. This tropical transect had a very moist lapse rate resulting in the least change in atmospheric temperature and humidity with elevation. Nonetheless, by projecting these data as a function of partial pressure we provide a physical mechanism to explain variations in moss carbon isotopic values globally and we help reconcile the previously reported empirical relationships, such as elevation, temperature, and over-story, all of which tend to be covariates of decreasing partial pressure with elevation. While differences in microclimate and lapse rate are clearly important factors in regulating

$\Delta^{13}C_{moss}$, these factors contribute to the global error in our model for predicting p_a and ultimately to uncertainties in our estimates of atmospheric CO_2 concentrations during the Pliocene.

Our reconstructions of CO_2 concentration for this mid-Pliocene interval are within the range of previously reported CO_2 estimates, tending to agree with alkenone estimates from Pagani et al. (2010). This suggests that CO_2 concentrations during this warm Pliocene interval were above 400 ppm. In fact our mean Pliocene value (410 ± 50 ppm) is not statistically different from the alkenone based estimates (357 ± 47 ppm) previously reported by Pagani et al. (2010) and our theoretical predictions based on BRYOCARB calibrated to modern $\Delta^{13}C_{moss}$ values indicate CO_2 concentrations of approximately 510 ppm, albeit highly variable due to the sensitivity of the model simulations. Generally, our estimates showed sustained atmospheric CO_2 estimates of slightly higher than 400 ppm with only two anomalously low values (Fig. 4). These estimates could represent an actual reduction in atmospheric CO_2, or they might be artefacts of sampling or analysis. It should be noted that poor preservation and a possible shift in dominant moss species to *Drepanocladius spp*. was evident in samples corresponding to these two anomalously low CO_2 estimates. While one of these samples contained only 0.17 mg/C and a $\delta^{13}C$ value of -20.9 ‰, the other contained 0.88 mg/C and a $\delta^{13}C$ value of -25.0 ‰. Thus, it is conceivable that the sample corresponding to the atmospheric CO_2 estimate of 335 ppm, might be approaching our minimum detection limit and should be verified in subsequent studies.

It should also be noted that changes in growth rate due to phosphorus availability and biases in shell size are known to contribute uncertainty to alkenone-derived CO_2 concentration estimates (Seki et al., 2010). Similar assumptions may affect boron-derived estimates of CO_2 concentrations. For instance, a recent update on the global boron cycle estimates the mean residence time of boron to be ~ 1.5 Ma and suggests that boron isotopes may not be sensitive to ocean pH on timescales less than 1 Ma (Schlesinger and Vengosh, 2016). This may help explain the apparent lack of variability in boron isotope based CO_2 estimates during the Pliocene (Hönisch et al., 2009; Tripati et al., 2009); however, boron isotopes do seem to reproduce the CO_2 variability measured in ice cores over the Pleistocene (Hönisch et al., 2009). Overall, our estimates using two independent approaches suggest that Pliocene CO_2 concentrations during this interval ranged between 400 and 500 ppm are consistent with recent estimates derived from both alkenones and from boron isotopes (Martinez-Boti et al., 2015; Seki et al., 2010).

There are numerous assumptions based on known uncertainties in our CO_2 reconstruction approach. First of all, our empirically based approach requires some estimate of the isotopic ratio of atmospheric CO_2 during this time, which we derive from C3 vegetation (Fletcher et al., 2008; White et al., 1994). Here we estimate the isotopic composition of the atmosphere over the Pliocene to be $\delta^{13}C$ = -6.23 ± 0.9 ‰, which is within the range of values recorded over glacial-interglacial intervals in ice cores $\delta^{13}C$ = -6.2 to -7.0 ‰ (Bauska et al., 2016) and consistent with estimates derived from carbon isotope measurements of foraminifera (Ravelo et al., 2004). If we assume that the isotopic composition of atmospheric CO_2 was -8.2 ‰ during the Pliocene and similar to today due to greater transfer of lighter carbon from the terrestrial reservoir to the atmospheric reservoir, that would result in reduced $\Delta^{13}C_{moss}$ and decreases in our mean estimate of atmospheric CO_2 to approximately 390 ppm. This adjustment to our original estimate of $\delta^{13}C$ of atmospheric CO_2 would bring our atmospheric CO_2 estimate more in line with previous reconstructions, but is still within the range of error of our original estimate.

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Another critical assumption of our approach is that the total pressure of the atmosphere has not changed at the BP site since the Pliocene either through increased partial pressure of constituent gases or more likely through changes in elevation due to dynamic isostacy. The current elevation of the site is approximately 380 MASL with a summertime total atmospheric pressure of approximate 88.5 kPa. If we assume that the site was at 0 m during the Pliocene that would increase the total summertime atmospheric pressure to 93.9 kPa and would decrease our Pliocene CO_2 estimates to about 390 ppm. However, estimates of dynamic eustacy since the Pliocene from paleoshorelines at lower latitudes are between 5 and 20 m (Rovere et al., 2014), suggesting that our assumptions regarding elevation at the site probably have a negligible impact on our estimates of Pliocene atmospheric CO_2 concentrations, especially given the uncertainty of the proxy approach. Therefore, the assumptions to our approach in estimating past CO_2 may be leading to estimates that are biased slightly high relative to previous estimates. When these assumptions are considered, our estimates still suggest atmospheric CO_2 concentrations around 400 ppm or above during this Pliocene warm interval.

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4.3 Fire, vegetation, climate

Wildfire is a key driver of ecological processes in modern boreal forests (Flannigan et al., 2009; Ryan, 2002), and although historically rare, is becoming more frequent in the tundra in recent years (Mack et al., 2011). The modern increase in fire frequency is likely as a consequence of atmospheric CO_2 driven climate warming and feedbacks such as reduced sea ice extent (Hu et al., 2010), because the probability of fire is highest where temperature and moisture are conducive to growth and drying of fuels followed by conditions that favor ignition (Whitman et al., 2015). Young et al. (2017) confirmed the importance of summer warmth and moisture availability patterns in predicting fire across Alaska, highlighting a July temperature of ~13.5 °C as a key threshold for fire across Alaska.

The abundance of charcoal at BP demonstrates that climatic conditions were conducive for ignition and that sufficient biomass available for combustion existed across the landscape. brGDGTs derived temperature estimates suggest mean summer temperatures at BP exceeded the ~13.5 °C threshold (Young et al., 2017) that drastically increases the chance of wildfire. An increase in atmospheric convection has been simulated in response to diminished sea-ice during warmer intervals (Abbot and Tziperman, 2008), but this study did not confirm if this increase in atmospheric convection was sufficient to cause lightning ignitions. An alternative ignition source for combustion of biomass on Ellesmere Island during the Pliocene is coal seam fires, which have been documented to be burning at this time (Estrada et al., 2009). However, given the interaction of summer warmth and ignition by lightning within the same climate range as posited for BP, we consider lightning the most likely source of ignition for Pliocene fires in the High Arctic.

Fire return intervals cannot be calculated from the BP charcoal counts due to the absence of a satisfactory age-depth model and discontinuous sampling. As strong interactions are observed between fire regime and ecosystem assemblage in the boreal forest (Brown and Giesecke, 2014; Kasischke and Turetsky, 2006), and in response to climate, comparison with modern fire regimes for areas with shared species compositions and climates may inform a potential range of mean fire return interval (MFRI).

Matthews and Fyles (2000) indicated that the Pliocene BP environment was characterized by an open larch dominated forest-tundra environment, sharing most species in common with those now found in three regions,

including central Alaska to Washington in western North America, the region centered around the Canadian/US border in eastern North America, as well as Fennoscandia in Europe. The modern area with the most species in common with BP is central northern Alaska (Fig. 7A). The area over which shared species were calculated is largely tundra, but includes the ecotone between tundra and boreal forest. Other zones that share many species with BP are continuous with Alaska down the western coast of North America to the region around the border of Canada and the United States, the eastern coast of North America in the region around the border of Canada and the United States (~50°N), and central Fennoscandia. Of these zones, the MST of Alaskan tundra sites (6–9°C) are less similar to BP (15.4°C) than ~50°N on both western and eastern coastal North American sites and central Fennoscandia (12–18°C, Fig. 7B). The eastern coast of North America has higher rainfall during the summer (>270 mm), than the west coast and Alaska (Fig. 7C), which correlates to the timing of western fires. The low summer precipitation for much of the west (<200 mm), is consistent with previously published summer precipitation estimates for BP (~190 mm). As a result, the fire regime of the west coast ~50°N may be a better analogue for BP than the east coast of North America. In central Fennoscandia there is also a west vs. east coastal variation in summer precipitation with the western, Nordic part of the region experiencing higher summer precipitation (252– >288 mm), than the more similar eastern, Swedish part of the region (~198 mm).

Comparison to modern fire detection data (Fig. 7D) suggests that the two regions most climatically similar to BP, ~50°N western North America and central Sweden, have radically different fire regimes. This is likely caused by historical fire suppression in Sweden that limits utility of very modern data for comparison in this study (Brown and Giesecke, 2014; Niklasson and Granström, 2004). To understand the fire regimes as shaped by climate and species composition rather than human impacts, we considered both the modern and recent Holocene reconstructions for these regions (Table 1). This shows that, a) within any region variation arises from the complex spatial patterning of fire across landscapes, and b) that the regions most similar to BP (~50°N western North American and eastern Fennoscandian reconstructions for the recent Holocene) have shorter fire return intervals than the cooler Alaskan tundra or wetter summer ~50°N eastern North American coast.

While the shared species for Siberia appears low, the number of observations in the modern biodiversity database used is likewise low – perhaps causatively so. Given the similar climate to BP on the Central Siberian Plateau and some key aspects of the floras in Siberia such as the dominance of larch, we considered the fire regime of the larch forests of Siberia. Kharuk et al. (2016; 2011) studied MFRIs across Siberia, from 64°N to 71°N, the northern limit of larch stands. They found an average MFRI across that range of 110 years, with MFRI increasing from 80 years in the southern latitudes to ~300 in the north (Table 1). Based on similarity of the climate variables, the more southerly MFRIs (~80 years) may be a better analogue. Key differences between boreal fires in the North America compared to Russia are a higher fire frequency with more burned area in Russia, but a much lower crown fire and a difference in timing of disturbance, with spring fires prevailing in Russia compared to mid-summer fires in western Canada (de Groot et al., 2013; Rogers et al., 2015).

The pollen-based vegetation reconstruction derived in this study indicates that open *Larix-Betula* parkland persisted in the basal (300.3-300.4 MASL) parts of the sequence. Groundcover was additionally dominated by shrub birch, ericaceous heath and ferns. While the regional climate may have been somewhat dry, the record suggests that, locally,

a moist fen environment dominated by Cyperaceae, existed near the sampling location. Shrubs including *Alnus* and *Salix* likely occupied the wetland margins.

The corresponding relatively low concentration of charcoal may reflect lower severity fires or higher sedimentation rates. If the former, it is posited that a surface fire regime existed. This premise is supported by the fire ecology characteristics of the dominant vegetation. *Larix* does not support crown fires due to leaf moisture content (de Groot et al., 2013) and self-pruning (Kobayashi et al., 2007). The persistence and success of larch in modern-day Siberia appears to be driven by its high growth rate (Jacquelyn et al., 2017) tolerance of frequent surface fire due to thick lower bark (Kobayashi et al., 2007) and tolerance of spring drought due to its deciduous habit (Berg and Chapin III, 1994). Arboreal *Betula* are very intolerant of fire and easily girdled. However, they are quick to resprout and are often found in areas with short fire return intervals. Like *Larix*, arboreal *Betula* have high moisture content of their foliage and are not prone to crown fires. *Betula* nana L., an extant dwarf birch, is a fire endurer that resprouts from underground rhizomes or roots (Racine et al., 1987) thus regenerating quickly following lower severity fires (de Groot et al., 1997). The vegetation and fire regime characteristics are similar further up the sequence at 301.15-301.25 MASL, with the exception that ferns increased in abundance while heath decreased.

In the upper part of the sequence (301.35-301.45 MASL), where charcoal was abundant, the *Larix-Betula* parkland was replaced by a mixed boreal forest assemblage with a fern understory. Canopy cover was more closed compared to the preceding intervals. The forest was dominated by *Larix* and *Picea*, with lesser amounts of *Pinus*. While *Betula* remained part of the forest, it decreased in abundance possibly due to increased competition with the conifers. Based on exploratory CRACLE analyses of climate preferences using GBIF occurrence data (GBIF.org, 2018a, b, c, d) of the dominant taxa (*Larix-Betula* vs. *Larix-Picea-Pinus*), the expansion of conifers could indicate slightly warmer summers (MST ~15.8 ºC vs. 17.1 ºC). This result differs from the stable MST estimated by bacterial tetraethers, although within reported error, and the small change is certainly within the climate distributions of both communities. The analyses also suggest that slightly drier conditions may have prevailed during the three wettest months (249-285mm vs. 192-219mm). While the interaction between climate, vegetation and fire is complex, small changes is MST and precipitation could have directly altered both the vegetation and fire regime, which in turn further promoted fire adapted taxa. In addition to regional climatic factors, community change at the site may have been further influenced by local hydrological conditions, such as channel migration, pond infilling and ecosystem engineering by beaver (*Cantor spp*.).

The high charcoal content suggests that fire was an important disturbance mechanism, although it could also reflect a slow sedimentation rate. If the former, it is likely that frequent, mixed severity fires persisted. While *Larix* is associated with surface fire, *Picea* and *Pinus* are adapted to higher intensity crown fires. A crown fire regime may have established as conifers expanded, altering fuel loads and flammability. For example, black spruce sheds highly flammable needles, its lower branches can act as fuel ladders facilitating crown fires (Kasischke et al., 2008), and it was previously tentatively identified at BP (Fletcher et al., 2017). While it has thin bark and shallow roots maladapted to survive fire (Auclair, 1985; Brown, 2008; Kasischke et al., 2008), it releases large numbers of seeds from semi-serotinous cones, leading to rapid re-establishment (Côté et al., 2003). The documentation of Onagraceae pollen at the

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top of the sequence could potentially reflect post-fire succession. For example, the species *Epilobium angustifolium* L. is an early-seral colonizer of disturbed (i.e. burnt) sites, pollinated by insects.

It is possible that the *Larix-Betula* parkland dominated intervals correspond to the peat- and sand-stratigraphic Units II and III described by Mitchell et al. (2016), whereas the mixed boreal forest in the upper part of the sequence is contemporaneous with Unit IV, described as peat and peaty sand, coarsening upwards. While it is clear that the vegetation and fire regimes changed through time at this Arctic site, CO_2 and temperatures appear more stable, or at least to have no apparent trend. Thus, it is suggested that the fire regime at BP was primarily regulated by regional climate and vegetation, and perhaps additionally by changing local hydrological conditions. Regarding climate, MST remained high enough (> ~13.5ºC) throughout the sequence to allow for fire disturbance and the pollen suggests that temperatures may have marginally increased in the upper part of the sequence. Alternatively, other climate variables, such as the precipitation regime, or local hydrological change may have initiated the change in community. Up-sequence changes in vegetation undoubtedly influenced fine fuel loads and flammability. Indeed, the fire ecological characteristics of the vegetation are consistent with a regional surface fire regime yielding to a crown fire regime.

Betula and *Alnus*, which occurred earlier in the depositional sequence, are favored by beaver in foraging (Busher, 1996; Haarberg and Rosell, 2006; Jenkins, 1979). Moreover, the presence of sticks cut by beaver in Unit III reveals that beavers were indeed at the site, moistening the local land surface. The lack of beaver cut sticks and changes in sediment in Unit IV may indicate that the beavers abandoned the site, possibly in response to changes in vegetation (i.e. increased conifers and decreased *Betula*) limiting preferred forage or due to lateral channel migration, as evidenced by the coarsening upward sequence described by Mitchell et al. (2016). As a result, the local land surface may have become somewhat drier, contemporaneous with the change towards *Larix-Picea-Pinus* forest and a mixed severity fire regime.

Critically, the charcoal record suggests that there was substantial biomass burning that could have been a feedback mechanism amplifying or dampening warming during the Pliocene due to its prevalence through time, and the complex direct impacts on the surface radiative budget and direct and indirect effects on the top of the atmosphere radiative budget (Feng et al., 2016). Further investigation is warranted to better characterize the fire regime to improve accuracy of fire simulations in earth system models of Pliocene climate.

5. CONCLUSION

The record of high CO_2 supports the hypothesis that Pliocene Arctic terrestrial fossil localities probably represent periods of higher warmth that supported higher productivity. The novel temperature estimates presented here suggest that summer temperatures were considerably warmer during the Pliocene (~15.4°C) compared to modern day Eureka, Canada (~4.1°C; Fig. 2). This highlights the increasing influence of arctic amplification of temperatures as CO_2 exceeds modern levels. Our reconstruction of the paleovegetation and ecology of this unique site on Ellesmere Island suggests an assemblage similar to forests of the western margins of North America and eastern Fennoscandia. The evidence of recurrent fire and concurrent changes in taxonomic composition suggests that fire played an active role in Pliocene Arctic forests, shaping the environment as it does in the boreal forest today. The importance of fire in the modern boreal forest suggests that fire may have had direct and indirect impacts on Earth's radiative budget at high

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The abundance of charcoal at BP demonstrates that climatic conditions were conducive for ignition and that sufficient biomass available for combustion existed across the landscape. Mean summer temperatures at BP likely exceeded the ~13.5 °C threshold (Young et al., 2017) that drastically increases the chance of wildfire as demonstrated here from brGDGT derived temperatures and corroborated by previous studies with a seasonal component (Csank et al., 2011b; Fletcher et al., 2017). An increase in atmospheric convection has been simulated in response to diminished sea-ice during warmer intervals (Abbot and Tziperman, 2008), but this study did not confirm if this increase in atmospheric convection was sufficient to cause lightning ignitions. An alternative ignition source for combustion of biomass on Ellesmere Island during the Pliocene is coal seam fires, which have been documented to be burning at this time (Estrada et al., 2009). However, given the interaction of summer warmth and ignition by lightning within the same climate range as posited for BP, we consider lightning the most likely source of ignition for Pliocene fires in the High Arctic.¶
Fire return intervals cannot be calculated from the BP charcoal counts due to the absence of a satisfactory age-depth model and discontinuous sampling. As strong interactions are observed between fire regime and ecosystem assemblage in the boreal forest (Brown and Giesecke, 2014; Kasischke and Turetsky, 2006), and in response to climate, comparison with modern fire regimes for areas with shared species compositions and climates may inform a potential range of mean fire return interval (MFRI).¶
The modern area with the most species in common with BP is central northern Alaska (Fig. 7A). The area over which shared species were calculated is largely tundra, but includes the ecotone between tundra and boreal forest. Other zones that share many species with BP are continuous with Alaska down the western coast of North America to the region around the border of Canada and the United States, the eastern coast of North America in the region around the border of Canada and the United States (~50°N), and central Fennoscandia. Of these zones, the MST of Alaskan tundra sites (6–9°C) are less similar to BP (15.4°C) than ~50°N on both western and eastern coastal North American sites and central Fennoscandia (12–18°C, Fig. 7B). The eastern coast of North America has higher rainfall during the summer (>270 mm), than the west coast and Alaska (Fig. 7C), which correlates to the timing of western fires. The low summer precipitation for much of the west (<200 mm), is consistent with previously published summer precipitation estimates for BP (~190 mm). As a result, the fire regime of the west coast ~50°N may be a better analogue for BP. [1]

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latitudes during the Pliocene, although the net impact of the component process remains unknown. Collectively, these reconstructions provide new insights into the paleoclimatology and paleoecology of the Canadian High Arctic, ~3.9 Ma.

Data Availability. The data generated and used in this analysis are available in the supplemental information associated with this article.

Sample Availability. Samples used in this analysis are curated by the Canadian Museum of Nature. Sample numbers used for each analysis are given in the supplemental information (Table S1 and S2).

Supplemental Link. To be provided by Copernicus Publishing

Author Contribution. Conceptualization: A.P.B. with modification by other authors; Methodology: A.P.B., J.G., J.S.S.D., K.J.B., T.F.; Formal analysis: All authors; Investigation: A.P.B., J.G., K.J.B., L.W., T.F.; Resources: A.P.B., J.G., J.S.S.D., K.J.B.; Data curation: A.P.B., J.G., K.J.B., L.W., T.F.; Writing—Original draft: All authors; Writing—Review and editing: All authors; Supervision: A.P.B., J.S.S.D., K.J.B., N.R.; Project administration: A.P.B., N.R., T.F.; Funding acquisition: A.P.B., J.G., J.S.S.D., K.J.B., N.R., T.F. (Definitions as per the CRediT Taxonomy)

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

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Table 1. Modern and recent Holocene fire return interval reconstructions for the candidate analogous regions considered in this study.

Region	Modern		Reference	Recent Holocene		Reference
Alaskan Tundra	Seward Peninsula	273*	Kasischke et al. (2002)	Up-Valley	263	Higuera et al. (2011)
	Nulato Hills	306*		Down-valley	142	
Alaskan Boreal	Porcupine/ Upper Yukon (Central)	~100	Yarie (1981)			
	Sites near Fairbanks, and Delta Junction (Central)	70130	Johnstone et al. (2010a); Johnstone et al. (2010b); Johnstone and Kasischke (2005)			
	Kenai Peninsula		Lynch et al. (2002)	Interior Alaska and Kenai Peninsula	198 ± 90	Lynch et al. (2002)
	Yukon river Lowlands	120	Kasischke et al. (2002)	Brooks Range	145	Higuera et al. (2009)
	Kuskokwim Mountains	218				
	Yukon-Tanama Uplands	330				
	Tanana-Kuskokwim Lowlands	178				
	Kobuk Ridges and Valleys	175				
	Davidson Mountains	403				
	North Ogilive Mountains	112				
	Ray Mountains	109				
	Yukon-Old Crow Basin	81				

Western North America	Darkwoods, British Columbia	~69	Greene and Daniels (2017)			
	Cascade Mountains, Washington	~27	Wright and Agee (2004)			
	Desolation Peak, Washington Coastal type	108-137				
	Desolation Peak, Washington Interior type	~52				
Eastern North America	Quebec – west	~270*	Bouchard et al. (2008)	Maine	≥ 800	Lorimer (1977)
	Quebec – east	>500*				
				Quebec – "Spruce zone"	570	de Lafontaine and Payette (2011)
				Quebec – "Fir zone"	>1000	
	Quebec – Abitibi northwest	418*	Bergeron et al. (2006 post-1940)^	Quebec – Abitibi northwest	189	Bergeron et al. (2006 post-1940)^
	Quebec – Abitibi southwest	388*		Quebec – Abitibi southwest	165	
	Quebec – Abitibi east	418*		Quebec – Abitibi east	141	
	Quebec – Abitibi southeast	2083*		Quebec – Abitibi southeast	257	
	Quebec – Temiscamingue north	2083*		Quebec – Temiscamingue north	220	

	Quebec – Temiscamingue south	2777*		Quebec – Temiscamingue south	313	
	Quebec – Waswanipi	418*		Quebec – Waswanipi	128	
	Quebec – Central Quebec	388*		Quebec – Central Quebec	150	
	Quebec – North Shore	645*		Quebec – North Shore	281	
	Quebec – Gaspésia	488*		Quebec – Gaspésia	161	
	Quebec – northwestern - lakeshore	99’	Bergeron (1991)	Quebec – northwestern - lakeshore	63’	Bergeron (1991)
	Quebec – northwestern – lake island	112’		Quebec – northwestern – lake island	74’	
Fennoscandia	Sweden	*	Niklasson and Drakenberg (2001); Niklasson and Granström (2004)	North Sweden	50-150	Niklasson and Granström (2004); Niklasson and Granström (2000)
				Southern Sweden	20	Niklasson and Drakenberg (2001)
	Central Sweden	*	Brown and Giesecke (2014)	Central Sweden - Klotjärnen	180	Brown and Giesecke (2014)
				Central Sweden - Holtjärnen	240	
Siberian Plateau	Northern	300	Kharuk et al. (2016); Kharuk et al. (2011)			
	Southern	80				
	Mean (64-71°N)	110				

^ = The reciprocal converted from burn rate (%) (see Van Wagner et al., 2006)

* = Estimates likely effected in some areas by human activity. In such instances Recent Holocene is preferred.

‘ = Fire cycle

†=‘Recent’ here refers to records that (or have distinct sections that) begin after the end of the Holocene Climate Optima and end near present

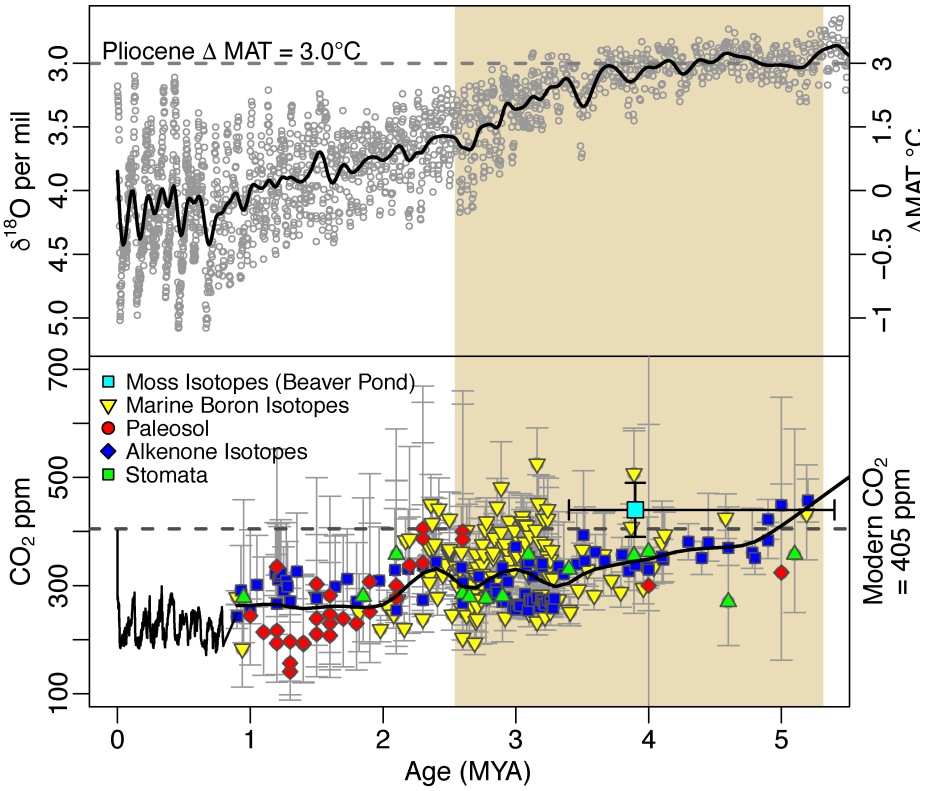

Figure 1: Global temperatures and atmospheric CO2 concentration spanning the last 5 million years of Earth's history. Mean annual temperatures (MAT) are inferred from compiled $\delta^{18}O$ foraminifera data (Lisiecki and Raymo, 2005) and plotted as anomalies from present (top panel). Modern atmospheric CO_2 measurements (NOAA/ESRL), and ice core observations from EPICA (Luthi et al., 2008) are compared with proxy estimates (bottom panel; see Table S1) for the Pliocene Epoch indicated with beige shading. Smoothed curves have been fit to highlight trends in pCO_2 and temperature during the Pliocene. The results from this paper (BP) are included with both age and pCO_2 error.

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Figure 1. Atmospheric CO_2 concentration and global temperatures and spanning the last 5 million years of Earth's history. Modern atmospheric CO_2 measurements (NOAA/ESRL), and ice core observations from EPICA (Luthi et al., 2008) are compared with proxy estimates (top panel) for the Pliocene Epoch highlighted in red (Hönisch et al., 2009; Pagani et al., 2010; Royer, 2006; Tripati et al., 2009). Mean annual temperatures (MAT) are inferred from compiled $\delta^{18}O$ foraminifera data (Lisiecki and Raymo, 2005) and plotted as anomalies from present (bottom panel). Smoothed curves have been fit to highlight trends in pCO_2 and temperature during the Pliocene. The results from this paper (BP) are included with both age and pCO_2 error.¶

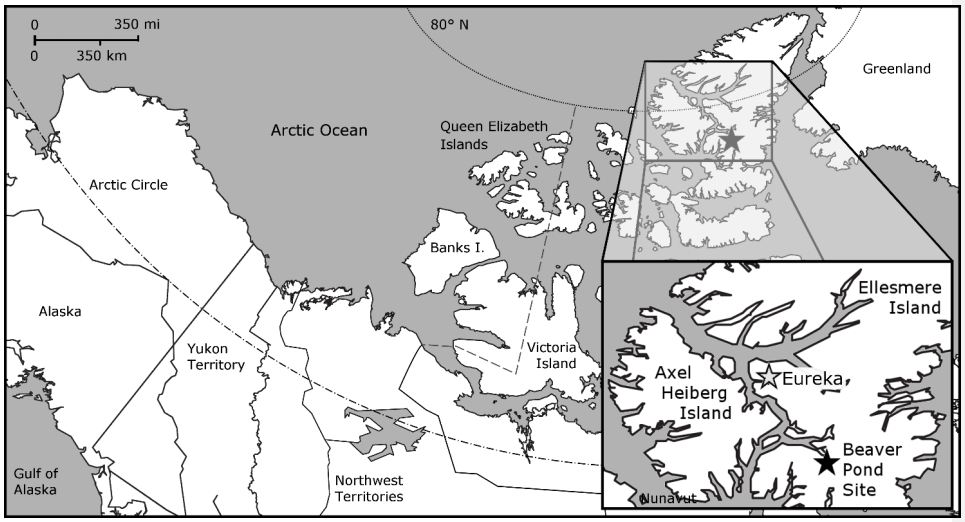

Figure 2. Map of the Canadian Arctic Archipelago, highlighting the location of the Beaver Pond Site (Black Star; 78° 33′ N; 82° 25′ W) and Eureka Climate Station (Grey Star; 80° 13’ N, 86° 11’ W ¬ used for modern climate comparison) on west-central Ellesmere Island.

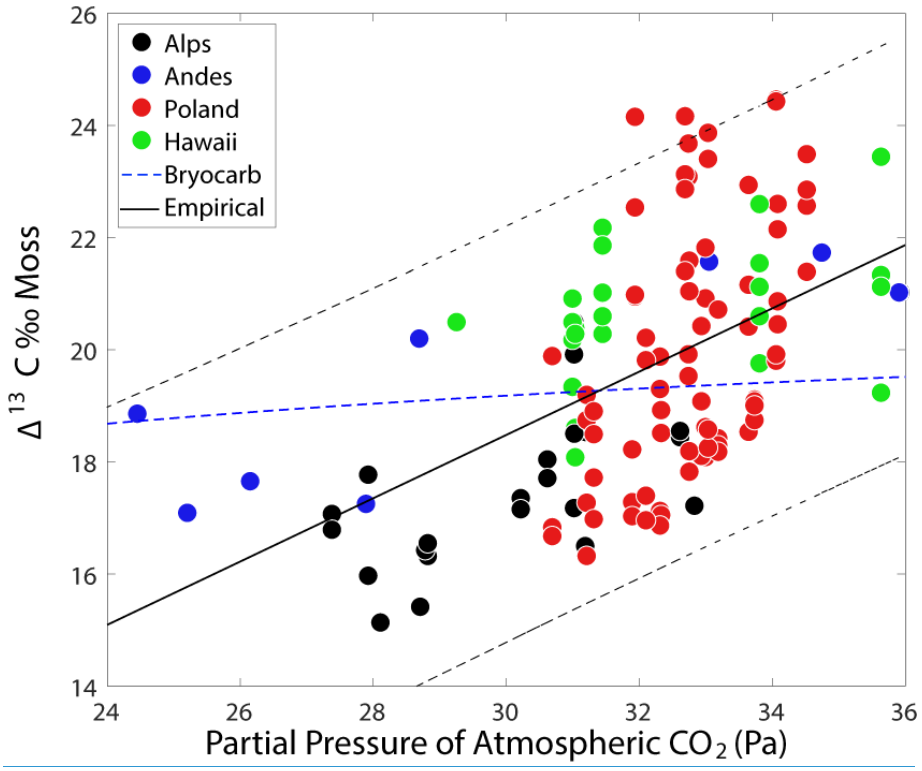

Figure 3. Sensitivity of carbon isotopic discrimination to the partial pressure of atmospheric CO_2 in mosses sampled from different elevational transects. Moss carbon isotope data collected from an elevational transects in the Swiss Alps (black dots; Ménot and Burns, 2001), the Peruvian Andes (blue dots; Royles et al., 2014), the mountains of Poland (red dots; Skrzypek et al. 2007), and Hawaii (green dots; Waite and Sack 2011). Partial pressure of atmospheric CO_2 calculated from atmospheric surface pressure reanalysis data (Dee et al., 2011) combined with atmospheric CO_2 observations from year moss samples were collected. All carbon isotopic measurements of mosses have been normalized to cellulose based on published regression of cellulose and whole moss values (Ménot and Burns, 2001) and reported as discrimination (Δ) from atmospheric $\delta^{13}CO_2$ (GlobalGlobal View-CO_2, 2013) from the year mosses were collected in units of ‰. Empirical model fit (black line) is plotted with prediction intervals (black dashed) compared with predictions from the BRYOCARB model (Fletcher et al. 2008) with parameters optimized to match observations.

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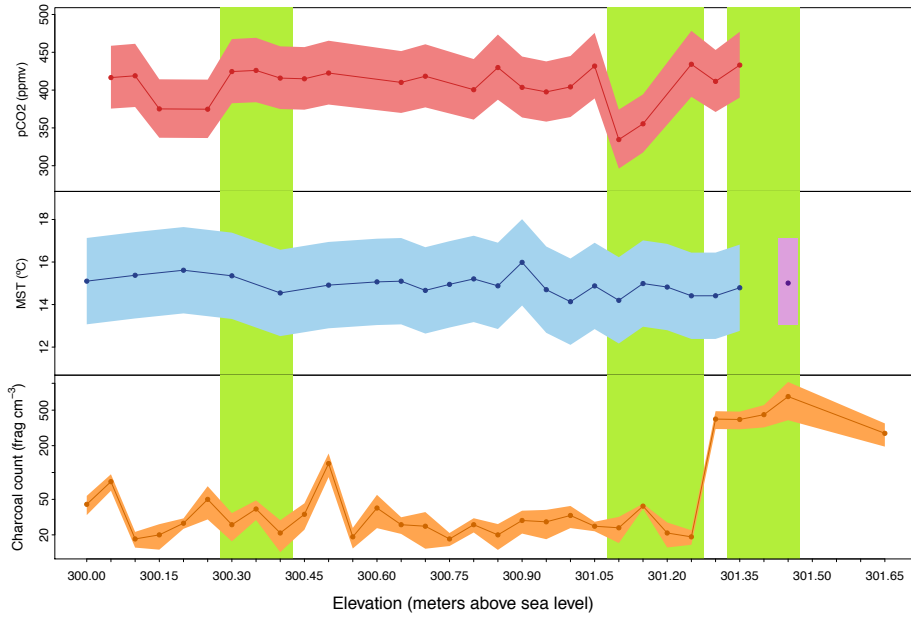

Figure 4. Reconstruction of atmospheric CO_2, mean summer temperature, and fire for the Canadian High Arctic during the Pliocene. Atmospheric CO_2 concentrations estimated from carbon isotopic measurements of mosses and plants (red; ± 2 σ). Mean summer temperature reconstructed from a brGDGT based proxy (blue; ± 2 σ) and relative 2010 data point in approximate relative position (purple; ± 2 σ). Charcoal counts reported as the number of fragments per volume (fragments cm^{-3}) of peat (Orange ± 2 σ). Green boxes indicate relative depths of pollen sampling. Elevation of the deposit is reported as meters above sea level. (Data: Table S2)

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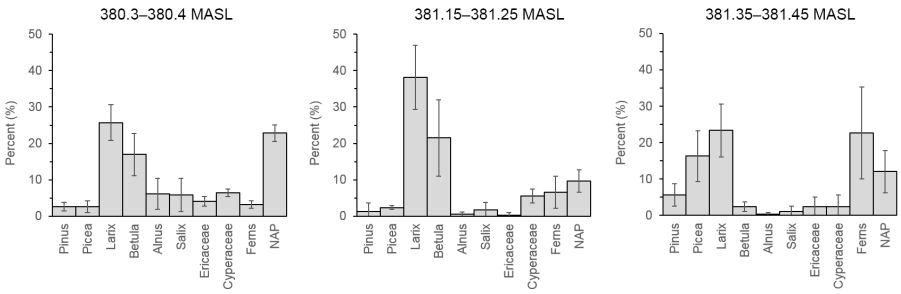
(A)



(B)

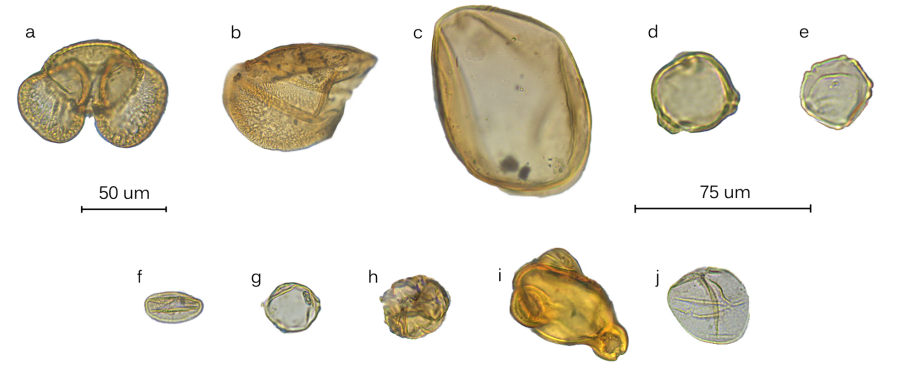


Figure 5. (A) Bar charts showing the relative pollen abundance in each portion of the section (error bars = 95% confidence intervals; MASL- Meters Above Sea Level). (B). Pollen plate of select grains encountered in the BP section: (a) *Pinus*, (b) half a *Picea* grain, (c) *Larix*, (d) *Betula*, (e) *Alnus*, (f) *Salix*, (g) *Myrica*, (h) ericaceous grain, (i) *Epilobium*, and (j) Cyperaceae. 50um scale = (a–c), 75um scale = (d–j).

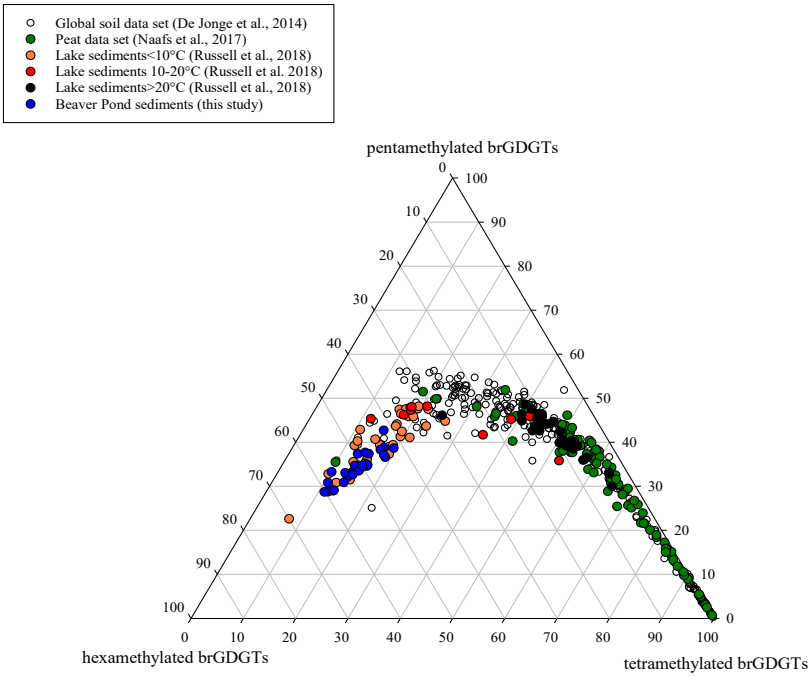


Figure 6. A ternary plot illustrating the fractional abundances of the tetra- (Ia-c), penta (IIa-c and II′a-c), and hexamethylated (IIIa-c and III′a-c) brGDGTs. The global soil dataset (open circles; De Jonge et al., 2014), the global peat samples (green circles; Naafs et al., 2017), and lake sediments from East Africa (black circles indicate samples from lakes >20°C, red circles indicate samples from lakes between 10–20°C and orange circles designate samples from lakes <10°C; Russell et al., 2018) are included for comparison with the Beaver Pond sediments (blue circles; this study).

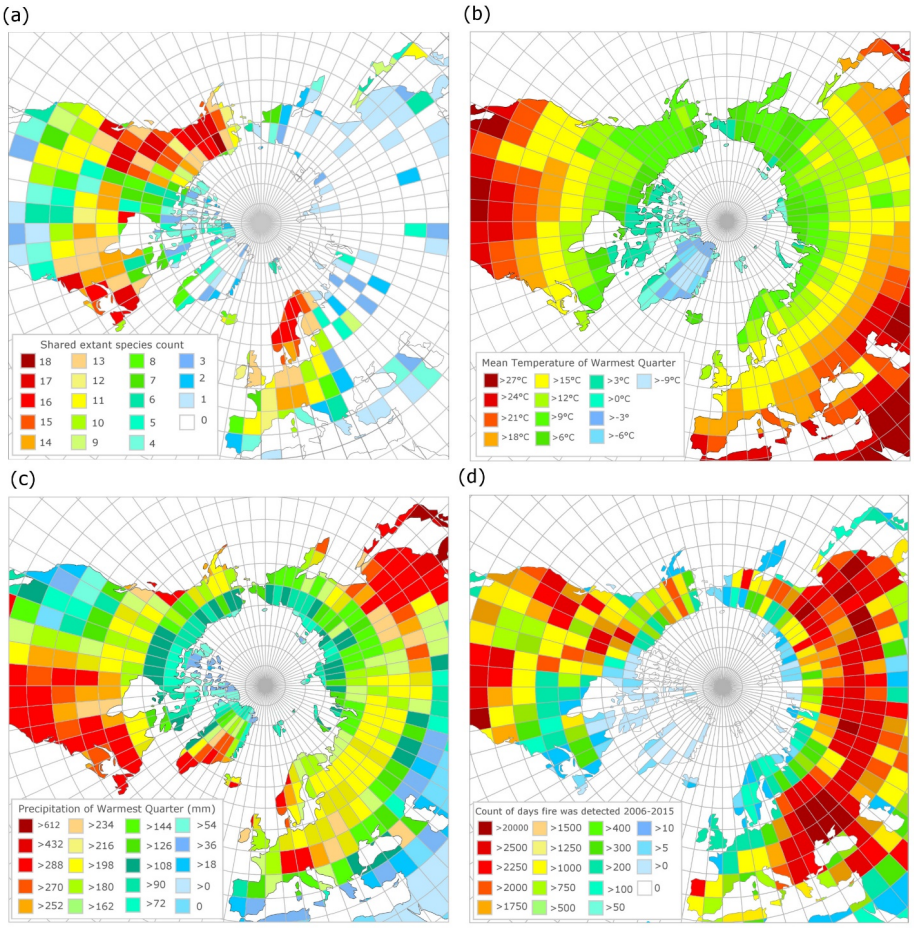


Figure 7. (a) Modern geographic distribution of observed occurrences of species common to the Beaver Pond species list, (b) Mean temperature of the warmest quarter (summer average) derived from WorldClim, (c) Mean precipitation of the warmest quarter (summer rain) derived from WorldClim, (d) Count of unique fire pixels detected per day, over 10 years from MODIS 6 Fire Product, normalized by area of the latitude by longitude grid.

Supplementary Information

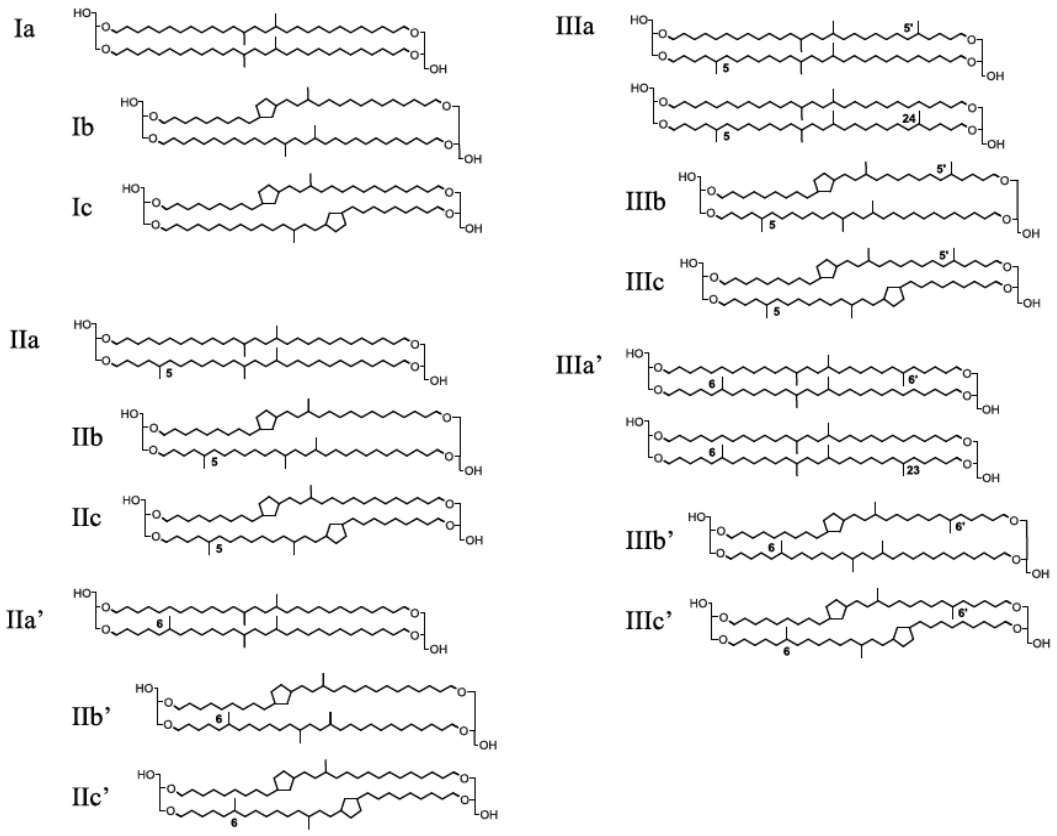


Figure S1. Molecular structures of all 15 brGDGTs (I-III). The molecules designated with a prime symbol are referred to as the 6-methyl brGDGTs.

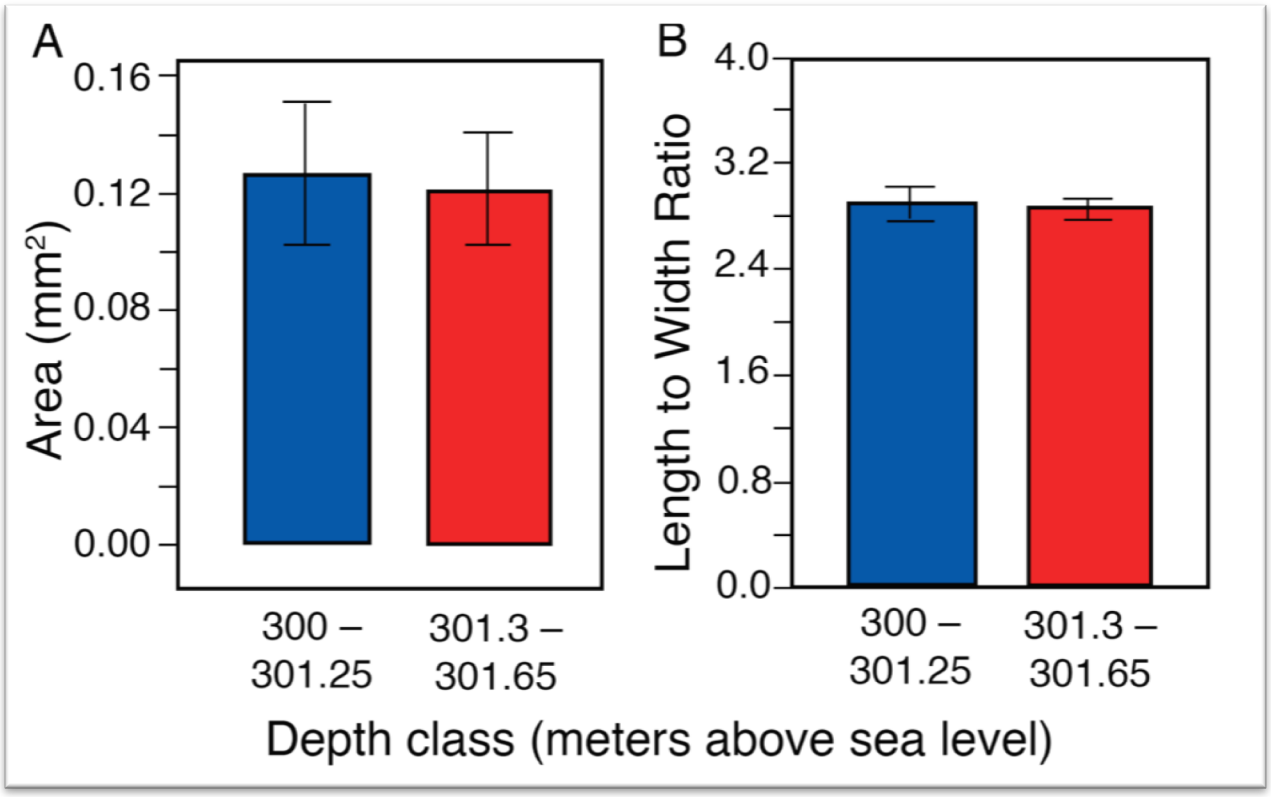


Figure S2. A comparison of the (A) area and (B) shape (length to width) of the uppermost samples (130–165) that have a higher mean charcoal concentration, and the lowermost samples (0–125)

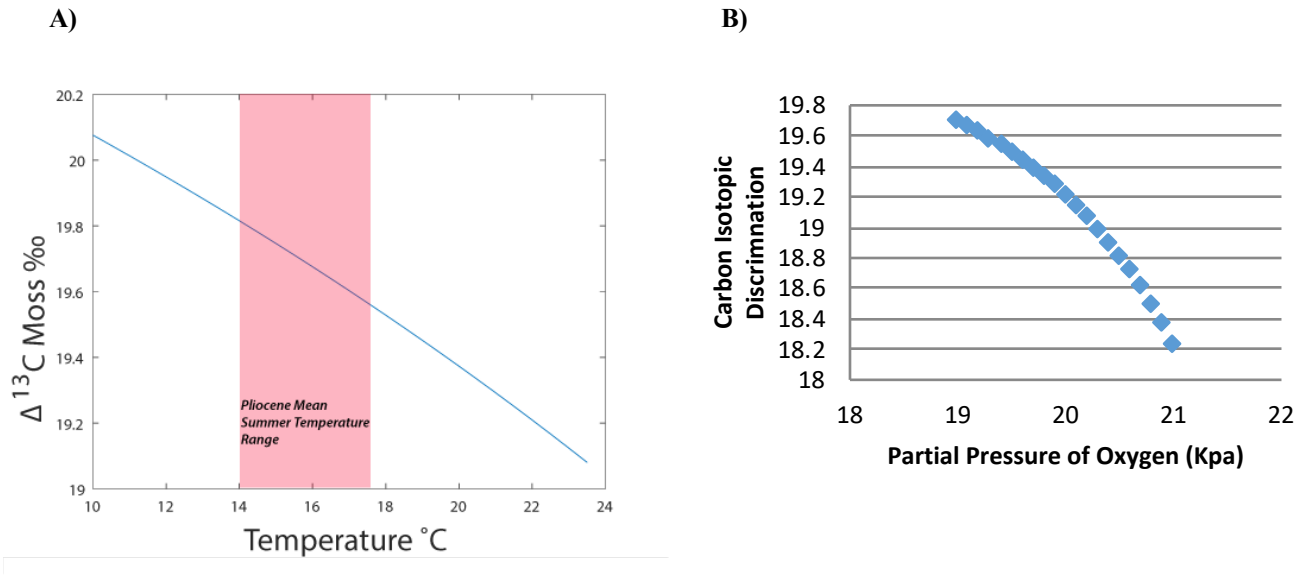
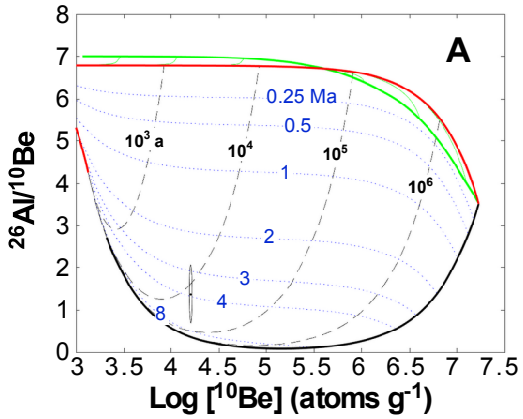


Figure S3. A) Changes in moss $\Delta^{13}C$ as a function of temperature. Where the entire range of temperatures represents the lapse rate of moss samples from the Peruvian Andes and the red box represents the mean summer temperature range for the Arctic derived from our tetraether measurements. B) Changes in moss $\Delta^{13}C$ only as a function of pO_2 showing an increase in $\Delta^{13}C$ with elevation (i.e. reduced pO_2).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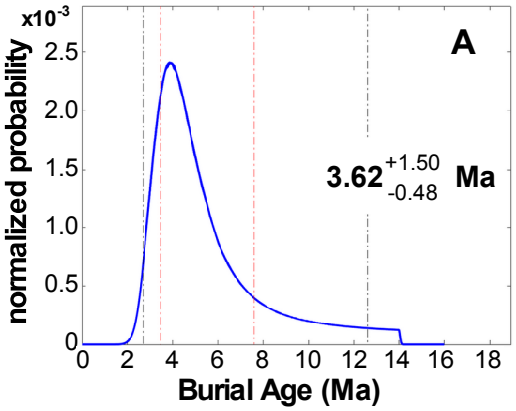
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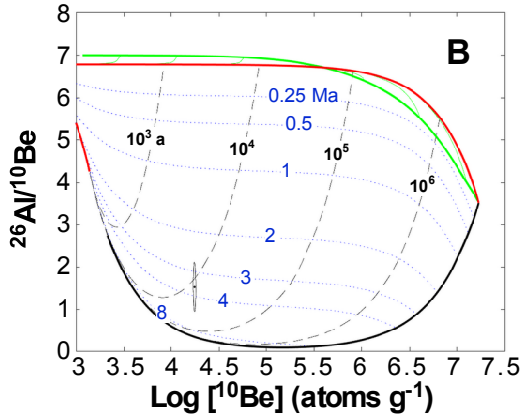
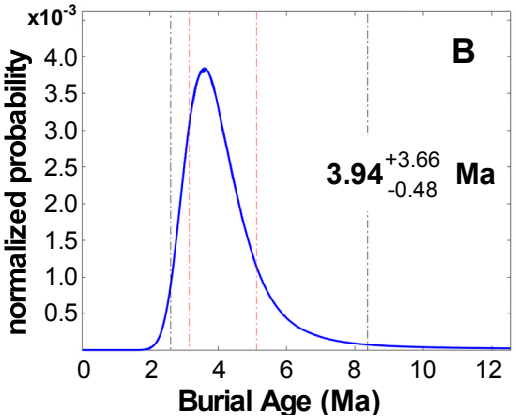
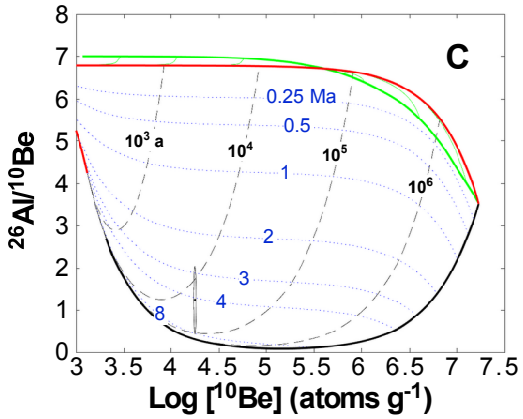
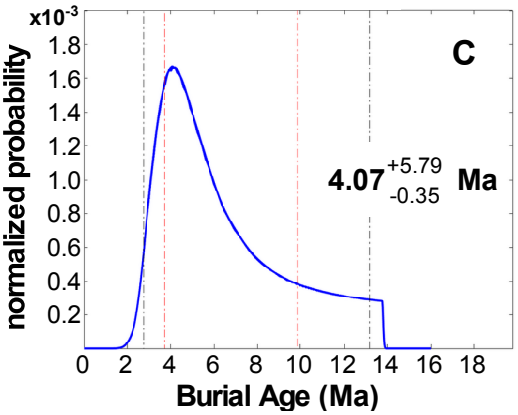
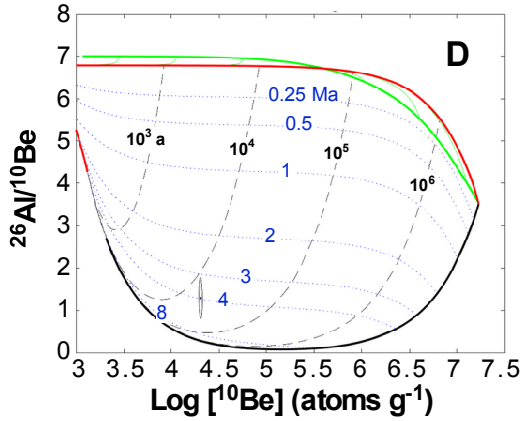
3.62 +1.50 -0.48 Ma
Burial Age (Ma)


B
0.25 Ma
0.5
1
2
3
4
8
10^3 a
10^4
10^5
10^6
26Al/10Be
Log [10Be] (atoms g-1)


B
x10-3
normalized probability
3.94 +3.66 -0.48 Ma
Burial Age (Ma)


C
0.25 Ma
0.5
1
2
3
4
8
10^3 a
10^4
10^5
10^6
26Al/10Be
Log [10Be] (atoms g-1)


C
x10-3
normalized probability
4.07 +5.79 -0.35 Ma
Burial Age (Ma)


D
0.25 Ma
0.5
1
2
3
4
8
10^3 a
10^4
10^5
10^6
26Al/10Be
Log [10Be] (atoms g-1)

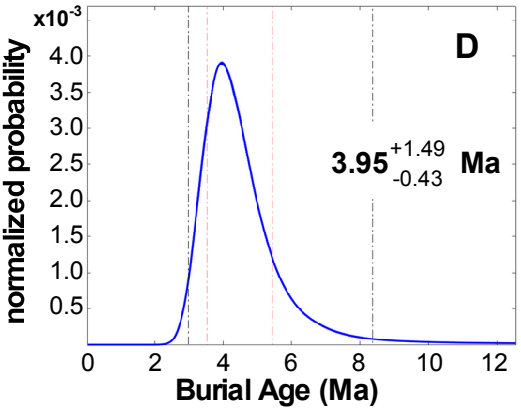

D
x10-3
normalized probability
3.95 +1.49 -0.43 Ma
Burial Age (Ma)

Figure S4. Burial age results. depthID: average initial sample depth, pdfmaxage: the most probable age as determined from the probability density function, sigma1plus and sigma1minus: the +/-1sigma errors in pdfmaxage, sigma2plus and sigma2minus: the +/- 2sigma errors in pdfmaxage, exposure_meanvalue: mean value of pre-buildup exposure age taken from FMINLBFGS optimization algorithm, burial_meanvalue: mean value of burial age taken from FMINLBFGS optimization algorithm. The 26Al/10Be vs. log10 10Be plots are unique for each mass depth. The generally horizontal dotted curves are burial isochrons, from top to bottom 0.25, 0.50, 1.0, 2.0, 3.0, 4.0. and 8.0 Ma, and the near-vertical dashed lines are pre-burial minimum exposure duration isochrons, from left to right 103, 104, 105, and 106 years. The PDF plots are probability distributions of 2000 solutions of the data.

Table S1. Fractional abundances of the brGDGTs found in the Beaver Pond sediments.

Sample Name	Ia	Ib	Ic	IIa	IIb	IIc	IIIa	IIIb	IIIc	IIa′	IIb′	IIc′	IIIa′	IIIb′	IIIc′
BP-A-02	0.08	0.05	0.01	0.15	0.05	0.00	0.27	0.01	0.00	0.12	0.05	0.00	0.21	0.01	0.00
BP-A-03	0.07	0.03	0.00	0.11	0.03	0.00	0.26	0.00	0.00	0.15	0.05	0.00	0.30	0.01	0.00
BP-A-04	0.12	0.04	0.00	0.13	0.03	0.00	0.23	0.00	0.00	0.15	0.04	0.00	0.25	0.01	0.00
06BP01	0.10	0.04	0.01	0.13	0.03	0.00	0.31	0.00	0.00	0.12	0.03	0.00	0.22	0.00	0.00
06BP16	0.11	0.04	0.00	0.15	0.03	0.00	0.30	0.00	0.00	0.13	0.03	0.00	0.20	0.00	0.00
06BP18*	0.11	0.03	0.00	0.15	0.02	0.00	0.29	0.00	0.00	0.14	0.03	0.00	0.21	0.00	0.00
06BP18*	0.11	0.03	0.00	0.15	0.02	0.00	0.29	0.00	0.00	0.13	0.03	0.00	0.21	0.00	0.00
BP-F-73*	0.10	0.04	0.00	0.19	0.03	0.00	0.26	0.00	0.00	0.14	0.03	0.00	0.19	0.01	0.00
BP-F-73*	0.11	0.05	0.00	0.22	0.04	0.00	0.21	0.00	0.00	0.16	0.04	0.00	0.15	0.01	0.00
BP-A-06	0.11	0.04	0.01	0.15	0.03	0.00	0.24	0.00	0.00	0.15	0.04	0.00	0.22	0.01	0.00
BP-A-07	0.13	0.04	0.01	0.16	0.03	0.00	0.23	0.00	0.00	0.15	0.04	0.00	0.20	0.01	0.00
06BP03	0.11	0.04	0.00	0.15	0.03	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.22	0.01	0.00
06BP05*	0.10	0.04	0.00	0.15	0.03	0.00	0.29	0.00	0.00	0.12	0.03	0.00	0.23	0.01	0.00
06BP05*	0.11	0.04	0.01	0.16	0.03	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.19	0.00	0.00
06BP07	0.13	0.04	0.00	0.19	0.03	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.16	0.00	0.00
06BP09	0.09	0.03	0.00	0.14	0.03	0.00	0.30	0.00	0.00	0.13	0.03	0.00	0.23	0.01	0.00
06BP11*	0.11	0.04	0.01	0.16	0.03	0.00	0.31	0.00	0.00	0.12	0.03	0.00	0.19	0.00	0.00
06BP11*	0.10	0.04	0.00	0.15	0.03	0.00	0.29	0.00	0.00	0.14	0.03	0.00	0.21	0.00	0.00
06BP13	0.11	0.04	0.00	0.15	0.03	0.00	0.30	0.00	0.00	0.13	0.03	0.00	0.20	0.00	0.00
06BP14*	0.10	0.04	0.00	0.15	0.03	0.00	0.29	0.00	0.00	0.13	0.03	0.00	0.21	0.00	0.00
06BP14*	0.10	0.04	0.00	0.14	0.03	0.00	0.30	0.00	0.00	0.14	0.03	0.00	0.21	0.01	0.00
06BP15*	0.10	0.03	0.00	0.12	0.03	0.00	0.31	0.00	0.00	0.12	0.03	0.00	0.23	0.00	0.00
06BP15*	0.09	0.04	0.01	0.15	0.03	0.00	0.35	0.00	0.00	0.11	0.02	0.00	0.20	0.00	0.00
06BP17*	0.09	0.03	0.00	0.14	0.03	0.00	0.31	0.00	0.00	0.15	0.04	0.00	0.20	0.00	0.00
06BP17*	0.10	0.04	0.01	0.15	0.03	0.00	0.29	0.00	0.00	0.12	0.03	0.00	0.21	0.00	0.00
06BP19	0.11	0.04	0.00	0.14	0.03	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.22	0.00	0.00
06BP20	0.11	0.03	0.00	0.16	0.03	0.00	0.30	0.00	0.00	0.12	0.03	0.00	0.20	0.00	0.00
06BP21	0.08	0.04	0.01	0.14	0.03	0.00	0.38	0.00	0.00	0.09	0.02	0.00	0.19	0.00	0.00
06BP22*	0.12	0.04	0.00	0.17	0.03	0.00	0.29	0.00	0.00	0.12	0.03	0.00	0.19	0.00	0.00
06BP22*	0.11	0.04	0.00	0.16	0.03	0.00	0.31	0.00	0.00	0.12	0.03	0.00	0.19	0.00	0.00
06BP23*	0.15	0.04	0.01	0.21	0.03	0.00	0.37	0.00	0.00	0.14	0.03	0.00	0.00	0.00	0.00
06BP23*	0.12	0.04	0.00	0.17	0.03	0.00	0.30	0.00	0.00	0.12	0.03	0.00	0.18	0.00	0.00
BP-F-78*	0.09	0.04	0.00	0.13	0.03	0.00	0.26	0.00	0.00	0.14	0.04	0.00	0.24	0.01	0.00
BP-F-78*	0.10	0.05	0.00	0.15	0.04	0.00	0.22	0.00	0.00	0.17	0.04	0.00	0.20	0.01	0.00

06BP24*	0.12	0.04	0.00	0.17	0.03	0.00	0.30	0.00	0.00	0.12	0.03	0.00	0.18	0.00	0.00
06BP24*	0.13	0.03	0.00	0.16	0.02	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.20	0.00	0.00
BP-A-16	0.13	0.05	0.01	0.14	0.03	0.00	0.23	0.00	0.00	0.15	0.04	0.00	0.20	0.01	0.00
06BP25*	0.12	0.03	0.00	0.16	0.02	0.00	0.28	0.00	0.00	0.13	0.03	0.00	0.21	0.00	0.00
06BP25*	0.12	0.03	0.00	0.16	0.02	0.00	0.29	0.00	0.00	0.13	0.03	0.00	0.20	0.00	0.00
BP-A-17	0.14	0.04	0.01	0.16	0.03	0.00	0.26	0.00	0.00	0.14	0.03	0.00	0.18	0.00	0.00
06BP26*	0.10	0.03	0.00	0.15	0.03	0.00	0.34	0.00	0.00	0.11	0.02	0.00	0.20	0.00	0.00
06BP26*	0.12	0.03	0.00	0.16	0.02	0.00	0.30	0.00	0.00	0.12	0.03	0.00	0.20	0.00	0.00
06BP27*	0.08	0.03	0.00	0.12	0.02	0.00	0.33	0.00	0.00	0.11	0.03	0.00	0.25	0.00	0.00
06BP27*	0.08	0.03	0.00	0.12	0.02	0.00	0.33	0.00	0.00	0.12	0.03	0.00	0.26	0.00	0.00
BP-A-18 1	0.15	0.04	0.00	0.18	0.03	0.00	0.24	0.00	0.00	0.14	0.03	0.00	0.17	0.00	0.00
06BP28*	0.06	0.02	0.00	0.08	0.02	0.01	0.33	0.00	0.00	0.10	0.03	0.01	0.32	0.01	0.00
06BP28*	0.08	0.03	0.00	0.09	0.02	0.00	0.29	0.01	0.00	0.13	0.04	0.00	0.31	0.00	0.00
06BP28*	0.08	0.03	0.00	0.10	0.03	0.00	0.30	0.01	0.00	0.13	0.04	0.00	0.28	0.00	0.00
BP-A-20*	0.07	0.03	0.00	0.10	0.03	0.00	0.31	0.01	0.00	0.11	0.05	0.00	0.28	0.01	0.00
BP-A-20*	0.08	0.03	0.00	0.11	0.03	0.00	0.29	0.01	0.00	0.12	0.05	0.00	0.26	0.01	0.00
BP-A-69*	0.08	0.04	0.00	0.13	0.03	0.00	0.35	0.00	0.00	0.10	0.02	0.00	0.23	0.00	0.00
BP-A-69*	0.09	0.04	0.00	0.14	0.03	0.00	0.32	0.01	0.00	0.11	0.02	0.00	0.23	0.00	0.00

*indicates polar fraction was re-analyzed

Table S2. Data used to generate Figure 4.

Sample	Elevation	pco2	pco2min	pco2max	MST	MSTmin	MSTmax	CharCount	Charmin	Charmax	1 std	Pollen
06BP01	300				15.10	13.07	17.13	44.00	33.31	54.69	10.69	
06BP02	300.1	417	376	459				79.00	62.44	95.56	16.56	
06BP03	300.1	419	378	461	15.38	13.35	17.41	18.00	14.39	21.61	3.61	
06BP04	300.2	375	337	414				20.00	13.65	26.35	6.35	
06BP05	300.2				15.61	13.58	17.64	27.00	23.39	30.61	3.61	
06BP06	300.3	375	337	414				50.00	29.79	70.21	20.21	
06BP07	300.3				15.35	13.32	17.38	26.00	16.93	35.07	9.07	1
06BP08	300.4	425	383	467				39.00	29.36	48.64	9.64	1
06BP09	300.4	426	384	469	14.55	12.52	16.58	21.00	12.81	29.19	8.19	1
06BP10	300.5	416	375	458				34.00	22.85	45.15	11.15	
06BP11	300.5	415	374	457	14.91	12.88	16.94	126.00	88.73	163.27	37.27	
06BP12	300.6	423	381	465				19.00	14.07	23.93	4.93	
06BP13	300.6				15.07	13.04	17.10	40.00	23.80	56.20	16.20	
06BP14	300.7	410	370	452	15.10	13.07	17.13	26.00	20.49	31.51	5.51	
06BP15	300.7	418	377	461	14.67	12.64	16.70	25.00	13.98	36.02	11.02	
06BP16	300.8				14.95	12.92	16.98	18.00	15.00	21.00	3.00	
06BP17	300.8	401	361	441	15.21	13.18	17.24	26.00	21.27	30.73	4.73	
06BP18	300.9	430	387	473	14.88	12.85	16.91	20.00	13.57	26.43	6.43	
06BP19	300.9	404	364	444	15.98	13.95	18.01	29.00	20.67	37.33	8.33	
06BP20	301	398	358	438	14.70	12.67	16.73	28.00	17.88	38.12	10.12	
06BP21	301	404	364	445	14.14	12.11	16.17	33.00	23.83	42.17	9.17	
06BP22	301.1	432	389	476	14.88	12.85	16.91	25.00	21.94	28.06	3.06	
06BP23	301.1	335	296	374	14.20	12.17	16.23	24.00	16.19	31.81	7.81	
06BP24	301.2	355	318	394	14.99	12.96	17.02	42.00	39.35	44.65	2.65	1
06BP25	301.2				14.82	12.79	16.85	21.00	14.44	27.56	6.56	1
06BP26	301.3	434	391	478	14.41	12.38	16.44	19.00	15.49	22.51	3.51	1

06BP27	301.3	412	371	453	14.41	12.38	16.44	397.00	307.28	486.72	89.72	
06BP28	301.4	433	390	477	14.79	12.76	16.82	393.00	304.74	481.26	88.26	1
06BP29	301.4							445.00	320.14	569.86	124.86	1
06BP30	301.5							710.00	385.49	1034.51	324.51	1
06BP34	301.7							275.00	195.00	355.00	80.00	
BP-A-20*	~301.35				15.01	12.98	17.04					

*2010 field season sample

Table S3. $^{26}Al/^{10}Be$ burial ages

PDF max age (yr)	1s error +ve (yr)	1s error -ve (yr)	meanvalue (yr)
3.62E+06	1.50E+06	4.78E+05	3.58E+06
3.94E+06	3.66E+06	4.77E+05	NA
4.07E+06	5.79E+06	3.51E+05	NA
3.95E+06	1.49E+06	4.31E+05	3.95E+06

Table S4. Input for burial modeling

Depth cm	Bulk Density g cm^{-3}	Latitude deg	Longitude deg	Surface elevation m	^{10}Be conc atoms g^{-1}	^{10}Be conc err atoms g^{-1}	^{26}Al conc atoms g^{-1}	^{26}Al conc err atoms g^{-1}	Eros Rate cm ka^{-1}
18050	2.2	78.550	-82.373	333	17665	402	26986	7335	2.25
18097	2.2	78.550	-82.373	333	16163	376	22263	7889	2.25
18155	2.2	78.550	-82.373	333	17853	387	22322	10215	2.25
18222	2.2	78.550	-82.373	333	20505	598	26508	7147	2.25

Notes:

1. We attempted a depth-profile type isochron burial date (Balco and Rovey, 2010), however the differences in the measured concentrations were too small and uncertainties in ^{26}Al were too large to define an isochron curve. Therefore we used the more common method of simple burial dating using the relationship of $^{26}Al/^{10}Be$ vs. $\log^{10}Be$, which requires the assumption that the pre-burial ratio of the sand samples was the production ratio of $^{26}Al/^{10}Be$ (6.75).

2. We computed the $^{26}Al/^{10}Be$ burial ages that best fits the measured concentrations, assuming a simple surface buildup and burial history. The depositional environment in the Pliocene was an alluvial fan and pebble-braided-stream system along a mountainous piedmont. Therefore it is reasonable to assume that there was little opportunity for long-term (Ma) deep burial (>20 m) of stored sediment during transport from the nearby mountains. Thus we assume, like most other applications of the simple burial method, that the initial ratio in the sand grains was 6.75 and that only one significant burial event affected the grains in the past 8 Ma (by that time, both isotopes have effectively reached saturation). We use the Lifton et al. (2014) constraints and approach for scaling the production rates in the catchment (buildup) and in the sampled section (post-depositional). The calculations include post-depositional muon production (cosmic ray influx according to Lifton et al., 2014) and erosion of the surface.

3. While the current depth of the samples is approximately 10 m below gravel and till, muons can still penetrate to produce cosmogenic ^{26}Al and ^{10}Be. In other words, the samples are not completely shielded. However, we estimate that there was more sediment and ice above the samples during the Pliocene and Pleistocene as follows: The surface of the Beaufort Formation is more than 40 m higher in elevation across Strathcona Fiord than at BP. Furthermore the region has been significantly eroded since the Pliocene sediment was deposited, as stream paleoflow indicators reveal that the fiord was filled at the time of BP deposition. The amount of erosion in the fiord is much greater than 400 m. We conservatively estimate that 50 m of post-Pliocene erosion occurred above the fiord on the opposite side, or 90 m of sediment loss on the BP side. This would equate to an erosion rate over 4 Ma of 2.25 cm ka^{-1}. Besides sediment with a bulk density of 2.2 g cm^{-3} for coarse sand and sandy gravel, BP would have been covered by ice for the majority of the Quaternary, given its close proximity to the second largest ice field in Canada, Prince of Wales Ice Field. Plateau ice thicknesses are currently >200 m (Kinnard et al., 2008) (i.e. equivalent to approximately 81.8 m of sandy gravel), whereas the ice field is much thicker in valleys and would have been even thicker during much of the Pleistocene. While the mass depth that shielded the samples at any time remains uncertain, our most reasonable estimate is 90.0 +81.8 or 171.8 m of average gravel cover (mass depth = 37.8 x 103 g cm^2). We added 171.8 m to each of the modern sample depths (8.70, 9.17, 9.74, and 10.42 m). While the uncertainty in the actual mass depth is large, a greater shielding thickness does not change the age significantly once depths are greater than 50 m (much deeper burial would yield a slightly younger mean age of 3.6 Ma, while a shallower depth estimate will significantly increase the age beyond the ca. 8 Ma saturation limit. Therefore, we prefer the revised calculated mean burial age over the minimum ages reported in Rybczynski et al. (2013) which were derived using unreasonably great depths and zero erosion (needed for no muogenic production) and a superseded production rate systematics.

4. The mean age of the four samples is 3.9 +1.5/-0.5 Ma. The final most probable ages are therefore our best estimate of burial duration of the Beaver Pond layer. The burial age and error (1σ and 2σ shown in Figure S3) is determined using a systematic parameter search and chi-squared statistic to create a continuous probability density function. We also calculate a burial mean-value age using the FMINLBFGS optimization algorithm (from Matlab file exchange) for comparison

with the probability distribution function most probable burial age approach. Those burial mean-value ages were not available for two of the samples because the tails of their pdfs reach beyond the saturation value.

Supplemental References

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