I personally would consider it an honour to be fossilized-

-Richard Dawkins, *The Blind Watchmaker*
RAPID CLIMATE CHANGE AND SPRUCE-CLIMATE DISEQUILIBRIUM IN LATEGLACIAL NOVA SCOTIA

by

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ABSTRACT

A network of 25 lateglacial-aged lakes was sampled in Nova Scotia to examine the relationship between climate and vegetation at the end of the last Ice Age. Nova Scotia lake sediments are sensitive to the rapid climate events that typify the lateglacial and possess some of the best records for millennial-decadal scale climate change outside the Greenland Ice Core records. The presence of spruce in lateglacial Nova Scotia has been previously estimated using the 20% spruce pollen limit. Using fossil stomates and plant macrofossils, I have refined this limit to 8%, potentially eliminating significant underestimates of the distribution of spruce. The 8% limit was then used to determine if spruce was in equilibrium with climate during the lateglacial at 4 time intervals. Evidence for spruce was found at numerous sites with chironomid inferred July temperatures as low as 13°C. Sites where the temperature was at, or above, 13°C but spruce was absent indicated that spruce was not in equilibrium with climate at that time interval as the result of a migration or population expansion lag. Significant portions of northeastern Nova Scotia and all of Cape Breton Island remained in vegetation-climate disequilibrium during the lateglacial, while a small portion of southwestern Nova Scotia also exhibited disequilibrium conditions.

While reconstructing the thermal regime in lateglacial Nova Scotia, evidence for the Older Dryas (GI-1d) and the recently discovered GI-1c2 was observed and confirmed with chironomid inferred temperature reversals >3 °C. The discovery of these two events indicates that brief cooling events typical of NW European records also occurred in North America. Furthermore, chironomid and LOI records from Nova Scotia also provide
evidence for a potential decade-scale cooling event analogous to the current slowdown in the rate of global warming. These data, indicating cooling between 1.6-6.4 °C, have been correlated to over 30 marine and terrestrial sites throughout the North Atlantic and beyond.
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Ian, my husband and best friend, demonstrated unlimited patience and generosity in his support for me. Without him, this project would never have been completed.

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Chapter 1

Introduction

Sediment cores from lakes in Nova Scotia contain a rich archive of fossil evidence for rapid climate change during the lateglacial (~14000-11600 yr). The Younger Dryas (12850-11650 yr BP) was the largest climate reversal in the lateglacial record (Mott et al. 1986; Lowe et al. 2008) and for a long time, the only one recognized in North America (Yu and Eicher 2001). Previous studies using pollen and plant macrofossils examined the impact of the Younger Dryas cooling event on the vegetation and surrounding landscape (Livingstone and Livingston 1958; Mott et al. 1986; Green 1987; Stea and Mott 1989; Stea and Mott 1998; Jetté and Mott 1989; Mayle and Cwynar 1995a). Changes in loss-on-ignition (LOI), an index of organic carbon, were found to reflect accurately changes in climate leading up to and during the Younger Dryas event (Dean et al. 1974; Levesque et al. 1993; Mayle et al. 1993; Mayle and Cwynar 1995a; Mayle and Cwynar 1995b; Heiri et al. 2001; Whitney et al. 2005). The utility of LOI for identifying rapid changes in climate was tested by Levesque (et al. 1993), when the apparent presence of the Gerzensee Oscillation based on LOI analysis was confirmed using fossil chironomids, aquatic insects whose distribution and abundance are strongly correlated with summer surface water temperatures (Walker et al. 1997; Lotter et al. 1999; Brooks and Birks 2001; Francis et al. 2006). This event came to be known as the Killarney Oscillation in North America. More recently, Quaternary climate events have been correlated to the Greenland Ice Cores (Björck et al. 1998; Lowe et al. 2008;
Rasmussen et al. 2014) where the Killarney/Gerzensee Oscillation is identified as the Greenland Interstadial 1b (GI-1b). Similarly the Younger Dryas is identified as the Greenland Stadial 1 (GS-1).

This early research using pollen, LOI and chironomids identified Nova Scotia as a location that is sensitive to lateglacial climate change. Its long coastline and low relief topography (outside the Cape Breton and Cobequid Highlands) result in the coastal influence being most significant to local climate variation.

The Wisconsinan ice sheet began to retreat from Nova Scotia approximately 14000 yr ago (Stea and Mott 1998; Shaw et al. 2006), allowing colonization by plants that survived the Ice Age south of the ice margins (Livingstone and Livingstone 1958; Green 1987; Mott and Stea 1993; Stea and Mott 1989; Mayle and Cwynar 1995a). The newly deglaciated terrain, in close proximity to retreating ice, was marked by strong north to south temperature gradients (Levesque et al. 1997) that resulted in the formation of ecotonal boundaries similar to the present day treeline of the Boreal Forest. Climate-driven vegetation shifts at ecotonal boundaries left a distinct signature in the pollen record as more temperate taxa replaced the shrub tundra vegetation that had colonized the earliest deglaciated landscapes. With the onset of Younger Dryas cooling, forests that had formed ecotonal boundaries south of the ice retreated from many sites, with vegetation shifting from forest back to shrub-tundra as the limits of trees shifted south to warmer temperatures (Mayle and Cwynar 1995a).
Spruce in Nova Scotia

White spruce and black spruce are distributed throughout modern-day Nova Scotia. Both are cold tolerant species that form the northern treeline of the boreal forest and likely represent the earliest spruce in Nova Scotia. Red spruce, although present today in Nova Scotia, is much less cold tolerant and is limited to lower altitudes. Green (1987, Fig. 1) suggested that spruce migrated across the exposed Georges Bank and Browns Bank, and colonized the area around Curry Pond in southwestern Nova Scotia as early as 12,000 years ago. This chronology was based on bulk sediment radiocarbon dates that appear too old based on the sediment lithology, pollen profiles and more recent accelerator-mass-spectrometer (AMS) dates. AMS-dated plant macrofossils obtained from Little Lake, to the northeast of Curry Pond, indicate that spruce was present by 11800 $^{14}$C yr BP (Mayle and Cwynar 1995a). By this time, lowland Nova Scotia had been ice-free for about 700 years. Plants had likely migrated into the province from New Brunswick across the head of the Bay of Fundy and possibly along the exposed shelf on the south shore of the province from Maine (Green 1987; Mayle and Cwynar 1995a). In a multi-site investigation, Mayle and Cwynar (1995a) found that spruce remained limited to the central third of the province until the onset of the Younger Dryas around 10800 $^{14}$C yr BP (~12850 cal yr BP; Rasmussen et al. 2014). They suggested that spruce had not migrated to the north because of latitudinal cooling, supported by pollen evidence for shrub tundra. The absence of spruce in southwestern Nova Scotia was attributed to a possible migration lag. The idea of a migration lag in southwestern Nova Scotia, which
would also be indicative of vegetation-climate disequilibrium, was the inspiration for this dissertation.

**Thesis organization**

Before I could determine if spruce was in equilibrium with climate, I needed to establish a criterion for inferring when spruce was present on the landscape at each of the 25 sites I examined. Previous authors had proposed a 20% spruce pollen limit (Anderson et al. 1991), but spruce macrofossils had been found in samples containing significantly smaller abundances of pollen (Mayle and Cwynar 1995a). Chapter 2 is dedicated to refining the pollen limit used to determine when spruce was present on the landscape using plant macrofossils and conifer stomates, collected from 25 lakes, as evidence for the presence of local spruce in lateglacial Nova Scotia. This chapter has been published in Review of Palaeobotany and Palynology (Vincent and Cwynar 2015).

Chapter 3 addresses the question of whether a condition of migration lag, and hence a vegetation-climate disequilibrium, existed between spruce and climate prior to the Younger Dryas. Using the refined pollen limit from Chapter 2 and chironomid inferred air temperatures, I compared the distribution of spruce to the thermal regime in Nova Scotia at 4 discrete time intervals, using LOI curves as chronostratigraphic markers. Using 2 cold events, the Killarney Oscillation and the Younger Dryas, as markers, I was able to select 4 contemporaneous time slices at each of my 25 lake sites and develop snapshots of the distribution of spruce in relation to temperatures. I plan to submit this chapter to Ecology.
The creation of 25 loss-on-ignition curves, which illustrate general temperature trends, provided me with an abundance of climate information. Although I expected to find the well documented Younger Dryas and Killarney Oscillations (Mott et al. 1986; Wright 1989; Peteet et al. 1990; Levesque et al. 1993; Mayle and Cwynar 1995a; Whitney et al. 2005) in the lake records, I was surprised to find evidence for 3 smaller cold events. Chapters 4 and 5 are each a detailed examination of these cold events using chironomid-inferred temperatures as a proxy to lake water temperatures.

Chapter 4 examines two cold events: the GI-1c2 and the Older Dryas (GI-1d). The GI-1c2 has only recently been discovered in Europe (Brauer et al. 2000) and the Older Dryas has a weak record in North America (Lowe and Walker 1997; Lowe et al. 1994; Lowe et al. 1995; Björck et al. 1998; Yu 2007; Yu and Eicher 2001). Sediment collected for this thesis from Thin Ice Pond and Veinot Lake is also being used in a multi-author study investigating microtephra for developing a lateglacial tephrachronology in North America. Preliminary results from that paper, of which I am a coauthor, indicate that the evidence for the Younger Dryas and GI-1c2 can be well constrained between radiocarbon dates and a tephra horizon. Once the microtephra paper is published, I will use the results to confirm the timing of the Older Dryas and GI-1c2, and submit chapter 4 as an article to Geology for publication.

Chapter 5 examines an unrecognized climate event during the rapid warming that initiated the Holocene (our current warm period beginning ~11650 cal BP) that may be analogous to the modern hiatus in global surface warming (Chen and Tung 2014). Loss-on-ignition and chironomid inferred temperatures are compared to more than 30
terrestrial and marine records from around the North Atlantic, East Pacific, and Arabian Seas where this brief event is apparent but has often gone unnoticed or unremarked upon. Given the widespread occurrence of this event and the possibility that it is analogous to the pause in our modern climate warming, I hope to publish Chapter 5 in a first-rank international journal.

**Contributions**

The idea for spruce-climate disequilibrium in Nova Scotia was first conceived by Francis Mayle and Les Cwynar (1995). We developed a basic plan for testing the hypothesis that spruce was not in equilibrium with climate by comparing pollen and chironomid records at discrete time intervals. As a geology major who did not know a spruce from a pine, I found Les to be a valuable mentor who taught me not only how to identify a pollen grain, but to appreciate its significance to the reconstruction of past vegetation and climate. I am the principal author on all manuscripts within this article-format thesis. Occasionally, I required assistance identifying individual specimens while I was learning to identify stomates and chironomids. I have acknowledged that assistance in the relevant chapters. Reference formats for chapters 1, 2, 4 and 5 adhere to the guidelines of the Review of Palaeobotany and Palynology. Chapter 3 is formatted for Ecology.
Notes on dating

The paleoecological literature uses various methods of reporting the age of samples. $^{14}$C years (carbon 14 years) represent the age of the sample calculated directly from the decay of $^{14}$C. The production of atmospheric $^{14}$C has not been constant over time (de Vries, 1958) and so dates reported in this fashion represents a first approximation of the age of the sample. Radiocarbon dates can be further refined by correlation to other independently dated material such as corals, tree rings and stalagmites, then calculating the probability distribution of the samples true age. This is known as the “calibrated age” and is reported as cal BP. BP stands for before present and is equivalent to 1950 AD, when atmospheric $^{14}$C levels were greatly altered by atomic bomb testing. More recently, the Greenland Ice Core Chronology 2005 (GICC05) has reported ages as yr b2k (Rasmussen et al. 2006). These dates are based on annual layer counting of the ice cores, reported before AD 2000; ice core years are not affected by variations in atmospheric $^{14}$C production.

Samples dated in this thesis have been calibrated using the CALIB program developed by Stuiver and Reimer (1993). When referring to dated material from studies conducted by different authors, I will generally report their findings as I found them in the original publication.
References


Chapter 2
Pollen whiskers, rational limits and cryptic refugia: determining the local presence of spruce on lateglacial landscapes of eastern Canada

Abstract
The 20% spruce isopoll has traditionally been used to identify the range limit of spruce in eastern North America. Using plant macrofossils and stomates in a network of 25 lateglacial-aged sites throughout Nova Scotia, I identified sites where spruce was locally present. My data suggest that 8% pollen abundance is a more reasonable limit that consistently indicates the local presence of trees, although spruce was present at a site where its pollen percentages were only 4%. Using 8% to define the presence of spruce, I find that the number of sites likely to contain spruce increases from 3 to 11 as compared to using the 20% isopoll. The resulting increase in the estimated distribution of spruce leads to a reduction in apparent migration lags and migration rates of spruce into the region. Reducing the error in distribution histories will provide better-constrained estimates of climate change and its impact on the local vegetation.

Introduction
Identifying the point at which trees are first present on the landscape in the paleovegetation record is important for several reasons. These trees may represent the
earliest arrivals, marking the onset of conditions suitable to support them and help define the climate of the time. Determining with greater precision when the first migrants arrived can reduce the apparent migration lag between trees and climate, a critical step for understanding the effect of climate on the movement of trees (Webb 1986; Latałowa and van der Knaap 2006). Of similar importance are the trees surviving in suitable microhabitats within an otherwise inhospitable landscape. In some cases, environmental conditions may not be conducive to sexual reproduction and trees will persist in a vegetative state. These trees, undetectable in the pollen record (Hicks 2006), can rapidly begin seed production when conditions improve. Persisting in cryptic refugia, surviving trees can explain rapid recolonization rates found after the retreat of glaciers; missing them results in unrealistic migration and colonization rates (Payette et al. 1985; Bennett 1986; Kullman 2002; Latałowa and van der Knaap 2006; Kullman 2008; Ammann et al. 2014). When conditions deteriorate, small numbers of trees can mark the limits of suitable shrinking habitat (Cwynar and Spear 1991) and constrain the timing of apparent climate change. Furthermore, trees themselves can significantly affect climate by changing the albedo of the landscape, so knowing when trees are present and physically modifying the landscape is important for climate modelers (Bonan 2008).

Unfortunately, identifying isolated groups and individual trees is difficult using traditional pollen analysis. Pollen has evolved to be dispersed in two main ways. Insect-pollinated trees produce sticky pollen that is poorly dispersed and therefore seldom deposited in lakes, the most commonly studied repository of fossil pollen. Wind-pollinated taxa have evolved to be easily dispersed over large distances, and may be
carried into areas where the plant does not occur. Their pollen may be deposited in advance of a migrating front of trees and appear in a pollen diagram as a whisker, that is indistinguishable from pollen deposited by small numbers of local trees (Davis et al. 1991; Moore et al. 1991). Larger numbers of local trees can also create an apparent whisker in the sediment record if they are poor pollen producers. Pollen production varies among taxa by several orders of magnitude (Birks and Birks 1980; Faegri et al. 1989; Moore et al. 1991; David 1997).

Trees migrate slowly and may take decades to reach sexual maturity. Consequently, a considerable lag may occur between the earliest arrival of trees on the landscape and the first detectable occurrence of their pollen in lake sediments, although the lag may be small relative to radiocarbon dating errors (Webb III 1986; Williams et al. 2002). Considering delayed reproduction, variation in pollen production and widespread dispersal of pollen into sites that do not contain trees, the question then becomes, “How much pollen must be present at a site in order to conclude that a taxon was actually growing there?”

**Pollen Limits**

Since the 1940s, palynologists have been trying to identify the point in the pollen record that indicates a tree is present on the landscape. Different techniques have yielded mixed results that were often later rejected. A thorough history of the presence/absence challenge has been presented by Ammann (et al. 2014) and includes the use of the “rational limit”, a term discussed as early as 1949 by Firbas and later defined by Smith and Pilcher (1972) as “the point at which the pollen curve begins to rise to sustained
values” indicating the arrival of a taxon. Although the rational limit is often a good reference for the arrival of trees, early colonizing trees living near their physiological limits can occur in very small numbers and be indistinguishable from the whisker preceding the arrival of trees. By the time the “rational limit” has been reached, some trees have arguably been present for a long time (Bennett 1985; Kullman 2008). For some moderate pollen producers, the beginning of the continuous pollen curve (the empirical limit) has been suggested (Ammann et al. 2014) as indicative of the arrival of a species at a site.

In other studies, a numerical limit has been set to indicate the presence of trees. MacDonald and Cwynar (1985) used abundances increasing through 15%, rather than the attainment of “sustained values”, to mark the arrival of lodgepole pine (Pinus contorta ssp. latifolia) in Yukon Territory. This value has since been refined to 5%, resulting in arrival estimates ~500-2500 years earlier than previously suggested (Strong and Hills 2013). In the Northwest Territories, Ritchie (1974) found that spruce contributed >15% to the pollen spectra of lake sediments in the forest tundra and 10-14% in the tundra. Spruce (Picea spp.) tree line in eastern North America has been shown to correlate with the 20% isopoll (Anderson et al. 1991). This work was based on a 100 km grid using pollen from lakes that reflect regional vegetation (Jacobson and Bradshaw 1981). Although the 20% limit conforms well to the general position of treeline, itself a broad area (up to 100km wide) of fluctuating tree abundance, it cannot resolve the limits of individual trees or isolated stands on the landscape. To determine if trees were present on the local (100’s of
meters or less) landscape scale, we must look to evidence that is locally sourced, such as plant macrofossils and stomates.

**Macrofossils**

Size and fragility prevent macrofossils from being transported far from their source (Birks 1980; Birks and Birks 1980), except in rare instances of transport by forest fire convection (Pisaric 2002) or in mountainous regions with open vegetation that permits transport by wind (Glaser 1981). Plant macrofossils have been used to improve vegetation reconstructions on the local scale, often at treeline, minimizing apparent migration lags found in the pollen data alone (MacDonald and Ritchie 1986; Peteet 1986; Davis et al. 1991; Peteet 1991; Prentice et al. 1991; Clayden et al. 1996; Kullman 1996, 1998; Pisaric et al. 2000; Gervais and MacDonald 2001; Ali et al. 2003; Pisaric et al. 2003; Tinner and Theurillat 2003; Leitner and Gajewski 2004). For example, Kullman (1995, 2001) dated fossil spruce wood at 2 sites in the Scandes Mountains and found it to be 2000 and 5000 years older than previous estimates for spruce arrival based on pollen. Lodgepole pine macrofossils have been found with 2% pine pollen in southeastern Alaska (Peteet 1991), suggesting that pine may be present at pollen values significantly below the 15% proposed by MacDonald and Cwynar (1985). *Larix* and *Pinus* plant imprints were found at a high altitude travertine site in the French Alps 100 m higher than previous estimates for treeline history and 2600 years prior to estimates based on pollen evidence (Ali et al. 2003).
Stomates

Despite the reliability of macrofossils as indicators of the local presence of trees, macrofossils are often absent or rare in lake sediments (McLachlan and Clark 2004, Ammann et al. 2014). The absence of a macrofossil is negative evidence that is subject to several interpretations (Ammann et al. 2014); the taxon may indeed have been locally absent; the taxon was present but no plant macrofossil reached the coring site, or the taxon was present and its macrofossils reached the core location but were destroyed by insects or decomposition. The addition of stomate analysis can often fill in the gaps where macrofossils are absent (Gervais et al. 2002). Stomates are generally found in large numbers on the surface of leaves. The lignified guard cells of conifer stomates preserve well, even after the needle from which they came has been degraded, making them potentially more abundant than fossil leaves - the same is not true for the unlignified stomates of deciduous, broadleaved trees. Limited by the same dispersal factors as leaves (a plant macrofossil), stomates provide strong evidence that the source plant was locally present (Hansen 1995; Clayden et al. 1996; Clayden et al. 1997), although stomates have been found beyond treeline in situations where reworking of older sediments is possible (Leitner and Gajewski 2004; Clayden et al. 1996). Despite the relative abundance of stomates compared to macrofossils, the interpretation of negative stomate evidence can also be ambiguous and should be approached with caution (Gervais and MacDonald 2001; Pisaric et al. 2001; Leitner and Gajewski 2004, Ammann et al. 2014).
Spruce in Nova Scotia

Black spruce and white spruce (*Picea mariana* (Britton, Sterns and Poggenburg), *Picea glauca* (Moench) respectively) are found throughout present day Nova Scotia whereas red spruce (*Picea rubens* (Sargent) is limited to lowland regions. Spruce were among the first trees on the landscape after deglaciation (Mott et al. 1986; Jetté and Mott 1989; Mayle and Cwynar 1995a).

The 20% spruce isopoll applied to lateglacial Nova Scotia sites suggests that spruce trees were present in central Nova Scotia (Stea and Mott 1989; Mayle and Cwynar 1995a) during the last few centuries before the onset of the Younger Dryas (12700 cal BP) with limited northern expansion immediately before Younger Dryas cooling began (Jetté and Mott 1989). This distribution is very restricted, given that ice had been absent from most of the province since 14000 cal. BP (Shaw et al. 2006), and may be a minimal distribution considering that spruce macrofossils have been found in the area associated with pollen values <15% (Mayle and Cwynar 1995a).

As part of a larger study examining the distribution of spruce during the lateglacial period, I sought to refine the limits of spruce distribution in Nova Scotia prior to the onset of the Younger Dryas. Here I present data that support the use of plant macrofossils and stomates as a means to refine significantly the pollen limit for the presence of spruce on the local landscape in lateglacial Nova Scotia. These results also confirm that spruce can be present while contributing surprisingly small relative amounts of pollen to the local pollen rain.
Methods

Field Methods

Twenty-five lakes, evenly spaced throughout Nova Scotia, were selected for this study (Fig. 2.1, Table 2.1.). Small lakes (~200m diameter) without inflowing streams were selected to ensure the pollen input is dominated by local sources within several hundred meters of the site (Jacobson and Bradshaw 1981). The Cape Breton Highlands were not sampled because ice covered this region until the end of the Younger Dryas (Stea and Mott 1998).

All sites were cored using a modified Livingstone piston sampler (Wright 1967). Cores were taken from the deepest part of each lake for maximum sediment recovery with the exception of Skating Bench Pond and Brier Island Bog; they were cored from the bog surface after sounding for bedrock or clay with coring rods to determine the deepest part of the basin. Cores exhibiting typical, visible, lithological changes associated with the Younger Dryas were selected for the study (Levesque et al. 1993; Mayle et al. 1993; Mayle and Cwynar 1995b). Selected cores were wrapped in plastic film and aluminum foil, then transported to the lab where they were stored at 4°C.

Laboratory methods

The lateglacial and earliest Holocene sections of each core were sliced at 0.5 cm increments, except in basal clay and sand where 1.0 cm increments were used. From each slice a 0.5 ml sample was removed for loss-on-ignition analysis (Dean 1974). Lateglacial
lake sediments in Nova Scotia have a characteristic loss-on-ignition profile (Mott et al. 1986; Levesque et al., 1993, Stea and Mott 1998). The Younger Dryas event (or Greenland Stadial-1 (GS-1) according to Björck et al. 1998) has been shown to occur simultaneously throughout the Province (Levesque et al. 1993; Mayle et al. 1993). Consequently, I used the LOI curve to select contemporaneous sediment from each site.

Two slices were chosen from the peak in LOI values representing the warm interval Greenland Interstadial-1a (GI-1a) between the Greenland Interstadial-1b (GI-1b) and the GS-1, except at Brier Island Bog where only one slice was used; this sample was added late in the analysis. To confirm that my choice of samples represents the GI-1a, radiocarbon dates were obtained from 5 sites (one site shown in Fig. 2.2). Each slice from each GI-1a peak was subsampled for pollen and stomate analyses. After the subsamples were removed, all remaining material from the 2 slices centered on the peak plus material from 1 cm above and 1 cm below the peak (totaling 6 slices or 3cm) was sieved for plant macrofossil analysis using screens with 0.25, 0.5 and 1.0 mm meshes. Plant macrofossils were identified using modern reference material at the University of New Brunswick.

Material for Chase Pond was sieved and identified by Mayle and Cwynar (1995a).

Pollen was processed according to Faegri et al. (1989) and Cwynar et al. (1979). Samples were not coarse-sieved so as to preserve bits of epidermis bearing stomates. Slides were examined at 400X magnification using keys from McAndrews et al. (1973) and Moore et al. (1991) as well as modern reference material at the University of New Brunswick. Pollen analysis was completed at all 25 sites. Percent abundance of each
taxon was calculated using a minimum sum of 300 pollen grains and spores of terrestrial plants. Pollen grains of *Picea* were not resolved to species for this study.

Pollen slides were scanned for fossil stomates. The original purpose for these samples did not require the addition of an exotic marker. If stomates were not found during the pollen count, then triple the amount of material examined for pollen was scanned for stomates (~900 terrestrial pollen grains). Stomate frequencies are reported as number of stomates/300 pollen grains in the sum. Fossil stomates of *Picea, Pinus spp.* and *Thuja spp.* were identified using modern reference material at the University of New Brunswick and the key of Hansen (1995). *Thuja* is easily separated from *Pinus* and *Picea* by its small size (mean length 25-27 µm), narrow medial border and round outline. *Picea* and *Pinus* are much larger (32-40 µm and 44-54 µm, respectively) and more oval. Size comparisons are helpful for separating the very small *Thuja* from larger stomates such as *Picea* and *Pinus*, however, size alone cannot generally be used to discriminate among taxa (Hansen 1995). An acute angle of attachment is characteristic between the upper woody lamellae and the stem in *Pinus* forms an acute angle. In *Picea* this angle is much smaller and the upper woody lamellae appear completely attached to the stem. The lower woody lamellae are often obscured in fossil *Picea* - note poor resolution in Fig. 2.3. In *Pinus*, the lower woody lamellae are quite broad relative to the upper woody lamellae.

In this study I consider the presence of spruce macrofossils or stomates in any concentration to be evidence for the local presence of at least one individual spruce tree.
Results and Discussion

Loss-on-Ignition and Radiocarbon dates

Thin Ice Pond (Fig. 2.2) exhibits a typical loss-on-ignition curve for lateglacial Nova Scotia (Mayle et al. 1993). The middle of the GI-1a at Thin Ice Pond (TO-10420) has been accelerator-mass-spectrometer (AMS) dated to 10,800 ±100 yr B.P. and calibrated using a 2σ error estimate with CALIB version 7.0 and IntCal13 to 12549-12929 cal BP (Stuiver and Reimer 1993, Reimer et al. 2013). This date is consistent with the onset of the GS-1 in the Greenland Ice Core Chronology 2005 (GICC05) dated to 12896 yr b2k (Lowe et al. 2008). AMS results for samples taken from the GI-1a at four other lakes are also listed in Table 2.2 and confirm the use of the loss-on-ignition curve as a chronostratigraphic marker.

Pollen

The relative abundance of *Picea* pollen in each of the two GI-1a samples from each site is listed in Table 2.1, sorted by latitude. Spruce pollen values range from 2.3% (Spider Pond) to 30.2% (Croskill Lake; rounded to the nearest 0.1%) with the lowest pollen values found at the northernmost sites and the highest pollen values concentrated in the geographic centre of the province (Fig. 2.1). Low spruce pollen values in the north and south are not surprising as Mayle and Cwynar (1995a) show that these regions were covered by shrub tundra preceding the GS-1 with high values of *Betula, Salix*, Ericales, and Cyperaceae.
Stomates

Spruce stomate counts range from 0 to 10.5 stomates per 300 pollen grains (Table 2.1). Stomates were found in 25 of the 48 slices and indicate the presence of spruce at 14 of 25 sites with pollen values ranging from 4.2%-30.2%. Stomates were found most frequently at sites in the central portion of the province (Fig. 2.1).

Spruce Macrofossils

Macrofossil analysis indicated the presence of spruce at 4 southwestern sites with 5 spruce needles and 1 spruce sterigma co-occurring with spruce pollen values ranging from 7-20% (Table 2.1). At one site, Lac à Magie, a macrofossil was found in the absence of stomates. At China Lake, one sample contained stomates while the adjacent sample contained a macrofossil. The remaining three samples with macrofossils also contain stomates.

Presence/absence

Here I consider spruce trees to be locally present if spruce macrofossils or stomates were found in the sample regardless of spruce pollen values. Not surprisingly, the frequency with which I found spruce stomates and macrofossils increased with the abundance of spruce pollen, but only very generally. The highest concentration of spruce
stomates was found in a sample from Li’l Jess Pond containing only 17% spruce pollen, whereas some samples with higher spruce pollen values (26% and 30%) contained only 7.1 and 3.9 stomates per 300 pollen grains respectively. The stomates in the Li’l Jess samples were found within a larger piece of needle. Given that stomates can occur in large numbers on the surface of a needle that may or may not be disaggregated before deposition, the interpretive value of stomates lies primarily in their presence or absence, not their frequency.

Similarly, the frequency of plant macrofossils was greatest in samples with the highest pollen values, but their presence was sporadic, occurring in only 5 of the 25 samples containing stomates.

A comparison between spruce pollen values and the presence or absence of *Picea* macrofossils and stomates is summarized graphically in figure 2.4. Clear evidence for the presence of spruce trees (spruce macrofossils and/or stomates) occurs consistently in all samples containing more than 8.3% spruce pollen. By lumping together the evidence for both samples at each site I find that spruce macrofossils and/or stomates consistently occur at all sites where one or both samples contain at least 8% spruce pollen, with one exception. Lac à Magie samples show contradictory evidence for the presence of spruce with pollen values near 8% occurring in the absence of macrofossils and stomates in one sample, and 6.7% spruce pollen values occurring with a spruce sterigmata in the adjacent sample. Given the ambiguity of the absence of macrofossils or stomates and considering that the adjacent sample at Lac à Magie does contain spruce macrofossils (indicated by “m” in figure 2.4) associated with 6.7% pollen, it seems reasonable to conclude that the
8.1% spruce pollen found at Lac à Magie is indicative of the presence of spruce. In samples with less than 8% spruce pollen I find locally sourced evidence for the presence of spruce in only 19% of samples. Interestingly, spruce stomates were found in samples that contained as little as 4.2% spruce pollen.

The beginning of the continuous curve (empirical limit of Ammann et al. 2014) for spruce pollen has been suggested as indicative of the presence of spruce (Picea abies) in Europe (Giesecke and Bennett 2004; Ammann et al. 2014). The new 8% limit falls well within the continuous limit and slightly precedes the rational limit of previously published Nova Scotia spruce pollen curves (Livingstone and Livingstone 1958; Mott et al. 1986; Jetté and Mott 1989; Mayle and Cwynar 1995a) where the continuous curve was considered a whisker preceding the arrival of distant trees. These results suggest that the rational limit, at least for spruce, may be inappropriate for identifying the presence of spruce on the landscape and that spruce is likely present while contributing an apparent whisker to the pollen record.

The 15% spruce pollen limit found by Ritchie (1974) was based on samples from large lakes, whose records are dominated by the regional pollen rain. The lower limit I found may be explained by the smaller size of the lakes I sampled. Trees within 1.0-4.5 km are better represented by pollen collected from small basins and the contribution from distant sources tends to be less than in larger basins (Bradshaw and Webb 1985). The relevant source area of vegetation represented by my data is also much smaller than the 100 km sampling grid used by Anderson (et al. 1991) to define the 20% spruce pollen limit (Sugita 1994).
**Distribution of spruce**

Pollen, macrofossil and stomate data have been summarized on a map (Fig. 2.1) to highlight spatial trends. The long axis of Nova Scotia runs approximately southwest to northeast. Sites containing 8% or more spruce pollen as well as stomates and/or macrofossils are concentrated in the south-central area of the province, forming a relatively contiguous zone. At sites to the southwest and northeast ends of the Province, spruce pollen values decrease and both stomates and macrofossils are less frequent. Although the 20% isopoll also indicates that the central portion of the Province contained spruce trees earliest, the 8% isopoll greatly increases the area inhabited by spruce. Using the 8% limit to define the presence of spruce would have a significant impact on apparent migration routes and rates in future discussions of forest history, reducing migration lags by up to 100 km.

**Conclusions**

I suggest that the pollen limit for the presence of spruce on the landscape in Nova Scotia be refined to 8%. The consistent presence of spruce stomates and/or macrofossils at or above this limit indicates that 8% is not a whisker deposited in advance of distant approaching trees, but is the contribution from trees within the catchment. Using the 8% pollen limit should significantly reduce the underestimation of the distribution of spruce. Given the evidence for the local presence of spruce associated with pollen levels as low
as 4%, we should interpret future pollen records more cautiously, especially when using the resulting interpretation of plant distribution for studies of plant migration rates or climate change. Furthermore, this study supports previous suggestions that pollen analysis can easily be supported with locally sourced stomate evidence resulting in significantly fewer ambiguous results than using macrofossils alone (Birks 1980; Hansen 1995; Clayden et al. 1996; Ammann et al. 2014). Stomate analysis requires no further processing than samples for pollen, relatively little extra counting and, in this study, yielded nearly triple the number of samples indicating the local presence of spruce than plant macrofossils alone, resulting in a significant reduction in ambiguity. Future pollen analyses seeking to identify the earliest possible arrivals of conifers on the landscape ought to use stomate analysis to refine the limits of tree distribution. With the addition of stomate analysis, we may be able to confirm that the continuous spruce pollen curve is indicative of the local presence of spruce in lateglacial Nova Scotia.

Acknowledgments

This research was supported by a Natural Sciences and Engineering Research Council Discovery Grant to Les Cwynar. I gratefully acknowledge the helpful comments of Alwynne Beaudoin and 3 anonymous reviewers. I also thank Ray Spear, Mark Landry, Josh Kurek, Tara Warren, Cy Pedersen and Ian Vincent for their assistance in the field as well as Susan Clayden and Barbara Hansen for assistance with stomate identification.
Table 2.1. Site summary: Locations of sites and evidence for the presence of spruce based on Picea stomates or macrofossils. P indicates spruce is present at that site. Unofficial site names marked with *.

<table>
<thead>
<tr>
<th>#</th>
<th>Site name</th>
<th>Coordinates</th>
<th>Sample Depth (cm)</th>
<th>% Picea pollen</th>
<th># stomates/300 pollen grains</th>
<th>Picea macrofossils</th>
<th>Picea</th>
<th>P=present</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Campbell’s Lake</td>
<td>43.76°N, 65.27°W</td>
<td>589-589.5</td>
<td>19.4</td>
<td>4.8</td>
<td>2 needles</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Thin Ice Pond*</td>
<td>43.91°N, 65.87°W</td>
<td>569.5-570.5</td>
<td>6.6</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Finale Pond*</td>
<td>43.98°N, 65.63°W</td>
<td>410-410.5</td>
<td>9.1</td>
<td>1</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Brier Island Bog</td>
<td>44.21°N, 66.37°W</td>
<td>584.3-585</td>
<td>8.5</td>
<td>0.66</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>China Lake</td>
<td>44.25°N, 64.56°W</td>
<td>416.3-417</td>
<td>12.5</td>
<td>-</td>
<td>1 needle base</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Lac a Magie</td>
<td>44.27°N, 66.08°W</td>
<td>821-821.5</td>
<td>6.7</td>
<td>-</td>
<td>1 sterigmata</td>
<td></td>
<td></td>
</tr>
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<td>7</td>
<td>Skating Bench Pond*</td>
<td>44.40°N, 65.07°W</td>
<td>754-754.5</td>
<td>19.9</td>
<td>2.0</td>
<td>1 needle</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Little Lake</td>
<td>63.94°W</td>
<td>652.5-652.5</td>
<td>10.5</td>
<td>2.3</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>Mud Lake</td>
<td>44.72°N, 65.11°W</td>
<td>478.4-478.5</td>
<td>5</td>
<td>0.5</td>
<td>-</td>
<td></td>
<td></td>
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<tr>
<td>10</td>
<td>Veinot Lake</td>
<td>44.74°N, 64.54°W</td>
<td>817.5-818</td>
<td>25.9</td>
<td>7.1</td>
<td>-</td>
<td></td>
<td></td>
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<td>11</td>
<td>Li’l Jess Pond</td>
<td>63.32°W</td>
<td>441-441.5</td>
<td>16.8</td>
<td>10.5</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Midconnor Lake</td>
<td>44.84°N, 64.73°W</td>
<td>630.5-631</td>
<td>15.5</td>
<td>0.33</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Spider Pond*</td>
<td>44.87°N, 62.57°W</td>
<td>385-385.5</td>
<td>2.6</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
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<tr>
<td>14</td>
<td>Marsh Lake</td>
<td>44.97°N, 62.10°W</td>
<td>792.5-792.5</td>
<td>6.3</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Croskill Lake</td>
<td>45.13°N, 63.18°W</td>
<td>493.5-494</td>
<td>28</td>
<td>3.8</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>Long Lake</td>
<td>45.23°N, 62.38°W</td>
<td>494.5-494.5</td>
<td>30.2</td>
<td>8.7</td>
<td>-</td>
<td></td>
<td></td>
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<tr>
<td>17</td>
<td>McInnis Lake</td>
<td>45.34°N, 62.09°W</td>
<td>562.5-563</td>
<td>8.3</td>
<td>0.24</td>
<td>-</td>
<td></td>
<td></td>
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<tr>
<td>18</td>
<td>Western Pond</td>
<td>45.34°N, 61.50°W</td>
<td>356.5-357</td>
<td>4.9</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>19</td>
<td>Little Lake (C.C.)</td>
<td>45.41°N, 63.03°W</td>
<td>544-544.5</td>
<td>3.8</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>20</td>
<td>Leak Lake</td>
<td>45.43°N, 64.34°W</td>
<td>261-261.5</td>
<td>7.2</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>21</td>
<td>Stillman Pond</td>
<td>45.51°N, 62.52°W</td>
<td>626.5-627</td>
<td>12.5</td>
<td>0.33</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td>Borden’s Lake</td>
<td>45.59°N, 61.62°W</td>
<td>1055.5-1056.5</td>
<td>6.3</td>
<td>0.33</td>
<td>-</td>
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<td>23</td>
<td>Hector Lake</td>
<td>45.65°N, 61.36°W</td>
<td>331.5-332</td>
<td>5.1</td>
<td>-</td>
<td>-</td>
<td></td>
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<tr>
<td>24</td>
<td>Chase Pond</td>
<td>45.65°N, 60.66°W</td>
<td>536.5-537.5</td>
<td>3.9</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>25</td>
<td>Main à Dieu Pond</td>
<td>45.98°N, 69.84°W</td>
<td>498-498.5</td>
<td>3.2</td>
<td>-</td>
<td>-</td>
<td></td>
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</tr>
</tbody>
</table>
Table 2.2. Radiocarbon dates of GI-1a at 5 representative sites. All dates have been calibrated using CALIB version 7.0, (Stuiver and Reimer 1993) and IntCal13 (Reimer et al. 2013).

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{14}$C age yr BP</th>
<th>Cal. Age (2σ estimate range)</th>
<th>Lab</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thin Ice Pond</td>
<td>10800±100</td>
<td>12739(12549-12929)</td>
<td>TO-10420</td>
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<tr>
<td>Lac a Magie*</td>
<td>10480±80</td>
<td>12363(12095-12631)</td>
<td>TO-1433*</td>
</tr>
<tr>
<td>Brier Island Bog</td>
<td>10815±35</td>
<td>12717(12683-12752)</td>
<td>UCI-76878</td>
</tr>
<tr>
<td>Main à Dieu</td>
<td>10680±1190</td>
<td>12491(9300-15683)</td>
<td>TO-10413</td>
</tr>
<tr>
<td>Western Pond</td>
<td>11360±850</td>
<td>13317(10870-15765)</td>
<td>TO-10423</td>
</tr>
</tbody>
</table>

*Calibrated after Mayle et al.1993.
Figure 2.1. Site map and distribution of spruce during GI-1a in Nova Scotia, i.e. shortly before the start of the YD. Numbers correspond to sites in Table 2.1. Trees indicate the presence of spruce based on macrofossils and/or stomates. Grey trees <8% spruce pollen, black tree >8%. Hatched area indicates estimates of spruce distribution based on 20% isopoll (Anderson et al. 1991).
Figure 2.2. Loss-on-ignition profile for Thin Ice Pond. Dated GI-1a indicated by block and AMS dated to 10,800 +/- 14C yr. BP. Arrow marks peak used for sample selection.
Figure 2.3. Photomicrographs of Picea spp. stomata with distinguishing features (x400, scale in µm). Note broad oval shape of stomata. Lower woody lamellae are narrow and difficult to distinguish from upper woody lamellae. Upper woody lamellae appear completely attached to the stem. Stem is approximately 5 µm wide.
Figure 2.4. Bar graph representing the relative abundance of *Picea* pollen in each of two samples from each site sorted by abundance. Bar length indicates % *Picea* pollen. Black bars indicate the occurrence of *Picea* macrofossils, stomates or both (s=stomate, m=macrofossil).
References


Bennett, K.D., 1985. The spread of *Fagus grandifolia* across eastern North America during the last 18 000 years. Journal of Biogeography, 12, 147-164.

Bennett, K.D., 1986. The rate of spread and population increase of forest trees during the postglacial. Philosophical Transactions of the Royal Society of London, B. 314, 523-531.


Hicks, S., 2006. When no pollen does not mean no trees. Vegetation History and Archaeobotany, 15, 253-261.


Strong, W.L. and Hills, L.V., 2013. Holocene migration of lodgepole pine (Pinus contorta var. latifolia) in southern Yukon, Canada. The Holocene, 23(9), 1340-1349.


Chapter 3

Widespread spruce-climate disequilibrium in lateglacial Nova Scotia

Abstract

Pollen-based paleoclimate reconstructions assume that vegetation is in equilibrium with climate. If this is not the case, then the inferred timing of climate events is invalid. Previous studies have suggested that spruce occurred in the central third of Nova Scotia during the lateglacial, but was absent in the southwest due to disequilibrium with climate and in the northeast due to latitudinal cooling. I tested the hypothesis that spruce was not in equilibrium with climate using 25 lateglacial-aged sites throughout Nova Scotia during 4 time slices (start of GI-1b: ~13250 ca BP, start of GI-1a: 13000 ca BP, start of GS-1a: ~12700 ca BP and start of Holocene: ~11600 ca BP). Spruce distribution was determined from pollen and stomate data, and compared to maps of chironomid inferred mean July temperatures to determine if a disequilibrium existed between spruce and climate in southwestern Nova Scotia during the lateglacial. The alternative hypothesis H2: “Southwestern Nova Scotia was not warm enough to support spruce” was rejected because spruce occurred at temperatures as low as 13 °C in central Nova Scotia while southwestern Nova Scotia temperatures were higher than 13 °C. Moreover, northeastern Nova Scotia summer temperatures were also suitable for spruce occurrence but spruce was absent, indicating that a significant disequilibrium also existed.
in the northeast. Similarly, disequilibrium between spruce and climate occurred throughout much of the 500 years prior to the onset of the GS-1.

Introduction

Paleoecologists using fossil pollen to reconstruct climate must assume that the amount of time required for trees to respond is small relative to the time scale of climate change and that the distribution and abundance of plants are in equilibrium with climate (Davis 1976, 1986; Davis and Botkin 1985; Webb III 1986). The existence of a lag would create pollen-based paleoclimate reconstructions that underestimate the time of climate change. Moreover, temperature transfer functions assume that the vegetation used in the model is in equilibrium with climate. While most researchers would agree that trees move relatively slowly across the landscape, compared to other proxies such as chironomids which can fly, and that there must be a lag between when a new landscape becomes suitable to support a tree species and the time it takes for that species to arrive, this lag has not previously been documented over long timescales on a fine spatial scale. Studies tracking the movement of trees across North America and Europe show that forest generally keeps pace with, and is indicative of, climate (Ammann et al. 2000; Williams et al. 2002) within decades to centuries. Species specific responses at individual sites have not previously been examined and differing responses among species presents a further complication (Ritchie 1986).

Boreal Forest expansion reduces the albedo of the landscape leading to a positive feedback for climate warming, consequently, it is important to understand how the future Boreal Forest will respond to projected climate (Betts 2000; Bonan 2008). Part of that
understanding is determining the magnitude of migration lag for various taxa within the forest. One taxon of interest is spruce. Together, black and white spruce (*Picea mariana* and *Picea glauca*, respectively) are the dominant taxa in the Boreal Forest, distributed from Alaska to Newfoundland (Payette 1993). Spruce is sensitive to changes in climate, particularly at ecotonal boundaries such as northern treeline and in lateglacial Nova Scotia (Caccianiga and Payette 2006, Mayle and Cwynar 1995a, Payette 1993, Cwynar and Spear 1991). White spruce is a pioneer species that can rapidly expand into available habitat, while black spruce can persist beyond treeline as krummholz where it can hasten recolonization if conditions improve (Payette 1993; Payette et al. 1985). Future climate warming may result in a rapidly expanding, spruce dominated, northern treeline. Alternatively, some taxa may not keep up with the rapidly changing climate and are at risk of extinction or local extirpation (Davis 1989).

**Setting**

The topography and location of Nova Scotia make it an ideal region for studying lateglacial climate and vegetation (~14650-11650 yr; Lowe et al. 2008). Situated on the east coast of Canada, the region is sensitive to temperature and oceanic circulation changes in the North Atlantic. Wisconsinan ice began its retreat from Nova Scotia around 15600 years ago (Stea and Mott 1998; Shaw et al. 2006), exposing new landscapes for the colonization of plants migrating from the south and west. Steep thermal gradients behind retreating ice led to well-defined ecotones that are clearly recorded in various proxies within lake sediments (Mott 1991; Wilson et al. 1993; Levesque et al. 1997;
Whitney et al. 2005). Consequently, climate events such as the Younger Dryas and Killarney Oscillation are clearly recorded in lateglacial lake sediments (Green 1976; Mott 1991; Levesque et al. 1993; Mayle and Cwynar 1995b; Spooner et al. 2005; Whitney et al. 2005; Lennox et al. 2010).

**Observation**

The lateglacial distribution of spruce in Nova Scotia has attracted much attention (Green 1976,1987; Mott 1991, Mayle and Cwynar 1995a; Stea and Mott 1998). The distribution of spruce appears to have been limited to the central third of Nova Scotia prior to the onset of the Younger Dryas (or GS-1 according to Björck et al. 1998 and Lowe et al. 2008) cooling event, ~12,700-11,600 cal yr BP (all dates cited in calibrated years before present (cal yr BP) according to Stuiver and Reimer 1993 (v.6); Stuiver et al. 1998; Mayle and Cwynar 1995b). Mayle and Cwynar (1995) suggest that spruce migrated via the Chignecto Isthmus, into central mainland Nova Scotia before spreading to the northeast and southwest, but was halted due to the rapid onset of GS-1 cooling. Assuming southwestern Nova Scotia was as warm as or warmer than central Nova Scotia, due to latitude and distance from ice (Levesque et al. 1997; Stea and Mott 1989; Shaw et al. 2006), the spreading forest appears not to have had enough time to reach all available habitats in the southwest. Shrub-tundra vegetation in northeastern Nova Scotia has been interpreted as indicating that conditions may have been too cold for spruce at that time (Mayle and Cwynar 1995a).
Hypotheses

Based on the observed distribution of spruce in lateglacial Nova Scotia, I can suggest several hypotheses for the absence of spruce in southwestern Nova Scotia.

H1: spruce may not have reached southwestern Nova Scotia due to physical barriers to migration.

H2: southwestern Nova Scotia was too cold to support spruce.

H3: southwestern Nova Scotia was warm enough to support spruce but spruce migration lagged behind climate.

Spruce migration was likely not hindered by physical barriers. A relatively gentle topography left few barriers to plant migration; the exceptions being the Cape Breton Highlands, which remained ice covered until at least 13 000 years ago, and parts of the Cobequid Highlands (Stea and Mott 1998). Exposed shorelines and lower sea levels provided ample migration paths past remnant ice (Stea and Mott 1998; Shaw et al. 2006), which was likely more limited than previously suggested, given the widespread distribution of complete lateglacial records in my study. With no barriers to migration, I can reject H1 as an explanation for the distribution of spruce in lateglacial Nova Scotia.

If we assume that the shrub tundra vegetation observed in southwestern Nova Scotia accurately reflects the climate during the lateglacial, then hypothesis H2: southwestern Nova Scotia was too cold to support spruce will hold true. This explanation is unlikely as southwestern Nova Scotia lies at equal and lower latitudes than sites to the north and east. Furthermore, southwestern Nova Scotia is further from retreating ice
where steep temperature gradients near the ice are most likely to prevent the establishment of spruce (Levesque et al. 1997; Stea and Mott 1989; Shaw et al. 2006). Conversely, H3: southwestern Nova Scotia was warm enough to support spruce, implies a migration lag and suggests that spruce was not in equilibrium with climate in southwestern Nova Scotia.

**Predictions**

I predict that inferred lateglacial air temperatures in the southwest would have been as warm as, or warmer than, the inferred temperatures from central and northeastern Nova Scotia and spruce was not in equilibrium with climate. To test this prediction, I determined if a migration lag existed between spruce and climate during four different time slices: start of GI-1b: ~13250 ca BP, start of GI-1a: 13000 ca BP, start of GS-1a: ~12700 ca BP and start of Holocene: ~11600 ca BP, by comparing the distribution of spruce from a high-resolution network of sites to the temperatures at each of those sites. Sites that are warm enough to support spruce but where spruce is absent will be considered indicative of disequilibrium between spruce and climate. Spruce is considered present at a site when the relative abundance of spruce pollen reaches 8% and/or stomates are present (Vincent and Cwynar 2015). Stomates are a reliable indicator of the local presence of conifers on the landscape (Hansen 1995; Clayden et al. 1996, 1997; Hansen et al. 1996; David 1997; Gervais & MacDonald 2001). Vincent and Cwynar (2015) also found stomates associated with spruce pollen values less than 8% in Nova Scotia, and acknowledge that spruce may also be present at sites with lower spruce pollen values.
A chironomid temperature transfer function was used to infer mean July air temperatures (Francis et al. 2006). June-August temperatures are the most influential climate parameters on the growth of spruce at treeline (Payette 2007); however, mean July temperatures have been used as a parameter to define the distribution of spruce (Thompson et al. 1999). Chironomid life cycles are rapid and their response to climate is independent of the distribution of spruce, avoiding the circularity inherent in climate studies using fossil pollen to infer both past vegetation and past climate.

**Methods**

I collected sediment cores from the deepest part of 25 lakes (Fig. 3.1) using a modified Livingstone piston corer (Wright 1967). Lakes were selected for small size (<10 ha), circular shape and the absence of inflowing and out flowing streams. Pollen collected from sediments in lakes this size represents trees that are local – about 20m to a few 100s of meters from the lake (Jacobson and Bradshaw 1981).

Cores were sliced at 0.5cm intervals and subsampled for loss-on-ignition (LOI) analysis (Dean 1974; Heiri et al. 2001). Lateglacial Nova Scotia sediment lithology is distinct and the LOI curve can be used as a chronostratigraphic marker (Mayle et al. 1993; Levesque et al. 1997, Vincent and Cwynar 2015). From the LOI curve I selected time slices of interest for pollen, stomat and chironomid analyses from each of the 25 cores (Fig. 3.2). These time-intervals include the last point before decreasing LOI values marking the start of GI-1b, the last point before rising LOI values marking the onset of the GI-1a, the last point before decreasing LOI values at the onset of GS-1, and the last
point before the rise in LOI marking the beginning of the Holocene. These distinctive points in the LOI curve ensure that samples are contemporaneous between lakes, while minimizing costly radiocarbon dates that require well-placed macrofossils for accuracy.

Pollen was processed according to Faegri et al. (1989) and Cwynar et al. (1979). Chironomids were prepared according to Walker (2001) with minimum head capsule count of 50 (Heiri and Lotter 2001; Larocque 2001; Quinlan and Smol 2001). Chironomid identifications were made using Weiderholm (1983), Walker (1988), Oliver and Roussel (1983), Brooks (et al. 2007) and Walker (2007) as well as a reference collection at the University of New Brunswick. When a suitable chronology is available, evidence for the presence of spruce from other authors is included in the figures, results and discussion.

Results

Chronostratigraphy

Representative LOI curves for two sites are shown in Fig. 3.2. Thin Ice Pond is located in southwestern Nova Scotia in the region where spruce has been considered absent prior to the Younger Dryas (Mayle and Cwynar 1995a). Radiocarbon dates at Thin Ice Pond support the use of the LOI as a chronostratigraphy (Table. 3.1) with the exception of the onset of the Holocene. My date of 10729-10229 cal yr BP (Before Present = before 1950) is more than 1000 years younger than the GICC modeled age of 11703 b2k. Dating at the onset of the Holocene is problematic due to the presence of a
radiocarbon plateau (Lotter 1992) spanning that interval; thus, the young result is not surprising. It does, however, fall within the range of dates obtained for the GS-1/Holocene boundary found elsewhere in Atlantic Canada (Mayle and Cwynar 1995a); moreover, the sediment lithology and stratigraphy are typical of the GS-1 throughout the region (Mayle and Cwynar 1993, 1995a, 1995b; Levesque et al. 1993; Whitney et al. 2005; Vincent and Cwynar 2015). The GI-1b event is obscured in sites where the highest abundance of spruce pollen has been found (Sk, Mu, MC and Cr; Fig. 3.1). This may be the result of climate moderation through the GI-1b because of surrounding forest (Bonan 2008). The date obtained from nearby Veinot Lake confirms that a small inflection in the LOI is consistent with the age of the GI-1b event. Where the GI-1b is not apparent, I have not provided samples for the time intervals pre GI-1b and mid GI-1b.

Pollen, chironomid and stomate data are summarized in Tab. 3.2-3.5 and were used to develop maps for each time interval.

**Climate parameters**

What minimum limiting temperature should be used to indicate an environment with summer temperatures suitable for spruce? I have chosen 13°C as this is the lowest inferred July temperature at a site (Li) where spruce occurred (Fig. 3.3), well within the temperature limits defined by Thompson et al. (1999; 11.4-24.4 °C) and the 12-15 °C optimum defined by Anderson et al. (1991). Arguably, spruce can survive at temperatures below 13 °C, but the selection of 13 °C allows for the error in the chironomid temperature transfer function (RMSEP=1.53 °C), reducing the likelihood of a false identification of
disequilibrium and making the hypothesis, H2: southwestern Nova Scotia was too cold to support spruce, more difficult to reject.

**Start of the GI-1b (~13250 cal BP; Fig. 3.4)**

Chironomid inferred temperatures indicate that most of the Province was at least 13°C and therefore warm enough to support spruce at the onset of GI-1b, with the Chignecto Isthmus and the easternmost shore of Nova Scotia being the only significant areas that remained cooler than 13 °C (Fig. 3.4). These areas may have been partially ice-covered before the onset of the Younger Dryas (Stea and Mott 1998; Spooner et al. 2005) resulting in cold temperatures in the vicinity of the ice (Levesque et al. 1997). Green (1987) found that spruce did not reach this area until after the start of the GS-1.

Spruce was present in the central portion of the Province, as proposed by Mayle and Cwynar (1995a), with the highest pollen values, surprisingly, on the south shore at Little Lake (Li). Given the evidence for spruce on the Bay of Fundy shore at BI (Table 3.2-3.5, Fig. 3.4) as well as at Schubenacadie Game Farm (S), Lantz (L), Truro (T) and Blomidon (B) (Stea and Mott 1989; Mott and Stea 1993), I have interpolated the sites where the GI-1b is not clearly evident (see Chronostratigraphy) as part of the area in which spruce is established. Other undated work from Sk (Vincent and Cwynar, unpublished) has yielded abundant spruce stomates at least 10 cm below the onset of the Younger Dryas, suggesting that spruce was likely present in the area by the onset of the GI-1b. Interpolating around Cr, a site without a defined GI-1b, is more difficult without
dated samples. To minimize the risk of overestimating the extent of disequilibrium, I have left Cr out of the region where spruce is considered present.

No evidence for the presence of spruce was found in central and eastern mainland Nova Scotia as well as Cape Breton Island. At LC, spruce pollen values reached 19%, despite the absence of stomates and likely persistence of late ice and snow in the region (Stea and Mott 1998; Spooner et al. 2005; Shaw et al. 2006). Other taxa in the sample are more typical of shrub tundra, with Cupressaceae, Betula, and Dryas occurring in the sample. This assemblage is more consistent with a desert periglacial site (Richard pers comm; Richard 1978; Richard 1994), where local pollen accumulates on nearby glaciers, mixing with pollen that is transported long distances by wind, such as spruce and pine. The pollen on the ice is eventually deposited in lakes during ice retreat or summer melting. Spruce pollen is well adapted to disperse great distances and its contribution from a distance can be much greater than local taxa near the ice, many of which are insect pollinated (Faegri et al. 1989).

**Start of the GI-1a (~13000 cal BP; Fig. 3.5)**

This interval marks the end of a cold event identified as the Killarney Oscillation (Levesque et al. 1993) in which temperatures in southwest New Brunswick decreased as much as 5.0 ºC (Levesque et al. 1993; Whitney et al. 2005). My data indicate that temperatures decreased 0.3-3.8 ºC at all but 3 sites (Fig. 3.5); those sites warmed slightly (Bd, BI and MD; 0.8 ºC, 1.0 ºC and 0.7 ºC respectively). Given that I sampled from the end of the GI-1b, my samples may not reflect the most significant cooling associated with
the event. At Li, pollen values reach 29.9% for spruce, suggesting that it was abundant on
the local landscape. Abundant pollen from other arboreal taxa indicate the presence of
forest while the temperature decreased to 13 °C. Temperatures in southwestern Nova
Scotia, Cape Breton Island and most of the northeastern mainland Nova Scotia remained
at or above 13 °C.

Start of the GS-1 (~12700 cal BP; Fig. 3.6)

In the 300 years between the GI-1b and the onset of the GS-1, temperatures
increased 1-2 °C and the regions suitable to support spruce increased slightly (Fig. 3.6).
At this time spruce has expanded 150km to the northeast whereas the two sites in Cape
Breton Island (Cs and MD) continue to indicate an absence of spruce despite suitable
temperatures (16.0 °C). In the southwest, at TI, temperatures remain above 13 °C with
little evidence for spruce (4% pollen, no stomates). The distribution of spruce at this time
interval is confirmed at S, L, T, B, and C (Mott and Stea 1993; Stea and Mott 1989).

Start of the Holocene (~11600 cal BP; Fig. 3.7)

By the end of the GS-1 (YD), nearly the entire south shore remained colder than
13 °C while many inland sites maintained temperatures above 13 °C. One site, Fn,
contains spruce stomates.
Discussion

Identifying when a species arrives on the landscape can be challenging; consequently, defining a migration lag, or the time between when climate is suitable to support a taxon and when it arrives, can be difficult. Arguably, low pollen values in the pollen record could be indicative of population expansion, not arrival (Vincent and Cwynar 2015; Ammann et al. 2014; Davis et al. 1991; Bennett 1985; Kullman 2008). Earlier trees may have been present on the landscape but left no indication of their presence in the sedimentary record (Peteet 1991; Kullman 1995; Kullman 2001; Vincent and Cwynar 2015). By defining spruce “absent” when pollen values are below 8% and stomates absent, I recognize that spruce may, in fact, be present at levels undetectable by my criteria. However, the fact that clear evidence for spruce at other equally suitable sites (warmer than 13 °C) indicates that either a migration lag has occurred, if spruce is absent, or a population expansion lag has occurred, if spruce is present. In either case, spruce is not present in the densities that could be supported at that site given the suitable temperature and absence of barriers to migration, and this would indicate some degree of disequilibrium between climate and vegetation. Throughout the discussion I will refer to both possible scenarios collectively as a “lag”. Furthermore, I can identify obvious signs of vegetation-climate disequilibrium, using my criteria and comparisons among sites. Equilibrium, however, can be difficult to define and identify (Webb 1986) and is not within the scope of my study. It is possible, for example that a site has a minimum temperature of 13°C and spruce is present, but not at carrying capacity.
Start of GI-1b (Fig. 3.4)

Spruce is established in the central region of the province by the onset of the GI-1b with temperatures near 14-16 °C where spruce is present. Temperatures in the southwest reach 17 °C but sites in this region contain no evidence for spruce. This is much warmer than the central sites where spruce is present at temperatures as low as 14 °C. To the northeast, spruce is absent, as previously noted by Mayle and Cwynar (1995a), although they attributed this to unsuitable conditions due to latitudinal cooling. Temperatures throughout much of the northeast are, in fact, warm enough to support spruce (14-17 °C), resulting in a large area (>200km) into which spruce has not migrated or expanded. This large area is indicative of vegetation-climate disequilibrium. High pollen abundances along the South Shore of the Province support the theory that spruce migrated across the lower Bay of Fundy onto exposed coastline before becoming established in the province (Green 1987). The absence of spruce is confirmed at Collins Pond (C) and Lismore (LIS) (Stea and Mott 1989).

Start of GI-1a (Fig. 3.5)

Vegetation in New England has been shown to respond to climate change at the centennial scale (Shuman et al. 2004). The GI-1b cooling lasted for about 200 years (Rasmussen et al. 2014) and, at this snapshot in time, was not cold enough to prevent spruce from persisting at most sites. In fact, despite decreased temperatures during this cold event and the growth of areas unsuitable for the persistence of spruce in easternmost
mainland Nova Scotia, spruce continued to expand in the Province, particularly toward the southwest, highlighting the degree of migration lag prior to the GI-1b. Even if spruce was present at sites in the northeast at levels undetectable by my investigation, an apparent lag continues to exist as reflected in the higher abundance of spruce in central Nova Scotia. Data from S, L, T, B, C and LIS confirm these results (Mott and Stea 1993; Stea and Mott 1989).

Start of GS-1 (Fig. 3.6)

Spruce expanded 150km to the northeast relative to the previous time slice, leaving only a narrow region of the northeastern mainland in apparent spruce-climate disequilibrium that persisted since before the GI-1b. Spruce populations continued to expand in the central region with spruce values approaching 30%. Further south TI remains free of evidence for spruce, despite suitable temperatures that had persisted for over 500 years. Green (1987) indicated that spruce arrived at Curry Pond, near TI, sometime before the onset of the GS-1. A careful examination of the pollen record from Curry Pond shows no evidence of the significant cooling associated with the GS-1 as found elsewhere in Nova Scotia. Nor does the sediment profile indicate a period of cooling; perhaps an indication that the GS-1 is found in stratigraphically lower sediment than the sample that was collected. This interpretation may be the result of bulk dating inaccuracies that resulted overestimated the early arrival at that site. At LIS and C, the pollen profile does not extend to the end of the GS-1, as that was not the focus of those studies.
Start of the Holocene (Fig. 3.7)

Spruce is absent throughout most of the Province by the end of the YD. The temperatures at this time slice are generally suitable to support spruce, but are not likely indicative of the coldest part of the GS-1. Complete GS-1 profiles from Ws, Cs, MD, TI and Vn (Vincent 2015) indicate that temperatures in the middle of the GS-1 were colder than at the end (8.5 °C, 9.5 °C, 11 °C, 10.6 °C, and 9.6 °C, respectively). By the time the LOI begins to increase, and define the last temperature interval, temperatures have already increased above the YD minimums. Studies elsewhere in the Atlantic Canada indicate that chironomid-inferred water temperatures were lower during the middle of the GS-1 as well, reaching 10-12 °C at the coldest point (Wilson et al. 1993; Levesque et al. 1993; Whitney et al. 2005). Summer water temperature is generally warmer than mean air temperature (Livingstone et al. 1999) placing the middle of the GS-1 well below the 13 °C air temperature limit for the presence of spruce.

Another contributing factor to the absence of spruce in the GS-1 could be the presence or re-expansion of ice as suggested by Stea and Mott (2005). They indicate that an ice lobe crossed the Chignecto Isthmus and expanded in the central and northeastern mainland and Cape Breton Island. Evidence for the presence of ice is apparent at Piper Lake (Spooner et al. 2005) as well as some of my sites. For example, similar to LC at the start of the GI-1b, Cs contains pollen that resembles a desert periglacial assemblage (pers comm; Richard 1978; Richard 1994) suggesting that the site may have been surrounded by ice and snow. LC in this interval contains very little pollen and the LOI record through
the GS-1 approaches 2% organic content. The remaining sites do not resemble periglacial deserts, but persistent ice cover could result in a complete hiatus of sedimentation, which can be difficult to identify.

Stomates are present at the end of the GS-1 at Fn. Reworked material may explain the presence of spruce stomates at Fn as spruce was well established in the area by the onset of the GS-1. However, Mayle and Cwynar (1995a) have suggested that spruce persisted throughout the GS-1 in isolated stands in the south-central area contributing to the rapid recolonization of spruce at the onset of the Holocene. Vincent (2015) indicates that the chironomid inferred air temperatures at Veinot Lake likely remained above 10.5 °C throughout the GS-1. Thompson et al. (1999) indicate that only 10% of white spruce trees surveyed occur at sites where July temperatures are below 11.8 °C while below 7.1 °C, white spruce is entirely absent; black spruce can tolerate even lower temperatures. Without a complete GS-1 temperature record I cannot be certain that spruce survived the entire event at the site, but it does seem likely that Fn and other sites in central Nova Scotia remained warm enough to support spruce throughout the GS-1 and contributed to the rapid recolonization noted by Mayle and Cwynar (1995a).

Without an LOI record, the pollen analyses of Stea and Mott (1989) and Mott and Stea (1993) are difficult to compare to our sites for this time interval. Pollen samples were collected at wide intervals and do not clearly resolve the initial spruce rise typical of the early Holocene. At two of their sites, C and LIS, the pollen records do not extend to the end of the GS-1, as this was not the focus of their study. Where the records are
sufficiently long (L, S, T, B) spruce pollen values plummet at the onset of the GS-1 and remain low throughout, indicating that spruce probably did not return until the Holocene.

Large areas throughout the Province were in vegetation-climate disequilibrium at the end of the GS-1. It is apparent that warming preceded the increase in LOI and the samples chosen in this study may reflect the earliest stages of warming, which have been shown to be extremely rapid (Taylor et al. 1997; Gravech and Severinghaus 2005) and therefore most likely to result in a migration lag or population expansion lag. Without a dated GS-1 temperature profile, the duration of lag at this interval remains unknown.

Conclusions

I can reject H2 that southwestern Nova Scotia was too cold to support spruce prior to the onset of the Younger Dryas. Temperatures in southwestern Nova Scotia were, at every time interval except the start of the Holocene, as warm as or warmer than those at the central Nova Scotia sites where spruce trees were present, supporting the hypothesis (H3) that southwestern Nova Scotia was warm enough to support spruce prior to the Younger Dryas; spruce was not in equilibrium with climate prior to the Younger Dryas in southwestern Nova Scotia. Furthermore, northeastern sites, previously considered too cold to support spruce, also exhibited a spruce-climate disequilibrium. Despite warm temperatures in northeastern Nova Scotia, there existed a considerable lag between spruce and climate through each of the four time intervals (start of GI-1b, start of GI-1a, start of GS-1 and start of Holocene). Spruce could have been present at densities too low to detect, and sites where I have identified spruce is identified as being
present may actually represent expanding populations. This problem was previously acknowledged by Green (1987) in lateglacial Nova Scotia pollen records and observed by Payette and Filion (1984) for white spruce at treeline. But the same argument can be made for expanding populations. If 13 °C is warm enough to support an expanding population at one site but not at a second site, then the second site is indicative of a spruce-climate disequilibrium.

Equilibrium conditions are difficult to define (Prentice 1986), let alone identify, so it remains possible that sites both with and without evidence for spruce are in a state of disequilibrium; in this study comparative evidence for disequilibrium is provided. This comparison indicates that a large area in Nova Scotia experienced vegetation-climate disequilibrium for over 500 years. Given relatively uniform temperatures across the Province, migration toward the higher latitudes progressed more rapidly (500 m/yr) than migration to the southwest. Spruce has been shown to migrate 200-300 m/yr in eastern Canada, behind the retreating ice sheet, and up to 2000 m/yr in western Canada, possibly hastened by seed transport in rivers (Ritchie and MacDonald 1986). Ice was not a barrier to migration in NS prior to the GS-1, and there was no long network of rivers to facilitate the migration of spruce. Williams et al. (2002) have indicated that vegetation can respond to climate change within 100-200 years, much more quickly than the estimated 500 years estimated in this study.

It is interesting to note that, although the GI-1b resulted in 2-3 °C cooling in the northeast and 1 °C in the southwest, spruce continued to increase in abundance with limited southwestward expansion through the GI-1b. Between the GI-1b and the GS-1,
when temperatures increased, spruce migrated steadily into the northeastern unforestected regions. The resulting pattern suggests that spruce can persist and increase in abundance at 13 °C, but temperatures closer to 15 °C are required to facilitate migration. Mature trees can persist in conditions that are not suitable for seedling survival, resulting in differential mortality during periods of climatic stress and reduced migration rates while maintaining existing populations. Payette (1993) has also found that spruce in postglacial Quebec and Labrador expanded its populations without changing its latitudinal position.

The original theory that spruce was absent from northeastern Nova Scotia due to latitudinal cooling was based on a paleovegetational reconstruction that interpreted shrub tundra as indicative of cold climate (Mayle and Cwynar 1995). These results highlight the problems associated with an incomplete understanding of migration lag and strongly support the use of a proxy that is independent of vegetation to identify small-scale paleoclimate events. The use of multiple vegetation proxies to identify the presence of trees beyond forest limits indicated by low pollen values is also recommended (Ritchie 1995, Kullman 1996, 1998, Clayden et al. 1997, Birks and Birks 2000, Gervais et al. 2002, Ali et al. 2003, Birks 2003, Vincent and Cwynar 2015). In the absence of stomate analysis, spruce may have been missed at sites where it is underrepresented by pollen, resulting in an interpretation of limited spruce distribution and greater migration lags that would require faster migration rates to explain the ultimate establishment of forest in Nova Scotia.

In the absence of barriers to migration, vegetation can lag climate and remain in a state of disequilibrium for several centuries over relatively short distances (<50km). The
magnitude of migration lag can also vary over short distances and may be the result of feedback between the migrating plant and its local environment, with trees moderating their own microclimate. The potential for local extinctions of plants that cannot keep pace with changing climate is significant and will be difficult to predict.

Acknowledgments

I would like to thank Ray Spear, Josh Kurek, Tara Warren, Mark Landry, Cy Pedersen and Ian Vincent for fieldwork assistance, Bronwen Whitney for a myriad of contributions, Francine McCarthy for the use of a lab and microscope, Barbara Hansen and Susan Clayden who assisted with stomate identification, and Jack Williams for many helpful comments on an earlier draft of this paper. This research was supported by a Natural Sciences and Engineering Research Council of Canada Discovery Grant to L.C.C.

<table>
<thead>
<tr>
<th>Thin Ice Pond</th>
<th>¹⁴C date</th>
<th>Lab ID</th>
<th>Calibrated age BP</th>
<th>GICC 05 Model b2k*</th>
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<tr>
<td>Start GI-1b</td>
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<td>TO-10421</td>
<td>13223-12710</td>
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Table 3.2. Start of GI-1b: *Picea* pollen, stomate and chironomid inferred temperature (CIT) data. *Picea* pollen values greater than 8% and/or the presence of *Picea* stomates are used to define the presence of spruce. (*Indicates 1 cm slice, ** indicates low sum ~100 grains).

<table>
<thead>
<tr>
<th>Site name</th>
<th>Location</th>
<th>Sample depth (cm)</th>
<th>Rel. Abund.</th>
<th>Spruce pollen (%)</th>
<th>Spruce stomate(s)</th>
<th>CIT (°C) (Air)</th>
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<td>15</td>
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Table 3.3. Start of GI-1a: *Picea* pollen, stomate and chironomid inferred temperature (CIT). *Picea* pollen values greater than 8% and/or the presence of *Picea* stomates are used to define the presence of spruce. (*Indicates 1 cm slice).

<table>
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<th>Site name</th>
<th>Location</th>
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<th>Spruce stomate(s)</th>
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<td>Long Lake (Lo)</td>
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<td>666-666.5</td>
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<td>Mud Lake (Mu)</td>
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<tr>
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Table 3.4. Start of GS-1: *Picea* pollen, stomate and chironomid inferred temperature (CIT). *Picea* pollen values greater than 8% and/or the presence of *Picea* stomates are used to define the presence of spruce.

<table>
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<tr>
<th>Site name</th>
<th>Location</th>
<th>Sample depth (cm)</th>
<th>Rel. Abund.</th>
<th>Spruce stomate(s)</th>
<th>CIT (°C)</th>
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<td>Campbell’s Lake (Ca)</td>
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<td>Chase Pond (Cs)</td>
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<td>China Lake (Ch)</td>
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<tr>
<td>Croskill Lake (Cr)</td>
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</table>
Table 3.5. Start of Holocene: *Picea* pollen, stomate and chironomid inferred temperature (CIT). *Picea* pollen values greater than 8% and/or the presence of *Picea* stomates are used to define the presence of spruce.

<table>
<thead>
<tr>
<th>Site name</th>
<th>Location</th>
<th>Sample depth (cm)</th>
<th>Rel. Abund. Spruce pollen (%)</th>
<th>Spruce stomate(s)</th>
<th>CIT °C (Air)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Borden Lake (Bd)</td>
<td>45.59°N 61.62°W</td>
<td>1019.5-1020</td>
<td>2.1</td>
<td>N</td>
<td>9.6</td>
</tr>
<tr>
<td>Brier Island Bog (BI)</td>
<td>44.21°N 66.37°W</td>
<td>579.5-580</td>
<td>2.2</td>
<td>N</td>
<td>15.7</td>
</tr>
<tr>
<td>Campbell’s Lake (Ca)</td>
<td>43.76°N 65.27°W</td>
<td>565.5-566</td>
<td>5.7</td>
<td>N</td>
<td>11.2</td>
</tr>
<tr>
<td>Chase Pond (Cs)</td>
<td>45.65°N 60.66°W</td>
<td>515-516</td>
<td>13*</td>
<td>N</td>
<td>12</td>
</tr>
<tr>
<td>China Lake (Ch)</td>
<td>44.25°N 64.56°W</td>
<td>400-400.5</td>
<td>3.1</td>
<td>N</td>
<td>12.5</td>
</tr>
<tr>
<td>Croskill Lake (Cr)</td>
<td>45.13°N 63.18°W</td>
<td>464-464.5</td>
<td>3.9</td>
<td>N</td>
<td>16.3</td>
</tr>
<tr>
<td>Finale Pond (Fn)</td>
<td>43.98°N 65.63°W</td>
<td>394.5-395</td>
<td>4.5</td>
<td>Y</td>
<td>15.6</td>
</tr>
<tr>
<td>Hector Lake (Hc)</td>
<td>45.65°N 61.36°W</td>
<td>327.5-328</td>
<td>1.9</td>
<td>N</td>
<td>13.2</td>
</tr>
<tr>
<td>Leak Lake (Lk)</td>
<td>45.43°N 64.34°W</td>
<td>245.5-246.5</td>
<td>4.5</td>
<td>N</td>
<td>12.5</td>
</tr>
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<td>Lac a Magie (LM)</td>
<td>44.27°N 66.08°W</td>
<td>803-803.5</td>
<td>6.8</td>
<td>N</td>
<td>14.5</td>
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<tr>
<td>Li’l Jess Pond (LJ)</td>
<td>44.75°N 63.32°W</td>
<td>416-416.5</td>
<td>4.3</td>
<td>N</td>
<td>12.4</td>
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<tr>
<td>Little Lake (Li)</td>
<td>44.67°N 63.94°W</td>
<td>630-630.5</td>
<td>2.1</td>
<td>N</td>
<td>10.8</td>
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<td>Little Lake C.C. (LC)</td>
<td>45.41°N 63.03°W</td>
<td>528.5-529</td>
<td>-</td>
<td>N</td>
<td>12.6</td>
</tr>
<tr>
<td>Long Lake (Lo)</td>
<td>45.23°N 62.38°W</td>
<td>647.5-648</td>
<td>5.2</td>
<td>N</td>
<td>10.8</td>
</tr>
<tr>
<td>Main à Dieu (MD)</td>
<td>45.98°N 59.84°W</td>
<td>487-487.5</td>
<td>3</td>
<td>N</td>
<td>11</td>
</tr>
<tr>
<td>Marsh Lake (Mr)</td>
<td>44.97°N 62.10°W</td>
<td>775.5-776</td>
<td>2.7</td>
<td>N</td>
<td>10.5</td>
</tr>
<tr>
<td>McInnis Lake (MI)</td>
<td>45.33°N 62.09°W</td>
<td>544.5-545</td>
<td>2.9</td>
<td>N</td>
<td>8.9</td>
</tr>
<tr>
<td>Midconnor Lake (MC)</td>
<td>44.84°N 64.73°W</td>
<td>617-617.5</td>
<td>7</td>
<td>N</td>
<td>12.3</td>
</tr>
<tr>
<td>Mud Lake (Mu)</td>
<td>44.72°N 65.11°W</td>
<td>469-469.5</td>
<td>6.7</td>
<td>N</td>
<td>15.3</td>
</tr>
<tr>
<td>Skating Bench Pond (Sk)</td>
<td>44.40°N 65.07°W</td>
<td>737.5-738</td>
<td>7.2</td>
<td>N</td>
<td>15.2</td>
</tr>
<tr>
<td>Spider Lake (Sp)</td>
<td>44.87°N 62.57°W</td>
<td>379.5-380</td>
<td>2</td>
<td>N</td>
<td>15.2</td>
</tr>
<tr>
<td>Stillman Pond (St)</td>
<td>45.51°N 62.52°W</td>
<td>604.5-605</td>
<td>3.5</td>
<td>N</td>
<td>14.4</td>
</tr>
<tr>
<td>Thin Ice Pond (TI)</td>
<td>43.91°N 65.87°W</td>
<td>548-548.5</td>
<td>1.5</td>
<td>N</td>
<td>11.5</td>
</tr>
<tr>
<td>Veinot Lake (Vn)</td>
<td>44.74°N 64.54°W</td>
<td>797-797.5</td>
<td>5.6</td>
<td>N</td>
<td>13.7</td>
</tr>
<tr>
<td>Western Lake (Ws)</td>
<td>45.34°N 61.50°W</td>
<td>334.5-335</td>
<td>2.0</td>
<td>N</td>
<td>9.1</td>
</tr>
</tbody>
</table>
Figure 3.1. Study sites in Nova Scotia. See Table 3.2 for site name codes.
Figure 3.2. Loss-on-ignition, chronology and sample selection for a) Thin Ice Pond and b) Veinot Lake. All years (printed vertically) are reported in calibrated years before present (cal BP). Samples used for dating marked by circles. Where multiple samples were used, range is marked by black bar. Time intervals selected from start of GI-1b, start of GI-1a, start of GS-1 and start of Holocene, marked by vertical lines. * Indicates moss layer.
Figure 3.3. % *Picea* vs. chironomid-inferred-temperature temperatures (CIT) °C, air). Sites where spruce is present (▲). Sites with no evidence for spruce (〇). All sites with evidence for spruce fall above 13° C.
Figure 3.4. Start of GI-1b chironomid inferred air temperatures in °C. Areas where spruce is present indicated by light grey. Presence is defined by >8% relative abundance of spruce pollen and/or the presence of spruce stomates (stomates indicated by *). Areas too cold to support spruce are indicated by stripes. White land regions indicate where July temperatures were warm enough for spruce (>13 °C) but spruce was not present (spruce-climate disequilibrium). Inset indicates the time slice in relation to the chronostratigraphy. Data for sites (S), Lantz (L), Truro (T) and Blomidon (B), Lismore (LIS) and Collins Pond (C) from (Stea and Mott 1989; Mott and Stea 1993).
Figure 3.5. Start of GI-1a chironomid inferred air temperatures in °C. Areas where spruce is present indicated by light grey. Presence is defined by >8% relative abundance of spruce pollen and/or the presence of spruce stomates (stomates indicated by *). Areas too cold to support spruce are indicated by stripes. White land regions indicate where July temperatures were warm enough for spruce (>13 °C) but spruce was not present (spruce-climate disequilibrium). Inset indicates the time slice in relation to the chronostratigraphy. Data for sites (S), Lantz (L), Truro (T) and Blomidon (B), Lismore (LIS) and Collins Pond (C) from (Stea and Mott 1989; Mott and Stea 1993).
Figure 3.6. Start of GS-1 chironomid inferred air temperatures in °C. Areas where spruce is present indicated by light grey. Presence is defined by >8% relative abundance of spruce pollen and/or the presence of spruce stomates (stomates indicated by *). Areas too cold to support spruce are indicated by stripes. White land regions indicate where July temperatures were warm enough for spruce (>13 °C) but spruce was not present (spruce-climate disequilibrium). Inset indicates the time slice in relation to the chronostratigraphy. Data for sites (S), Lantz (L), Truro (T) and Blomidon (B), Lismore (LIS) and Collins Pond (C) from (Stea and Mott 1989; Mott and Stea 1993).
Figure 3.7. Start of Holocene chironomid inferred air temperatures in °C. Areas where spruce is present indicated by light grey. Presence is defined by >8% relative abundance of spruce pollen and/or the presence of spruce stomates (stomates indicated by *). Areas too cold to support spruce are indicated by stripes. White land regions indicate where July temperatures were warm enough for spruce (>13 °C) but spruce was not present (spruce-climate disequilibrium). Inset indicates the time slice in relation to the chronostratigraphy. Data for sites (S), Lantz (L), Truro (T) and Blomidon (B), Lismore (LIS) and Collins Pond (C) from (Stea and Mott 1989; Mott and Stea 1993).
References


ages for the Younger Dryas in Atlantic Canada. Quaternary Research, 39, 355-360.


(http://www.paleolab.ca/wwwguide/)


Chapter 4

Evidence for the Older Dryas (GI-1d) and GI-1c2 in the eastern
North American lateglacial event stratigraphy

Abstract

European paleoclimate records indicate that the warming trend at the end of the last ice age was interrupted by several cold events such as the Younger Dryas (GS-1), Older Dryas (GI-1d), GI-1c2 and the Killarney/Gerzensee Oscillation (GI-1b). In North America, the lateglacial climate record has only provided evidence of the GS-1 and GI-1b with sparse evidence for the GI-1d, suggesting that the region was less sensitive to the rapid climate oscillations than were recorded in Europe. Using chironomid inferred temperatures and loss-on-ignition records from 2 sites in Nova Scotia, a region highly sensitive to lateglacial climate change, I present evidence for all four lateglacial events. The magnitude of cooling associated with the GI-1d and GI-1c2 in Nova Scotia was similar to records in Europe (2.6-6 °C) and the GI-1d was likely colder than the previously recognized GI-1b at these sites. These results suggest that sites in North America were sensitive to rapid climate events during the lateglacial and that the rapid lateglacial climate oscillations are not unique to sites east of the Atlantic Ocean.
Introduction

The lateglacial (14700-11650 yr) marks the transition between the full glacial conditions of the last Ice Age and our current warm period, the Holocene. This gradual warming trend was interrupted by several short-lived cold events (Lowe et al. 1995; Björck et al. 1998; Rasmussen et al. 2006) that were first identified in European paleoclimate records. The most prominent event was the Younger Dryas cold event (12900-11600 cal BP) first discovered in Europe then, nearly 90 years later, in North America (Mott et al. 1986, Wright 1989). Meanwhile, the European lateglacial record yielded evidence for numerous cold oscillations during the period of glacial retreat and warming that included the Bølling and Allerød Interstadials (BOA), separated by the Older Dryas Stadial. As the number of lateglacial records increased around the North Atlantic, and the complexity of lateglacial climate change became apparent, an event stratigraphy based on the Greenland ice core records that could be used with various proxies, locations, and dating techniques, was proposed by the INTIMATE group (Björck et al. 1998). Lateglacial records in Europe, including Greenland, currently indicate that before the onset of the Younger Dryas (GS-1), the warm Bølling-Allerød (GI-1) was interrupted by the Older Dryas (GI-1d), a recently recognized GI-1c2 (Brauer et al. 2000) and the Killarney/Gerzensee Oscillation (GI-1b). In contrast, North American terrestrial sediments appeared to record only the Younger Dryas Stadial and the Killarney Oscillation (equivalent to the GI-1b: Mott et al. 1986; Peteet et al. 1990; Lowe et al. 1994; Lowe et al. 1995; Wright 1989; Wilson et al. 1993; Mayle and Cwynar 1995; Spooner et al. 2005; Whitney et al. 2005).
Despite differences in estimated timing and intensity, the Older Dryas/GI-1d has been generally correlated to the Older Dryas and the Aegelsee Oscillation in Europe (van Raden et al. 2012; Diefendorf et al. 2006; Heiri et al. 2007; Lotter et al. 1992,) where cooling led to a shift toward arctic herbs, such as *Dryas octopetala*, from which the event earned its name. The Older Dryas/GI-1d cooling resulted in the readvance of glaciers (Ingólfsson et al. 1997), shifts in temperate Mediterranean forest taxa (Fletcher et al. 2010), and oxygen isotope shifts in both the Atlantic and Pacific Oceans (Benson et al. 1997; Hendy et al. 2002; Hong et al. 2009).

The GI-1c2 was not identified as a cold interval in the original GRIP event stratigraphy, although it is conspicuous, and remained unnamed until its discovery in Meerfelder Maar in Germany (Brauer et al. 2000). This cold event interrupts the warm interval GI-1c in the Greenland ice core records and it is clearly visible in the Venezuelan marine records, where it had been previously correlated to the Older Dryas (Hughen et al. 1996). Following the event stratigraphy proposed by Hughen (et al. 1996), Yu and Eicher (2001) and Yu (2007) correlated isotopic shifts in sediments from 2 lakes in North America with the Older Dryas. In Yu’s (2007) records the GI-1c2 is absent and the correlation of an event with the Older Dryas/GI-1d is not clearly established. These appear to be the only records to date in which events correlative with the Older Dryas (GI-1d) and GI-1c2, have been identified in North America to date, despite much work on the Younger Dryas and Killarney Oscillations (Mott et al. 1986; Wright 1989; Peteet et al. 1990; Levesque et al. 1993; Mayle and Cwynar 1995; Spooner et al. 2005; Whitney et al. 2005). Neither the Older Dryas nor the GI-1c2 was definitively recognized as
Lake sediment collected in Nova Scotia, Canada has yielded several high resolution, detailed records of lateglacial climate change (Mott et al. 1986; Wright 1989; Levesque et al. 1993; Levesque et al. 1997; Mayle and Cwynar 1995; Spooner et al. 2005; Whitney et al. 2005). The proximity of these sites to treeline made them especially sensitive to fluctuating temperatures. Studies using pollen, plant macrofossils, conifer stomates and fossil insects all record changes associated with the Younger Dryas (GS-1) cold event and the earlier Killarney Oscillation (GI-1b), but not the Older Dryas (GI-1d) and GI-1c2. Each of these events is well dated within the Greenland Ice Core Chronology 2005 (GICC05) that synchronizes all the Greenland Ice core records into a single, well-dated event stratigraphy (Lowe et al. 2008).

Here I present data from 2 sites in Nova Scotia adjacent to the lateglacial forest limit that provide evidence for the Older Dryas (GI-1d) and GI-1c2.

**Study sites and methods:**

Thin Ice Pond (43° 54’ 38” N, 65° 51’ 52” W) and Veinot Lake (44° 44’ 09” N, 64° 32’ 18” W) were cored using a 7.5 cm modified Livingstone piston corer (Wright 1967) in the deepest portion of each lake (Fig. 4.1). Loss-on-ignition analysis was carried out at 0.5 cm intervals (Dean 1974; Heiri et al. 2001). The organic content of lake
sediiments is an approximate index of lake water temperature (Heiri et al. 2001) and this relationship holds especially true in lateglacial Nova Scotia where LOI has been successfully used an indicator of cooling events and as a chronostratigraphic marker (Levesque et al. 1993, 1997, Whitney et al. 2005). Midge processing and head capsule minima follow Walker (2001). A minimum equivalent of 50 midge head capsules was counted per sample (Heiri et al. 2001 and Quinlan and Smol 2001). Midge head capsules were identified according to Oliver and Roussel (1983) and Brooks et al. (2007) as well as a reference collection at the University of New Brunswick. Midge-inferred July lake surface water temperatures were calculated using a chironomid temperature transfer function developed by Francis et al. (2006).

Accelerator mass spectrometry 14C dates were collected from plant macrofossils for 6 samples from Thin Ice Pond and 2 samples from Veinot Lake. All 14C dates were calibrated using Calib version 6.0 with 2-sigma error estimate (Stuiver and Reimer 1993; Reimer et al. 2009). Events of interest were dated directly for comparison to the GICC05 (Lowe et al. 2008).

Results

Midge analysis

Midge taxa from Thin Ice Pond and Veinot Lake are well represented by the assemblages in the modern calibration dataset developed by Frances et al. (2006) for air temperature, indicating that the transfer function should perform well at these sites (Birks
The proportion of taxa in individual samples not represented by the transfer function at Thin Ice Pond ranged from 0-10.4% (average 4.3%). At Veinot Lake unrepresented taxa composed 0-7.3% (average 3.3%) of the total assemblage in each sample. Chironomid-inferred-temperatures (CIT) for both lakes have been plotted below the Greenland Ice Core Chronology 05 (GICC05) for ease of comparison (Fig. 4.2 a-c).

**Younger Dryas**

At both sites the Younger Dryas (GS-1) event is clearly indicated by a large reduction in the organic content of the lake sediments (Fig. 4.2). This LOI profile of the YD is typical of lateglacial sites in Nova Scotia (Mott and Stea 1993; Wilson et al. 1993, Levesque et al. 1993, Mayle and Cwynar 1995, Whitney et al. 2005) and provides a stratigraphic marker for lateglacial climate studies. The same trend is found in the chironomid-inferred temperatures that decrease 7-12 ºC through the Younger Dryas (GS-1).

**Killarney Oscillation (GI-1b)**

The Killarney Oscillation (GI-1b) is clearly visible in the Thin Ice Pond LOI (Fig. 4.2 b) record as a 70% decrease in the sediment organic content from 582-570 cm (Fig. 4.2) but is more subdued at Veinot Lake (Fig. 4.2c; 827-823cm), despite the widespread evidence for cooling throughout Nova Scotia (Levesque et al. 1993). Cooling associated with the Killarney Oscillation (GI-1b) at Thin Ice pond approaches 5ºC. At Veinot Lake, there is no clear temperature response to the Killarney Oscillation.
Older Dryas (GI-1d)

A smaller, prominent, reduction in organic content is evident prior to the Killarney Oscillation (GI-1b), reaching its minimum at 591 cm and 861 cm at Thin Ice Pond and Veinot Lake respectively (Fig. 4.2 b,c). Radiocarbon dates place this event just after 13480-14880 cal BP at Thin Ice Pond and 13780-14050 cal BP at Veinot Lake, which correlate well with the Older Dryas (GI-1d) in the GICC05 dated to 13954-14075 yr b2k. Chironomid-inferred temperatures indicate that cooling at both sites is greater than 4 ºC.

GI-1c2

Between the Older Dryas (GI-1d) and the Killarney (GI-1b), LOI values continue to climb, representing the warm interval in the GICC05 (Fig. 4.2a) identified as the GI-1c, which is also interrupted by a brief cooling event called the GI-1c2 (Brauer et al. 2000; Lowe et al. 2008). At Veinot Lake (Fig. 4.2c), a small inflection beginning near 848 cm is associated with a decline in temperature of 6.2 ºC that reaches a minimum just after the LOI curve begins to increase again. At Thin Ice Pond (Fig. 4.2 b), there is no obvious inflection in the LOI and temperatures decline a modest 2.6 ºC at 583.5 cm, before the onset of the Killarney Oscillation (GI-1b). Stratigraphically situated between the Killarney (GI-1b) and Older Dryas (GI-1d) events, this smaller reversal correlates with the GI-1c2. AMS dates from Veinot Lake confirm this interpretation, placing the
end of the event at 13415-13730 cal BP, well within the GICC05 dates for the GI-1c2 at 13660-13600 yr b2k.

Following the GI-1c2, temperatures at Veinot Lake do not reach previous maximum values, indicating that this site may have experienced the general cooling trend typical of the lateglacial found elsewhere (Brooks et al. 2012; Yu and Eicher 2001; Grootes et al. 1993; Johnsen et al 1992; Dansgaard et al. 1993; von Grafensteine et al. 1999).

**Discussion**

The magnitude of cooling during the Older Dryas (GI-1d) event in Nova Scotia is similar to that found outside the region. The Older Dryas (GI-1d) in Europe and Greenland resulted in cooling of 1.5-6 °C (Lotter et al. 1992; Johnsen et al. 1995; Brooks and Birks 2000; Watson et al. 2010; van Asche et al. 2012; Brooks et al. 2012) most notably at sites most strongly influenced by coastal climate. There has been little evidence in North America for the Older Dryas (GI-1d). Yu (2007) and Yu and Eicher (2001) reported multiple climate oscillations, one of which is the Older Dryas (GI-1d), at Crawford Lake in Ontario and at White Lake in the northeastern USA. They found chironomid inferred temperatures cooled 1-2 °C at Crawford Lake, nearly 1000 km from the Atlantic Ocean, and 3 °C at White Lake, within 100 km of the ocean. Given that lateglacial climate is driven, at least in part, by changes in ocean circulation (Clark et al. 2001; Pearce et al. 2013), it is not surprising to see the cooling effects muted at sites further from the coast. Assuming relatively constant sedimentation rates, the Killarney Oscillation (GI-1b) appears to have endured longer than the Older Dryas (GI-1d) but was
not as cold. The Gerzensee Oscillation in European records is longer and less abrupt, but is not consistently warmer than the Older Dryas (Lotter et al. 1992; Johnsen et al. 1995; Brooks and Birks 2000; Heiri et al. 2007; Watson et al. 2010; Brauer et al. 2012; Van Asche and Hoek 2012; Van Asche et al. 2012; Brooks et al. 2012). The Older Dryas may have not been identified in previous paleoclimate studies in Nova Scotia because of its brief duration (Levesque et al. 1993; Whitney et al. 2005).

The GI-1c2 event has not been recognized in many climate reconstructions although its signal is apparent in a number of marine records and the Greenland Ice core stratigraphy (Hughen et al. 1996; Björck et al. 1998; van Asche and Hoek 2012; van Asche et al. 2012; van Raden et al. 2012). Brauer et al. (2000) recognized this event as being short and not severe, which seems typical of its record in Europe (Brauer et al. 2000; Wall et al. 2010). Van Asche et al. (2012) found a small reversal in $\delta^{18}O$ from a site in Ireland that corresponds to ~0.5 °C cooling. Cooling in Nova Scotia was much more dramatic with a 2.6-6 °C reversal.

The GI-1c2 has not been previously identified in North America and was misidentified in Yu (2007) and Yu and Eicher (2001). Yu (2007) found “the chronology at White Lake places the major climate transitions consistently 300–400 yr younger than the timing indicated from Greenland and elsewhere”. This lake record is similar to the GICC05 record, but appears to be misaligned with the stratigraphy in Björck et al. (1998) and then refined by Lowe et al. (2008). Most notably, the GI-1c2 was likely been misidentified as the Older Dryas (GI-1d) and the Older Dryas was designated as the intra-Bølling Cold Period. If Yu (2007) had correlated his findings to the GRIP stratigraphy as
laid out by Björe k et al. (1998) rather than the Venezuelan marine records (Hughen et al. 1996) the 300-400 year discrepancy would be significantly reduced and the GI-1c2 would have been apparent, as the GRIP stratigraphy provides a better chronological fit to the European terrestrial records and, now, North American records as well.

The clearer record of rapid events found at Thin Ice Pond relative to Veinot lake is likely due to differences in vegetation surrounding the sites. Previous vegetation studies indicate that forest had not reached southwestern Nova Scotia, near Thin Ice Pond, prior to the Younger Dryas, but had migrated as far as Veinot Lake (Railton 1972; Green 1987; Mayle and Cwynar 1995; Vincent and Cwynar 2015). Maps of spruce distribution support this interpretation (Vincent 2015, chapter 3). Pollen evidence from nearby Little Lake (Mayle and Cwynar 1995) as well as observed stomates in samples from unpublished work from Veinot Lake suggests that trees arrived in the area after the Older Dryas (GI-1d) and before the GI-1c2. Forest continued to expand in the area of Veinot Lake through the Older Dryas (GI-1d), until the onset of the Younger Dryas. The Killarney Oscillation (GI-1b) has been recorded in Atlantic Canada with temperature reversals near 5 ºC. The establishment of boreal forest, which has been shown to increase local air temperatures by 5 ºC (Bonan et al. 1992), may have caused each subsequent climate event to be progressively more obscured, with the Killarney Oscillation (GI-1b) being completely absent from the sediment record at sites surrounded by forest.

**Conclusions**

The proximity of Thin Ice Pond and Veinot Lake to both the retreating ice sheets and the Atlantic Ocean has resulted in clear records of cooling associated with the Older
Dryas (GI-1d) and the GI-1c2 cold events. The clarity of climate records can be significantly altered by the presence of forest on the landscape, suggesting that the most sensitive locations for climate study lie near the boundaries of vegetation, where temperatures are warm yet forest is not yet established and creating its own microclimate. The discovery of these events should lay to rest the assumption that the lateglacial climate oscillations that are typical of European terrestrial and Greenland ice core records did not affect sites west of the Atlantic Ocean. Furthermore, the Older Dryas (GI-1d) was colder, and possibly briefer than the Killarney Oscillation (GI-1b) event.

Recognizing that lateglacial climate events were experienced throughout the North Atlantic, with similar magnitude and timing, improves our ability to understand the effects of future climate events and their impacts in North America. Rapid climate responses to changes in ocean circulation are not limited to sites east of the North Atlantic.
Figure 4.1. Site locations for Veinot Lake and Thin Ice Pond.
Figure 4.2. a) GICC05 20-year average plot (Rasmussen et al. 2006) with stratigraphy following Björck (et al. 1998) plus the GI-1c2 (Brauer et al. 2000). b) Thin Ice Pond and c) Veinot Lake % loss-on-ignition (LOI; black) and chironomid-inferred-temperatures (CIT) lake surface temperature (°C; grey) with calibrated dates (Stuiver and Reimer 1993) marked by wedges above the x-axis.
References


Fletcher, W.J., Sanchez Goñi, M.F., Peyron, O. and Dormoy, I., 2010. Abrupt climate changes of the last deglaciation detected in a Western Mediterranean forest record. Climate of the Past, 6, 246-264.


temperature reconstruction from southwestern Nova Scotia. Canadian Journal of Earth Science, 42, 2051-2057


Chapter 5

Rapid warming at the Younger Dryas (YD) – Holocene transition

was interrupted by a cooling event

Abstract

The current slow down in the rate of global surface warming has been attributed to changes in atmospheric circulation and heat absorption in the deeper oceans, particularly in the Northern Hemisphere. The Younger Dryas-Holocene transition has often been used as analogue to modern climate change but has also been viewed as proceeding progressively from low to higher values without interruption. If the same forces are acting on the climate today as during the transition, then we might expect to find a similar slow down or hiatus in warming in the paleoclimate record. Using high-resolution lake water temperature records from North Atlantic coastal sites, I show that the Younger Dryas-Holocene transition was in fact interrupted by a brief period of cooling (1.6–6.4 °C). The Younger Dryas-Holocene slow down has also been noted in numerous terrestrial and marine records from around the North Atlantic and beyond, suggesting that the current hiatus in warming is not unique, but quite possibly an expected feature of rapid global warming events.
**Introduction**

Current research into the state of global warming has suggested that the recent, unexpected slowdown in the rate of surface warming may be the result - at least in part - of increased heat storage in the deeper oceans coupled with a weakening stratospheric polar vortex (Chen and Tung 2014; Kim et al., 2014). Millennial-scale oscillations in surface temperatures have been found in the marine and terrestrial records that span the Pleistocene, triggered by changes in complex interactions between melt water flux, ocean circulation and the atmosphere (Clark et al. 2001; Monnin et al. 2001; Kirby et al 2002; Chen and Tung 2014; Kim et al. 2014). However, no slowdown has been recognized during a period of rapid warming analogous to the recent one, where the climate response is sensitive to not only the magnitude, but also the rate of changing circulation and melt water inputs (Liu et al. 2009).

The warming trend that followed the Younger Dryas cooling (also GS-1 following terminology of Björck et al. 1998 and Lowe et al. 2008) and ushered in the Holocene interglacial was rapid and the latest of a number of steps representing the transition from the last Ice Age to the present interglacial (Dansgaard et al. 1989). This rapid transition has often been used as an analogue for our current state of global warming and seems a good starting point in a search for an analogous hiatus. The Greenland ice core records have been used as a template for climate change in the Northern Hemisphere (Björck et al. 1998) and for formally defining the start of the Holocene (Walker et al. 2009). These records indicate that temperatures in Greenland rose 7°C in less than 15 years (Taylor et al. 1997) and ultimately 10°C (Gravech and
Severinghaus 2005), terminating the 1300-year-long Younger Dryas (YD) cold event. They also indicate that the transition to the Holocene was interrupted by a brief climate reversal (Stuiver and Grootes 2000; Grootes and Stuiver 1997; Stuiver et al. 1995; Meese et al. 1994) that has not previously been examined.

I collected sediment records (Fig. 5.1) from eastern Canada and reviewed previously published terrestrial and marine records spanning the YD-Holocene transition to examine the possibility that the current hiatus in warming is not unique but maybe a characteristic feature of rapid warming in response to the interactions between the atmosphere and oceans. I have produced new reconstructions of surface lake water temperatures using fossil chironomids and sediment organic content analyses from a dense network of sites located near the North Atlantic Ocean. Here I found that the YD-Holocene transition was interrupted by a decadal scale hiatus in warming at ca. 11550 that produced local cooling of 1.6-6.4 °C. This cooling was also apparent in records from throughout the North Atlantic, with some evidence for cooling in the eastern Pacific and Arabian Sea as well. These results contradict previous notions that the transition was rapid and uninterrupted and suggest that brief slowdowns or hiatuses may be a characteristic feature of rapid warming because of the complex interactions between the oceans and atmosphere.

**Chronology and Loss-on-ignition profiles**

Loss-on-ignition (LOI) has been used to identify century scale climate events in lateglacial Nova Scotia (Levesque at al. 1993, Mayle et al. 1993, Whitney et al. 2005). As
such, LOI is likely to be a suitable technique for finding an hiatus during a period of climate warming. Thin Ice Pond (Fig. 5.2a) exhibits a typical LOI profile for lateglacial-aged lakes in Nova Scotia. Using the LOI profiles, I identified the YD at 26 sites. The portion of the lateglacial curve starting at the transition into the Holocene (the first LOI points consistently increasing) and ending within the Holocene (the point at which LOI values begin to level off and often decrease slightly) for each of 18 lakes is shown in Figure 5.2b. These lakes were selected because they highlight a pattern of warming at the end of the YD that presents itself as a “step” in the LOI profile. Seven of the 26 sites originally cored are not included in this study due to low sedimentation rates (1.5-4.0 cm compared to mean transition length of 30 cm), resulting in an inability to resolve the details of the YD-Holocene transition. The sixth site, with suitable resolution (8.5 cm), lacks a step and lies in an area that was likely ice-covered throughout the YD (Stea and Mott 1998).

The step at each site forms a plateau or reversal in LOI values ranging from 6.5-30 cm in length with similarities within regions. Some sites exhibit broad steps, (TI, LM, Ca, LJ, Sk, Li) while others are narrower (MC, MD, Fg, Lk, Cs), likely reflecting differences in local sedimentation rates. At several sites (LM Ca, LJ, Vn, LC, Fg, Ws. Fig. 5.2b) the step is double-pointed and at three sites from central Nova Scotia the step has a subtler concave profile before rising into the Holocene (Figs. 5.2b; Ch, Sk, Li). The northern sites generally have simpler, narrower steps. Despite the differences in the shape of the step, there is clearly a widespread phenomenon that is causing a shift in the organic content of lake sediment during the YD/Holocene transition. Given that the organic
content of lateglacial sediments in Nova Scotia fluctuates with temperature (Levesque et al. 1993), the step provides indirect evidence for a pause or small reversal in the warming trend during the YD/Holocene transition.

**Chironomid-inferred temperature profiles**

Direct evidence for changes in temperature during the YD/Holocene transition is found in the chironomid inferred temperature reconstructions for 5 lakes representing southern, central and northern Nova Scotia (Fig. 5.2c). Details can be found in Supplemental Information. At Main à Dieu Pond (MD), the temperature profile parallels the LOI curve and forms a small plateau in the otherwise steep warming curve. At Thin Ice Pond (TI), Veinot Lake (VL), Chase Pond (Cs) and Western Pond (Ws) the chironomid-inferred temperatures decline in samples from the LOI inflection by 2.2°C, 1.6°C, 3.8°C and 6.4°C respectively. Together the LOI and chironomid inferred temperatures suggest a brief period during the YD/Holocene transition when the warming trend temporarily stalled and often reversed. The widespread prevalence of the step at my sites indicates that this event was regional in extent and driven by large scale forcing factors rather than local ones.

**The Younger Dryas to Holocene transition outside Nova Scotia**

Numerous examples of inflections and reversals occur in various proxy records from both marine and terrestrial sites throughout the North Atlantic (Figs. 5.3, 5.4, 5.5).
Sites lacking evidence for a step are not included in this figure as this short event may either be absent or simply missed due to sampling resolution. The transition into the Holocene was rapid (Taylor et al. 1997, Gravech and Severinghaus 2005) and the step in my records may have been deposited in as little as 10 years (Ralska-Jasiewiczowa et al. 2003).

**Terrestrial Evidence**

The continuous sample plot (high resolution) of the GISP2 $\delta^{18}O$ record and the Greenland Ice Core Chronology 2005 (GICC05) records both indicate a pause in warming in the YD-Holocene transition (Fig. 5.4; Walker et al. 2009; Stuiver and Grootes 2000; Grootes and Stuiver 1997; Stuiver et al. 1995; Meese et al. 1994; Lowe et al. 2008). Not surprisingly, I have found the same step elsewhere (Fig. 5.5), as the Greenland ice core records are considered the stratotype for North Atlantic climate change. For example, Schwander (et al. 2000) have correlated the transition, including two peaks, in Swiss lakes Gerzensee and Leysin to the GRIP record (Fig. 5.5d). Several of the Nova Scotia sites also exhibit a double peak, which may indicate even finer scale climate fluctuations that would require very high-resolution sampling to identify. The double peak is also evident at the lateglacial Bølling Sø site in Denmark, generated from organic content and biogenic silica analysis (Sarmaja-Korjonen et al. 2006), and in the pollen abundance of several tree taxa in southern central Norway (Paus et al. 2011) suggesting that conditions briefly deteriorated and prevented the continued expansion of
birch, pine and juniper. A similar reversal in pine pollen values is seen on the Mid-Atlantic region of the USA (Yu 2007).

At Whitrig Bog, Scotland, a temperature reversal of nearly 2°C at the YD/Holocene transition reflects a brief increase in the cold tolerant chironomid, *Heterotrichocladius* (Mayle et al. 1997, Brooks and Birks 2000a, Brooks and Birks 2000b). A similar excursion was noted at Kråkenes Lake, Norway (Brooks and Birks 2001- Fig. 5.5b). However, the authors suggest that, although the chironomid temperature transfer function being used was to infer air temperature, cold melt water from a nearby cirque glacier had a more significant impact on the chironomid assemblage than did air temperature. Although this explanation is plausible, the timing of the Kråkenes reversal and its similarity to my chironomid temperature profiles is intriguing, especially in light of similar Norwegian chironomid temperature and pollen profiles (Paus et al. 2011).

Further south, Peñalba et al. (1997) also recorded a 10 ºC temperature reversal in a Spanish lateglacial pollen sequence (Fig. 5.5f).

Similar stepped profiles have been found in numerous other terrestrial studies (Fig. 5.3 and Fig. 5.5) using oxygen isotopes, fossil beetle remains, pollen, testate amoeba accumulation rates, chironomids and sediment geochemistry, as well as the inferred temperatures generated from some of these proxies (Atkinson et al. 1987; McCarthy et al. 1995; Nolan et al. 1999; Schwander et al. 2000; Brooks and Birks 2001; Cwynar and Spear 2001; Ralska-Jasiewiczowa 2003, Lücke and Brauer 2004; Wall et al. 2010; Verbruggen et al. 2010). In most of these studies, the presence of a cold reversal
during the transition into the Holocene was not discussed as the focus of the research lay elsewhere.

**Marine evidence**

A large body of evidence to support the occurrence of a cold reversal during the YD/Holocene transition also lies in the marine records. Numerous marine records (Fig. 5.3) possess YD/Holocene profiles with a step, often remaining unnoticed or unremarked upon by the authors (Keigwin and Jones 1989; Lehman and Keigwin 1992; Behl and Kennett 1996; Rasmussen et al. 1996; Kroon et al. 1997; Leuschner and Sirocko 2000; Cacho et al. 2001; Pailler and Bard 2002; Martrat et al. 2003; McManus et al. 2004; Piotrowski et al. 2004; Praetorius et al. 2008; Rodrigues et al. 2010). The step appears in marine records using various water mass tracers and climate proxies from the North Atlantic, Mediterranean Sea, Arabian Sea and Antarctic (Fig. 5.5) suggesting changing SSTs, salinity, deep-water production rates and shifting oceanic fronts (Pearce et al. 2013; Bard et al. 1987, Ruddiman et al. 1977). Brief changes in ocean currents that influence Nova Scotia climate are apparent in a high resolution record from Placentia Bay, Newfoundland, as small plateaux, peaks and reversals in diatom taxa in the middle of the transition between the Younger Dryas and the Holocene (Pearce et al. 2013).
Timing and duration

Although the YD/Holocene transition is simple to correlate stratigraphically, previous efforts to date the YD/Holocene transition in Nova Scotia have yielded very large error estimates -11399-11976 cal BP (Mayle and Cwynar 1995) due to a radiocarbon plateau that spans the end of the YD. The key to dating this event lies in annually resolved sediments, such as ice core records and varved lakes. Ralska-Jasiewiczowa (et al. 2003) found that the YD termination, dated 11550-11460 cal BP was made up of two distinct warming periods separated by a decade long transition, characterized by increased precipitation, around 11520 cal BP in varved sediments from Poland. The frequently published 20-year resolution plot of GISP2 $\delta^{18}$O suggests that YD/Holocene warming was uninterrupted, however, a continuous sample plot of the GISP2 $\delta^{18}$O record shows the YD/Holocene transition from 11650-11600 and includes a double peak with the initial and largest reversal lasting 10 years from 11640-11630 (Fig. 5.4) (Stuiver and Grootes 2000; Grootes and Stuiver 1997; Stuiver et al. 1995; Meese et al. 1994). Similarly, the GICC05 high-resolution record indicates a pause in warming at 11725 ka b2k (Walker et al. 2009). The Swiss lakes Gerzensee and Leysin sediment records (Schwander et al. 2000) have been correlated to the GRIP core and date the transition to 11535-11487 cal BP (Fig. 5.5d). The differences in timing between these studies are within the timescale offset between GICC05 and calibrated $^{14}$C dates (Lohne et al. 2013) and within the 99 year counting error of the NGRIP record (Lowe et al. 2008; Rasmussen et al. 2006). Despite these differences and difficulties, stratigraphically correlated sources do suggest that the transition into the Holocene spanned ~ 50-100
years (Lücke and Brauer 2004; Severinghaus et al. 1998) and that the cooling associated with this step began in the incipient stages of warming and lasted about a decade.

A detailed study of the YD/Holocene transition by Pearce et al. (2013) suggests that the ocean led the atmosphere at the end of the YD. Their data indicate a brief pause in warming marked by a plateau in an otherwise increasing abundance of warm indicator diatoms (*T. nitzschioides*), producing a step beginning at 11600 and lasting until 11500 cal BP, falling between 2 periods of significant warming.

**Synthesis**

Regional climate trends emerge when the marine records are considered together with the terrestrial evidence. During the incipient Holocene, Arctic sea ice initially decreased in extent, and then briefly increased, before a final retreat to modern levels (Fig. 5.5g, Piotrowski et al. 2004; Pearce et al. 2013). Stratigraphically coincident, during the early Holocene, the expanded circumpolar vortex briefly contracted around 11,600 B.P. then returned to its median Holocene position (Kirby et al. 2002). Sea surface temperatures increased, briefly decreased, and then continued to warm to current levels (see marine evidence). Sea surface salinity appears to have followed the same trend as temperature, as did terrestrial lake water temperatures in northern Europe and pollen inferred temperatures from Spain, the latter attributed to the brief decline in SSTs (see terrestrial evidence).

The current warming hiatus has also been tentatively attributed to a slower AMOC driven by salinity and SST changes in the North Atlantic, with excess heat being
stored in deeper waters (Chen and Tung 2014). Kim et al. (2014) have suggested that dramatic losses in sea ice may have contributed to recent low temperatures in the mid latitude Northern Hemisphere due to weakening of the polar vortex.

The similarity between the contemporary surface cooling, reduced salinity and circulation along with an unstable circumpolar vortex just decades into a period of rapid warming is intriguingly similar to the cold reversal at the YD/Holocene transition and suggests that the same mechanism is responsible, i.e., increased summer warming at the equator (Berger and Loutre 1991) leading to an increased AMOC that carries warm water to the subpolar Atlantic. The excess heat leads to a rapid reduction in Arctic sea ice (Pearce et al. 2014), which in turn reduces surface salinity and density at high latitudes and allows for the destabilization of the circumpolar vortex (Kim et al. 2014). The reduced density difference eventually slows the AMOC, resulting in the brief cooling evident in so many North Atlantic records while the contracted circumpolar vortex leads to colder mid latitude temperatures and an increase in cold continental storms in North America (Kirby et al. 2002). An investigation into the changing temperatures of deepwater in the paleoceanographic record may shed some light onto whether rapid surface cooling with deep water heat storage is a typical part of warming or not.

The widespread evidence for a brief cold reversal that reveals itself as a step in terrestrial and marine records throughout the North Atlantic and beyond dispels previous notions that the YD termination was a brief period of rapid, uninterrupted warming. While our modern experience of global warming is often doubted in the face of occasional periods of bitter cold winters and disappointing summers, the Nova Scotia
paleoclimate records suggest that cold periods may be a normal part of rapid warming trends.

Acknowledgments

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Figure 5.1. Site map showing position of lakes cored for this study. Site names: Lac à Magie (LM), Thin Ice Pond (TI), Campbell’s Pond (Ca), Skating Bench Pond (Sk), China Lake (Ch), Midconnor Lake (MC), Veinot Lake (Vn), Little Lake (Li), Li’l Jess Pond (LJ), Leak Lake (Lk), Little Lake (Colchester County) (LC), Foghorn Pond (Fg), Long Lake (Lo), McInnis Lake (MI), Western Pond (Ws), Borden’s Lake (Bd), Chase Pond (Cs), Main à Dieu Pond (MD).
Figure 5.2. Lateglacial loss-on-ignition curves for Thin Ice Pond, Nova Scotia, Canada. This sediment profile is typical of lateglacial-aged sediment in Nova Scotia and is used to chronostratigraphically correlate between sites. Ages are reported using median cumulative probability for calibrated years BP (cal BP) according to Stuiver and Reimer (1993). All dates fall within the high probability region of the radiocarbon calibration curve. Steps indicated by thick grey bar. Box provides context for figure 5.2b. Loss on ignition profile for the Younger Dryas to Holocene transition for 18 Nova Scotia sites. The context of these profiles is indicated by box in figure 5.2a. Cooling event, or step is indicated by thick grey bar. Moss layer marked by plus (+). c) Chironomid inferred temperatures (CIT) for 5 Nova Scotia sites. Step is indicated by grey bar. Temperature profiles are plotted on the same x-axis scale as the LOI for the same lake, shown immediately above in figure 5.2b.
Figure 5.3. Map of sites in which YD/ Holocene transition steps and/or temperature reversals can be seen. Complete references in supplemental material.
Figure 5.4. a) GISP2 d18O data for YD/Holocene transition (Stuiver and Grootes 2000; Grootes and Stuiver 1997; Stuiver et al. 1995; Meese et al. 1994). Transition 11650-11600 indicated by asterisks (*). Step indicated by thick grey bar. b) Position of circumpolar vortex by latitude during YD/Holocene transition (Kirby et al. 2002). Shoulder indicated by thick grey bar.
Figure 5.5. Evidence of step at marine and terrestrial sites from various authors. Step in each proxy is indicated by an arrow. a) $\delta^{18}$O benthic foraminifera from Site 984 on the Reykjanes Ridge in the North Atlantic, after Praetorius (et al. 2008). b) Chironomid inferred mean July air temperatures from Kråkenes Lake, Norway. After Brooks and Birks (2001). c) SSTs based on alkenone $U^{k_{37}}$ from shallow core D13882 on the western Iberian Margin. After Rodrigues (et al. 2010). d) Oxygen isotope analyses of bulk sediment samples from Lake Gerzensee, Switzerland. After Schwander (et al. 2000). e) SST based on alkenone $U^{k_{37}}$ from core MD95-2043 in the Alboran Sea. (After Cacho et al. 2001). f) Mean July temperature presented as an anomaly from modern July temperature (17 °C) based on pollen from Quintanar de la Sierra, Spain. After Peñalba (et al. 1997). g) Nd isotope ratios of Antarctic core RC11-83 correlated to Western North Atlantic sea ice cover from core 91-045-094. After Piotrowski (et al. 2004). h) Changing dust supplies indicated by CaCO$_3$ % from core 70KL in the Arabian Sea. After Leuschner and Sirocko (2000).
References


Clark, P.U., Shakun, J.D., Baker, P.A., Bartlein, P.J., Brewer, S., Brook, E., Carlson,


during the last deglaciation. Nature, 356, (63720), 757-762.


Supplemental information

Methods

Field methods

I collected complete lateglacial records from cores extracted from the deepest part of 24 small (<10 ha) lakes without inflowing streams throughout Nova Scotia (Fig. 5.1) using a modified Livingstone piston corer (Wright 1967). Lateglacial sediments in Nova Scotia possess distinct lithologies that permit detailed sampling of visually evident events such as the Killarney Oscillation (GI-1b), the YD and the transition from the YD to the Holocene (Levesque et al. 1993; Mayle et al. 1993).

Laboratory methods

Each core was analyzed for LOI (Dean 1974, Heiri et al. 2001) at 0.5 or 1.0 cm contiguous increments from the base of the core into the early Holocene so that the entire YD was apparent in the plotted LOI curve. Based on the LOI curves, samples were selected for chironomid analysis and for macrofossils for radiocarbon dating. Chironomid samples were prepared as outlined in Walker (2001) with a minimum head count of 50 (Quinlan and Smol 2001). Chironomid head capsules were identified using Wiederholm, (1983), Walker (1988), Oliver and Roussel (1983) Brooks (et al. 2007) and Walker (2007), as well as a reference collection at the University of New Brunswick. Maximum surface water temperature estimates were calculated for each sample using the temperature transfer function of Walker et al. (1997).
Methods References:


References for Figure 5.3


31) J. Schwander, U. Eicher, B. Ammann, Oxygen isotopes of lake marl at Gerzensee and Leysin (Switzerland), covering the Younger Dryas and two minor oscillations, and their


Chapter 6

Synthesis and Conclusions

Using fossil spruce pollen, macrofossils and stomates I have refined the pollen limit for identifying the local presence of spruce to 8%. With the refined limit, I find that spruce had a larger distribution in the province than previously thought based on the 20% isopoll (Anderson et al. 1991). In fact, spruce can be present and contribute as little as 4% pollen to the pollen sum, but above 8% I can be reasonably certain that the pollen is from local trees. Fossil stomates, which are sourced from local trees, are simple to count while conducting pollen analysis and require little extra preparation. I recommend the addition of stomate analysis to plant migration studies to provide the most accurate timing of the earliest arrivals on the landscape or the presence of trees within cryptic refugia.

Despite relatively warm temperatures, spruce lagged climate by up to 500 years prior to the Younger Dryas. Where spruce was present, it expanded in place while surrounded by suitably warm habitat. These data provide a significant example of migration or population expansion lag at the 10s of kilometers scale and highlight the problems associated with using palynological data for inferring climate. Consequently, I can reject the hypothesis H2: “Southwestern Nova Scotia was too cold to support spruce” and confirm that spruce was not in equilibrium with climate throughout Nova Scotia prior to the Younger Dryas.

Summer temperatures during the Younger Dryas were likely too cold to support spruce (although the selection of the start of the Holocene time slice does not illustrate
this). As a result, spruce disappeared from most sites in Nova Scotia during the prolonged cold event. I found some evidence to suggest that isolated stands of spruce may have survived the Younger Dryas and led to rapid recolonization during the early Holocene as originally proposed by Mayle and Cwynar (1995).

I present the first well-dated evidence for the Older Dryas in North America at Thin Ice Pond and Veinot Lake that indicates cooling of ~4 °C, similar to previous records found in Europe (Lotter et al. 1992; Johnsen et al. 1995; Brooks and Birks 2000; Watson et al. 2010; van Asche et al. 2012; Brooks et al. 2012). Similarly, the GI-1c2 left its mark with a 2.6-6 °C cooling. This is the largest signal for the GI-1c2 found to date. The addition of the GI-1c2 and the Older Dryas as well as the Killarney Oscillation (Levesque et al. 1993) to the paleoclimate record in North America dispels the notion that the rapid, shorter duration climate events of the lateglacial were only recognizable in Europe (Lowe and NASP members 1994; 1995). Knowing that millennial scale events typical of European lateglacial records are more widespread than previously observed should also improve our ability to predict the magnitude of future climate events.

The recent slow down in ocean surface warming has led to doubts about global warming as humans in the Northern Hemisphere have faced a series of severe winters and cool summers from a general slowdown in the rate of surface warming. However, the excess heat from global warming is likely being stored in deeper ocean waters (Chen and Tung 2014) suggesting that global warming projections are still valid. The consequence of this heat storage, however, remains unclear. My data at the transition into the Holocene also indicate a slow down and reversal in terrestrial temperatures that is
analogous to the current hiatus. If similar conditions led to the climate reversal in the early Holocene as today, then, potentially, decadal-scale cold periods are an expected result of rapid climate warming. If that is true, global warming may continue at its previously forecasted pace in the near future. Further study into the cause of the reversal, including an examination of fossil proxies from the deeper oceans is necessary to determine if these two events, separated by over 11 000 years, are the result of similar oceanic and atmospheric conditions. Such research is critical to our understanding of future ocean temperatures and their effect on our climate.
References


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