Chironomid-based paleotemperature reconstruction of the North Slave Region, Northwest Territories during the Late Holocene

By

Colin Robertson, B.Sc.

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Colin Robertson

Abstract

I reconstructed late Holocene mean July air temperatures using a chironomid-based inference model based on lake sediment collected from two shallow lakes within the North Slave Region, Northwest Territories. No significant shifts were identified in the chironomid species assemblages, which indicated that these lakes experienced very little environmental change over the last approximately 500 years. Discrepancies were identified between the chironomid-temperature records which may be due to site-specific catchment characteristics obscuring the regional chironomid-temperature signal. The significant Chan Lake temperature reconstruction had similar late Holocene climate events as other regional climate records, such as brief warming events throughout the 1700s and the current warming trend (~ 1850 – present day). A paleofire-climate analysis was also conducted, however, there was no relationship between reconstructed fire events and warmer chironomid inferred mean-July air temperatures. These findings contribute to the characterization of environmental change in Subarctic lakes of northwestern Canada.

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1. Introduction

Northwestern Canada is experiencing an unprecedented warming trend due to climate change (IPCC, 2018; Porter et al., 2013). Instrumental records indicate that the Canadian Arctic and adjacent Subarctic region are warming at between 0.1 – 0.5 °C per decade, and models suggest that this region may undergo accelerated warming in the 21st Century due to positive feedback mechanisms (Government of Canada, 2019; National Center for Atmospheric Research, 2020; Przybylak and Wyszyński, 2020). There is a need for greater paleoclimate records across northwestern North America to better understand how the current, abrupt warming trend fits within the long-term pattern of natural climate variability for the region and to provide a historical baseline for recent warming (Kaufman et al., 2016; Dalton et al., 2018). Paleorecords, including ice cores, lake sediments and tree rings have proven useful in reconstructing past climate events over various time scales across global sites (Bradley, 2014). These natural archives have played a vital role in advancing our knowledge of climatology because they are capable of extending instrumental climate records where long-term climate data may be absent or sparse (Pisaric, 2009).

Our knowledge of the long-term climate variability within the Subarctic Northwest Territories, Canada is incomplete due to short-term observational meteorological data and the poor spatial coverage of long-term climate datasets (Upiter et al., 2014; Porter et al., 2019). Recent studies conducted in the central Northwest Territories provide evidence that this region has experienced similar Holocene (~ 10,000 years ago to the present) climatic events to other polar regions, however, there is variability in the timing and magnitude of these events due to varying proxy-climate sensitivities (Macdonald et al., 2009; Dalton et al., 2018; Upiter et al., 2014; Porter et al., 2019). In addition, there are only a few long-term climate records available across this northern region (Upiter et al., 2014; Dalton et al., 2018). Recent developments in paleoclimatology have highlighted the need to increase the temporal resolution of quantitative climate reconstructions (von Gunten et al., 2012). Dated lake records have emerged as useful

proxy tools because they have a high-temporal resolution and they often contain an array of biological and chemical proxies that can be utilized to infer both long-term climate and environmental change (Fritz, 2008; von Gunten 2012). Chironomid head capsules, whose remains are commonly found in northern lake sediment, are a favourable biological proxy for deciphering past climate due to their high sensitivity to changes in mean July air temperature (Eggermont and Heiri, 2012). Also, chironomid-based paleo-studies have proved effective at reconstructing mean July temperatures at northern sites where longer-term climate records are absent (Brooks and Birks, 2001; Walker, 2001).

To date, a centennial scale temperature record for the Yellowknife region, Northwest Territories has not been reported. The primary objective of my thesis is to apply chironomid assemblage data to reconstruct late Holocene (approximately the last 500 years) mean July air temperatures for the North Slave Region of the Northwest Territories. This study will provide new information on the long-term climate trends for the Yellowknife region, and may provide further insights on the magnitude of the current warming trend in the context of the region's natural climate variability.

Studies have demonstrated that forest fires burn frequently across the Subarctic region of the central Northwest Territories (Sulphur et al., 2016; Pelletier et al., 2020). Research has shown that the primary cause of forest fires within the Northwest Territories are lightning-ignition events associated with summer storms (Kochtubajda et al. 2006). However, severe drought has also been linked to recent burning events in the northwestern boreal forests of Canada, such as the record-breaking 2014 wildfire season within the Great Slave Region, Northwest Territories (Whitman et al., 2018). In terms of long-term fire dynamics, a recent paleolimnological study reported less frequent forest fires within the central Northwest Territories during periods of wetter, colder conditions throughout the Holocene, and suggested that the recent increase in wildfire activity is linked to 20th Century warming (Sulphur et al.,

2016). Furthermore, models predict that further warming will result in longer fire seasons and an increase in area burned within the Northwest Territories (Flannigan et al. 2008).

Although studies have presented findings on the climate-fire relationship within the boreal forests of the Northwest Territories, more research is required to gain a better understanding on the interplay between climate and fire activity within the Great Slave Region (Whitman et al., 2019). The secondary objective of my thesis will be to analyze the chironomid-inferred temperature record in relation to available long-term Yellowknife precipitation data (Pisaric et al., 2009) and paleofire data (Pelletier et al., 2020) to examine how climate influences the fire regime of the North Slave Region. The findings from this project will generate important new information on the historical and contemporary influence of climate on wildfire activity within this Subarctic region.

2. Literature Review

2.1 Chironomid-based paleoclimatology approaches

2.1.1 Proxy records for paleoenvironmental studies

Climate change events occur simultaneously across temporal scales. Gradual changes in climate occur over multi-millennial time periods due to Earth's orbital changes, which mark the onset of glacial periods (Bradley, 2014). Simultaneously, climate variability also occurs on centennial and inter-annual scales due to changes in solar irradiance and atmospheric-ocean teleconnections (Dalton et al., 2018). Anthropogenic activities, such as the burning of fossil fuels for power generation, transportation and agricultural activities, are the leading cause of the current global warming trajectory (IPCC, 2018). In recent decades there has been substantial research on the mechanisms driving the current warming trend in circumpolar regions and the impacts of this abrupt climate change event on northern environmental systems (Macdonald, 2010). Observational data shows that regions in northwestern Canada are

warming at a rapid rate, with Subarctic and High Arctic temperatures increasing at a rate of 0.1 °C and 0.3 – 0.5 °C per decade, respectively (Government of Canada, 2019). Since the year 2000, the rate of temperature increase has accelerated across northwestern Canada, and climate models are projecting future warming trends for these regions in the coming decades (IPCC, 2018). Canada's Arctic and adjacent Subarctic region are well documented as "climate sensitive", however, the natural climate variability of this region is not fully understood due to limited short-term instrumental climate records (~ 50 – 100 years), and a paucity of long-term climate datasets (Pisaric et al., 2009; Porter et al., 2019). A more extensive archive of climate records across spatio-temporal scales is critical for enhancing our understanding of northwestern Canada's climate system and to establish a historical context for the current warming trend and predicted trajectories of anthropogenic climate change (von Gunten et al., 2012; Macumber et al., 2018; Woelders et al., 2018).

In order to generate long-term records of climate variability researchers must measure temperature by proxy. The quantification of Holocene climate trends using proxy sources has been a focus of interest in recent years within northwestern Canada (Tillman et al., 2010; Briner et al., 2016; Kaufman et al., 2016; Macumber et al., 2018). The proxy records most commonly applied to reconstruct past climates in this region include tree rings (annual temporal resolution, century scale temporal range), lake sediment (annual-decadal temporal resolution, century-millennial temporal range) and peat cores (variable temporal resolution, century-millennial temporal range) (Bradley, 2014; Young et al., 2019). The variety of proxies within these environments are able to provide a wealth of information on the paleoclimate of this region across time scales, with some available archives capable of capturing the full extent of the Holocene climate record (Irvine et al., 2012; Porter et al., 2019).

Late-Holocene reconstructions (3500 yr BP – present day) across northwestern Canada have revealed a fairly stable climate with intermittent climatic perturbations, such as the Medieval Climate Anomaly, Little Ice Age and the 19th Century to present day warming trend

(Upiter et al., 2014; Macumber et al., 2018). Although, environmental reconstructions generated from tree rings (Porter et al., 2013), peat (Porter et al., 2019) and lake sediments (Rühland and Smol, 2005; Upiter et al., 2014) have captured the recent warming period across northwestern Canada the magnitude of this warming trend varies across proxy records and is absent in some (Tomkins et al., 2008; Dalton et al. 2018, Macdonald et al., 2009).

Northern studies have demonstrated that proxy data within natural archives of stratigraphic material are capable of providing long-term reconstructions of environmental change across various temporal scales where long-term quantitative climate records are limited (Kaufman et al., 2004; Smol et al., 2012; Briner et al., 2016). In recent years, paleoclimate reconstructions derived from glacial records (Zdanowicz et al., 2014), peat cores (Tillman et al., 2010; Viau and Gajewski, 2009), lake sediments (Bunbury and Gajewski, 2009, 2012; Macumber et al., 2018) and tree rings (Tomkins et al., 2008; Pisaric et al., 2009) have provided valuable insights regarding Holocene climate variability across northwestern Canada, however, the temperature reconstructions do not always display synchronous temperature fluctuations (Briner et al., 2016; Kaufman et al., 2016). These discrepancies highlight the need for more diverse paleoclimate records across Arctic and Subarctic sites in an effort to better understand the local and regional climate dynamics of northern Canada (Upiter et al., 2014). Increasing the sampling of lake sediment across Arctic and Subarctic locations has the potential to improve the spatial-temporal climate coverage of northern regions (Kurek et al., 2009; Upiter et al., 2014; Smol, 2016; Burge et al., 2018).

Lake sediments are excellent natural archives of climate change because they are common across the northern landscape, they often contain climate-sensitive proxies that are capable of reconstructing past climatic conditions and they can be dated (Cohen, 2003; von Gunten et al., 2012). Producing a precise chronology is essential for paleoenvironmental studies because it provides us with a time frame for when environmental changes occurred and it allows us to estimate rates of environmental processes and fluxes (Cohen, 2003). In order to

produce a reliable chronology, lake sediment layers are dated radiometrically. Radiocarbon (^{14}C) and lead-210 (^{210}Pb) dating techniques are suitable for late-Holocene paleoenvironmental reconstructions based on their half-life values (^{14}C = 5730 years, ^{210}Pb = 22.3 years) and their range of time scales (^{14}C = 500 – 40,000 years, ^{210}Pb = 1 – 150 years) (Smol, 2008). Lake sediments have proven useful in producing robust records of environmental change at a range of temporal resolutions, such as near-annual (Hughen et al., 2000; Larocque et al., 2009), multi-decadal (Fritz, 2008; Larocque-Tobler et al., 2015; Woelders et al., 2018) and centennial-millennial (Michelutti et al., 2007; Macdonald et al., 2009; Irvine et al., 2012). In recent surveys, Briner et al. (2016) and Kaufman et al. (2016) identified several lake sediment proxies utilized in northern Holocene climate change studies. These include diatoms for tracking changes in ice-cover (Smol, 1988; Roberts et al., 2015), pollen for reconstructing temperature, shifts in vegetation and forest fire histories (Gajewski et al., 1985; Viau and Gajewski, 2009; Prince et al., 2018), and chironomids for tracking changes in temperature (Eggermont and Heiri, 2012).

2.1.2 Lake sediments and paleolimnological reconstructions

Lakes are complex ecosystems whose development are influenced by a range of environmental processes across spatio-temporal scales. Although lakes vary in size they are all characterized as open systems as they continuously exchange energy and matter with their surrounding environment. A principal input of a lake system is sediment, which can be classified as autochthonous or allochthonous (Smol, 2008). Autochthonous sediment is composed of materials produced from inside the water column, such as the remains of aquatic plants and algae, and animals or precipitates from aquatic biochemical processes. Alternatively, allochthonous sediments are derived from natural or anthropogenic sources outside the lake basin. These include soil and rock sediments, terrestrial plant and animal remains, including pollen, charcoal, and other organic materials, and environmental contaminants (Smol, 2008; Smol et al., 2012; Mwamburi, 2018). Sediment transport within lakes is intermittent and is variable across lake basins due to differences in catchment morphology and catchment

hydrological processes (Gilbert, 1978; Szmytkiewicz and Zalewska 2014; Crann et al., 2015). Over time, autochthonous and allochthonous sediments, gradually accumulate on the lake bed forming a vertical profile of lake sediment, as the deeper, older layers are progressively overlain by younger sediment (Thompson, 1984; Smol, 2008). Calm lacustrine environments with minimal post-depositional disturbances, such as bioturbation and water currents which can dislocate layered sediment, are required for the build-up and preservation of lake sediment (Tylmann et al., 2012).

In the absence of quantitative environmental data, paleolimnological studies have demonstrated that well-preserved lake sediments are capable of reconstructing precise historical environmental conditions for lakes and their surrounding regions (Eggermont and Heiri, 2012; Fortin et al., 2015; Burge et al. 2018). Lake sediments are a common environmental archive used in paleoenvironmental studies because they are a widespread landscape feature that can be cored, collected and analyzed with relative ease, and they contain a suite of proxies which can be used to track a variety of environmental variables of interest across both spatial and temporal scales (Cohen, 2003; Smol, 2008; Smol et al., 2012; Larocque-Tobler et al., 2015). Initially, early paleolimnology studies focused on investigating acid deposition and eutrophication, however, since the 1980s quantitative techniques have been developed for an array of proxy environmental reconstructions (Catalan et al., 2013; Burge et al., 2018). Common biological indicators include the use of diatoms to track changes in lake pH, salinity and nutrient levels, pollen records to reconstruct catchment-scale vegetation changes and charcoal to reconstruct regional fire histories (Smol, 2008; Chipman et al., 2015; Burge et al., 2018). Similarly, lake sediment profiles of pollutants, such as heavy metals, have been examined to investigate the magnitude and timing of natural and anthropogenic disturbances that most likely deposited them, and to better understand their mobility within lake catchments (Smol et al., 2012).

Chironomids found in lake sediments are a popular biological proxy used in paleoenvironmental reconstructions due to their sensitivity to numerous environmental variables. For instance, chironomid assemblages have been applied to generate quantitative reconstructions of lake depth, dissolved oxygen and surface water and air temperatures in numerous northern and southern hemisphere environments (Heiri and Lotter, 2010; Eggermont and Heiri, 2012; Bajolle et al., 2018; Engels et al., 2019). In recent decades, chironomids have emerged as a leading tool for reconstructing regional July air temperatures due to their strong response to temperature (Walker and Pellatt, 2003).

2.1.3 Chironomids as a useful paleoclimatology tool

Chironomidae fossils are widely recognized as an important biological proxy for quantifying and understanding past environmental change (Walker, 1995; Brooks and Birks, 2001; Walker, 2001; Cohen, 2003; Velle et al., 2005; Smol, 2008). The dipteran family, Chironomidae, inhabit a wide range of global freshwater systems, including Arctic and Subarctic lakes (Armitage et al., 1995; Walker, 1995; Cohen, 2003; Smol, 2008). Chironomid remains are commonly found in Holocene paleorecords and their remains are highly abundant relative to other benthic fossils (Walker, 1995; Eggermont and Heiri, 2012). Fossil chironomids are well preserved in lake sediment records due to their robust, chitinous head capsules, which shed during the molting stage of larval development. Chitinous remains are of special interest to paleoclimatologists because a down-core profile of chironomid assemblages can be used to produce reliable environmental reconstructions (Cohen, 2003; Smol, 2008).

In the early 20th Century, researchers produced chironomid-based environmental reconstructions that were qualitative in nature. For instance, lakes were classified as “oligotrophic Tanytarsus lakes” or “eutrophic Chironomous lakes” and chironomid-based temperature reconstructions were often reported as “cooler period” or “warmer period” (Walker, 1995; Smol, 2008; Juggins and Birks, 2012). In the decades that followed,

developments occurred in the field of chironomid paleoecology, including biogeographical investigations to elucidate the role of climatic parameters on the distribution of chironomid communities and the application of chironomids to investigate anthropogenic water quality issues such as eutrophication and acid rain (Cohen, 2003; Smol, 2008; Eggermont and Heiri, 2012). In the 1990s, considerable progress was made in the field of paleoclimatology with the development of the first chironomid-based transfer function capable of reconstructing quantitative temperature reconstructions (Walker et al., 1991; Walker, 1995; Walker et al., 1997). The transfer function approach relies on the assertion that modern chironomid distributions are influenced by particular in-lake or regional environmental variables (e.g. July air temperature, dissolved oxygen) and that fossil chironomid distributions were sensitive to these environmental changes in the past (Telford and Birks, 2009; Smol et al., 2012). Thus, the collection of modern chironomid distributions, derived from surface-sediments across an environmental gradient, is an integral step in the development of a paleoenvironmental transfer function.

In recent decades, a variety of calibration datasets have been generated across various regions for multiple environmental variables of interest (Smol, 2008; Juggins and Birks, 2012). Since the development of the first chironomid-temperature transfer function, chironomids have been widely used in paleoclimate investigations to reconstruct Holocene temperature variability across tropical, temperate, and Subarctic and Arctic regions of the world (Brooks and Birks, 2001; Eggermont and Heiri, 2012; Heiri et al., 2015; Lang et al., 2018). Chironomid assemblages are also frequently used to track in-lake variables such as hypolimnetic oxygen (Little and Smol, 2001), dissolved organic carbon (Larocque et al., 2006), total nitrogen (Brodersen and Anderson, 2002), and lake trophic status (Brodersen and Quinlan, 2006). Recently, several authors have questioned the reliability of chironomid-based Holocene temperature reconstructions, and have re-opened the paleolimnological debate concerning the nature of the chironomid-temperature relationship. Most of the debate concerns the role of temperature versus other environmental variables on determining chironomid populations.

2.1.4 The nature of the chironomid-temperature relationship

Chironomids are a widely used proxy to infer long-term Holocene temperature trends (Brooks and Birks, 2001; Larocque et al., 2006). In 1991, Walker et al. developed the first chironomid-based transfer function, which pioneered the use of chironomid-based numerical models to infer past temperatures and established a quantitative relationship between chironomid populations and temperature (Walker, 1995; Cohen, 2003; Smol, 2008; Eggermont and Heiri, 2012). Since Walker et al.'s (1991) pioneering work, the transfer function technique has become a well-established method in the field of chironomid paleolimnology, and as a result, has allowed researchers to develop numerous robust, quantitative models of environmental change (Velle et al., 2005; Eggermont and Heiri, 2012; Juggins and Birks, 2012). In recent decades numerous chironomid-based transfer functions have been developed to infer both surface water temperatures (Walker et al. 1997) and air temperatures (Larocque et al., 2006; Barley et al., 2006; Porinchu et al., 2009), with air temperature inference models emerging as the most robust (Porinchu et al., 2009; Eggermont and Heiri, 2012). This is because chironomid assemblages often respond to mean July air temperatures more strongly than any other environmental variable, including surface water temperatures (Brooks and Birks, 2001; Eggermont and Heiri, 2012).

July air temperature is most likely the dominant climate variable governing northern chironomid populations because during the summer months lakes typically experience the highest primary productivity (e.g. algae and macrophyte growth), which provide food sources for chironomid larvae (Larocque et al., 2001; Eggermont and Heiri, 2012). Chironomid-based inference models are also increasingly being trained on air temperature rather than surface-water temperature because climate data can provide more accurate estimates of lake site thermal conditions than field measurements of surface water temperature (Porinchu et al., 2009). Furthermore, studies have indicated that models coupled with regional air temperature

data seem to perform better than those coupled with spot measurements of water temperature data (Porinchu et al., 2009; Eggermont and Heiri, 2012).

2.1.5 Evidence of the strong chironomid-temperature relationship across global sites

The correlation between chironomid assemblages and summer air temperature is well-established in paleolimnology literature, however, the extent to which temperature influences chironomid communities has been disputed (Walker, 1995; Cohen, 2003). For example, in the late 20th Century, there was much debate concerning the influence of temperature over other environmental variables (e.g. turbidity, substrate composition and sedimentation rates) in determining Arctic and northern Europe chironomid assemblages (Cohen, 2003; Smol, 2008). Although chironomid-based temperature modelling was met with initial criticism, many investigations have demonstrated the effectiveness of chironomid-based temperature modelling (Smol, 2008; Eggermont and Heiri, 2012; Fortin et al., 2015; Luoto and Ojala, 2017). The small errors bars associated with chironomid-based July air temperature inferences suggests that these are robust records of environmental change (Cohen, 2003; Brooks and Birks, 2001; Brooks et al., 2012). Laboratory studies have also established that temperature plays a role in the functioning and growth of chironomidae during all four stages of development (Péry and Garric, 2006). In addition, the chironomid-temperature signal is global in scope and studies across various regions have concluded that chironomid communities are more strongly correlated to average July air temperatures in relation to other environmental parameters (Luoto et al., 2014 – Subarctic sites, Finland; Nazarova et al. 2011 – Arctic, tundra and boreal sites, Russia; Eggermont et al. 2010 – Tropical sites, Kenyan and Uganda; Matthews-Bird et al. 2015 - Tropical sites within South America). Temperature is an important driver influencing chironomid diversity across northern Canada (Larocque et al., 2006). In 2015, Fortin et al. surveyed chironomid data from numerous northern North America calibration datasets, 435 in lakes total, and concluded that temperature was the most prominent limiting factor explaining the abundance and distribution of chironomid populations across boreal, Subarctic

and Arctic sites. In another major study, Engels (2019) found that July air temperature (T_{Jul}) was more strongly correlated with chironomid diversity in relation to lake depth, pH, conductivity, total organic carbon (TOC) or total phosphorous (TP) for a northern Canadian dataset (Engels, 2019).

Despite the increasing research on chironomid-environmental relationships, only a limited number of studies have assessed the sensitivity of chironomids to July air temperature within Subarctic Canada (Larocque et al., 2006 – northern Quebec; Macdonald et al., 2009 – central Northwest Territories, Upiter et al., 2014 – central Northwest Territories). The paleorecords generated by Macdonald et al. (2009) and Upiter et al. (2014) reveal a common pattern of Holocene climate variability seen in other proxy records, notably the sequence of climate events such as the Medieval Warming Period, Little Ice Age and Twentieth Century warming trend (Macdonald et al., 2009; Upiter et al., 2014). However, the magnitude and timing of these climate events is not wholly synchronous across these investigations. The results from these studies provided valuable insights on the long-term climate variability of the region, but they also indicated that down-core chironomid assemblages are affected by both temperature and non-climatic variables, such as increased lake stratification in response to recent warming (Macdonald et al., 2009) and tree-line migration (Upiter et al., 2014).

2.1.6 The validity of chironomid-based temperature models

In an effort to gain a better understanding of the mechanism driving chironomid distribution, studies have tested the validity of chironomid-based temperature models (Telford, 2019). For example, Lotter et al. (1999) tested the performance of a North American, a European and a combined chironomid-based inference model using both North American and European chironomid fossil assemblages. Although there are differences among the climate records regarding the magnitude and rate of climate change, all three models produced similar Lateglacial climate patterns. Similarly, Lang et al. (2018) produced an accurate 200-year

temperature reconstruction from a eutrophic UK lake using a Norwegian transfer function. These studies underscore the strong chironomid-temperature relationship within these regions and the value of chironomid records as a proxy tool for reconstructing mean July air temperatures. In recent years, near-annual European chironomid temperature inferences have also been tested against meteorological records to assess their reliability. The strong correlations between meteorological records and inferred temperatures suggest that chironomids can reliably reconstruct July air temperature changes in these regions, however, the authors recommend that further studies are required to validate the chironomid-temperature signal in other regions during this time period (Larocque and Hall, 2003; Larocque et al., 2006; Larocque-Tobler et al., 2015).

2.1.7 Discrepancies regarding the chironomid-temperature relationship

Chironomid-based reconstructions rely heavily on the assertion that chironomid assemblages are strongly influenced by temperature, however, this assumption has been challenged by recent research findings (Velle et al., 2010; Brooks et al., 2012; Eggermont and Heiri, 2012; Velle et al., 2012). Producing reliable chironomid-inferred temperature reconstructions for certain time periods within the Holocene epoch is difficult because the temperature variability is often too similar to the prediction error of the transfer function (Brooks, 2006). In addition, it has been reported that during some stages of the Holocene, secondary environmental variables play a greater role in governing chironomid distributions than July-air temperature (Brooks, 2006). Multivariate analysis techniques have revealed that chironomid assemblages are sensitive to a range of environmental gradients, as a result, reliable chironomid reconstructions have been produced for dissolved oxygen (Quinlan and Smol, 2001a; Luoto et al., 2017), total phosphorous (Brooks et al., 2001), lake depth (Korhola et al., 2000; Engels et al., 2012) and other variables of interest (Velle et al., 2010). Laboratory studies have also reported chironomid responses to various environmental gradients. For example, Brodersen et al. (2004, 2008) found that cold-water chironomids respond to changes

in both dissolved oxygen and increasing temperature. In addition, researchers have suggested that it might be challenging to untangle the aggregated effects of temperature, depth and dissolved organic carbon on chironomid-based inferences generated from a high boreal-Arctic region in northwestern Quebec (Larocque et al., 2006). Furthermore, inconsistencies have been found in chironomid-inferred Holocene temperature reconstructions across spatial scales in northern Europe suggesting that other environmental parameters may play a role in governing chironomid distributions (Brooks, 2006; Velle et al., 2010, 2012). In recent years, there has been much discussion concerning the accuracy of chironomid-based Holocene temperature records and the nature of the chironomid-temperature relationship (Cohen, 2003; Eggermont and Heiri, 2012; Fortin et al., 2015; Luoto and Ojala, 2017). For instance, there has been vigorous debate among Velle et al. (2010) and Brooks et al. (2012) concerning the reliability of chironomid-based Holocene temperature records and the degree to which chironomids respond to temperature versus other environmental variables, such as pH, dissolved oxygen and trophic state which have been found to co-vary with temperature in some datasets. Although there is disagreement concerning the reliability of chironomid-temperature reconstructions, they both agree that further research into the chironomid-temperature relationship is necessary to improve the accuracy and utility of chironomid-based temperature models (Velle et al., 2010, 2012; Brooks et al., 2012; Eggermont and Heiri, 2012; Luoto and Ojala, 2017).

2.2 Quantitative Paleoclimatology Techniques

2.2.1 Transfer Function Development

The development of transfer functions is arguably the most important recent advancement in chironomid-based paleoclimatology because they allow researchers to produce quantitative models, based on modern biological-environmental relationships, which when applied to fossil assemblages can produce robust reconstructions of past environmental change (Juggins and Birks, 2012). Since the establishment of the transfer function technique, inference

models have been developed for a range of variables, however, the most common inference models are trained on the chironomid-temperature relationship (Larocque et al., 2009; Porinchu et al., 2009; Eggermont and Heiri, 2012). The transfer function approach is predicated on the principle of uniformitarianism, which states that environmental processes that are operating in the present have also operated in the past (Cohen, 2003). In terms of chironomid paleoclimatology studies, transfer functions rely on modern chironomid-environmental responses to quantitatively infer past environmental change (Birks et al., 2010). The mathematical foundation of a transfer function is presented in Equation 1, where the chironomid assemblage (Y) is some ecological response function $f()$ of the environmental variable (X) of interest (e.g. July air temperature) plus a residual error (Telford and Birks, 2009; Juggins and Birks, 2012). The empirical relationship between modern chironomid assemblages (Y_m) and a contemporary environmental predictor (X_m) is then used to solve $f()$. The output, which is the chironomid abundance response to the modern environmental parameter, is then applied to downcore chironomid assemblages to quantitatively reconstruct an environmental parameter of interest (Telford and Birks, 2009; Juggins and Birks, 2012).

$$Y_m = f(X_m) + \text{error}$$

Y_m = a matrix of modern chironomid abundances
 X_m = a matrix of contemporary environmental predictors
 $f()$ = a set of ecological response functions

Equation 1: An equation used that relate a set of modern biological responses to a modern environmental variable (Juggins and Birks, 2012).

The science of reconstructing robust mean July air temperature values involves three steps, (1) developing of a modern environmental training set, (2) building a numerical model to solve Equation 1 and (3) applying the model to a chironomid fossil record and evaluating the derived reconstruction (Brooks and Birks, 2001; Juggins and Birks, 2012) (Figure 1). Developing a high-quality training set, a record of modern fossil assemblages with associated temperature

measurements, is an important step in generating reliable temperature reconstructions (Telford and Birks, 2009). The quality of temperature inferences is largely dependent on the choices made during the development of the calibration data set (Birks, 1998; Brooks and Birks, 2001; Juggins and Birks, 2012; Fortin et al., 2015). The training set and chironomid fossil sediment cores should be obtained from similar lacustrine environments in an effort to reduce the effects of sediment deposition variability. Training sets typically range from 50-100 study lakes, with a recommended minimum of 30 (Birks et al., 2010, Juggins and Birks, 2012). They contain a suite of contemporary lake chemistry and climate data, and relative abundance data on the distribution of chironomid taxa (50 – 100 species) across the study lakes (Birks et al., 2010). Also, the samples should span the temperature gradient represented by the fossil chironomids in an effort to maximize the chironomid-temperature response (Birks, 1998; Juggins and Birks, 2012; Fortin et al., 2015). Study lakes should be carefully selected when developing a training set for an inference model (Smol, 2008; Patterson et al., 2012). In an effort to maximize the chironomid-temperature signal and minimize the effects of other environmental variables, study lakes that are isolated, undisturbed, and receive minimal inputs are preferential sites compared to other lacustrine environments (Barley et al., 2006).

Figure 1: A systems diagram outlining the steps involved in deriving an environmental reconstruction from biota data using a modern training or calibration data set (based on a diagram in Juggins and Birks, 2012).

The next step involved in developing a paleoenvironmental reconstruction is the selection, application and evaluation of a numerical model. Each numerical technique has its strengths and limitations and selecting a numerical model for a particular study is largely dependent on the nature of the investigation and the numerical properties of the paleolimnological data (Birks et al., 2010; Juggins and Birks, 2012). Although there a wide variety of paleolimnological modelling techniques, this discussion will focus on the evaluation of two common approaches used in paleoenvironmental modelling — weighted averaging partial least squares and modern analogue techniques.

Weighted-averaging is a numerical technique which calculates the average by placing greater weight on some values compared to others. For chironomid-based temperature

reconstructions, this technique is used to identify the temperature optimum for a range of chironomid taxon within a training set (Juggins and Birks, 2012; Fortin et al., 2015; Luoto and Ojala 2017) (Equation 2). This formula relies on the assumption that species occupy different niches in environmental space and that species will be most abundant in sites closest to their environmental optimums (Juggins and Birks, 2012). The formula is:

$$\hat{u}_k = \frac{\sum_{i=1}^n y_{ik} x_i}{\sum_{i=1}^n y_{ik}}$$

Equation 2: A weighted-averaging equation to estimate temperature optimum of a chironomid taxon ($\hat{u}_k$) where the temperature optimum ($\hat{u}_k$) of taxon k , y_{ik} is the relative abundance of taxon k in sample i , x_i is the temperature value at site i , and n is the number of sites (Juggins and Birks, 2012).

Weighted averaging techniques are then utilized to predict historical temperatures based on the relative abundance of chironomid fossils present in lake sediment (Fortin et al., 2015; Luoto and Ojala 2017) (Equation 3). This function relies on the assumption that temperature is a primary limiting factor governing chironomid communities and that abundant chironomid fossils were deposited during optimum climatic conditions (Juggins and Birks, 2012). The equation used to estimate the past environmental temperature ($\hat{x}_0$) is:

$$\hat{x}_0 = \frac{\sum_{k=1}^m y_{ik} \hat{u}_k}{\sum_{k=1}^m y_{ik}}$$

Equation 3: A weighted-averaging equation to estimate past environmental temperatures ($\hat{x}_0$) where $\hat{u}_k$ is the temperature optimum of taxon k , y_{ik} is the abundance of taxon k in sample i , $\hat{x}_0$ is the estimated temperature value, and m is the number of taxon (Juggins and Birks, 2012).

The use of weighted averaging techniques has gained popularity in recent years because they are rooted in sound ecological theory and have strong predictive capabilities (Birks et al. 2010; Juggins and Birks, 2012). Weighted averaging approaches are based on the underlying concept that species have a unimodal response to environmental limiting factors, which are frequently observed in real data. Studies have demonstrated that weighted-averaging regression is a powerful quantitative method for reconstructing environmental variables based on assemblage data, and they are particularly useful for developing transfer functions based on training sets that span a long environmental gradient (Braak and Juggins, 1993). Also, environmental reconstructions based on weighted averaging can still be robust in a no or poor analog scenario assuming the taxon temperature optimum estimates are reliable (Birks et al. 2010). As a result, weighted averaging techniques are useful tools for estimating chironomid taxa optima and for reconstructing paleotemperatures, and they have been proven to perform just as well or outperform other numerical methods (Juggins and Birks, 2012; Fortin et al., 2015; Luoto and Ojala 2017). However, weighted averaging techniques have limitations, such as edge effect errors which generate non-linear distributions at the edges of ecological gradients leading to inaccurate chironomid-inferred temperature reconstructions (Birks et al., 2010). In addition, weighted averaging analyzes environmental variables separately and often does not take into account the residual linear correlations between species which may be a function of secondary environmental variables (Braak and Juggins, 1993).

In 1993, Braak and Juggins presented an enhancement to the weighted-averaging technique called Weighted-Averaging Partial Least Squares (WAPLS) regression. Partial least squares modelling involves the useful extraction of predictor components between the modern biological and environment data, which maximizes the linearity between the biological data (e.g. chironomid data) and the environmental variable of interest (e.g. average July air temperature) within a transfer function (Juggins and Birks, 2012). Selecting the number of components is a key step in the development of a robust WAPLS transfer function. The addition of components improves the fit of the WAPLS generated predicted variable to the modern

environmental data (Birks et al., 2010). However, in the context of chironomid-based air temperature models, it is important to choose an appropriate number of components because the addition of too many components may introduce unwanted “noise” - variability in the chironomid taxa that do not fit the chironomid-temperature model (Juggins and Birks, 2012). Selecting the number of components should be determined using cross-validation techniques, such as bootstrapping (random sampling with replacement) or jack-knifing (leave-one-out cross validation), which assesses the predictive power of the model based on a varying number of components chosen using performance statistics (r^2 , RMSEP and model bias) (Barley et al., 2006; Birks et al., 2010). WAPLS modelling techniques often generate lower predictive error statistics relative to weighted-averaging techniques because they are able to improve the fit of outliers to the chironomid-temperature model, and they are able to reduce or eliminate edge effects which constrain weighted averaging models. In addition, WAPLS techniques may be able to transform the residual linear correlations, generated from secondary environmental gradients acting upon chironomid distributions, to better reconstruct temperature estimates (Juggins and Birks, 2012). Numerous studies have demonstrated improved performance statistics when comparing WAPLS modelling techniques to weighted-averaging methods (Birks et al., 2010; Heiri and Lotter, 2010; Juggins and Birks, 2012).

The modern analogue technique (MAT) follows a simplistic approach to environmental modelling through employing direct space-for-time substitution (Juggins and Birks, 2012). Under the assumption that fossil and modern chironomid assemblages were deposited under similar environmental conditions, researchers reconstruct climate by matching fossil assemblages with analogous modern samples (Birks et al., 2010; Juggins and Birks., 2012). The environmental reconstruction is based on the modern analog that most closely resembles the fossil assemblage, however, if several modern analogs are similar to the fossil chironomid assemblage then the reconstruction is developed using the weighted mean of the closest analogs (Birks et al., 2010; Juggins and Birks., 2012). The modern analogue technique is able to model local-species environmental relationships in large data sets. It is widely used in

paleoceanography but rarely used by paleolimnologists because it requires a highly populated series of samples across the environmental gradients of interest (Juggins and Birks., 2012). Also, the ability to generate a reliable environmental reconstruction may be challenging when analogous samples cannot be found. Studies have also demonstrated that modern analogue techniques are sensitive to spatial autocorrelation errors, which generates erroneous high model performance statistics (Telford and Birks, 2009). Nevertheless, robust environmental reconstructions can be generated using moderately-sized training sets (Olander et al., 1999; Eggermont et al., 2010; Juggins and Birks, 2012).

There are a range of numerical methods within the field of paleoclimatology, each with their unique strengths and weaknesses. Several studies that have investigated the efficacy and performance of these numerical techniques have concluded that there is no optimal method (Telford and Birks, 2009). The most appropriate numerical techniques are those with the ecological and statistical properties that best suit the particular investigation (Telford and Birks, 2009; Juggins and Birks, 2012). Nevertheless, in 2009 Telford and Birks demonstrated that modern analogue techniques are primarily chosen for ocean-based reconstructions, such as modelling sea-surface temperatures using fossil diatom assemblages, because they often outperform other models due to strong spatial autocorrelation (Telford and Birks, 2009). In contrast, techniques within the weighted averaging family usually outperform modern analog techniques for paleolimnology reconstructions (Telford and Birks, 2009). For chironomid-based modelling, studies reveal that transfer functions developed using WAPLS techniques usually outperform transfer functions trained on other modelling procedures and are therefore chosen more frequently (Barley et al., 2006; Porinchu et al., 2009; Fortin et al., 2015; Luoto and Ojala 2017). A summary of numerous chironomid-temperature transfer functions supports this assertion (Eggermont and Heiri, 2012).

2.2.2 Fortin combined training set model

In recent years, there has been growing interest in the application of quantitative modelling techniques to reconstruct past temperatures in areas within Canada where long-term climate data records are absent. A strong chironomid-temperature relationship has been reported across Canadian study sites and, as a result, chironomid-temperature transfer models have been developed across a range of climate gradients (Walker et al., 1997 – southernmost Canada to the Canadian High Arctic, eastern Canada; Barley et al., 2006 – boreal, lichen woodland, alpine tundra and Arctic tundra sites, northwestern North America; Larocque et al., 2006 – boreal, forest-tundra and tundra sites, northern Quebec; Porinchu et al., 2009 – boreal, forest-tundra, low Arctic tundra and mid Arctic tundra sites, central Canadian Arctic). The validity of transfer functions is assessed through the use of cross-validation performance statistics (e.g. r^2 , RMSEP, bias). The application of transfer functions to lake core fossil assemblages has been successful in deriving quantitative estimates of mean July air temperature (Bunbury and Gajewski, 2012; Bajolle et al., 2018). For instance, Upiter et al. (2014) applied both the Barley et al., (2006) and Porinchu et al., (2009) inference models to fossil chironomid assemblages to reconstruct July air temperatures from a lake in the central Northwest Territories. Both transfer functions are based on a WAPLS-2 component approach with robust performance statistics (Barley et al., (2006), $r^2_{\text{boot}} = 0.82$, $\text{RMSEP}_{\text{boot}} = 1.46^\circ\text{C}$; Porinchu et al., (2009), $r^2_{\text{jack}} = 0.77$, $\text{RMSEP}_{\text{jack}} = 1.03^\circ\text{C}$), and as a result, similar temperature reconstructions were produced.

Robust temperature transfer functions have been developed and published for study regions across northwestern Canada (Barley et al., 2006; Larocque et al., 2006; Porinchu et al., 2009). Each transfer function is composed of a training set, a unique group of lakes which best represent the environmental parameter range for chironomid taxa in that particular study region (Brooks, 2006). Although this approach serves to maximize the chironomid-environmental relationship, studies have questioned the validity of this method as it may not

capture the full ecological gradient of interest, and as a result, reconstructions may be based on incomplete chironomid-environmental relations (Fortin et al., 2015). Also, in the context of chironomid-based temperature modelling, it is important to evaluate each transfer function with care because temperature optima can vary between training sets. This was demonstrated when Fortin et al. (2010) generated variable chironomid-inferred Holocene temperature reconstructions in the Canadian High Arctic using different training sets. In addition, transfer functions based on a poorly represented training set have been known to produce less reliable temperature reconstructions (Fortin et al., 2015). Although regional transfer functions have been successfully applied to lake sediments derived from other environments, this is not a recommended approach when modelling small-scale Holocene temperature change (Holmes et al., 2011). There has been increased focus on investigating the performance of transfer functions based on combined training sets, however, most of the research has taken place outside of North America (Holmes et al., 2011).

Fortin et al. (2015) aggregated existing boreal, tundra, Subarctic and Arctic training sets in northern Canada and Alaska to generate an extensive 435-lake training set for northern North America. Although the combined training set contains more environmental gradients, the redundancy analysis suggests that air temperature is the primary variable governing chironomid distributions across northern North America. As a result, a chironomid-based transfer function (WAPLS-2 component approach) for reconstructing mean July air temperatures was produced and applied to existing chironomid records spanning this region (Fortin et al., 2015). The Fortin et al. (2015) temperature reconstructions fit those previously reported, and the extensive training set estimated more reliable temperature threshold and optima values for many of the taxa (Fortin et al., 2015). The results from this study suggest that the combined Fortin et al. (2015) inference model, which contains greater chironomid diversity and a larger number of modern analogues, is capable of producing robust temperature reconstructions when applied to northern fossil chironomid assemblages (Telford, 2016). Recently, errors associated with the geographical coordinates of lakes within the Fortin et al. (2015) training set were rectified by

Medeiros (2020). The improved Fortin et al. (2015) transfer function model has the following robust performance statistics: $r^2_{\text{boot}} = 0.74$, root mean squared error of prediction ($\text{RMSEP}_{\text{boot}} = 1.81$ °C and maximum bias ($\text{max_bias}_{\text{boot}} = 2.16$ °C (Medeiros, 2020).

2.2.3 Taxonomic resolution and head counts

Consistent, accurate identification procedures are crucial for determining the taxonomy of the modern training set and of the fossil chironomid remains. Since chironomid-temperature optimum vary at the genus level, identifying chironomid remains to the highest taxonomic resolution is most often preferred (Brooks et al., 2007). The first chironomid-based inference model was based on midge identification at the tribe, sub-tribe and genus level, however since this pioneering study, many investigations have identified individual taxa at the species and sub-species level, resulting in an expanded taxonomic classification of chironomid larvae (Heiri and Lotter, 2010). Although advancements have been made in chironomid fossil taxonomy and detailed dichotomous keys have been published, it is still very challenging to identify all head capsules in a fossil assemblage at the species and sub-species level (Brooks et al., 2007; Heiri and Lotter, 2010). The primary reason is poor head capsule preservation. For example, fossil remains of the sub-tribe *Tanytarsini* are often missing key components, such as spurs or antennal pedestals, that prevent them from being identified at a higher taxonomic level (Brooks et al., 2007). In addition, poorly preserved chironomid specimens that cannot be identified are labelled as “Unidentifiable” and are subsequently not included in the inference model (Brooks et al., 2007). The chironomid taxa within the modern North American chironomid training sets are identified at the highest taxonomic level but for many this is at the genus level (e.g. *Tanytarsina*-other, *Cricotopus/Orthocladius*, *Chironomus*, *Polypedilum*) (Barley et al., 2006; Fortin et al., 2015).

In 2010, Heiri and Lotter examined the effects of varying taxonomic resolution techniques on the performance of a chironomid-temperature model derived from Central

European data sets. Their results revealed that the transfer function model performance is sensitive to taxonomic resolution. The RMSEP (root mean square error prediction) declined from 1.59 – 1.41 °C, and the r^2 value between chironomid-inferred temperatures and observed July air temperatures increased from 0.78 to 0.90 with the use of a higher taxonomic resolution (Heiri and Lotter, 2010; Telford, 2016). Heiri and Lotter (2010) concluded that the highest achievable taxonomic resolution should be utilized for chironomid-based temperature transfer functions, however, they also recognized the benefits of using an intermediate resolution to reduce the likelihood of misidentification errors (Heiri and Lotter, 2010; Velle et al., 2010). In recent decades, the use of higher taxonomic resolution techniques and the increased use of weighted averaging models has lowered the RMSEP in paleotemperature reconstructions (Eggermont and Heiri, 2012).

Researchers have also demonstrated the impacts of low chironomid head counts on the reliability of quantitative temperature reconstructions. Using simulated counts within existing data sets, Heiri and Lotter (2001) and Quinlan and Smol (2001b) reported that the reliability of inferred temperatures decreases with lower chironomid sample sizes (Heiri and Lotter, 2001). The performance statistics revealed that for temperature reconstructions using 50 chironomid head capsules, the standard deviation of the temperature inferences was about 40% of the RMSEP. It improved to 20% for temperature inferences derived from 100 chironomid head capsules. The variability of inferred temperatures decreased slightly for reconstructions containing 200 chironomid heads, however, the variability and errors increased significantly for head counts of less than 50 (Heiri and Lotter, 2001; Telford, 2016). Telford and Birks (2011) also demonstrated that the probability of Type 2 errors (a “false negative” - incorrectly accepting that there is no relationship between the proxy and the environmental variable of interest) increases with species-poor assemblages or limited fossil records. Therefore, having a large chironomid fossil record is important because the occurrence of a Type 2 error may cause researchers to mistakenly draw conclusions about proxy-environmental relationships. Results from the rigorous testing of chironomid-based temperature inferences have concluded that a

minimum of 50 head capsules per sample is required to generate a reliable temperature reconstruction (Heiri and Lotter, 2001; Brooks et al., 2007; Telford, 2016). It is common practice to omit data based on low head counts well below 50, as these low counts contribute to increased uncertainty in the climate reconstruction (Porinchu et al., 2009; Telford, 2019).

2.2.4 Paleoreconstruction significance test

Chironomid-based inference models have proved effective at reconstructing mean July air temperatures, however, some researchers have critiqued these methodological approaches as a bit of a “black box” with little methods to test the validity of these temperature reconstructions. Testing and evaluating paleoclimate models is a crucial step in the development of a reliable transfer function (Juggins and Birks, 2012). A statistically significant reconstruction should explain more of the fossil assemblage variability than other random environmental variables. Telford and Birks (2011) presented a method for assessing the statistical significance of a transfer function. Published reconstructions generated from a range of proxy-climate transfer functions were utilized for this study. The test involved estimating the proportion of variance in each biota assemblage from a single reconstruction and then comparing it to reconstructions inferred from random environmental variables (Telford and Birks, 2011). If the observed environmental variable explained more variance than 95% of the random reconstructions then it was deemed significant (Telford and Birks, 2011). The analysis revealed that the Lateglacial chironomid-inferred July temperature reconstructions from two Norwegian lakes were deemed significant, while the Bjørnfjelltjørn chironomid reconstruction was not deemed significant. This suggests that environmental factors may influence Holocene chironomid populations more strongly than summer temperatures at the Bjørnfjelltjørn site (Telford and Birks, 2011). Since its inception, the Telford and Birks (2011) validation test has been used to assess the significance of chironomid-based temperature inferences. Although not all of the published reconstructions have been deemed significant, they are still capable of

providing important paleoclimate information (Telford and Birks, 2011; Upiter, 2014). Nevertheless, these reconstructions should be interpreted with caution (Brooks et al., 2012).

2.3 Paleoclimatology and paleoenvironmental reconstructions in northwestern Canada

2.3.1 The climate variability of the central Northwest Territories

Our knowledge of this region's long-term climate variability remains incomplete due to the limited surface coverage and short-term record of the available climate datasets. There is a small network of weather stations across central Northwest Territories and many have only been regularly collecting meteorological data since the early to mid-twentieth century (Environment Climate Change Canada, 2021). For instance, the instrumental climate records for station YELLOWKNIFE A span from 1942 to the present day. Paleoclimatology studies conducted across the central Northwest Territories have provided researchers with greater insight into the climate patterns and events that occurred throughout the Holocene, however, the timing and magnitude of these events do vary throughout the region (Table 1).

Table 1: Paleoclimatology studies conducted across the central Northwest Territories.

Location	Proxy	Time Series of Study (Max years BP)	Publication
Carleton Lake (64° 15' N, 110° 06' W)	chironomids	~ the last 6,500 years	Upiter et al., 2014
Danny's Lake (63° 28' N, 112° 32' W)	diatoms	ca. 3300 yr BP to present-day	Dalton et al., 2018
Lakes across the central Northwest Territories	Lake sediment grains	ca. 8052 yr BP to present-day	Macumber et al., 2018

Slipper Lake and Lake S41 (63° 43' N, 109° 19' W)	chironomids and biogenic-silica content	~ 0 – 2000 Year AD	Macdonald et al., 2009
Slipper Lake (64° 35' N, 110° 50' W)	diatoms	~ the last 5,600 years	Rühland and Smol, 2005
Lakes CL-1 (61° 53' N, 116° 32' W) and HL-1 (62° 30' N, 113° 40' W), The North Slave Region, Subarctic Northwest Territories	chironomids	~ 1470 – 2016 Year CE	This study
Mirror Lake (62° 02' N, 128° 15' W), Southwestern Northwest Territories	tree rings and annually laminated lake sediment	~ 1704 – 1996 Year AD	Tomkins et al., 2008
Several locations throughout boreal Canada, including sites throughout central Northwest Territories	pollen	~ the last 2000 and 12,000 years	Viau and Gakewski, 2009

Most Arctic and Subarctic paleoclimate reconstructions reveal a period of rapid warming during the early Holocene, just following the glacial-interglacial transition (circa 12 000 - 8000 cal yr BP), which marks the initiation of the Holocene Thermal Maximum (HTM) (Dalton et al., 2018). In a review of paleo-Arctic studies Kaufman et al. (2004) suggests a uniform onset of the HTM across northwestern Canada (circa 11 000 – 9000 cal yr BP), however, in more recent

reviews conducted by Briner et al (2016) and Kaufman et al. (2016) the timing of this warming period varies across sites. The results from a multiproxy paleolimnology study conducted in central Northwest Territories reveal a shift in pollen assemblages and an increase in organic sedimentation during the early Holocene, which provide strong evidence for the onset of the HTM at 11 000 cal yr BP (Rühland et al., 2009). Paleoreconstructions from surrounding regions reveal a similar timing for the occurrence of the HTM. For instance, quantitative temperature inferences from Alaska and northwest Canada identify the onset of the HTM from 12 700 – 11 500 cal yr BP, while a temperature reconstruction from the southern Yukon mark the beginning of the HTM around 10,800 cal yr BP (Irvine et al., 2012; Kaufman et al., 2004). A chironomid analysis conducted by Upiter et al. (2014) within the central Northwest Territories recorded the tail end of the HTM in the late-middle Holocene (6 – 4 cal kyr BP). A similar warming trend was identified during this time period from a peat core study conducted in the southern Northwest Territories (Tillman et al., 2010). Following the HTM there is a general cooling trend observed in the paleo climate records across the Arctic, including central Northwest Territories sites, however, the timing and magnitude of these cooling periods are asynchronous across the region (Dalton et al., 2018; Kaufman et al., 2016; Upiter et al., 2014; Bunbury and Gajewski, 2012).

Late Holocene climate events, such as the Medieval Warming Period (MWP), have been identified within climate records derived from Arctic and Subarctic studies (Kaufmann et al., 2016; Clegg et al., 2010). Temperature reconstructions show that the MWP occurred in the central Northwest Territories at around 800-1000 cal yr BP (Dalton et al., 2018; Upiter, 2014; Tillman et al., 2010) with the exception of one climate record that suggests it occurred earlier in between 1350-1100 cal yr BP (Macdonald et al., 2009). The Little Ice Age (LIA) is a common cooling trend following the MWP that has been identified in a wide array of paleorecords across northern latitude regions, yet its age of initiation and length varies from site to site (Cohen, 2003). For instance, studies conducted in the central Northwest Territories have reported cooling trends attributed to the LIA lasting across a range of time-series (e.g. 750 – 200 cal yr

BP (Dalton et al., 2018), 700 – 350 cal yr BP (Upiter et al 2014), 500-100 cal yr BP (Tillman et al., 2008), 650-100 cal yr BP Century (Macdonald et al., 2009)).

The decadal-centennial climate trends over the past few hundred years within northwestern Canada region are variable. Numerous proxy reconstructions across this region show a common pattern of cooling during the end of the LIA to rapid warming from the onset of the industrial era to the present. Temperature reconstructions from a chironomid-based study provide evidence that the central Northwest Territories has experienced a rapid warming trend in the last 200 years (Upiter et al., 2014). In addition, evidence of a current warming period in this region has been marked by an increase in planktonic diatom assemblages beginning in the 19th Century (Rühland and Smol, 2005). Surrounding northern regions have shown similar trends. High resolution tree ring data indicates that the Mackenzie Delta Region is warming at a rapid rate of 0.2° per decade and that the 2000 – 2009 decade has been the warmest on record since AD 1245 (Porter et al., 2013). Also, a recent $\delta^{18}\text{O}$ precipitation analysis study reported that Eastern Beringia (Alaska and Yukon) has experienced accelerated warming since the 19th Century, with an increasing rate of 2°C over the last 100 years (Porter et al., 2019).

Although numerous climate records show a warming trend following the LIA to the present, not all proxies in this region are sensitive to this climate pattern (Tomkins et al., 2008). For example, a chironomid-based temperature record for the central Northwest Territories does not show a trend of industrial-era warming (Macdonald et al., 2009). Also, a high-resolution diatom record extracted within the Subarctic Northwest Territories provided no evidence of a current warming trend (Dalton et al., 2018). Furthermore, some studies have reported brief cooling and warming periods in recent centuries. For instance, chironomid remains from sites near treeline in the Northwest Territories indicate a short warming event in the 1700s (Macdonald et al., 2009), while brief cooling events have been observed throughout

the 1800s in annually resolved tree ring and peat records from this region (Porter et al., 2013; Tomkins et al., 2008; Tillman et al., 2010).

The long-term variability of precipitation within the central region of the Northwest Territories has not been studied extensively. Cook et al. (2004) reconstructed the summer PDSI (Palmer Drought Severity Index) over most of North America (a 2.5° gridded network of 286 sites) using 835 tree-ring chronologies spanning the past 1000 years. The northern extent of the grid, which contains three points within the Great Slave Region (points 62, 77, 93), had very weak PDSI verification statistics due to a limited network of tree-ring records, and as a result, produced less reliable drought reconstructions. Pisaric et al., (2009) reconstructed Yellowknife precipitation data from 1679 – 2003 using twelve jack-pine tree-ring chronologies. June precipitation values were modelled because the tree-ring chronologies correlated more strongly with this climatic variable. The tree-ring precipitation record compared well to the instrumental record and provided a statistically robust model of precipitation variability from 1819 – 2005 (Pisaric et al., 2009). In addition, the inferred precipitation record showed good correspondence to other long-term precipitation-based records, such as the reconstructed summer PDSI derived from grid point 93 (Cook et al., 2004) and the Pacific Decadal Oscillation (PDO) reconstruction. The results from this study suggest that the Yellowknife Region experienced lower than average precipitation from 1698 – 1739, 1776 – 1796, 1801 – 1821, 1842 – 1865, 1880 – 1893 and 1927 – 1979. Distinct periods of positive precipitation occurred during 1687 – 1697, 1756 – 1775, 1822 – 1841, 1890 – 1926 and 1980 – 1995 (Pisaric et al., 2009). Over the last decade, the Great Slave Region of the Northwest Territories have experienced reoccurring drought events (Korosi et al., 2017; Gaboriau et al., 2020). The summer of 2014 was marked with extreme drought conditions, which resulted in most the severe fire season on record (Olesinski and Brett, 2016).

2.3.2 The paleofire regime of the central Northwest Territories

Wildfires are a common and important landscape disturbance in northern boreal forest environments, such as the Great Slave Region which lies within the boreal-tundra transition zone (Kochtubajda et al., 2019; Government of the Northwest Territories, 2008, 2009). Ecologically, wildfire activity plays a significant role in influencing the composition of northern vegetation, in controlling insect and pest populations and in cycling essential elements such as carbon, but wildfires may also have profound negative impacts on surrounding communities (Kochtubajda et al. 2006; Flannigan et al., 2008). The fire regime in any ecosystem is a function of complex interactions between a range of climatic and geophysical variables (Gaboriau et al., 2020). Although wildfire activity is ubiquitous across the boreal landscape, burn rates are particularly high across the western and northwestern boreal forest (Taiga Shield) compared to central and eastern boreal forest ecozones (Bergeron et al. 2004; Kochtubajda et al. 2006). This is in part due to a generally drier climate, which facilitates the ignition and spread of the wildfire, but it is also due to the abundant coniferous vegetation across the landscape, which has a higher flammability than other vegetation types (Campbell and Flannigan 2008).

The wildfire season of the Northwest Territories typically starts in May and ends in early September, and the majority of the wildfires in the region are ignited by cloud-to-ground lightning flashes associated with summer thunderstorm activity (Kochtubajda et al. 2006). The pattern between the number of lightning-ignited fire occurrences and area burned by lightning is unusual because the number of lightning-ignited fires peak in July, while area burned by lightning peaks in June (Kochtubajda et al. 2006; Stocks et al. 2002). Researchers attribute this pattern to monthly differences in climatic conditions. June normally receives less rainfall and experiences stronger winds compared to July, which may generate increased wildfire activity during that month (Kochtubajda et al. 2006). The Government of the Northwest Territories has generated a detailed contemporary record of fire events for the northern Great Slave region dating back to 1965 (Government of the Northwest Territories, 2018). The fire history reveals

that the number of wildfires is highly variable from year to year with increased wildfire activity in the year 1973 and in 1994 (Kochtubajda et al. 2006). It also reveals that the strongest wildfire season on historical record was the summer of 2014, where a record 3.4 million ha of forest burned across the Northwest Territories (Darwent, 2016; Pelletier et al., 2020).

Paleolimnological approaches have demonstrated the ability to reconstruct regional paleofire regimes within northwestern North America (Chipman et al., 2015; Sulphur et al., 2016; Prince et al., 2018). Investigations reveal that wildfires burned frequently across Subarctic Northwest Territories since the onset of the Holocene (Sulphur et al., 2016; Bothwell et al. 2004). Recently, Sulphur et al. (2016) used pollen and microscopic charcoal preserved in lake sediment, located just north of Yellowknife, to report low-resolution changes in vegetation and regional fire dynamics. Their results revealed that the fire regime has responded to past changes in climate over the past few millennia (Sulphur et al., 2016). Following deglaciation (ca. 8610 – 8390 to ca. 7430 – 7220 cal. yr BP), a warm, dry climate punctuated with frequent forest fire activity was reported, however, fewer forest fire events were identified during the Middle - Late Holocene portion of the record which was a time characterized by wetter and cooler conditions (Sulphur et al., 2016). Also, increased wildfire frequency was documented for the 20th Century, with recent warming being the likely driver (Sulphur et al., 2016). In addition, Pelletier et al., (2020), reported that the fire interval for the North Slave Region is roughly 35 – 80 years, and they identified that lakes within this region have been impacted by two to six wildfire events since the early 19th Century (Pelletier et al., 2020).

3. Study site and Methods

3.1 The climate and landscape of the Yellowknife Region, Northwest Territories

The Yellowknife Region lies within the Taiga Plains and Taiga Shield ecozones (Government of the Northwest Territories, 2008) (Figure 2). This Subarctic region is

characterised by a dry continental climate with short, cool summers and long, very cold winters with a mean annual air temperature ranging from -3.5°C to -9°C and an average annual precipitation total in between 230 and 350 mm (Government of the Northwest Territories, 2008, 2009; Galloway et al., 2015). The climate of this region is influenced by a range of factors including latitude, incoming solar radiation, the albedo of the landscape, lake effects, the position and strength of pressure systems, such as the Pacific High and Aleutian Low, and the Pacific Decadal Oscillation (Flannigan et al., 2008; Pisaric et al., 2009; Government of the Northwest Territories, 2008, 2009; Dalton et al., 2018). Our knowledge concerning the climate variability of the Yellowknife Region is constrained due to the limited spatial coverage (a single weather station - “Yellowknife A”, Station ID 2204101) and the short temporal coverage of the instrumental records (1942 – present). The city of Yellowknife has an average annual temperature range of -3 to -6°C and its mean annual precipitation varies from 280 – 360 mm (Government of the Northwest Territories, 2008). The coldest month is typically January with a mean temperature range of -26 to -28°C while July, the warmest month, has a mean monthly temperature of 16.6°C (Government of the Northwest Territories, 2008; Environment and Climate Change Canada, 2019). Also, the average annual temperature for the City of Yellowknife varies annually with an increasing trend since 1942.

The Yellowknife Region is located along the northern extent of Great Slave Lake. Undulating topography, coniferous forest stands, extensive discontinuous permafrost and numerous lakes of varying size characterize the physical landscape of this Subarctic region (Government of the Northwest Territories, 2008, 2009; Flannigan et al., 2008). Although the climate is fairly similar across the North Slave Region, there is a high degree of landscape heterogeneity among the Great Slave Plain and Great Slave Upland High Boreal ecoregions.

Figure 2: Field map of the study lakes Chan Lake (CL-1) and Hidden Lake (HL-1) and study lakes from previous paleoenvironmental studies within the central region of the Northwest Territories, Canada. The pink shaded area of the map indicates the Great Slave Plain High Boreal Ecoregion and the blue shaded area of the map indicates the Great Slave Upland High Boreal Ecoregion.

In July of 2016, a total of ten lake sediment cores and four peat cores were collected as part of a larger project that aimed to examine the impact of local and regional forest fire activity on heavy metal deposition across a wildfire gradient (Pelletier et al., 2020). This research project investigated two out of the ten collected lake cores (Chan Lake (CL-1) and Hidden Lake (HL-1)) that were sampled in the North Slave Region. CL-1 and HL-1 were chosen due to their minimal inflows and outflows, small surface area ($\sim 0.2 - 0.82 \text{ km}^2$) and shallow depth ($< 3 \text{ m}$). The lakes are also characterized with low gradient slopes and riparian vegetation, which can reduce erosional inputs (Meyer et al., 2001). In addition, the study lakes

are also located more than 30 km from Yellowknife anthropogenic sources of pollution where low levels of heavy metals generated from nearby smelting activities have been detected (Houben et al. 2016; Pelletier et al., 2020). These lakes are ideal sites for paleoclimatology studies because their catchment characteristics dampen their sensitivity to non-climate environmental variables (Catalan et al., 2013; Smol, 2016).

3.2 Geographic description of the study lakes

CL-1 (61° 53' N, 116° 32' W) lies within the Great Slave Plain High Boreal Ecoregion of the Taiga Plains and is located roughly 130km southwest of Yellowknife (Figure 2). It has a surface area of ~ 0.84 km² and a maximum depth of ~ 1.5 m. The Great Slave Plain High Boreal (HB) Ecoregion is a part of the Western Interior Platform geologic province, which is primarily composed of sedimentary rock overlain by smooth and gently undulating poorly-drained glacial till deposits (Galloway et al., 2015; Government of the Northwest Territories, 2009). The sedimentary rock formation is characterized by carbonate sandstone, mudstone, dolomite and limestone, which are from the Phanerozoic eon and range in age from the Cambrian-Ordovician to the Devonian Period (Galloway et al., 2015). Underlying the sedimentary rock deposits are the Proterozoic and even older Archean craton basement rocks (Galloway et al., 2015). This landscape exhibits little relief, and as a result, it is dominated by wetlands, bogs and shallow marl ponds (Thienpont et al., 2019; Wolfe and Kokelj., 2019). *Pinus banksiana* (jack-pine) and *Picea glauca* (white spruce) forest stands are abundant forest communities across the region, with proportionally higher white spruce growth across poorly drained sites. Mixedwood forests, featuring *Populus tremuloides* (trembling aspen) are more common in southern areas of this region near Great Slave Lake (Government of the Northwest Territories, 2009). In addition, this region is currently characterized with charred woodland as a result of the recent 2014 wildfire activity (Pelletier et al., 2020).

HL-1 (62° 30' N, 113° 40' W) is located roughly 35 km east of Yellowknife along the Ingraham Trail and it is located within the Great Slave Upland High Boreal Ecoregion of the Taiga Shield (Figure 2). It has a surface area of ~ 0.2 km² and a maximum depth of ~ 2 m. The geology of this region is composed of flat to gently undulating hills of exposed Slave Province bedrock (Archean craton), which is dominated by coarse grained plutonic rock but also contains metamorphic rock and kimberlite pipes (Pell, 1997). Overlying this Precambrian bedrock is a thin veneer of coarse-textured glacial till (Wolfe and Kokelj., 2019; Galloway et al., 2015). Lakes and rivers are dominant landform features across this ecoregion, and peatlands have developed in low-lying areas (Government of the Northwest Territories, 2008). The forest cover is scattered across the region and is characterized as discontinuous due to the presence of rocky outcrops (Government of the Northwest Territories, 2008). Woodlands composed of jack pine, black spruce and paper birch commonly grow in areas with a thin veneer of soil covering bedrock and on well-drained glacial till, while deeper, moist soils allow for the development of black spruce stands with a thick moss understory level. (Pisaric et al., 2009; Government of the Northwest Territories, 2008). Young jack pine and trembling aspen stands dominate post-fire forest communities within this region, and charred open woodlands can be identified throughout the landscape as evidence of recent wildfire activity (Government of the Northwest Territories, 2008; Whitman et al., 2018; Pelletier et al., 2020).

3.3 Sediment core collection

The lake sediment cores for this study were obtained in July of 2016 in the central basin of each lake (CL-1 and HL-1) where it is the deepest and flattest, and where the potential of lake sediment disturbance is low (Cohen, 2003; Smol, 2008). Also, sediment accumulation tends to be the greatest at the lake basin center, which makes it an ideal coring site for long-term paleolimnological studies (Van Metre et al., 2004). The lake sediment was collected from an inflatable raft using an 8.6 cm diameter UWITEC gravity corer at a depth of 1.5 and 1.8 m respectively for lakes CL-1 and HL-1. The sediment cores were then transported to a laboratory

in Yellowknife for sub sampling. The lengths of the CL-1 and HL-1 sediment cores spanned 39 cm and 37 cm respectively. In the lab, the lake sediment from each lake core was mounted and sliced at 0.5 cm intervals, except for the uppermost layer that was sectioned at a depth of 1 cm due to the high water content. The sediment from each segment was transferred into sterile Whirlpak® bags which were kept frozen and shipped to Carleton University for future macro charcoal, loss on ignition (organic matter content), chironomid and ^{210}Pb and ^{14}C radioisotope analysis.

3.4 Sedimentology analysis and age-depth model

The water and organic matter content of each lake sediment segment was estimated using loss-on-ignition (LOI) techniques. Approximately 1 – 2 cm³ of wet sediment (2 cm³ for the top 0 – 10 cm depths which were waterlogged and contained primarily organic matter, 1 cm³ for the lower depths) was transferred into a crucible and placed in a drying oven at 100°C for 24 hours. The samples were then removed and weighed to determine the water content for each lake sediment slice. The organic matter content of each contiguous lake sediment increment was then measured as weight loss following combustion in a muffle oven at 550°C for 4 hours (Heiri et al., 2001). Downcore profiles of LOI and water content were generated for each lake core using the ggplot2 package (v. 3.3.3) (Wickham, 2016) within R Studio(v. 1.3).

The chronologies of the CL-1 and HL-1 sediment cores were generated during the analysis stage of a larger paleolimnology research project (Pelletier et al. 2021) using ^{210}Pb dating techniques for the upper portion of the cores (0 – 30 cm) and ^{14}C dating techniques for the lower portion of the cores (>30 cm). The chronology for the CL-1 core was based on the radio-isotope dates from the duplicate CL-2 core because there was a good match between the CL-1 and CL-2 LOI profiles, which suggests that the lake cores have similar sediment deposition trends and lake histories (Figure 3) (Alberta Environment, 2007). Previous studies have shown similarities in LOI trends from multiple cores collected from the same shallow lake (Shuman,

2003). Also, there may be good agreement among lake cores collected within a small, simple lake with minimal inflows, which is characteristic of the small CL lake basin (Engstrom and Rose, 2013).

Estimates of ^{210}Pb sediment age were generated by Flett Research Ltd (Manitoba, Canada) by measuring ^{210}Po activity (Flynn, 1968) and ^{226}Ra activity (Appleby & Oldfield, 1978) through the use of alpha spectrometry techniques. Twelve and fifteen measurements of ^{210}Pb activity were performed for the HL-1 and CL-2 cores respectively, while ^{226}Ra activity estimates were carried out three times throughout each core. The ^{210}Pb and ^{226}Ra measurements were then converted to estimated ^{210}Pb dates using the constant-rate-of-supply model (Appleby & Oldfield, 1978). Radiocarbon ages obtained on plant macrofossils were generated for each core at the A. E. Lalonde AMS Laboratory. Plant macrofossils were identified and enumerated with the lake sediment using a dissection microscope, a Bogorov counting chamber and forceps. The CL-2 core was radiocarbon dated using spruce needles picked from the 39-39.5cm sediment layer, while the HL-1 core was dated using sedge stems/leaves identified within the 36.5-37 cm sediment layer (Pelletier et al., 2021). The estimated ^{210}Pb and ^{14}C dates were included in the final age depth model for each core, which were generated using the R package Bacon package (v.2.3.9.1) (Blaauw and Christen, 2011) (Pelletier et al., 2021).

Figure 3: A comparison of LOI (percent organic matter) over depth between the CL-1 (circles) and CL-2 (triangles) cores.

3.5 Chironomid sampling, identification and analysis

In each core, chironomids were identified and enumerated using the standard protocols outlined in Walker (2001). Chitinous midge head capsules were analyzed for contiguous 1 cm slices down each core. The volume of subsampling sediment varied for each sediment slice between 2 – 5 cm³, which is higher than the recommended 1 cm³ for chironomid analysis with small, shallow lakes Walker (2001). For the HL-1 core, a range of 0.32 – 1.4 g of dry sediment were subsampled for each segment, while 2.5 – 14 g of wet sediment were subsampled for each CL-1 sediment slice. The subsampled sediment was placed in a beaker containing 5% KOH solution and heated for 30 minutes to deflocculate the sediment. The treated sediment was

then gently sieved through a 100 μm filter using deionized water to remove fine organic material. The remaining sediment was then transferred to a Bogorov counting chamber where it was examined under a Leica EZ4 dissection microscope (35x magnification). Each subsample of sediment was examined four times to ensure all of the chironomid heads were identified and picked. Chironomid heads were picked from the sediment using forceps and placed in a drop of water on a cover slip. A maximum of fifteen chironomid heads were placed on a single cover slip. The cover slips containing chironomid heads were then mounted on microscope slides using Eukitt® quick-hardening mounting medium for future identification.

Chironomid head fossils were identified to the highest taxonomic resolution using a compound microscope (x400 magnification) and a Palaearctic Chironomid identification guide (Brooks et al., 2007). Past studies have outlined that a minimum of 50 chironomid head capsules should be counted and identified per increment of lake sediment (Heiri and Lotter, 2001, Quinlan and Smol, 2001b). This minimum head count guideline was attempted but it was not achieved for all of the subsamples due to low head count recovery and limited sediment availability. A total of 40-50 head capsules has been shown to be sufficient for paleoenvironmental reconstructions, and even head counts as low as 30 are capable of providing reliable information about past environmental change (Porinchu et al., 2009; Quinlan and Smol, 2001). Sediment increments that did not meet the minimum head capsule count threshold were still included in the analysis. Within the HL-1 core, low chironomid head counts were counted for the 20 – 20.5, 24 – 24.5 and 30 – 30.5 cm depths for which 38 , 39 and 33 head capsules were tallied respectively. In addition, <5 head capsules were enumerated within the limited quantity of the HL-1 1 cm sediment interval, as a result, the HL-1 0 cm and 1 cm chironomid data were aggregated. The majority of the CL-1 sediment layers contained >50 chironomid head capsules, however, the 1 – 1.5, 2 – 2.5, 3 – 3.5 and 12 – 12.5 cm depths contained only 33, 34, 44 and 37 head capsules respectively.

The downcore relative abundance data was calculated for each lake sediment profile using the chironomid count data. Also, number of chironomid head capsules per gram of dry sediment was calculated for each lake sediment increment. A stratigraphic profile for each of these measures was generated in R Studio using the ggplot2 software. In addition, a constrained incremental sum of squares (CONISS) analysis, using a broken stick model, was applied to the downcore relative abundance data for each of the lake cores to cluster the chironomid assemblage into significant zones. The rioja package (v. 0.9-26) was utilized to perform this CONISS analysis (Juggins, 2020). Downcore trends in species richness (Hill's N2 Diversity Index) were also estimated for each core using the vegan package (v.2.5-6) (Oksanen et al., 2019) implemented in R Studio.

The vegan package was also utilized to run a principal components analysis (PCA) for each sediment profile to identify the principal components of variance within the species abundance data. First, a Hellinger transformation (square root function on relative abundance data) was applied to the chironomid relative abundance data, and then a detrended correspondence analysis (DCA) was performed to determine the amount of species turnover across and environmental gradient (Hill and Gauch, 1980). Since the results from the DCA summary revealed that the extent of species turnover (length of gradient axes) was less than 2 standard deviations for each core (Table 2), a PCA was performed to investigate the distribution and ordination of the species relative abundance data (Upiter et al., 2014). The PCA plots for each sediment core were generated using the package ggbiplot (v.0.55) (Vu, 2019).

Table 2: DCA axes 1- 4 lengths for the CL-1 and HL-1 species relative abundance data.

Lake Core	DCA Axes			
	1	2	3	4
CL-1	1.19726	0.84776	0.88147	0.95140
HL-1	1.24136	1.37724	0.87488	1.14318

3.6 Charcoal analysis and fire events

The charcoal accumulation rate (CHAR) was calculated for both cores using the macroscopic charcoal count data from Pelletier et al. (2020) (Equation 4). The macroscopic charcoal analysis was conducted by Nicolas Pelletier and his research team using the methods outlined in Pelletier et al. (2020). In summary, 1 cm³ of sediment was subsampled from each contiguous slice of lake sediment. It was then immersed in a weak aqueous solution containing a deflocculant and 11% sodium hypochlorite to bleach the non-charcoal organic material. The bleached sediment was then sieved using a 100 µm filter and the remaining contents were poured into a petri dish and analyzed under a dissection microscope where pieces of macroscopic charcoal were identified and enumerated. Sediment slices with charcoal measurements that exceeded a threshold value were used to identify charcoal peaks and paleo fire events. In addition, historical fire maps for the study region were surveyed to identify contemporary fire events Pelletier (2020).

$$\text{Charcoal accumulation rate (\# of pieces of macroscopic charcoal cm}^{-2} \text{ yr}^{-1}) = \frac{\# \text{ of pieces of charcoal}}{\text{AgeTop} - \text{AgeBot} * d}$$

Equation 4: Charcoal accumulation rate (# of pieces of macroscopic charcoal cm⁻² yr⁻¹) equation, where *# of pieces of charcoal* equal the amount of charcoal fragments found in the sediment layer (# cm⁻³), *AgeTop* and *AgeBot* equal the median modelled ages at the bottom and top of the layer and *d* is the depth of the layer (modified equation from Pelletier et al., 2021).

3.7 Temperature reconstruction development

First, the modern analogue technique was employed to determine how well the fossil chironomid assemblages in my sediment cores were represented in the Fortin et al. (2015) modern training set. In order to do this, a Hellinger transformation was applied to the chironomid relative abundance data for each core. Each chironomid dataset was then compared to the Fortin modern assemblage data by conducting a squared chord distance analysis which was applied using the analogue package v.0.17-4 (Simpson & Oksanen, 2020). The squared chord distances (SCD) were calculated between each fossil assemblage and its 10 closest modern analogues. Fossil assemblages with SCD values exceeding 5% of the SCD values among the training set were identified as having poor modern analogues (Fortin et al., 2015).

The Fortin et al. (2015) inference model (Medeiros, 2020) was applied to the transformed relative abundance chironomid data for each lake to generate July air temperature estimates over the past few centuries. Fossil chironomid taxa that were absent from the training set were not included in the respective temperature reconstructions. The Fortin et al. (2015) WA-PLS transfer function was built and cross-validated (leave-one-out techniques) using the rioja package (v. 0.9-26) (Juggins, 2020). The two-component WA-PLS model was selected because of its significance value and its strong r^2_{boot} , $RMSEP_{boot}$ and max_bias_{boot} performance statistics. July air temperature reconstructions were then generated for each core along with their sample-specific errors. The significance of each temperature reconstruction was tested using the method presented by Telford and Birks (2011) within the R package paleoSig (Telford, 2011). A chironomid-based environmental reconstruction is deemed significant if it can explain more variance in the chironomid assemblage than 95% of the randomized environmental variables (Telford and Birks, 2011). Each WA-PLS (2-component) July temperature reconstruction was smoothed by applying a Generalized Additive Model (GAM) using the mgcv package (v. 1.8-34) (Wood, 2011). Upper and lower 95% confidence intervals were also generated for each GAM model. The model summary and gam.check diagnostics were then

analysed to ensure a suitable number of basis functions were chosen and to check that the GAM model fit to the data. The GAM models and confidence interval data were fitted and plotted to the July temperature reconstructions for each core using the brms package (v. 2.14.4) (Bürkner, 2018) available in R Studio.

3.8 Comparing the chironomid-inferred temperature record to other regional climate records

A contemporary record of Yellowknife average July air temperatures spanning 1942 – 2016 (Station ID YELLOWKNIFE A, 2204101) was obtained from the Adjusted and Homogenized Canadian Climate Data website (Environment Climate Change Canada, 2021) at <https://www.canada.ca/en/environment-climate-change/services/climate-change/science-research-data/climate-trends-variability/adjusted-homogenized-canadian-data.html>. An analysis was then conducted by comparing the trends of each temperature reconstruction, spanning from 1942 – 2016, with the instrumental temperature record. Strong correlations between near-annual chironomid-based temperature inferences and meteorological data have been reported in previous studies conducted in Sweden (Larocque and Hall, 2003), Switzerland (Larocque et al., 2009) and Poland (Larocque-Tobler et al., 2015). Due to a lack of near-annually resolved sediment, we were not able to perform these quantitative analyses. Instead, a qualitative analysis was performed between the temperature estimates and the 5-year moving average Yellowknife instrumental temperature record.

The CL-1 and HL-1 temperature reconstructions were then placed in a regional context by comparing them to other paleoenvironmental records from nearby lakes (e.g. Carleton Lake, Danny's Lake and Lake S41) within the central region of the Northwest Territories (Figure 2). The chironomid-based temperature data from Carleton Lake (Upiter et al., 2014) were derived using the Barley et al. (2006) and Porinchu et al. (2009) transfer functions and were accessed through the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center at

<https://www.ncei.noaa.gov/pub/data/paleo/insecta/chironomidae/northamerica/canada/nwt/carleton2014cit.txt>. The chironomid-inferred July air temperature data from Lake S41 (Macdonald et al., 2009) was also accessed through NOAA's National Climatic Data Center at <ftp://ftp.ncdc.noaa.gov/pub/data/paleo/paleolimnology/northamerica/canada/nwt/s41-2008.txt>. The end-member sediment datasets from Danny's Lake and Carleton Lake were extracted from the Macumber et al. (2018) study using DigitizeIt v. 2.5.3.

In addition, the CL-1 and HL-1 temperature reconstructions, charcoal accumulation and fire event data (Pelletier et al., 2020) and available Yellowknife precipitation data were analyzed in an effort to investigate the relationship between climate and wildfire activity within the Yellowknife Region over the past few hundred years. A Pearson's *r* correlation test was performed using R studio in order to assess the significance of the relationship between each temperature reconstruction and the charcoal accumulation data. The reconstructed Yellowknife June precipitation record (1680-2003 CE) published by Pisaric et al., 2009 was obtained for this analysis. To date, this is the only available long-term precipitation record for the Yellowknife Region. A record of June precipitation data from 2004 – 2016, obtained from the YELLOWKNIFE A station instrumental record at https://climate.weather.gc.ca/index_e.html, was also included in the analysis.

4. Results

4.1 Core chronology and sediment stratigraphy

The age-depth model for the CL-2 core (Figure 4) was built using fifteen ^{210}Pb activity measurements and a single AMS radiocarbon date using a spruce needle found at the 39 – 39.5 cm interval (Table 3). The predicted age of this bottom interval was 1104 +/- 74 cal year BP, while the uppermost sediment was dated to 2016 (Year CE). The final chronology for the full length of the CL-2 core (39 cm) was predicted to span from 1391 – 2016 CE, with the 37 cm

depth dated to 1471 CE. Similarly, the age-depth model for the HL-1 core (Figure 5) was produced using twelve fifteen ^{210}Pb activity measurements and a single AMS radiocarbon date using sedge stems and leaves enumerated from the 36.5 – 37 cm interval (Table 3). The bottom of the HL-1 core was radiocarbon dated to 785 +/- 23 cal year BP, while the uppermost sediment was dated to 2016 (Year CE). The final chronology for this lake profile was predicted to span from 1387 – 2016 CE, with the 35 cm depth dated to the year 1472 CE.

Figure 4: The Bayesian age-depth model for CL-2 Lake. The top left panel shows that the Markov Chain Monte Carlo runs were fairly stable (>5000 iterations), the middle plot shows the prior (curve) and posterior (filled distribution curve) distribution for the accumulation rate (year/cm), and the rightmost plots show the prior and posterior for the unique accumulation rates or “memory”. The primary plot shows the Bacon age-depth model for CL-2 Lake with the light blue regions representing the ^{210}Pb dates and the purple regions represent ^{14}C dates. 0 cal yr BP represents the year 1950 CE (Pelletier et al., 2021).

Figure 5: The Bayesian age-depth model for HL-1 Lake. The top left panel shows that the Markov Chain Monte Carlo runs were fairly stable (>3000 iterations), the middle plot shows the prior (curve) and posterior (filled distribution curve) distribution for the accumulation rate (year/cm), and the rightmost plots show the prior and posterior for the unique accumulation rates or “memory”. The primary plot shows the Bacon age-depth model for HL-1 Lake with the light blue regions representing the ^{210}Pb dates and the purple regions represent ^{14}C dates. 0 cal yr BP represents the year 1950 CE (Pelletier et al., 2021).

Table 3: AMS radiocarbon (^{14}C) ages for CL-2 and HL-1 Lake.

Lake	Core length (cm)	Interval Depth (cm)	Material	Lab ID	^{14}C BP	$\pm$	F^{14}C	$\pm$
CL-2	39	39-39.5	Spruce needle	UOC-5037	1104	74	0.8716	0.008
HL-1	37	36.5-37	Sedge stems/leaves	UOC-5038	785	23	0.9069	0.0026

The loss-on-ignition (LOI) results revealed that the CL-1 and HL-1 cores were primarily comprised of organic matter, with most intervals consisting of >50% organic matter (Figures 6 and 7). The top sediment intervals in both the CL-1 and HL-1 cores contained the highest amount of organic matter. Specifically, the top 5 cm within the CL-1 and HL-1 cores contained 9 and 3 intervals, respectively, that exceeded 60% organic matter. The loss-on-ignition (LOI) results are more variable within the CL-1 core than the HL-1 core. The CL-1 organic content ranged from 37 – 69% and is the highest within the uppermost sediment, with a maximum value of 69% at the 2.5 cm interval (2012 CE). Comparatively, the HL-1 organic matter content ranged from 56 – 63% within the uppermost sediment to the 26 cm depth (1907 CE) and then decreased to ~50% where it remains fairly steady throughout the basal portion of the core. The peak organic matter value of 83% was at the 23 – 23.5 cm interval (1945 CE) may be an anomaly as it does not fit the overall LOI trend.

The water content range was similar between the two lacustrine cores (CL-1 = 82 – 98%, HL-1 = 86 – 99%). Within the CL-1 profile, water content was highest in the upper sediment layers, with a maximum value of 99% at the 10 cm depth. Following the 10 – 10.5 cm interval, water content values declined to ~ 88% and remained fairly stable. The HL-1 core had a similar trend in water content, with the higher values in the upper sediment intervals and lower values in the bottom layers of the core. Between the 18 – 29 cm interval, water content was linearly interpolated due to analysis errors (Figure 7).

Based on the lacustrine sedimentology profile, sediment accumulated more slowly within CL-1 (average sedimentation rate of $0.002 \text{ g cm}^{-2} \text{ y}^{-1}$) relative to HL-1 (average sedimentation rate of $0.05 \text{ g cm}^{-2} \text{ y}^{-1}$). Higher sedimentation rates occurred within the uppermost layers of the CL-1 core, with the highest sediment rate ($0.008 \text{ g cm}^{-2} \text{ y}^{-1}$) occurring at the 14.5 cm interval (1963.5 CE). Following this peak, sedimentation rates gradually declined with core depth. Similarly, within the HL-1 core, sedimentation rates were highest within the

top intervals of sediment and then gradually declined. A maximum sedimentation rate of $0.2 \text{ g cm}^{-2} \text{ yr}^{-1}$ occurred within the HL-1 record at a depth of 6.5 cm (~ 2008 CE).

The age-depth models revealed that the lake sediment cores had a similar temporal resolution. The average temporal resolution of the CL-1 core was 14.5 yr cm^{-1} while the average temporal resolution of the HL-1 core was 15.3 yr cm^{-1} . Maximum rates of 3.1 yr cm^{-1} (CL-1 core) and 1.2 yr cm^{-1} (HL-1 core) occurred in the uppermost sediment layers, while lower temporal resolution values (CL-1 = 42 yr cm^{-1} , HL-1 = 59 yr cm^{-1}) occurred in the basal portion of each core due to sediment compaction.

Figure 6: Organic matter data derived from Loss-On-Ignition (LOI), sedimentation rate and water content data for the CL-1 core.

Figure 7: Organic matter data derived from Loss-On-Ignition (LOI), sedimentation rate and water content data for the HL-1 core. Water content was linearly interpolated between the 18 – 29 cm interval.

4.2 Chironomid diversity and ordination

From the CL-1 core, a total of 26 chironomid taxa were identified. The most common taxa identified were *Psectrocladius sordidellus*-type, *Procladius*, *Tanytarsini* spp. and *Tanytarsus* spp. (Table 4). *Dicretodipes nervosus*-type, *Chironomus plumosus*-type, *Polypedilum* spp. and *Indeterminable* chironomid heads were also common. The most evenly distributed chironomid taxa, with a high effective occurrence across the sediment intervals (Hill's N2 value), were

Tanytarsini spp. (34), *Procladius* (32) and *Psectrocladius sordidellus*-type (31). Of the 26 taxa identified, 24 were included in the Fortin et al. (2015) inference model, and 8 were identified as rare (Hill's N2 value < 5).

In total, 31 chironomid taxa were identified within the HL-1 sediment profile. The most common taxa were *Dicrotendipes nervosus*-type, *Microtendipes pedellus*-type, *Tanytarsini* spp., *Polypedilum* spp., *Procladius* and *Tanytarsus* spp. (Table 5). Other common taxa include: *Cricotopus/Orthocladius*, *Chironomus plumosus*-type, *Glyptotenpides barbipes*-type, *Psectrocladius sordidellus*-type and *Indeterminable* chironomid heads. The chironomid taxa that were both highly abundant and most evenly distributed throughout the lake core were *Microtendipes pedellus*-type (Hill's N2 value = 31), *Tanytarsini* spp., *Polypedilum* spp., *Dicrotendipes nervosus*-type (Hill's N2 value = 30). Seven of the 31 chironomid taxa were poorly represented (Hill' N2 value < 5) in the Fortin et al. (2015) inference model, and 2 were excluded from the chironomid-temperature model.

Table 4: Chironomid species identified, occurrence of chironomid species and the Hill's diversity index data for the CL-1 lake core (RA = Relative Abundance).

Taxa identified within the Sediment	Occurrences within the Sediment	Hill's N2	Min RA	Max RA	Mean RA	Included in the Fortin Model
<i>Chironomus anthracinus</i> -type	19	13	0	9	3	Yes
<i>Chironomus plumosus</i> -type	36	27	0	14	7	Yes
<i>Cladopelma lateralis</i> -type	33	25	0	14	7	Yes
<i>Cricotopus/Orthocladius</i>	27	20	0	7	3	Yes
<i>Cryptochironomus</i>	13	12	0	3	2	Yes
<i>Dicrotendipes nervosus</i> -type	37	28	0	16	7	Yes
<i>Dicrotendipes notasus</i> -type	1	1	0	1	1	Yes
<i>Einfeldia dissidens</i> -type	3	3	0	2	2	Yes
<i>Endochironomus</i> spp.	2	2	0	2	2	Yes
<i>Glyptotenpides barbipes</i> -type	15	13	0	4	3	Yes

<i>Glyptotenpides pallens</i> -type	1	1	0	2	2	Yes
<i>Microtendipes pedellus</i> -type	30	17	0	25	8	Yes
<i>Pagastiella</i>	12	9	0	6	3	Yes
<i>Parakiefferiella bathophila</i> -type	4	4	0	2	2	Yes
<i>Polypedilum</i> spp.	36	26	0	18	8	Yes
<i>Psectrocladius sordidellus</i> -type	38	31	2	27	15	Yes
<i>Psectrocladius barbatipes</i> -type	31	24	0	9	5	Yes
<i>Psectrocladius septentrionalis</i> -type	2	2	0	2	2	Yes
<i>Tanypodinae</i> spp.	25	18	0	12	5	No
<i>Macropelopia</i>	10	9	0	4	3	No
<i>Procladius</i>	38	32	2	19	10	Yes
<i>Tanytarsini</i> spp.	38	34	6	41	21	Yes
<i>Tanytarsus</i> spp.	38	29	2	19	10	Yes
<i>Cladotanytarsus mancus</i> -type	20	11	0	13	5	Yes
<i>Lauterborniella</i> <i>Zavreliella</i> spp.	1	1	0	1	1	Yes
<i>Zalutschia</i> type-B	2	2	0	2	2	Yes
<i>Indeterminable</i>	37	31	0	11	6	No

Table 5: Chironomid species identified, occurrence of chironomid species and the Hill's diversity index data for the HL-1 lake core (RA = Relative Abundance).

Taxa identified within the Sediment	Occurrences within the Sediment	Hill's N2	Min RA	Max RA	Mean RA	Included in the Fortin Model
<i>Chironomus anthracinus</i> -type	17	15	0	5	1	Yes
<i>Chironomus plumosus</i> -type	29	22	0	13	5	Yes
<i>Cladopelma lateralis</i> -type	28	20	0	8	2	Yes
<i>Cricotopus/Orthocladius</i>	30	26	0	8	4	Yes
<i>Cryptochironomus</i>	7	6	0	3	0	Yes
<i>Cryptotendipes</i>	4	3	0	4	0	Yes
<i>Dicortendipes nervosus</i> -type	35	30	3	18	9	Yes
<i>Dicortendipes notasus</i> -type	8	5	0	9	1	Yes
<i>Einfeldia dissidens</i> -type	18	13	0	9	2	Yes
<i>Endochironomus</i> spp.	10	8	0	4	1	Yes

<i>Glyptotenpides barbipes</i> -type	29	23	0	8	3	Yes
<i>Glyptotenpides pallens</i> -type	6	5	0	4	0	Yes
<i>Glyptotenpides severini</i> -type	11	10	0	3	1	Yes
<i>Microchironomus</i>	1	1	0	2	0	Yes
<i>Microtendipes pedellus</i> -type	35	31	2	28	15	Yes
<i>Pagastiella</i>	5	5	0	2	0	Yes
<i>Parachironomus</i>	1	1	0	2	0	Yes
<i>Parakiefferiella bathophila</i> -type	7	7	0	3	0	Yes
<i>Polypedilum</i> spp.	35	30	7	31	17	Yes
<i>Psectrocladius sordidellus</i> -type	32	23	0	11	3	Yes
<i>Psectrocladius barbatipes</i> -type	11	10	0	2	0	Yes
<i>Psectrocladius septentrionalis</i> -type	2	2	0	3	0	Yes
<i>Tanypodinae</i> spp.	28	22	0	9	3	No
<i>Macropelopia</i>	9	7	0	5	1	No
<i>Procladius</i>	34	26	0	8	3	Yes
<i>Tanytarsini</i> spp.	35	30	5	31	16	Yes
<i>Tanytarsus</i> spp.	34	24	0	16	6	Yes
<i>Cladotanytarsus mancus</i> -type	6	5	0	3	0	Yes
<i>Corynocera ambigua</i>	2	2	0	2	0	Yes
<i>Corynocera oliveri</i> -type	4	4	0	3	0	Yes
<i>Laut/Zavreliella</i>	3	3	0	3	0	Yes
<i>Indeterminable</i>	30	25	0	12	5	No

There was little variability in the composition of chironomid communities over time for each lacustrine environment, as a result, the CONISS analysis yielded no statistically significant stratigraphic zones (Figure 8, Figure 9). However, there is variability in relative abundance of dominant taxa over time within both the CL-1 and HL-1 cores.

An average of 60 heads were enumerated from each sediment layer within the CL-1 core (Figure 10). A lower number of chironomid heads were found at the top of the core (12 – 0 cm, 49 heads) in relation to the lower layers of the core (37.5 – 12 cm, 66 heads). The interval from 37.5 – 36 cm (1471 – 1525 CE) had high relative abundance values for *Microtendipes pedellus*-type and *Tanytarsini* spp. with peak values of 18% and 16% respectively (Figure 8). There was very little change in the chironomid assemblages from 36 – 18 cm (1525 – 1949 CE).

During this interval the chironomid community was primarily comprised of *Tanytarsini* spp., which had an average relative abundance of 18% and attained a peak value of 31%. *Psectrocladius sordidellus*-type was the second most dominant taxa during this time period attaining an average relative abundance of 16% and reaching a maximum value of 27%. The relative abundance of *Microtendipes pedellus*-type declined during this interval with values ranging from 0 – 7%. Between 18 - 10 cm (1949 – 1983 CE), *Tanytarsini* spp. increased in relative abundance with an average value of 28%. The presence of *Psectrocladius sordidellus*-type slightly decreased but it still frequently attained relative abundance values >10%. There was a slight increase in the relative abundance of *Dicrotendipes nervosus*-type from 5% (37.5 – 18cm) to 9% and a slight decrease in relative abundance from 7% (37.5 – 18cm) to 4% for *Chironomus plumosus*-type and *Cladopelma lateralis*-type. The uppermost sediment (10 – 0 cm, 1983 – 2016 CE) intervals were primarily composed of *Tanytarsini* spp. which had an average relative abundance of 21% and peaks of 30%. In addition, the relative abundances for *Microtendipes pedellus*-type and *Polypedilum* spp. increased and had noticeable peaks of 25% and 18%, respectively, within the 9 – 6 cm interval. *Cladotanytarsus mancus*-type was also more prevalent in the upper sediment layers of the CL-1 core.

An average of 55 chironomid head capsules were identified within the HL-1 lake sediment samples (Figure 11). Head capsule concentrations were relatively consistent throughout the upper layers (12 – 0.75 cm) of the sediment core with a range of 46 – 60 heads found per sample. The variability in the number of head capsules found increased in the lower layers of the core (35.5 – 12 cm) where a range of 33 – 70 head capsules were picked from each sample.

The dominant taxa throughout the HL-1 core were *Microtendipes pedellus*-type, *Tanytarsini* spp., *Polypedilum* spp. and *Dicrotendipes nervosus*-type (Figure 9). Within the bottom layers of the sediment (35.5 – 27 cm, 1472 – 1884 CE) these taxa had average relative abundance values of 20%, 18%, 12% and 10% respectively. The sediment interval from 27 – 7

cm (1884 – 2007 CE) was characterized by a continued dominance of these four taxa with a slight increase in the relative abundance of *Polypedilum* spp. from 12% – 17%, and a slight decrease in the presence of *Microtendipes pedellus*-type (20% to 14%). This interval was also characterized by intermittent peaks in relative abundance (>10%) for *Tanytarsus* spp. The uppermost sediment (7 – 0.75 cm, 2007-2016 CE) is marked by an increase in *Polypedilum* spp. to a maximum relative abundance of 31%. *Tanytarsini* spp., *Microtendipes pedellus*-type, *Dicrotendipes nervosus*-type were still abundant in this interval but their average relative abundance values declined to 16%, 10%, and 7% respectively. This layer was also characterized by a marked increase in *Chironomus plumosus*-type which attained a peak relative abundance value of 12%.

Figure 8: Chironomid relative abundance ($\geq 2\%$) and CONISS cluster data for the CL-1 core. Chironomid taxa are ordered left to right in alphabetical order based on the Fortin et al. (2015) training set. Quantitative chironomid-based average July air temperature (AJAT) reconstructions based on the CL-1 chironomid assemblage data. Significant shifts in the chironomid assemblage data were assessed by applying a cluster analysis model (CONISS, broken stick model) but no significant zones were found.

Figure 9: Chironomid relative abundance (chironomid taxa that have an abundance of $\geq 2\%$ in at least one sample were included in the analysis) and CONISS cluster data for the HL-1 core. Chironomid taxa are ordered left to right in alphabetical order based on the Fortin et al. (2015) training set. Quantitative chironomid-based average July air temperature (AJAT) reconstructions based on the HL-1 chironomid assemblage data. Significant shifts in the chironomid assemblage data were assessed by applying a cluster analysis model (CONISS, broken stick model) but no significant zones were found.

Figure 10: Chironomid concentration (number of chironomid head capsules per 1 gram of dry sediment) and chironomid count data for the CL-1 core. Chironomid taxa are ordered left to right in alphabetical order based on the Fortin et al. (2015) training set.

Figure 11: Chironomid concentration (number of chironomid head capsules per 1 gram of dry sediment) and chironomid count data for the HL-1 core. Chironomid taxa are ordered left to right in alphabetical order based on the Fortin et al. (2015) training set.

The Principal Component Analysis (PCA) revealed that axes 1 and 2 explained 38.3% of the variance in the chironomid abundance data for the CL-1 core (Figure 12). Axis 1 explained 24.8% of the variability in the chironomid assemblage data, and there was a strong, positive association between PC axis 1 and *Microtendipes pedellus-type*. *Polypedilum* spp., *Cladotanytarsus mancus-type* and *Cricotopus/Orthocladus* were also positively correlated to axis 1, while *Chironomus plumosus-type*, *Tanypodinae* spp., *Psectrocladius sordidellus-type*, *Cladopelma lateralis-type* and *Psectrocladius barbatipes-type* negatively associated with it. Principal component 2 explained 13.5% of the variance with *Polypedilum* spp., *Psectrocladius*

barbatipes-type, *Cladopelma lateralis*-type, *Psectrocladius sordidellus*-type and *Chironomus plumosus*-type positively related it. The chironomid taxa that were negatively related to PCA axis 2 were *Tanypodinae* spp., *Tanytarsini* spp., *Dicrotendipes nervosus*-type and *Cladotanytarsus mancus*-type.

Principal components 1 and 2 explained 26.4% of the chironomid assemblage variance within the HL-1 core (Figure 13). The first component explained 13.4%, and the chironomid taxa positively associated with it included *Tanypodinae* spp., *Chironomus anthracinus*, *Cricotopus/Orthocladius* and *Tanytarsini* spp. Alternatively, *Chironomus plumosus* was strongly, negatively related to the first principal component. Other taxa negatively correlated to PCA axis 1 included *Polypedilum* spp., *Dicrotendipes notatus*-type and *Glyptotendipes barbitipes*-type. The second component explained 13% of the variance in the chironomid abundance data, with *Glyptotendipes barbitipes*-type, *Tanypodinae* spp., *Chironomus anthracinus* and *Parakiefferiella bathophila*-type being scored positively. The chironomid taxa that had negative scores along PCA axis 2 included *Tanytarsini* spp., *Microtendipes pedellus*-type and *Dicrotendipes nervosus*-type.

Figure 12: Principal components analysis (PC1 and PC2) of for the relative abundance chironomid data within core CL-1. Each arrow indicates a chironomid species while the numbers correspond to the top sediment depth of each slice (cm).

The modern analogue analysis revealed that there were good modern analogues for a majority of the fossil chironomid samples in both the CL-1 and HL-1 cores (see Figures A1 and A2 in the Appendix). Since no fossil chironomid samples within the CL-1 or HL-1 dataset had a squared chord distance (measure dissimilarity) exceeding the 95% threshold interval (poor modern analogues) the Fortin et al. (2015) transfer function was deemed suitable and was applied to reconstruct the temperatures for each study site.

The chironomid-based temperature reconstructions for the CL-1 and HL-1 cores extended from the late Holocene to the present (CL-1 ~ 1471 – 2016 CE , HL-1 ~ 1472 – 2016 CE) (Figure 14). The temporal resolution for both cores was variable over time. Due to compaction, the bottom sediment layers present a lower resolution time series than the uppermost sediment in both cores. The chironomid-based temperatures varied from 10.8 °C at 1572 CE to a maximum of 14.7 °C in late 20th Century (1990 CE) for lake CL-1 and yielded a mean temperature of 12.4 °C. The sample-specific errors for the CL-1 temperature reconstructions ranged from 1.83 – 1.85 °C. The chironomid temperature inferences for the HL-1 site were slightly warmer with a range of 13.3 °C (1472 CE) to 16.2 °C (1797 CE) and a mean temperature of 14.6 °C. The margin of error for the HL-1 temperature inferences ranged from 1.83 – 1.87 °C.

The trajectories of temperature change were determined by plotting a Generalized Additive Model (GAM) against the chironomid-inferred temperature profiles for each lake system (Figure 14). The temperature trends vary over time and between the lacustrine environments. The CL-1 core estimated warmer temperatures from 1471 – 1525 CE while the HL-1 dataset inferred the coldest temperatures during this time period. In general, both temperature reconstructions inferred an increase in temperature from the late 16th Century to the late 18th Century, which is when the highest HL-1 temperature inference occurred. The temperature reconstructions followed different trajectories from the 19th Century to the present day. From ~ 1800 CE to 2016, lake CL-1 experienced a steady increase in temperature, while the HL-1 temperature record suggests the occurrence of a decreasing temperature trend. The recent temperature reconstructions for both lake systems were characterized by a high degree of temperature variability. For instance, the temperature estimates ranged from 11.9 – 14.7 °C in between 1989 – 2006 CE for the CL-1 core. Similarly, the HL-1 temperature record had a large range of temperature inferences from 13.5 – 15.9 °C in between 2007 – 2016 CE.

Figure 14: Quantitative chironomid-based July air temperature reconstructions for both the CL-1 (a) and HL-1 (b) cores. The black line represents the GAM model fit to the temperature inferences and the purple shaded area represents the 95% pointwise confidence limits of the GAM model. The shaded grey area indicates the sample-specific error for each temperature estimate (black circles).

The CL-1 temperature reconstruction was deemed significant (p value = 0.034) because it explained more variance in the chironomid fossil record than 999 other randomly generated variables (Figure 15a). Alternatively, the HL-1 temperature reconstruction failed the Telford and Birks (2011) significance test (p value = 0.78) suggesting that the temperature trend does not explain a majority of the variance in the chironomid data (Figure 15b).

Figure 15: Results from the Telford and Birks (2011) significance test for the CL-1 (a) HL-1 (b) chironomid-inferred temperature reconstruction (based on the Fortin et al. (2015) transfer function model). The red-dashed line represents the significance threshold and the black solid line represents the amount of variance explained by the temperature reconstruction. Environmental reconstructions are deemed significant if they explain more variance in the fossil record than 95% of the 999 randomly generated variables. The dashed black line represents amount of variance explained by the PC1 axis.

4.4 Comparison between the temperature reconstructions and the instrumental data

The validity of the contemporary CL-1 and HL-1 temperature reconstructions (1945 – 2016 CE) were also assessed by comparing them to the instrumental Yellowknife, Northwest Territories July air temperature series (1942 – 2016) (Figure 16). The CL-1 temperature reconstruction is comprised of 20 temperature measurements for this time period while the HL-

1 temperature profile is comprised of 23 temperature measurements. Both of the CL-1 and HL-1 trained transfer functions generated chironomid-inferred July air temperature estimates (CL-1 = 12.6 °C, HL-1 = 14.7 °C) that were below the average Yellowknife July air temperature value based on instrumental data (16.6 °C). Despite the difference in absolute temperature measurements, there is good agreement between the chironomid-based CL-1 temperature measurements and the smoothed instrumental record (5-year moving average, $k = 5$) (Figure 16a). For instance, there are similar trends in the interannual-variability of both records from ~ 1942 to 1970, such as the increasing temperature trend during the 1950s and the marked decrease in temperature during the 1960s. In addition, there are similar patterns of temperature change from the 1990s to the present day, such as an increasing trend during the mid 1990s and a decline in temperatures between 2000 and 2010. There was also good agreement between the overall temperature trend and the magnitude of warming between the instrumental record and the CL-1 temperature reconstruction.

The HL-1 core chironomid-based temperature record and the measured July air temperatures shared a similar increasing temperature trend between from ~ 1942 – 1962, but in recent decades, the chironomid-inferred record did not present similar patterns of temperature change as the instrumental record (Figure 16b).

Figure 16: Mean July air temperatures measured (instrumental data) (blue line) for Yellowknife, Northwest Territories from 1942 – 2016 versus the CL-1 (a) and HL-1 (b) chironomid-inferred

temperatures for the Yellowknife Region, Northwest Territories from 1945 – 2016 CE (dashed, black line). The red line represents a moving average ($k = 5$) of the instrumental data.

4.5 Comparison between the temperature reconstructions and other proxy data

A correlation analysis between reconstructed temperatures and charcoal accumulation for the CL-1 watershed determined there was no significant relationship between these two variables (p -value = 0.27). This result was somewhat expected as most inferred peaks in temperature do not correspond with peaks in charcoal accumulation (Figure 17). It is important to note though that one of the warmest temperatures on record (2009 CE, 13.6° C) is preceded by a peak in charcoal accumulation (2010 CE, 600 (# of charcoal $\text{cm}^{-2} \text{yr}^{-1}$) and a known fire event (2010). There is little evidence to suggest that fire events were strongly influenced by June precipitation, as there is variability in the reconstructed (grey dashed line) and instrumental (solid black line) June precipitation values during years with fire activity. Interestingly, four of the six years with known fire events (1924 - 2016) experienced higher than normal precipitation levels while the other two years experienced near mean June precipitation levels (Figure 17).

Figure 17: The lower panel shows quantitative chironomid July air temperature inferences (black lines, white circle) and charcoal accumulation rate data (solid blue bars) (Pelletier et al. 2020) obtained from the CL-1 core (1471 – 2016 CE). The upper panel presents tree-ring based June precipitation reconstructions (z-scores) from 1680 – 2003 CE (grey dashed line) and a 25-year moving average for the precipitation reconstruction time series (Pisaric et al., 2009). The solid line in the upper panel represents average June precipitation values (z-score values) from 2004 – 2016 derived from Yellowknife Station, Northwest Territories (instrumental record). The red dashed vertical lines represent known fire events that occurred within the lake catchment (Pelletier et al., 2020). The black dashed vertical line marks the end of the reconstructed June precipitation record and the beginning of the instrumental record.

Similarly to the CL-1 dataset, there was not a significant correlation ($p\text{-value} = 0.06$) between HL-1 reconstructed July air temperatures and charcoal accumulation rates. The variability throughout the charcoal record does not appear to vary in step with changes in temperature, however, the highest reconstructed temperature within the 20th Century (2007, 15.98° C) is associated with the largest annual charcoal accumulation value (128 # of charcoal $\text{cm}^{-2} \text{yr}^{-1}$) (Figure 18). Also, there does not appear to be a clear pattern between precipitation levels and fire incidence, however, the fires events of 1904, 1996 and 2014 occurred following a year or during a year with lower than average June precipitation values (Figure 18).

Figure 18: The lower panel shows quantitative chironomid July air temperature inferences (black lines, white circle) and charcoal accumulation rate data (solid blue bars) (Pelletier et al. 2020) obtained from the HL-1 core (1472 – 2016 CE). The upper panel presents tree-ring based June precipitation reconstructions (z-scores) from 1680 – 2003 CE (grey dashed line) and a 25-year moving average for the precipitation reconstruction time series (Pisaric et al., 2009). The solid line in the upper panel represents average June precipitation values (z-score values) from 2004 – 2016 derived from Yellowknife Station, Northwest Territories (instrumental record). The red dashed lines in both the lower and upper panels represent known fire events that occurred within the lake catchment (Pelletier et al., 2020). The black dashed vertical line marks the end of the reconstructed June precipitation record and the beginning of the instrumental record.

Discussion

The following analysis begins with an interpretation of the CL-1 and HL-1 sedimentology data and chironomid assemblage data. Within the chironomid assemblage analysis, I outline the potential reasons for the differences between the CL-1 and HL-1 midge communities and we compare the chironomid assemblage data presented in this study to other regional records. In the proceeding few sub chapters, I examine the CL-1 and HL-1 chironomid-temperature trends and compare them to other proxy-based climate records in the central region of the Northwest Territories. In the penultimate subchapter, I discuss plausible reasons for the CL-1 and HL-1 paleotemperature discrepancies, such as differences in catchment characteristics and processes and limitations within the transfer function. The final aspect of the discussion is an analysis of the relationships between the North Slave Region paleofire data and the proxy-based climate inferences.

5.1 Sedimentology analysis

The CL-1 and HL-1 sediment stratigraphy analyses demonstrated that both lakes had a high organic matter content (LOI) and a high water content (Figures 6 and 7). Paleolimnology studies from the central region of the Northwest Territories also reported high organic matter (Macumber et al., 2018; Crann et al., 2015) in the uppermost sediment of their lake cores. In addition, it has been reported that Finnish lakes, within a similar boreal environment to this study, contain a high amount of organic matter and sediment water content (Rantala et al., 2015). The sedimentation rates for both lakes (CL-1 = $0.002 \text{ g cm}^{-2} \text{ y}^{-1}$, HL-1 = $0.05 \text{ g cm}^{-2} \text{ y}^{-1}$) fall within the range of sedimentation rates of other shallow lakes (Alexie and Grace lakes) in the Yellowknife Region (Sivarajah et al., 2020). Hidden Lake has a faster sedimentation rate in relation to Chan Lake, which suggests that this boreal lake system is more productive (Crann et al., 2015). Similar fast sedimentation rates were also measured in shallow lakes within the Mackenzie Delta (Deison et al. 2012). The upper-mid portions of both cores experienced very

different trends in sedimentation, which suggests that unique lake catchment characteristics and processes, such as precipitation, riparian buffer condition, trophic state and within-lake disturbances, played an important role governing the input and deposition of sediment within each lake environment (Sivarajah et al., 2020; Fiskal et al., 2019; Pilgrim et al., 2015). The deposition times of CL-1 (14.5 yr cm^{-1}) and HL-1 (15.3 yr cm^{-1}) are in line with deposition times estimated across temperate lacustrine sites within southern Canada and northeastern USA (21 yr cm^{-1}) (Goring et al., 2012). They are also similar to the rapid deposition times ($< 50 \text{ yr cm}^{-1}$) of other highly productive boreal lakes within the central Northwest Territories (Crann et al., 2015). In addition, the rapid deposition rates in the uppermost sediment may be explained by the high water content within these layers which resulted in less sediment compaction (Crann et al., 2015).

5.2 Chironomid ecological interpretation

No major shifts occurred in the CL-1 or HL-1 chironomid assemblages over time (Figures 8 and 9), which suggests that these lake catchments did not undergo substantial environmental changes over the last few hundred years (Engels et al., 2019; Walker and Macdonald, 1995). Similar stable conditions were identified during this time period within chironomid lake records in Alaska (Clegg et al., 2010), southwest Yukon (Bunbury and Gajewski, 2012) and the central region of the Northwest Territories (Upiter et al., 2014, Macdonald et al., 2009). In addition, other studies have reported little variability in chironomid assemblages over time within shallow, Subarctic lakes due to a lack of lake stratification (Engels and Cwynar, 2011).

Both sediment cores were dominated by *Tanytarsini* spp. taxa (Figures 8 - 11). *Tanytarsini* are cold water adapted and are commonly found across Subarctic environments (Gandouin et al. 2006), however, it is important to note that the *Tanytarsini* tribe is comprised of 28 different species, each with unique ecological preferences (Brooks et al. 2007; Larocque et al., 2006). For instance, *Tanytarsus* spp. are typically found in warm, productive lake

environments while *Corynocera oliveri*-type are mostly associated with cold, oligotrophic lakes (Brooks et al. 2007). We were unable to identify the *Tanytarsini* spp. specimens at the species taxonomic level because the specimens were poorly preserved and missing key components, such as premandibles and antenna, that are integral in the identification process (Brooks et al., 2007). Similar paleoclimatology studies also classified poorly preserved *Tanytarsina* head capsules as *Tanytarsini* other/*indeterminable* (Upiter et al., 2014; Bunbury and Gajewski, 2012; Saulnier-Talbot and Pienitz, 2010). *Psectrocladius sordidellus*-type and *Microtendipes* also had high relative abundance values throughout the CL-1 record (Figure 8). *Psectrocladius sordidellus*-type are commonly associated with warm, shallow waters and are abundant across boreal and tundra lakes (Upiter et al., 2014; Bunbury and Gajewski, 2012; Engels and Cwynar, 2011; Larocque et al., 2006; Walker and MacDonald 1995). The taxa *Microtendipes* and *Polypedilum* also dominated the HL-1 record suggesting that this aquatic environment has stayed relatively warm and shallow over the past few hundred years (Engels and Cwynar, 2011; Macdonald et al., 2009; Brooks et al., 2007) (Figure 9).

Although no significant shifts in chironomid assemblages were identified throughout CL-1 over time, *Microtendipes* is a dominant species that experienced substantial changes in relative abundance. The increased dominance of this taxa within the bottom layers and uppermost sediment of the CL-1 record may indicate that changes in lake depth occurred since *Microtendipes* are often found in the sublittoral – littoral zones of lakes (Brooks et al., 2007). Moreover, Millet et al. (2007) have identified *Microtendipes* as sensitive to low lake levels, and Engels and Cwynar (2011) demonstrated that *Microtendipes* have a statistically significant relationship to lake water depth, with an ecological preference to shallow waters. The increased presence of *Microtendipes* might be a result of the recent lake expansion (~ 51% increase in surface area) that CL-1 and other nearby lakes experienced between 1986 - 2011 within the Mackenzie Bison Sanctuary, Northwest Territories (Korosi et al., 2017). The timing of the increase in *Microtendipes* (~ 1989 CE) within the CL-1 record coincides with the lake expansion in this region. Korosi et al. (2017) demonstrated that this recent lake expansion

activity was outside the natural variability of lake level change in this region. Also, their analyses showed that these catchment changes are significantly correlated to July-October positive Pacific/North American teleconnection patterns (e.g. warmer and wetter conditions) and summer sea-surface temperature anomalies, which suggests that global climate change is a contributing factor to the recent lake expansion in this region (Korosi et al., 2017). Although climate is linked to lake surface area change, it is likely not the sole driver since recent warming temperatures and drought conditions have not resulted in substantial lake level change in the Mackenzie Bison Sanctuary (Korosi et al., 2017).

Other paleoclimatology studies conducted in the central Northwest Territories reported predominantly cold-water chironomid taxa throughout their lake records. For instance, during the late-Holocene, the chironomid record from Lake S41 (Macdonald et al. 2009) and from Carleton Lake (Upiter et al., 2014) (Figure 2) was dominated by *Corynocera ambigua*-type which is often found in cold, oligotrophic lakes north of the treeline (Walker and MacDonald et al., 1995; Brooks et al., 2007). *Sergentia*, which is commonly found in cold, low-nutrient Subarctic lakes, was also a common species throughout the Carleton Lake core. The genera *Tanytarsus* spp. also had a high relative abundance in both the Lake S41 and Carleton Lake chironomid during the last few hundred years. This genus is composed of 10 different morphotypes that are commonly found in the littoral of warm, productive, boreal lakes, except for *Tanytarsus lugens*-type which is a cold stenotherm that inhabits the profundal zone of oligotrophic lakes and the littoral of cold, Subarctic lakes (Self et al., 2015; Brooks et al., 2007).

The CL-1 principal component analyses revealed that axis 1 and 2 explained 24.8% and 13.5% of the chironomid assemblage variation (Figure 12). The chironomid species related with PCA axis 1 were *Microtendipes*, which had the strongest (positive) association, *Polypedilum*, *Cladotanytarsus* and *Chironomus plumosus*-type. These species are characterized as warm species that are common in boreal lakes south of the treeline (Upiter et al., 2014; Macdonald et al., 2009; Walker and Macdonald, 1995). It is important to note that *Microtendipes* and

Polypedilum are commonly found in the littoral, shallow zone of lakes (Gouw-Bouman et al. 2019; Engels and Cwynar, 2011). Also, *Chironomus plumosus*-type, *Polypedilum* and *Cladotanytarsus* are associated with productive (eutrophic) lake environments (Bajolle et al. 2019; Lemmen and Lacourse, 2018; Brooks et al., 2007).

For HL-1, axis 1 and axis 2 explained 13.4% and 13% of the chironomid assemblage variation over time (Figure 13). The chironomid taxa that was most strongly associated with PCA axis 1 was *Chironomus plumosus*-type. *Polypedilum*, *Dicrotendipes*, *Glyptotendipes*, *Chironomus anthracinus*-type and *Cricotopus/Orthocladius* were also related with PCA axis 1. Similar to the CL-1 taxa, these chironomid species are associated with relatively warm conditions across northern lakes (Upiter et al., 2014; Bunbury and Gajewski, 2012; Brooks et al., 2007; Axford et al., 2009; Macdonald et al., 2009; Walker and Macdonald, 1995). Also, they are commonly found in the sediment of littoral shallow waters across mesotrophic – eutrophic lakes (Gouw-Bouman et al. 2019; Araneda et al., 2013; Engels and Cwynar, 2011; Brooks et al., 2007).

Tanypodinae spp. also had positive scores along axis 1. It was difficult to identify the specific taxa within this subfamily because many of the head capsules lacked key components (e.g. ligula) or had components that were too difficult to classify (e.g. unclear ventral setation). A paleoclimatology study conducted in Baffin Island, Arctic Canada identified *Tanypodinae* spp. as a thermophilous taxa (Axford et al., 2009), however, the subfamily Tanypodinae are comprised of over 60 species which have a range of temperature and environmental sensitivities (Brooks et al., 2007).

5.3 Chironomid community differences between Chan and Hidden Lake

Although the study lakes were dominated by both *Tanytarsini* spp. and thermophilous, littoral taxa, species diversity differed between the two lakes (Figure 8 and Figure 9). In

particular, HL-1 had a greater number of chironomid species with a relative abundance of $\geq 2\%$. Whilst temperature is the main driver behind chironomid distributions, other environmental factors have been attributed to chironomid spatial variability, such as sediment morphology. For instance, species within the subfamily *Chironominae* are often found in soft, silty substrate, while *Orthoclatiinae* taxa are preferential to rock and gravel beds (Brooks et al., 2007). The geological and sedimentological characteristics of the Taiga Plains and Taiga plains may have affected the chironomid assemblages within each lake, however, our data does not strongly support this assertion since a majority of the dominant species are not associated with a particular substrate type (Brooks et al., 2007). Also, *Microtendipes*, which is commonly found in sandy-silty sediments, was found in both the Taiga Plains (composed of sedimentary rock, overlain by smooth glacial till) and Taiga Shield (composed of undulating exposed bedrock, overlain by coarse-textured glacial till) despite their geological differences (Galloway et al., 2015; Brooks et al., 2007).

Chironomid larvae consume a range of food types, which include bacteria, algae, aquatic plant matter and other invertebrates (Brooks et al., 2007; Callisto et al., 2007). Macrophyte type has also been identified as an important factor governing chironomid distribution and abundance across lake environments. Previous research has demonstrated that macrophyte availability and diversity influence chironomid communities, and have established positive relationships between macrophyte and chironomid species (Tóth et al., 2012; Brooks et al., 2007). It is plausible that variations in macrophyte populations influenced the unique chironomid communities within each lake system since *Microtendipes* and *Polypedilum*, which are associated with macrophytes, had higher relative abundances in HL-1 than CL-1 (Bajolle et al., 2018). Furthermore, HL-1 also contained higher concentrations of *Glyptotendipes* taxa which are also associated with the presence of aquatic plant populations (Brooks et al., 2007).

Chironomid species occupy lakes of different trophic status and over the years numerous taxa have been categorized as oligotrophic, mesotrophic or eutrophic based on their

environmental sensitivities (Brooks et al., 2007). As a result of these trophic status sensitivities, studies have demonstrated that chironomids are capable of indicating changes in lake nutrient levels, such as total phosphorous, over time (Brodersen and Quinlan, 2006; Brooks et al., 2007). Differences in lake trophic state may explain the variability in chironomid assemblages between CL-1 and HL-1. It is possible that HL-1 is in a more eutrophic condition relative to CL-1. This would explain why it contained more littoral taxa adapted to higher phosphorous levels, such as *Glyptotendipes* and *Dicrotendipes*, and a greater dominance of *Polypedilum*, which is associated with eutrophic lakes (Brodersen and Quinlan, 2006; Brooks et al., 2007). However, *Procladius*, which is commonly found in eutrophic environments, and often the last species to survive anoxic conditions, had a higher relative abundance throughout the CL-1 record (Brooks et al., 2007). In addition, contemporary total phosphorous levels in HL-1 (0.01 mg/L) and CL-1 (0.03 mg/L) are below eutrophic standard levels (Table A3) (Brodersen and Quinlan, 2006; Liu et al., 2019).

5.4 Quantitative chironomid temperature reconstructions

The CL-1 and HL-1 chironomid assemblages had excellent modern analogues with the contemporary fossil record within the Fortin et al. (2015) training set (Figure A1 and A2), indicating that the CL-1 and HL-1 chironomid data are similar to modern chironomid communities (Simpson, 2007). This was surprising as lake systems within the central Northwest Territories were not well represented in this extensive dataset (Fortin et al., 2015). Despite the strong modern analogue metrics for both chironomid records, only the CL-1 temperature reconstruction was deemed significant (Figure 15) (Telford and Birks, 2011). This suggests that the CL-1 chironomid-based temperature record is robust and that July air temperature is the primary variable governing chironomid assemblages in this lake (Brooks and Birks, 2001; Eggermont and Heiri, 2012). This result was particularly interesting as there were no significant shifts in chironomid communities within our record (Figure 8) or other regional chironomid records during this time period (Upiter et al., 2014; Macdonald et al., 2009). Alternatively, the

HL-1 temperature record did not pass the significance test, indicating that secondary variables may be confounding the July air temperature-chironomid response in this catchment. Although, the HL-1 temperature reconstruction was not deemed significant, it was still included in the following regional paleoclimate analysis as chironomid-based studies have reported important climatic information derived from non-significant reconstructions (Lemmen and Lacourse, 2018; Upiter et al., 2014).

The July air temperature estimates were consistently warmer throughout the HL-1 lake record than to the CL-1 lake record (Figure 14). This is most likely due to the dominance of colder boreal taxa *Tanytarsini* spp. (temperature optimum of 9.83 °C) and *Psectrocladius sordidellus*-type (temperature optimum of 10.44 °C) within the CL-1 chironomid record (Brooks et al., 2007, Fortin et al., 2015; Bajolle et al., 2019). Along with the significance testing, the validity of the contemporary July air temperature estimates (1945 – 2016 CE) were also assessed by comparing them to the contemporary Yellowknife temperature instrumental record (1942 – 2016) (Figure 16). Previous studies conducted in Sweden (Larocque and Hall, 2003), Switzerland (Larocque et al., 2009) and Poland (Larocque-Tobler et al., 2015) have demonstrated the effectiveness of this transfer function validation technique. The CL-1 temperature profile matched more closely to the instrumental record than the HL-1 temperature model, which was not surprising since the CL-1 temperature reconstruction was statistically significant (Telford and Birks, 2011). Although there were similar patterns in July air temperature variability between the CL-1 record and instrumental data, the Fortin et al. (2015) transfer function underestimated the mean July air temperatures by ~ 1.9 °C for the Yellowknife Region. Similarly, underestimated mean August air temperature values were recorded in a recent boreal study (Bajolle et al., 2018). The researchers concluded that this was due to the use of the Fortin et al. (2015) dataset, which has low temperature optimums for each taxa due to the many Arctic and High Arctic training set lakes (Bajolle et al., 2018; Fortin et al. 2015). Lower temperature reconstructions, compared to the original temperature inferences, were also generated in the assessment of the Fortin et al. (2015) model for lakes within the treeline

transition zone due to the influence of Arctic lakes in the expanded training set (Fortin et al. 2015). Based on previous studies, the discrepancy between the measured and modelled Yellowknife temperatures may be due to the cold temperature optima in the Fortin et al. (2015) model (4.4 °C – 13.05 °C), which are well below the mean July air temperature for Yellowknife (16.6 °C) (Environment Climate Change Canada, 2021). The addition of boreal lakes within the central Northwest Territories to the Fortin et al. (2015) training set may increase the accuracy of contemporary and late Holocene temperature estimates for this region.

The contemporary CL-1 (12.6 °C) and HL-1 (14.7 °C) July air temperature estimates are closer to the Yellowknife mean August air temperature value (14.3 °C) than the mean July air temperature value (16.6 °C) (Environment Climate Change Canada, 2021). This interesting result suggests that other thermal parameters, such as August air temperature, may be affecting the chironomid assemblages in this region (Eggermont and Heiri, 2012). Although there are few studies available, it has been demonstrated that mean August air temperature is a primary variable governing chironomid assemblages in lake environments (Larocque et al., 2006, Larocque-Tobler et al., 2015; Bajolle et al., 2018).

5.5 Regional comparison of proxy-based temperature records

The mean July air temperatures estimates for CL-1 (12.4 °C) and HL-1 (14.6 °C) from ~ 1470 – 2016 CE are warmer than other regional chironomid-based temperatures estimates (Figure 19). For example, Upiter et al. (2014) reported a mean July air temperature of 11.71 °C (based on the Barley et al. (2006) transfer function) and 11.58 °C (based on the Porinchu et al. (2009) transfer function) at Carleton Lake (Figure 2) during this time period. A cooler midge-inferred temperature estimate (10.44 °C) was generated from Lake S41 (Macdonald et al. (2009)). The warmer temperature estimates from CL-1 and HL-1 Lake are not surprising since the other paleotemperature estimates were generated from more northerly lakes within the forest-tundra zone of central Northwest Territories. A comparison between the

paleotemperature reconstructions from this study and the other regional studies revealed both similarities and differences in July air temperature trends across decadal to centennial time scales. There are limited paleotemperature estimates within the CL-1 and HL-1 temperature records preceding the 19th Century, therefore, the paleoenvironmental interpretations during this time interval should be examined with caution.

The HL-1 record suggests that there was a cooling period in between 1472 – 1617 CE within the North Slave Region, which may coincide with the tail end of the Little Ice Age (LIA) (Figure 19b). A decline in chironomid-inferred air temperatures associated with the LIA was also reported from Lake S41 (1300 – 1850 CE, Macdonald et al., 2009) and Carleton Lake (1310 – 1650 CE, Upiter et al., 2014) (Figures 19cdh). In addition, marked changes in lake sedimentation reflect cooler conditions during this time period (Tomkins et al., 2008). For example, the high percentage of coarse grained end members within Carleton and Danny's Lake suggests a decrease in lake productivity and perhaps enhanced permafrost cover during the LIA (Figures 19ef) (Macumber et al., 2018). The CL-1 temperature record revealed a decrease in temperatures at approximately 1570 CE, however, the temperature estimates that precede this brief cooling event are much warmer and do not reflect the LIA period (Figure 19a). Although this climate reconstruction was deemed significant, research has demonstrated that chironomid-based temperature models based on large training sets can be influenced by secondary gradient effects (Medeiros et al., 2015). Previous research findings have concluded that chironomid assemblages are sensitive to changes in lake depth over time (Kurek and Cwynar, 2009; Engels and Cwynar, 2011). Lake CL-1 has experienced lake expansion in recent decades in response to changes in catchment hydrological process (Korosi et al., 2017). The warmer temperature estimates from ~ 1470 – 1570 CE may be a function of the chironomid fauna responding to changes in environmental factors other than temperature, such as lake depth or surface area. It is important to recognize that this time period was marked by a dominance in *Microtendipes*, which is an indicator of littoral habitat expansion (Brooks et al., 2007; Kurek and Cwynar, 2009; Engels and Cwynar, 2011).

There is a general warming trend within the CL-1 and HL-1 records during the 1600s followed by a peak in temperature at ~ 1700 CE, which was also recorded in the Lake S41 climate record (Macdonald et al., 2009). A temperature peak can also be observed at ~ 1750 – 1790 CE within the CL-1, HL-1 and Carleton Lake (Upiter et al., 2014) chironomid-based records. These brief post-LIA warming periods were also captured in northwestern Canada July temperature tree-ring records (Szeicz and MacDonald 1995; Tomkins et al., 2008), and they may be attributed to changes in radiative forces within the northern hemisphere (Macdonald et al., 2009).

There was a brief decline in temperatures within the CL-1 and HL-1 reconstructions followed by a temperature divergence from ~ 1800 CE to the present day (Figures 19ab). The widely recognized warming trend can be observed in the CL-1 and Carleton Lake (Upiter et al., 2014) chironomid-inferred paleotemperature trends from the 19th century onwards (Figures 19acd). In addition, the substantial decline in coarse-textured end-members observed in the Carleton Lake and Danny Lake sediment records may be indicative of changing sedimentation processes due to enhanced warming (Macumber et al., 2018) (Figures 19efg). In contrast, the HL-1 and Lake S41 chironomid-based records do not present a warming trend from ~ 1800 to 2016 CE (Figures 19bh). Instead, the HL-1 transfer function modelled a general decrease in temperature over this time period, and the Lake S41 climate record suggests temperatures were relatively stable during this time period within the central Northwest Territories. Macdonald et al. (2009) suggested that the recent stable temperature trend at Lake S41 was as a result of enhanced lake stratification, driven by global warming effects, which maintained the presence of cold-water chironomid taxa within the lake sediment. Similarly, Dalton et al. (2018) did not observe a warming shift in the Danny's Lake diatom assemblages from benthic to planktic taxa during the 19th century, that was previously reported from Slipper Lake, Northwest Territories (Rühland and Smol, 2005). Whilst there are similarities between the

regional paleotemperature records within the central Northwest Territories, the differences indicate that the proxy-climate responses are affected by site-specific catchment factors.

Figure 19: Chironomid-based mean July air temperature inferences using the Fortin et al. (2015) transfer function for lakes HL-1 (a) and CL-1 (b) within the North Slave Region (over the last ~ 550 years). Chironomid-based July temperature reconstructions for Carleton Lake, central Northwest Territories using the Barley et al. (2006) transfer function (c) and the Porinchu et al. (2009) transfer function (d) (over the last ~ 1500 years). A sediment record of end-member data from Carleton Lake, central Northwest Territories (e, (dark grey = very fine sand, light grey = fine silt)) and Danny's Lake, central Northwest Territories (f, (dark grey = fine sand, light grey = very coarse silt) and g (dark grey = coarse silt)). A chironomid-inferred July air temperature record for Lake S41, central Northwest Territories over the last ~1500 years (h).

5.6 Potential influences confounding the chironomid-temperature signal

The inconsistencies between the CL-1 and HL-1 temperature reconstructions suggests that other variables, aside from July air temperature, may be influencing the chironomid assemblages in this region. Similarly, discrepancies amongst chironomid-based late Holocene temperature inferences were reported in Fennoscandia (Velle et al., 2005, 2010) and northern Sweden (Bigler et al., 2003). Although chironomid fauna are mostly sensitive to changes in temperature (Walker and Macdonald, 1995), studies have demonstrated that other catchment and limnological factors also play a role in governing the composition and distribution of chironomid assemblages (Velle et al., 2010; Larocque et al., 2006; Larocque and Bigler, 2004). Other environmental variables that have been shown to influence chironomid communities include pH, lake depth, dissolved oxygen levels and lake sedimentology (Berntsson et al., 2014; Michelutti et al., 2011). In addition, many of these environmental factors co-vary with climate, and as a result, obscure the chironomid-temperature signal in certain lake sites (Velle et al., 2010, Brodersen and Quinlan, 2006). In this investigation, we propose a number of factors which may have caused the discrepancy between the CL-1 and HL-1 Lake temperature reconstructions. These factors include environmental variables such as lake stratification, trophic state, lake expansion and precipitation, and limitations associated with the Fortin et al. (2015) model, including the transfer function RMSEP, the low taxonomic resolution of the training set and the short temporal record of the sediment cores.

The most notable difference between the two temperature reconstructions was the absence of a twentieth-century warming trend within the HL-1 temperature reconstruction (Figures 14ab and 19ab). Macdonald et al. (2009) reported a similar twentieth century temperature trend for Lake S41 (Figure 2) and suggested it was due to a decoupling of the chironomid-temperature relationship as a result of climate-induced lake stratification. The Lake S41 biogenic-silica (BSi) content and fossil diatom results from the nearby Slipper Lake suggested that warmer temperatures over the last two centuries increased lake productivity

and enhanced lake stratification within Lake S41. These changes in lake structure may have generated a cold, low oxygen bottom layer which could have confounded the chironomid-temperature signal within this lake (Macdonald et al., 2009; Upiter et al., 2014). Macdonald et al.'s (2009) assertion of this chironomid-temperature decoupling was further supported by a decrease in chironomid head count within the uppermost sediment since recent research has demonstrated that low oxygen conditions stress chironomid communities (Brodersen et al., 2008). HL-1 and S41 are both shallow, small lakes located in the Subarctic, central region of the Northwest Territories. Therefore, it is possible that the lack of a significant warming trend within the HL-1 temperature reconstruction was also a result of lake thermal stratification. However, it is important to recognize that the coring location of Lake HL-1 was shallower than that of the sampling site in Lake S41 (Macdonald et al., 2009), as a result, the impacts of thermal stratification on the chironomid-temperature relationship would have been reduced. In addition, the HL-1 chironomid count per sediment layer did not indicate signs of low oxygen stress on the chironomid community structure over time.

The HL-1 temperature reconstruction was marked by a recent decreasing trend in temperatures, which is not in line with regional climate measurements for this region (National Center for Atmospheric Research, 2020; Environment Climate Change Canada, 2021) (Figure 19b). One hypothesis to explain this discrepancy is the influence of lake productivity on the chironomid assemblage. *Chironomus plumosus*-type, which is associated with shallow, eutrophic lakes, was the chironomid species most strongly, negatively related to the first principal component within the HL-1 sediment record, and it experienced an increase in relative abundance within the uppermost sediment (Bajolle et al., 2018; Brooks et al., 2007). Research has demonstrated that lake productivity and temperature co-vary and therefore it may be difficult to separate the influence from each respective variable on the distribution of chironomid populations (Velle et al., 2010). Lemmen and Lacourse (2018) reported a temperature divergence over the last 100 years between two chironomid-based temperature reconstructions (one based on the Barley et al. (2006) model and the other based on the Fortin

et al. (2015) model) from a low elevation, temperate lake in British Columbia. It was suggested that the recent cooling trend within the Fortin et al. (2015) temperature record was a result of the substantial increase in *Chironomus* head capsules enumerated from the upper sediment, which was most likely due to increased anthropogenic eutrophication. A shift in lake ecology from a mesotrophic to a eutrophic trophic state was also reported by a marked presence in *Chironomus* in a Late Holocene chironomid-inference temperature series from the Netherlands (Gouw-Bouman et al. 2019). Although *Chironomus* is common in warm lakes, it has a lower temperature optimum of 11.13 °C relative to other warm, littoral taxa such as *Microtendipes* (12.46 °C) within the Fortin et al. (2015) training set. Therefore, it is plausible that the recent cooling trend observed in the HL-1 temperature reconstruction was associated with an ecological transition in trophic status to a more eutrophic state.

The marked increase in *Chironomus* within the uppermost sediment of the HL-1 sediment may also be a result of recent wildfire activity in the Great Slave Upland High Boreal Ecoregion (Whitman et al., 2018; Pelletier et al., 2020). Previous studies have reported changes in chironomid abundance as a direct consequence of enhanced nutrient loading from forest fire activity (Delettre, 1994; Giles, 1994). In addition, Francis (2001) found noticeable differences in pre and post-fire chironomid assemblages of Douglas Lake, Michigan, with increases in eutrophic taxa such as *Chironomus*, *Dicoretendipes*, *Glyptotendipes* and *Polypedilum*. Similar results were reported in lacustrine sediment from Chile suggesting that chironomid communities are impacted by perturbation-associated changes in water chemistry and nutrient availability (Araneda et al., 2013). Although this is not conclusive, it is possible that the chironomid-based temperature inferences within this catchment are influenced by environmental gradients associated with fire disturbances.

Precipitation, geology, vegetation and watershed hydrological processes vary across the ecoregions of the central Northwest Territories (Government of the Northwest Territories, 2008). Lake CL-1 is located in the Taiga Plains High Boreal Ecoregion, which receives between

300 and 390 mm of precipitation through the year, whereas Lake HL-1 is located in the Taiga Plains High Boreal Shield, which has a mean annual precipitation range of 280 – 360 mm (Government of the Northwest Territories, 2008). Furthermore, lake HL-1 is surrounded by discontinuous jack pine and aspen-dominated forests and exposed granite outcrops, while lake CL-1 lies within a limestone escarpment and is surrounded by low lying, jack pine-dominated forests (Government of the Northwest Territories, 2008). Differences in these localized factors may have explained the differences in the chironomid diversity and the subsequent variability in the temperature reconstructions between the CL-1 and HL-1 lake records (Palmer et al., 2002). Berntsson et al. (2014) examined the influence of site-specific environmental gradients, such as precipitation and within-lake processes, on the late Holocene chironomid-temperature response within lakes across northwest Sweden. Their research findings suggested that both July air temperatures and precipitation played a role in governing the chironomid assemblages within this northern region. Specifically, they proposed that regional shifts in sedimentation rates, driven by varying precipitation patterns and topographical differences, influenced the composition of chironomid communities within each unique lake environment. This study concluded that varying precipitation patterns are most likely the key driver confounding the chironomid-temperature response within this region, however, this does not preclude the influence of other localized environmental factors, such as topography, lake bottom substrate and in-lake processes, on chironomid-based temperature records (Berntsson et al. 2014). These findings suggest that the spatial heterogeneity and precipitation differences between the CL-1 and HL-1 lake environments may be responsible for the differences in the temperature records presented in this study.

The performance statistics of the Fortin et al. (2015) transfer function may also explain the variability between the CL-1 and HL-1 paleotemperatures. Although this inference model is capable of producing reliable temperature reconstructions across High Arctic, Arctic and boreal regions, it has a high RMSEP_{boot} value of 1.81 °C which may have resulted in less precise temperature reconstructions within each core. Furthermore, the Fortin et al. (2015) model

combined many of the taxon (e.g. *Tanytarsus* and *Paratanytarsus* species) within the *Tanytarsini* tribe into a single category called *Tanytarsina-other* which had a low temperature optimum value of 9.83 °C. The coarse resolution of the Fortin et al. (2015) training set may have resulted in less precise temperature reconstructions since there are a range of individual species temperature tolerances within the broad category *Tanytarsini-other* (Walker, 2001; Engels and Cwynar, 2011). Research has demonstrated that transfer functions built on training sets with a more detailed classification of *Tanytarsini* species result in more accurate taxon temperature estimates and more precise temperature reconstructions (Larocque et al., 2006).

In addition, the temperature fluctuations within the late Holocene have been small ($\sim 1 - 2$ °C) relative to the glacial-interglacial transition and they are fairly similar to the errors (e.g. $RMSEP_{boot}$) associated with chironomid-based inference models (Velle et al., 2005, 2010; Engels et al., 2019). Recently, studies have concluded that chironomid-based inference models may not be able to successfully reconstruct small-scale Holocene temperature fluctuations (Kurek et al., 2004; Birks and Birks, 2006; Engels et al., 2019). These conclusions, along with the fact that there were no significant changes in the chironomid record, may explain why certain late Holocene climate change events were absent from the CL-1 and HL-1 temperature reconstructions, and why the HL-1 temperature record suggests that the region did not undergo any substantial climatic change over the last ~ 500 years.

5.7 Climatic influences on the paleofire signal

The long-term trends in charcoal accumulation rate are fairly similar between the two lake sites, however, CL-1 has a substantially higher charcoal accumulation rate than HL-1 (Figures 17 and 18). Pelletier et al. (2020) explained that this variability was as a result of differences in lake basin morphology and catchment processes between the two lakes. The charcoal accumulation and fire event record data suggest that the North Slave Region experienced low magnitude wildfire activity from ~ 1470 to the twentieth-century and more

frequent wildfires from the twentieth century to the present day (Figures 17 and 18). Sulphur et al. (2016) reported similar late Holocene paleofire trends for the Waite Lake boreal forest region located within the central Northwest Territories. Due to the steady charcoal accumulation rate and low resolution of the age-depth model, it was not possible to identify specific fire events before approximately 1850 CE (Pelletier et al., 2020). Neither of the reconstructed temperature series had a significant correlation to charcoal accumulation, implying that summer temperature may not be the sole variable driving wildfire activity in this region. The correlation statistic was near significant for HL-1 though (p -value = 0.06), suggesting that July air temperature is a more important factor in the fire behaviour of this catchment compared to CL-1. Although this study did not establish a statistical relationship between mean July air temperature estimates and charcoal accumulation, temperature is a key factor governing forest fire activity within the Northwest Territories and surrounding regions, due to its direct influence on fuel flammability (Chipman et al., 2015; Coogan et al., 2019). In recent decades, it has been reported that the high burn severity and more frequent forest fires is a result of the twentieth-century warming trend (Sulphur et al., 2016). In particular, high temperatures and prolonged drought conditions are responsible for the severe 2014 wildfire season that caused more than 380 fires and burned 3.4 million hectares of forests (Gaboriau et al., 2020).

The analysis between the tree-ring-based spring precipitation data and known fire events suggests that there is no clear correlation between these two variables in the North Slave Region (Figures 17 and 18). In the CL-1 record, four of the six known fire events experienced higher than normal spring precipitation, which is not surprising since lightning strikes associated with thunderstorm activity ignite the majority of the forest fires within the Northwest Territories (Kochtubajda et al., 2006). Observational data have demonstrated that lightning strikes correlate with temperature and precipitation in the northern boreal forest (Coogan et al., 2019), and Veraverbeke et al. (2017) reported that lightning ignited several wildfire events during the 2014 (Northwest Territories) and 2015 (Alaska) wildfire season. Three

of the known HL-1 fire events occurred during years with lower-than-average precipitation, including severe 2014 wildfire which was driven by both prolonged drought conditions and lighting strikes (CBC News, 2014; Veraverbeke et al. (2017); Whitman et al., 2018). It is likely that HL-1 forest fire events are associated with moderate – severe drought conditions since these weather conditions enhance forest moisture loss and promote burning (Kochtubajda et al. 2006; Gaboriau et al., 2020). In addition, northern boreal forest late Holocene paleofire records have revealed that periods of dry, warm conditions are associated with more frequent burning (Morris et al. 2015; Sulphur et al., 2016).

6. Conclusion and Future Work

For this investigation we generated late Holocene (~ 1470 – 2016 CE) chironomid-based reconstructions of mean July air temperature for the North Slave Region. The fossil chironomid records utilized in this study were collected from two shallow boreal lakes and they were characterized with warm, littoral taxa. No substantial changes in chironomid assemblages were observed in either of the lake sediment records. The paleotemperature estimates were warmer than those previously reported from more northern lake systems within the central Northwest Territories. However, the contemporary chironomid-temperature inferences underestimated the instrumental mean July air temperature data for this region. This result indicated that applying the Fortin et al. (2015) transfer function to shallow, boreal lake records might not be the best paleoclimatological approach for reconstructing accurate July air temperatures. Despite both chironomid records having strong modern analogues, the CL-1 chironomid-temperature data was deemed significant while the HL-1 chironomid-temperature record was not. Although there were some similar climate events between the two climate reconstructions from approximately 1600 – 1800 CE there were also notable discrepancies. The most striking deviation was the absence of the recent global warming trend in the HL-1 climate record. Although not wholly synchronous, similar late Holocene climate events were identified between the temperature reconstructions presented in this study and other regional proxy-

based temperature records. These include, a cool period associated with the latter portion of the LIA, brief temperature peaks within the 1700s CE and the warming trend that has persisted since approximately 1850 CE (Upiter et al., 2014). The observed deviations among the CL-1 and HL-1 temperature trends indicated that the chironomid temperature-response may be influenced by localized catchment characteristics and processes, such as lake morphometry, lake trophic status, fire impacts, and precipitation (Francis, 2001; Larocque et al., 2001; Velle et al., 2005, 2010; Berntsson et al., 2014). In addition, this result suggested that chironomids may not be a suitable proxy for reconstructing the small-scale temperature variability observed in late Holocene climate records.

The paleofire data presented by Pelletier et al. (2020) indicated that the North Slave Region experienced an increase in fire frequency since the onset of the twentieth century. Our paleofire-climate analysis did not yield a significant correlation between charcoal accumulation and reconstructed July air temperatures, nor did it establish a clear relationship between known fire events and June precipitation, however, it did provide further insights on the complex influence of climate on fire occurrence within this boreal region.

Through the use of paleoclimatological approaches we have been able to generate quantitative chironomid temperature estimates for the North Slave Region. However, the variability within our climate records highlight the need for additional proxy-based studies to increase our understanding of the natural long-term climate variability of this region, and to provide further insights into the utility of the chironomid transfer approach. Future studies should involve a multi-proxy approach in an effort to gain a more extensive understanding of the paleoenvironmental changes that have occurred in this boreal environment during the Holocene (Burge et al., 2018). An analysis of multiple July air temperature reconstructions derived from different proxy-based transfer functions (e.g. chironomids, pollen, cladocera) has proven useful in developing a comprehensive record of climate change (Birks and Birks, 2006).

Adopting this approach in the North Slave Region may help resolve some of the climate discrepancies observed in this study.

Non-climatic variables may be impacting the chironomid-temperature reconstructions presented in this study, however, due to a lack of available limnological data we were not conclusively able to identify these environmental variables. Larocque et al. (2006) investigated the factors that affect the distribution of chironomid communities in northwestern Quebec using 33 environmental characteristics. They concluded that August air temperature has the largest impact, and they identified that other environmental parameters (e.g. lake depth and DOC) are most likely influencing the chironomid-temperature signal in this region. Similar research should also be aimed at better understanding the factors that influence chironomid assemblages in the central Northwest Territories in an effort to improve the interpretation of midge paleotemperature estimates. In addition, to reduce the interference of lake-specific characteristics on the chironomid-temperature response, multiple lake sites should be sampled when generating future regional paleoclimate reconstructions. This approach will help increase the precision and robustness of chironomid-based temperature records within this region (Birks and Birks, 2006).

Due to the coarse resolution of the age-depth model, the record of North Slave Region fire events only extends back to the late nineteenth century (Pelletier et al., 2020). Previous studies have employed multiple proxies derived from lake sediments to successfully reconstruct Holocene paleofire records and vegetative histories (Brown et al., 2019; Gaboriau et al. (2020)). Similar paleolimnological approaches should be applied to North Slave Region lakes to better characterize the temporal climate-vegetation-fire dynamics within this boreal region.

Incorporating these recommendations into future research will improve the reliability of proxy-based environmental modelling, and will lead to a more nuanced understanding of the long-term climate and environmental variability of the North Slave Region. The natural

variability captured in these paleodata is vital for evaluating the magnitude and effects of the current global warming trend impacting this region.

7. References

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Appendix

Figure A1: The squared-chord distances between the CL-1 fossil assemblages and the Fortin et al. (2015) modern analogues over time.

Figure A2: The squared-chord distances between the HL-1 fossil assemblages and the Fortin et al. (2015) modern analogues over time.

Table A3: Unpublished total phosphorous data for CL-1 and HL-1 that was generated from collected water samples.

Lake Core	Total Phosphorous (mg/L)
CL-1	0.03
HL-1	0.01