1 **Boreal Peatland Water Table Depth and Carbon Accumulation During the**
2 **Holocene Thermal Maximum, Roman Warm Period, and Medieval Climate**
3 **Anomaly**

4

5 James R Holmquist

6 Smithsonian Environmental Research Center

7 P.O. Box 28, 647 Contees Wharf Rd

8 Edgewater, MD 21037

9 Email: HolmquistJ@si.edu | Phone: 01 443 482 2267

10

11 Robert K Booth

12 Earth and Environmental Science Department, Lehigh University

13

14 Glen M MacDonald

15 Department of Geography, University of California, Los Angeles

16

17

18

19

20

21

22

23

Abstract

The Boreal Shield and James Bay Lowland regions of Ontario have few Holocene reconstructions specific to surface moisture despite their potential importance to documenting the geographic patterns and teleconnections of Holocene pluvials and droughts, as well as their role in the global carbon cycle. We reconstructed water table depth using preserved testate amoebae in four stratigraphic peatland cores, supplemented with a literature review to investigate the effects of the Holocene Thermal Maximum (HTM) and Medieval Climate Anomaly (MCA) on surface moisture and long-term apparent carbon accumulation (LARCA) in the regions. A 7320 calendar years before AD 1950 (yrBP) length record registered a wet shift at 4600 yrBP in concert with the HTM and a later post-2550 yrBP shift towards wetter and more stable conditions, possibly related to the Roman Warm Period. During the late-Holocene the three most southern boreal records indicated a lack of droughts during the MCA. Medieval pluvials in boreal Ontario would be consistent with North American precipitation patterns under La Niña conditions. Of the three wet-warm periods identified (HTM, 2550 yrBP, MCA), the 2550 yrBP shift had the most consistent positive influence on LARCA regionally. Although past precipitation was positively correlated with temperature in the region, recent drying may be indicative of a fundamental change in evaporation-precipitation balance.

47 **Keywords**

48 Testate Amoebae; Peatlands; Boreal; James Bay Lowland; Holocene Thermal Maximum;

49 Medieval Climate Anomaly; Roman Warm Period

50

51

52

53

54

55

56

57

58

59

60

61

62

63

64

65

66

67

68

69

70

71

72

73 **Abbreviations**

74 ^{14}C - Radiocarbon

75 C - carbon

76 HBL – Hudson Bay Lowland

77 HTM – Holocene Thermal Maximum

78 JBL – James Bay Lowland

79 LARCA – Long-term apparent rate of carbon accumulation

80 LIA – Little Ice Age

81 MAP – Mean Annual Precipitation

82 MAT – Mean Annual Temperature

83 MCA – Medieval Climate Anomaly

84 NAO – North Atlantic Oscillation

85 RWP – Roman Warm Period

86 SST – Sea Surface Temperatures

87 TA – Testate Amoebae

88 TJJA – Average June, July, August Temperature

89 WTD – Water Table Depth

90 yrBP – years before AD 1950

91

92

I. Introduction

Peatland deposits serve the dual role as records of past climate change, and as stores of global soil carbon (C; Clymo et al., 1998; Bridgeham et al., 2006; Dise, 2009). Climate reconstructions by proxy records are vital in understanding the sensitivity and responses of boreal and subarctic systems where climate change is likely to be more pronounced than in lower latitudes (Price et al., 2013). Peatland C storage is dependent on saturated conditions, and low temperatures (Yu et al., 2009), yet in some instances may be enhanced by high plant productivity during warm episodes (Beilman et al., 2009; Loisel et al., 2012; Charman et al., 2013), and therefore it is important to understand the effect past warming has had on surface moisture and C storage in the world's large and vulnerable northern peatlands. The Holocene Thermal Maximum (HTM), which had variable timing in North America (Kaufman et al., 2004; Ressen et al., 2009), and the Medieval Climate Anomaly (MCA; 1000 - 700 yrBP; Mann et al., 2009) provide natural experiments for the effect of positive temperature anomalies on surface moisture and C sequestration (Booth et al., 2010a). Remote regions of the Boreal Shield and James Bay Lowland (JBL) of Ontario are currently data sparse despite their climatically sensitive location relative to phenomena such as Pacific forcing (Shabbar et al., 1997; Seager et al., 2008) and their importance as a major global peatland C store (Loisel et al., 2010; van Bellen et al., 2011; Bunbury et al., 2012; Lamarre et al., 2012). In this paper we present new water table depth (WTD) reconstructions from four peatlands in central and northern Ontario, one spanning mid-Holocene to modern, and three spanning late-Holocene to

modern and combine these with previously published data to assess the relationship between large-scale warming, hydroclimatology and C sequestration in this region.

Testate amoebae (TA) are useful indicators of past environmental change in peatlands, particularly water table depth (WTD) and pH (Charman et al., 2001; Booth, 2008; Mitchell et al., 2008; Sullivan and Booth, 2011; Amesberry et al., 2013). In oligotrophic peatlands, TA community composition is sensitive to surface moisture, and their tests remain well preserved (Booth, 2008; Amesberry et al., 2013). Studies in the Quebec JBL and elsewhere in North America have used TA to investigate mid and late-Holocene WTD variability and related hydrologically sensitive phenomena (van Bellen et al., 2011; Booth et al., 2012; Clifford and Booth, 2013). Other studies in the JBL region, and the Quebec JBL have used more recent peat deposits to constrain WTD fluctuations during the MCA and Little Ice Age (LIA; Loisel and Garneau, 2010; Lamarre et al., 2012; Bunbury et al., 2012).

The Hudson Bay Lowlands (HBL), JBL, and Boreal Shield of Ontario encompass the second largest peatland complex in the world (Riley, 2011; Tarnocai and Bockheim, 2011). In total, Ontario is estimated to store 26.0 Pg of C in the HBL-JBL, and 9.3 Pg in the Boreal Shield (Far North Science Advisory Panel, 2010). Climate change is already occurring in boreal areas with temperature increasing 0.5-3.0°C and mean annual precipitation (MAP) increasing by 10-20% over the instrumental record (Price et al., 2013). Likewise, longer arctic growing seasons caused increases in plant productivity by 0.46-0.67% per year between 1986-2006 (Jia et al., 2009). By 2100 the Boreal Shield and HBL-JBL may warm as much as 8.0°C and 5.0°C in winter and summer respectively. Precipitation is predicted to increase modestly in the Boreal Shield (11% by 2100) and

more in the HBL-JBL (20-40% by 2100) with both areas shifting towards more winter precipitation (Price et al., 2013).

Precipitation changes may influence the direction and magnitude of C cycle feedbacks in peatlands (Yu et al., 2009). C accumulation in bogs can be highly dependent on moisture thresholds of *Sphagnum* species (Strack et al., 2009). Some studies indicate that WTD drawdown decreases *Sphagnum* productivity (Robroek et al., 2009), and increases peat decay through the increased activity of aerobic bacteria (Fenner and Freeman, 2011). Likewise, wetter conditions can cause higher net ecosystem productivity in peatlands because of lower decay rates of submerged peat (Laiho, 2006).

Warm periods during the Holocene (Kaufman et al., 2004; Mann et al, 2009; Renssen et al., 2009, Wang et al., 2012) may provide natural experiments that can be examined to assess the effects of elevated temperatures on WTD in the region. The HTM period was delayed in some regions of Canada because of the influence of residual Laurentide ice, and later effects of open water and ice in the Hudson and James Bays. During the HTM the dry arctic front in central Canada was at its most northern extent between 6000 and 4000 yrBP along with the northernmost position of the tree line dividing boreal and tundra biomes (Edwards et al., 1996; MacDonald et al., 1998; Beringer et al., 2001) and conditions in the study area were likely warm and moist.

Post-HTM a pollen record from the JBL records a warm-wet shift post-2700 yrBP following cool-dry conditions (O'Reilly et al., 2014). This shift could potential be related to the Roman Warm Period (RWP; 2500–1600 yrBP; Wang et al., 2012). In some regions, such as coastal Scotland, increased moisture during this time was associated with eastern Atlantic cooling (Wang et al. 2012). Although most evidence for warming and/or

moisture shifts for the RWP is accruing in Europe, Greenland and elsewhere (Hass, 1996; Ji et al., 2005; Seidenkrantz et al., 2007; Martín-Puertas et al., 2009; Wang et al., 2012) its nature and causes are not as clearly understood as the HTM and MCA, especially in North America.

In the late-Holocene the warm MCA (1000 - 700 yrBP) and subsequent cool LIA (550 – 250 yrBP) are associated with respective positive and negative temperature anomalies and were likely caused by a combination of variations in solar intensity and volcanic aerosol forcing (Shindell et al., 2003). Northern Hemisphere proxy records and models indicate as high as a 1.8°C anomaly between MCA and LIA (Mann et al., 2009). During the MCA sea surface temperatures (SST) in the eastern tropical Pacific were cool, suggesting persistent La Niña conditions at 1000 yrBP coupled with positive-phase North Atlantic Oscillation (NAO) and Arctic Oscillation (MacDonald and Case, 2005; Trouet et al., 2009; Oglesby et al., 2012; Ortega et al., 2015). North American precipitation today is, and during the MCA was, sensitive to tropical Pacific forcing (Herweijer and Seager, 2008) and coupled climate models and empirical data indicate that droughts extending through portions of the North American Southwest and Plains during the MCA were exacerbated by persistent La Niña-like conditions (Cook et al., 2004; MacDonald and Case, 2005; Feng et al., 2008; Seager et al., 2008; Mann et al., 2009). However, winter precipitation in the Pacific Northwest and portions of boreal Canada are typically enhanced under these conditions (Shabbar et al., 1997; Herweijer and Seager, 2008; Steinman et al., 2012). The current study area lies on the hinge point between a negative and positive hydrological response to La Niña-like conditions (Shabbar et al., 1997; Seager et al., 2008) and documenting its response to the MCA will aid in understanding

the teleconnections between climate warming, the Pacific Ocean and North American hydrology.

Peatland initiation and related C sequestration was driven by deglaciation, marine retreat due to isostatic uplift in the JBL, and possibly by increased moisture during the HTM in regions south of the JBL (Holmquist et al., 2014; Packalen et al., 2014; Holmquist and MacDonald, 2014). LARCA can be highly variable within individual peatland sites due to autogenic forcing, but regionally synchronous changes may be linked with temperature, growing season length, photosynthetically active radiation (Holmquist et al., 2014), and isostatic uplift (Glaser et al., 2004; Packalen et al., 2014). In a previous synthesis from the study region vertical peat accumulation was significantly and positively correlated with MAP over the last 2000 yrBP, although MAP was highly intercorrelated with temperature (Holmquist et al., 2014). While the ultimate formation and productivity of peatlands depends on moisture thresholds (Yu et al., 2009), it is not clear whether these thresholds were crossed in the past with moisture deficit having a negative effect on LARCA, or whether they will be crossed in the future due to geographic or temporal changes in precipitation or evaporation.

In this study, we reconstructed WTD for four peatlands in the Boreal Shield and JBL of Ontario and incorporated these into a larger paleohydrology network to determine: 1. Did the JBL region experience hydrological shifts coincidental with warming during the mid-Holocene, MCA, and recent climate change? 2. Did WTD shifts during the HTM in our record coincide with changes seen in nearby reconstructions of WTD, relative humidity, and precipitation? 3. Is the larger geographic patterning of MCA hydroclimatic changes consistent with Pacific SST – North American teleconnections and

broader geographies of pluvials and droughts associated with La Niña-like conditions? 4.
Did LARCA increase during high-temperature, high-precipitation events?

2. Study Sites

The four peat cores used in this study (JBL4, JBL7, JBL2, and JBL8) were collected in the summer of 2008 as part of an eight-core transect of ombrotrophic peat bogs (Holmquist et al., 2014; Fig. 1) in central and northern Ontario. All four cores presented here were dominated by *Sphagnum fuscum*. JBL4, JBL7, and JBL8 occurred in larger bog systems with *Sphagnum angustifolium* hollows and intermediate *Sphagnum magellanicum* sections, while JBL2 was from an ombrotrophic *Sphagnum*-dominated section within a larger poor-fen complex. All four cores were permafrost-free at the time of collection, however the two most northern cores (JBL7 and JBL4) came from sites with subarctic *Cladonia* spp. occurring discontinuously. JBL4 is located within the mapped extent of discontinuous permafrost (Brown et al., 2001).

3. Materials and Methods

3.1 Stratigraphic Information

The top 0.8 – 1.0 m of peat was collected with a box corer, with subsequent depths being sampled with a Russian auger. Cores were frozen and subsampled on a band saw in 1-cm increments. We analyzed cores for bulk density, loss-on-ignition, and dominant macrofossil components. Full reports of these measurements as well as additional methods are available in Holmquist et al. (2014) and Holmquist and MacDonald (2014). We established well-constrained chronologies of the cores by using multiple AMS radiocarbon (^{14}C) dates (Holmquist et al., 2014) and Bayesian age-depth

modeling implemented with Bacon 2.2 (Blaauw and Christen, 2011; Reimer et al., 2013; R-Core Development Team, 2013; Holmquist and MacDonald, 2014).

3.2 Testate Amoebae Processing and Identification

Initial analysis indicated that JBL7 was dominated largely by *Sphagnum* throughout its 7400 yrBP stratigraphy and was thus selected to reconstruct WTD throughout both the mid- and late-Holocene. We used 4-cm resolution for TA analysis pre-2550 yrBP variability in JBL7. We varied post-2550 yrBP TA sampling resolution for all four cores to account for changes in the slope of the age-depth model and effectively capture WTD variability during the MCA and LIA. Counts were halted earlier in JBL8 and JBL2, because TA were not abundant in deeper older peat.

Processing and identification of TA followed protocols modified from Charman (2001) and Booth et al. (2010). We separated TA from peat using two stages of filtration (Booth et al., 2010b), filtering samples through a 250- μ m filter and reverse-filtered them through a 7- μ m filter. Common North American TA test sizes typically fall between 10 - 200 μ m in length (Charman, 2001; Booth, 2008). We counted 150 tests per sample except for samples that had fewer than ~ 6000 tests cc^{-1} , for which we halted the analysis at 50 tests (Payne and Mitchell, 2009). We identified TA following Charman et al. (2000) and Booth (2008), including shells of the rotifer *Habrotrocha angusticollis*. Several taxa were combined into single taxonomic groupings in this study due to their ecological or morphological similarity, and for consistency with the calibration dataset (Tab. 1). We summarized temporal patterns in TA percentages using nonmetric multidimensional scaling (NMS), as well as constrained cluster analysis (Bennett, 1996; McCune et al., 2002; Juggins, 2009; R Core Development Team, 2013). We reconstructed WTD using a

weighted-average transfer function approach and 776 modern samples from North America (Booth, 2008; Markel et al. 2010). *Diffflugia pristis* type were included in multivariate statistics, zonation statistics and discussion of zones, but were not included in the transfer function analysis because of the lack of information about the group's morphological variability, taxonomy, and relatively poor characterization of environmental optima in the calibration dataset.

Testate amoebae data from JBL7 and JBL2 have been discussed in other publications focusing on diatom records (Hargan et al., 2015), and patterns in vegetation and peat decay (Philben et al., 2015). In this publication they are combined with additional new and review data to focus on mid to late-Holocene surface moisture fluxes and possible relations to allogenic drivers such as climate.

3.3. Synthesis Review Data

To provide a broader climatic context we synthesized the results of other published regional and continental precipitation, relative humidity, and surface moisture reconstructions, as well as mapped North American precipitation anomalies for the seven most extreme La Niña periods in the observational record. Additional WTD (Bunbury et al., 2012), MAP and Mean Annual Temperature (MAT; O'Reilly et al., 2014), and relative humidity data (Edwards et al., 1996) were plotted relative to JBL7 for mid to late-Holocene comparisons. Additional age-depth interpretation was required for relative humidity reconstructions from Edwards et al., (1996), as the ages were reported in ¹⁴C-years. For these data we converted ¹⁴C ages to yrBP by using the median calibrated age from INTCAL13 (Blaauw, 2010; R-Core Development Team, 2013; Reimer et al., 2013). We mapped precipitation anomalies relative to the new climate normal period (1981-

2010) for the seven most extreme La Niña events according to the Multivariate El Niño Southern Oscillation Index (1949-51, 1954-56, 1964-66, 1970-1972, 1973-75, 1988-90, 2010-2011; NOAA, 2013a, 2013b; Fig. 1). We collated existing paleohydrology records related to the MCA and mapped the geographic pattern of drought or pluvial conditions during that period and determined the overall North American pattern and assessed the consistency of the JBL records with that pattern.

3.4. Long-term apparent rate of carbon accumulation

LARCA for the cores we collected, as well as four others from the same latitudinal transect of Ontario's Boreal Shield and JBL, were compiled and published in Holmquist and MacDonald (2014). LARCA was estimated as the product of bulk density, loss-on-ignition at 550°C, and an assumption of 0.48 g C / 1 g organic matter, and accumulation rate for each 1-cm increment inferred from the age-depth model. Regional compiled LARCA values were graphed using 500-year bins of mean-corrected, detrended LARCA values from the catotelms of all eight cores (Holmquist and MacDonald, 2014).

4. Results

Taxa that had a concentration of > 20% in at least one sample from each record are displayed in the corresponding stratigraphic plot and summary table (Figs 2-5; Tab. 2). Constrained cluster analysis determined that three to five stratigraphic zones were apparent in the TA assemblages (Figs 2-5; Tab. 2). NMS ordinations were 2-dimensional for all sites, with final stress values ranging from 11.0-15.3 (Tab. 3). In all four cores the WTD reconstructions correlated significantly with NMS axis 1, 2 or both (Tab 3; Fig 6), indicating that the transfer function adequately summarizes past variability in community composition.

299 The TA record from JBL7 extends from 7320 yrBP to modern times, and is
300 divided into five assemblage zones (Fig. 2; Tab. 2). JBL7-A spans both the post-glacial
301 and HTM periods ending at 2550 yrBP. Relatively low concentrations of *Archerella*
302 *flavum* relative to later time periods, and relatively high percentages of *Heleopera* spp.,
303 *Hyalosphenia papilio*, and *Phryganella acropodia* characterize the mid-Holocene
304 stratigraphy (Fig. 2; Tab. 2). Low levels of the relatively dry indicator *Trigonopyxis* spp.
305 occur throughout the mid-Holocene, and high percentages of the taxon occur at 2600
306 yrBP indicating a significant drying event. Despite having data gaps from low TA
307 preservation, perhaps due to shifts to herbaceous-dominated peat, the record captures
308 general shifts from dry to slightly wetter conditions from 7320 to 4000 yrBP. Peak
309 wetness around 4600 yrBP is still dry relative to later Holocene wet shifts. The record
310 also captures transient dry shifts around 6000 and 2600 yrBP.

311 Persistent occurrences of *Diffflugia pulex* and *A. flavum* dominate the stratigraphy
312 in JBL7 from 2550 yrBP to modern (Fig. 2; Tab. 2), consistent with relatively wet but
313 likely variable conditions (Sullivan and Booth, 2011). From 2550 to 1280 yrBP (JBL7-B)
314 WTD was relatively wet compared to the rest of the late-Holocene (11 ± 2 cm). From
315 1280 to 660 yrBP (JBL7-C), a shift in dominance from *A. flavum* to *D. pulex* occurred,
316 along with the minor occurrences of *Arcella catinus* type and *Arcella vulgaris* type. There
317 was little change in WTD despite statistical shifts in community composition. *A. flavum*
318 and *D. pulex* occupy similar WTD in the transfer function used, however *D. pulex* may be
319 an indicator of more variable conditions. From 660 to 260 yrBP (JBL7-D) higher
320 concentrations of *A. flavum* and decreasing amounts of *D. pulex* occur. During the most

recent period (JBL7-E) *A. flavum* decreased, while *D. pristis* type, *D. pulex*, *H. papilio*, and *Nebela militaris* increased, indicating generally drier, but variable conditions.

JBL4 is the most northern late-Holocene record and extends from 1610 yrBP to modern times, and was divided into four zones (Fig. 3; Tab. 2). *Diffflugia pulex* was dominant from 1610 to 1180 yrBP (JBL4-A), whereas *Archerella flavum* was dominant from 1180 to 800 yrBP (JBL4-B). *Bullinularia indica* spikes in JBL4-B suggest drier conditions. From 800 to 270 yrBP (JBL4-C) *A. flavum* decreased, and *D. pulex* increased, indicating a more variable WTD. However JBL4-C WTD reconstructions are more uncertain than other sites because high concentrations of *Diffflugia pristis* type, a taxonomically uncertain group not well characterized in the calibration dataset, occurred in the core. During the last three centuries (JBL4-D), there was an increase in *A. flavum*, *Hyalosphenia elegans*, and *Hyalosphenia papilio*, and a decrease in *D. pulex*.

The record of JBL2 extends from 1050 yrBP to modern times and was divided into four zones (Fig. 4; Tab. 2). From 1050 to 500 yrBP (JBL2-A and JBL2-B) *Amphitrema wrightianum* type was dominant indicating very wet conditions, likely standing water throughout much of the growing season. From 500 to 160 yrBP (JBL2-C) *A. flavum* replaced *A. wrightianum* in dominance, and *Heleopera* spp. occurred at lower concentrations indicating a shift to drier, but still moderately wet conditions. During the last century (JBL2-D) *A. flavum* and *D. pulex* declined, while *H. elegans* and *H. papilio* increased. The recent century was dry relative to the rest of the late-Holocene.

The TA record from JBL8 extends from 1340 yrBP to modern times and was divided into three zones (Fig. 5; Tab. 2). From 1340 to 620 yrBP (JBL8-A) *A. flavum*, *D. pulex*, and *H. angusticollis* were common indicating moderate to wet conditions relative

to the rest of the late-Holocene. From 1140-1040 yrBP there was a slight increase in *Trigonopyxis* spp. indicating a short dry event. From 620 to 210 yrBP (JBL8-B), *Trigonopyxis* spp. and *Heleopera* spp. increased while *A. flavum* decreased indicating relatively dry conditions. From 210 yrBP to modern times (JBL8-C) *A. flavum* declined and *Assulina muscorum*, *D. pulex*, and *H. papilio* increased, indicating relatively dry and variable conditions.

5. Discussion

5.1 Mid-Holocene Hydrology in Boreal Canada

The WTD reconstruction from JBL7 records dry early post-glacial/post emergence conditions, an HTM with a peak in moisture about 4600 yrBP, and wet stable conditions occurring post-2550 yrBP (Fig. 2; Fig. 7).

Peak moisture at 4600 yrBP, may have been related to the HTM. Edwards et al. (1996) reported a HTM moisture peak with similar timing for Southern and Central Canada (Fig. 7). They cite a northward shift of the summer position of the arctic front as a possible reason for this pattern (Edwards et al., 1996). Records of WTD, total precipitation, and MAT from Victor Mine area show moisture and temperature exceeding mean Holocene levels, but not approaching recent modern levels during this time period (Bunbury et al., 2012; O'Reilly et al., 2014). The terminus of the HTM was marked by a shift to dry conditions, characterized by a decrease in relative humidity post-4.6 ka, and lower WTD in JBL7 and VC04-06 (Bunbury et al., 2012). Pollen studies in the region also suggest a drop in temperature from 3400 – 2700 yrBP coinciding with drier conditions (O'Reilly et al., 2014).

An increase in moisture post-2550 yrBP at JBL7 is consistent with a warm-wet period reconstructed from other studies in the region, and possibly with the RWP. TA community composition changes in both JBL7 and VC04-06 may be related to general increases in mean annual precipitation recorded by Viau and Gajewski (2009), although the WTD change is much more rapid in JBL7. Similar to the HTM 4600 yrBP wet shift, the RWP wet shift occurs during wet-warm periods inferred from nearby pollen records (O'Reilly et al., 2014; Fig. 7). Although dating precision precludes a firm conclusion, this warming may be related to increased insolation of the RWP (Wang et al. 2012).

5.2 Regional Hydrology patterns during Late-Holocene

The patterns of the MCA and LIA are much better resolved than the RWP and warrant more detailed comparative discussion. All four records show synchronous multi-centennial changes in WTD or TA community composition centering on the MCA (1000 - 700 yrBP), LIA (550 – 250 yrBP; Fig. 8), and the natural and anthropogenic warming of the most recent centuries (Mann et al., 2009). Ecological transitions recorded by JBL4-B, JBL7-C, JBL2-A and B, and JBL8-A correspond to the MCA, whereas JBL4-C, JBL7-D, JBL2-C, and JBL8-B represent fluctuations during the LIA (Fig. 8; Tab. 2), and all four cores show dry variable conditions during the last 1-3 centuries.

JBL4, the most northerly record, contrasts to the three more southerly sites and VC04-06 (Bunbury et al., 2012) by indicating a dry MCA relative to the LIA (Fig. 8). Likewise during the LIA the reconstruction from JBL4 differs from sites to the south, recording drier conditions. The pattern at JBL4 differs from more southerly sites, which record little change, or relatively wet conditions during the MCA. This could reflect the boundary of pluvial versus drought response to Pacific Ocean forcing in the central

Canadian subarctic (Fig. 1; Ladd and Gajewski, 2010), or alternatively the local influence of permafrost on surface moisture and peat growth (Lammare et al., 2012).

In JBL7, there was little difference in WTD between the pre-MCA and MCA, however the MCA was wet relative to the LIA and recent centuries. The WTD during the LIA was more variable than during the MCA, and more uncertain due to dominance by *D. pulex*, which has a wider WTD tolerance than *A. flavum*. Bunbury et al., (2012) also report a wet MCA and VC04-06 displays trends similar to JBL7, however differing sampling resolutions make a direct comparison between absolute maximum and minimum WTD inadvisable (Figs 4, 5, 8; Tab. 2). The two most southern sites at JBL8 and JBL2 indicate a MCA characterized by relatively wet periods, with a subsequent relatively dry LIA. The relative changes were much more extreme in JBL2 where surface pooling switched to a dry subsurface water. Similar shifts occurred but were more moderate in JBL8. Unfortunately, both cores lack a record of the mid-Holocene, so comparisons can only be made to later hydrological fluctuations. In JBL2 this shift from wet to dry is coincident with succession from fen to bog vegetation (Hargan et al., 2014; Fig. 4).

During recent centuries all four sites as well as VC04-06 (Bunbury et al., 2012) record drying. This may reflect the influence of unprecedented warming reported for the Canadian arctic and subarctic resulting in high rates of evaporation relative to precipitation over the growing season during the past one to two centuries (Rühland et al., 2003; Schindler and Smol, 2006; Paul et al., 2010). This is especially the case for the three most northern cores (JBL4, JBL7, and VC04-06; Fig. 8). Previous studies have downweighted taxa to accommodate differential preservation and found no significant

affect on WTD reconstruction (Mitchell et al., 2008); drying conditions in the upper stratigraphies of TA records are likely not the result of a bias introduced by differential preservation of tests. It is also possible that other anthropogenic forces, such as increased sulphate deposition could alter testate amoebae community composition by increasing the proportion of bacteriavores (Payne et al., 2010). Future studies could focus on TA records from the last 1-3 centuries to determine if recorded drying is from recent changes in evaporation-precipitation balance or another source of variation that has not been accounted for. However, if these changes are registering climate variation, then present warming may be having a fundamentally different effect on surface moisture than it did during the HTM, RWP, and MCA in boreal Canada.

5.3. The geographic distribution of MCA and LIA droughts and pluvials

Modern data indicate pluvial conditions for much of the Boreal Shield, and JBL north of the Great Lakes during La Niña years (Fig. 1; Tab. 4) extending east to the Quebec portion of the JBL. Areas that are typically moist during recent La Niña events show evidence of increasing *Sphagnum* colonization, decreased WTD, and regional dune stabilization during the MCA (Filion, 1987; Loisel and Garneau, 2010). South of the Great Lakes there is evidence of increased aridity and fire activity from peatland records extending from the northern Great Plains to eastern Ontario during the MCA (Booth et al., 2006; Laird and Cumming, 2009; Booth et al. 2012).

The geographic pattern exhibited in the JBL and surrounding region is consistent with a general, but varied, continental pattern of precipitation changes during the MCA being similar to recent major La Niña events. Both modern La Niña precipitation anomalies and MCA hydrology show a pattern of droughts in the US Southwest, North

American Northeast, and Great Plains, paired with pluvial conditions in portions of the Pacific Northwest, the northern Rocky Mountains and north of the Great Lakes (Fig. 1; Tab. 4). In the Great Plains proxy records indicate widespread medieval droughts (Vance et al., 1992; Vance et al., 1997; Cambell et al., 2000; Fritz et al., 2000; Tian et al., 2006; Laird and Cumming, 2009; Galloway et al., 2011), supporting dendrochronologically derived records of stream activity (Case and MacDonald, 2003). These droughts have also been recorded in the American Southwest (Stine, 1994; Benson et al., 2002; Yuan et al., 2004; Kleppe et al., 2011). In the far Pacific Northwest, and portions of the northern Rockies, a number of studies provide evidence for regional medieval wetness (Karst-Riddoch et al., 2005; Edwards et al., 2008; Bracht-Flyr and Fritz, 2012; Steinman et al., 2012).

5.4. Warming, Water Table Depth and Long-Term Apparent Carbon Accumulation

During the mid-Holocene increases in WTD detected by JBL7 and corroborated by other records, occur at the same time as regional increases in LARCA at 4600 yrBP and 2550 yrBP (Fig. 7). The HTM peak in moisture was associated with only a modest increase in LARCA values in JBL7 (Fig. 2). However, a compilation of mid-Holocene LARCA values from a number of sites spanning the JBL and Boreal Shield regions indicates anomalously high LARCA in aggregate during the HTM (Fig. 8; Holmquist and MacDonald, 2014). Data are still sparse, and high LARCA in the JBL during this time period may alternatively be associated with synchronous autogenic events.

The JBL7 the 4600 yrBP peak in moisture was also associated with a transition from *Sphagnum*- to herbaceous-dominated peat. This is a rare form of reverse succession but has been observed in one other paleoecological record in the Quebec sector of the

JBL (van Bellen et al., 2013), and in modern observations (Bauer et al., 2003) and peatland manipulations (Mitchell and Niering et al., 1993; Lamers et al., 2002). Pool expansion, transitions away from forested wetlands to open bog and poor fens, and other types of non-linear ecosystem behavior may be more common in the future if MAP increases and permafrost melt becomes more expansive. If *Carex*-dominated sections of ombrotrophic bogs expand they could potentially release more methane (Turetsky et al., 2014). Future studies should model this behavior as hysteresis between alternative stable states in order to determine its potential likelihood under future wetter conditions. Between the 4600 yrBP wet period and the 2550 yrBP wet shift, a dry period resulted in high oxygen exposure and high peat decay (Philben et al., 2015).

Increased moisture and temperature following 2550 yrBP may have had more of an effect on LARCA than earlier temperature and moisture anomalies during the HTM. Holmquist and MacDonald (2014) documented increased LARCA following 2550 yrBP especially in cores from the JBL, as opposed to the Boreal Shield. This time period is generally thought to be a point where precipitation increased (Bunbury et al., 2012; O'Reilly et al., 2014). According to pollen records it may have also been relatively warm in the JBL starting about 2400 yrBP (O'Reilly et al., 2014). This possibly indicates drivers similar to the 4600 yrBP warm peak acting on peatlands during the RWP that were more developed, more of them were bogs, and likely to be sensitive to external drivers rather than autogenic forces.

WTD and temperature fluctuations during the MCA and LIA did not have an apparent effect on LARCA (Fig. 8). Only in JBL4 did WTD and TA zones seem to indicate low LARCA during the cool LIA and high LARCA associated with the MCA

before droughts set in at that particular site (Fig. 4). Globally the MCA has statistically higher LARCA than the LIA (Charman et al., 2013). However Holmquist and MacDonald (2014) did not report any statistical difference in the eight-core transect from which the four cores presented here originate. It may be that past WTD fluctuations did not move outside the temperature-precipitation space required for optimal peat formation, and autogenic factors were responsible for peat accumulation and preservation at these time scales.

6. Summary and Conclusion

Increased precipitation in the HBL and JBL during the HTM, 2550 yrBP, and MCA periods is consistent with current knowledge about temperature-precipitation correlation in the boreal zones. Based on the 7320 yrBP-long reconstruction, the HTM was generally wet, although not as wet as the RWP. This record shows high synchronicity with the other reconstructions in the region. The MCA was also generally wet especially for the two most southern sites (JBL8 and JBL2), and excluding the most northern site (JBL4). This is consistent with patterns of tropical Pacific forcing on North American precipitation anomalies in both proxy and instrumental datasets. Although wet conditions dominated during these warm periods, LARCA increases during the HTM are not apparent in all records, indicating that local autogenic forces may have dominated LARCA at some sites. A long-term establishment of stable wet conditions following 2550 yrBP may be indicative of a broader regional pattern that increased carbon accumulation and storage. The MCA, despite being wet and warm, did not have an apparent effect on LARCA. It could be that the temperature and WTD fluctuations in the peatlands were not significant enough to perturb peat production and/or preservation. If

the future climate is as wet as it was during the HTM and MCA, then there is the potential for peatlands to grow at average rates and perhaps even faster and store similar amounts or more soil C due to longer growing seasons and more available photosynthetic radiation. Under this scenario increased methane emissions due to increased anaerobic activity, dissolved and frozen methane released from permafrost thaws, and potential ecosystem transitions to more poor-fens could decrease or reverse any possible cooling a more active CO₂ sink in bogs would create. Alternatively, if recent drying recorded by northern sites especially is indicative of a change in precipitation-evaporation balance caused by increased moisture deficit over the growing season, then present and future warming may reduce peatland C stocks as drying increases aerobic decay over growing seasons. Additional data from this region, synthesis, and modeling efforts are needed to predict whether peatlands of the Boreal Shield and JBL of Ontario will act as positive, negative, or net-neutral feedbacks to future warming under potential precipitation changes.

Acknowledgements

This research was made possible by grants from the National Science Foundation (NSF 0843685; NSF-0628598). The authors thank to David Beilman for collaboration on the 2008 fieldwork and to Konstantine Kreminitsky, Nicolai Kondov, Karly Wagner, Alvin Li, Julianne Lee, Michelle Lim, and Loren Quintana for laboratory assistance. The authors also thank two anonymous reviewers for their contributions towards improving this manuscript.

Works Cited

- Amesbury, M. J., Mallon, G., Charman, D. J., Hughes, P. D., Booth, R. K., Daley, T. J., & Garneau, M. (2013). Statistical testing of a new testate amoeba- based transfer function for water- table depth reconstruction on ombrotrophic peatlands in north- eastern Canada and Maine, United States. *Journal of Quaternary Science*, 28(1), 27-39.
- Anderson, L. (2011). Holocene record of precipitation seasonality from lake calcite $\delta^{18}O$ in the central Rocky Mountains, United States. *Geology*, 39(3), 211-214.
- Bauer, I. E., Gignac, L. D., & Vitt, D. H. (2003). Development of a peatland complex in boreal western Canada: lateral site expansion and local variability in vegetation succession and long-term peat accumulation. *Canadian Journal of Botany*, 81(8), 833-847.
- Beilman, D. W., MacDonald, G. M., Smith, L. C., & Reimer, P. J. (2009). Carbon accumulation in peatlands of West Siberia over the last 2000 years. *Global Biogeochemical Cycles*, 23(1).
- Bennett, K. D. (1996). Determination of the number of zones in a biostratigraphical sequence. *New Phytologist*, 132(1), 155-170.
- Benson, L., Kashgarian, M., Rye, R., Lund, S., Paillet, F., Smoot, J., . . . Lindström, S. (2002). Holocene multidecadal and multicentennial droughts affecting Northern California and Nevada. *Quaternary Science Reviews*, 21(4), 659-682.
- Beringer, J., Tapper, N. J., McHugh, I., Chapin, F., Lynch, A. H., Serreze, M. C., & Slater, A. (2001). Impact of Arctic treeline on synoptic climate. *Geophysical Research Letters*, 28(22), 4247-4250.

550 Blaauw, M. (2010). Methods and code for ‘classical’ age-modelling of radiocarbon
551 sequences. *Quaternary Geochronology*, 5(5), 512-518.

552 Blaauw, M., & Christen, J. A. (2011). Flexible paleoclimate age-depth models using an
553 autoregressive gamma process. *Bayesian Analysis*, 6(3), 457-474.

554 Booth, R., Jackson, S., & Notaro, M. (2010a). Using peatland archives to test
555 paleoclimate hypotheses. *Science*, 22, 209-221.

556 Booth, R., Lamentowicz, M., & Charman, D. (2010b). Preparation and analysis of testate
557 amoebae in peatland paleoenvironmental studies. *Mires and Peat*, 7(02), 1-7.

558 Booth, R. K. (2008). Testate amoebae as proxies for mean annual water- table depth in
559 Sphagnum- dominated peatlands of North America. *Journal of Quaternary*
560 *Science*, 23(1), 43-57.

561 Booth, R. K., Jackson, S. T., Forman, S. L., Kutzbach, J. E., Bettis, E., Kreigs, J., &
562 Wright, D. K. (2005). A severe centennial-scale drought in midcontinental North
563 America 4200 years ago and apparent global linkages. *The Holocene*, 15(3), 321-
564 328.

565 Booth, R. K., Jackson, S. T., Sousa, V. A., Sullivan, M. E., Minckley, T. A., & Clifford,
566 M. J. (2012). Multi-decadal drought and amplified moisture variability drove
567 rapid forest community change in a humid region. *Ecology*, 93(2), 219-226.

568 Booth, R. K., Notaro, M., Jackson, S. T., & Kutzbach, J. E. (2006). Widespread drought
569 episodes in the western Great Lakes region during the past 2000 years: geographic
570 extent and potential mechanisms. *Earth and Planetary Science Letters*, 242(3),
571 415-427.

572 Bracht-Flyr, B., & Fritz, S. C. (2012). Synchronous climatic change inferred from diatom
573 records in four western Montana lakes in the US Rocky Mountains. *Quaternary*
574 *Research*, 77(3), 456-467.

575 Bridgham, S. D., Megonigal, J. P., Keller, J. K., Bliss, N. B., & Trettin, C. (2006). The
576 carbon balance of North American wetlands. *Wetlands*, 26(4), 889-916.

577 Brown, J., Ferrians Jr, O., Heginbottom, J., & Melnikov, E. (1998). Circum-Arctic map
578 of permafrost and ground-ice conditions, revised February 2001. National Snow
579 and Ice Data Center/World Data Center for Glaciology. *Digital Media, Boulder,*
580 *CO.*

581 Bunbury, J., Finkelstein, S. A., & Bollmann, J. (2012). Holocene hydro-climatic change
582 and effects on carbon accumulation inferred from a peat bog in the Attawapiskat
583 River watershed, Hudson Bay Lowlands, Canada. *Quaternary Research*, 78(2),
584 275-284.

585 Campbell, I., Last, W. M., Campbell, C., Clare, S., & McAndrews, J. H. (2000). The late
586 Holocene paleohydrology of Pine Lake, Alberta: a multiproxy investigation.
587 *Journal of Paleolimnology*, 24(4), 427-441.

588 Case, R. A., & MacDonald, G. M. (2003). Tree Ring Reconstructions Of Streamflow For
589 Three Canadian Prairie Rivers. *Journal of the American Water Resources*
590 *Association*, 39(3), 703-716.

591 Charman, D., Beilman, D., Blaauw, M., Booth, R., Brewer, S., Chambers, F., . . . Hughes,
592 P. (2013). Climate-related changes in peatland carbon accumulation during the
593 last millennium. *Biogeosciences*, 10(2), 929-944.

594 Charman, D. J. (2001). Biostratigraphic and palaeoenvironmental applications of testate
595 amoebae. *Quaternary Science Reviews*, 20(16), 1753-1764.

596 Clifford, M. J., & Booth, R. K. (2013). Increased probability of fire during late Holocene
597 droughts in northern New England. *Climatic change*, 119(3-4), 693-704.

598 Clymo, R., Turunen, J., & Tolonen, K. (1998). Carbon accumulation in peatland. *Oikos*,
599 368-388.

600 Cook, E. R., Woodhouse, C. A., Eakin, C. M., Meko, D. M., & Stahle, D. W. (2004).
601 Long-term aridity changes in the western United States. *Science*, 306(5698),
602 1015-1018.

603 Dise, N. B. (2009). Peatland response to global change. *Science*, 326(5954), 810.

604 Edwards, T. W. D., Birks, S. J., Luckman, B. H., & MacDonald, G. M. (2008). Climatic
605 and hydrologic variability during the past millennium in the eastern Rocky
606 Mountains and northern Great Plains of western Canada. *Quaternary Research*,
607 70(2), 188-197.

608 Edwards, T. W. D., Wolfe, B. B., & Macdonald, G. M. (1996). Influence of Changing
609 Atmospheric Circulation on Precipitation $\delta^{18}\text{O}$ -Temperature Relations in Canada
610 during the Holocene. *Quaternary Research*, 46(3), 211-218.

611 Ersek, V., Clark, P. U., Mix, A. C., Cheng, H., & Edwards, R. L. (2012). Holocene winter
612 climate variability in mid-latitude western North America. *Nature*
613 *communications*, 3, 1219.

614 Far North Science Advisory Panel (2010). *Science for a Changing North: Report to the*
615 *Ontario Ministry of Natural Resources*.

616 Feng, S., Oglesby, R. J., Rowe, C. M., Loope, D. B., & Hu, Q. (2008). Atlantic and
617 Pacific SST influences on Medieval drought in North America simulated by the
618 Community Atmospheric Model. *Journal of Geophysical Research: Atmospheres*
619 (1984–2012), 113(D11).

620 Fenner, N., & Freeman, C. (2011). Drought-induced carbon loss in peatlands. *Nature*
621 *Geoscience*, 4(12), 895-900.

622 Filion, L. (1987). Holocene development of parabolic dunes in the Central St. Lawrence
623 Lowland, Québec. *Quaternary Research*, 28(2), 196-209.

624 Fritz, S. C., Ito, E., Yu, Z., Laird, K. R., & Engstrom, D. R. (2000). Hydrologic variation
625 in the northern Great Plains during the last two millennia. *Quaternary Research*,
626 53(2), 175-184.

627 Galloway, J. M., Lenny, A. M., & Cumming, B. F. (2011). Hydrological change in the
628 central interior of British Columbia, Canada: diatom and pollen evidence of
629 millennial-to-centennial scale change over the Holocene. *Journal of*
630 *Paleolimnology*, 45(2), 183-197.

631 Glaser, P. H., Hansen, B., Siegel, D. I., Reeve, A. S., & Morin, P. J. (2004). Rates,
632 pathways and drivers for peatland development in the Hudson Bay Lowlands,
633 northern Ontario, Canada. *Journal of Ecology*, 92(6), 1036-1053.

634 Hallett, D. J., Mathewes, R. W., & Walker, R. C. (2003). A 1000-year record of forest
635 fire, drought and lake-level change in southeastern British Columbia, Canada. *The*
636 *Holocene*, 13(5), 751-761.

637 Hargan, K. E., Rühland, K. M., Paterson, A. M., Holmquist, J., MacDonald, G. M.,
638 Bunbury, J., . . . Smol, J. P. (2015). Long-term successional changes in peatlands

639 of the Hudson Bay Lowlands, Canada inferred from the ecological dynamics of
 640 multiple proxies. *The Holocene*, 25(1), 92-107.

641 Hass, H. C. (1996). Northern Europe climate variations during late Holocene: evidence
 642 from marine Skagerrak. *Palaeogeography, Palaeoclimatology, Palaeoecology*,
 643 123(1), 121-145.

644 Herweijer, C., & Seager, R. (2008). The global footprint of persistent extra- tropical
 645 drought in the instrumental era. *International Journal of Climatology*, 28(13),
 646 1761-1774.

647 Holmquist, J. R., & MacDonald, G. M. (2014). Peatland succession and long-term
 648 apparent carbon accumulation in central and northern Ontario, Canada. *The*
 649 *Holocene*, 24(9), 1075-1089.

650 Holmquist, J. R., MacDonald, G. M., & Gallego-Sala, A. (2014). Peatland initiation,
 651 carbon accumulation, and 2 ka depth in the James Bay Lowland and adjacent
 652 regions. *Arctic, Antarctic, and Alpine Research*, 46(1), 19-39.

653 Ji, J., Shen, J., Balsam, W., Chen, J., Liu, L., & Liu, X. (2005). Asian monsoon
 654 oscillations in the northeastern Qinghai–Tibet Plateau since the late glacial as
 655 interpreted from visible reflectance of Qinghai Lake sediments. *Earth and*
 656 *Planetary Science Letters*, 233(1), 61-70.

657 Jia, G. J., Epstein, H. E., & Walker, D. A. (2009). Vegetation greening in the Canadian
 658 Arctic related to decadal warming. *Journal of Environmental Monitoring*, 11(12),
 659 2231-2238.

660 Juggins, S. (2009). Rioja: analysis of Quaternary science data. R package version 0.5-6.
 661 URL [http://cran.r-project.org/package= rioja](http://cran.r-project.org/package=rioja).

662 Karst-Riddoch, T. L., Pisaric, M. F., Youngblut, D. K., & Smol, J. P. (2005). Postglacial
 663 record of diatom assemblage changes related to climate in an alpine lake in the
 664 northern Rocky Mountains, Canada. *Canadian Journal of Botany*, 83(8), 968-982.
 665 Kaufman, D., Ager, T., Anderson, N., Anderson, P., Andrews, J., Bartlein, P., . . . Duvall,
 666 M. (2004). Holocene thermal maximum in the western Arctic (0–180 W).
 667 *Quaternary Science Reviews*, 23(5), 529-560.
 668 Kleppe, J., Brothers, D. S., Kent, G. M., Biondi, F., Jensen, S., & Driscoll, N. W. (2011).
 669 Duration and severity of Medieval drought in the Lake Tahoe Basin. *Quaternary*
 670 *Science Reviews*, 30(23), 3269-3279.
 671 Ladd, M. J., & Gajewski, K. (2010). The North American summer Arctic front during
 672 1948–2007. *International Journal of Climatology*, 30(6), 874-883.
 673 Laiho, R. (2006). Decomposition in peatlands: reconciling seemingly contrasting results
 674 on the impacts of lowered water levels. *Soil Biology and Biochemistry*, 38(8),
 675 2011-2024.
 676 Laird, K. R., & Cumming, B. F. (2009). Diatom-inferred lake level from near-shore cores
 677 in a drainage lake from the Experimental Lakes Area, northwestern Ontario,
 678 Canada. *Journal of Paleolimnology*, 42(1), 65-80.
 679 Lamarre, A., Garneau, M., & Asnong, H. (2012). Holocene paleohydrological
 680 reconstruction and carbon accumulation of a permafrost peatland using testate
 681 amoeba and macrofossil analyses, Kuujjuarapik, subarctic Québec, Canada.
 682 *Review of Palaeobotany and Palynology*, 186, 131-141.
 683 Lamers, L. P., Smolders, A. J., & Roelofs, J. G. (2002). The restoration of fens in the
 684 Netherlands. *Hydrobiologia*, 478(1-3), 107-130.

685 Loisel, J., Gallego-Sala, A., & Yu, Z. (2012). Global-scale pattern of peatland Sphagnum
 686 growth driven by photosynthetically active radiation and growing season length.
 687 *Biogeosciences Discussions*, 9(2), 2169-2196.

688 Loisel, J., & Garneau, M. (2010). Late Holocene paleoecohydrology and carbon
 689 accumulation estimates from two boreal peat bogs in eastern Canada: Potential
 690 and limits of multi-proxy archives. *Palaeogeography, Palaeoclimatology,*
 691 *Palaeoecology*, 291(3), 493-533.

692 MacDonald, G. M., & Case, R. A. (2005). Variations in the Pacific Decadal Oscillation
 693 over the past millennium. *Geophysical Research Letters*, 32(8).

694 MacDonald, G. M., Szeicz, J. M., Claricoates, J., & Dale, K. A. (1998). Response of the
 695 central Canadian treeline to recent climatic changes. *Annals of the Association of*
 696 *American Geographers*, 88(2), 183-208.

697 Mann, M. E., Zhang, Z., Rutherford, S., Bradley, R. S., Hughes, M. K., Shindell, D., . . .
 698 Ni, F. (2009). Global signatures and dynamical origins of the Little Ice Age and
 699 Medieval Climate Anomaly. *Science*, 326(5957), 1256-1260.

700 Markel, E. R., Booth, R. K., & Qin, Y. (2010). Testate amoebae and $\delta^{13}\text{C}$ of Sphagnum
 701 as surface-moisture proxies in Alaskan peatlands. *The Holocene*, 20(3), 463-475.

702 Martín-Puertas, C., Valero-Garcés, B. L., Mata, M. P., González-Sampériz, P., Bao, R.,
 703 Moreno, A., & Stefanova, V. (2008). Arid and humid phases in southern Spain
 704 during the last 4000 years: the Zonar Lake record, Cordoba. *The Holocene*, 18(6),
 705 907-921.

706 McCune, B., Grace, J. B., & Urban, D. L. (2002). *Analysis of ecological communities*
 707 (Vol. 28): MjM software design Gleneden Beach, Oregon.

708 Mitchell, C. C., & Niering, W. A. (1993). Vegetation change in a topogenic bog
 709 following beaver flooding. *Bulletin of the Torrey Botanical Club*, 136-147.

710 Mitchell, E. A., Payne, R. J., & Lamentowicz, M. (2008). Potential implications of
 711 differential preservation of testate amoeba shells for paleoenvironmental
 712 reconstruction in peatlands. *Journal of Paleolimnology*, 40(2), 603-618.

713 NOAA. (2013). GPCC Precipitation V6 Combined Precipitation (mm) Composite
 714 Anomaly, 1981-2010 climate. Retrieved 17 April 2013

715 NOAA. (2013). Multivariate El Niño Index (MEI). Retrieved 12 June 2013

716 O'Reilly, B. C., Finkelstein, S. A., & Bunbury, J. (2014). Pollen-Derived Paleovegetation
 717 Reconstruction and Long-Term Carbon Accumulation at a Fen Site in the
 718 Attawapiskat River Watershed, Hudson Bay Lowlands, Canada. *Arctic, Antarctic,
 719 and Alpine Research*, 46(1), 6-18.

720 Oglesby, R., Feng, S., Hu, Q., & Rowe, C. (2012). The role of the Atlantic Multidecadal
 721 Oscillation on medieval drought in North America: Synthesizing results from
 722 proxy data and climate models. *Global and Planetary Change*, 84, 56-65.

723 Ortega, P., Lehner, F., Swingedouw, D., Masson-Delmotte, V., Raible, C. C., Casado,
 724 M., & Yiou, P. (2015). A model-tested North Atlantic Oscillation reconstruction
 725 for the past millennium. *Nature*, 523(7558), 71-74.

726 Packalen, M. S., Finkelstein, S. A., & McLaughlin, J. W. (2014). Carbon storage and
 727 potential methane production in the Hudson Bay Lowlands since mid-Holocene
 728 peat initiation. *Nature Communications*, 5(4078). doi: doi:10.1038/ncomms5078

729 Paul, C., Douglas, M., & Smol, J. (2010). Diatom-inferred Holocene climatic and
 730 environmental changes in an unusually subsaline high Arctic nunatak pond on
 731 Ellesmere Island (Nunavut, Canada). *Journal of Paleolimnology*, 44(4), 913-929.
 732 Payne, R., Gauci, V., & Charman, D. J. (2010). The impact of simulated sulfate
 733 deposition on peatland testate amoebae. *Microbial ecology*, 59(1), 76-83.
 734 Payne, R. J., & Mitchell, E. A. (2009). How many is enough? Determining optimal count
 735 totals for ecological and palaeoecological studies of testate amoebae. *Journal of*
 736 *Paleolimnology*, 42(4), 483-495.
 737 Philben, M., Holmquist, J., MacDonald, G., Duan, D., Kaiser, K., & Benner, R. (2015).
 738 Temperature and oxygen controls on decomposition in a James Bay peatland.
 739 *Global Biogeochemical Cycles*. doi: 10.1002/2014GB004989
 740 Price, D. T., Alfaro, R., Brown, K., Flannigan, M., Fleming, R., Hogg, E., . . .
 741 McKenney, D. (2013). Anticipating the consequences of climate change for
 742 Canada's boreal forest ecosystems 1. *Environmental Reviews*, 21(4), 322-365.
 743 R Core Development Team (2012). R: a language and environment for statistical
 744 computing. ver. 2.13.0: R Foundation for Statistical Computing Vienna, Austria.
 745 Reimer, P. J., Bard, E., Bayliss, A., Beck, J. W., Blackwell, P. G., Ramsey, C. B., . . .
 746 Hajdas, I. (2013). IntCal13 and Marine13 radiocarbon age calibration curves 0–
 747 50,000 years cal BP. *Radiocarbon*, 55(4), 1869-1887.
 748 Renssen, H., Seppä, H., Heiri, O., Roche, D., Goosse, H., & Fichet, T. (2009). The
 749 spatial and temporal complexity of the Holocene thermal maximum. *Nature*
 750 *Geoscience*, 2(6), 411-414.

751 Riley, J. (2011). Wetlands of the Ontario Hudson Bay Lowland: a regional overview.
752 *Nature Conservancy of Canada: Toronto, Ontario, Canada, 156.*

753 Robroek, B. J., Schouten, M. G., Limpens, J., Berendse, F., & Poorter, H. (2009).
754 Interactive effects of water table and precipitation on net CO₂ assimilation of
755 three co- occurring Sphagnum mosses differing in distribution above the water
756 table. *Global Change Biology, 15*(3), 680-691.

757 Rühland, K., Pienitz, A., & Smol, J. P. (2003). Paleolimnological evidence from diatoms
758 for recent environmental changes in 50 lakes across Canadian Arctic treeline.
759 *Arctic, Antarctic, and Alpine Research, 35*(1), 110-123.

760 Schindler, D. W., & Smol, J. P. (2006). Cumulative effects of climate warming and other
761 human activities on freshwaters of arctic and subarctic North America. *AMBIO: A*
762 *Journal of the Human Environment, 35*(4), 160-168.

763 Seager, R., Burgman, R., Kushnir, Y., Clement, A., Cook, E., Naik, N., & Miller, J.
764 (2008). Tropical Pacific Forcing of North American Medieval Megadroughts:
765 Testing the Concept with an Atmosphere Model Forced by Coral-Reconstructed
766 SSTs. *Journal of Climate, 21*(23), 6175-6190.

767 Seidenkrantz, M.-S., Aagaard-Sørensen, S., Sulsbrück, H., Kuijpers, A., Jensen, K. G., &
768 Kunzendorf, H. (2007). Hydrography and climate of the last 4400 years in a SW
769 Greenland fjord: implications for Labrador Sea palaeoceanography. *The*
770 *Holocene, 17*(3), 387-401.

771 Shabbar, A., Bonsal, B., & Khandekar, M. (1997). Canadian precipitation patterns
772 associated with the Southern Oscillation. *Journal of Climate, 10*(12), 3016-3027.

773 Shindell, D. T., Schmidt, G. A., Miller, R. L., & Mann, M. E. (2003). Volcanic and solar
774 forcing of climate change during the preindustrial era. *Journal of Climate*, 16(24),
775 4094-4107.

776 Steinman, B. A., Abbott, M. B., Mann, M. E., Stansell, N. D., & Finney, B. P. (2012).
777 1,500 year quantitative reconstruction of winter precipitation in the Pacific
778 Northwest. *Proceedings of the National Academy of Sciences*, 109(29), 11619-
779 11623.

780 Stevens, L. R., Stone, J. R., Campbell, J., & Fritz, S. C. (2006). A 2200-yr record of
781 hydrologic variability from Foy Lake, Montana, USA, inferred from diatom and
782 geochemical data. *Quaternary Research*, 65(2), 264-274.

783 Stine, S. (1994). Extreme and persistent drought in California and Patagonia during
784 mediaeval time. *Nature*, 369(6481), 546-549.

785 Strack, M., Waddington, J., Lucchese, M., & Cagampan, J. (2009). Moisture controls on
786 CO₂ exchange in a Sphagnum- dominated peatland: results from an extreme
787 drought field experiment. *Ecohydrology*, 2(4), 454-461.

788 Sullivan, M. E., & Booth, R. K. (2011). The potential influence of short-term
789 environmental variability on the composition of testate amoeba communities in
790 Sphagnum peatlands. *Microbial ecology*, 62(1), 80-93.

791 Tarnocai, C., & Bockheim, J. (2011). Cryosolic soils of Canada: Genesis, distribution,
792 and classification. *Canadian Journal of Soil Science*, 91(5), 749-762.

793 Tian, J., Nelson, D. M., & Hu, F. S. (2006). Possible linkages of late-Holocene drought in
794 the North American midcontinent to Pacific Decadal Oscillation and solar
795 activity. *Geophysical Research Letters*, 33(23), L23702.

796 Trouet, V., Esper, J., Graham, N. E., Baker, A., Scourse, J. D., & Frank, D. C. (2009).
 797 Persistent positive North Atlantic Oscillation mode dominated the medieval
 798 climate anomaly. *Science*, 324(5923), 78-80.
 799 Turetsky, M. R., Kotowska, A., Bubier, J., Dise, N. B., Crill, P., Hornibrook, E. R., . . .
 800 Nykänen, H. (2014). A synthesis of methane emissions from 71 northern,
 801 temperate, and subtropical wetlands. *Global Change Biology*.
 802 van Bellen, S., Dallaire, P. L., Garneau, M., & Bergeron, Y. (2011). Quantifying spatial
 803 and temporal Holocene carbon accumulation in ombrotrophic peatlands of the
 804 Eastmain region, Quebec, Canada. *Global Biogeochemical Cycles*, 25(2).
 805 van Bellen, S., Garneau, M., Ali, A. A., Lamarre, A., Robert, É. C., Magnan, G., . . .
 806 Pratte, S. (2013). Poor fen succession over ombrotrophic peat related to late
 807 Holocene increased surface wetness in subarctic Quebec, Canada. *Journal of*
 808 *Quaternary Science*, 28(8), 748-760.
 809 Vance, R., Last, W., & Smith, A. (1997). Hydrologic and climatic implications of a
 810 multidisciplinary study of late Holocene sediment from Kenosee Lake,
 811 southeastern Saskatchewan, Canada. *Journal of Paleolimnology*, 18(4), 365-393.
 812 Vance, R., Mathewes, R., & Clague, J. (1992). 7000 year record of lake-level change on
 813 the northern Great Plains: A high-resolution proxy of past climate. *Geology*,
 814 20(10), 879-882.
 815 Viau, A., & Gajewski, K. (2009). Reconstructing millennial-scale, regional paleoclimates
 816 of boreal Canada during the Holocene. *Journal of Climate*, 22(2).
 817 Wang, T., Surge, D., & Mithen, S. (2012). Seasonal temperature variability of the
 818 Neoglacial (3300–2500BP) and Roman Warm Period (2500–1600BP)

819 reconstructed from oxygen isotope ratios of limpet shells (*Patella vulgata*),
820 Northwest Scotland. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 317,
821 104-113.

822 Yu, Z., Beilman, D. W., & Jones, M. C. (2009). Sensitivity of northern peatland carbon
823 dynamics to Holocene climate change. *Carbon cycling in northern peatlands*, 55-
824 69.

825 Yuan, F., Linsley, B. K., Lund, S. P., & McGeehin, J. P. (2004). A 1200 year record of
826 hydrologic variability in the Sierra Nevada from sediments in Walker Lake,
827 Nevada. *Geochemistry, Geophysics, Geosystems*, 5(3).

828

829 **Figure 1:** A map of North America including the USA and Canada displays the locations
830 of JBL4, JBL7, JBL2, and JBL8 as well as other reports of medieval surface moisture
831 anomalies in the vicinity of the JBL and Boreal Shield of Ontario, the Quebec JBL, and
832 the Great Lakes area. The results of a brief review of studies citing medieval surface
833 moisture anomalies (Tab. 4) are also displayed, along with modern data on La Niña-year
834 precipitation anomaly (NOAA, 2013a, 2013b) and the modern location of the Arctic front
835 (Ladd and Gajewski, 2009).

836

837 **Figure 2:** Stratigraphy data for JBL7 during the mid- and late-Holocene including: depth,
838 age, notable TA species (%), WTD reconstructions with upper and lower standard error
839 estimates, NMS axis 1 and 2, the absolute values of the residuals of a linear model with
840 NMS axis 1 as the independent variable and WTD as the dependent variable, and TA
841 zones based on a cluster analysis. Acrotelm and catotelm LARCA, dominant

macrofossils, and acrotelm-catotelm boundaries were originally published in Holmquist and MacDonald (2014). **D. pristis* type was excluded from transfer function analyses. Grey bars indicate areas with very low-test counts.

Figure 3: Stratigraphy data for JBL4 including: depth, age, notable TA species (%), WTD reconstructions with upper and lower standard error estimates, NMS axis 1 and 2, the absolute values of the residuals of a linear model with NMS axis 2 as the independent variable and WTD as the dependent variable, and TA zones based on a cluster analysis. Acrotelm and catotelm LARCA, dominant macrofossils, and acrotelm-catotelm boundaries were originally published in Holmquist and MacDonald (2014). **D. pristis* type was excluded from transfer function analyses.

Figure 4: Stratigraphy data for the mid and late-Holocene for JBL2 including: depth, age, notable TA species (%), WTD reconstructions with upper and lower standard error estimates in grey, NMS axis 1 and 2, the absolute values of the residuals of a linear model with NMS axis 1 as the independent variable and WTD as the dependent variable, and TA zones based on a cluster analysis. Acrotelm and catotelm LARCA, dominant macrofossils, and acrotelm-catotelm boundaries were originally published in Holmquist and MacDonald (2014). **D. pristis* type was excluded from transfer function analyses.

Figure 5: Stratigraphy data for JBL8 including: depth, age, notable TA species (%), WTD reconstructions with upper and lower standard error estimates in grey, NMS axis 1 and 2, the absolute values of the residuals of a linear model with NMS axis 1 as the

independent variable and WTD as the dependent variable, and TA zones based on a cluster analysis. Acrotelm and catotelm LARCA, dominant macrofossils, and acrotelm-catotelm boundaries were originally published in Holmquist and MacDonald (2014). **Figure 6:** NMS Axes and WTD reconstructions were correlated to verify that WTD was the main factor affecting testate amoebae community composition. The amount of variance explained by axis 1 or axis 2 is reported. Trend-lines indicate significant correlations ($p < 0.05$). P-values and R^2 values are reported in Tab. 3.

Figure 7: JBL7 WTD shows some synchronicities with the one other mid-Holocene TA-based WTD record in the region (Bunbury et al., 2012), as well with a JBL pollen-based reconstruction of MAT and TAP (O'Reilly et al., 2014), and Mean June, July, August Temperature (TJJA; Bunbury et al., 2012). Broken black lines indicate TA zone transitions. Triangles indicate notable wet or dry shifts discussed in the text. Both TA records show increased moisture and/or ecological transitions occurring during peak relative humidity with respect to present ($\Delta RH\%$ wrt present) in Central Canada (Edwards et al., 1996). These are compared to a 500-year binned compilation of the catotelm LARCA z-scores from eight JBL and Boreal Shield peatlands (Holmquist and MacDonald, 2014).

Figure 8: Mean WTD reconstructions for JBL4, JBL7, JBL2, and JBL8 are plotted over the late-Holocene. These results are compared to VC04-06 (Bunbury et al., 2012). The approximate timing of the MCA and LIA are presented. Broken lines indicate TA zone transitions.

888

889 **Table 1:** A list of taxa (Booth, 2008) that were combined in this analysis because of
890 ecological and morphological similarity, and for consistency with the calibration dataset.

891

892 **Table 2:** Zones, age-ranges, dominant taxa, and WTD descriptive statistics for four cores
893 in the JBL. **D. pristis* type, were omitted from transfer function analyses.

894

895 **Table 3:** Statistics for linear models with NMS axis 1 and 2 as the independent variable
896 and reconstructed WTD as the dependent variable.

897 **Table 4:** The results of a literature review of reported moisture conditions across various
898 proxies in North America during the MCA.

Figure 1

Figure 2

JBL7 (54°23'43" N, 89°31'20" W)

Figure 3

JBL4 (55°16'09" N, 88°55'50" W)

Figure 4

JBL2 (52°01'07" N, 90°07'53" W)

Figure 5

JBL8 (50°27'33" N, 89°55'42" W)

Figure 6

Figure 8

Table 1

Booth, 2008	Combined Categories
<i>Heleopera rosea</i> , <i>Heleopera sphagni</i> , and <i>Heleopera perticola</i>	<i>Heleopera</i> spp.
<i>Amphitrema wrightianum</i> , <i>Amphitrema stenostoma</i>	<i>Amphitrema wrightianum</i> type
<i>Trig o nopyxis minuta</i> , <i>Trigonopyxis arcula</i>	<i>Trigonopyxis</i> spp.
<i>Trinema lineare</i> , <i>Corythion dubium</i> type	<i>Corythion dubium</i> type
<i>Nebela bohémica</i> , <i>Nebela collaris</i>	<i>Nebela bohémica</i> type

Table 2

Zone	Age Range (yrBP)	Major Taxa	WTD range (cm)	WTD mean (cm)
JBL4-A	1610 - 1180	<i>Archerella flavum</i> , <i>Diffflugia pulex</i>	9 - 14	11 ± 2
JBL4-B	1180 - 800	<i>A. flavum</i> , <i>Bullinularia indica</i> , <i>D. pulex</i>	9 - 22	13 ± 4
JBL4-C	800 - 270	<i>Diffflugia pristis</i> type*, <i>D. pulex</i>	10 - 20	13 ± 3
JBL4-D	270 to -58	<i>A. flavum</i> , <i>Hyalosphenia elegans</i> , <i>Hyalosphenia papilio</i> , <i>Nebela militaris</i>	10 - 23	15 ± 3
JBL7-A	7320 - 2550	<i>Assulina muscorum</i> , <i>Centropyxis platystoma</i> , <i>Cyclopyxis arcelloides</i> , <i>Heleopera</i> spp., <i>Hyalosphenia papilio</i> , <i>Phryganella acropodia</i> , <i>Trigonopyxis</i> spp.	7 - 36	18 ± 7
JBL7-B	2550 - 1280	<i>Amphitrema wrightianum</i> type, <i>A. flavum</i> , <i>D. pulex</i> , <i>Heleopera</i> spp.	7 - 14	11 ± 2
JBL7-C	1280 - 660	<i>Arcella catinus</i> type, <i>Arcella vulgaris</i> , <i>A. flavum</i> , <i>D. pristis</i> type*, <i>D. pulex</i>	7 - 15	12 ± 2
JBL7-D	660 - 260	<i>A. flavum</i> , <i>D. pulex</i> , <i>D. pristis</i> type*	9 - 22	13 ± 3
JBL7-E	260 to -58	<i>A. muscorum</i> , <i>H. papilio</i> , <i>N. militaris</i>	13 - 28	19 ± 5
JBL2-A	1050 - 870	<i>A. wrightianum</i> type	-6 to -2	-3 ± 1
JBL2-B	870 - 500	<i>A. wrightianum</i> type, <i>A. flavum</i>	-2 to 13	4 ± 4
JBL2-C	500 - 160	<i>A. flavum</i> , <i>D. pulex</i> , <i>Heleopera</i> spp.	9 - 17	12 ± 2
JBL2-D	160 to -58	<i>Diffflugia pristis</i> type*, <i>D. pulex</i> , <i>Hyalosphenia elegans</i> , <i>H. papilio</i>	8 - 18	14 ± 3
JBL8-A	1340 - 620	<i>A. flavum</i> , <i>D. pulex</i> , rotifer: <i>Habrotrocha angusticollis</i> , <i>Trigonopyxis</i> spp.	11 - 25	16 ± 4
JBL8-B	620 - 210	<i>A. flavum</i> , <i>H. angusticollis</i> , <i>Heleopera</i> spp., <i>Trigonopyxis</i> spp.	17 - 30	25 ± 4
JBL8-C	210 to -58	<i>Assulina muscorum</i> , <i>H. papilio</i> , <i>Trigonopyxis</i> spp.	23 - 37	29 ± 4

Table 3

Table 3

				WTD~NMS Axis 1		WTD~NMS Axis 2	
Core	NMS Axis 1 (%)	NMS Axis 2 (%)	Final Stress	P-Value	R ²	P-Value	R ²
JBL4	47	44	12.5	0.39	0.01	4.28E-07	0.37
JBL7	40	30	11	6.29E-09	0.33	6.32E-09	0.33
JBL2	63	28	13.6	4.72E-13	0.65	7.34E-09	0.49
JBL8	65	25	15.3	7.71E-05	0.29	<2.2E-16	0.77

Table 4

Table 4

Site Name	Lat	Long	Local	Citations	MCA (+ Wet or - Dry)	Explanation
Hole Bog	47.3	-94.24283	Great Lakes	Booth et al., 2005	-	Short droughts. 1000, 800, 700 yrBP
Minden Bog	43.606333	-82.83342	Great Lakes	Booth et al., 2005; Booth et al. 2012	-	High multi-decadal moisture variability characterizes the MCA. Regional and local wildfires are coincident with the drought inferred from Beech pollen declines.
Irwin Smith Bog	45.030839	-83.61706	Great Lakes	Booth et al., 2012	-	Bog surface burned in the early MCA. Later MCA was characterized by relatively high decadal variability and charred <i>Picea</i> spp. needles accompanying droughts.
Rawson Lake	49.666667	-93.73333	Great Lakes	Laird and Cummings, 2009	-	1060–820 yrBP: decline in lake level of 2.5 to 3.0 m.
Chappice Lake	50.166667	-110.3667	Great Plains	Vance et al., 1992	-	Laminated sediments record low-water stands related to significant droughts
Coldwater Lake	46.016667	-99.06667	Great Plains	Fritz et al., 2009	-	930 - 870 yrBP there was high diatom-inferred salinity followed by alternations between low and high.
Felker Lake	51.95	-122	Great Plains	Galloway et al. 2011	-	Reconstructed lake levels low during MCA
Kenosee Lake	49.816667	-102.3	Great Plains	Vance et al., 1997	-	1150-650 yrBP: Lake-level decline associated with episodes of increased salinity.
Moon Lake	46.85	-98.15	Great Plains	Fritz et al, 2009	-	Diatom inferred salinity reaches its peak ~850 yrBP. Prolonged aridity from 930 - 790 yrBP.
Pine Lake	52.066667	-113.45	Great Plains	Cambell et al., 2000	-	Water-level decline (<1 m) is suggested by the absence of <i>Typha</i> .
Rice Lake	48	-101.5333	Great Plains	Fritz et al., 2009	-	Mg/Ca correlates w/ salinity, and has a peak ~930-810 yrBP.
Steel Lake	46.966667	-94.68333	Great Plains	Tian et al, 2006	-	850-500 yrBP. ¹⁸ O record suggests that the region was relatively dry.