The Effects of Shoreline Retrogressive Thaw Slumping on the Hydrology and Geochemistry of Small Tundra Lake Catchments

by

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Supervisory Committee

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Abstract

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The overall goal of this study was to examine the hydrological and geochemical linkages between the contributing landscape and small tundra lakes affected by shoreline retrogressive thaw slumping (SRTS) in the upland region north east of Inuvik, NT. In 2007, 2008, and 2009, detailed hydroclimatological and geochemical data were obtained from a pair of representative tundra lake catchments (Lake 5A: Control; Lake 5B: Affected by SRTS). This was supplemented with less detailed data obtained from 10 regional small tundra lake catchments (control and affected by SRTS). The hydrology and geochemistry of Lake 5A and Lake 5B exhibited strong seasonal variability that was characterized by spring snowmelt. For the three study years, Lake Level (LL) peaked during spring snowmelt, when the addition of melt water from the contributing landscape led to a rapid rise in LL that was enhanced by snow and ice damming the outlet channel. The addition of this relatively dilute runoff water led to a decrease in the concentration of most major ions and nutrients in the study lakes over the spring months. Notably, the concentration of nutrients increased at the beginning of spring snowmelt, due to the mobilization of surficial organic materials by runoff, before decreasing as runoff to the lake became more diluted. Recent changes in key hydroclimatic factors have likely affected the hydrology and geochemistry of the study lakes. The examination of a suite of hydroclimatic indicators, derived from historical climate data, indicated that the annual May 1st snowpack in Tuktoyaktuk has been increasing at a significant rate over the past half century. Furthermore, detailed snow survey data suggested that the capture of snow by SRTS-affected terrain increases the snowmelt contributions to small tundra lakes. An increase in the contribution of snowmelt inputs to the lake water balance could lead to a higher peak LL and more dilution of lake water. In addition to hydro-climatic drivers, the geochemistry of the study lakes was also driven by SRTS. SRTS-affected lakes had significantly higher concentrations of major ions than unaffected study lakes, due to the addition of relatively ion-rich runoff from SRTS-affected terrain during the spring and summer months. The outlet channels draining the SRTS-affected

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Chapter 1: General Introduction

1.1 Introduction

The upland region east of the Mackenzie Delta is characterized by the presence of thousands of small tundra lakes (**Figure 1.1**). The hydrology of small tundra lakes in this region is directly influenced by climatic and landscape-level factors (e.g., temperature, precipitation, mineral earth hummocks, and near surface permafrost) (Pohl et al., 2009; Quinton and Marsh, 1998; Quinton and Marsh, 1999). Recent climate warming has led to changes in ambient air temperature, precipitation, and permafrost extent. Since the 1940s, the mean annual air temperature in the Inuvik region has increased by approximately 3°C and is projected to increase by an additional 4 to 7°C over the next century (AMAP, 2012; Government of the Northwest Territories (GNWT), 2008). Notably, the rate of climate warming will be the greatest during the autumn and winter months. For instance, the mean autumn and winter air temperature of the Inuvik region is projected to increase by between 3 and 6°C by 2080 (AMAP, 2012). This has significant implications for arctic freshwater systems, because autumn and winter air temperature controls key hydrological processes, such as snow and ice formation (Ashton, 1983; Ashton, 1986; AMAP, 2012).

Recent climate warming has led to changes in a number of key climatic and hydrologic drivers of the small tundra lake water balance, which include decreases in snow cover extent, earlier and more intense spring snow and ice melt, longer open-water periods, and increases in the rate of permafrost degradation (AMAP, 2012; Bonsal and Prowse, 2003; Burn, 2008; Lantz and Kokelj, 2008; Lesack et al., 2013). Another key issue facing arctic regions is increases in precipitation. Over the past century, precipitation has increased by 5 to 8% and is projected to increase by up to 35% by the year 2100 (ACIA, 2005; AMAP, 2012). The greatest change in precipitation will be observed in autumn and winter (AMAP, 2012).

Evidence suggests that recent climate warming has led to an increase in the rate of permafrost degradation across the circumpolar arctic, an effect that has important implications for the hydrology and geochemistry of arctic freshwater systems (AMAP, 2012; Frey and Mclelland, 2009; Lantz and Kokelj, 2008; Smith et al., 2005). The hydrology and geochemistry of small tundra lakes in regions of continuous permafrost is directly controlled by seasonal active layer depth (Hinzman et al., 1991; Quinton and Marsh, 1999; Quinton and Pomeroy, 2006; Woo, 2012). The active layer is the zone of seasonally unfrozen ground lying above the permafrost that directly controls the vertical infiltration and water residence times of runoff and determines the importance of

Figure 1.1. A photo of the upland region east of the Mackenzie

subsurface flow, relative to surface flow, to freshwater systems (Hinzman et al., 1991; Quinton and Marsh, 1999). Permafrost degradation leads to thicker summer active layers, which could act to increase the vertical infiltration, water residence times, and subsurface storage of runoff, all of which have important implications for the geochemistry of small tundra lakes (Keller et al., 2010; Lantz and Kokelj, 2008; Quinton and Marsh, 1999; Quinton and Pomeroy, 2006; Smith et al., 2005). For instance, Quinton and Marsh (1999) found that the geochemistry of runoff pathways was directly controlled by the importance of subsurface flow relative to surface flow and postulated that permafrost degradation would likely increase the amount of ion-rich subsurface flow to freshwater systems.

In extreme cases, permafrost degradation can lead to shoreline retrogressive thaw slumping (SRTS). SRTS is a notable outcome of permafrost degradation in this region that affects approximately 8% of lakes with a surface area greater than 1ha (Lantz and Kokelj, 2008). SRTS occurs when the ice-rich surface sediments making up the shoreline thaw, become unstable, and slump into the adjacent lake (Burn and Friele, 1989). SRTS activity has increased in this region since the 1970's as a result of warmer ambient air and ground temperatures (Lantz and Kokelj, 2008). In other extreme cases, permafrost degradation can lead to the rapid drainage of lakes. The morphology of the outlet channels draining small tundra lakes in the study region is often defined by the presence of permafrost. The melting of the ice-rich permafrost within the outlet channel can lead to the rapid drainage of small tundra lakes.

This study focuses on small tundra lakes affected by SRTS. In recent years, the effects of SRTS on small tundra lakes has become of increasing interest to the scientific community as an analogue for the effects of permafrost degradation on arctic freshwater systems. Studies have focused on the effects of SRTS on lake catchment geomorphology, lake catchment vegetation, lake geochemistry, and lake biota (Kokelj et al., 2005; Kokelj et al., 2009a; Kokelj et al., 2009b; Lantz et al., 2009; Mesquita, 2008; Moquin, 2011; Moquin et al., 2012; Thompson, 2009; Thompson et al., 2012).

Studies suggest that SRTS modifies the geochemistry of tundra lakes in the upland region east of the Mackenzie Delta, initiating a number of in-lake biological responses (Kokelj et al., 2005b; Kokelj et al., 2009b; Mesquita, 2008; Thompson, 2009; Thompson et al., 2013). Lakes affected by SRTS typically have higher ionic concentrations than unaffected catchments. The elevated ionic concentrations associated with SRTS-affected lakes appear to be related to lower concentrations of dissolved organic carbon (DOC), which results in less colour and less light attenuation within the water column (Mesquita, 2008; Thompson, 2009; Thompson et al., 2012). Moquin (2011), Moquin et al. (2014), and Thompson et al. (2012) found that charged clay particles within the water column, deposited into the lake via SRTS, bind with DOC, causing it to fall out of the water column to the bottom of the lake. By doing this, SRTS directly controls the processes driving production, such as the availability of photosynthetically active radiation throughout the water column. Shifts in the foodweb associated with these changes are of particular interest to local communities because they may affect fish populations.

The effects of SRTS on the geochemistry and ecology of small tundra lakes has been explored, but the landscape-level hydrological processes driving the observed hydro-bio-geochemical effects are still largely unstudied. In addition to the impacts of climate change, the effects of SRTS on the water balance of small tundra lakes are still largely unknown.

The overall goal of this study is to further our understanding of the hydrogeochemical response of small tundra lakes to climate variability and change and SRTS in the upland region northeast of Inuvik, NT.

1.2 Literature Review

The purpose of this literature review is to provide an overview of current knowledge on the hydrology and geochemistry of small tundra lakes in regions of continuous permafrost.

1.2.1 The Hydrology of Small Tundra Lakes in a Region of Continuous Permafrost

A water balance approach is a good way to review what is understood about the hydrology of small tundra lakes because it describes how key hydrologic and climatic factors act together to affect the lake water level.

Water Balance

The water balance of small tundra lakes in the study region consists of three main components: inputs (i.e., precipitation and runoff), storage (i.e., surface water and subsurface water), and outputs (i.e., evaporation and discharge). Pohl et al. (2009) developed the following water balance equation for small tundra lakes located at the heart of the study region:

Equation 1.1. The summer water balance of small tundra lakes

$$LL = (Q_{in} - Q_{out}) + (P - E) +/- S$$

where LL is lake level in m, Q_{in} and Q_{out} are water inflows and outflows in mm d^{-1}, P is precipitation onto the lake in mm d^{-1}, E is evaporation from the lake in mm d^{-1}, and S is the change in lake storage in mm d^{-1}.

Inputs

The timing and magnitude of water inputs (spring snowmelt and summer rainfall runoff) to small tundra lakes are directly controlled by climatic factors (temperature and precipitation), which exhibit high seasonal variability. In early winter (late September and early October), ambient air temperature decreases to below 0°C. Precipitation is typically in the form of snow, which is stored on the surface of the lake catchment until spring snowmelt. The annual snowpack develops over the approximately 8 to 10 month winter period, reaching a maximum snow water equivalent in late April. Water runoff to small tundra lakes is minimal over the winter period, as the ephemeral rills that feed most small tundra lakes are either dry or frozen to the bottom (Pohl et al., 2009; Quinton and Marsh, 1999; Quinton and Pomeroy, 2006).

In early-spring, ambient air temperature increases to above 0°C. At this time, precipitation that was originally stored in the form of snow and ice is rapidly transported to freshwater systems via runoff (Quinton and Marsh, 1999). Surface runoff is the dominant runoff pathway because the active layer is still frozen (Quinton and Marsh, 1999). The frozen active layer prevents the vertical infiltration of water runoff into the soil profile, preventing subsurface runoff (Hinzman et al., 1991; Quinton and Marsh, 1999). Spring snow and ice melt, referred to as the spring freshet, is an important source of water recharge to arctic freshwater systems. In regions of continuous permafrost, the spring freshet is often the most significant hydrologic event of the entire year (Quinton and Marsh, 1999). Stream discharge during the spring period is directly controlled by antecedent winter and spring melt climatic conditions (Quinton and Marsh, 1999).

Research suggests that recent climate warming has led to earlier spring freshet periods. Bonsal and Prowse (2003) identified a significant decreasing trend in the timing of the spring 0°C isotherm, or the first day that a 31-day running mean of ambient air temperature increases to above 0°C, for a number of locations north of 60° N, thereby correlating recent climate warming to earlier spring freshet periods. Similarly, Burn (2008) found a significant decreasing trend in the timing of the spring freshet for a number of locations along the Mackenzie River. Notably, the effect of recent climate warming on the timing of the spring freshet at small tundra lake catchments in this region is still largely unknown.

Studies also suggest that recent climate warming has led to more intense spring snowmelt periods. For the East Channel of the Mackenzie River, at Inuvik, Lesack et al. (2013) found that the period of time falling between the initiation of the spring freshet and peak river discharge has shortened by 8 days since 1964. Earlier and more intense spring snowmelt periods have important implications for small tundra lakes. For instance, earlier and more rapid snow and ice melt could lead to longer open-water periods. Longer open-water periods, in conjunction with warmer summer air temperatures, could lead to greater rates of evaporation, which is a key driver of the summer water balance of small tundra lakes (Pohl et al., 2009). Furthermore, more intense spring snowmelt periods could lead to a more rapid rise in lake level in early-spring.

Notably, high water levels are one of the main factors contributing to the rapid drainage of thaw lakes (Pohl et al., 2009).

At Trail Valley Creek, located approximately 50 Km northeast of Inuvik, the active layer deepens with the progression of the spring and summer months. As a result, the vertical infiltration of runoff increases and the majority of runoff is transported through the upper and lower peat layers . The active layer is comprised of organic and mineral soil. The organic layer is extremely porous and has long water residence times. Consequently, stream discharge typically decreases during the late spring and summer months (Quinton and Marsh, 1999). Seasonal variability in active layer depth directly controls runoff pathways and is integrally linked to stream discharge (Hinzman et al., 1991; Quinton and Marsh, 1999).

Lake Storage

The water balance of small tundra lakes in winter is not well known, largely because extreme weather conditions make them difficult to access (Woo et al., 2008). It is generally accepted that lake storage does not fluctuate by much over the winter months, because the inputs and outputs to and from the lake are minimal (Pohl et al., 2009; Quinton and Marsh, 1999; Woo et al., 2008). Similar to stream and river systems, lake storage typically increases during the spring freshet, which typifies the water balance of many small tundra lakes (Pohl et al., 2009; Quinton and Marsh, 1999). In the absence of spring snowmelt, the summer water balance of small tundra lakes is mainly driven by rainfall and evaporation (Pohl et al., 2009). In arctic regions, evaporation typically exceeds precipitation during the summer months, leading to a decline in lake storage. For this reason, the summer is often referred to as a period of drying for small tundra lakes (Marsh and Bigras, 1988; Pohl et al., 2009; Rouse et al., 2003).

The storage capacity of small tundra lakes is determined by the elevation of the outlet channel (Spence and Woo, 2006). When LL reaches the mouth of the outlet channel, the storage capacity of the lake has been reached and lake drainage is initiated. In early spring, the outlet channel of small tundra lakes is typically blocked by snow and ice, which prevents lake drainage and allows LL to rise to above the storage capacity of the lake (Woo, 1980). Conversely, LL may decrease below the storage capacity of the outlet channel during the summer months, causing lake drainage to cease.

Outputs

In the upland region east of the Mackenzie Delta, the two primary outputs from the small tundra lake water balance are evaporation and discharge via the lake outflow channel (Pohl et al., 2009). In subarctic and arctic regions, lake evaporation exhibits high seasonal variability, which is primarily driven by the length of the open-water period and the surface area and depth of the lake (Oswald and Rouse, 2004; Rouse et al., 1997; Schindler and Smol, 2006). During winter and early spring, lake evaporation is impeded by the presence of a thick ice cover (Marsh and Bigras, 1988; Oswald and Rouse, 2004; Quinton and Marsh, 1999; Pohl et al., 2009). For small lakes, evaporation typically increases from the onset of the open-water period and peaks in late summer. Whereas, for large lakes, evaporation typically peaks in late fall and early winter (Oswald and Rouse, 2004). Potential increases in mean annual summer air temperatures associated with climate warming will likely increase the amount of water lost from small tundra lakes through evaporation, due to changes in the surface energy balance and longer ice-free seasons (AMAP, 2012; Burn, 2002; Prowse et al., 2009). AMAP (2012) predicted that arctic inland regions, such the upland region east of the Mackenzie Delta, will become drier in the upcoming years, due to longer open-water seasons and warmer summer air temperatures.

In winter and early spring, discharge from small arctic lakes is typically minimal, because snow and ice that has accumulated in the outflow channels draining small arctic lakes impedes lake drainage (Kane et al., 1991; Pohl and Marsh, 2006; Woo, 1980). Even after spring snowmelt increased LL to the elevation of the outflow channel, the presence of snow and ice in the outflow channel can prevent lake drainage (Kane et al., 1991; Pohl and Marsh, 2006; Woo, 1980; Woo, 2012). This process is often referred to as snow and ice damming (Woo, 1980; Woo, 2012). Once the lake water carves a trench through the snow and ice dam, lake drainage is initiated. Lake drainage is typically minimal during the summer months because LL generally decreases to below the outlet of the lake (Spence and Woo, 2006).

Lake drainage is primarily driven by LL and the physiology of the outflow channel (Marsh and Neumann, 2001; Spence and Woo, 2006). In the upland region northeast of Inuvik, the physiology of the outlet channel is typically controlled by the presence of ice-rich permafrost. For some small tundra lakes in the region, the melting of that ice-rich permafrost, associated with elevated LL and warm ambient air and ground temperatures, has led to the rapid drainage of the associated lake. Rapid drainage is one of the ways permafrost degradation in the

Northwestern Arctic, associated with recent climate warming, has impacted the hydrology of small tundra lakes.

Shoreline Retrogressive Thaw Slumping

Another way permafrost degradation has affected small tundra lake catchments is SRTS. SRTS typically results in physical modifications to the contributing lake catchment, which include: the removal of near-surface permafrost; thickening of the active layer; the development of large depressions on the lake shoreline (**Figure 1.2**); the removal of vegetation, the organic litter layer, and the organic soil horizons; warmer ground temperatures; and the expansion of the contributing lake catchment (Burn and Friele, 1989; Kokelj et al., 2009a). These physical modifications have a number of implications for the water balance of small tundra lakes. Burn and Friele (1989) found that the summer active layer within an area affected by SRTS can be up to 3m deep. That is approximately 4 times deeper than that of unaffected soils. This delays active layer freeze-back in winter. Lantz et al. (2009) found that soils affected by SRTS can take 51 to 139 days longer to freeze-back in winter than unaffected soils. By increasing the depth of the active layer, SRTS could increase the vertical infiltration of runoff into the soil profile and, in turn, the amount of water stored in the subsurface component of the lake water balance.

Figure 1.2. A photo of an active shoreline retrogressive thaw slump that formed on the shoreline of a small tundra lake on Richard’s Island.

The effects of SRTS are not confined to the portion of the lake catchment that is slumping. Hinzman et al. (1991) observed that soil moisture migration, evapotranspiration, sublimation, and freezing or thawing directly control the thermal regime of the active layer for the Alaskan North Slope. This suggests that increases in the vertical infiltration of melt water into the soil profile, associated with SRTS, will likely increase the transport of heat to adjacent soils. Kokelj et al. (2009a) found that the unaffected terrain located next to SRTS typically develops deeper active layers than unaffected terrain that is not located next to SRTS. Furthermore, they found that the lateral transport of heat from SRTS-affected soils created large depressions in the lake bottom adjacent to the slump. Kokelj et al. (2009a) observed that more than 90% of shoreline slumps in the upland region east of the Mackenzie Delta are multi-aged and occur within regions of previous slumping. They attributed this to the destabilization of the landscape that occurs when these large depressions form in the lake bottom adjacent to SRTS.

Lantz et al. (2009) found that SRTS-affected terrain had deeper winter snowpacks than adjacent unaffected terrain. In the Trail Valley Creek Research Basin, Pomeroy et al. (1997) and Pohl et al., (2006) found that the largest snow water equivalent (SWE) estimates were typically associated with snowdrifts, which tend to form in depressions, such as those characteristic of SRTS. This suggests that SRTS may increase the snowmelt contribution of the contributing lake catchment. It is important to note, however, that the effects of SRTS on the SWE of affected lake catchments is still largely unknown. Since spring snowmelt is typically the most significant hydrological event for arctic freshwater water systems, SRTS may have significant implications for the water balance and geochemistry of affected lake catchments (Quinton and Marsh, 1999).

1.2.2 The Geochemistry of Small Tundra Lake Catchments in a Region of Continuous Permafrost

In the upland region east of the Mackenzie Delta, landscape-level hydrological processes have been found to directly control the geochemistry of runoff and stream pathways (Quinton and Pomeroy, 2006; Keller et al., 2010). Stream discharge is driven by the addition of “new” water in spring and “old” water in summer (Quinton and Pomeroy, 2006; Woo et al., 2008). “New” water reaches streams and lakes via *surface* flow pathways over the frozen ground, and thus, does not have long residence times within the soil and has had little time to interact chemically with the organic and mineral soils that dominate the active layer (Woo et al., 2008). “Old” water travels via *subsurface* flow pathways in the active zone, and thus, has resided in the soil for long periods of time. "Old" water is often composed of melt water, rain water, and water that has recently been liberated from near-surface permafrost (Keller et al., 2007; Quinton and Pomeroy, 2006; Woo et al., 2008). The relative contribution of "new" and "old" water runoff to Arctic rivers, lakes, and streams drives the seasonal geochemistry of these freshwater ecosystems (Quinton and Pomeroy, 2006; Woo et al., 2008).

Arctic tundra lakes are generally oligotrophic to ultra-oligotrophic and thus, landscape-level processes can have a significant impact on their geochemistry (Pienitz et al., 1997). Two potential sources of major ions include precipitation and runoff (Pienitz et al., 1997; Quinton and Pomeroy, 2006). At Trail Valley Creek, located approximately 50 Km away from Inuvik, Quinton et al. (2006) found that Na^+ and Cl^- were the dominant ions in runoff at the beginning of the spring freshet. As the spring and summer months progressed and the importance of subsurface runoff relative to surface runoff increased, Ca^{2+} and Mg^{2+} became the dominant ions in runoff. This suggests that Na^+ and Cl^- in small tundra lakes in this region are likely derived from atmospheric deposition via precipitation, where as Ca^{2+} and Mg^{2+} are more likely derived from the mineral soil layers that make up the active layer. This is in line with the work of Kokelj et al. (2009b) and Pienitz et al. (1997), who found that the concentration of Na^+ and Cl^- in small tundra lakes in the study region was correlated with proximity to the Beaufort Coast.

Precipitation and runoff are also a potential source of nutrients to arctic freshwater systems. Quinton and Pomeroy (2006) found that the concentration of nutrients in runoff increased at the beginning of spring snowmelt. They proposed that this was likely due to the mobilization of organic materials that occurs when runoff off is initiated. Similarly, MacIntyre et al. (2006) found that heavy rainfall events often led to nutrient loading into Toolik Lake, Alaska. They proposed that increases in summer rainfall, associated with projected climate warming, could lead to increases in nutrient loading to arctic freshwater systems. This suggests that runoff initiated by precipitation can be a source of nutrients to small tundra lakes. Changes in the meltwater contribution of the contributing lake catchment, associated with climate change and SRTS, will likely affect the concentration of major ions in freshwater systems during the spring freshet period.

Research suggests that near-surface permafrost is also a source of major ions and nutrients to arctic freshwater systems (Hobbie et al., 1999; Keller et al. 2007; Kokelj and Burn, 2005; Kokelj et al., 2009b; Keller et al., 2010). In regions of continuous permafrost, chemical interactions between the soil profile and runoff are confined to the active layer. As a result, major ions, such as Ca^{2+}, Mg^{2+}, Na^{+}, and K^{+}, and nutrients, such as Phorphorus, leach out of the active layer over time in runoff (Keller et al., 2007; Kokelj et al., 2005b). This is one of the reasons that the active layer typically has significantly lower concentrations of major ions and nutrients than near surface permafrost (Keller et al., 2007). As near-surface permafrost degrades, the vertical infiltration of runoff increases, which allows runoff water to interact chemically with the newly liberated ion and nutrient-rich soils, potentially increasing the concentration of major ions and nutrients in water runoff. It's important to note that little is known about the physical and chemical interactions between subsurface and surface runoff in regions of continuous permafrost and how projected permafrost degradation will impact aquifers.

In recent years, the effects of SRTS on small tundra lakes have been used used as an analogue for the potential effects of permafrost degradation on arctic freshwater systems. Using data obtained from 11 paired lakes (unaffected vs. affected by SRTS), Kokelj et al. (2005; 2009b) found that small tundra lakes affected by SRTS typically have higher ionic concentrations than unaffected lakes (Kokelj et al., 2005; Kokelj et al., 2009b). They postulated that SRTS acts to liberate mineral particles from previously frozen soils, which affects the water quality of the runoff pathways supplying water to shallow tundra lakes. Kokelj et al. (2009b), who found that the ionic concentration, hardness, and alkalinity of affected lakes tended to decrease with the relative age of the disturbance, suggesting that ions leach out of slumped soils over time. Increases in the deposition of charged mineral particles into the lake water column, associated with SRTS, has significant implications for the lake ecosystem.

For the same 11 paired lakes, Thompson et al. (2012) found that small tundra lakes affected by SRTS have significantly lower concentrations of Total Phosphorus and Total Dissolved Nitrogen than unaffected lakes. This was partially attributed to sedimentation. That is, dissolved organic matter binds with charged mineral particles in the water column and settles at the bottom of the lake. This is in line with the work of Thompson et al. (2008). Thompson et al. (2008) put varying amounts of slump sediments into small mesocosms with humic lake water. They found that the concentration of dissolved organic matter in the lake water decreased over time. Furthermore, the mesocosms that had more slump sediments had lower concentrations of dissolved organic matter than the mesocosms that had less slump sediments. This has significant implications for the lake ecosystem. For instance, Mesquita (2008) found that sedimentation, associated with SRTS, promotes macrophyte growth. Furthermore, Moquin et al. (2014) found that SRTS affected lakes had significantly different macroinvertebrate communities than unaffected lakes. This suggests that changes in the geochemistry of small tundra lakes, associated with SRTS, directly affects ecosystem structure.

Although the effects of SRTS on the geochemistry of shallow tundra lakes is well known, the landscape-level hydrological processes driving the observed effects and resulting effects on ecosystems structure and function are still largely unknown (Keller et al., 2007; Kokelj et al., 2005; Kokelj et al., 2009b).

1.3 Purpose

The purpose of this thesis project is to investigate the hydrological and geochemical linkages between the contributing landscape and small tundra lakes affected by SRTS in the upland region adjacent to the Mackenzie Delta. This will be achieved by examining hydrometric and geochemical data obtained from representative small tundra lake catchments.

1.4 Broad Objectives

The objectives of this study expand on previous work done by Kokelj et al. (2005; 2009b), Thompson et al., (2008; 2012), Thompson (2009), Mesquita (2008), Moquin (2011), and Moquin et al. (2014).

Objective 1: Examine key hydroclimatic drivers of the small tundra lake water balance to assess how historical climate variability and change and the presence of SRTS affects the hydrology of small tundra lake catchments.

Objective 2: Examine the geochemical signature of catchment runoff to and from small tundra lakes to assess the impacts of runoff from the contributing catchment, including terrain affected by SRTS, on the geochemistry of small tundra lakes.

Chapter 2: Study Area

2.1 Study Region

The Mackenzie Delta is located in the Northwest Territories, Canada, where the Mackenzie River drains into the Beaufort Sea. East of the Mackenzie Delta is an upland region that is typified by an abundance of thermokarst lakes. Surface water makes up more than 15% of the total surface area of this region (Kokelj et al., 2005; Marsh and Neumann, 2001). The hydrological processes that drive the water balance of thermokarst lakes in this region are unique because of the presence of near-surface permafrost and the potential enlargement due to the thawing of ground ice (Davis, 2001; Hinzman et al., 1991; Pohl et al., 2009; Quinton and Marsh, 1998; Quinton and Marsh, 1999). SRTS is a common feature in the study region that occurs along the shoreline of approximately 8% of small tundra lakes (Kokelj et al., 2009b; Lantz and Kokelj, 2008).

The proposed study focused on a pair of lakes located at the southern end of the upland region east of the Mackenzie Delta, near Noell Lake, supplemented with data collected at 10 additional lakes located at the northern end of the upland region east of the Mackenzie Delta and Richards Island (**Figure 2.1**). 11 of the 12 study lakes are part of an extensive International Polar Year/ArcticNet study that examined 66 paired lakes (i.e., unaffected and affected by SRTS) lying parallel to a transect of the proposed Mackenzie Valley Natural Gas Pipeline, which runs from as far south as Inuvik to as far north as Tuktoyaktuk, NT. Notably, the extension of the Dempster Highway that runs from Inuvik to Tuktoyaktuk will also transverse this region.

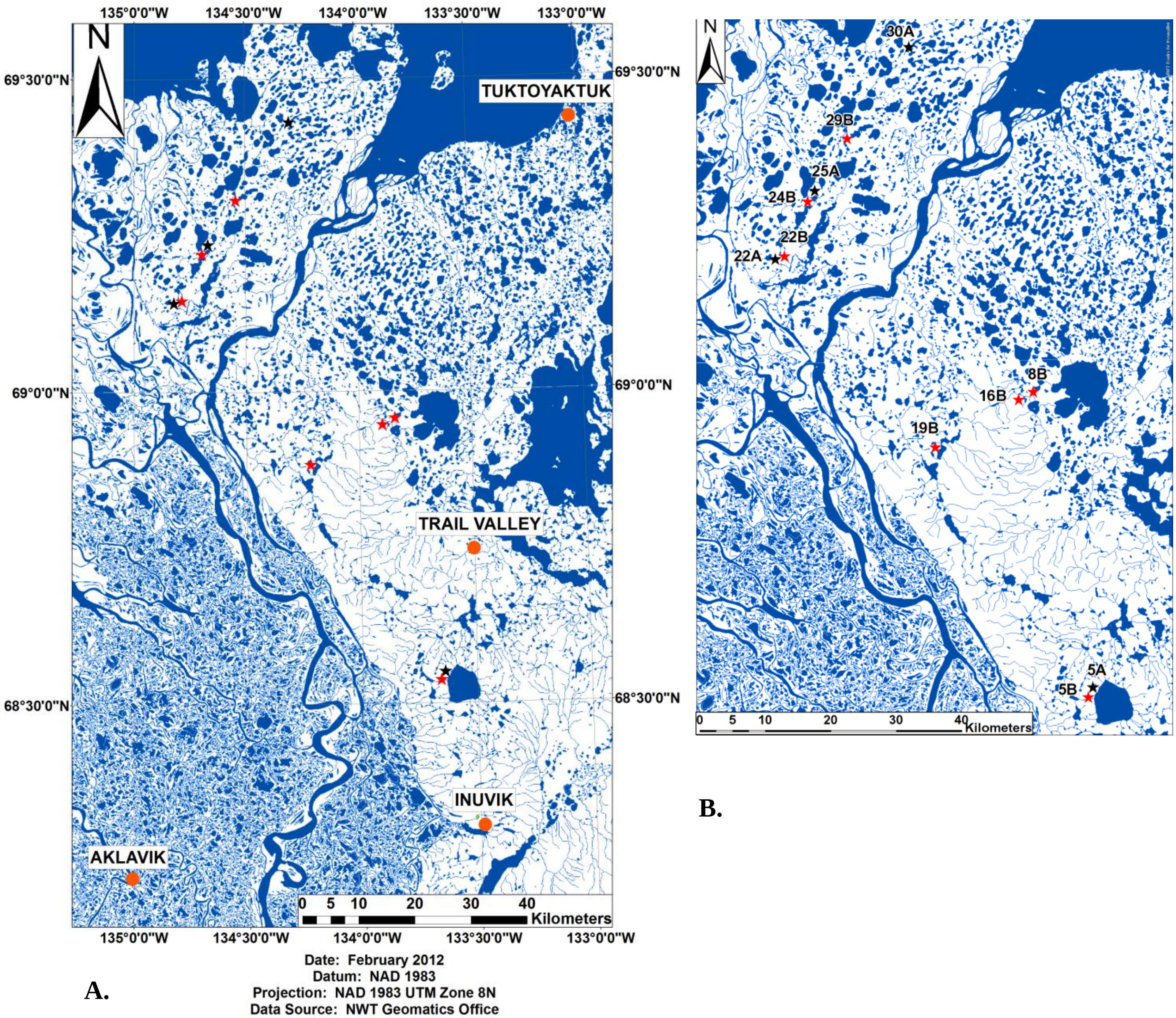


Figure 2.1. A) A map of the study region. The local communities are indicated using orange dots. The study lakes are indicated using stars (Red Stars: Affected by SRTS; Black Stars: Unaffected by SRTS). B) A more detailed map of the study region, indicating the location and name of all the study lakes (Red Stars: Affected by SRTS; Black Stars: Unaffected by SRTS). The data for the maps presented in Figure 2.1 A and Figure 2.1 B were obtained from the NWT Centre for Geomatics, Department of Environment and Natural Resources, Government of the Northwest Territories (2008).

2.2 Study Lakes

The two primary study lakes (5A and 5B) were chosen because they have similar physical characteristics (**Figure 2.2**). Importantly, this pair of lakes are accessible by snowmobile in winter and are only a short helicopter flight away from Inuvik during the summer months, making them logistically easier to study than many of the other lakes in the transect.

Lake 5A (unaffected) is used as a reference lake because it is unaffected by obvious permafrost degradation. Lake 5B (affected by SRTS) is used to assess the potential impact of permafrost degradation on surface runoff and receiving lake water quantity and quality because it has been impacted by obvious permafrost degradation (SRTS).

The 10 regional study lakes (unaffected: 22A, 25A, 30A; affected by SRTS: 8B, 16B, 19B, 22B, 24B, 29B, YaYa sub-catchment lake) are located in both the Mackenzie Uplands and Richards Island. Geochemical signature surveys were carried out at these 10 lakes, to assess how representative the two primary study lakes are of other small tundra lakes in the region.

For details on the physical characteristics of the two primary study lakes and the 12 regional study lakes, refer to **Appendix A**.

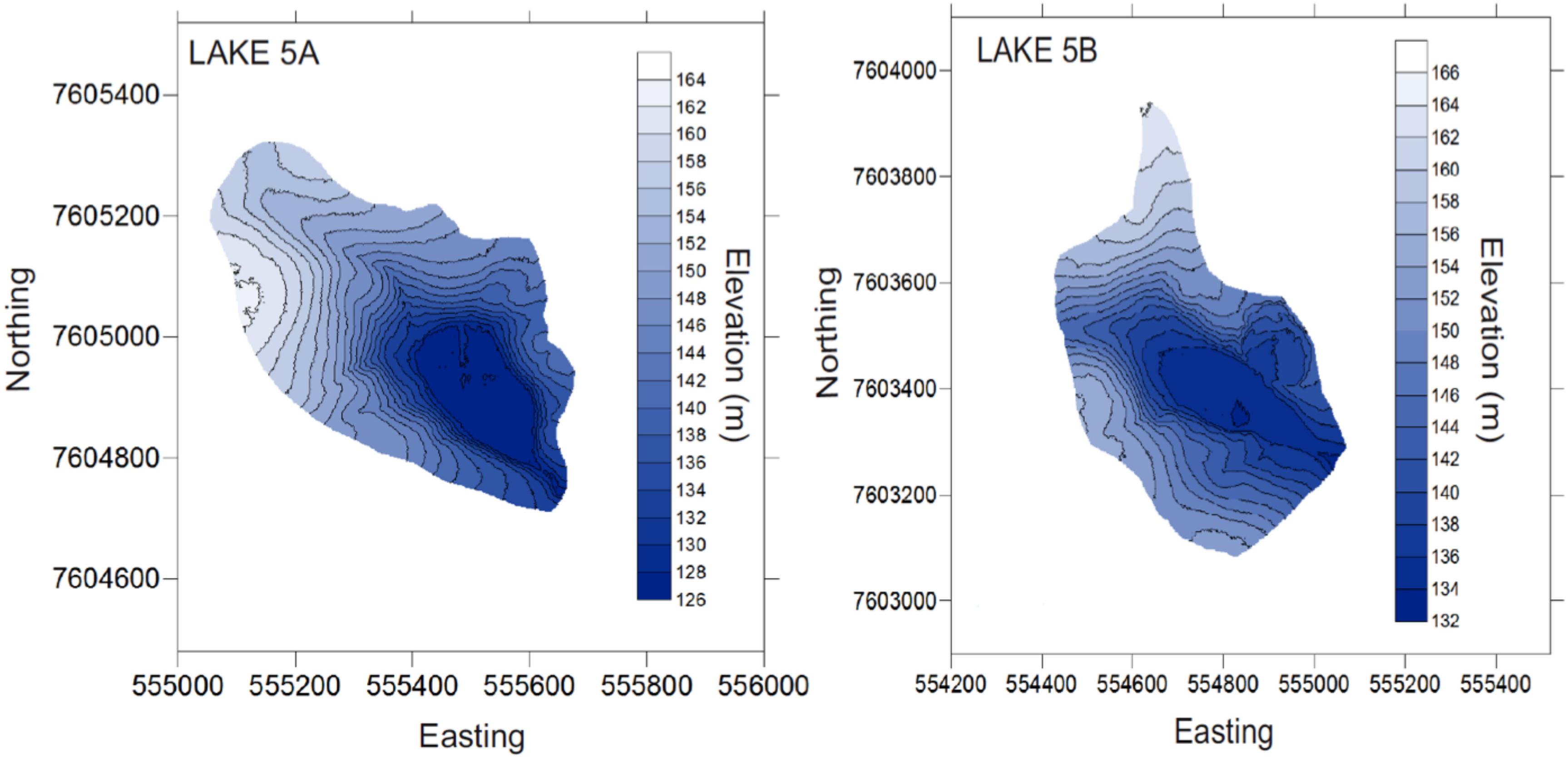


Figure 2.2. A Digital Elevation Model of the Lake 5A and Lake 5B catchments.

2.3 Surficial Geology

The upland region east of the Mackenzie Delta and Richards Island are located on the Arctic Continental Shelf. The surficial geology of this region is predominantly glacial till and aeolian sand (Hivon and Sego, 1993; Dallimore et al., 2000; Burn and Kokelj, 2009). The ground surface is dominated by mineral earth hummocks, which can range from 0.4 to 1 m in diameter and 0.1 to 0.4 m in height above the ground surface (Quinton and Marsh, 1998). The two primary study lakes are located on the northwest shore of Noell Lake, where the surficial geology is classified as hummocky moraine.

2.4 Vegetation

The majority of the upland region east of the Mackenzie Delta is located north of the tree line. The two primary study lakes are located in the forest transition zone (Burn and Kokelj, 2009). Here, vegetation is largely composed of low shrubs, Labrador tea, mosses, lichens, and berries (Quinton and Marsh, 1999). The variability in vegetation northwards, towards Richards Island and the Tuktoyaktuk Coastlands, is controlled by its proximity to both the Arctic Ocean and the Mackenzie River (Dallimore et al., 2000).

2.5 Climate

Arctic regions are typically characterized by extreme seasonal variations in solar radiation, which directly controls the climate in high-latitude regions (Rouse, 1993). According to the Koppen-Geiger classification, the study region is classified as a Polar Tundra Climate, which is characterized by long, cold winters and short, hot summers (Kottek et al., 2006). In December and January, the study region has 30 days without sunlight, leaving the region to be cold and dark (Town of Inuvik, 2010). In May, June, and July, the region has 56 days of 24-hour sunlight (Town of Inuvik, 2010). The contrast in sunlight hours, between winter and summer, results in strong seasonal variability in ambient air temperature and, in turn, strong seasonal variations in precipitation.

The study region experiences strong seasonal and spatial variability in air temperature and precipitation. Inuvik and Tuktoyaktuk are located at the southern and northern ends of the study region. The average mean annual air temperature for Inuvik and Tuktoyaktuk is -8.5 and

-10.2°C, respectively (**Figure 2.3 A**). Based on mean average daily air temperature data (1958 – 2009) for Inuvik and Tuktoyaktuk, the study region typically experiences 7 to 8 months of sub-zero temperatures. Snow and ice formation is normally initiated in late-September, when the average daily air temperature drops to below 0°C. Ambient air temperature is similar for Inuvik and Tuktoyaktuk over the winter months. In mid-May and early-June, the average daily air temperature starts to increase above 0°C, driving snow and ice melt. Inuvik generally experiences warmer spring and summer months than Tuktoyaktuk. As a result, snow and ice melt is generally initiated approximately two weeks earlier in Inuvik. For information regarding the retrieval and infilling of air temperature data see **Appendix C.1** and **Appendix D.1**.

The mean total annual precipitation for Inuvik and Tuktoyaktuk is 307mm and 167mm, respectively (**Figure 2.3 B**). Total monthly precipitation exhibits strong seasonal variability. Based on total monthly precipitation normals (1958 – 2009), the summer months (June, July, August, and September) receive more precipitation than the winter months. August receives the highest amount of precipitation and April receives the lowest amount of precipitation. The study region also experiences strong spatial variability in precipitation. By comparison, Tuktoyaktuk receives notably less precipitation than Inuvik. For more information regarding the retrieval and infilling of precipitation data see **Appendix C.2** and **Appendix D.2**.

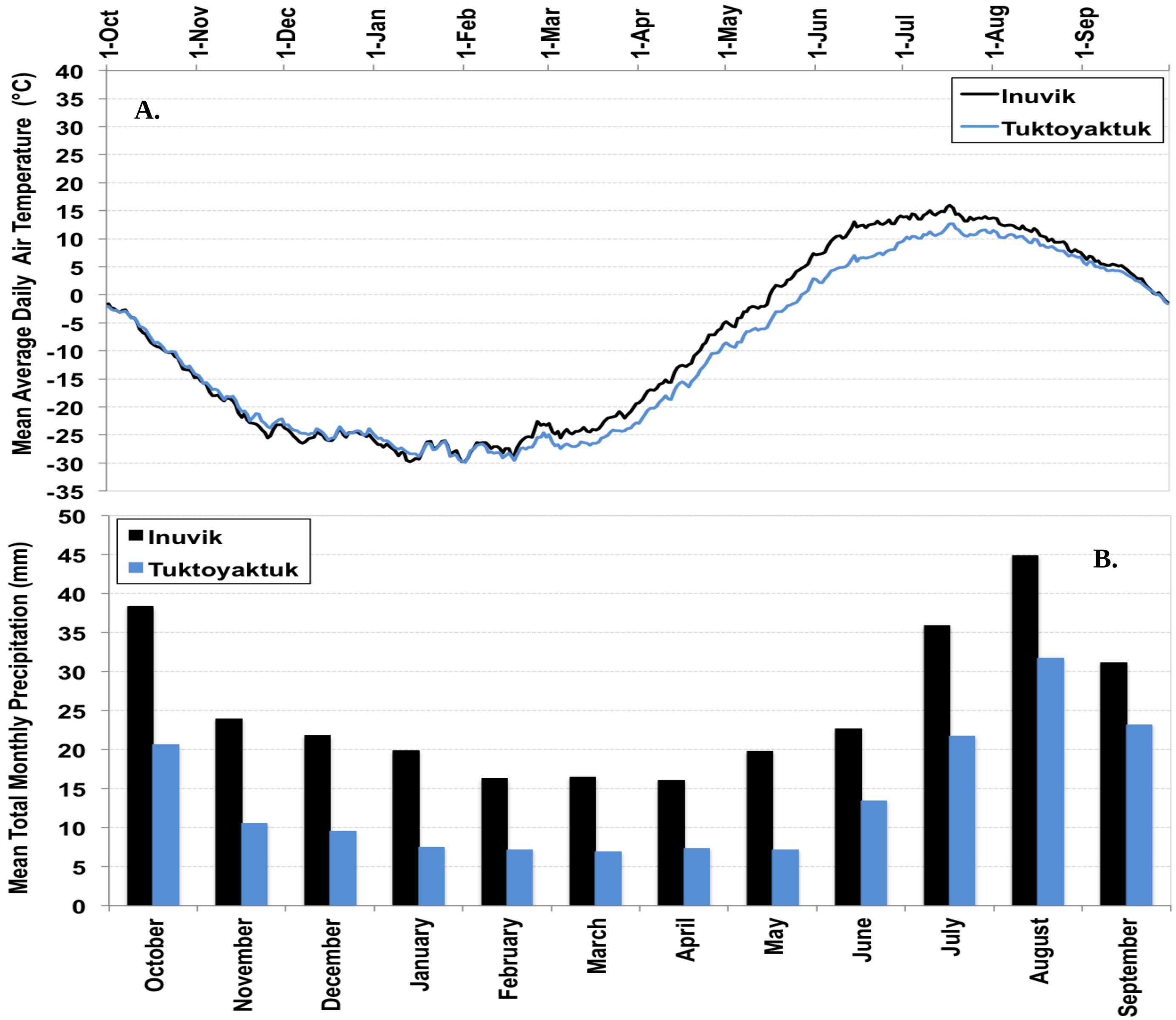


Figure 2.3. A) The mean average daily air temperature for Inuvik and Tuktoyaktuk, NT (1958 to 2009). B) The mean total monthly precipitation for Inuvik and Tuktoyaktuk, NT (1958 to 2009).

2.6 Permafrost

Permafrost thickness is highly variable across the study region. The vertical extent of permafrost is directly controlled by a number of factors, which include ambient air temperature, slope orientation and angle, vegetation, drainage, snow cover, soil cover, rock cover, and water content (French, 2010). The Lower Mackenzie Delta and the upland region east of the Mackenzie Delta are underlain by up to 100m of permafrost (Burn and Kokelj, 2009). The outer Delta and Richard's Island is underlain by 500 to 750m of permafrost (Taylor et al., 1996a; Burn and Kokelj, 2009). Overlying the permafrost is a zone of seasonally unfrozen ground called the active layer. In the upland region east of the Mackenzie Delta, the thickness of the active layer typically ranges from 0.4 to 0.8 m in thickness (Quinton and Marsh, 1999).

Chapter 3: The Effects of Recent Climate Change and Shoreline Retrogressive Thaw Slumping on the Water Balance of Small Tundra Lakes

3.1 Introduction

The hydrology of freshwater systems in the study region is directly controlled by climatic factors (e.g., temperature and precipitation) and landscape-level factors (e.g., near-surface permafrost, mineral earth hummocks), both of which are strongly influenced by climate variability and change. Since the late-1940s, the mean annual air temperature of the Mackenzie Delta region has increased by 3 to 4°C and is projected to increase by an additional 4 to 7°C by 2100 (ACIA, 2005; AMAP, 2012; GNWT, 2008). Recent climate warming has been coupled with a net increase in precipitation across the circumpolar arctic (Manabe and Stouffer, 1994; Rahmstorf and Ganopolski, 1999; Peterson et al., 2002). On average, rainfall and snowfall has increased over arctic land areas by 5% since 1950 (AMAP, 2012). Notably, the effects of recent climate warming on precipitation vary from region to region. Since the 1950s, snow cover depth has decreased in the North American Arctic (AMAP, 2012). In the Eurasian Arctic, however, snow cover depth has increased. Changes in the contributing snowpack have significant implications for small tundra lakes, because spring snowmelt is typically the most significant hydrologic event affecting arctic freshwater systems (Hinzman et al., 1991; Quinton et al., 1999, Quinton et al., 2006). Examining the implications of climate change on key climatic controls of the small tundra lake water balance, with a focus on the upland region northeast of Inuvik, is thus crucial for the development of appropriate freshwater monitoring programs in the upcoming years.

The upland region northeast of Inuvik and Richard's Island has thousands of small tundra lakes. The water balance of these small tundra lakes is controlled by a number of key hydroclimatic factors, including fall freeze-up, snow accumulation, snow damming, spring snow and ice melt, open-water duration, summer rainfall, and evaporation (Pohl et al., 2009; Quinton and Marsh, 1999; Woo, 1980; Woo, 2012). Recent climate warming has led to changes in the timing and magnitude of a number of these key water balance controls. Overall, studies suggest that historical climate change has led to later lake-ice formation, increases in precipitation, decreases in snowcover extent, earlier spring snow and ice melt initiation, and longer open-water periods, across the circumpolar arctic (AMAP, 2012; Bonsal and Prowse, 2003; Dibike et al., 2012; Lesack et al., 2013). In the Inuvik region, in parallel with other regions of the Arctic, winter and spring ambient air temperatures have been increasing, which has led to earlier, more intense spring snow and ice melt periods and an overall decline in snow cover extent (AMAP, 2012; Lesack et al., 2013; Marsh et al., 2002). It is important to note that, in contrast with other regions of the Arctic, there has be no apparent change in the timing of ice formation (Lesack et al., 2013).

In addition to climatic factors, the hydrology of small tundra lakes is also driven by landscape-level factors, which directly control snow distribution and runoff pathways (Pohl and Marsh, 2006; Quinton and Marsh, 1998; Quinton and Marsh, 1999). One example of this is Shoreline Retrogressive Thaw Slumping (SRTS). SRTS could affect the water balance of small tundra lakes in the study region by increasing the accumulation of snow within the contributing lake catchment. Observations made by Kokelj et al. (2009a) and Lantz et al. (2009) indicated that terrain affected by SRTS typically has deeper latent-winter snowpacks than unaffected terrain. Notably, the effect of SRTS on the SWE of the contributing lake catchment is still largely unknown. The modification of the SWE of tundra lake catchments by SRTS is one of several ways recent climate warming could impact the water balance of small tundra lakes in the study region. Further investigation into past hydroclimatic and landscape controls of the water balance of small tundra lakes in the SRTS-prone upland region northeast of Inuvik is needed to gain a better understanding of the potential effects of projected climate warming on arctic freshwater systems.

3.2 Purpose and Objectives

The purpose of this study is to investigate the historical impacts of climate variability and change (CVC) and SRTS on the key hydroclimatic factors and landscape-level features that drive the water balance of tundra lakes in the upland region east of the Mackenzie Delta. The three primary objectives of this chapter are to:

i. Analyze 3 years (2007, 2008, and 2009) of detailed hydrological and climatic field data collected at a pair of representative tundra lakes located in the upland region adjacent to the Mackenzie Delta (Lake 5A: unaffected; Lake 5B: affected by SRTS).

ii. Relate short-term, site specific hydroclimatic field data to historical climate data collected at the nearest Environment Canada weather station, located in Inuvik, NT, with the aim of extending the study period back to the late 1950s.

iii. Examine a suite of relevant hydroclimatic indicators, derived from historical climate data collected at Environment Canada weather stations in Inuvik, located in the southern part of the study region, and Tuktoyaktuk, located in the northern part of the study region.

Furthermore, the results presented here provide crucial hydrological information that will be used in the following chapter to examine how recent climate variability and change have affected the geochemistry of small tundra lakes in the study region.

3.3 Methodology

3.3.1 Field Monitoring

The following work was carried out by Environment Canada in 2007, 2008, and 2009, as part of a larger International Polar Year/ArcticNet project investigating the impacts of climate variability and change on arctic freshwater systems.

3.3.1.1 Lake Water Level

Lake Level (LL; measured in metres) was used to examine key hydroclimatic drivers of the water balance of the two primary study lakes (Lake 5A and Lake 5B). LL was measured continuously, every 30-minutes, using a PT2X and/or Onset Hobo pressure transducer installed at the near shore of each of Lake 5A and Lake 5B. LL was geodetically corrected to metres above sea level (m.a.s.l.) using GPS measurements taken in early spring and late summer. For information regarding the infilling of missing Lake Level data, see **Appendix B.1**.

3.3.1.2 Air Temperature

Air temperature was measured continuously, every 30 minutes, from two tripod towers located on the shoreline of Lake 5A and Lake 5B using a HMP45C Temperature (^{o}C) and Relative Humidity Probe (%) in a Gill shield mounted at 1.5m above the ground surface. For information regarding the infilling of missing air temperature data, see **Appendix B.2**.

3.3.1.3 Water Temperature

Water temperature (^{o}C) was sampled continuously, every 15 minutes, from an array of HOBO Water Temperature Pro Loggers installed at the centers of Lake 5A and Lake 5B from late April (prior to the spring freshet) to late September (prior to freeze-up) of each study year. The water temperature sensors were secured at 0.5m increments from the lake bottom to the lake surface.

3.3.1.4 Precipitation

Snow Water Equivalent

Comprehensive snow surveys were conducted at Lake 5A (undisturbed) and Lake 5B (affected by SRTS) in late-April of each study year, in order to estimate the maximum annual snow water equivalent (SWE) of the contributing lake catchment. At each lake catchment, a snow survey transect was conducted along each major slope, leading from the centre of the lake to the top of the slope (**Figure 3.1**). At Lake 5B, two additional transects were set up within the shoreline slump. The first transect ran from the mouth of the slump, at the lake, to the top of the headwall of the slump. The second transect ran through the centre of the slump, lying perpendicular to the first transect, leading from the top of the Northwest headwall to the top of the Southeast headwall.

A snow core spanning the depth of the snowpack was collected approximately every 25 m using a Metric Prairie Snow Sampler (Environment Canada ESC 30 Design). If the snowpack was deeper than 1.25m, then SWE was measured using a Federal Snow Sampler (US Forest Service Design). The SWE of the snow core was measured using a spring scale graduated in cm of water.

The density of the snow core was determined using **Equation 3.1**.

Equation 3.1: Snow Water Equivalent

$$SWE = \frac{\rho_s \times d_s}{\rho_w}$$

where SWE is in m, ρ_s is the density of snow (kg m^{-3}), d_s is the depth of snow (m), and ρ_w is the density of water (kg m^{-3}) (Woo, 2012). Between each SWE measurement, snow depth (d_s) was measured approximately every 5 m using a snow depth probe. SWE was then estimated for every 5 paces using d_s and the average of the two ρ_s measurements the d_s measurement fell between.

Each SWE value was classified by slope aspect (north slope, west slope, south slope, and slump), elevation (Lake 5A: upper (>145 masl) and lower (<145 masl); Lake 5B: upper (>144 masl) and lower (<144 masl)), and the presence or absence of shoreline retrogressive thaw slumping (unaffected, slump centre, slump headwall). The delineated Hydrological Runoff Units (HRU) are shown in **Figure 3.2**.

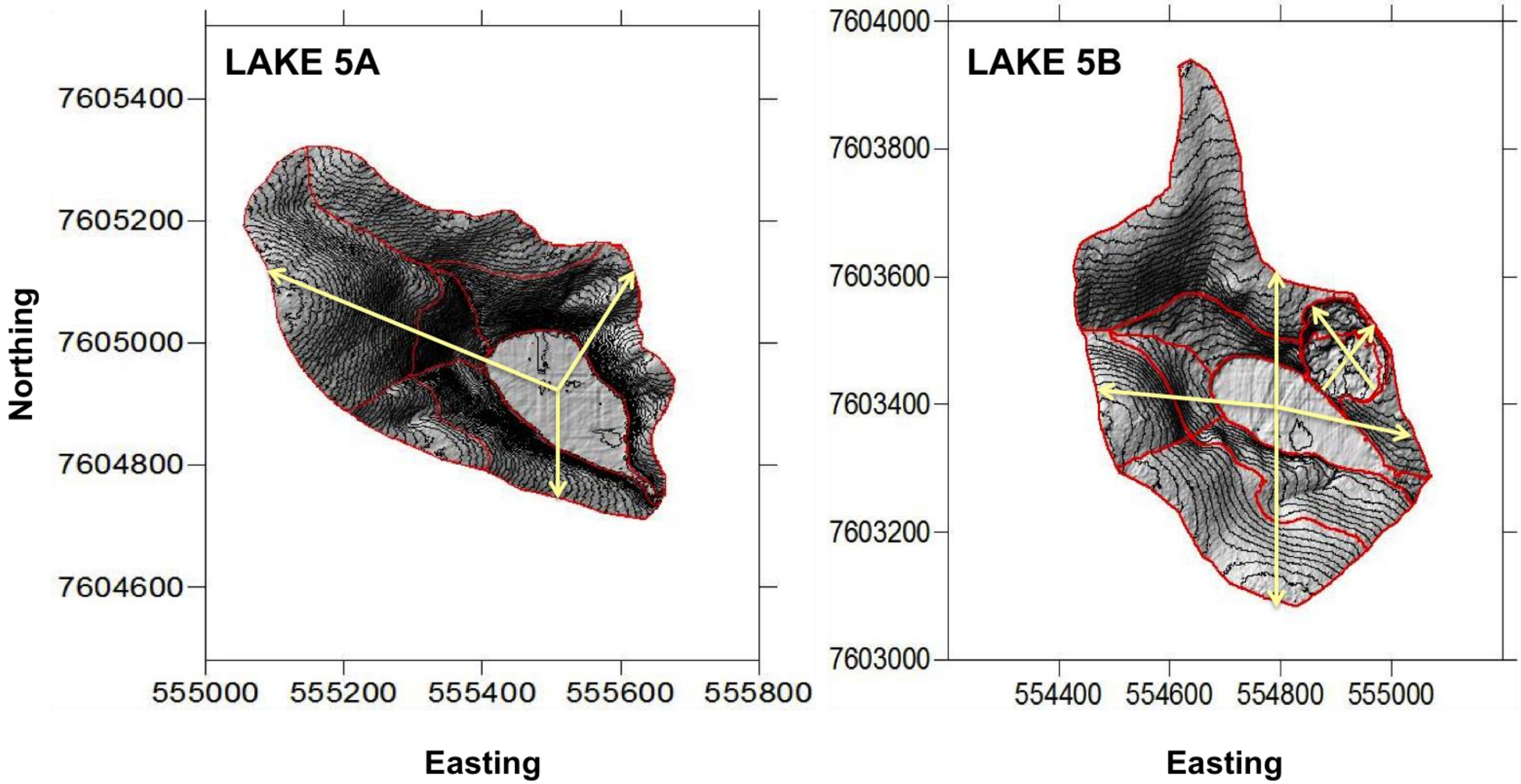


Figure 3.1. A digital elevation models of Lake 5A and Lake 5B. The black lines are contour lines at 2m intervals. Each area outlined in red represents a hydrological run-off unit, which is defined by slope aspect, elevation, and the presence or absence of shoreline retrogressive thaw slumping. The yellow arrows indicate the approximate location of each snow survey transect.

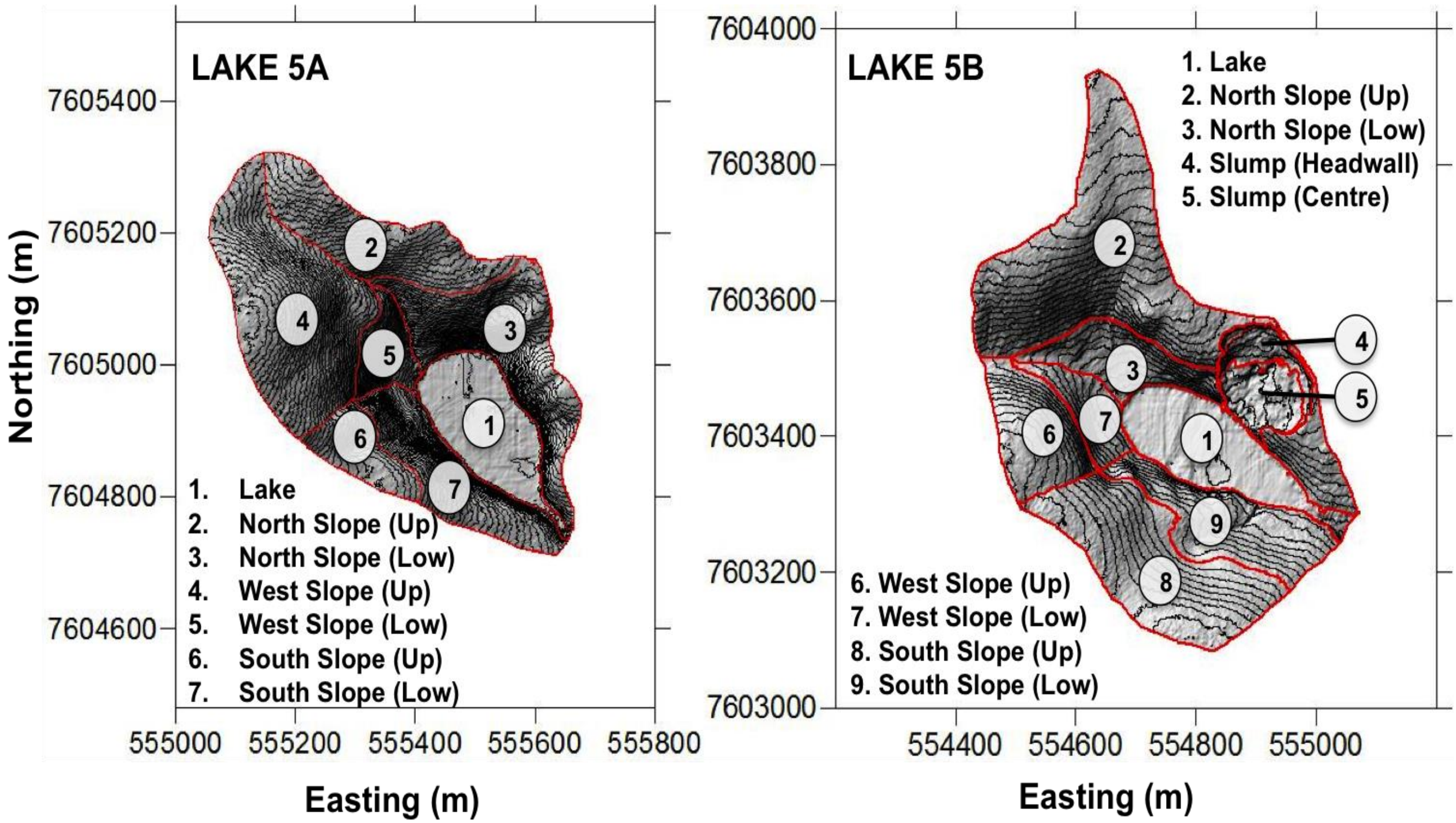


Figure 3. 2. A digital elevation model of the Lake 5A and Lake 5B catchments. The black lines are contour lines at 2m intervals. The areas outlined in red represent the Hydrological Run-off Units (HRU), which are defined by slope aspect, elevation, and the presence or absence of shoreline retrogressive thaw slumping. These HRUs were used to calculate the average weighted catchment SWE of Lake 5A and Lake 5B.

A digital elevation model of the lake catchment, loaded into Surfer 8 software (Golden Software Inc., 2002), was used to determine the surface area of the total lake catchment and each HRU. The fraction of the total lake catchment area occupied by each HRU is presented below in **Table 3.1**.

Table 3.1. The Hydrological Runoff Units (HRU), defined by slope aspect, elevation, and the presence or absence of shoreline retrogressive thaw slumping, for Lake 5A and Lake 5B. Each land cover type is listed here, along with the fraction of the total lake catchment that it occupies.

Hydrological Runoff Unit	Lake 5A Fraction of Catchment Area	Lake 5B Fraction of Catchment Area
North Slope (Upper)	0.18	0.33
North Slope (Lower)	0.22	0.12
West Slope (Upper)	0.32	0.10
West Slope (Lower)	0.05	0.05
South Slope (Upper)	0.07	0.20
South Slope (Lower)	0.16	0.13
Slump (Headwall)		0.03
Slump (Centre)		0.05

The average SWE for each HRU was multiplied by the fraction of the total lake catchment area it occupied. For each lake catchment, the weighted average SWE values for each HRU were summed, yielding a weighted average SWE for the contributing lake catchment.

An additional weighted average catchment SWE was estimated for the Lake 5B catchment to estimate catchment SWE without SRTS. To calculate the second weighted average catchment SWE for Lake 5B, the SWE values for the slumped area were replaced with SWE values from the adjacent unaffected terrain. The true weighted average catchment SWE for Lake 5B and the hypothetical weighted average catchment SWE were compared to assess the impact of SRTS on snowmelt contribution to Lake 5B.

Summer Rainfall

Rainfall was sampled continuously, every 15 minutes, for most of the open-water period from tripod towers located on the shoreline of Lake 5A and Lake 5B using a TR525USW Tipping Bucket Rain Gauge installed at 1.5m above the ground surface. For information regarding the infilling of missing rainfall data, see **Appendix B.3**.

3.3.1.5 Open-water Duration

The date of Ice-off was set as the first day when the temperature of the lake surface water, measured at 0.1m below the lake surface, increased to above 4°C. The date of Ice-on was set as the first day Freezing Degree Day (FDD) was equal to or less than -20°C. Marsh and Lesack (1993) validated this index for the Mackenzie Delta region. FDD were calculated by summing average daily air temperature at Lake 5A, starting on the first day ambient air temperature decreased to below 0°C. If FDD increased to above 0°C, it was assigned a value of 0.

3.3.1.6 Evaporation

A number of studies have successfully used the Priestley-Taylor model to estimate evaporation from small northern lakes. Some examples include Stewart and Rouse (1976), Roulet and Woo (1986), and Marsh and Bigras (1988). The Priestley-Taylor model for evaporation (E_{PT}) was used in this study to estimate evaporation from Lake 5A in 2007, 2008, and 2009. To this end, detailed micrometeorological data (i.e., net radiation, air temperature, and relative humidity) were collected every 15 minutes from a tripod tower set-up in the water at the near shore of control Lake 5A. E_{PT} (measured in m day^{-1}) is described in the following equation.

Equation 3.2: The Priestley-Taylor model for evaporation

$$E_{PT} = \frac{(\alpha \times s)}{(s \times \gamma)} \frac{(R - G)}{(\rho_w \times \lambda_v)}$$

where s (kPa °C^{-1}) is the slope of the saturation-vapour pressure versus air temperature curve, γ (kPa °C) is the psychrometric constant, R (W m^{-2} d^{-1}) is net radiation, G (W m^{-2} d^{-1}) is subsurface heat flux, ρ_W (Kg m^{-3}) is the density of water, and λ_V (MJ Kg^{-1}) is the latent heat of vaporization (Priestley and Taylor, 1972). α is an empirically derived evaporability factor, which was given a value of 1.26 in the study. Note that Stewart and Rouse (1976) found that α equals 1.26 for saturated surfaces. Furthermore, Marsh and Bigras (1988) assigned α a value of 1.26 when they used E_{PT} to estimate evaporation from small lakes in the Mackenzie Delta region.

G is typically defined as the difference between the change in the storage of heat by the water column (ΔG) and conduction of heat into the lake bed (Priestley and Taylor, 1972). Studies suggest that heat conduction into the lake bed is generally negligible for thermokarst lakes (Marsh and Bigras, 1988). As such, it is excluded from this analysis. In this study, G is assumed to equal ΔG and was estimated using the following equation:

Equation 3.3: Heat Storage by the water column

$$G = \Delta G = C \frac{\Delta T_W}{\Delta t} z$$

where ΔG (W m^{-2}) is the change in heat stored in the water column, C (J m^{-3} °$K^{+ -1}$) is the volumetric heat capacity of water, ΔT_W (°K) is the change in water temperature, Δt (s) is the change time, and z (m) is the depth of water. Water temperature data indicated that both Lake 5A and Lake 5B stratify during the summer months. The average water temperature of the water column was defined using the following equation:

Equation 3.4: The average temperature of the water column.

$$T_w = \sum_{i=1}^{n} T_{w,i} \; x \; \frac{V_i}{V}$$

where $T_{w,i}$ (°C) is the temperature of the given layer, V_i (m^3) is the volume of the given layer, and V (m^3) is the total volume of the lake. The volume of each layer and the total volume of the lake was determined using bathymetry data.

3.3.1.7 Digital Elevation and Bathymetry Models

Detailed elevation information for the two study lake catchments was obtained using LiDAR aerial surveys collected by Environment Canada in the summer of 2008. Bathymetric surveys of each of the study lakes were conducted using geo-referenced depth soundings. Elevation and depth data points were imported into Surfer 8 (Golden Software Inc., 2002) software package to produce a digital elevation model of the surface topography and bathymetric map of the lake bottom for the Lake 5A and Lake 5B.

3.3.2 Key Hydroclimatic Indices

3.3.2.1 Air Temperature

Adjusted and homogenized mean, maximum, and minimum daily air temperature data for Inuvik was obtained from Environment Canada's Adjusted and Homogenized Canadian Climate Data archive for the years 1958 to 2008 (Environment Canada, 2008). Supplementary air temperature data was obtained from Environment Canada's Historical Climate Data archive for the years 1958 to 2009 (Environment Canada, 2009). Historical air temperature data obtained from Inuvik was used to develop a suite of hydroclimatic indices (i.e., mean annual air temperature, spring freshet initiation, open-water duration, and evaporation) to examine how historical climate variability and change has affected the water balance of the two primary study lakes. Out of the three climate stations closest to Lake 5A (Trail Valley Creek, Inuvik, and Tuktoyaktuk), the climate station in Inuvik was chosen because the ambient air temperature conditions are the most representative of that of the two primary study lakes. For average daily temperature relationships between Lake 5A and the three closest climate stations, see **Appendix B.2**. For information regarding the retrieval and infilling of air temperature data for Inuvik, see **Appendix C.1**.

Adjusted and homogenized air temperature data was not available for Tuktoyaktuk. Unadjusted mean, maximum, and minimum daily air temperature data for Tuktoyaktuk was obtained from Environment Canada's Historical Climate Data archive for the years 1958 to 2009. Historical air temperature data obtained from Tuktoyaktuk was used to develop the same suite of hydroclimatic indices to examine how historical climate variability and change has affected the water balance of small tundra lakes at the northern end of the study region. For information regarding the retrieval and infilling of air temperature data for Tuktoyaktuk, see **Appendix D.1**.

3.3.2.2 Precipitation

Total Precipitation

Adjusted total monthly precipitation data for Inuvik and Tuktoyaktuk was obtained from Environment Canada's Adjusted and Homogenized Canadian Climate Data archive for the years 1958 to 2009. Historical precipitation data was used to develop a suite of hydroclimatic indices (i.e., Total Annual Precipitation, Annual Snowpack Index, Annual Rainfall Index, and Vertical Water Balance) to examine how historical climate variability and change has affected the water balance of the two primary study lakes, as well as small tundra lakes at the northern end of the study region. For information regarding the retrieval and infilling of total monthly precipitation data, see **Appendix C.2** and **Appendix D.2**.

Annual Snowpack Index

The annual snowpack index (mm), defined here as the total precipitation that fell between October 1st and May 1st, is typically a good indicator of the total amount of snow (in mm of water) that has fallen over the course of the winter months (e.g., Romolo et al., 2006). The annual snowpack index for Inuvik was determined for the years 1958 to 2009 to examine how historical climate variability and change has affected the SWE of the contributing catchment at Lake 5A and Lake 5B. The annual snowpack index for Inuvik was validated using field data collected at the two primary study lakes in 2007, 2008, and 2009 (**Appendix E.2**). The annual snowpack index for Tuktoyaktuk was determined for the years 1958 to 2009 to examine how historical climate variability and change has affected the SWE of the contributing catchment at the regional study lakes located further north along the study transect.

Annual Rainfall Index

The annual rainfall index (mm), defined here as the total precipitation that fell from June 1st and September 30th, is a good estimate of the amount of rainfall that falls during the summer months (e.g., Romolo et al., 2006). The annual rainfall index for Inuvik was determined for the years 1958 to 2009 to examine how historical climate variability and change has affected summer rainfall at Lake 5A and Lake 5B. The annual rainfall index for Inuvik was validated using field data collected at the two primary study lakes in 2007, 2008, and 2009, along with supplemental data for 2006 that was made available by Environment Canada (see **Appendix E.2**). The annual rainfall index for Tuktoyaktuk was determined for the years 1958 to 2009, in order to examine how historical climate variability and change has affected summer rainfall at the regional study lakes located further north within the study region.

3.3.2.3 Spring Freshet Initiation

The Spring Freshet refers to the rise in stream, river, and lake water level that occurs when snow and ice begins to thaw in early spring. Burn (2008) used river discharge data collected at various gauging stations along the Mackenzie River to examine long-term trends in the timing of the Spring Freshet within the Mackenzie River basin. He defined the Spring Freshet as the first day river discharge was 1.5 times greater than the average river discharge from the previous 16 days.

The method outlined by Burn (2008) was used to estimate the timing of the Spring Freshet at Lake 5A and Lake 5B using stream discharge data obtained by the Water Survey of Canada from nearby Trail Valley Creek for the years 2007 to 2009. The onset of the spring freshet at Lake 5A and Lake 5B, determined using stream discharge data, was used to validate the following two common temperature-based methods of estimating spring freshet initiation:

I. *Method 1*: Outlined by Bonsal and Prowse (2003), this method defines the onset of the spring freshet as the first day the 0° Isotherm is equal to or greater than 0°C. The 0° Isotherm is a 31-day running mean of average daily air temperature.

II. *Method 2*: Outlined by Pohl (Personal Communication, 2011), this method defines the onset of the spring freshet as the first day the average daily air temperature increased to above 0°C and stayed above 0°C.

Overall, *Method 2* provided the best temperature-based estimate of the timing of the spring freshet at Lake 5A and Lake 5B (**Appendix E.3**).

For the years 1958 to 2009, the timing of the spring freshet in Inuvik was defined as the first day average daily air temperature increased to above 0°C and stayed above 0°C. For the three primary study years, the timing of the spring freshet at Inuvik was a good indicator of the timing of the spring freshet at Lake 5A and Lake 5B (**Appendix E.3**). Historical trends in the timing of the spring freshet in Inuvik were used to examine how historical climate variability and change has affected the timing of the spring freshet at the two primary study lakes. For the years 1958 to 2009, the timing of the spring freshet at Tuktoyaktuk was estimated using the same method. Historical trends in the timing of the spring freshet in Tuktoyaktuk were used to examine how historical climate variability and change has affected the timing of the spring freshet at the regional study lakes located further north along the study transect.

3.3.2.4 Open-water Duration

In the absence of detailed physical measurements, Positive Degree Days (PDD), or the accumulation of days with an average daily air temperature above 0°C, can be used as an index of annual ice-off dates (Ashton, 1986). In this study, PDD was calculated by adding the average daily air temperature (°C) of a given day to the PDD value of the previous day, starting on October 1st of the given year. If PDD dropped below 0°C, it was assigned a value of 0. The timing of ice-off was set as the first day that the PDD for Inuvik was equal to or greater than 223°C. This value was determined using field data (**Appendix E.4**). It is the average PDD associated with the actual date of ice-off at Lake 5A and Lake 5B over the three primary study years.

The timing of ice-off in Inuvik was estimated for the years 1958 to 2009 to examine how historical climate variability and change has affected the timing of ice-off at Lake 5A and Lake 5B. The timing of ice-off in Tuktoyaktuk was also determined for the years 1958 to 2009 to examine how historical climate variability and change has affected the timing of ice-off at the regional study lakes located further north along the study transect.

The date of Ice-on was set as the first day Freezing Degree Day (FDD) was equal to or less than -20°C. FDD were calculated by summing average daily air temperature at Inuvik, starting on the first day ambient air temperature decreased to below 0°C. If FDD increased to above 0°C, it was assigned a value of 0. The timing of ice-on in Inuvik was estimated for the years 1958 to 2009. The timing of ice-on in Inuvik is a good indicator of the timing of ice-on at Lake 5A and Lake 5B (**Appendix E.4**) and was used to examine how historical climate variability and change has affected the timing of ice-on at the two primary study lakes. The timing of ice-on in Tuktoyaktuk was estimated for the years 1958 to 2009 to examine how historical climate variability and change has affected the timing of ice-on at lakes located further north along the study transect.

Open-water duration (days) in Inuvik and Tuktoyaktuk, defined as the difference between the date of *ice-off* and the date of *ice-on*, was estimated for the years 1958 to 2009.

3.3.2.5 Evaporation

The Hargreaves and Samani (1982) model for estimating evaporation (E_{HG}) is often used in hydrological studies when hydro-climatic data is limited (e.g., Jing et al., 2010; Sentelhas et al., 2010). This method was used in this study to estimate total daily evaporation from Inuvik and Tuktoyaktuk. Hargreaves model for evaporation is as follows:

Equation 3.5: Hargreaves model of evaporation

$$E_{HG} = 0.0023\ (T_{avg} + 17.8)\ (T_{max} - T_{min})^{0.5}\ R_a$$

E_{HG} is total daily evaporation (mm d^{-1}), T_{avg} is mean daily air temperature (°C), T_{max} is maximum daily air temperature (°C), T_{min} is minimum daily air temperature (°C), and R_a is extraterrestrial radiation (MJ m^{-2} d^{-1}). R_a was determined using the following equation:

Equation 3.6: Extraterrestrial Radiation

$$Ra = 15.392 \text{ x } d_r \text{ x } [w_s \text{ x } SIN(\Phi) \text{ x } SIN(\delta) + COS(\Phi) \text{ x } COS(\delta) \text{ x } SIN(w_s)]$$

where d_r is the relative distance between the earth and the sun, ws (radians) is the sunset hour angle, Φ (radians) is latitude, and δ (radians) is solar declination. dr was derived using the following equation:

Equation 3.7: The relative distance between the earth and the sun

$$dr = 1 + 0.033 \text{ x } COS[(2 \text{ x } \Pi)/(JD \text{ x } 365)]$$

where JD is Julian Day. W_s was determined using the following equation:

Equation 3.8: Sunset Hour Angle

$$w_s = \Pi/2 - ATAN[(-1xTAN(\Phi) \times TAN(\delta)) / X^{0.5}]$$

Finally, X was derived using the following equation:

Equation 3.9: Empirically derived constant X

$$X = 1 - [((TAN(\Phi))^2] \times [(TAN(\delta))^2]$$

The E_{HG} for Inuvik was validated using field data collected at Lake 5A and Lake 5B in 2007, 2008, and 2009, with supplementary data available for 2006 (**Appendix E.5**).

Total annual evaporation in Inuvik was estimated for the years 1958 to 2009 by summing the daily E_{HG} values from June 17th to September 30. June 17th was chosen because it is the average date of ice-off in Inuvik for the historical study period (1958 to 2009). Total annual evaporation in Tuktoyaktuk was estimated for the years 1958 to 2009 by summing the E_{HG} values from July 4th to September 30. July 4th was chosen because it is the average date of ice-off in Tuktoyaktuk for the historical study period (1958 to 2009).

3.3.2.6 Vertical Water Balance

The vertical water balance (P-E; measured in mm) for Inuvik and Tuktoyaktuk was estimated for the years 1958 to 2009 by subtracting total annual evaporation from total annual precipitation.

3.4 Statistical Analysis

3.4.1 Data Infilling

The primary weather stations used in this study were: Lake 5A, Inuvik, and Tuktoyaktuk. Missing air temperature data were infilled using data retrieved from one of the three weather stations closest to the primary weather station. An Independent-Samples T-Test was used to test whether or not there was a significant difference between the primary dataset and that of the closest weather station with available data ($\alpha = 0.05$). The datasets were compared, based on month, for the years the two datasets overlapped. If there was no significant difference between the two weather stations, then they were combined (L.A. Vincent, Personal communication, 2011). If there was a significant difference between the primary dataset and that of the closest weather station, the following analyses had to be performed.

Linear regression models between the primary dataset (the dependent variable) and the closest weather station with available data (the independent variable) were formed, based on month, for the years the two datasets overlapped. Depending on the month the missing data fell within, one of the linear regression models was used to infill the missing data. These tests were performed using SPSS Version 21 (IBM Corp., 2012).

3.4.2 Validation of Key Hydroclimatic Indices

In order to extend the study period back to the 1950's, a number of key hydroclimatic indices were examined (mean annual air temperature, total annual precipitation, annual snowpack index, annual rainfall index, spring freshet initiation, ice-off timing, ice-on timing, open-water duration, and evaporation). These indices were validated using field data collected at the two primary study lakes. An Independent-Samples T-Test was used to test whether or not there was a significant difference between the field data and the corresponding hydroclimatic index ($\alpha = 0.05$). These tests were performed using SPSS (Version 21).

3.4.3 Time Series Analysis

The Mann-Kendall (M-K) test is a non-parametric test that is often used in hydrological and climatological studies to detect trends in time series data (e.g., Gilbert, 1987; Connon et al., 2013; Kang et al., 2014; Lesack et al., 2013; Peters et al., 2013; Tondu et al., 2013). In this study, the M-K test was used to discern the presence of significant ($\alpha = 0.05$) trends in the hydroclimatic indices calculated for Inuvik and Tuktoyaktuk (i.e., air temperature, total annual precipitation, ice-on, annual snowpack index, spring freshet initiation, ice-off, open-water duration, annual rainfall index, evaporation, and the vertical water balance). Sen's method was then used to determine the slope of the trends identified using the M-K test (Gilbert, 1987).

Prewhitening is often performed on time series data to remove serial correlation. However, recent studies have noted that prewhitening is unnecessary and can increase the error of Sen's Slope Estimate (Fleming and Clarke, 2002; Peters et al., 2013; Yue and Wang, 2002). For this reason, all analyses were performed on non-prewhitened data using the MAKESENS excel template application (Salmi et al., 2002).

3.5 Results and Discussion

3.5.1 Lake Level

The water level at Lake 5A and Lake 5B exhibited strong seasonal variability, which was driven by a number of key hydrological processes (**Figure 3.3**). Over the winter months, defined here as the period of time falling between the first day that ice forms on the lake and the onset of spring snow and ice melt, there was generally a small increase in Lake Level (LL). At sub-zero temperatures, water balance inputs are typically stored within the contributing lake catchment in the form of snow and ice, and changes in LL are typically associated with either changes in atmospheric pressure or the weight of snow and ice on top of the lake, rather than runoff and precipitation (Pohl et al., 2009; Quinton and Marsh, 1999; Woo, 2012).

At the onset of spring snowmelt, there was a rapid increase in LL at both lakes for all study years. Between early-May and the annual spring freshet peak, the LL at Lake 5A increased by 0.35m, 0.45m and 0.23m for 2007, 2008 and 2009, respectively. Overall, spring snowmelt resulted in the greatest change in LL over the hydrological year. For the latter two years, the LL of Lake 5B increased by 0.47m and 0.49m, respectively. For all three study years, the annual LL peaked during the spring snowmelt period. These observations agree with other hydrological studies, which showed that the spring freshet was typically the most significant hydrological event acting on arctic freshwater systems (Pohl et al., 2009; Quinton and Marsh, 1999).

Lake drainage is typically initiated when LL reaches the elevation of the outflow channel (Woo, 2012). The rapid rise in LL associated with the spring snowmelt period was facilitated by a process known as “snow damming” (Woo, 2012). Narrow stream valleys, much like the outflow channels draining Lake 5A and Lake 5B, typically act as snow traps, forming deep winter snow drifts over the winter months (Kane et al., 1991; Pohl and Marsh, 2006). Woo (1980) found that thick snow drifts, which would form in the outlet channel draining Small Lake, near Resolute, acted as a temporary dam, preventing lake drainage in early spring (Woo, 2012). Similarly, Kane et al. (1991) found that snow drifts, common within the stream channels at Imnavait Creek, Alaska, can delay streamflow by 1 to 3 days. The results presented in **Figure 3.3** indicate that, in addition to snowmelt driven runoff, snow damming influenced the LL of Lake

5A and Lake 5B in early spring. Field observations also indicate that snow damming influenced the LL of Lake 5A and Lake 5B in the early spring of the three primary study years. Supplementary 2009 water level observations, obtained from the inflow and outflow channels at Lake 5A and Lake 5B, support the presence of snow damming (**Figure 3.4** and **Figure 3.5**). In 2009, the water balance of Lake 5B was positive from May 17^{th} to May 27^{th}, leading to a notable increase in LL. In other words, the water balance inputs (precipitation and runoff) were greater than the water balance outputs (evaporation and lake drainage). Runoff via the inflow channels peaked during this time period. Conversely, precipitation was minimal, indicating that the observed increase in LL was driven by snowmelt runoff. The progression of spring snowmelt at the inflow channel at Lake 5A is documented in **Figure 3.6**. Despite a substantial increase in LL, the outflow channels draining Lake 5A and Lake 5B were not flowing until May 30^{th} and May 28^{th}, respectively. Correspondingly, lake drainage wasn't initiated until 13 and 11 days after spring snowmelt began, respectively. Another indicator of the presence of snow damming was the rapid drop in LL that occurred following the annual spring freshet peak in 2009, as well as the two other study years.

For all three study years, LL continued to increase until late-May and early-June, then rapidly declined when lake drainage was initiated. On May 28^{th}, 2009, the water from Lake 5B had carved a channel through the snowdrift damming the outflow channel (**Figure 3.8**). Water that had pooled in the lake flooded through the outflow channel, causing LL to decrease. The LL of Lake 5A and Lake 5B decreased from May 28^{th} to August 2^{nd} due to lake drainage and evaporation. The decline in LL from May 28^{th} to June 16^{th} was driven by lake drainage. On June 17^{th}, when Lake 5A and Lake 5B became ice free, the decline in LL was driven by lake drainage and evaporation. On July 13^{th}, the water level of the outflow channel reached 0, indicating that lake drainage had ceased and that the decline in LL was driven by evaporation. LL reached an annual minimum on August 1^{st}, 2009 (Lake 5A = 127.88 m.a.s.l; Lake 5B = 134.29 m.a.s.l).

In 2008 and 2009, LL increased throughout August and September in response to summer rainfall events and runoff. In 2009, there were three major rainfall events. On August 1^{st}, the water level of the primary inflow channel to Lake 5A and Lake 5B increased from 0.09m to 0.20m and 0.01m to 0.13m, respectively, in response to 26mm of rainfall. This led to a 0.1m

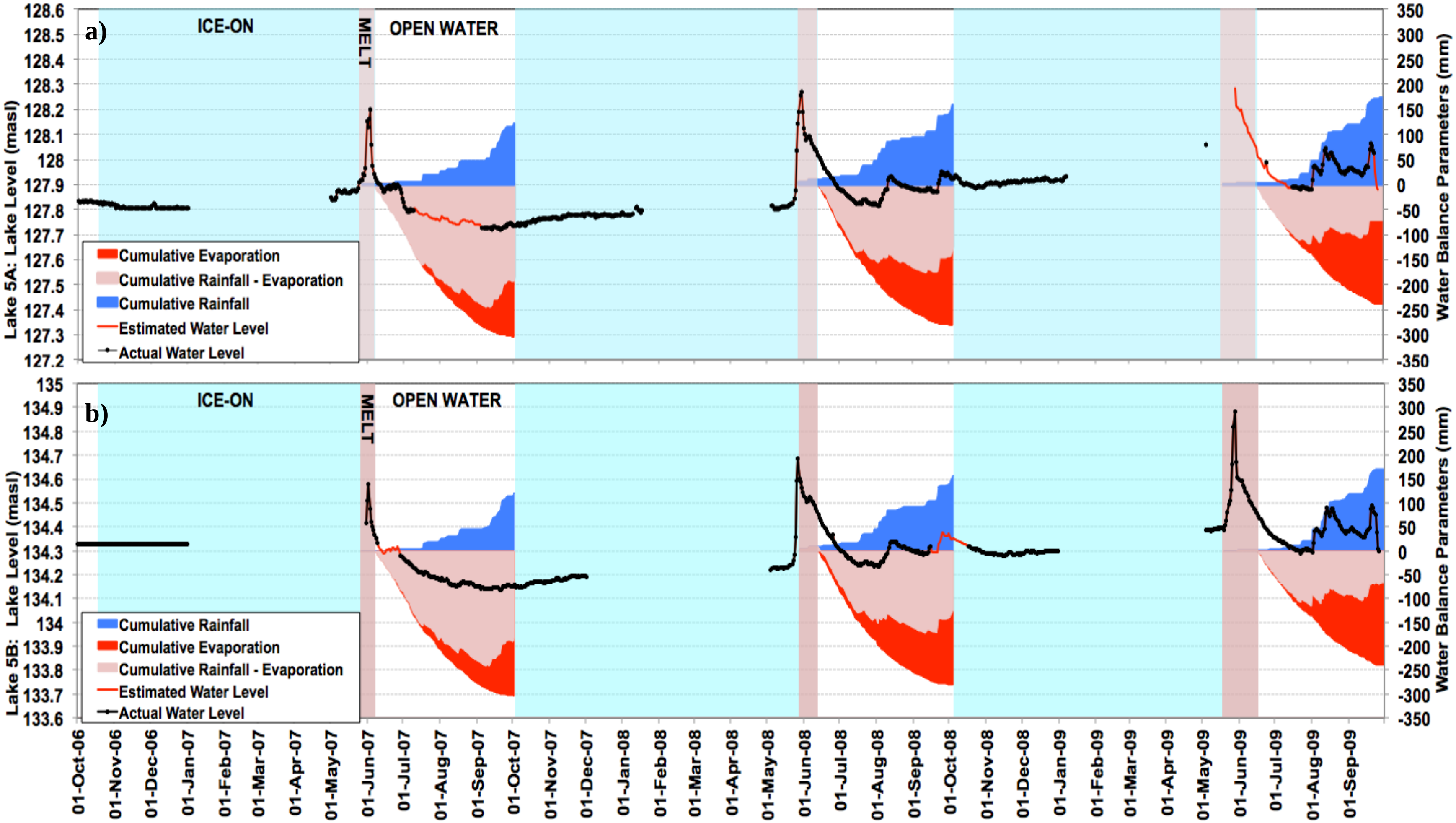


Figure 3.3. The Lake Level observed at a) Lake 5A and b) Lake 5B over the course of the 2007, 2008, and 2009 study years. Also presented on the right y-axis are three key water balance parameters: cumulative rainfall, cumulative evaporation, and cumulative rainfall minus evaporation measured at both study lakes.

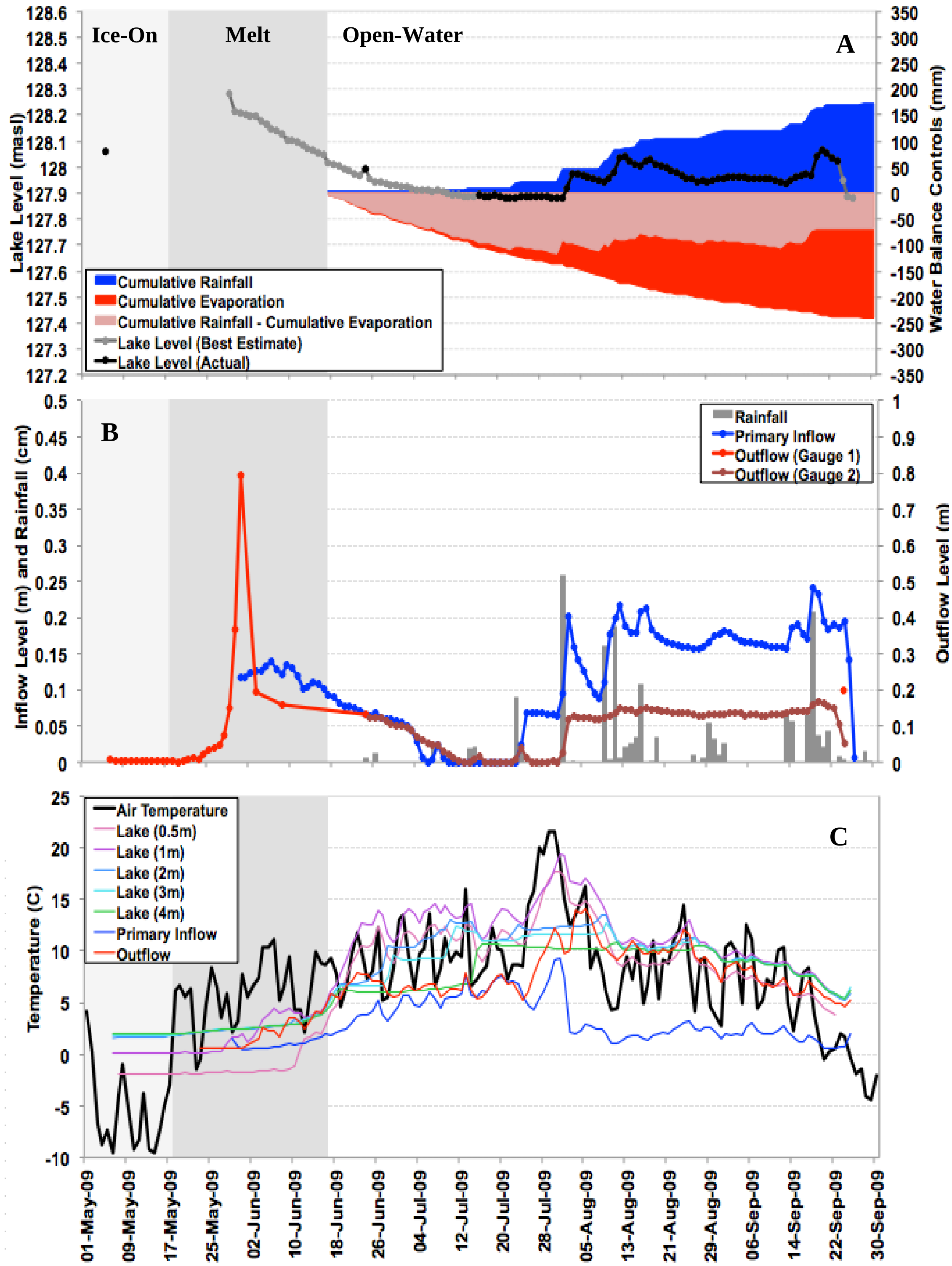


Figure 3.4. A) The Lake Level of Lake 5A from May 1st to September 30th, 2009, plotted with Cumulative Rainfall, Cumulative Evaporation, and Cumulative Rainfall minus Cumulative Evaporation. B) The water level of the Primary Inflow and the Outflow, plotted with Rainfall. C) Ambient Air Temperature, Lake Water Temperature (at 0.5m, 1m, 2m, 3m, and 4m), and the temperature of the Primary Inflow and Outflow.

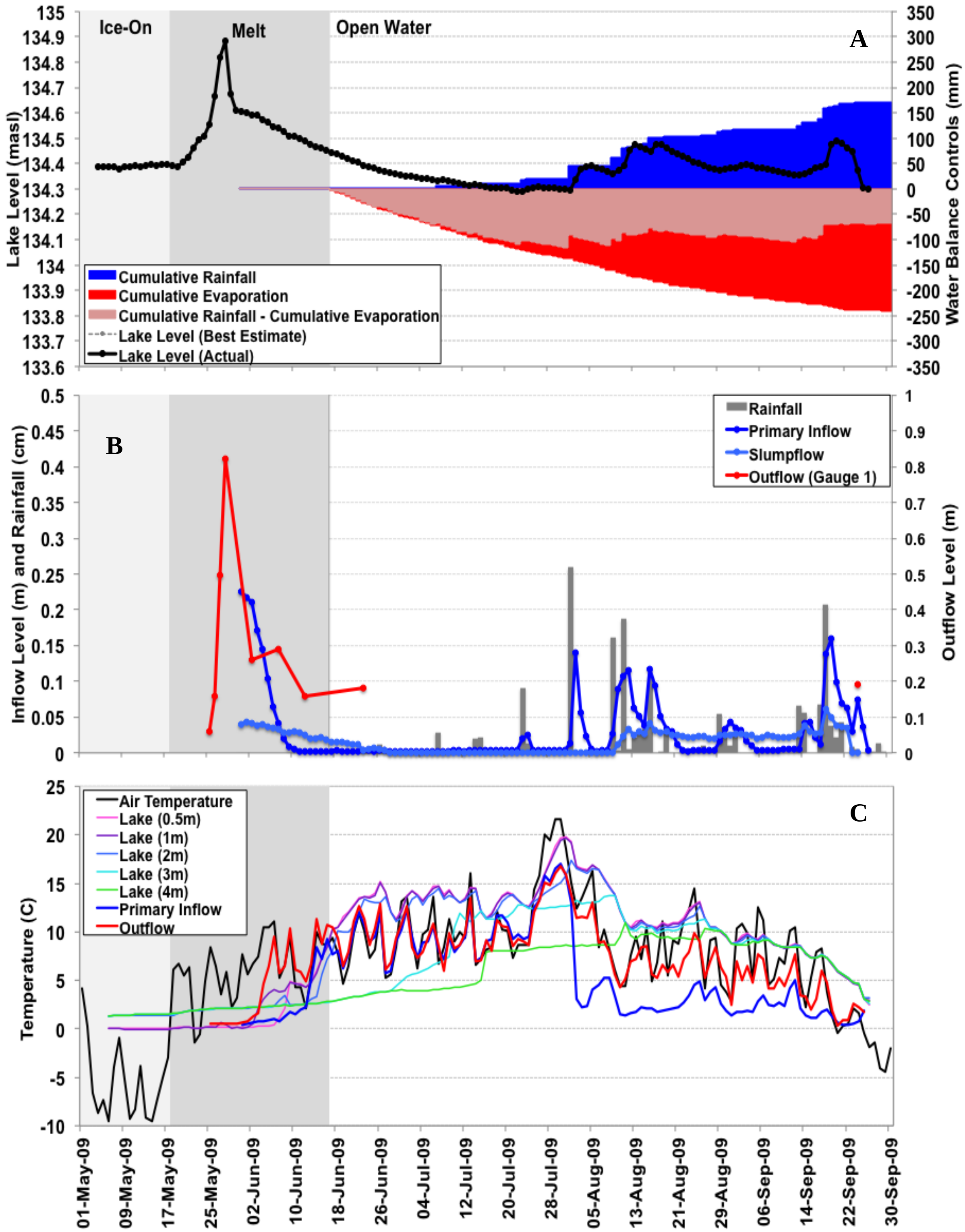


Figure 3.5. A) The Lake Level of Lake 5B from May 1st to September 30th, 2009, plotted with Cumulative Rainfall, Cumulative Evaporation, and Cumulative Rainfall minus Cumulative Evaporation. B) The water level of the Primary Inflow, Slumpflow, and Outflow, plotted with Rainfall. C) Ambient Air Temperature, Lake Water Temperature (at 0.5m, 1m, 2m, 3m, and 4m), and the temperature of the Primary Inflow and Outflow.

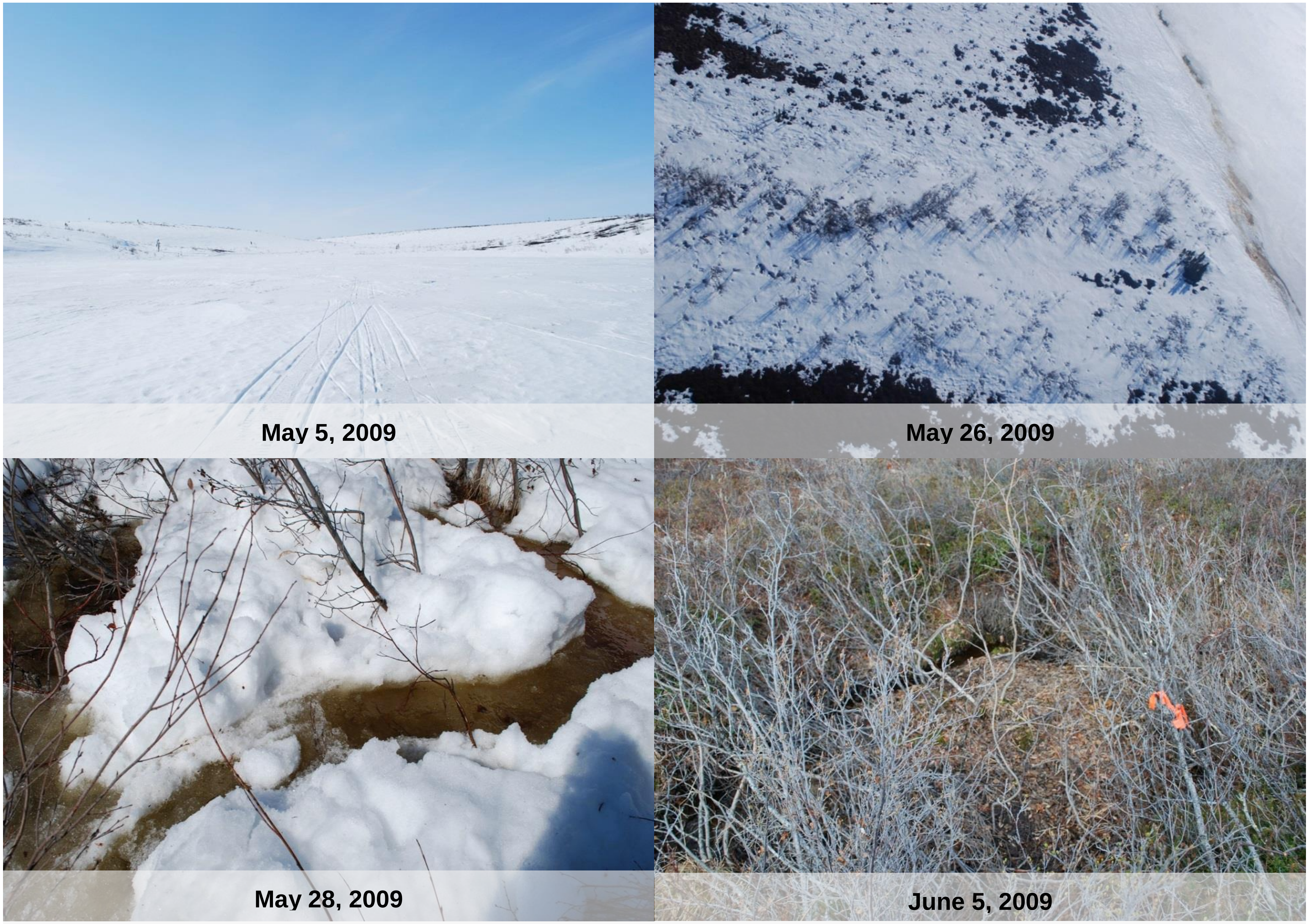


Figure 3.6. Field photos displaying progression of snowmelt at the inflow channel at Lake 5A, prior to and during spring snowmelt.

increase in the LL of the two primary study lakes that initiated lake drainage. Lake drainage continued for the remainder of the summer.

Between August 9th and August 13th, 2009, the water level of the primary inflow to Lake 5A and Lake 5B increased from 0.11m to 0.21m and 0.03m to 0.12m, respectively, in response to 38mm of rainfall. The water level of the slumpflow to Lake 5B increased from 0.0m to 0.03m. Overall, this led to a 0.1m increase in the LL of Lake 5A and Lake 5B.

Between September 13th and September 20th, 2009, the water level of the primary inflow to Lake 5A and Lake 5B increased from 0.16m to 0.23m and 0.004m to 0.16m, respectively, in response to 46mm of rainfall. The water level of the slumpflow increased from 0.03m to 0.06m. This led to a 0.13m and 0.14m increase in the LL of Lake 5A and Lake 5B, respectively.

The 2007 study year was not analogous to the 2008 and 2009 study years. Overall, the 2007 study year experienced the lowest mean annual LL of the three field study years (Lake 5A = 127.82 masl, 127.85 masl, and 127.95 masl; Lake 5B = 134.27 masl, 134.27 masl, and 134.37 masl), due to high cumulative summer evaporation and low cumulative summer rainfall. Cumulative evaporation was higher in 2007 than in 2008 and 2009 (304mm vs 281mm and 240mm, respectively). Furthermore, cumulative rainfall was lower in 2007 than the latter two study years (121mm vs 158mm and 172mm, respectively). Due to high rates of evaporation and the absence of major rainfall events, the LL of the two primary study lakes decreased throughout the entire summer, reaching a minimum in late-September (Lake 5A = 127.72 masl; Lake 5B = 134.14 masl, respectively). As a result, LL was lower at the end of the summer, relative to the subsequent two study years.

The above analysis clearly demonstrates that the water balance of Lake 5A and Lake 5B is directly controlled by a number of key hydrological processes, including spring freshet initiation, snow accumulation, open-water duration, summer rainfall, and evaporation. The timing and magnitude of these key hydrological controls is driven by climatic factors (temperature and precipitation), which have and will likely continue to fluctuate as climate changes.

Figure 3.7. Field photos displaying snowmelt progression at the outflow channel at Lake 5A, prior to and during spring snowmelt.

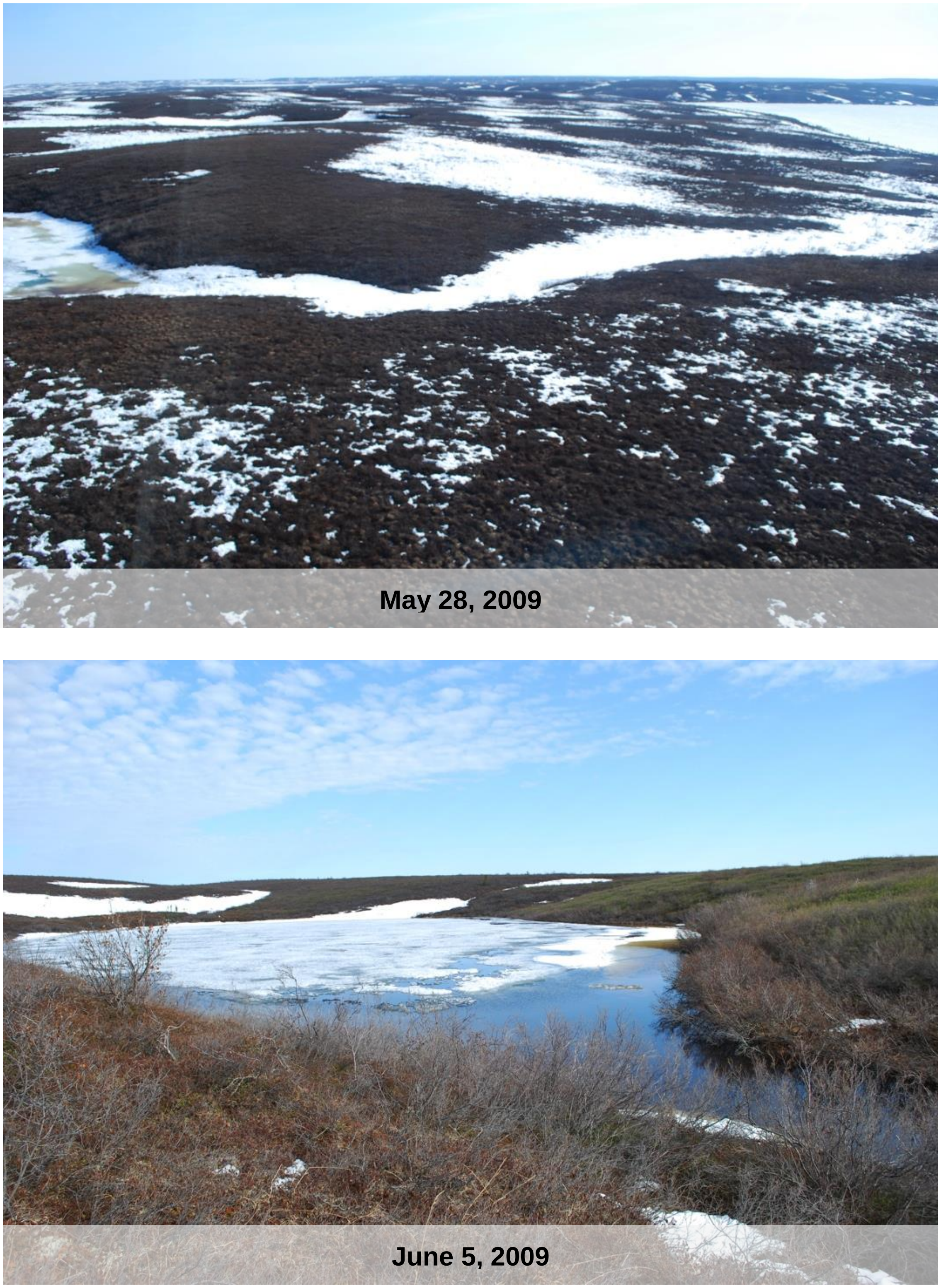


Figure 3.8. Field photos displaying the progression of snowmelt at the outflow channel at Lake 5A, from May 28th to June 5th.

3.5.2 Air Temperature

The average daily air temperature at Inuvik exhibited strong seasonal variability (**Figure 3.10**). Based on the mean for Inuvik (1958-2009), average daily air temperature remains below 0°C from about September 26th to about May 17th. In 2007, average daily air temperature increased to above 0°C and stayed above 0°C on May 18th, later than the average. In 2008, however, average daily air temperature increased above 0°C and stayed above 0°C on May 13th, earlier than the average. In 2009, average daily air temperature increased above 0°C as early as April 26th. Ambient air temperature peaked at 9.8°C, on April 27th, and did not decline below 0°C until May 3rd. Ambient air temperature did not rise above 0°C again until May 18th. Field observations indicate that air temperatures during late April 2009 were sufficiently high to initiate snowmelt. Upon arrival at the two primary study lakes on May 1st, 2009, a substantial portion of the snowpack on the east slopes of the two study lake catchments had ablated (**Figure 3.9**). Based on the maximum average daily air temperature for Inuvik (1958-2009), melting periods have occurred in April before. The average daily air temperature has increased to above 0°C at least once in April for 35 out of 52 years in the historical study period. It's noteworthy, however, that this study captured the highest late April mean daily average daily air temperature on record on April 27th, 2009.

Figure 3.9. A field photo taken on May 3, 2009 of the Lake 5B catchment. Note that the eastern slope, to the right of the shoreline slump, is almost completely bare.

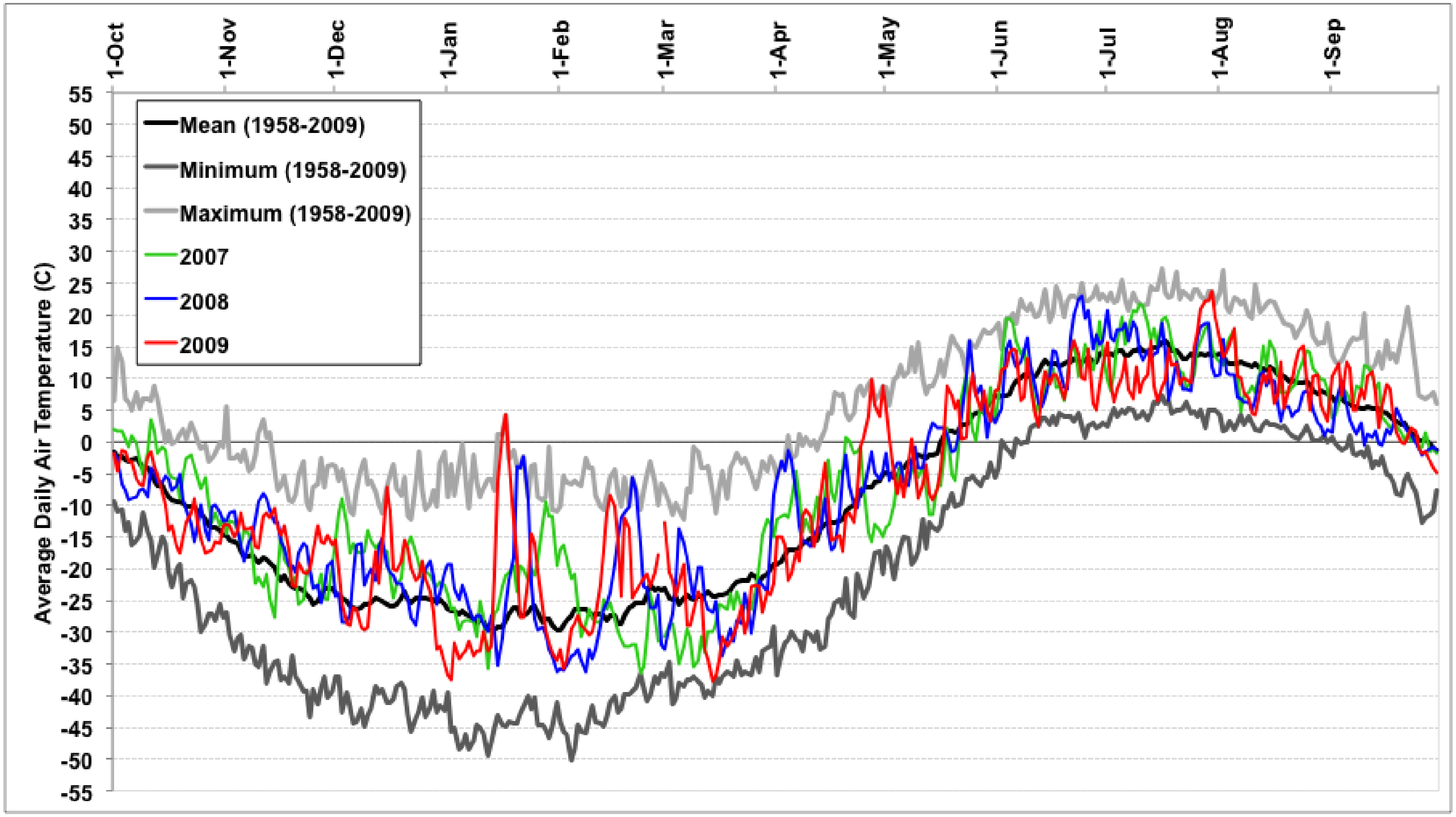


Figure 3.10. The average mean, minimum, and maximum daily air temperature for Inuvik for the years 1958 to 2009. Also presented here is the mean daily air temperature measured in Inuvik for the three primary study years (2007, 2008, and 2009).

For the three primary study years (2007, 2008, and 2009), the mean annual air temperature in Inuvik was -7.2°C, -7.6°C, and -7.8°C, respectively. For the historical study period, the maximum mean annual air temperature was -4.8°C in 1998 and the minimum mean annual air temperature was -11.2°C in 1964. All three study years were warmer than the historical average of -8.5°C, but were not warmer than the historical maximum. That being said, only 7 years had a mean annual air temperature that was warmer than 2007, making the 2007 study year one of the warmest on record.

In winter (October 1st – April 30th), the formation of lake ice is directly controlled by ambient air temperature. For the three primary study years (2007, 2008, and 2009), the mean winter air temperature in Inuvik was -18.1°C, -18.6°C, and -18.9°C, respectively. The maximum mean winter air temperature was -16.6°C in 1998 and the minimum mean winter air temperature was -23.9°C in 1971. For the three study years, the mean winter air temperature was warmer than the historical average of -20.5°C, but not warmer than the historical maximum. Similar to above, the 2007 study year had one of the warmest winters on record, with only 4 years having had a mean winter air temperature greater than -18.1°C.

The spring season was defined as May 1st to June 30th. With the exception of pre-freshet melting periods, snow and ice melt in the study region typically takes place in May and June (Lesack et al., 2013; Marsh et al., 2002). In Spring, ambient air temperature directly controls the timing and intensity of spring snow and ice melt (Ashton, 1983; Ashton, 1986). For the three primary study years (2007, 2008, and 2009), the mean spring air temperature in Inuvik was 4.7°C, 7.2°C, and 5.1°C, respectively. For 2007 and 2009, the mean spring air temperature was cooler than the historical average of 5.6°C, but not the coolest on record. The minimum mean spring air temperature was 1.8°C in 1978. Since 1958, 13 years have had a mean spring air temperature cooler than 2007. For the 2008 study year, the mean spring air temperature was warmer than the historical average, but not warmer than the historical maximum. The maximum spring air temperature was 9.9°C in 1998. Notably, only 5 years have had a mean spring air temperature that was warmer than 2008 since 1958.

In summer (July 1st to September 30th), ambient air temperature directly controls evaporation and the development of the active layer. For the three primary study years (2007, 2008, and 2009), the mean summer air temperature in Inuvik was 10.3°C, 7.8°C, and 9.0°C, respectively. In contrast with the spring months, the mean summer air temperature for 2007 was warmer than the historical average of 9.6°C, but not the warmest on record. The maximum mean summer air temperature was 12.5°C in 1979. Since 1958, 12 years have had a mean summer air temperature warmer than 2007. For 2008 and 2009, the mean summer air temperature was cooler than the historical average, but not cooler than the historical minimum. The minimum summer air temperature was 7.0°C in 1959. Notably, only 4 years have had a mean summer air temperature cooler than 2008 since 1958.

Based on a Mann-Kendall, non-parametric test for trend over the years 1958 to 2009, the mean annual air temperature in Inuvik increased at a significant rate of 0.05 °C per year (**Figure 3.11a**). Since 1958, the mean annual air temperature of the Inuvik region has increased by an average of 2.6°C. This agrees with a report released by the Department of Environment and Natural Resources (Northwest Territories, 2008), which indicated that the mean annual air temperature of the Mackenzie Delta region has increased by 3 to 4°C since the 1940s. Should ambient air temperature continue to increase at this rate, the mean annual air temperature of the Inuvik region will increase by 0.5°C over the next 10 years and by 5°C over the next 100 years. This agrees with the Arctic Climate Impacts Assessment (2004), which predicted that the mean annual air temperature in arctic regions will increase by 4 to 7°C by 2100.

The greatest rate of warming occurred during the winter months. Over the 52-year historical study period, the mean winter air temperature increased at a statistically significant rate of 0.08°C Year^{-1} (**Figure 3.12a**). Since 1958, the mean winter air temperature of the Inuvik region increased by an average of 4.2°C. Should ambient air temperature continue to increase at this rate, the mean winter air temperature of the Inuvik region will increase by an additional 5.7°C by the year 2080. This agrees with AMAP (2012), who projected that arctic inland areas will increase by 3°C to 6°C by 2080. Warmer ambient air temperatures in early winter may affect snow and ice formation, which is directly controlled by ambient air temperature. Additionally, warmer ambient air temperatures in late winter could lead to more frequent early melt periods, such as the one observed in April 2009. In contrast with the winter months, no statistically significant increasing or decreasing trends in mean spring and mean summer air temperatures were observed for the Inuvik region (**Figure 3.12b** and **Figure 3.12c**).

Recall that the study examined a set of lakes spanning from Inuvik to Richards Island near Tuktoyaktuk. The mean annual and mean winter air temperature for Tuktoyaktuk increased at a statistically significant rate of 0.04°C Year^{-1}and 0.05°C Year^{-1}, respectively, over the 52-year historical study period (**Figure 3.11b and Figure 3.13a**). Since 1958, the mean annual and mean winter air temperature for Tuktoyaktuk increased by an average of 2.1°C and 2.6°C, respectively. Interestingly, both the mean annual and the mean winter air temperature increased at a slower rate than Inuvik. In contrast with Inuvik, the mean spring air temperature for Tuktoyaktuk also increased at a significant rate of 0.04°C $Year^{-1}$ over the 52-year historical study period. Since 1958, the mean spring air temperature for Tuktoyaktuk increased by an average of 2.1°C. Warmer ambient air temperatures during the spring melt period could lead to earlier and more intense spring snow and ice melt in the northern part of the study region. In contrast with the winter and spring months, no statistically significant increasing or decreasing trend in mean summer air temperature was observed ($p > 0.05$).

Projected climate warming has significant implications for the two primary study lakes. A number of key hydrological processes controlling the water balance of Lake 5A and Lake 5B are directly controlled by ambient air temperature (i.e., ice formation, the timing and intensity of the spring freshet, ice ablation, open-water duration, and evaporation). The following sections explore how key hydroclimatic factors have changed in response to historical climate warming in the Mackenzie Delta Upland region and what effects these changes have had on the water balance of small tundra lakes.

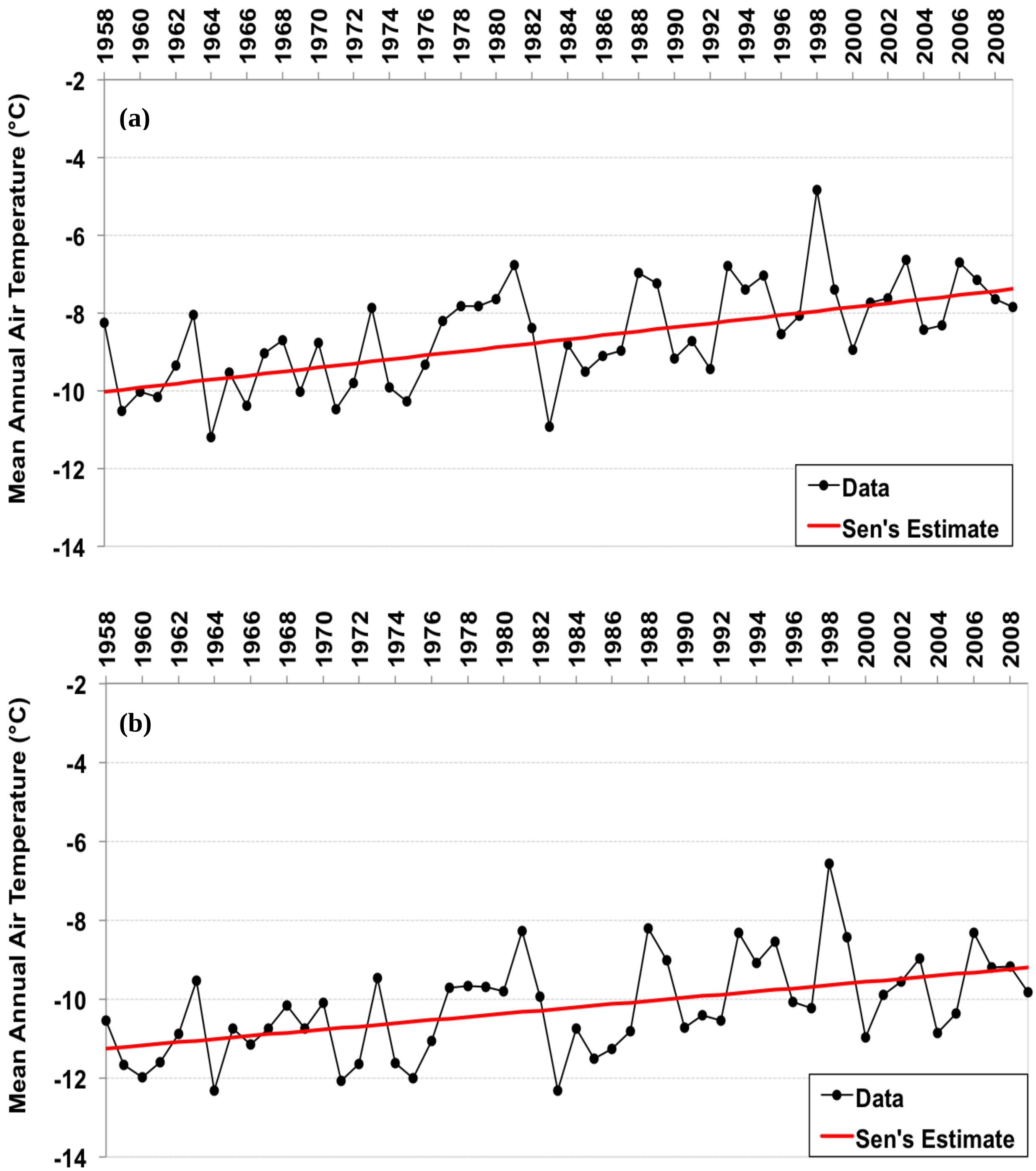


Figure 3.11. The mean annual air temperature for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009. Sen's estimate, displayed as a red line, was only presented if $p < 0.05$.

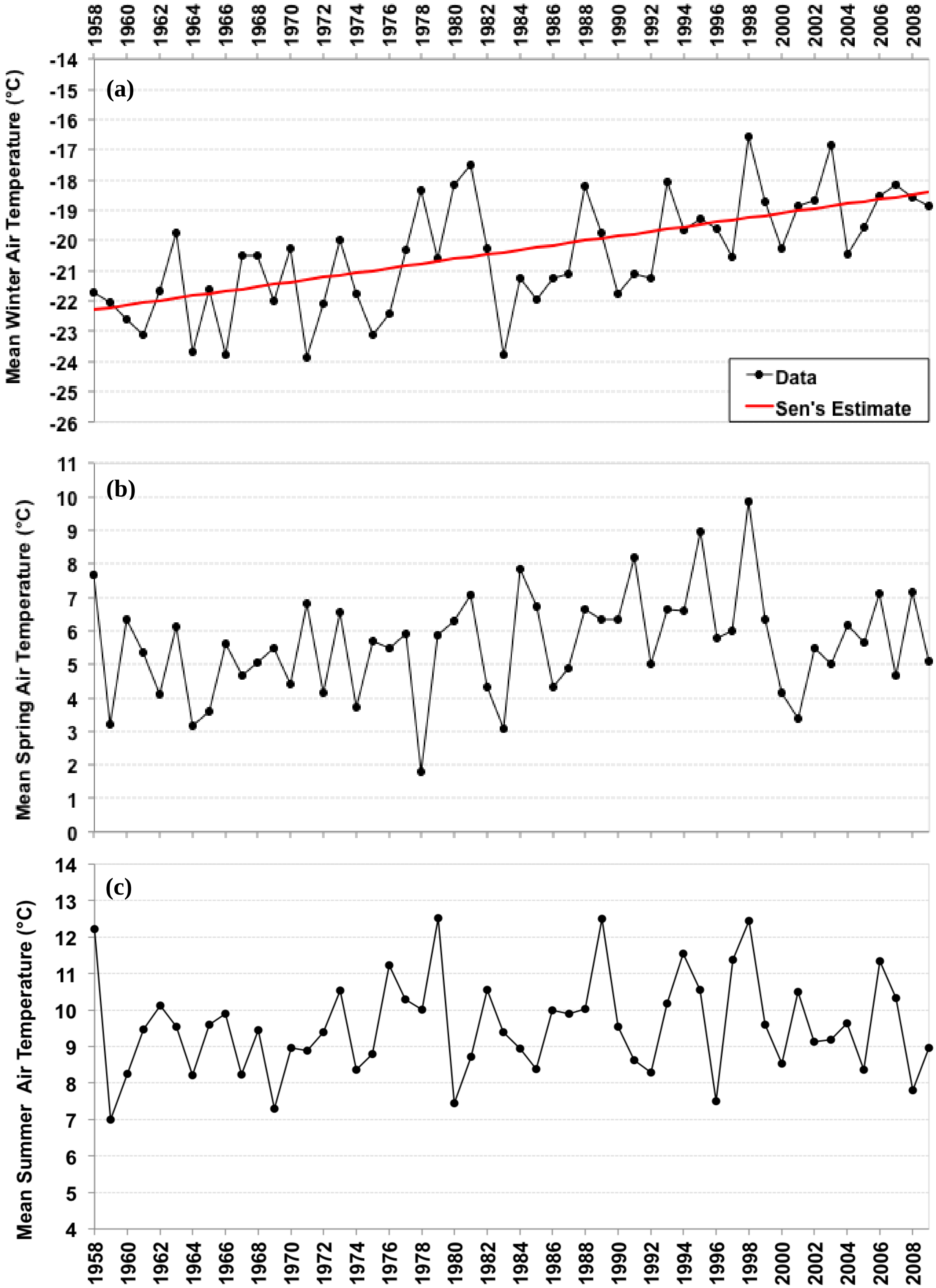


Figure 3.12. The mean winter air temperature (a), mean spring air temperature (b), and mean summer air temperature (c) for Inuvik for the years 1958 to 2009. Sen's estimate, displayed as a red line, was only presented if $p < 0.05$.

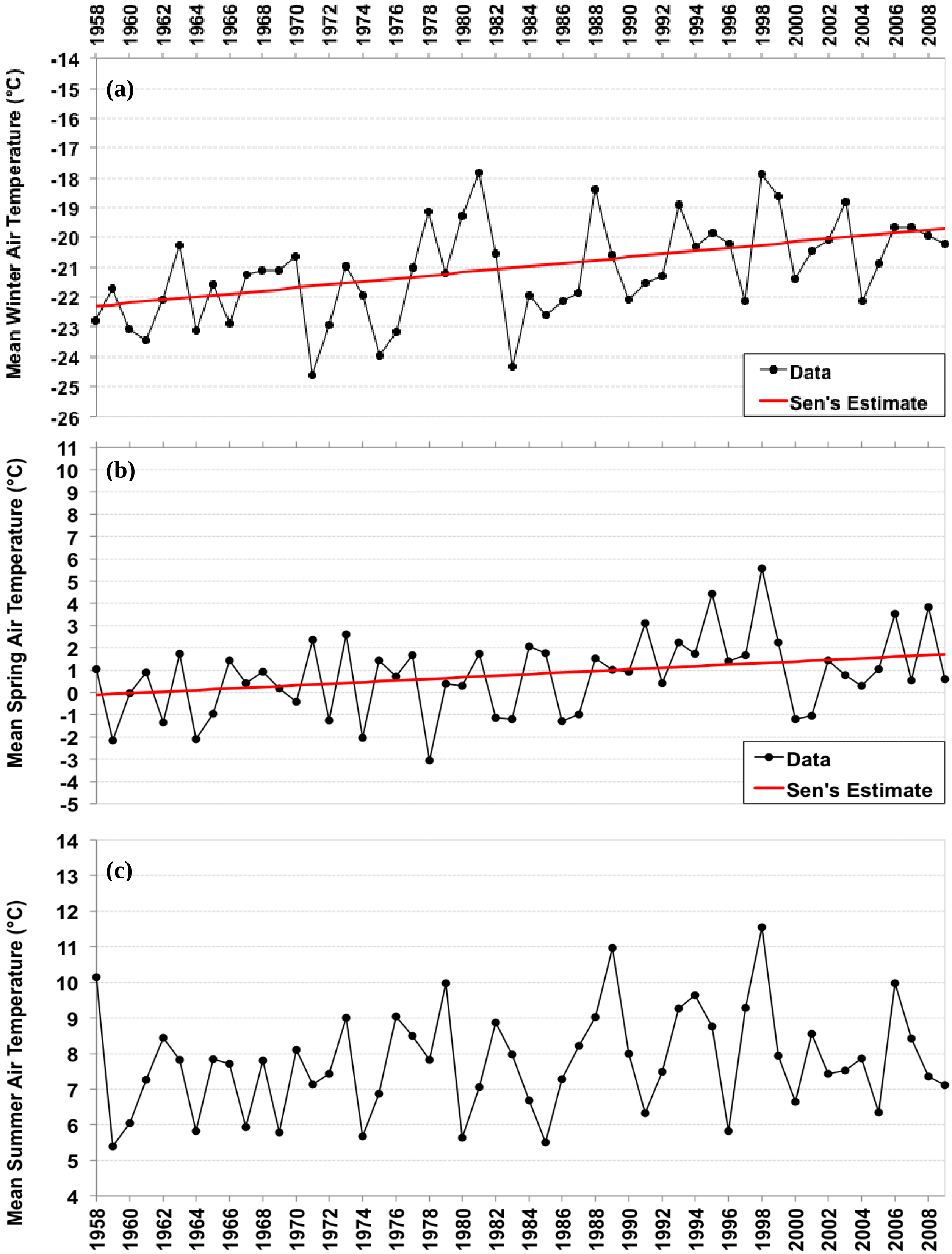


Figure 3.13. The mean winter air temperature (a), mean spring air temperature (b), and mean summer air temperature (c) for Tuktoyaktuk for the years 1958 to 2009. Sen's estimate, displayed as a red line, was only presented if $p < 0.05$.

3.5.3 Spring Freshet Initiation

As highlighted above in the water level analysis section, the spring freshet is the most significant hydrological event affecting the water balance of the two primary study lakes (see **Figure 3.3**). Ambient air temperature data, collected at Inuvik, were used to estimate the timing of the spring freshet for the years 1958 to 2009 using the method described by Pohl (personal communication, 2011). For the three study years (2007, 2008, and 2009), the spring freshet was initiated on May 21st, May 22nd, and May 17th, respectively. For all three study years, the spring freshet occurred earlier than the historical average of May 25th, but was not earlier than historical extremes. That being said, the 2009 study year had one of the earliest spring freshets on record. Only 8 of the 52 historical study years had a spring freshet earlier than 2009. Since 1958, the earliest spring freshet occurred on May 9th in 1998. It's important to note that 1998 had the warmest mean spring air temperature of the 52-year historical study period, which likely contributed to an earlier than normal spring snow and ice melt period. The latest spring freshet occurred on June 9th 1970, a range of 31 days.

There was no significant ($p > 0.05$) trend in the timing of the spring freshet in Inuvik over the years 1958 to 2009 (**Figure 3.14a**). This aligns with the results presented in the previous section, which indicated that the mean spring air temperature for Inuvik was not increasing or decreasing at a significant rate. Furthermore, this result agrees with those of Lesack et al. (2013) and Marsh et al. (2002). Based on air temperature data obtained from the Inuvik airport, Marsh et al. (2002) did not find a decreasing or increasing trend in the timing of the spring freshet. Similarly, using river discharge measurements obtained from the East Channel of the Mackenzie River, at Inuvik, Lesack et al. (2013) also did not find a decreasing or increasing trend in the timing of the spring freshet.

In contrast with the Inuvik region, there was a significant decreasing trend in the timing of the spring freshet in Tuktoyaktuk over the years 1958 to 2009 (**Figure 3.14b**). Over the historical study period, the timing of the spring freshet decreased at an average rate of 0.185 days per year. In other words, the average date of the spring freshet has decreased by approximately 10 days since 1958. This is also aligns with the results presented in the previous section, which indicated that the mean spring air temperature for Tuktoyaktuk was increasing at a significant rate. This suggests that the effects of recent climate warming on the timing of spring snowmelt may vary across the study region. This suggestion is supported by the work of Bonsal and

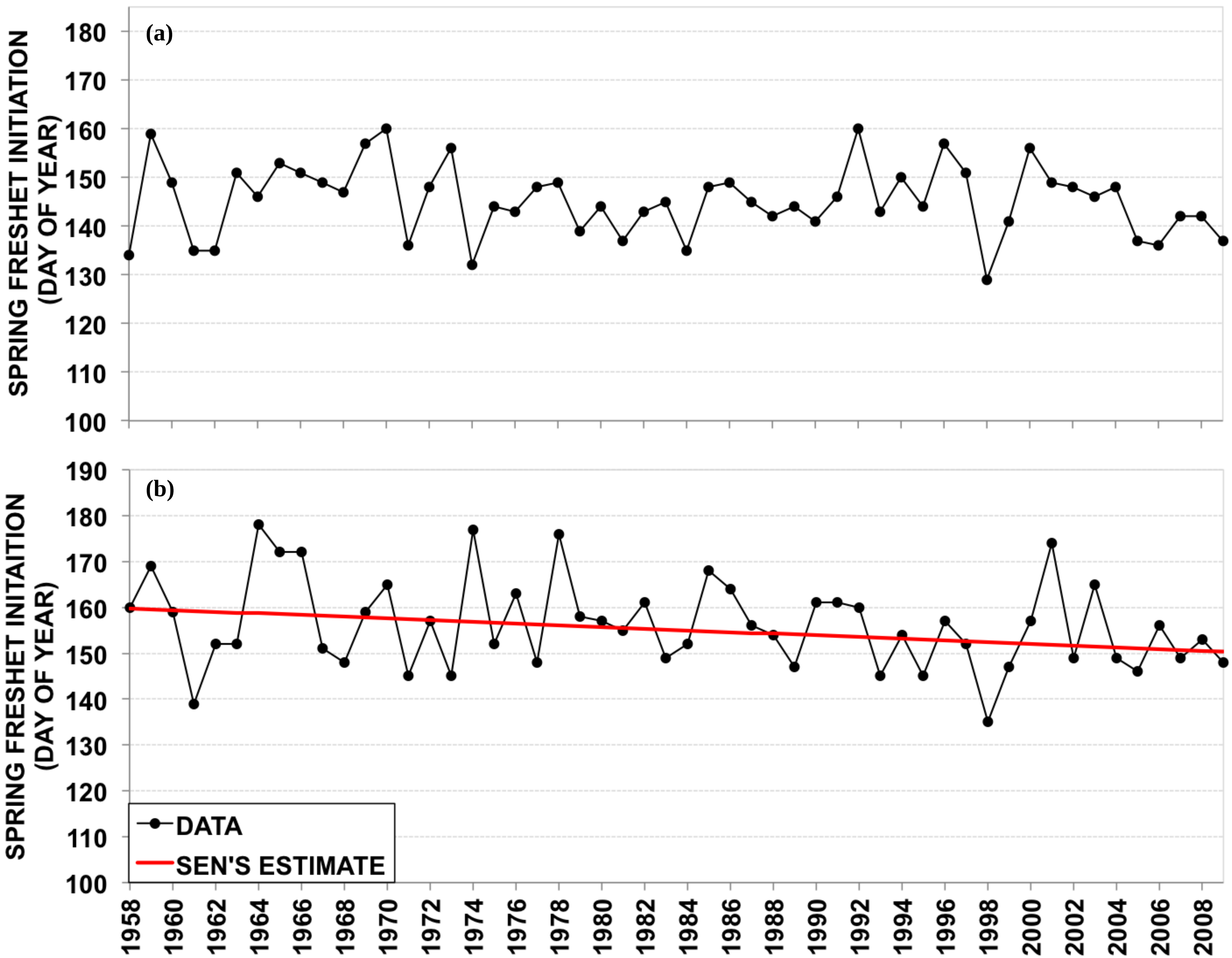


Figure 3.14. The timing of the spring freshet for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009. Sen's estimate, displayed as a red line, was only presented if $p < 0.05$.

Prowse (2003) who reported that long term, inter-annual trends in the timing of the Spring 0°C Isotherm exhibit considerable spatial variability, especially in western Canada.

3.5.4 Open-water Duration

For the three study years, the estimated timing of ice-off at the two study lakes was determined using surface water temperature data collected at the centre of each lake. In 2007, 2008, and 2009, the first day that the two study lakes were ice-free was June 8th, June 13th, and June 17th, respectively. The two study lakes became ice free later in 2009, in part because the ice was thicker than the other two study years. The maximum ice thickness of Lake 5A measured on May 1st was almost 0.5 m greater in 2009 than in 2007 and 2008 (1.57, 0.94, 1.08, and 1.00 m, respectively). Ice formation is directly controlled by ambient air temperature (Ashton, 1986). As highlighted in Section 3.5.2, the mean winter air temperature in 2009 was colder than the mean winter air temperature of the previous two study years. Colder winter ambient air temperatures may have led to a thicker lake ice cover.

An index of PDD (outlined in the Methodology section) was used to estimate the timing of ice-off at the two study lakes for the years 1958 to 2009 (**Figure 3.15a**). For the three study years (2007, 2008, and 2009), the date of ice-off was June 15th, June 14th, and June 15th. For all three study years, the two primary study lakes were ice-free earlier than the historical average of June 17th, but not earlier than historical extremes. Furthermore, ice-off was earlier than in 2008 for 17 of the 52 historical study years. The earliest estimated date of ice-off was June 1st in 1998. This is not surprising, since 1998 had the earliest spring freshet period on record, which was attributed to warmer than normal spring ambient air temperatures. The latest ice-off date on record was June 28th in 1978, a range of 27 days. Again, this is not surprising, because 1978 had coolest mean spring ambient air temperature in the 52-year historical study period. Based on a Mann-Kendall trend analysis, no significant trend in ice-off was observed over the historical study period.

By comparison, the mean date of ice-off in Tuktoyaktuk was over two weeks later than Inuvik at July 4th. The earliest date of ice-off occurred on June 18th in 1998 and the latest on July 19th in 1974 (**Figure 3.15b**). The timing of ice-off in Tuktoyaktuk decreased at a statistically significant rate of approximately 0.185 days each year over the 52-year historical study period. In other words, the average date of ice-off at the northern end of the study region decreased by approximately 10 days between the years 1958 and 2009.

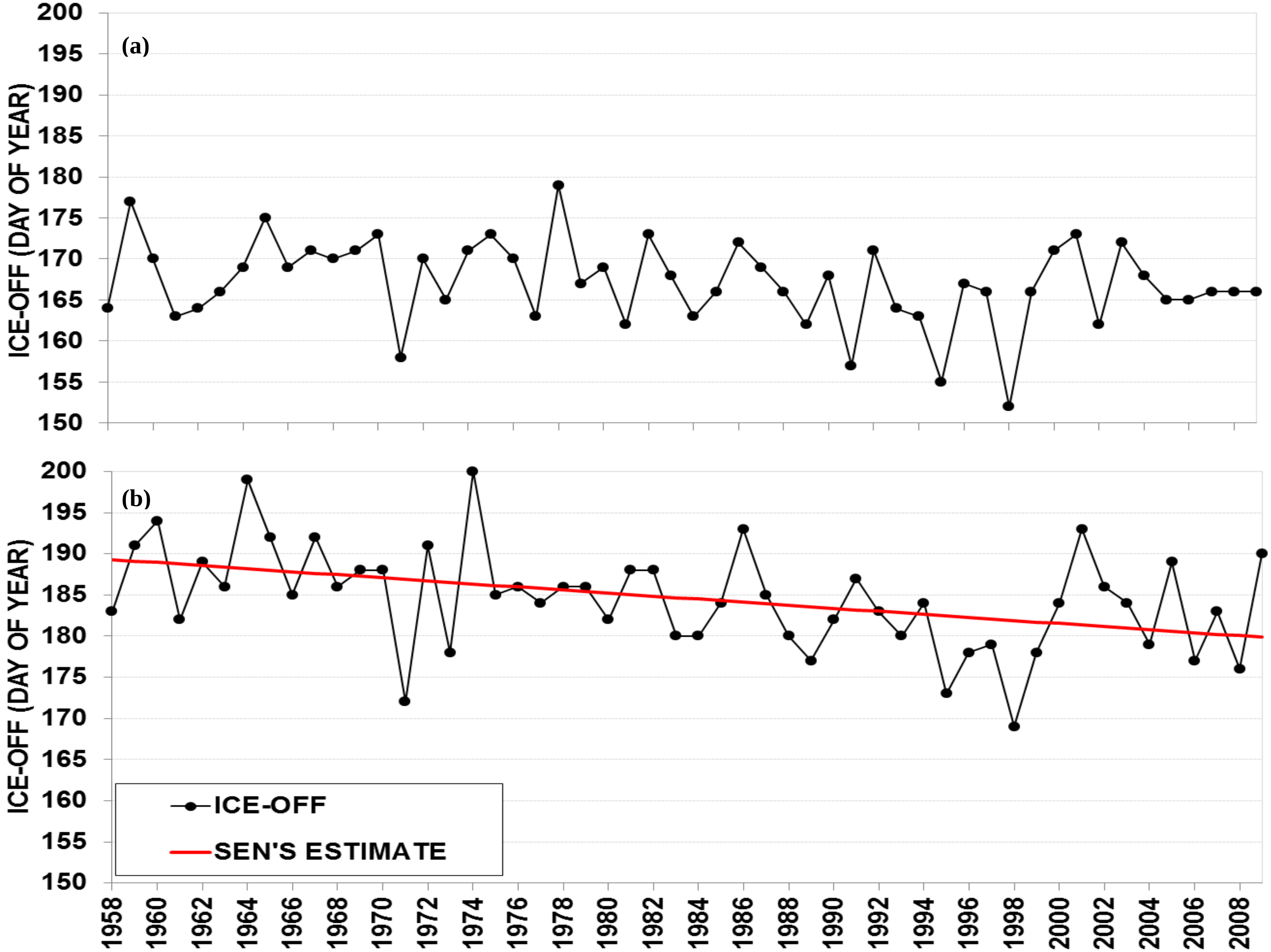


Figure 3.15. The timing of ice-off in (a)Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009. Sens's estimate, displayed in red, is only presented if p < 0.0.5.

An index of FDD (outlined in the Methodology section) was used to estimate the timing of ice-on at the two study lakes for the years 1958 to 2009. For the three study years (2007, 2008, and 2009), ice formation occurred on October 4th, October 5th, and October 1st. For all three study years, ice on occurred earlier than the historical average of October 6th, but not the earliest on record. The earliest date of ice-on was September 20th in 1992 and the latest was October 19 in 1969. For 12 of the 52 historical study years, ice-on occurred earlier than 2009. No significant trend in the timing of ice-on for the two study lakes was discerned over the historical study period (**Figure 3.16 a.**).

By comparison, the average date of ice-on for Tuktoyaktuk was October 6th, the same as the average for Inuvik. Also similar to Inuvik, the earliest ice-on date of ice-on was September 22nd in 1992 and the latest ice-on date was October 19th in 1969. No significant trend in the timing of ice-on was observed over the historical study period (**Figure 3.16b**).

For the three study years (2007, 2008, and 2009), the open-water period was 111 days, 113 days, and 108 days long, respectively. In 2007 and 2009, the open-water period was shorter than the historical average of 112 days, but was not the shortest on record. For 14 of the 52 historical study years, the open water period was shorter than 2009. Over the 52-year historical study period, the shortest open-water period was 92 days in 1992. In 2008, the open-water period was longer than the historical average, but was not the longest on record. For 23 of the 52 historical study years, the open water period was longer than 2008. The longest open-water period was 129 days in 1998, due to warm spring ambient air temperatures and an early spring snow and ice melt period. There was no significant trend in the length of the open-water period over the historical study period (**Figure 3.17a**).

By comparison, the average length of the open-water season in Tuktoyaktuk was 96 days, which is 16 days shorter than at the southern end of the study transect near Inuvik. The shortest open-water period was 69 days in 1974 and the longest open-water period was 115 days in 1998. No significant trend in the length of the open-water period was observed over the historical study period (**Figure 3.17b**).

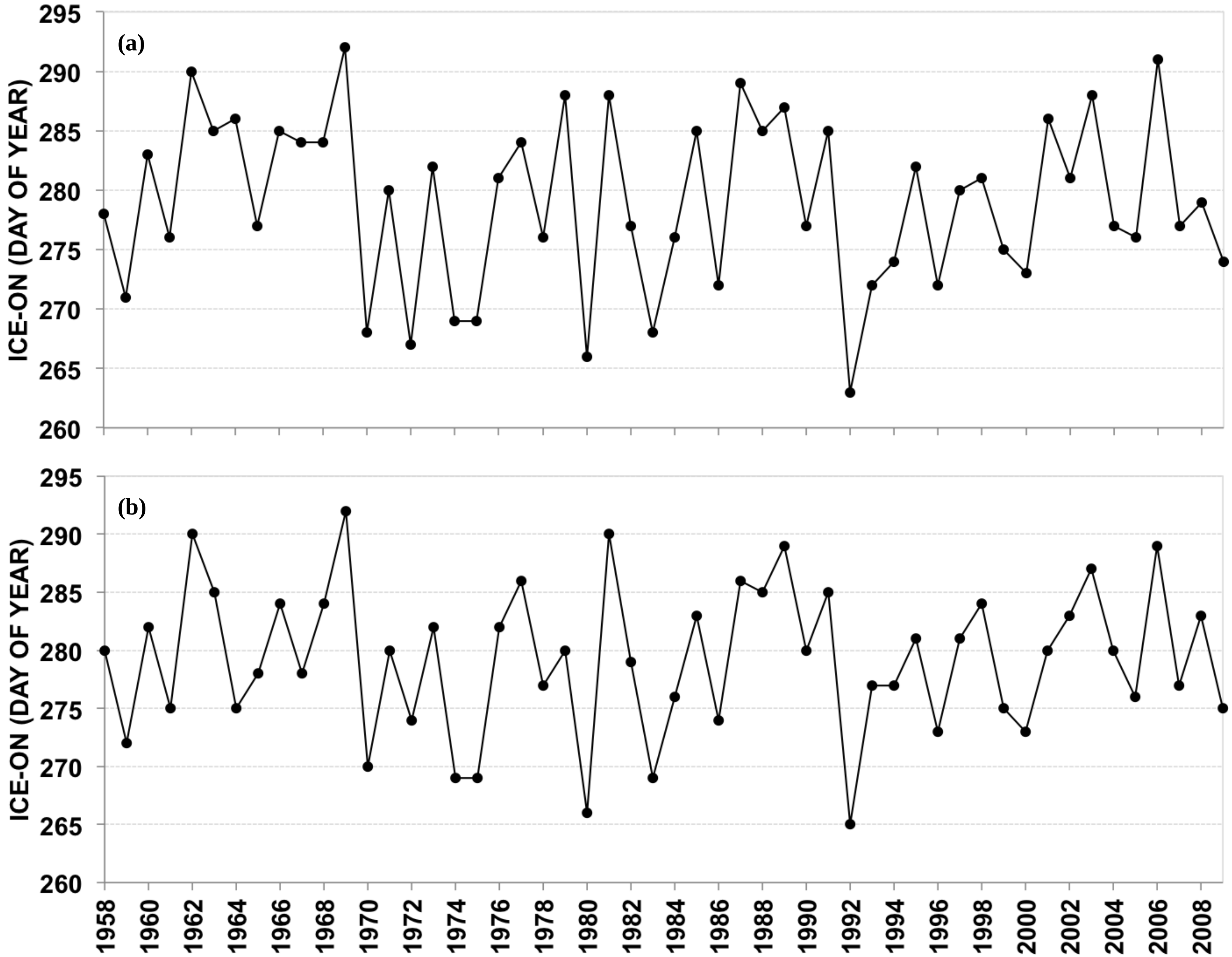


Figure 3.16. The timing of ice-on in Inuvik (a) and Tuktoyaktuk (b) for the years 1958 to 2009.

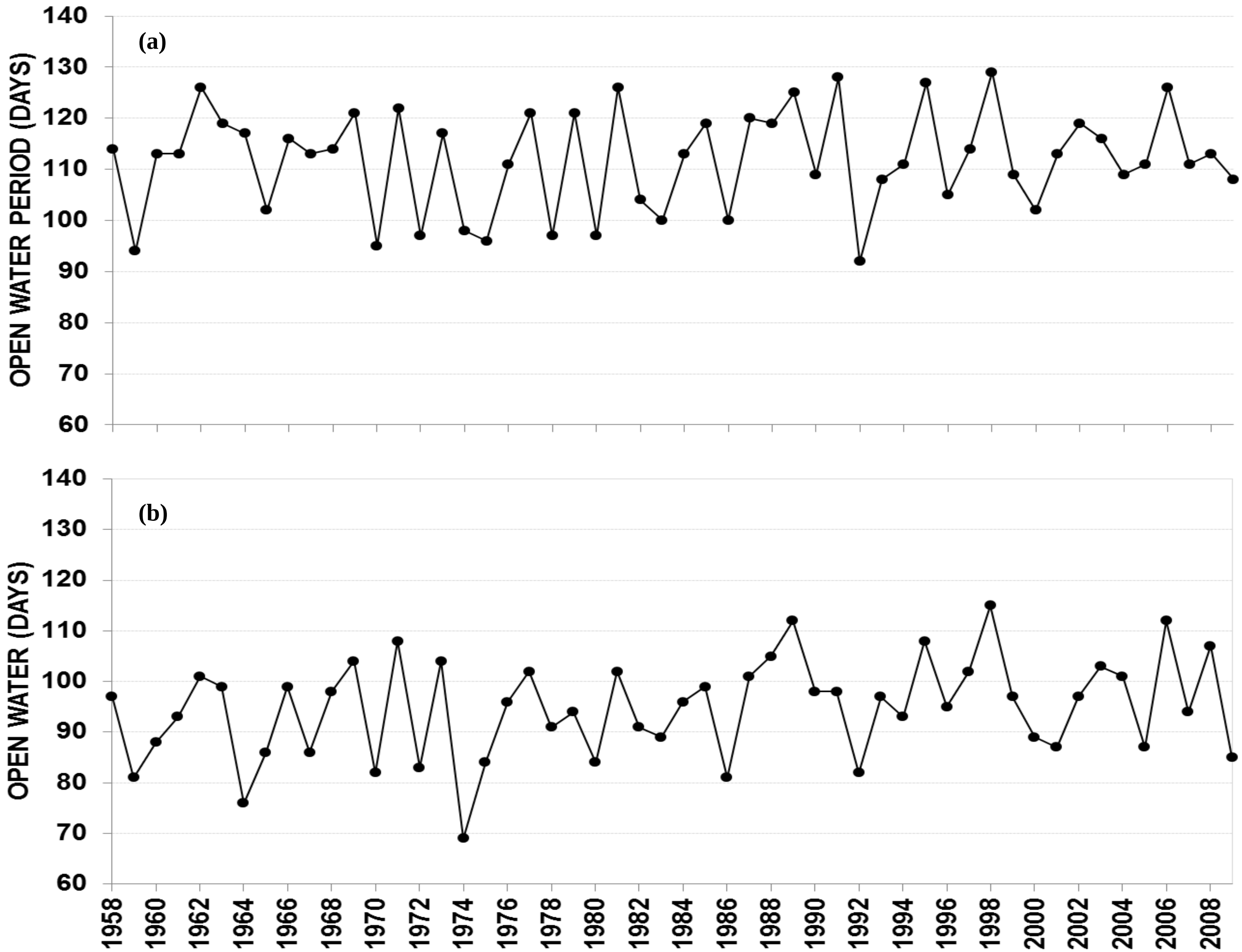


Figure 3.17. The length of the open-water period, in days, for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009.

3.5.5 Precipitation

3.5.5.1 Total Annual Precipitation

For the three study years (2007, 2008, and 2009), the Total Annual Precipitation at Inuvik was 216mm, 307mm, and 318mm, respectively (**Figure 3.18a**). The 2007 study year was a relatively dry year compared to the historical average of 307mm, close to the driest year on record, which was observed in1979 at 210mm. Notably, only 4 out of the 52 historical study years had less Total Annual Precipitation than 2007. In other words, 2007 was an abnormally dry year for this region, which led to a decrease in LL over the summer period. Conversely, the 2009 study year was a wet year, relative to the historical average, but not the wettest on record. 23 out of the 52 historical study years received more precipitation than 2009. The maximum Total Annual Precipitation was 485mm in 1972. There was not a significant trend in the Total Annual Precipitation in Inuvik over the 52-year historical study period.

By comparison, the average Total Annual Precipitation for Tuktoyaktuk was 167mm, which is140mm less than the average Total Annual Precipitation for Inuvik (**Figure 3.18b**). The minimum Total Annual Precipitation was 75 mm in 1958 and the maximum Total Annual Precipitation was 389 mm in 1995. It's important to note that no precipitation data was available 1994 to 1999. Data infilling could have contributed to led to higher than normal Total Annual Precipitation, Annual May 1st Snowpack, and Annual Rainfall Index for these years. Over the years 1958 to 2009, the Total Annual Precipitation in Tuktoyaktuk increased at a statistically significant rate of 1.044 mm year^{-1}. In other words, Total Annual Precipitation has increased by an average of 54 mm since 1958.

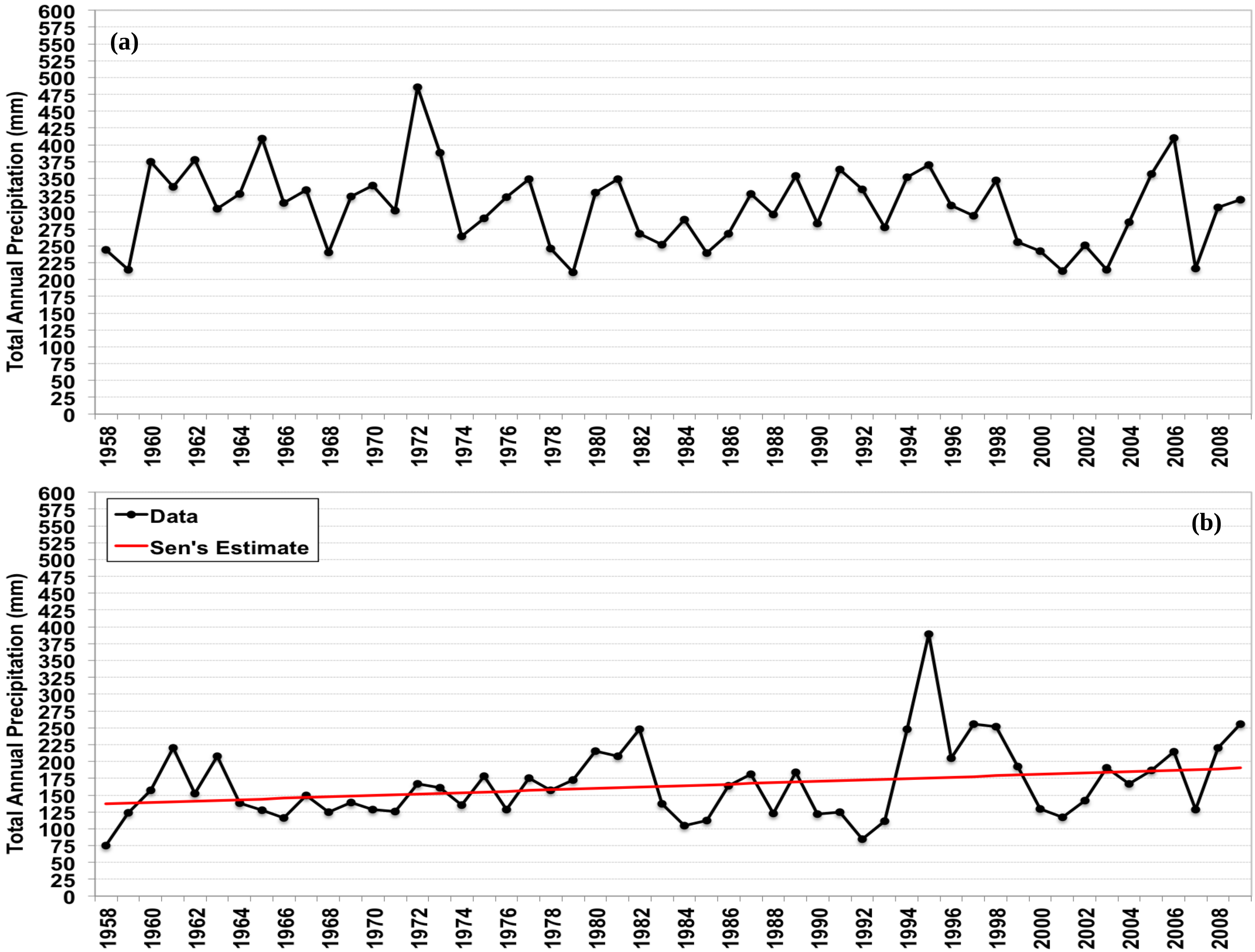


Figure 3.18. The Total Annual Precipitation for (a) Inuvik and (b) Tuktoyaktuk from 1958 to 2009. Sen's estimate, presented in red, is only displayed if $p < 0.05$.

3.5.5.2 Snow water equivalent

In 2007, 2008, and 2009, the average weighted SWE of the contributing catchment at Lake 5A was 0.136m, 0.173m, and 0.108m, respectively (**Table 3.2**). Over the three study years, the average weighted catchment SWE at Lake 5A was the highest in 2008 and the lowest in 2009. Notably, an early melt period in April 2009 melted a substantial portion of the snowpack at Lake 5A and Lake 5B before the snow survey took place, which could explain why the average weighted catchment SWE was so low. In 2008 and 2009, the average weighted catchment SWE of Lake 5B was 0.148m and 0.150m, respectively (**Table 3.3**). In contrast with Lake 5A, the average weighted catchment SWE at Lake 5B was higher in 2009 than in 2008. This is supported by the LL data presented in **Figure 3.3**.

Table 3.2. The Average Weighted Catchment SWE, Total Catchment Area, Volume of Snow Melt Water, and Lake Surface Area for the catchment contributing to Lake 5A, for the years 2007, 2008, and 2009. The Volume of Melt Water was divided by the Lake Surface Area to estimate the potential Water Level rise associated with spring snowmelt.

Year	Average Weighted Catchment SWE (m)	Total Catchment Area (m^2)	Volume of Melt Water (m^3)	Lake Surface Area (m^2)	Potential Water Level Rise (m)
2007	0.136	201,963	27,467	28,019	0.98
2008	0.173	201, 963	34, 940	28,019	1.25
2009	0.108	201, 963	21, 812	28,019	0.78
Average	**0.127**	**201, 963**	**28,073**	**28, 019**	**1.02**

Table 3.3. The Average Weighted Catchment SWE, Total Catchment Area, Volume of Snow Melt Water, and Lake Surface Area for the catchment contributing to Lake 5B, for the years 2008, and 2009. The Volume of Melt Water was divided by the Lake Surface Area to estimate the Water Level rise associated with spring snowmelt.

Year	Average Weighted Catchment SWE (m)	Total Catchment Area (m^2)	Lake Surface Area (m^2)	Volume of Melt Water (m^3)	Potential Water Level Rise (m)
2008	0.148	250,085	32,944	40,778	1.12
2009	0.150	250,085	32,944	37,393	1.14
Average	**0.149**	**287, 168**	**32, 944**	**40,060**	**1.13**

The distribution of snow within the study area is directly controlled by landscape-level factors (Lantz et al., 2009; Pohl and Marsh, 2006; Pomeroy et al., 1997). A recent study conducted by Lantz et al. (2009) indicated that terrain affected by SRTS typically has deeper winter snowpacks than adjacent unaffected terrain. Snow is typically transported from areas with a high fetch (e.g., hilltops) to areas with a low fetch (e.g., hillsides). For instance, Pohl and Marsh (2006) found that snow drifts tend to form within depressions and along ridges, which typically have a low fetch. SRTS generally occurs on hillsides, forming deep depressions with steep headwalls, or ridges. Based on the parameters outlined by Pomeroy et al. (1997) and Pohl and Marsh (2006), terrain affected by SRTS would be a sink for blowing snow. This is substantiated by the data presented in **Table 3.4**. In 2008 and 2009, the average SWE of the SRTS-affected terrain at Lake 5B was 21% and 28% greater, respectively, than the average SWE of the adjacent unaffected terrain. This suggests that SRTS could affect the water balance of small tundra lakes by increasing the snowmelt contribution of the contributing lake catchment.

Table 3.4. The average SWE of the shoreline slump at Lake 5B compared with the average SWE of the adjacent unaffected terrain.

	Average SWE (m)		Difference
	Slump	**Adjacent Unaffected Terrain**	**%**
2008	0.165	0.137	21
2009	0.122	0.96	28
	0.156	0.117	25

In order to examine the effect SRTS had on the water balance of Lake 5B, the actual average weighted catchment SWE of Lake 5B was compared to the hypothetical average weighted catchment SWE of Lake 5B, which was calculated by assuming the average SWE of the SRTS-affected terrain was equal to that of the adjacent unaffected terrain. In 2008 and 2009, SRTS at Lake 5B increased the snowmelt contribution of the contributing lake catchment by 708 m^3 and 632 m^3, respectively (**Table 3.5**). This led to an increase in lake level of 0.02 m, for both study years. The results presented here indicate that SRTS does affect the water balance of Lake 5B slightly, by increasing the snowmelt contribution of the contributing lake catchment and, subsequently, LL.

Table 3.5. The Average Weighted Catchment SWE, Total Catchment Area, Volume of Melt Water, and Lake Surface Area for Lake 5B (affected by SRTS) and Lake 5B (unaffected by SRTS), for the 2008 and 2009 study years. The Volume of Melt Water was divided by the Lake Surface Area to estimate the Water Level rise associated with spring snowmelt.

Study Year		Average Weighted Catchment SWE (m)	Total Catchment Area (m^2)	Volume of Melt Water (m^3)	Lake Surface Area (m^2)	Potential Lake Level Rise (m)
2008	**Affected**	0.148	250,085	36, 931	32,944	1.12
	Unaffected	0.145	250,085	36,223	32,944	1.10
	Difference	0.003	0	708	0	0.02
2009	**Affected**	0.150	250,085	37,393	32,944	1.14
	Unaffected	0.147	250,085	36,760	32,944	1.12
	Difference	0.003	0	633	0	0.02

Shoreline Retrogressive Thaw Slumping affects 8% of the contributing catchment at Lake 5B. For some lakes in the study region, however, SRTS can affect up to half the contributing lake catchment (Personal Observation, 2009). The Northern and Eastern slopes of Lake 5B make up approximately 53% of the contributing lake catchment. In order to examine how SRTS would affect the lake water balance if approximately half the contributing lake catchment were affected, the average weighted catchment SWE for Lake 5B was calculated by assuming the average SWE of the Northern and Eastern Slopes was equal to that of the shoreline slump. In this scenario, SRTS increased the volume of melt water from the contributing lake catchment by 6,403 m^3 and 5,042 m^3 in 2008 and 2009, respectively (**Table 3.6**). Furthermore, SRTS increased the potential rise in LL associated with spring melt by 0.19m and 0.15m in 2008 and 2009, respectively.

Table 3.6. The shoreline slump at Lake 5B occupies 8% of the contributing lake catchment. The values presented below represent what the Average Weighted Catchment SWE, Total Catchment Area, Volume of Melt Water, and Lake Surface Area for Lake 5B would be if the shoreline slump affected 53% of the catchment area. The Volume of Melt Water was divided by the Lake Surface Area to estimate the Water Level rise associated with spring snowmelt. Also presented here is what the Average Weighted Catchment SWE of Lake 5B would be if Lake 5B was unaffected by SRTS.

Study Year		**Average Weighted Catchment SWE (m)**	**Total Catchment Area (m^2)**	**Volume of Melt Water (m^3)**	**Lake Surface Area (m^2)**	**Potential Lake Level Rise (m)**
2008	**Affected**	0.171	250,085	42,626	32,944	1.29
	Unaffected	0.145	250,085	36,223	32,944	1.10
	Difference	0.026	0	6,403	0	0.19
2009	**Affected**	0.167	250,085	41,801	32,944	1.27
	Unaffected	0.147	250,085	36,759	32,944	1.12
	Difference	0.020	0	5,042	0	0.15

The annual snowpack index for Inuvik was used to examine historical variability in the SWE of Lake 5A and Lake 5B. In 2007, 2008, and 2009, the annual snowpack index for Inuvik was 100mm, 112mm, and 140mm, respectively (**Figure 3.19**). Notably, the 2007 study year had the lowest annual snowpack index, relative to 2008 and 2009, which contributed to lower spring water levels. For all three study years, the annual snowpack index was less than the historical average of 153mm, but not lower than historical extremes. That being said, 2007 had one of the smallest May 1st snowpacks on record. Only 3 of the 52 historical study years received less snow than 2007. The minimum annual snowpack index was 75mm in 2003 and the maximum annual snowpack index was 287mm in 1972. In contrast with the field data presented for Lake

5A, the annual snowpack index for Inuvik was greater in 2009 than in 2007 and 2008. This supports the theory presented earlier, which suggested that the SWE measured at Lake 5A in 2009 was lower than 2007 and 2008 due to the early melt period in late-April 2009. There was not a statistically significant trend in the annual snowpack index for Inuvik over the 52-year study period.

Over the 52-year historical study period, the average annual snowpack index for Tuktoyaktuk was 70mm, approximately 83mm less than Inuvik. The maximum annual snowpack index was 126 mm in 1997 and the minimum annual snowpack index was 23 mm in 1965. In contrast with Inuvik, there was a significant increasing trend in the annual snowpack index for Tuktoyaktuk over the 52-year historical study period. The annual snowpack index increased at an average rate of approximately 0.715 mm per year. This is in line with the current climate projections for mid-high latitudes, which suggest that warmer ambient air temperatures, associated with climate warming, will lead to more precipitation during the winter months (AMAP, 2012).

On open tundra, such as the hilltops at the two primary study lakes, a substantial amount of precipitation can be lost over the winter months due to snow transport and sublimation. For instance, Pomeroy et al. (1997) found that at Trail Valley Creek 28% and 18% of snow was lost over the winter months to sublimation and transport, respectively. Since snow surveys were obtained at the end of the winter season, the SWE estimates for Lake 5A and Lake 5B likely account for the effects of the snow transport and sublimation. It’s important to note, however, that the annual snowpack index does not account for the effects of snow sublimation and transport. It also doesn’t account for landscape-level drivers of SWE, such as landcover type. For this reason, it may not accurately represent the SWE and the two primary study lakes. That being said, it does represent inter-annual trends in the SWE the two primary study lakes (wet years vs. dry years), assuming that losses to snow sublimation and transport are relatively consistent year to year.

Historical trends in the annual snowpack index are not uniform across the study region. Overall, the results of this study suggest that the snowmelt contribution to small tundra lakes located at the northern end of the study region, near Tuktoyaktuk, has increased over the past half century.

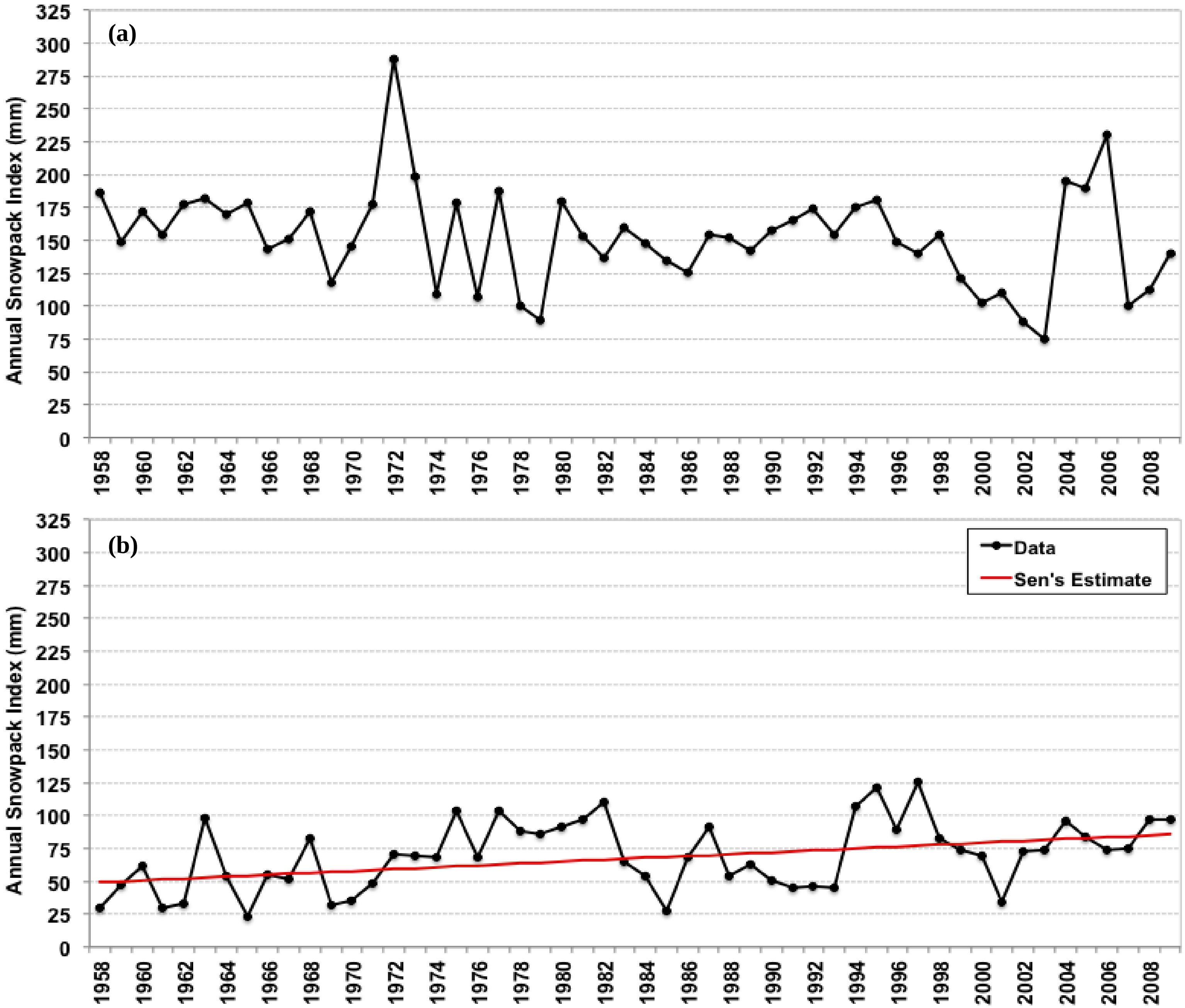


Figure 3.19. The annual snowpack index for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009. Sen's estimate, presented in red, is only displayed if $p < 0.05$.

3.5.5.3 Summer Rainfall

Overall, summer rainfall events were a vital source of water recharge to the two primary study lakes. For the three study years (2007, 2008, and 2009), the annual rainfall index for Inuvik was 104 mm, 183 mm, and 170 mm, respectively (**Figure 3.20a**). For 2007, the annual rainfall index was less than the historical average of 135mm, which led to a substantial draw down in LL over the summer months. Notably, 2007 was not the lower than historical extremes. That being said, only 9 of the 52 historical study years received less rainfall than 2007. The minimum annual rainfall index was 45 mm in 1968. This suggests that low summer water levels, similar to those observed in 2007, are not common to small tundra lakes in the study region. By comparison, the annual rainfall index in 2008 and 2009 was greater than the historical average, but not greater than historical extremes. Notably, only 6 of the 52 historical study years received more rainfall than 2008, which suggests that high summer water levels, such as those observed in 2008, are also not common to small tundra lakes in the study region. The maximum annual rainfall index was 193 mm in 1965. There was not a statistically significant trend in the annual rainfall index for Inuvik over the historical study period.

Over the 52-year historical study period, the mean total annual rainfall index for Tuktoyaktuk was 90 mm, approximately 45mm less than Inuvik. The minimum annual rainfall index was 36 mm in1992 and the maximum annual rainfall index was 246mm in 1995. Similar to Inuvik, there was not a statistically significant trend in the annual rainfall index for Tuktoyaktuk over the historical study period.

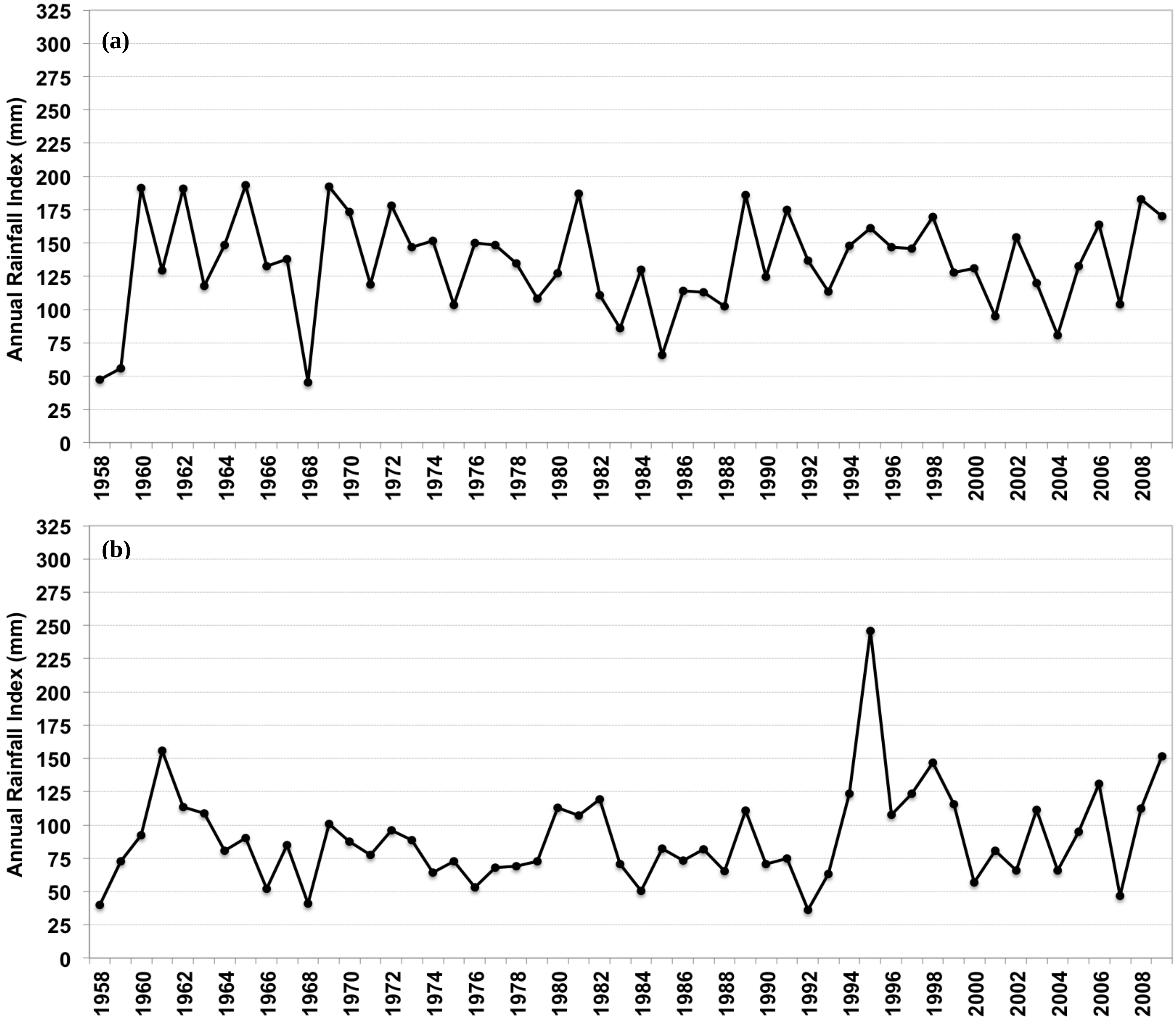


Figure 3.20. The annual rainfall index for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009.

3.5.6 Evaporation

Between June 29th and August 25th of 2007, 2008, and 2009, the Total Priestley-Taylor evaporation was 183mm, 174mm, and 150mm, respectively (**Table 3.7**). The maximum Total Priestley-Taylor evaporation was 183mm in 2007, due to high summer air temperatures and high incoming net radiation. The average daily net radiation at Lake 5A was 120 $W.m^{-2}$ in 2007, relative to 107 $W.m^{-2}$, and 116 $W.m^{-2}$ in 2008 and 2009, respectively. Interestingly, incoming net radiation was higher in 2009 than 2008, which would generally lead to higher rates evaporation. In contrast, Total Priestley-Taylor evaporation was lower in 2009 than 2008. This is likely due to low lake heat storage in 2008, which drove the rate of evaporation up, and high lake heat storage in 2009, which drove Priestley-Taylor evaporation down (-10 $W.m^{-2}$ and 8 $W.m^{-2}$, respectively). Note that Priestley-Taylor Evaporation is inversely proportional to heat storage.

Table 3.7. Average Daily Net Radiation, Heat Storage, and Priestley-Taylor Evaporation at Lake 5A for the 2007, 2008, and 2009 study years.

Year	Time Period		Net Radiation ($W.m^{-2}$)	Heat Storage ($W.m^{-2}$)	Priestley-Taylor Evaporation (mm)
2007	June 29	August 25	120	1	183
2008	June 29	August 25	107	-10	174
2009	June 29	August 25	116	8	150
Average			**114**	**0**	**169**

The model of evaporation outlined by Hargreaves and Samani (1982) was used to examine how historical climate variability and change has affected evaporation from the two primary study lakes over the historical study period. For the three study years (2007, 2008, and 2009), Total Hargreaves Evaporation for Inuvik was 276mm, 255mm, and 263mm (**Figure 3.21a**). Since the Hargreaves-Samani model for evaporation does not account for lake heat storage, evaporation was higher in 2009 than 2008. For all three study years, Total Hargreaves Evaporation was less than the historical average of 279mm, but not less than historical extremes. For 28 of the 52 historical study years, evaporation was greater than 2007. The minimum Total Annual Hargreaves Evaporation was 236 mm in 1960 and the maximum was 339 mm in 1958.

There was not a statistically significant trend in Total Annual Hargreaves Evaporation over the 52-year historical study period.

Total Annual Hargreaves Evaporation was notably less in Tuktoyaktuk than in Inuvik, due to cooler summer air temperatures. Over the 52-year historical study period, the average Total Annual Hargreaves Evaporation in Tuktoyaktuk was 165 mm. The minimum Total Annual Hargreaves Evaporation was 134 mm in 1959 and the maximum Total Annual Hargreaves Evaporation was 207 mm in 1989. Similar to Inuvik, there was not a statistically significant trend in calculated evaporation from the regional study lakes over the historical study period.

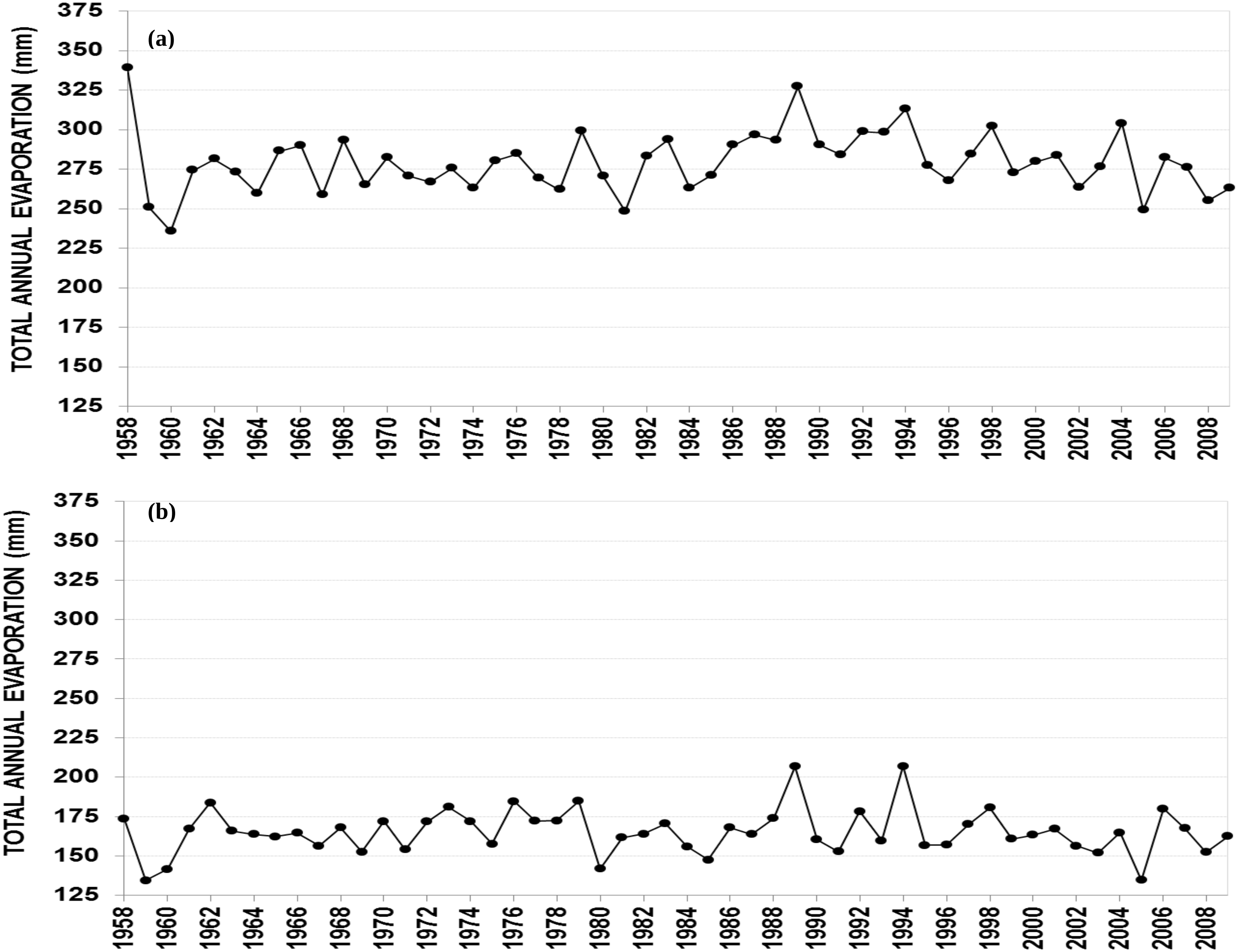


Figure 3.21. The Total Annual Hargreaves Evaporation for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009.

3.5.7 Vertical Water Balance

The vertical water balance of Inuvik in 2007, 2008, and 2009 was -60mm, 51mm, and 54mm, respectively (**Figure 3.22a**). In 2007, the vertical water balance was less than the historical average of 30mm. Although the vertical water balance in 2007 was not lower than the minimum vertical water balance of -95 mm in 1958, it was one of the lowest on record. Only 4 of the 52 historical study years had a vertical water balance that was lower than 2007. In 2008 and 2009, the vertical water balance was greater than the historical average, but not the highest on record. For 17 of the 52 historical study years, the vertical water balance was greater than 2009. The maximum vertical water balance was 219 mm in 1972, which is not entirely surprising given that 1972 received the most amount of precipitation of any year in the historical study period. There was not a statistically significant trend in the vertical water balance for Inuvik.

The vertical water balance for the regional study lakes was notably less than the two primary study lakes (**Figure 3.22b**). Over the 52-year historical study period, the average vertical water balance for Tuktoyaktuk was -3mm. In contrast with Inuvik, the vertical water balance for the regional study lakes at the northern end of the study region is typically negative, which suggests that these lakes are drying over time. The minimum vertical water balance was -98mm in 1958 and the maximum vertical water balance was 217mm in 1995. Similar to Inuvik, there was not a statistically significant trend in the vertical water balance for Inuvik.

It's important to note that, similar to the Annual May 1st Snowpack index, the Vertical Waterbalance index does not account for water lost to snow transport and sublimation, which could be significant. That being said, is does account for inter-annual variability in the vertical water balance (wet years vs. dry years).

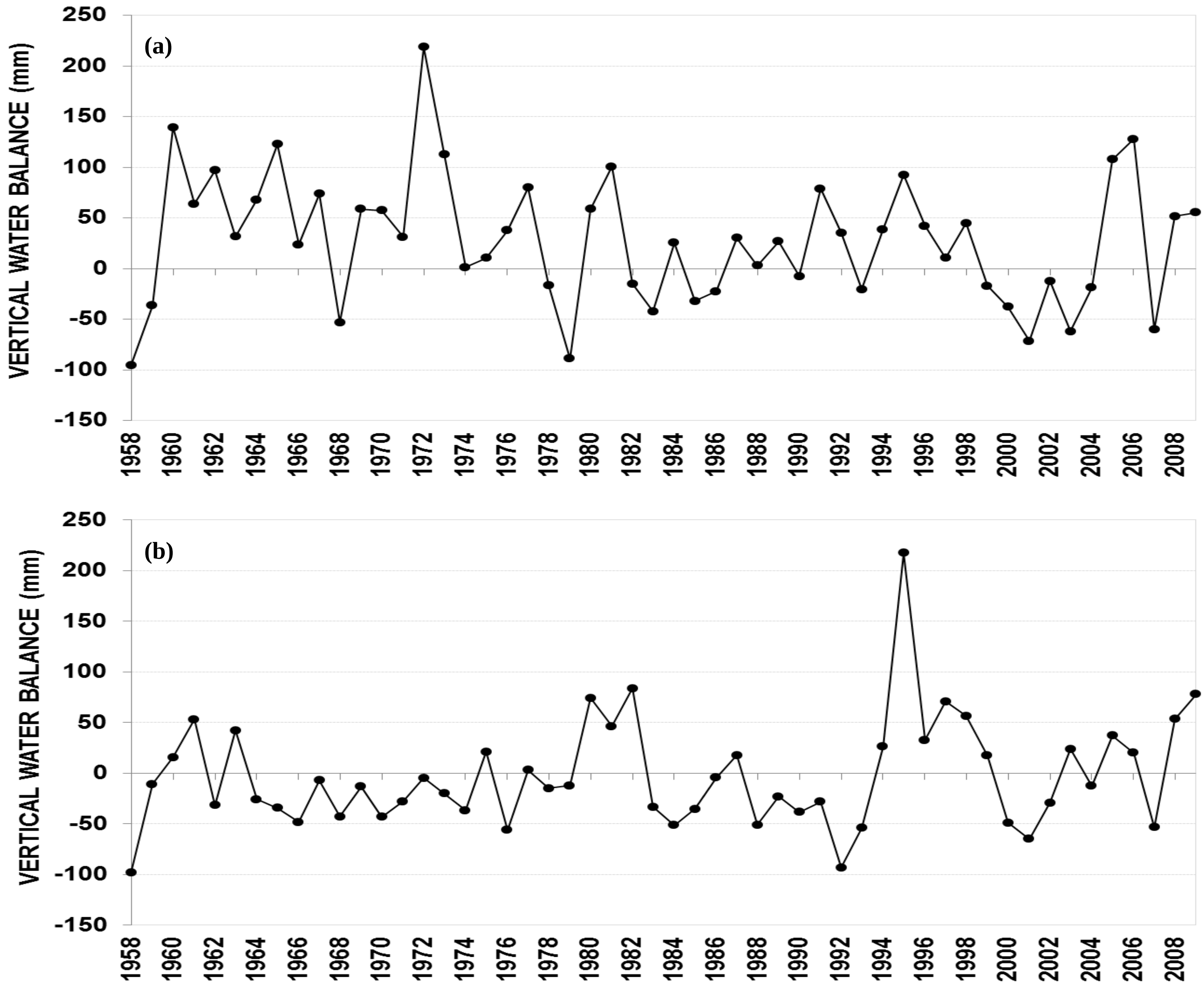


Figure 3.22. The vertical water balance for (a) Inuvik and (b) Tuktoyaktuk for the years 1958 to 2009.

3.6 Summary and Conclusions

The overall goal of this chapter was to investigate the historical impacts of climate variability and change (CVC) and SRTS on the key hydroclimatic factors and landscape-level features that drive the water balance of tundra lakes in the upland region east of the Mackenzie Delta. The primary factors influencing the water balance of the two primary study lakes included ice formation and ablation, spring snowmelt, snow and ice damming the outlet channel, open-water duration, evaporation, and summer rainfall events. The water balance of the two primary study lakes was typified by spring snowmelt. The Lake Level (LL) of Lake 5A and Lake 5B increased slightly over the winter months due to the weight of the overlying snow. At the onset of spring snowmelt, there was a rapid increase in LL that was amplified by the presence of snow and ice damming the outlet channel. As spring snowmelt progressed, lake water formed a trench through the snow and ice dam, initiating lake drainage and a rapid decline in LL. In the absence of summer rainfall events, the open-water period was typically a period of drying. In 2007, LL continued to decline over the open-water period, due to high cumulative evaporation and low cumulative rainfall. In 2008 and 2009, summer rainfall events led to high LL throughout most of July, August, and September.

Mean annual, winter, and spring air temperatures have been increasing at significant rates ($p < 0.05$) since the late 1950s, which has altered the timing and magnitude of a number of key hydroclimatic controls of the water balance of the two primary study lakes. Between the years 1958 and 2009, a statistically significant increasing trend in the annual May 1st snowpack for Tuktoyaktuk was observed. This has important implications for the water balance of the small tundra lakes located at the northern part of the study transect. More snowfall could lead to a greater rise in LL associated with spring snowmelt.

In addition to hydroclimatic factors, the water balance of the two study lakes is also influenced by landscape-level factors. This study suggests that SRTS-affected terrain accumulates snow over the winter months and has a greater SWE than the adjacent unaffected terrain. Overall, SRTS increases the SWE of the contributing lake catchment. Increases in the SWE of the contributing lake catchment associated with SRTS could lead to higher spring water levels that could influence the following summer and winter water levels.

Despite a significant decreasing trend in the timing of spring snow and ice melt in Tuktoyaktuk, no statistically significant trends in the length of the open-water period were observed for the study region over the historical study period. Furthermore, no significant trends in mean summer air temperature, total annual evaporation, or the vertical water balance were observed either. This suggests that more snowfall could lead to an overall increase in the LL of small tundra lakes in the northern part of the study region. Notably, SRTS and LL are key factors driving the rapid drainage of thaw lakes in the study region (Marsh and Neumann, 2001; Pohl et al., 2009).

The effects of recent climate warming appear to be amplified at Tuktoyaktuk, which is likely due to its close proximity to the Beaufort Sea. The rate of climate warming is typically greater in arctic coastal regions, relative to arctic inland regions, due to the rapid decline in sea ice extent in recent years (AMAP, 2012). This further emphasizes the need for more detailed, regionally specific water balance studies, to be used in the development of appropriate monitoring programs.

Overall, the results of this study suggest that recent climate warming and SRTS has modified key hydroclimatic processes, and these changes have directly affected the water balance of small tundra lakes in the study region. It's important to note that SRTS is an extreme form of permafrost degradation that only affects approximately 8% of lakes in the study region. Despite being unaffected by SRTS, the hydrology of other lakes in the study region has likely still been impacted by permafrost degradation. Active layer thickening, associated with recent climate warming in the region, increases the vertical infiltration and water residence times of runoff. For this reason, permafrost degradation in the study region will likely lead to increases in subsurface storage over the summer months. Changes in the hydrology of small tundra lake catchments, associated with recent climate warming and SRTS, has important implications for the geochemistry and ecology of small tundra lakes in the study region.

Chapter 4: Landscape-Level Factors Driving the Geochemistry of Small Tundra Lakes Affected by Shoreline Retrogressive Thaw Slumping

4.1 Introduction

The geochemistry of runoff in regions of continuous permafrost is driven by the hydrology of the contributing lake catchment, which is directly controlled by climatic factors (temperature and precipitation) and the presence of near-surface permafrost (Hinzman et al., 1991; Pohl et al., 2009; Quinton and Marsh, 1999; Quinton and Pomeroy, 2006). As highlighted in the preceding chapter, recent climate warming in the study region has modified a number of key hydroclimatic and landscape level factors, including mean winter and spring air temperatures, snow accumulation, the timing and intensity of spring snowmelt, and the depth of the active layer (AMAP, 2012; Bonsal and Prowse, 2003; Burn, 2008; Dibike et al., 2012). Given that water runoff is one of the primary inputs to the small tundra lake water balance, changes in the hydrology and geochemistry of runoff has important implications for the geochemistry and aquatic ecology of small tundra lakes.

The geochemistry of runoff typically exhibits strong seasonal variability that is typified by spring snowmelt. At the onset of spring snowmelt, melt water is rapidly delivered to runoff pathways via surface flow, which has relatively low concentrations of major ions and nutrients when compared with subsurface flow (Quinton and Pomeroy, 2006). The preceding chapter revealed increased snowfall and earlier spring freshet periods around Tuktoyaktuk since the late-1950's. This is consistent with other studies from the Northwestern Arctic (AMAP, 2012; Bonsal and Prowse, 2003; Burn, 2008; Lesack et al., 2013). A key question is: What impact will these hydroclimatic changes have on small tundra lakes?

With the progression of the summer months, the active layer thaws and the vertical infiltration and water residence time of runoff increases, allowing it to interact chemically with the soil profile. As a result, subsurface flow is dominated by major ions and nutrients derived from the organic and mineral layers comprising the active layer (Quinton and Pomeroy, 2006). Recent historical and projected future climate warming has and will continue to result in longer summers and accelerated permafrost degradation (i.e., active layer deepening and shoreline retrogressive thaw slumping) (AMAP, 2012; Bonsal and Prowse, 2003; Lantz and Kokelj, 2008; Smith et al., 2005). Studies suggest that permafrost degradation liberates major ions and nutrients to the overlying active layer, ultimately increasing the supply of major ions and nutrients to runoff and stream pathways (Hobbie et al., 1999; Keller et al., 2007; Kokelj and Burn, 2005; Kokelj et al., 2005). Reiterating the above question, what implications do these landscape level changes have on the geochemistry of runoff and subsequently, small tundra lakes?

Permafrost degradation has affected small tundra lakes in a number of ways, including overall active layer thickening, Shoreline Retrogressive Thaw Slumping (SRTS), and catastrophic drainage. SRTS has been used in a number of studies as a proxy for the potential effects of permafrost degradation on the geochemistry and biology of small tundra lakes (Kokelj et al., 2009b; Mesquita, 2008; Moquin, 2011; Moquin et al., 2012; Thompson, 2009; Thompson et al., 2012). Based on inter-lake comparisons, these studies found that lakes affected by SRTS typically have higher ionic concentrations than unaffected lakes (Kokelj et al., 2005; Kokelj et al., 2009b). Kokelj et al. (2009b) postulated that soluble ions are liberated from the near surface permafrost as it degrades, increasing the ionic concentration of SRTS-affected soils. Solutes then leach out of SRTS-affected soils in runoff, increasing the ionic concentration of the affected lake. It is important to note, however, that the relative roles of landscape-level hydrological processes in driving the geochemical response of small tundra lakes to SRTS is still largely unstudied.

The addition of charged particles from SRTS-affected terrain to adjacent lakes directly affects in-lake biological processes. Charged particles in the water column bind with dissolved organic matter and fall to the bottom of the lake, a process called sedimentation. Thompson et al. (2012) found that since the concentration of charged particles in SRTS-affected lakes is higher than that of unaffected lakes, the rate of sedimentation is also higher. Since organic matter is a significant source of nutrients to small tundra lakes in the study region, the concentration of nutrients in the water column of SRTS-affected lakes is lower than that of unaffected lakes. This is in line with the work of Moquin et al. (2014), who found that the macroinvertebrate community of SRTS-affected lakes was significantly different from that of unaffected lakes, due to elevated nutrient concentrations in their bottom sediments. Increases in the sedimentation of dissolved organic matter, associated with SRTS, affect other key biological processes. For instance, SRTS-affected lakes are typically clearer than unaffected lakes, increasing the penetration of photosynthetically active radiation into the water column. In summary, understanding the effects of recent climate change and SRTS on the geochemistry of runoff to small tundra lakes is crucial, because it has a number of implications for aquatic ecology.

4.2 Purpose and Objectives

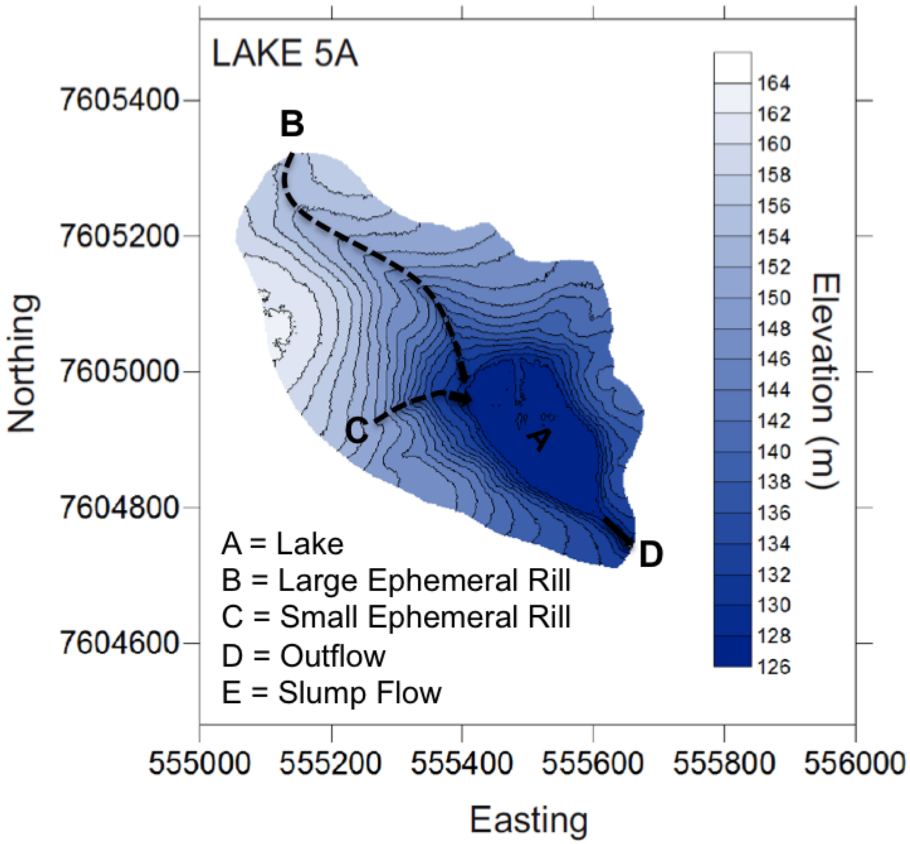
The purpose of this study is to evaluate the role of runoff in driving the geochemical response of Mackenzie Upland tundra lakes to shoreline retrogressive thaw slumping (SRTS). Two specific objectives are addressed:

i. Examine the seasonal geochemistry of the runoff pathways leading to and from a pair of shallow tundra lakes in the upland region adjacent to the Mackenzie Delta.
ii. Examine the geochemistry of the runoff pathways leading to and from 10 regional small tundra lakes located in the northern part of the upland region adjacent to the Mackenzie Delta and Richard's Island.

4.3 Methodology

This work was a part of a larger International Polar Year/ArcticNet project investigating the *Hydro-ecological responses of Arctic tundra lakes to climate change and landscape perturbation.* In 2008 and 2009, spatio-temporally detailed geochemical signature surveys were conducted by Environment Canada at two small tundra lake catchments (Lake 5A: Control; Lake 5B: Affected by SRTS). For the ice-on, spring melt, and open-water periods, water samples were collected from the lake centre (at a depth of 0.5 m), the primary and secondary inflow channels, an ephemeral rill running off of the SRTS at Lake 5B (referred to as slumpflow from here on), and the outflow channel (**Figure 4.1**). Limited spatio-temporally sampled water geochemistry data were also available for the years 2006, 2007, and 2010 at the two study sites. Additional geochemical data were available for 9 regional small tundra lake catchments (6 affected by SRTS; 3 unaffected). For the open-water period of the years 2006 through to 2008, water samples were collected by Environment Canada from the lake centre (at a depth of 0.5 m). Supplementary data were available for the ice-on and spring melt periods of the years 2006 and 2007. For the open-water period of the years 2009 and 2010, water samples were collected from the lake centre, as well as the primary inflow, the slumpflow at affected lakes, and the outflow channel at the 9 regional small tundra lake catchments, as well as at 1 additional small tundra lake catchment located within the YaYa lake subcatchment (referred to as YaYa Sub from here on).

Water samples were collected using clean, pre-labelled bottles. The samples were stored in chilled coolers, with freezer packs, and transported back to the Aurora Research Institute in Inuvik, NT. Within 24 hours of their retrieval, the samples were transported, via air, to the National Laboratory for Environmental Testing (Burlington and Saskatoon). Samples were analyzed for Ca^{2+}, Cl^{-}, K^{+}, Na^{+}, Mg^{2+}, SO_4^{2-}, Total Phoshorus (TP), Total Nitrogen (TN), and Total Dissolved Nitrogen (TDN). Note that at the 9 regional study lakes and the YaYa Sub lake, the samples collected from the inflow, slumpflow, and outflow channels were not analyzed for TP, TN, and TDN. All of the samples were collected, filtered, preserved, and analysed using standard protocols, as outlined by Environment Canada (2009).



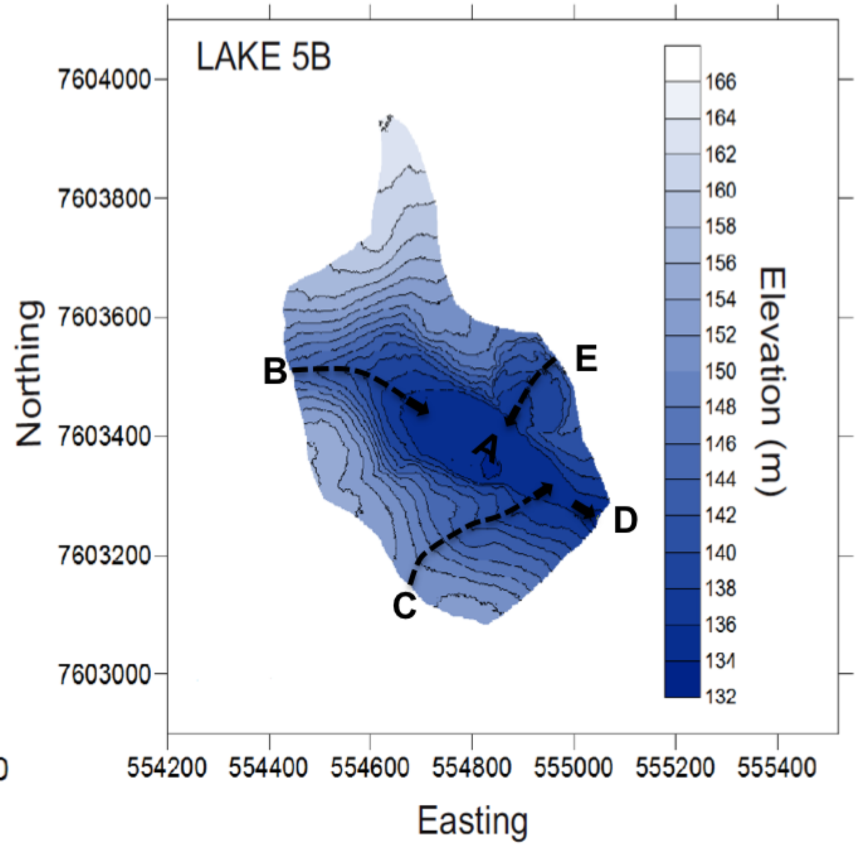


Figure 4.1. A map indicating the location of the major water sources, leading to and from Lake 5A and 5B, from which samples were collected.

4.4 Statistical Analyses

Prior to performing the following statistical analyses, all of the geochemistry data were logarithmically transformed to yield a normally distributed, homoscedastic dataset. All analyses were performed using SPSS Version 21 (IBM Corp., 2012).

4.4.1 Study Lake 5A and 5B

As with a one-way ANOVA, a repeated measures ANOVA tests the equality means among a number of groups. In contrast, however, it assumes that the same subject(s) makes up each group. In other words, a repeated measures ANOVA is used to test the same subject(s) across a number of different conditions. This is referred to as a “within-subject factor”. A mixed methods ANOVA, such as the one used here, can test both a “within-subject factor and a “between-subject factor”. A “between-subject factor” refers to two subjects that make up different groups.

A repeated measures ANOVA with a between-subject factor was used to test whether the geochemistry of the inflow channels, lake, and outflow channels:

1) varied significantly between seasons.
2) was significantly affected by SRTS.

The *dependent variables* included Major Ions (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, SO_4^{2-}) and Nutrients (Total Phosphorus, Total Nitrogen, and Total Dissolved Nitrogen). The *independent variables* included Season (within-subject factor; ice-on vs. spring melt and early open-water vs. mid to late open-water), Lake Type (between-subject factor; 5A vs. 5B), Catchment Flow (between subject factor; 5A Inflow vs. 5B Inflow vs. 5B Slumpflow), and Outflow Channel (between subject factor; 5A Outflow Channel vs. 5B Outflow Channel.

A repeated measures ANOVA with a between subject factor also tests for what is commonly referred to as an “Interaction Effect”. If a significant Interaction Effect is present, the relationship between the within-subject factor (Season) and the dependent variable (Major Ions and Nutrients) is dependent on the between-subject factor (Lake Type, Catchment Flow, and Outflow Channel) and vice versa.

4.4.2 Regional Study Lakes

A repeated-measures ANOVA with between-subject factors was used to test whether the geochemistry of the 9 regional study lakes:

1) varied significantly between year.
2) varied significantly between seasons.
3) was significantly affected by SRTS.

The *dependent variables* included Major Ions (Ca^{2+}, Cl^-, K^+, Mg^{2+}, Na^+, SO_4^{2-}) and Nutrients (Total Phosphorus, Total Nitrogen, and Total Dissolved Nitrogen). The *independent variables* included Year (within subject factor; 2006 to 2010), Season (within subject factor; ice-on vs. spring melt vs. open-water), and Lake Type (between-subject factor; unaffected terrain vs. SRTS-affected terrain).

For the 9 regional study lakes and the YaYa Sub lake, an independent-samples t-test was used to test whether the geochemistry of the inflow and outflow channels was significantly affected by SRTS. The *dependent variables* included Major Ions (Ca^{2+} , Cl^- , K^+ , Mg^{2+} , Na, SO_4). The *independent variable* was Shoreline Retrogressive Thaw Slumping (unaffected vs. SRTS-affected).

4.5 Results and Discussion

4.5.1 Lake Water

4.5.1.1 Lake 5A and 5B

Major Ions

In 2008 and 2009, detailed geochemical signature surveys were carried out at the major water sources to and from Lake 5A and Lake 5B. The ionic concentration of Lake 5A exhibited strong seasonal variability over the two study years, which was correlated with key hydrological events (**Figure 4.2a**). The ionic concentration of Lake 5A increased over the ice covered months, reaching a peak prior to spring snowmelt. This is attributed to the processes associated with ice formation. As lake-ice forms, the solutes that are initially held in surface water are excluded, increasing the ionic concentration of lake water (Hobbie, 1980; Lesack et al., 1991).

At the onset of spring snowmelt, there was a notable decrease in the ionic concentration of the lake water that corresponded with the annual spring freshet peak in LL. This is attributed to dilution by spring snowmelt water, which was delivered to the lake from the ice surface and surrounding landscape. The ionic concentration of the primary and secondary inflows was lower during and immediately following spring melt than during the mid to late open-water period (**Figure 4.2b and Figure 4.2c**). This agrees with the results of Quinton and Pomeroy (2006), who found that seasonal variability in the ionic concentration of runoff at Siksik Creek, located near the two study lakes, is typically controlled by snowmelt. During the spring melt and early open-water period, the majority of the active layer is still frozen, which limits the vertical infiltration and residence time of water runoff. Ultimately, the status of the active layer inhibits water runoff from interacting chemically with the soil profile. As a result, the ionic concentration of runoff is typically more dilute during spring melt and early open-water, relative to the mid to late open-water period. Notably, the ionic concentration of the primary and secondary inflow to Lake 5A was higher at the beginning of spring snowmelt and decreased as spring snowmelt progressed. This is consistent with the observations of Quinton and Pomeroy (2006), who found that the ionic concentration of runoff increased during the initial stages of spring snowmelt, but decreased as snowmelt progressed due to the dilution of runoff pathways. Overall, the addition of relatively dilute runoff from the contributing lake catchment led to a temporary decrease in the ionic concentration of Lake 5A.

In the absence of heavy rainfall events, the concentration of Ca^{2+}, Cl^{-}, K^{+}, Na^{+}, Mg^{2+} , and SO_4^{2-} in Lake 5A increased over the open-water period. Increases in the importance of subsurface flow relative to surface flow, evaporation from the lake surface, and lake mixing are key processes that would have contributed to this increase. As the active layer thaws, the vertical infiltration and residence time of water runoff increases, encouraging water runoff to interact chemically with the mineral soils that make up the soil profile. As a result, the ionic concentration of runoff increases, subsequently increasing the supply of soluble ions to associated lakes (Hinzman et al., 1991; Quinton and Marsh, 1999; Quinton and Pomeroy, 2006). As highlighted in Chapter 3, total Hargreaves Evaporation was 255mm and 263mm in 2008 and 2009, respectively. Taking into account the surface area and volume of Lake 5A, this equates to approximately 5% of the total lake volume. The loss of this water contributed to the enrichment of lake ions. As water is lost through evaporation over the course of the summer months, the major ions in the lake typically become more and more concentrated.

There were some points during the open-water period when the concentrations of Ca^{2+}, Cl^{-}, K^{+}, and Na^{+} decreased. These periods of decline can be attributed to dilution via rainfall. In 2009, the concentration of Ca^{2+}, Cl^{-}, K^{+}, and Na^{+} in the inflows to Lake 5A decreased between July 14th and September 25th in response to the addition of 162mm of rainfall (**Figure 4.2 a and b**). The addition of relatively diluted runoff led to a decrease in the concentration of Ca^{2+}, Cl^{-},

K^+, and Na^+ in the lake water. Notably, summer rainfall appears to be an important factor driving the geochemistry of Lake 5A that needs to be explored further.

The concentration of Ca^{2+}, K^+, Mg^{2+}, Na^+, and SO_4^{2-} was notably higher in Lake 5B than in Lake 5A (**Figures 4.3a**). This agrees with the work of Kokelj et al. (2005; 2009b), who found that the ionic concentration of SRTS-affected lakes was typically higher than that of unaffected lakes. By comparison, there was not an obvious difference in the concentration of Cl^- in Lake 5B, as compared to Lake 5A. This agrees with the work of Quinton and Pomeroy (2006), Kokelj et al. (2009b) and Pienitz et al. (1997), who found that the concentration of Cl^- in small tundra lakes in the Mackenzie Delta Uplands and Richard's Island was strongly driven by proximity to the Beaufort Coast and weakly driven by landscape-level processes, such as SRTS. They suggested that Cl^- was likely derived from marine aerosols, which are transported from the Beaufort Sea via precipitation.

Overall, seasonal variability in the ionic concentration of Lake 5B was similar to that of Lake 5A. The ionic concentration of Lake 5B was the highest during the ice-covered period and was the lowest during the spring snowmelt and early open-water period. In contrast with Lake 5A, however, the concentrations of all major ions in Lake 5B increased over the entire open-water period, with no evident dilution via rainfall.

In general, the trends observed at Lake 5A and Lake 5B in 2008 and 2009 correspond with the trends observed in the less spatio-temporally detailed sampling conducted at Lake 5A and Lake 5B in 2006, 2007, and 2010. Data were grouped based on Lake Type (Lake 5A and Lake 5B) and Season (ice-covered, spring melt and early open-water, and mid to late open-water) (**Figure 4.4**, **Figure 4.5**, and **Figure 4.6**). To summarize, the mean concentrations of Ca^{2+}, Cl^-, K^+, Mg^{2+}, Na^+, and SO_4^{2-} were significantly greater in Lake 5B than Lake 5A (ANOVA, $p < 0.05$, **Table 4.1**). The mean concentrations of Ca^{2+}, Cl^-, K^+, Na^+, Mg^{2+}, and SO_4^{2-} in Lake 5A and Lake 5B also varied significantly across the three hydrological periods. Lastly, there was a significant interaction effect between Lake Type and Season for the dependent parameter Cl^-, suggesting that the influence of SRTS on the concentration of Cl^- is dependent on Season and vice versa.

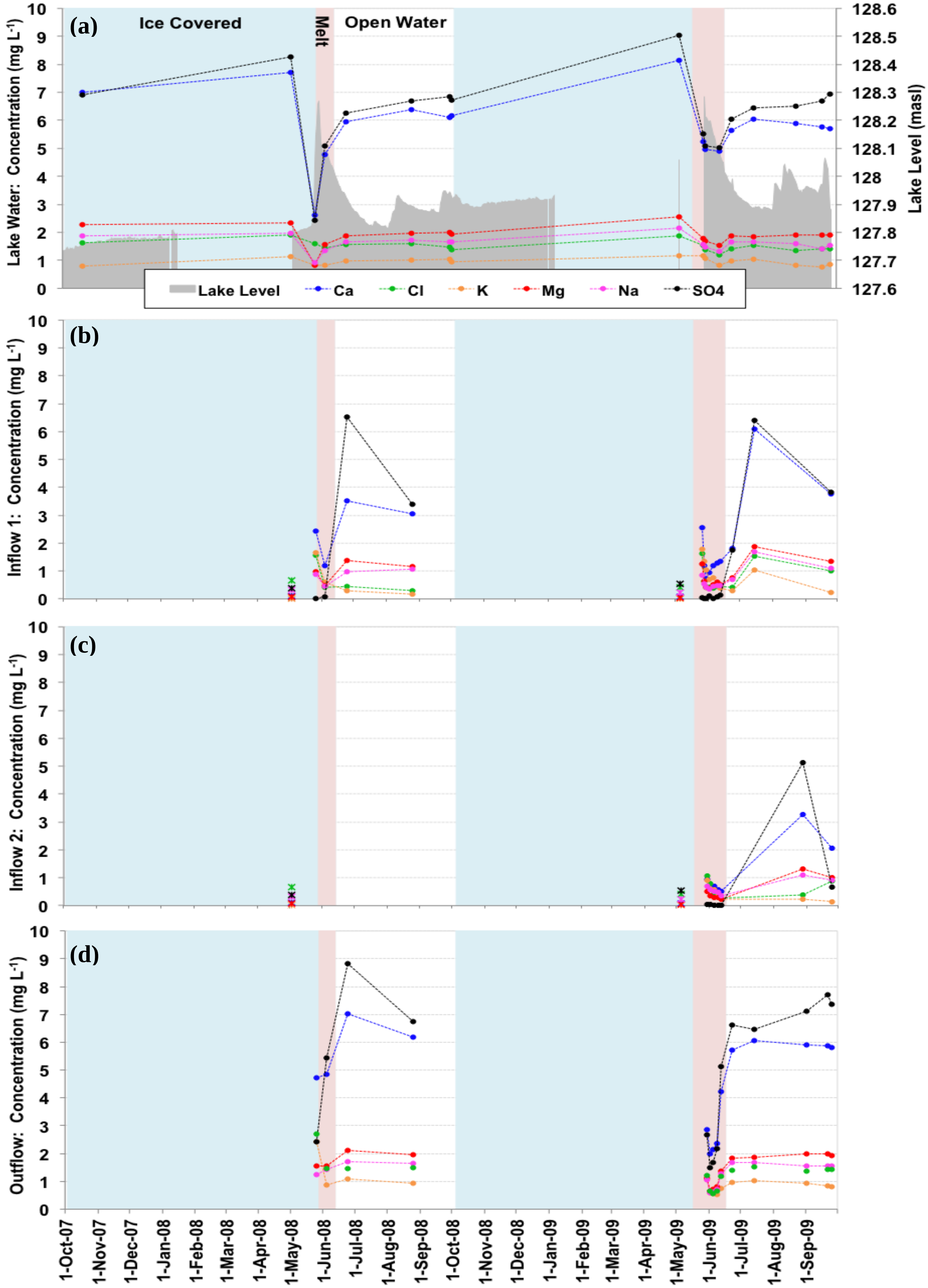


Figure 4.2. The concentrations of Ca^{2+}, Cl^-, K^+, Mg^{2+}, Na^+, and SO_4^{2-} in the (a) lake water, (b) inflow 1, (c) inflow 2, and (d) outflow at the Lake 5A study catchment over the 2008 and 2009 study years. The dots represent water and the stars represent snow. Also presented here is the corresponding Lake Level.

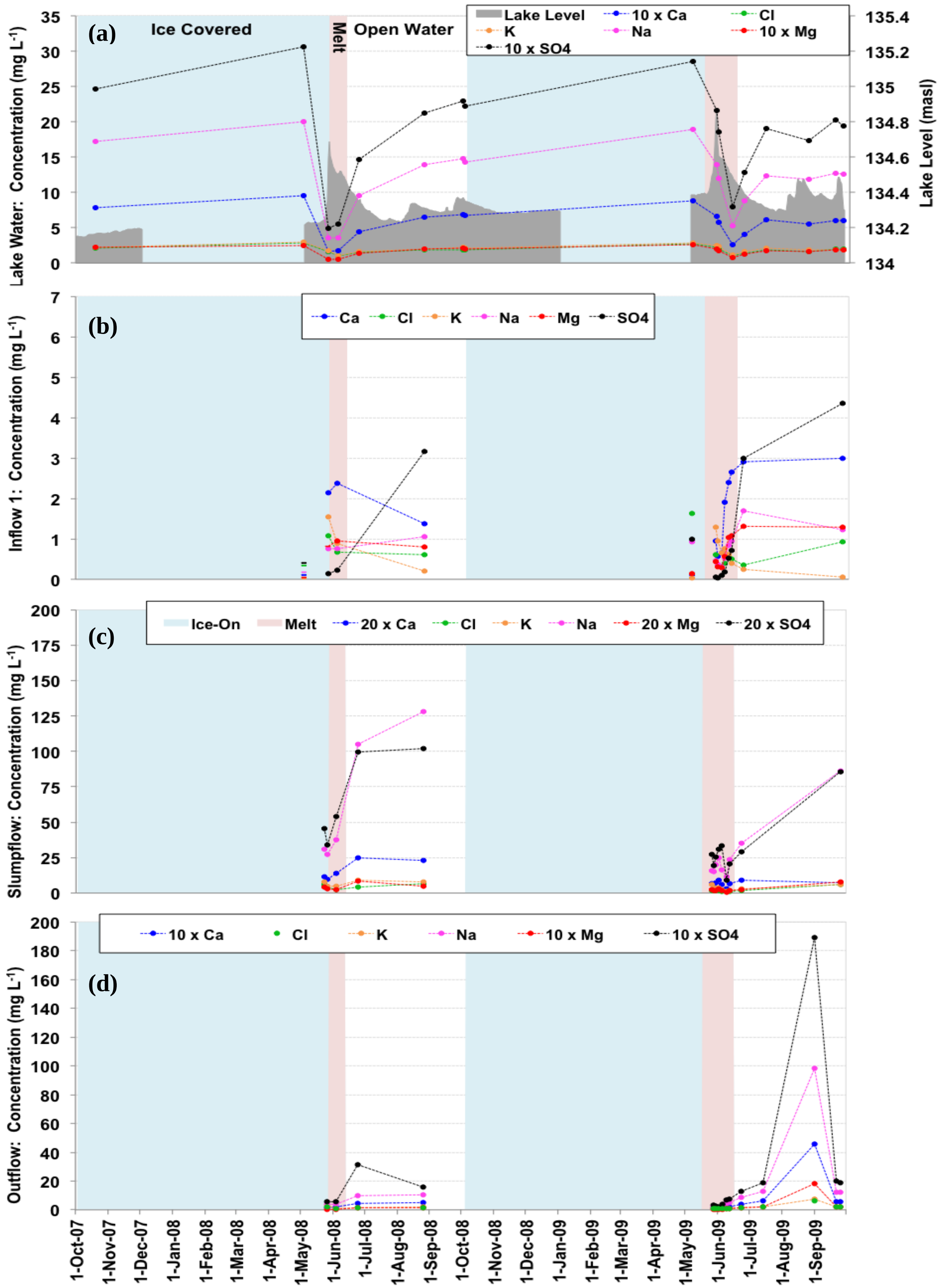


Figure 4.3. Lake 5B - The concentrations of Ca^{2+}, Cl^{-}, K^{+}, Na^{+}, Mg^{2+} and SO_4^{2-} in the (a) lake water, (b) Inflow 1, (c) Slumpflow, and (d) Outflow at the Lake 5B study catchment over the 2008 and 2009 study years. The dots represent water and the solid lines represent snow. Also presented here is the corresponding Lake Level.

Table 4.1. An ANOVA table for the parameters measured in the lake water obtained from Lake 5A and Lake 5B in the years 2006 to 2010. The dependent parameters (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) were tested according to Lake Type (5A vs. 5B) and Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameters	**Lake Type (5A vs. 5B)**			**Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water)**			**Lake Type * Season**		
	F	**df**	**P**	**F**	**df**	**P**	**F**	**df**	**P**
Ca^{2+}	279.974	1.000	**0.000**	14.347	1.085	**0.007**	1.775	1.085	0.230
Cl^{-}	20.896	1.000	**0.004**	35.440	2.000	**0.000**	7.254	2.000	**0.009**
K^{+}	173.118	1.000	**0.000**	17.746	2.000	**0.000**	3.197	2.000	0.077
Mg^{2+}	239.701	1.000	**0.000**	14.333	1.024	**0.009**	1.903	1.024	0.216
Na^{+}	226.010	1.000	**0.000**	19.632	1.064	**0.004**	4.047	1.064	0.087
SO_4^{2-}	451.420	1.000	**0.000**	15.191	1.033	**0.007**	1.575	1.033	0.256

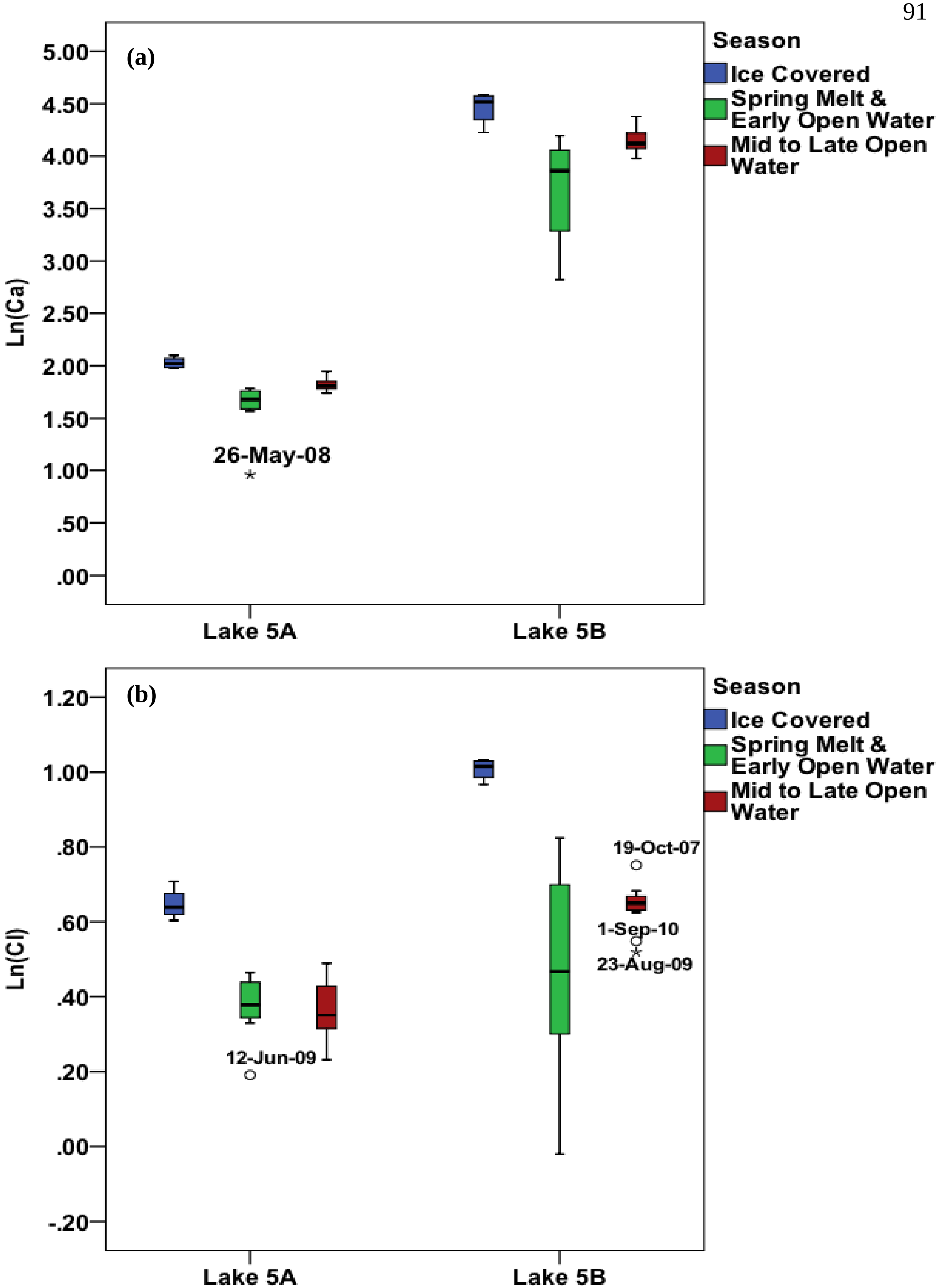


Figure 4.4. Box plots displaying the concentration of (a) Ca^{2+} and (b) Cl^{-} in Lake 5A and Lake 5B. The concentration of Ca^{2+} and Cl^{-} is grouped based on Lake Type (5A vs. 5B) and categorized based on Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

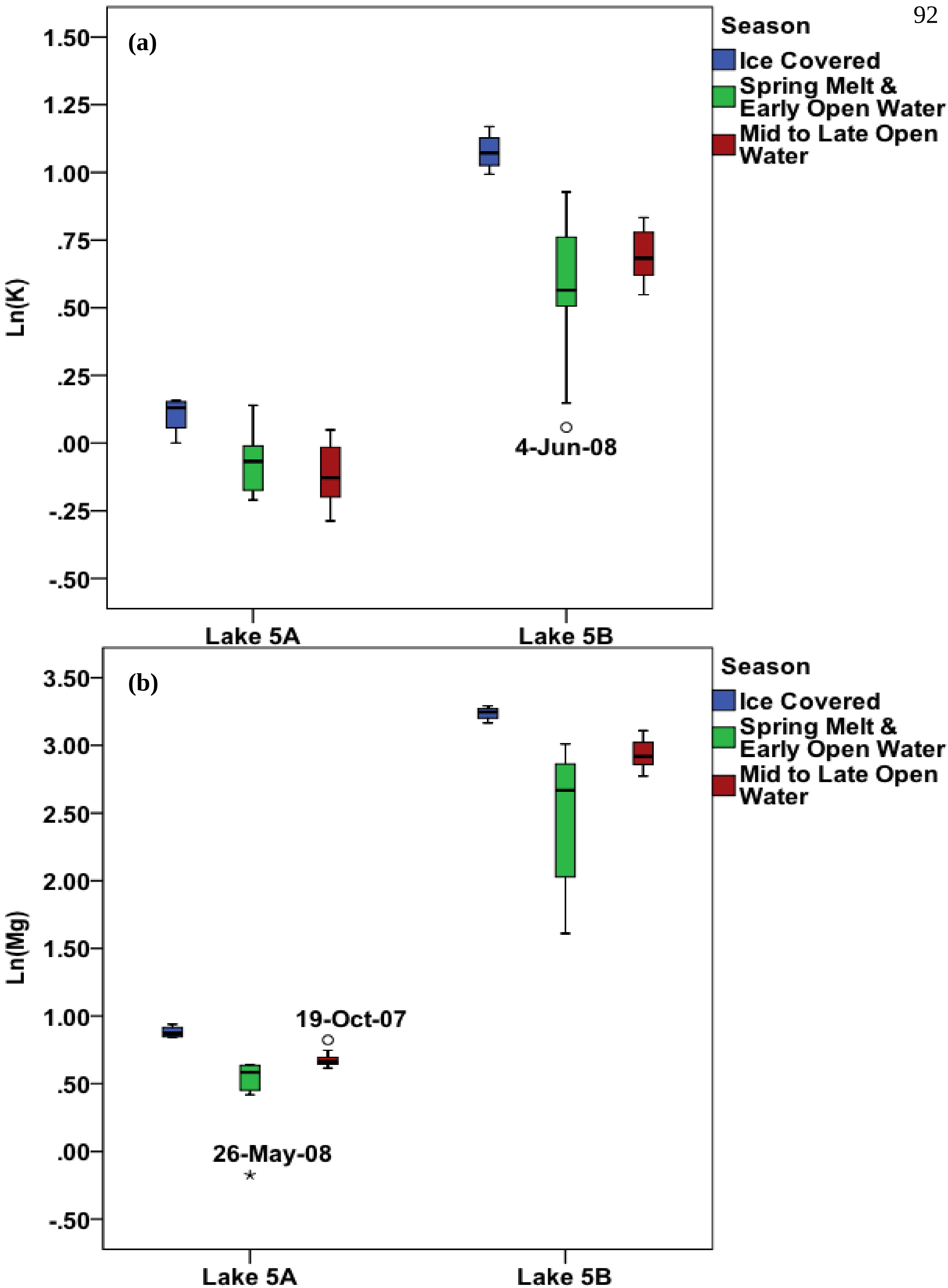


Figure 4.5. Box plots displaying the concentration of (a) K^+ and (b)Mg^{2+} in Lake 5A and Lake 5B. The concentration of Mg^{2+} and K^+ is grouped based on Lake Type (5A vs. 5B) and categorized based on Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

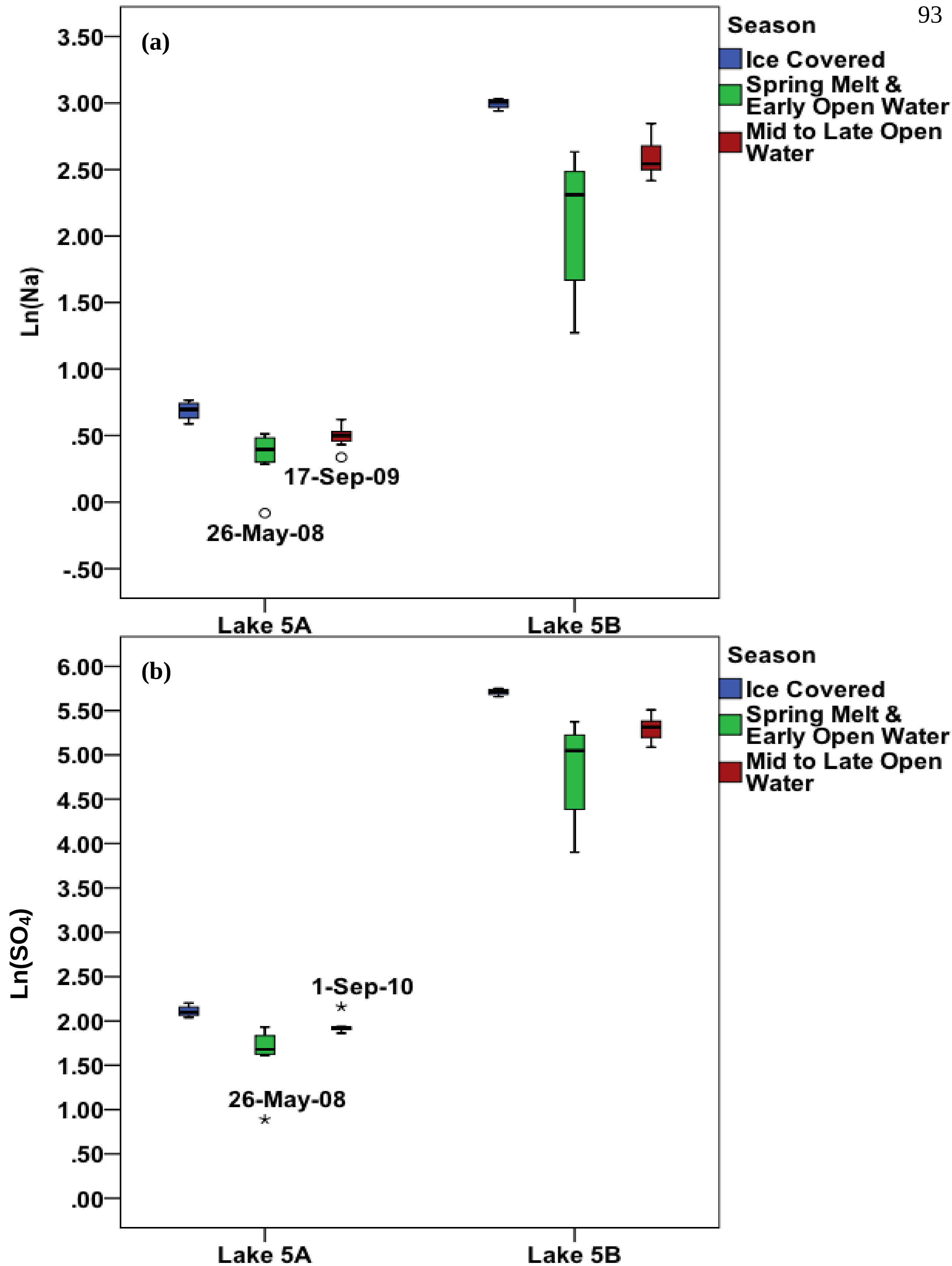


Figure 4.6. Box plots displaying the concentration of (a) Na^{+} and (b) SO_4^{2-} in Lake 5A and Lake 5B. The concentration of Na and SO_4 is grouped based on Lake Type (5A vs. 5B) and categorized based on Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

Nutrients

Over the 2008 and 2009 study years, the nutrient concentration of Lake 5A exhibited some seasonal variability (**Figure 4.7**). The concentration of Total Phosphorus (TP), Total Nitrogen (TN), and Total Dissolved Nitrogen (TDN) increased over the winter months. This is attributed to the freeze-out associated with ice formation (Hobbie, 1980; Lesack et al., 1991).

The concentration of TP, TN, and TDN in Lake 5A and Lake 5B increased at the beginning of spring snowmelt (late-May). This is attributed to the addition of nutrient-rich runoff. Spring snowmelt initiates runoff, which can liberate nutrient-rich surficial material from the contributing lake catchment and act as a source of Phosphorus and Nitrogen to small arctic lakes (Quinton and Pomeroy, 2006; Johnson and Luecke, 2012). As a result, the concentrations of TP, TN, and TDN of the inflow channels leading to Lake 5A and Lake 5B were higher during the early stages of spring snowmelt than during the later stages. As spring snowmelt progressed, the concentration of TP, TN, and TDN in the inflows and the lakes decreased. With the progression of spring snowmelt, the soil organic layer becomes saturated, forcing runoff to move on top of the soil organic layer rather than through it, causing the concentration of nutrients in runoff to decrease (Quinton and Pomeroy, 2006).

In general, the concentration of TP in Lake 5A and Lake 5B decreased over the open-water period. This is partially attributed to dilution via rainfall inputs. For instance, the concentration of TP in both the inflow and the lake decreased between June 24th and August 25th, 2008, in response to the addition of 80mm of rainfall onto the contributing lake catchment. Similarly, the concentration of TP in the inflow and the lake decreased between June 24th and August 23rd, 2009, in response to the addition of 104 mm of rainfall. The decrease of TP over the open water period is also partially attributed to higher rates of sedimentation and increases in biological uptake. In the previous section, it was noted that the concentration of charged particles in the water column increased over the summer months, due to the addition of relatively ion-rich runoff water from the contributing lake catchment. In turn, sedimentation rates within the lake water column would likely increase as well. Thompson et al. (2008) proposed that charged particles (clay and cations/anions), released into the lake water column from the contributing lake catchment, bind with humic material, causing it to become heavy and settle to the bottom. This is consistent with the results of Moquin (2011). Moquin (2011) found that the concentration of

TP in the water column of Lake 5B decreased over the open water period in 2010, which was partially attributed to increases in sedimentation rates.

In contrast with TP, the concentration of TN and TDN increased over the open water period. This is attributed to the addition of nutrient-rich run-off. The concentration of TN and TDN in the inflows and the lakes increased during periods of heavy rainfall in 2008 and 2009. Runoff, initiated by heavy rainfall events, typically liberates nitrogen from the contributing lake catchment during the summer months. For instance, Johnson and Luecke (2012) found that Nitrogen deficiency in Toolik Lake, Alaska, typically decreased during summers that had higher precipitation. They proposed that increases in precipitation, associated with projected climate warming, could lead to increases in nutrient loading to freshwater systems in the Northwestern Arctic.

The concentration of TP in runoff from SRTS-affected terrain was higher than that of runoff from the adjacent unaffected terrain, suggesting that SRTS increases the supply of TP to runoff. This is consistent with other studies from the Northwestern Arctic, which suggest that thermokarst activity could release stored nutrients into adjacent arctic freshwater systems (Hobbie et al. 1999;Keller et al. 2007). Despite increasing the TP concentration of runoff, SRTS does not appear to increase the concentration of TP in Lake 5B. This is in line with the work of Thompson et al. (2012), who found that SRTS-affected lakes had significantly lower concentrations of TP than unaffected lakes. As was mentioned earlier, this was partially attributed to higher rates of sedimentation associated with higher concentrations of charged particles in the water column. SRTS does not appear to affect the concentration of TN and TDN in runoff or lake water at Lake 5B.

The responses observed at Lake 5A and Lake 5B in 2008 and 2009 partially correspond with the trends observed in the less spatio-temporally detailed sampling conducted at Lake 5A and Lake 5B in 2006, 2007, and 2010. The nutrient data collected in the years 2006 to 2010, were grouped based on Lake Type (Lake 5A vs. Lake 5B) and Season (ice-on vs. spring melt and early open-water vs. mid to late open-water). To summarize, the mean concentrations of TP, TN, and TDN in Lake 5A were not significantly different from Lake 5B (ANOVA, $p > 0.05$, **Table 4.2, Figures 4.9** and **4.10**). The mean concentration of TP in Lake 5A and Lake 5B varied significantly across the three hydrological seasons ($p < 0.05$). On average, the concentration of TP in Lake 5A was the highest during spring snowmelt and the lowest during the open-water

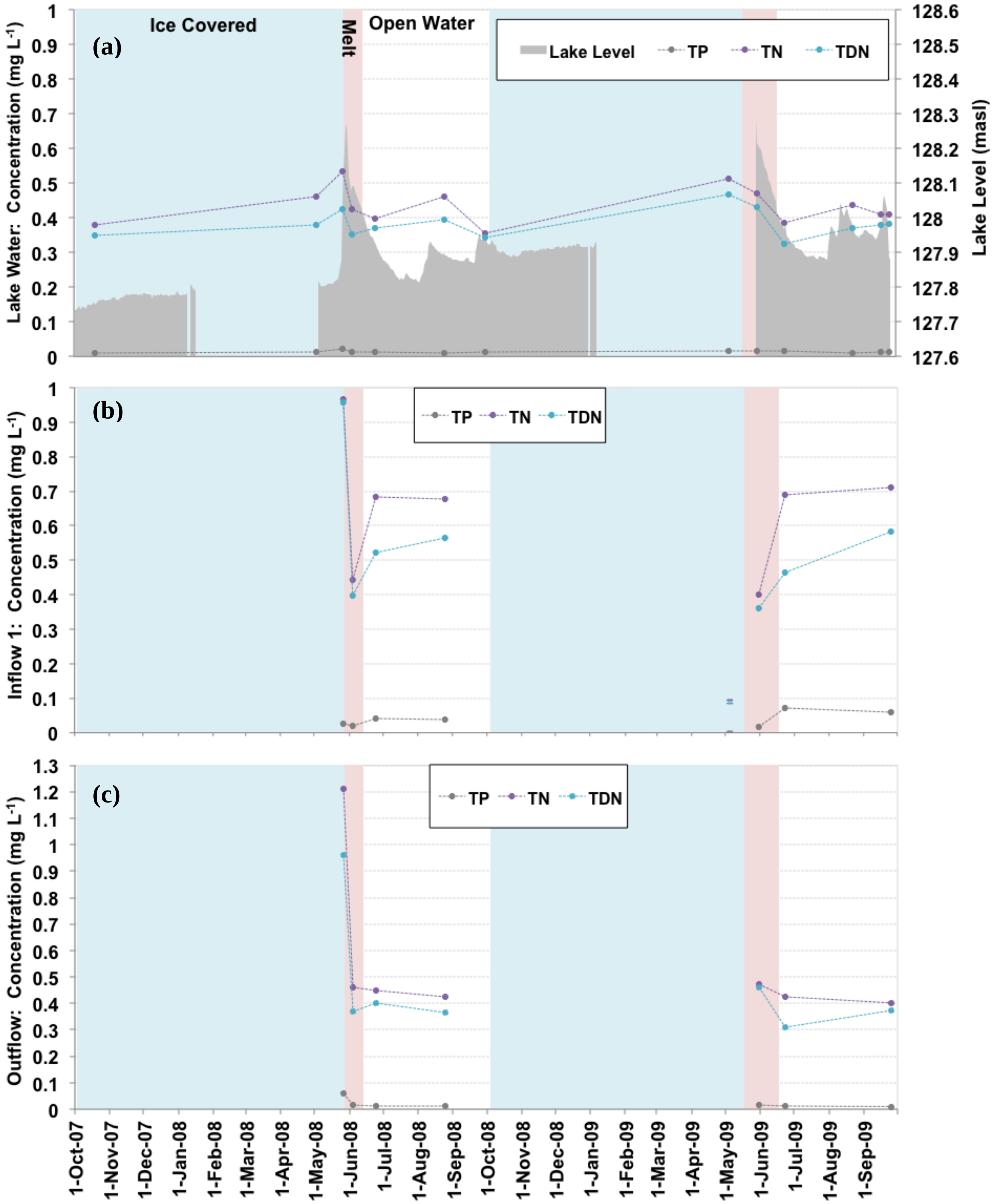


Figure 4.7. The concentration of Total Phosphorus (TP), Total Nitrogen (TN), and Total Dissolved Nitrogen (TDN) in the (a) lake water, (b) inflow 1, and (c) outflow at Lake 5A over the 2008 and 2009 study years. The dots represent water and the solid lines represent snowpack. Also presented here is the corresponding Lake Level.

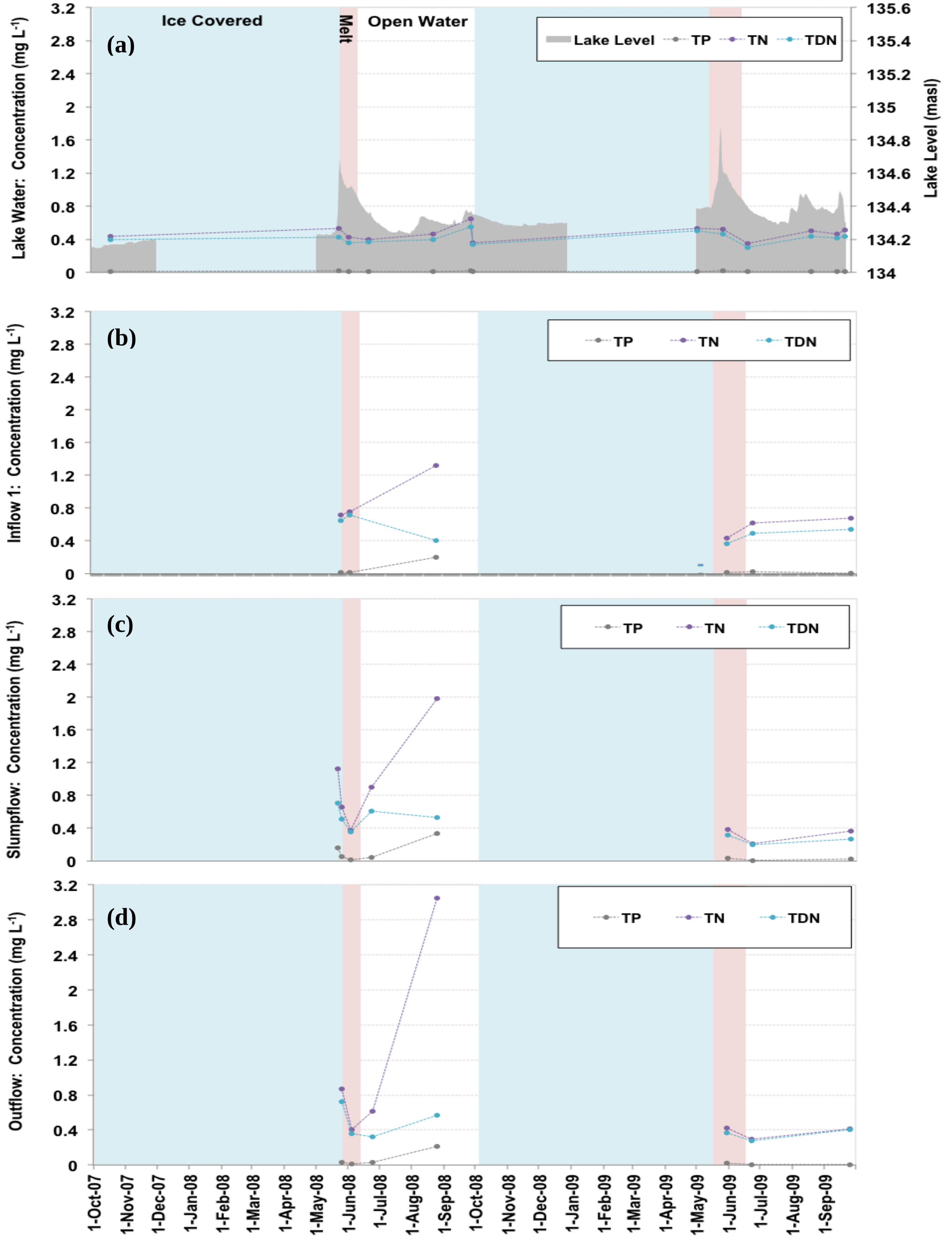


Figure 4.8. The concentration of Total Phosphorus (TP), Total Nitrogen (TN), and Total Dissolved Nitrogen (TDN) in the (a) lake water, (b) inflow 1, (c) slumpflow, and (d) outflow at Lake 5B over the 2008 and 2009 study years. The dots represent water and the solid lines represent snowpack. Also presented here is the corresponding Lake Level.

Table 4.2. An ANOVA table for the parameters measured in lake water obtained from Lake 5A and Lake 5B in 2006 to 2010. The dependent parameters (Total Phosphorus (TP), Total Nitrogen (TN), and Total Dissolved Nitrogen (TDN)) were tested according to Lake Type (5A vs. 5B) and Season (Ice Covered vs. Spring Melt and Early Open-water vs. Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameters	**Lake Type (5A vs. 5B)**			**Season (Ice Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water)**			**Lake Type * Season**		
	F	**df**	**P**	**F**	**df**	**P**	**F**	**df**	**P**
TP	2.288	1.000	0.205	6.123	2.000	**0.024**	1.168	2.000	0.359
TN	0.002	1.000	0.969	2.805	2.000	0.173	1.009	2.000	0.442
TDN	0.763	1.000	0.432	1.327	2.000	0.318	0.254	2.000	0.782

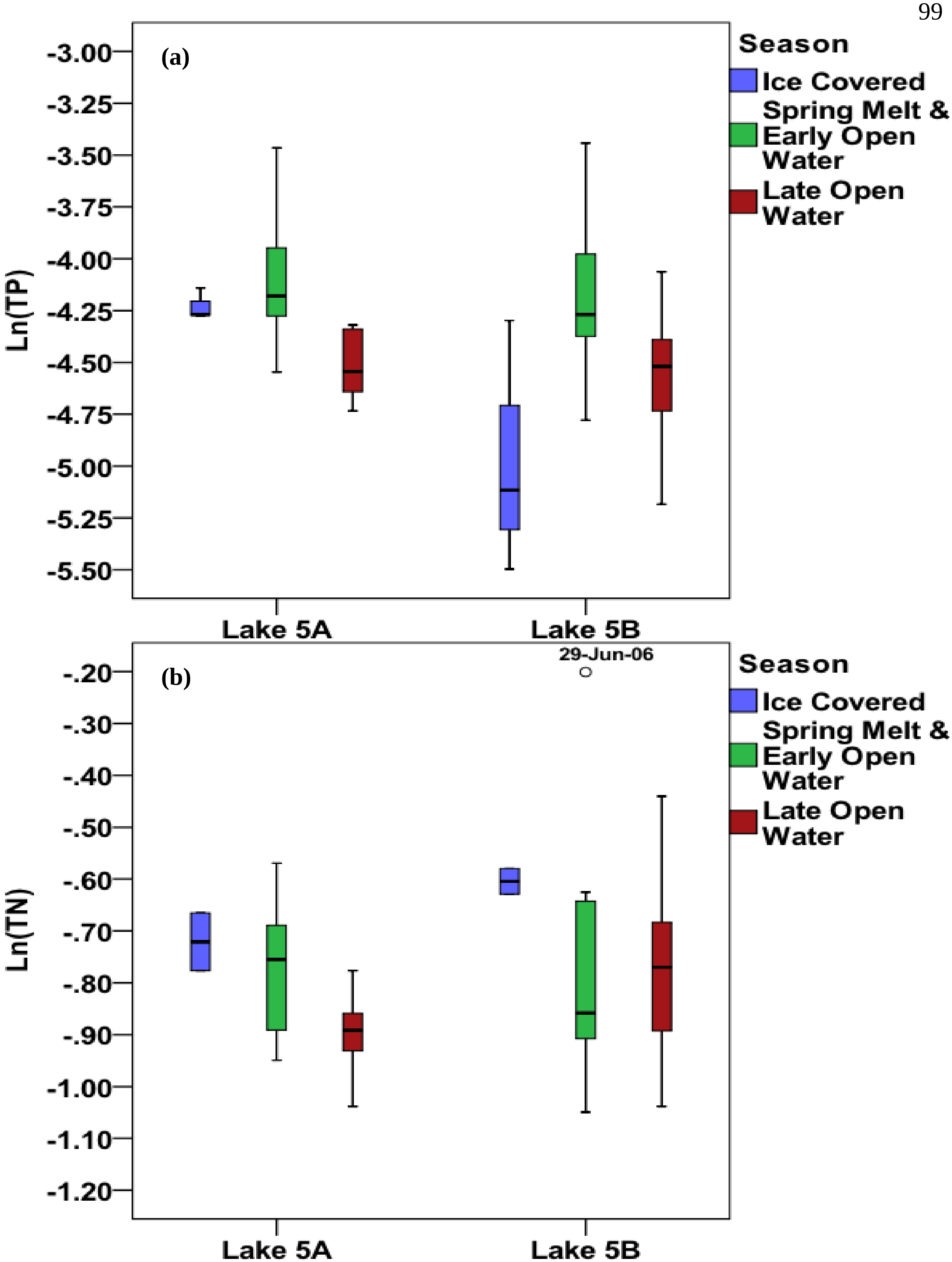


Figure 4.9. Box Plots displaying the concentration of (a) Total Phosphorus (TP) and (b)Total Nitrogen (TN) in Lake 5A and Lake 5B. The concentration of TP and TN is grouped based on Lake Type (5A vs. 5B) and categorized based on Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

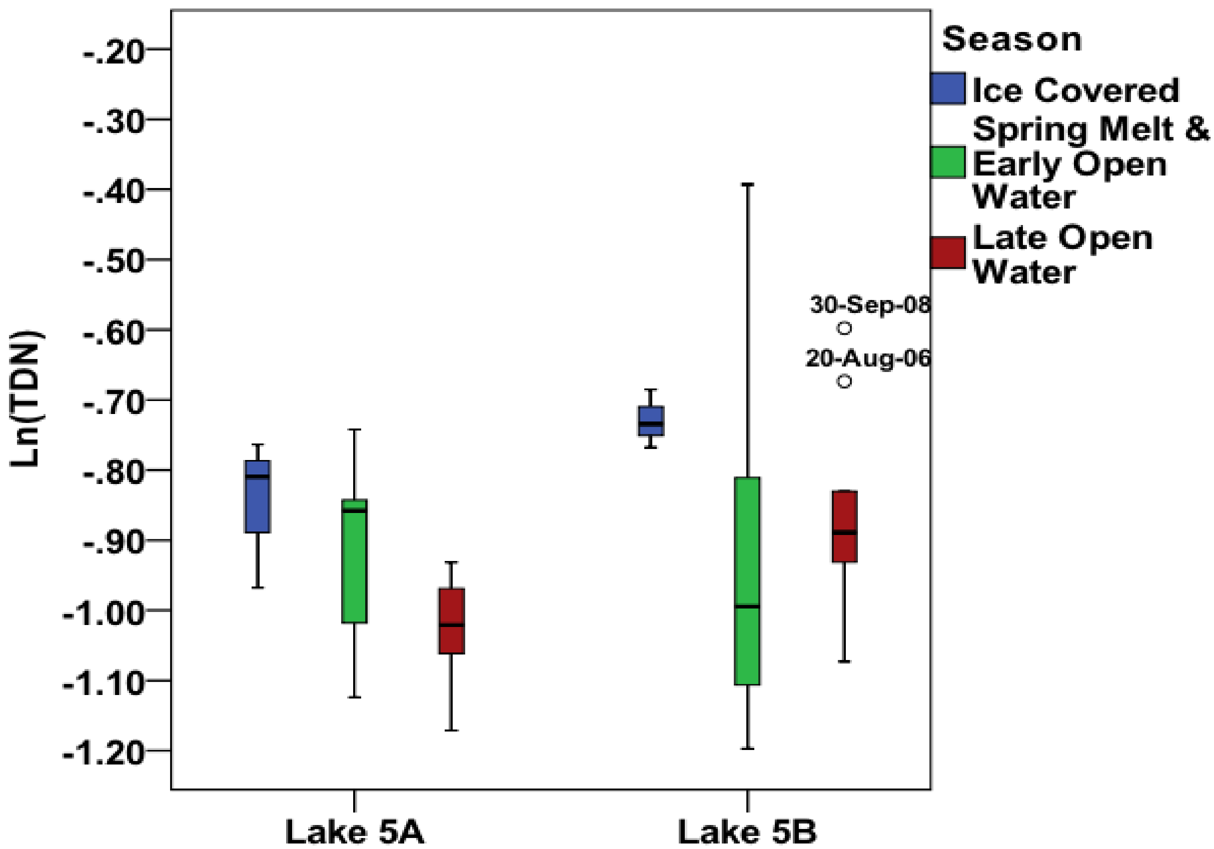


Figure 4.10. A box plot displaying the concentration of TDN in Lake 5A and Lake 5B. The concentration of TDN is grouped based on Lake Type (5A vs. 5B) and categorized based on Season (Ice-Covered vs. Spring Melt and Early Open-water vs. Mid to Late Open-water). All geochemistry data was measured in Mg.L^{-1} and then logarithmically transformed.

period. By comparison, the average concentration of TP in Lake 5B was the highest during spring snowmelt and the lowest during the ice covered period. This would suggest that the concentration of TP in Lake 5B decreases over the ice covered months. However, it's important to note that the detailed data obtained in 2008 and 2009 suggests that the concentration of TP in Lake 5B actually increased over the winter months. In contrast with TP, the mean concentrations of TN and TDN did not exhibit statistically significant variability between seasons. Lastly, there was not a statistically significant interaction effect between Lake Type and Season.

4.5.1.2 Regional Study Lakes

Major Ions

Overall, the results of the annual survey of the 9 regional lakes, conducted in late August of 2006 to 2010, are similar to the results reported for the two primary study lakes. Data were pooled based on Lake Type (Unaffected vs. SRTS-Affected) and Year (2006 to 2010). The mean concentrations of Ca^{2+}, K^{+}, and Mg^{2+} were significantly higher in SRTS-affected lakes than unaffected lakes (ANOVA, $p < 0.05$, **Table 4.3; Figure 4.11 to Figure 4.13**). Although not significant, the concentration of SO_4 in SRTS-affected lakes is generally higher than that of unaffected lakes ($p > 0.05$). These results agree with the results of Kokelj et al. (2005 and 2009b), who found that SRTS-affected lakes typically have higher ionic concentrations than unaffected lakes.

The mean concentrations of Cl^{-} and Na^{+} were not significantly different between SRTS-affected lakes and unaffected lakes. This is consistent with the results presented for the two primary study lakes. For instance, the mean concentration of Na^{+} was significantly greater in Lake 5B than Lake 5A, which was attributed to SRTS. This may be due to differences in the surficial geology of the contributing catchment at Lake 5B and that of the 9 regional study lakes.

It's important to note that the mean concentrations of Cl^{-} , Mg^{2+} , Na^{+}, and SO_4^{2-} varied significantly between study years. Despite exhibiting some inter-annual variability, there were no significant interaction effects between Lake Type and Year for the dependent parameters Ca^{2+} , Cl^{-} , K^{+} , Mg^{2+} , and Na^{+}. Conversely, there was a significant interaction effect between Lake Type and Year for the dependent parameter SO_4^{2-}. That is, the effect of SRTS on the concentration of SO_4^{2-} in lake water is affected by Year and vice versa. The correlation between SRTS and Year could explain why a statistically significant association between the concentration of SO_4 in lake water and SRTS was not observed.

Table 4.3. An ANOVA table for the parameters measured in lake water obtained from the 9 regional study lakes in late August of 2006 to 2010. The dependent parameters (Ca^{2+}, Cl^-, K^+, Mg^{2+}, Na^+, and SO_4^{2-}) were tested according to Lake Type (affected vs. unaffected) and Year. Significant results ($p < 0.05$) are bolded.

Parameter	Lake Type (affected vs. unaffected)			Year (2006 vs. 2007 vs. 2008 vs. 2009 vs. 2010)			Lake Type * Year		
	F	df	P	F	df	P	F	df	P
Ca^{2+}	9.115	1.000	**0.019**	3.295	2.120	0.063	0.804	2.120	0.473
Cl^-	0.048	1.000	0.832	10.078	4.000	**0.000**	2.395	4.000	0.074
K^+	5.756	1.000	**0.048**	3.954	1.467	0.063	0.806	4.000	0.436
Mg^{2+}	6.162	1.000	**0.042**	3.807	4.000	**0.014**	1.605	4.000	0.201
Na^+	0.273	1.000	0.618	13.024	4.000	**0.000**	0.441	4.000	0.778
SO_4^{2-}	5.517	1.000	0.051	5.000	4.000	**0.004**	3.423	4.000	**0.021**

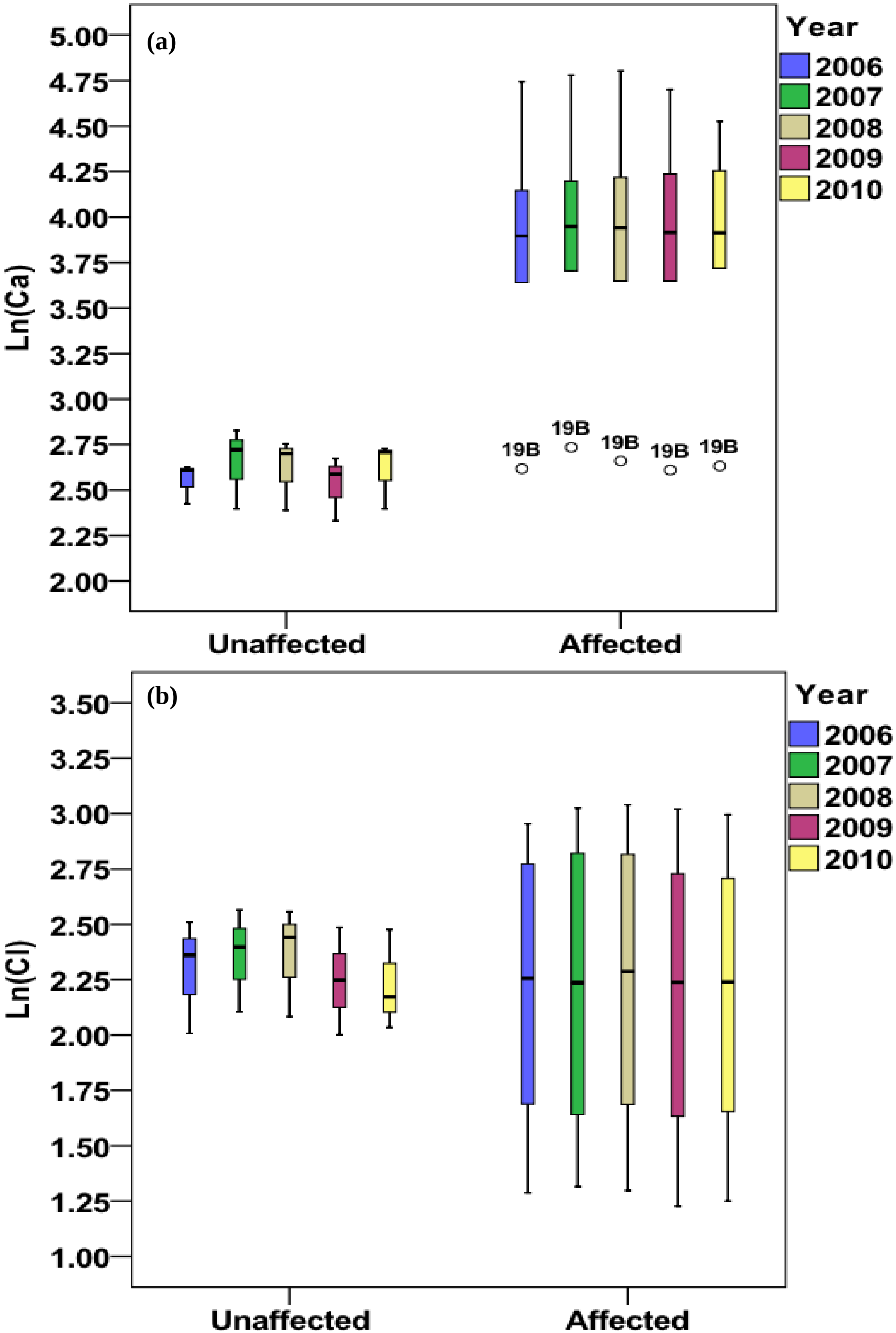


Figure 4.11. Box plots displaying the concentration of (a) Ca^{2+} and (b) Cl^{-} in the 9 regional study lakes. Concentrations are grouped based on Lake Type (unaffected vs. affected) and categorized based on Year (2006 to 2010). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

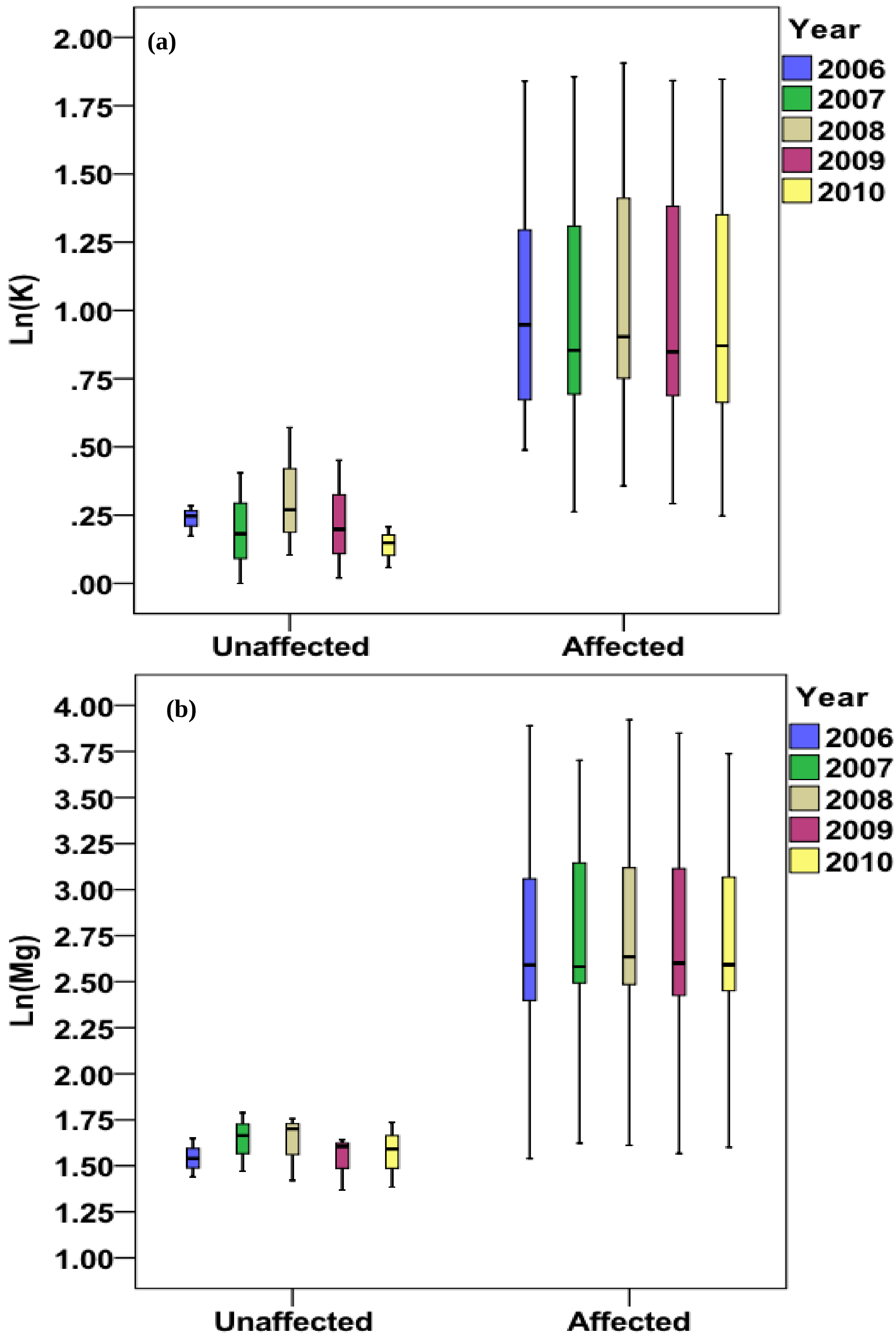


Figure 4.12. Box plots displaying the concentration of (a) K^+ and (b) Mg^{2+} in the 9 regional study lakes. Concentrations are grouped based on Lake Type (unaffected vs. affected) and categorized based on Year (2006 to 2010). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

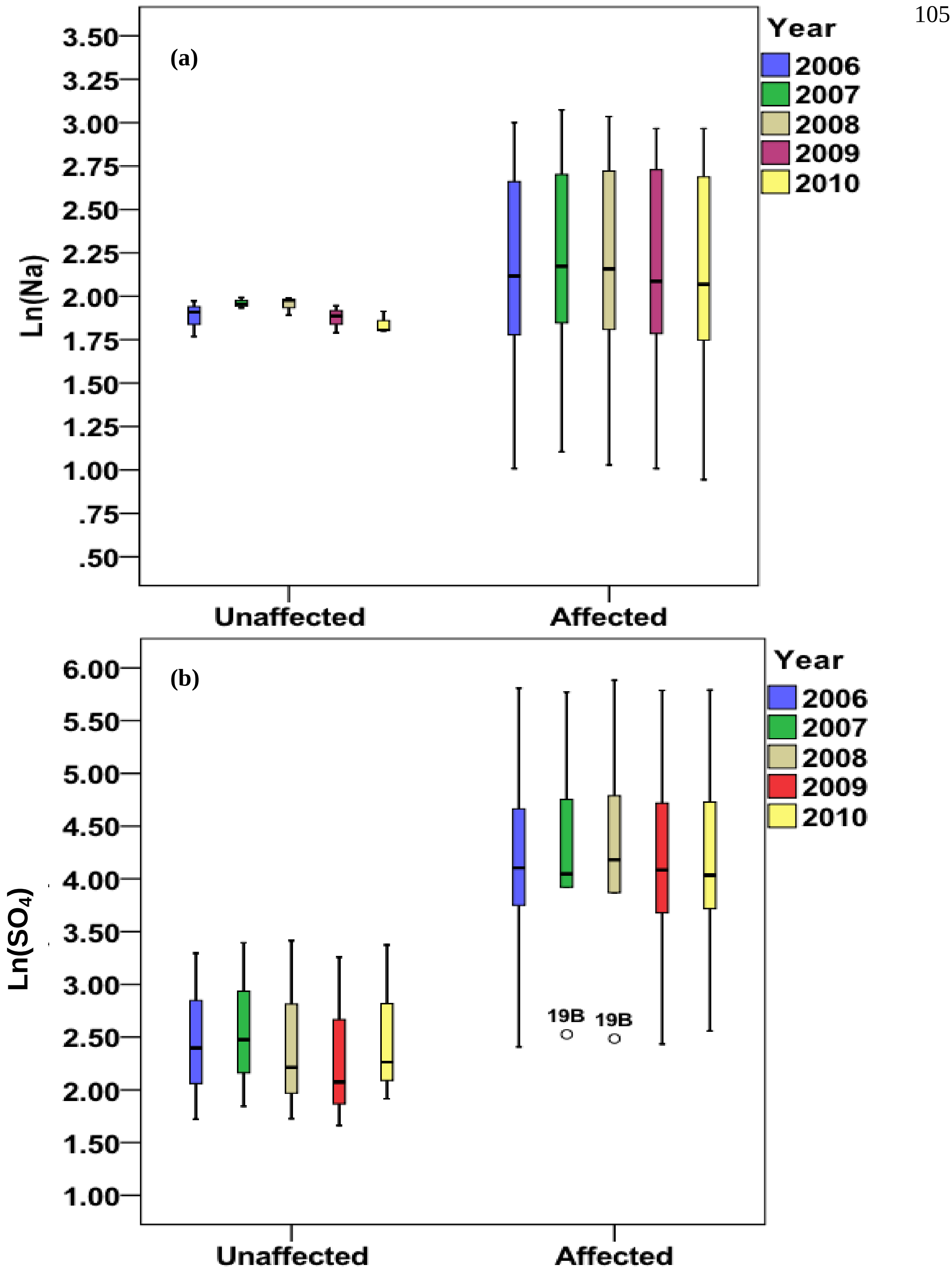


Figure 4.13. Box plots displaying the concentration of (a) Na and (b) SO_4 in the 9 regional study lakes. Concentrations are grouped based on Lake Type (unaffected vs. affected) and categorized based on Year (2006 to 2010). All geochemistry data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

In general, the results from a more temporally detailed survey of the 9 regional lakes, conducted in early May, late June, and late August of 2006 and 2007, support the results obtained from Lake 5A and Lake 5B. Across all three seasons (Ice-Covered, Early Open-water, and Late Open-water), the mean concentrations of Ca^{2+}, Mg^{2+}, Na^{+} and SO_4^{2-}were significantly greater in SRTS-affected lakes than unaffected lakes (ANOVA, $p < 0.05$, **Table 4.4**, **Figures 4.15 to 4.17**).

The mean concentrations of Cl^{-} and Na^{+} in SRTS-affected lakes were not significantly different than that of unaffected lakes ($p > 0.05$). This only partially supports the results presented for Lake 5A and Lake 5B, which indicated that SRTS increased the concentration of Na in Lake 5B.

The mean concentrations of Ca^{2+}, Cl^{-}, Mg^{2+}, K^{+}, Na^{+}, and SO_4^{2-} exhibited significant variability across the three seasons. The ionic concentration of lake water was typically the highest in late winter and the lowest during the early open-water period. The ionic concentration of the regional study lakes was typically higher during the late open-water period than during spring snowmelt. This is consistent with the results presented for Lake 5A and Lake 5B. That is, the ionic concentration of SRTS-affected and unaffected lakes exhibited significant seasonal variability, which appears to be driven by ice formation, spring snowmelt, subsurface runoff, and evaporation.

There were significant interaction effects between Lake Type and Season for the dependent variables Ca^{2+}, Cl^{-}, and SO_4^{2-}. In other words, the effect of season on the concentration of Ca^{2+}, Cl^{-}, and SO_4^{2-} is dependent on SRTS and vice versa. Notably, the mean difference in the concentration of Ca^{2+}, Cl^{-} , and SO_4^{2-} between early open-water and late open-water was greater in SRTS-affected lakes than unaffected lakes.

Table 4.4. An ANOVA table for the parameters measured in lake water obtained from the 9 regional study lakes in early-May, late-June, and late-August of 2006 and 2007. The dependent parameters (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) were tested according to Lake Type (affected vs. unaffected) and Season (Ice-Covered vs. Early Open-water vs. Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	Lake Type (affected vs. unaffected)			Season (Ice Covered vs. Early Open-water vs. Late Open-water)			Lake Type * Season		
	F	**df**	**P**	**F**	**df**	**P**	**F**	**df**	**P**
Ca^{2+}	17.366	1.000	**0.001**	46.869	2.000	**0.000**	3.586	2.000	**0.040**
Cl^{-}	0.417	1.000	0.528	132.674	2.000	**0.000**	3.489	2.000	**0.043**
K^{+}	11.813	1.000	**0.004**	98.836	2.000	**0.000**	1.515	2.000	0.236
Mg^{2+}	10.398	1.000	**0.006**	67.433	2.000	**0.000**	2.975	2.000	0.077
Na^{+}	0.156	1.000	0.699	112.273	2.000	**0.000**	2.660	2.000	0.086
SO_4^{2-}	12.541	1.000	**0.003**	94.620	2.000	**0.000**	5.063	2.000	**0.013**

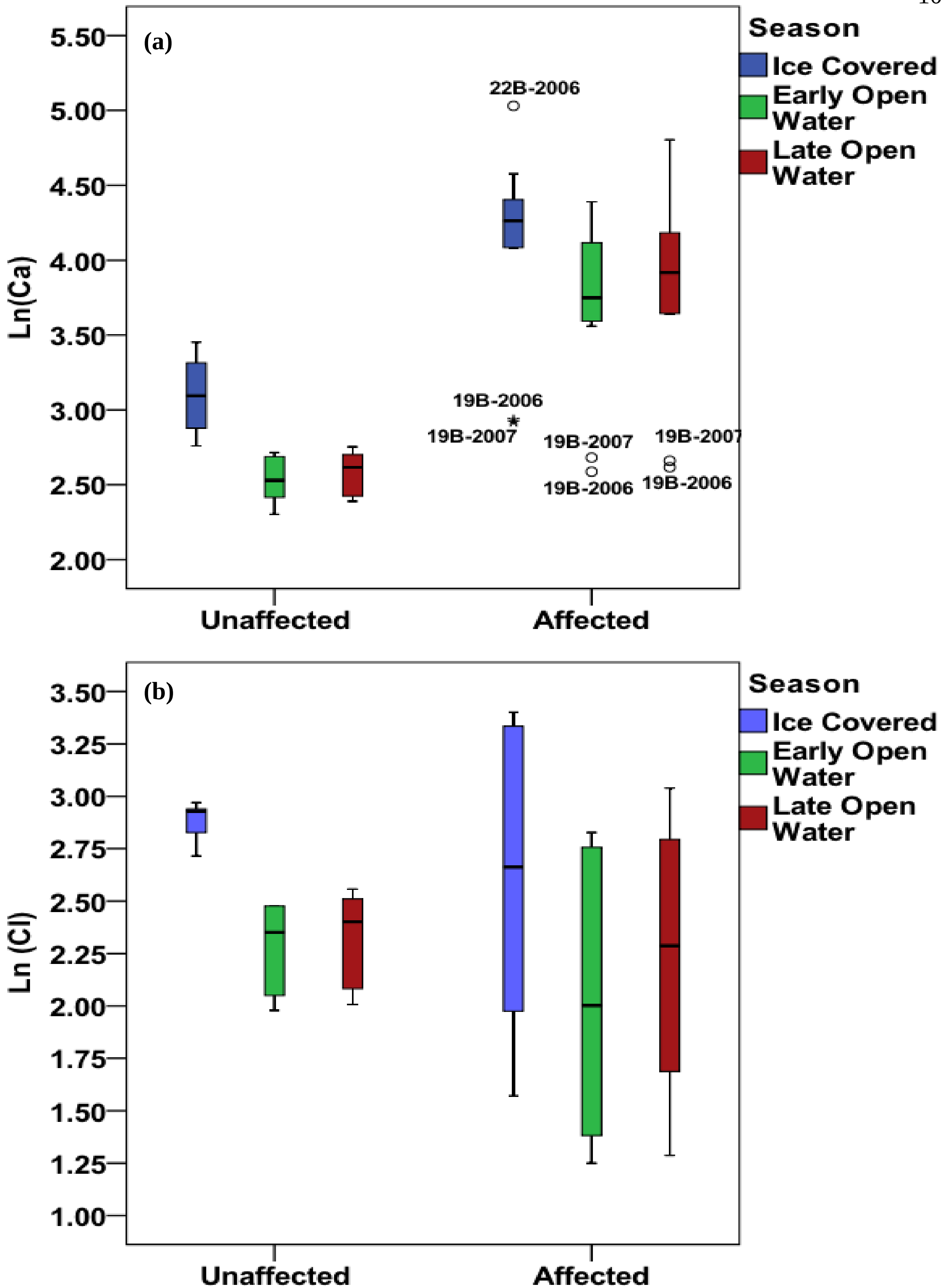


Figure 4.14. Box plots displaying the mean concentration of (a) Ca^{2+} and (b) Cl^{-} in the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Ice-Covered vs. Early Open-water vs. Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

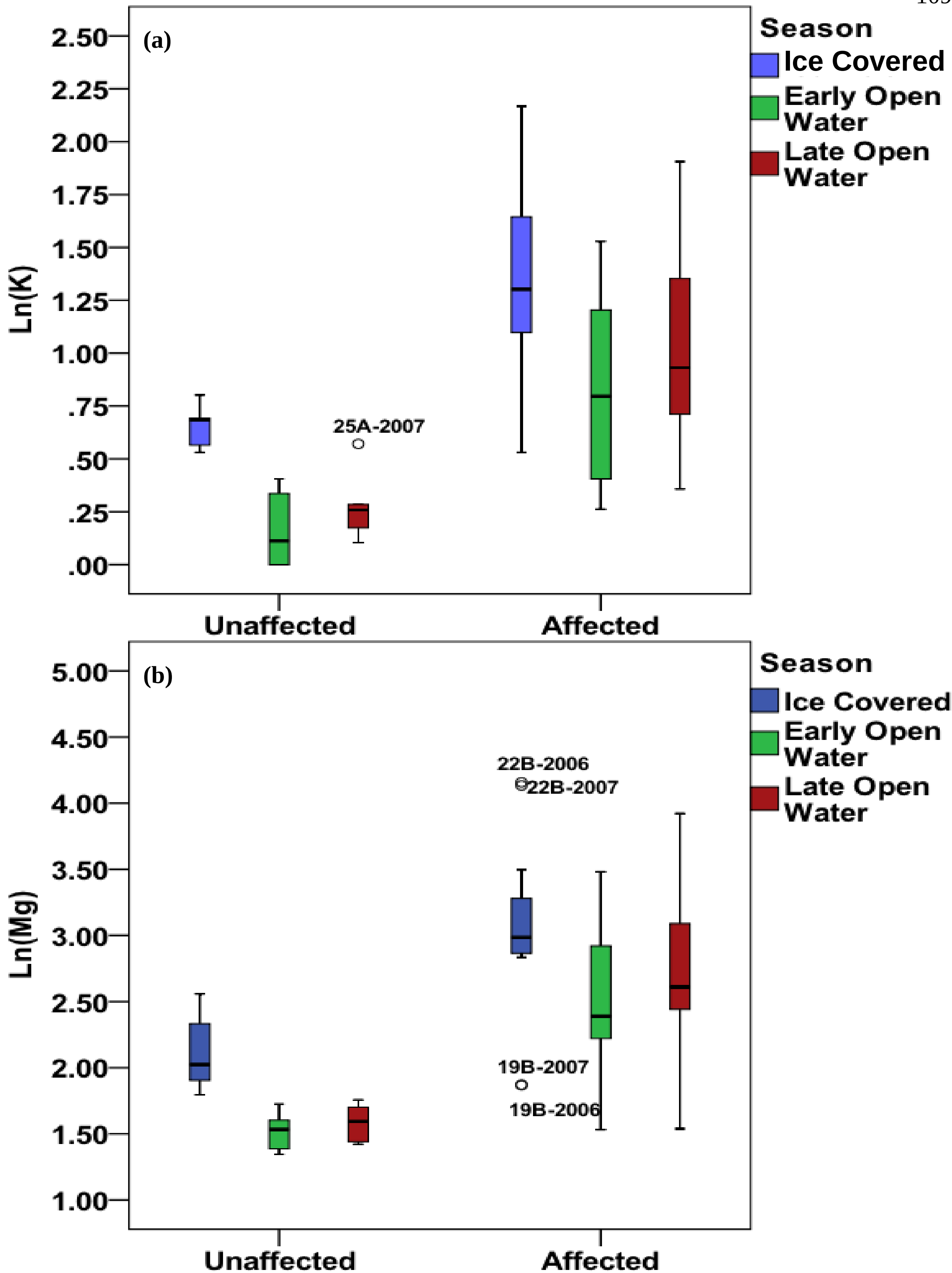


Figure 4.15. Box plots displaying the concentration of (a) K^{+} and (b) Mg^{2+} in the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Ice-Covered vs. Early Open-water vs. Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

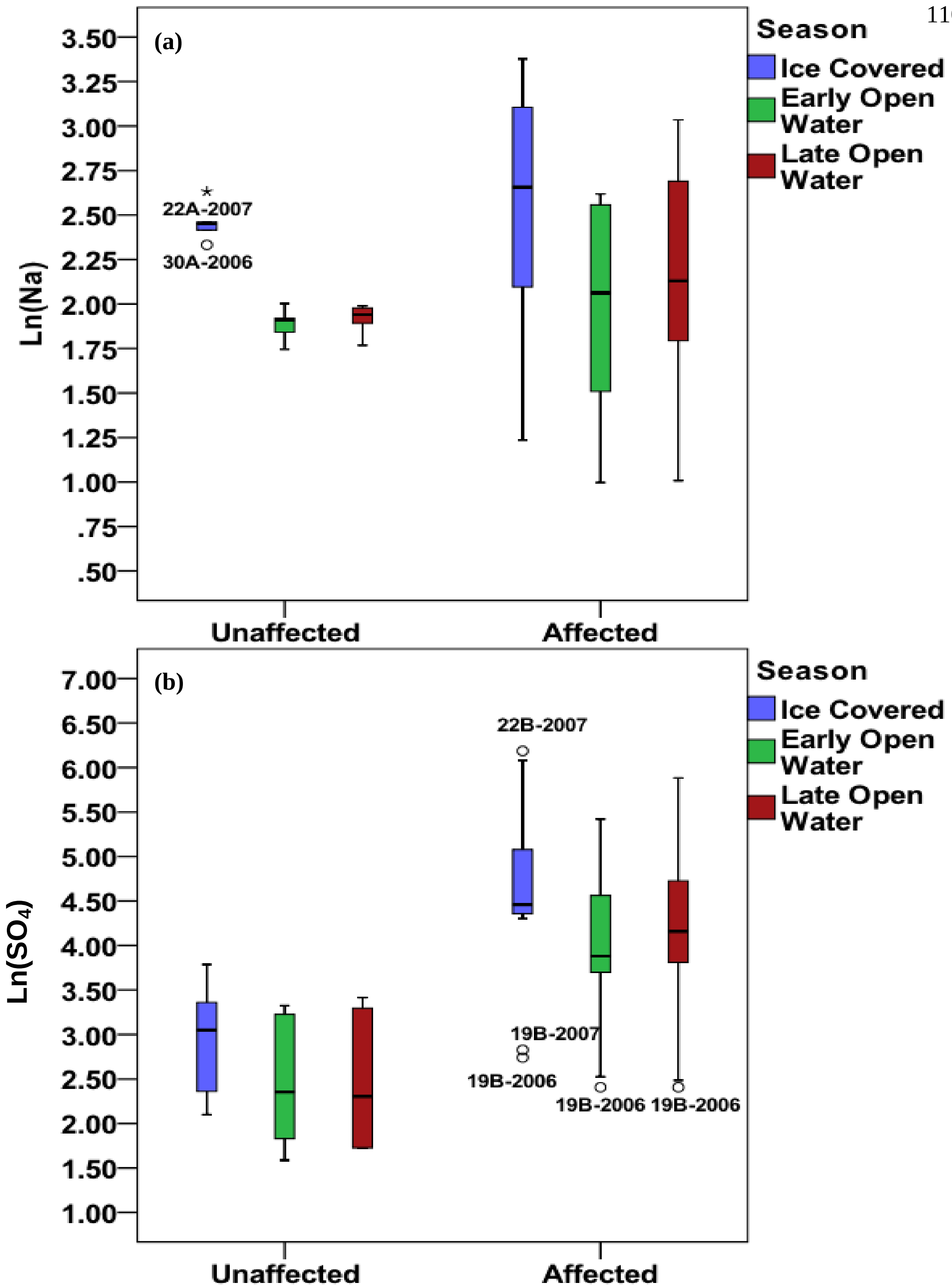


Figure 4.16. Box plots displaying the concentration of (a) Na^+ and (b) SO_4^{2-} (bottom) in the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Ice-Covered vs. Early Open-water vs. Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

Nutrients

Overall, the results of the annual 9-lake survey, conducted in late August of 2006 to 2010, support the results presented for Lake 5A and Lake 5B. Data were pooled based on Lake Type (Unaffected vs. Affected) and Year (2006 to 2010). The mean concentrations of TP, TN, and TDN in SRTS-affected lakes were not significantly different from that of unaffected lakes (ANOVA, $p > 0.05$, **Table 4.5**, **Figure 4.17** and **Figure 4.18**). The mean concentrations of TP, TN, and TDN did not exhibit significant variability over the five study years. Lastly, no significant interaction between SRTS and Year was observed.

The results of the more temporally detailed 9-lake survey, conducted in early May, late June, and late August of 2006 and 2007 are consistent with the results presented for Lake 5A and Lake 5B. Data were pooled based on Lake Type (Unaffected vs. Affected) and Season (Ice Covered vs. Early Open-water vs. Late Open-water). The mean concentrations of TP and TDN in SRTS-affected lakes were not significantly different than that of unaffected lakes (ANOVA, $p > 0.05$, **Table 4.6** and **Figure 4.19**). This is consistent with the results presented for Lake 5A and Lake 5B.

The mean concentration of TDN exhibited significant variability across the three hydrological seasons ($p < 0.05$). The mean concentration of TDN was the highest in late-winter and the lowest during the early open-water period. This is consistent with the results presented for Lake 5A and Lake 5B. The concentration of TDN in Lake 5A and Lake 5B increased over the open-water period, due to the addition of nutrient-rich runoff and concentration via evaporation. Following the open-water period, the concentration of TDN continued to increase over the winter months, due to the freeze-out associated with ice formation.

It is important to note that there was a significant interaction between Lake Type and Season for the dependent variable TP. This indicates that the effect of Season on the concentration of TP is dependent on SRTS. For instance, in unaffected lakes, the concentration of TP was the lowest in late winter and the highest in autumn. In SRTS-affected lakes, however, the concentration of TP was the lowest in late autumn and the highest during the early open-water period. The results of this study suggest that seasonal variability in the concentration of TP in the study lakes was different for unaffected lakes and SRTS-affected lakes.

Table 4.5. An ANOVA table for the parameters measured in lake water obtained from the 9 regional study lakes in 2006, 2007, and 2009. The dependent parameters (TP, TN, and TDN) were tested according to Lake Type (affected vs. unaffected) and Year (2006 vs. 2007 vs. 2009). Significant results ($p < 0.05$) are bolded.

Parameter	Lake Type (affected vs. unaffected)			Year (2006 vs. 2007 vs. 2009)			Lake Type * Year		
	F	df	P	F	df	P	F	df	P
TP	0.565	1.000	0.481	2.023	2.000	0.175	0.654	2.000	0.538
TN	0.146	1.000	0.715	3.812	2.000	0.052	0.167	2.000	0.848
TDN	0.847	1.000	0.393	0.483	2.000	0.628	0.919	2.000	0.425

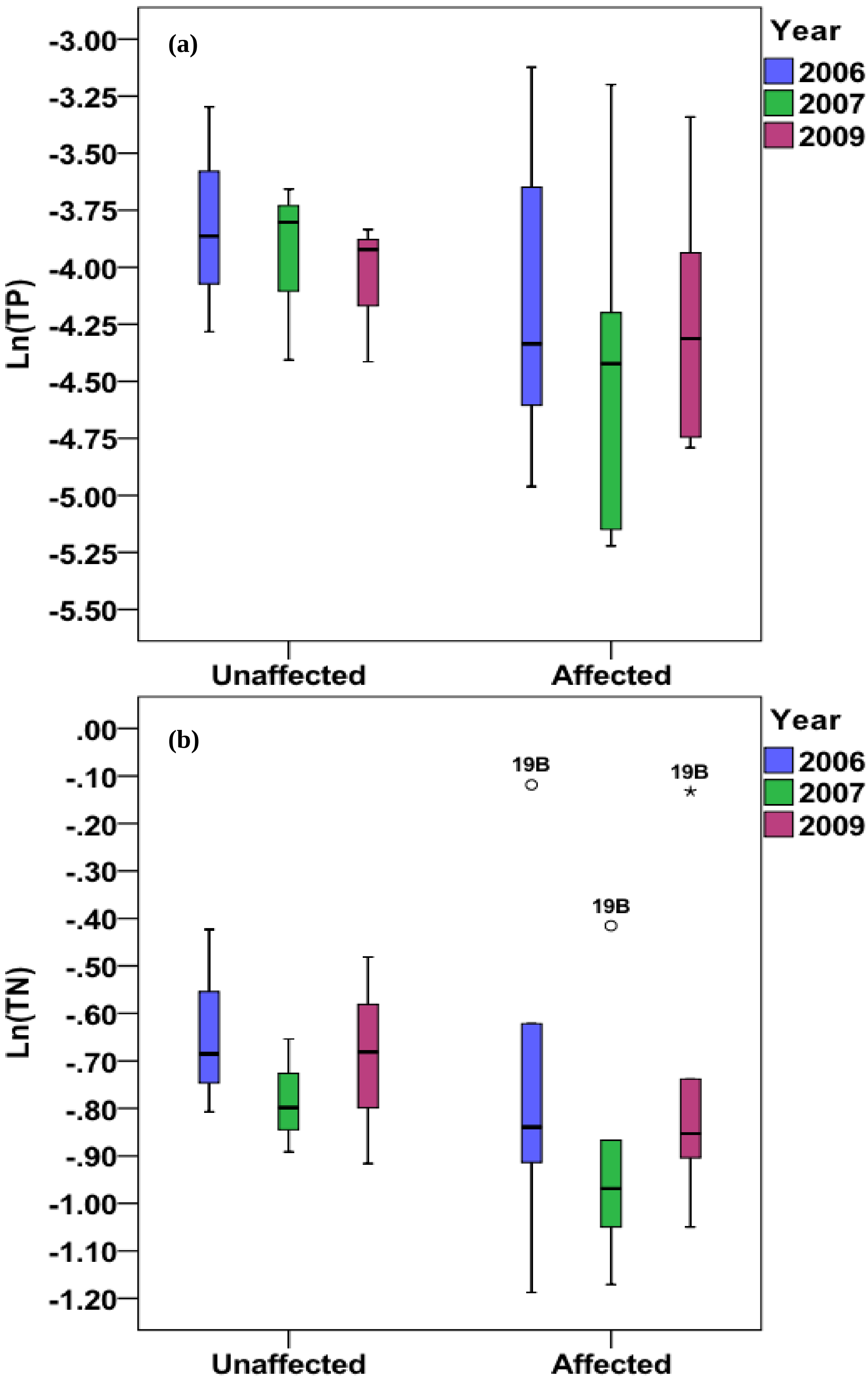


Figure 4.17. Box plots displaying the concentration of (a) TP and (b) TN in the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Ice Covered vs. Early Open-water vs. Late Open-water). All data was measured in Mg.L^{-1} and then logarithmically transformed.

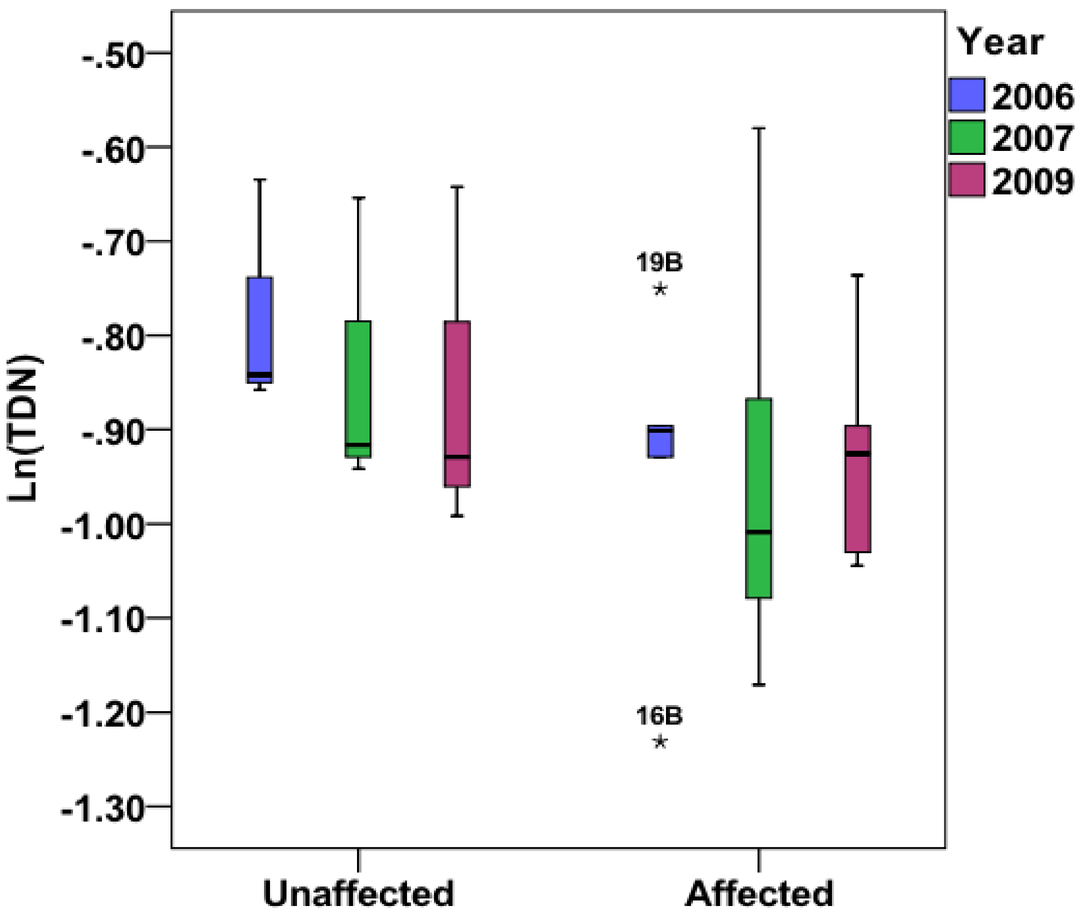


Figure 4.18. Box plots displaying the concentration of TN (top) and TDN (bottom) in the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Winter vs. Spring vs. Autumn). All data was measured in Mg.L^{-1} and then logarithmically transformed.

Table 4.6. An ANOVA table for the parameters measured in lake water obtained from the 9 regional study lakes. The dependent parameters (Total Dissolved Nitrogen (TDN) and Total Phosphorus (TP)) were tested according to Lake Type (affected vs. unaffected) and Season (Ice Covered vs. Early Open-water vs. Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	Lake Type (affected vs. unaffected)			Season (Ice Covered vs. Early Open-water vs. Late Open-water)			Lake Type * Season		
	F	df	P	F	df	P	F	df	P
TDN	4.448	1.000	0.053	21.192	1.411	**0.000**	0.424	1.411	0.591
TP	0.031	1.000	0.864	4.089	1.242	0.052	6.219	1.242	**0.018**

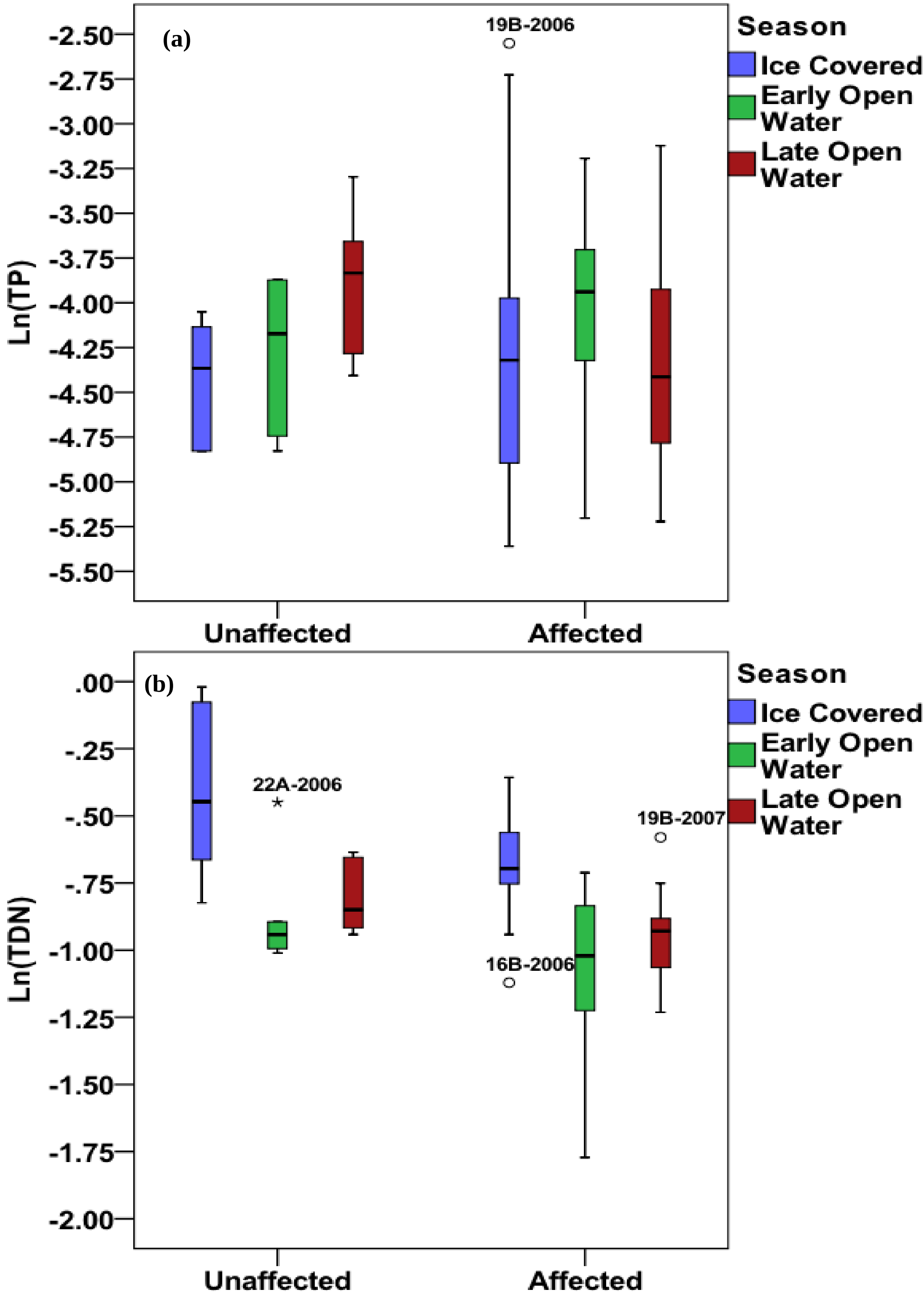


Figure 4.19. Box plots displaying the concentration of (a) TP and (b) TDN in the 9 lake survey. The dependent parameters are grouped based on Lake Type (affected vs. unaffected) and categorized based on Season (Ice Covered vs. Early Open-water vs. Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

4.5.2 Catchment Flow

4.5.2.1 Lake 5A and 5B

Major Ions

During the spring melt and open-water seasons of 2008 and 2009, detailed geochemical signature surveys were conducted at the surface flow pathways leading to and from Lake 5A and Lake 5B (see **Figure 4.1**). Data were pooled based on the Catchment Flow type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and Season (Spring Melt and Early Open-water vs. Middle to Late Open-water). The mean concentrations of Ca^{2+}, Cl^-, K^+, Na^+, Mg^{2+}, and SO_4^{2-} were not significantly different between the Inflow at Lake 5A and the Inflow at Lake 5B (ANOVA, $p > 0.05$, **Table 4.7, Figures 4.20 to 4.22**). They were, however, significantly higher in the 5B Slump flow than both the inflow at Lake 5A and the inflow at Lake 5B ($p < 0.05$). These results indicate that the mean concentrations of these ions were higher in Lake 5B, relative to Lake 5A, either partly or wholly due to landscape runoff inputs. This is consistent with the postulation made by Kokelj et al. (2005; 2009a) that major ions leach out of SRTS-affected soils over time.

The mean concentrations of Na^+ and SO_4^{2-} in runoff were significantly lower in the spring melt and early open-water period than during the middle and late open-water period. This partially agrees with other studies done in the study area. Quinton and Pomeroy (2006) found that runoff at nearby Siksik Creek was typically the most diluted during spring snowmelt. As the summer months progressed, the concentration of most major ions in runoff increased, which was attributed to an increase in the importance of subsurface runoff. As the active layer develops in summer, the vertical infiltration and residence time of water runoff increases, leading to an increase in the chemical interaction between water runoff with the soil profile and ultimately the concentration of major ions in runoff.

In contrast with Na^+ and SO_4^{2-}, the mean concentrations of Ca^{2+}, Cl^-, and Mg^{2+} in runoff were not significantly lower during the spring melt and early open-water period, relative to the middle and late open-water period. In August and September 2008, the concentrations of Ca^{2+}, Cl^-, and Mg^{2+} in runoff became heavily diluted due to rainfall inputs. Consequently, the mean concentration of these ions in runoff was lower during the middle to late open-water period than during the spring melt and early open-water period. Conversely, in 2009, the mean concentrations of Ca^{2+}, Cl^-, and Mg^{2+} in runoff were higher during the mid to late open-water period than during the spring melt and early open-water period, in spite of heavy rainfall events. This suggests that, in addition to spring snowmelt, summer rainfall has an important effect on the geochemistry of runoff to Lake 5A and Lake 5B that needs to be explored further.

There was a statistically significant interaction effect between Catchment Flow Pathway and Season for the dependent variables K^+ and SO_4^{2-}. Notably, the mean concentration of K^+ in runoff from unaffected terrain was higher during the spring melt and early open-water period than during the middle to late open-water period. In contrast with unaffected terrain, the mean concentration of K^+ in runoff from SRTS-affected terrain was lower during the spring melt and early open-water period than during the mid to late open-water period. This suggests that SRTS could be a source of K^+ to Lake 5B in summer.

Kokelj et al. (2002) found that K^+ was only present in the active layer in small amounts. In addition to being a major ion, K^+ is an essential nutrient that is required by vegetation communities in the processes associated with production. In summer and autumn, which is typically when productivity is the greatest, the uptake of K^+ by vegetation communities is likely greater than during the spring melt period. The results presented here suggest that, for unaffected terrain, K^+ might be taken up by vegetation at a faster rate than it can be liberated from the thawing active layer. As a result, the amount of K^+ in runoff decreases during the summer months. In contrast with the overlying active layer, Kokelj and Burn (2005) and Keller et al. (2007) found that K^+ is present in relatively large quantities in near surface permafrost. The results presented here suggest that the thawing of near surface permafrost, associated with SRTS, increases the supply of K^+ to the active layer at a faster rate than it is taken up by vegetation. As a result, the concentration of K^+ in runoff from SRTS-affected terrain is higher during the open-water period than during spring melt period.

Table 4.7. An ANOVA table for the parameters measured in catchment flow to Lake 5A and Lake 5B in 2008 to 2010. The dependent parameters (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) were tested according to Catchment Flow pathway (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	**Catchment Flow (5A Inflow vs. 5B Inflow vs. 5B Slumpflow)**			**Season (Spring Melt and Early Open-water vs. Mid to Late Open-water)**			**Catchment Flow * Season**			**Post-Hoc**
	F	**df**	**P**	**F**	**df**	**P**	**F**	**df**	**P**	
$\mathbf{Ca^{2+}}$	457.896	2.000	**0.00**	4.086	1.000	0.06	0.906	2.000	0.43	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*
$\mathbf{Cl^{-}}$	36.868	2.000	**0.00**	1.285	1.000	0.28	1.897	2.000	0.19	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*
$\mathbf{K^{+}}$	80.971	2.000	**0.00**	31.316	1.000	**0.00**	12.301	2.000	**0.00**	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*
$\mathbf{Mg^{2+}}$	81.807	2.000	**0.00**	0.696	1.000	0.42	2.040	2.000	0.17	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*
$\mathbf{Na^{+}}$	496.533	2.000	**0.00**	8.512	1.000	**0.01**	0.080	2.000	0.92	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*
$\mathbf{SO_4^{2-}}$	108.062	2.000	**0.00**	28.009	1.000	**0.00**	5.421	2.000	**0.02**	*5A Inflow* vs. 5B *Slumpflow* *5B Inflow* vs. 5B *Slumpflow*

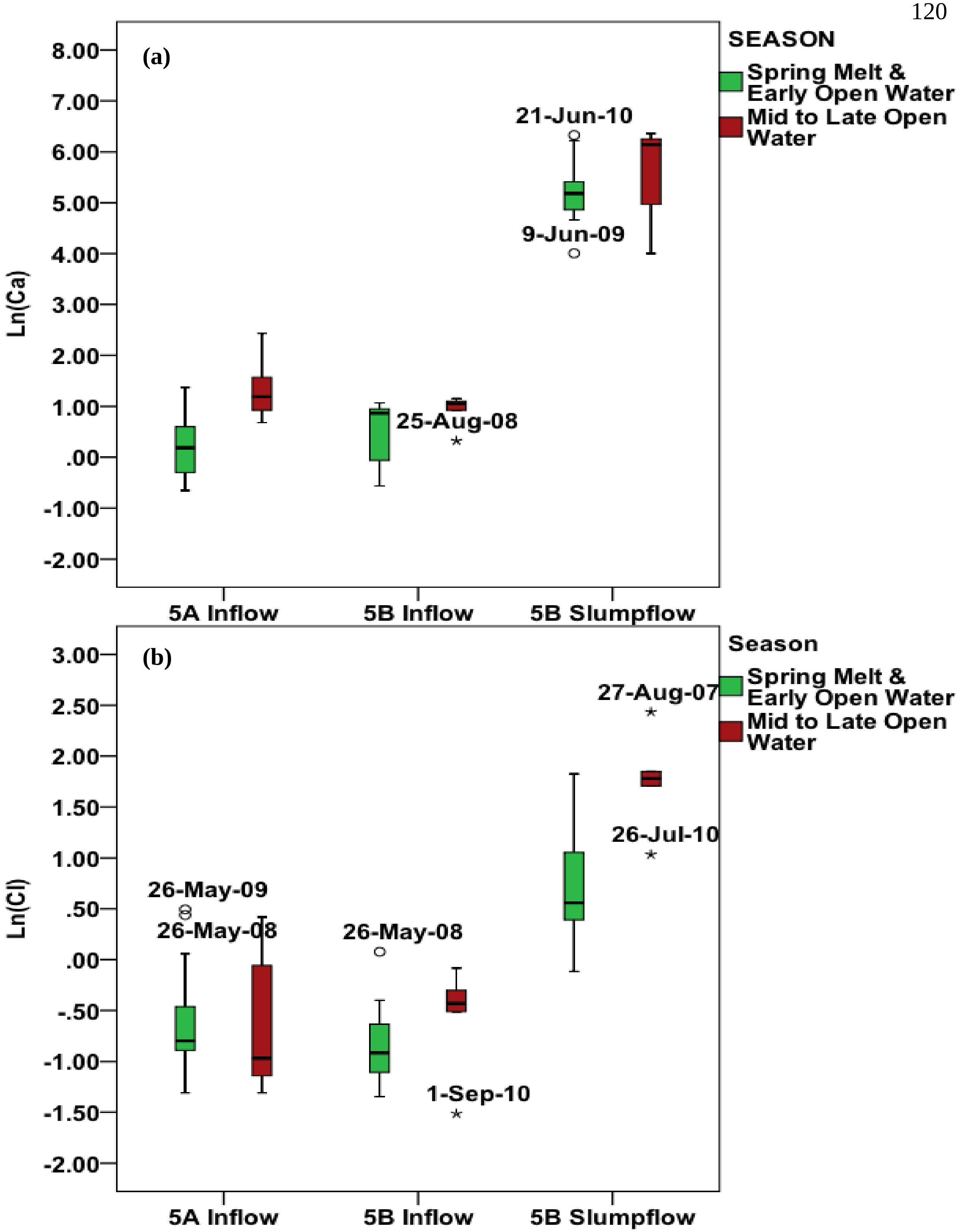


Figure 4.20. Box plots displaying the concentration of (a) Ca^{2+} and (b) Cl^{-} in catchment flow to Lake 5A and Lake 5B. The dependent parameters are grouped based on Catchment Flow Type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

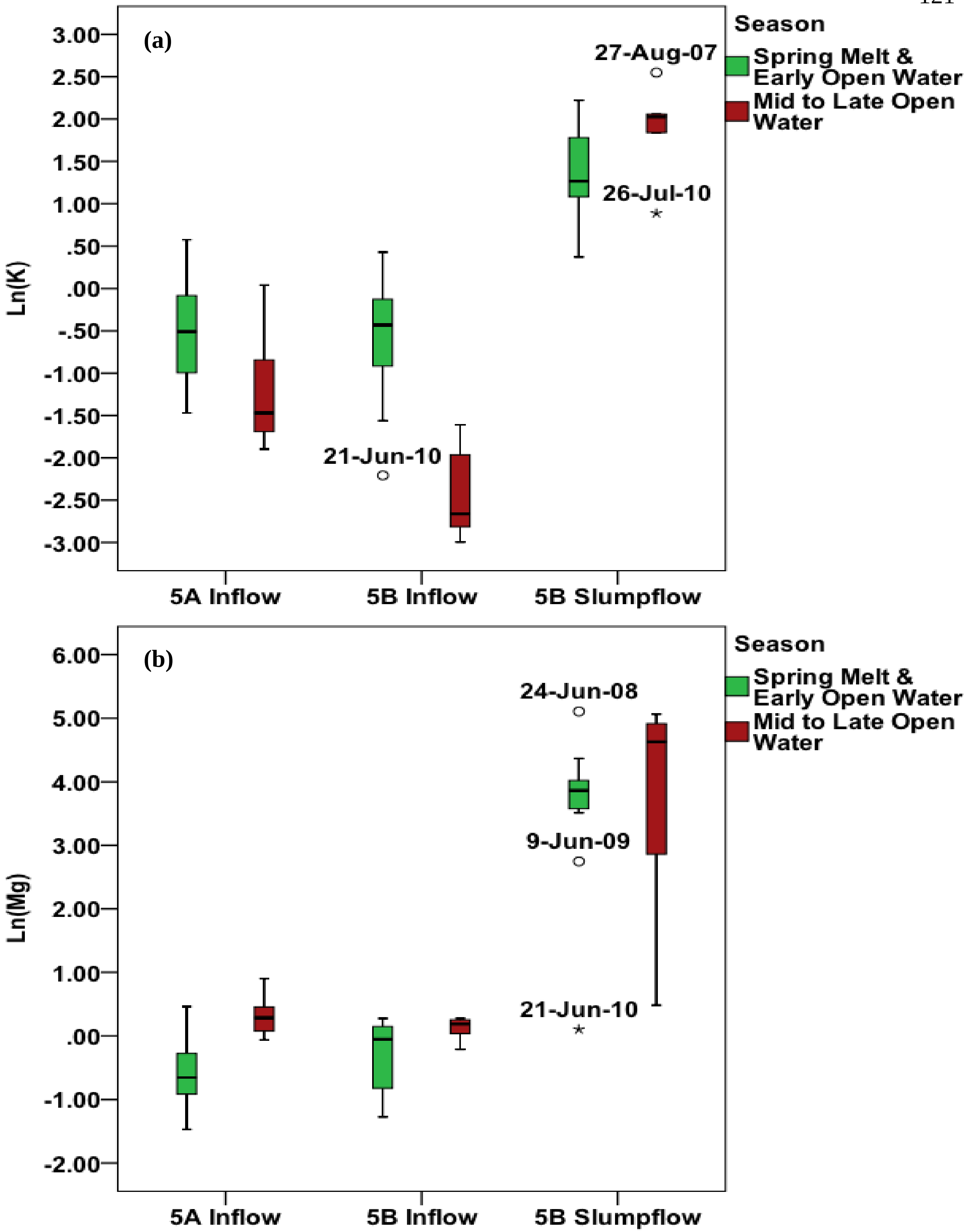


Figure 4.21. Box plots displaying the concentration of (a) K^+ and (b) Mg^{2+} in catchment flow to Lake 5A and Lake 5B. The dependent parameters are grouped based on Catchment Flow Type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

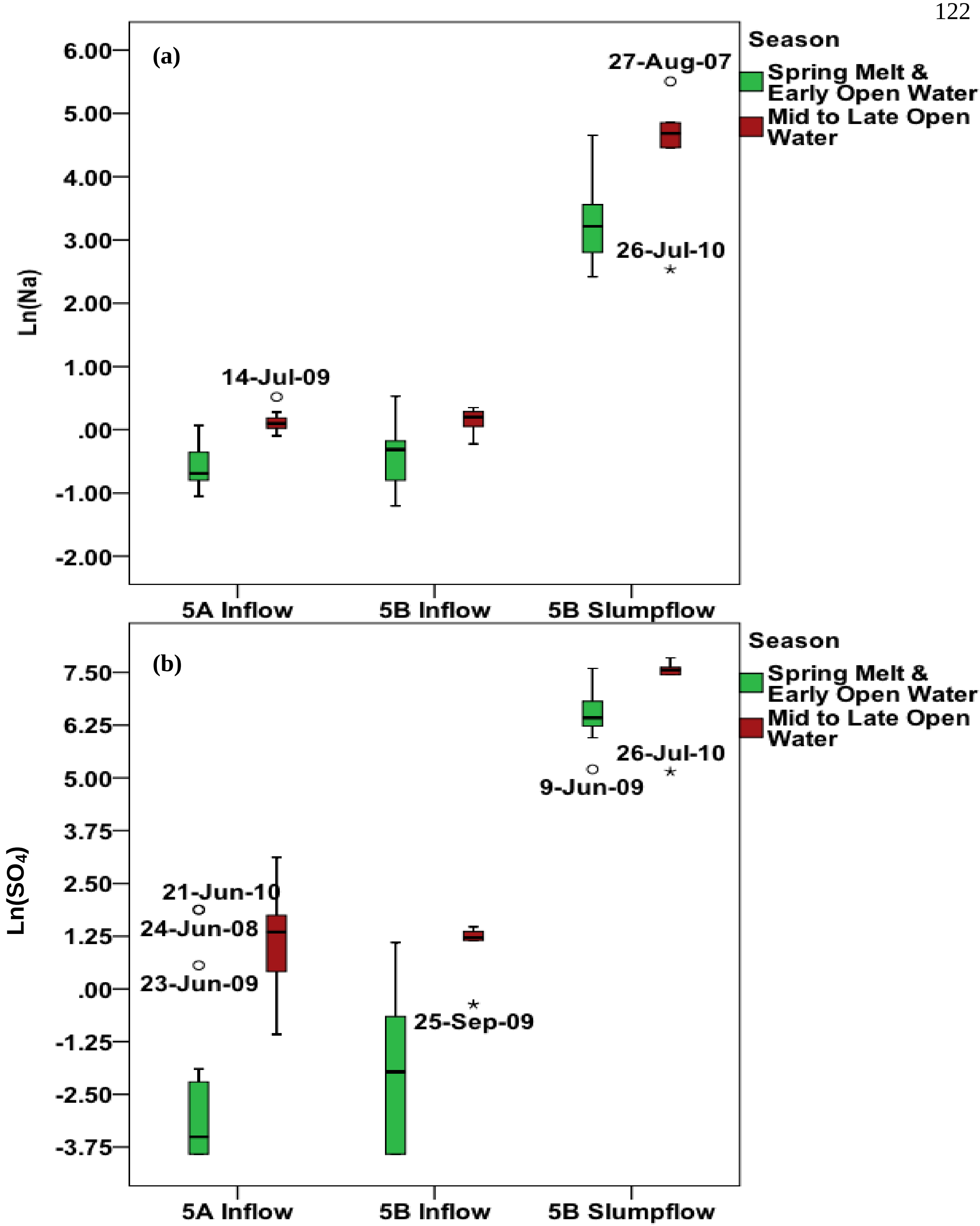


Figure 4.22. Box plots displaying the concentration of (a) Na and (b) SO_4 in catchment flow to Lake 5A and Lake 5B. The dependent parameters are grouped based on Catchment Flow Type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

Nutrients

The mean concentrations of TP, TN, and TDN were not significantly different between catchment flow pathways (ANOVA, $p > 0.05$, **Table 4.8, Figure 4.23,** and **Figure 4.24**). This is consistent with the results presented for lake water. The mean concentrations of TP, TN, and TDN in Lake 5B were not significantly different from Lake 5A. This suggests that SRTS does not increase or decrease the supply of nutrients to runoff and subsequently, Lake 5B. This does not agree with other studies, which suggest that permafrost degradation increases the supply of nutrients to runoff.

The mean concentrations of TN, TDN, and TP in runoff did not vary significantly between seasons. Interestingly, the concentration of TP and TDN in lake water did vary significantly between seasons. This suggests that in-lake biological processes could be driving the concentration of TP and TDN in the two primary study lakes, as opposed to catchment-level hydrological processes.

Table 4.8. An ANOVA table for the parameters measured in catchment flow to Lake 5A and Lake 5B in 2008 to 2010. The dependent parameters (Total Nitrogen (TN), Total Dissolved Nitrogen (TDN), and Total Phosphorus (TP)) were tested according to Catchment Flow pathway (5A Inflow vs. 5B Inflow vs. Slumpflow) and Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	Catchment Flow (5A Inflow vs. 5B Inflow vs. 5B Slumpflow)			Season (Spring Melt and Early Open-water vs. Late Open-water)			Catchment Flow * Season		
	F	**df**	**P**	**F**	**df**	**P**	**F**	**df**	**P**
TN	0.131	2.000	0.882	0.803	1.000	0.436	0.747	2.000	0.545
TDN	1.779	2.000	0.309	3.714	1.000	0.150	0.345	2.000	0.733
TP	0.138	2.000	0.877	2.248	1.000	0.231	0.198	2.000	0.830

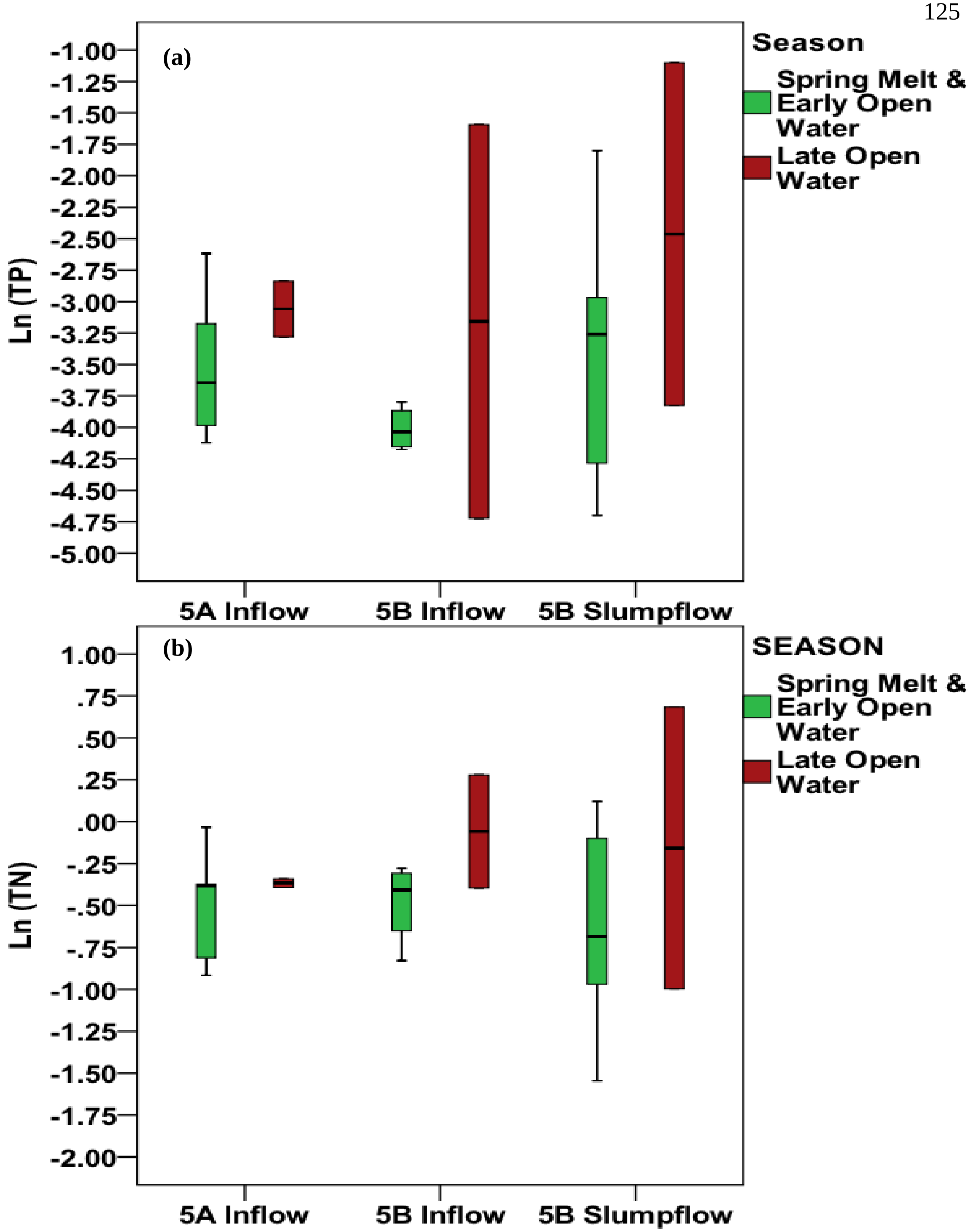


Figure 4.23. Box plots displaying the concentration of (a) Total Phosphorus (TP) and (b) Total Nitrogen (TN) in catchment flow to Lake 5A and Lake 5B. The dependent parameters are grouped based on Catchment Flow Type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

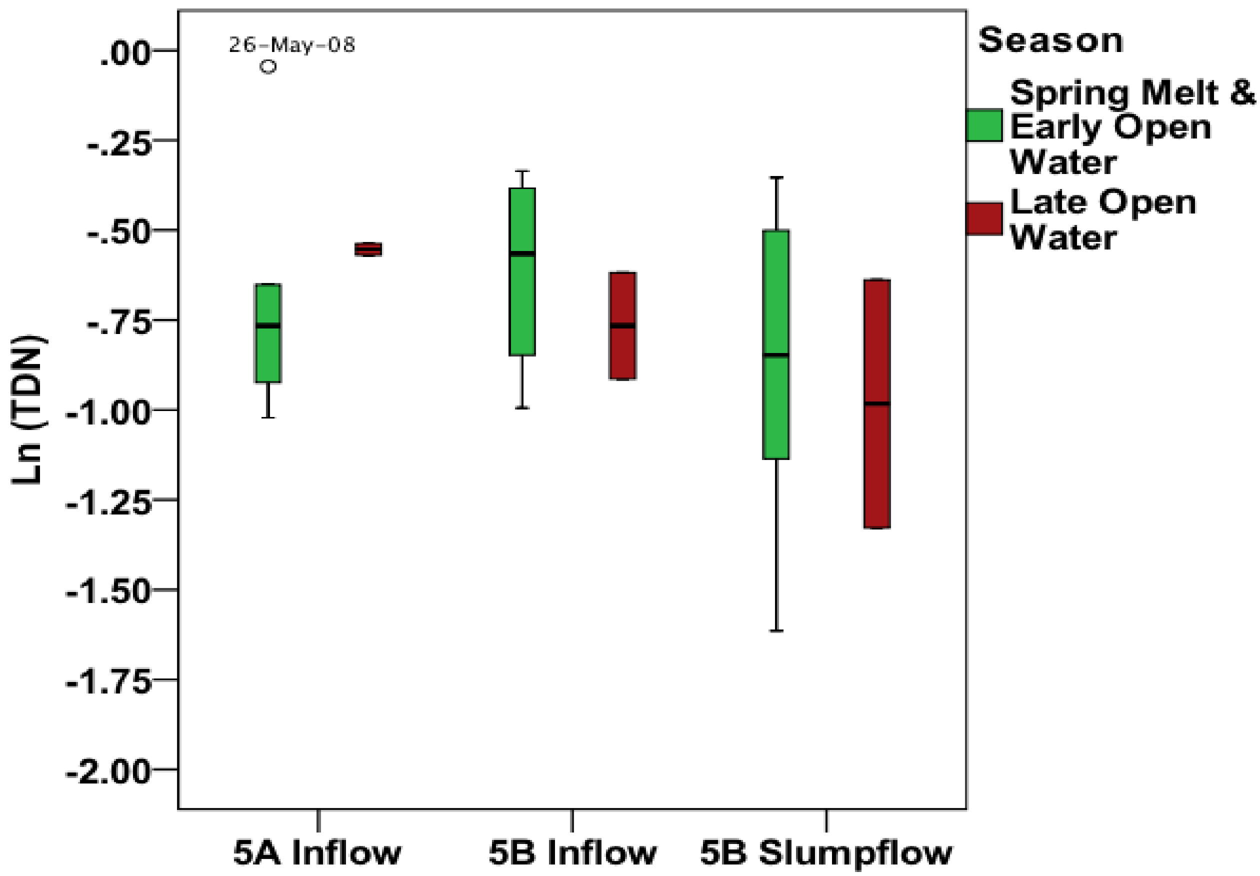


Figure 4.24. A box plot displaying the concentration of TDN in catchment flow to Lake 5A and Lake 5B. The dependent parameters are grouped based on Catchment Flow Type (5A Inflow vs. 5B Inflow vs. 5B Slumpflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

4.5.2.2 Regional Study Lakes

Major Ions

Data collected at the 9 regional lakes and the YaYa Subcatchment lake in August of 2009 and 2010 were pooled based on Catchment Flow Type (catchment flow vs. slumpflow). The mean concentrations of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} were significantly higher in runoff from SRTS-affected terrain than that of runoff from unaffected terrain (T-Test, $p < 0.05$, **Table 4.9**, **Figure 4.25**). This is partially consistent with the results reported for Lake 5A and Lake 5B, which indicated that SRTS increased the supply of these ions to runoff at the Lake 5B catchment. Overall, the results of this study suggest that SRTS increases the supply of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} to runoff, subsequently increasing the concentration of these ions in SRTS-affected lakes.

In contrast with the above, the mean concentrations of Cl^{-} and Na^{+} in runoff was not significantly affected by SRTS ($p > 0.05$). This was not unexpected, largely because the mean concentrations of Cl^{-} and Na^{+} in SRTS-affected lakes was not significantly different from that of unaffected lakes, which suggests that SRTS does not affect the concentration of Cl^{-} and Na^{+} in runoff and lake water at the 10 regional study lakes. It's important to note that this is not consistent with the results presented for Lake 5A and Lake 5B, which indicated that SRTS increased the supply of these ions to runoff at the Lake 5B catchment, subsequently increasing the concentrations of Cl^{-} and Na^{+} in Lake 5B. Overall, the results presented here support the postulation made earlier that the effect of SRTS on the concentration of Cl^{-} and Na^{+} in runoff and, subsequently, in small tundra lakes may vary across the study region depending on surficial geology.

Table 4.9. An Independent Samples T-Test for the parameters measured in catchment flow to the 9 regional study lake catchments and the YaYa Subcatchment Lake catchment in 2009 and 2010. The dependent parameters (Ca^{2+}, Cl^{-}, Mg^{2+}, Na^{+}, K^{+} , and SO_4^{2-}) were tested according to Catchment Flow pathway (Unaffected vs. SRTS-Affected). Significant results are bolded ($p < 0.05$).

Parameter	Catchment Flow (Unaffected vs. Affected)		
	t	df	P
Ca^{2+}	-2.713	16.000	**0.015**
Cl^{-}	-0.265	16.000	0.795
K^{+}	-2.364	16.000	**0.031**
Mg^{2+}	-2.580	16.000	**0.020**
Na^{+}	-2.022	16.000	0.060
SO_4^{2-}	-3.035	16.000	**0.008**

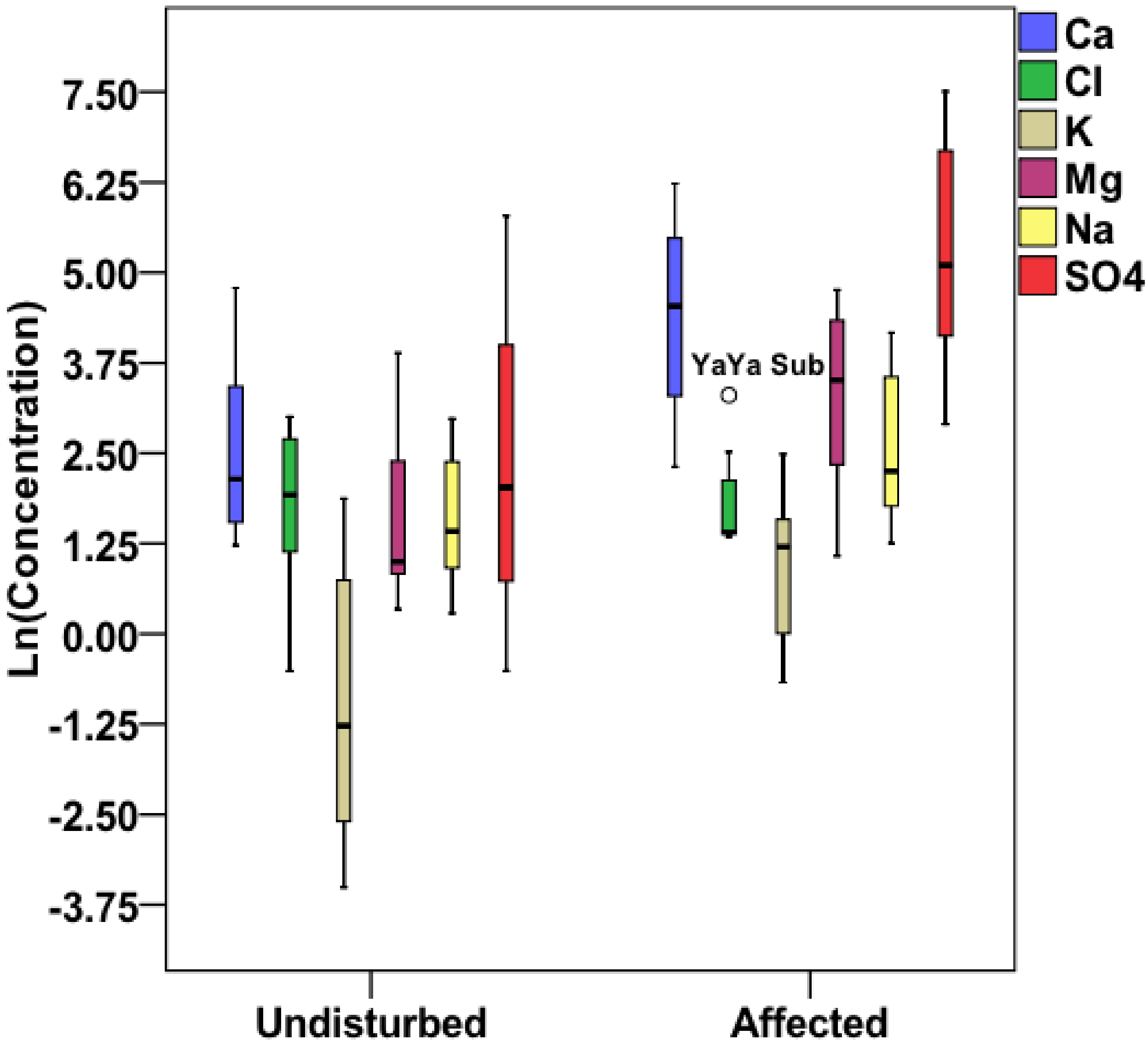


Figure 4.25. Box plots displaying the concentration of major ions (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) in catchment flow to the 9 regional study lakes. The dependent parameters are grouped based on Catchment Flow Type (Unaffected vs. Affected). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

4.5.3 Outflow

4.5.3.1 Lake 5A and 5B

Major Ions

The detailed geochemical data for the outflow draining Lake 5A and Lake 5B in 2008 and 2009 were pooled based on the Outflow Channel (5A Outflow vs. 5B Outflow) and Season (Spring Melt and Early Open-water vs. Mid to Late Open-water). The mean concentrations of Ca^{2+}, K^{+}, Mg^{2+}, Na^{+}, SO_4^{2-} in the outflow channel draining Lake 5B were significantly higher than that of the outflow channel draining Lake 5A (ANOVA, $p < 0.05$, **Table 4.10**, **Figure 4.26 to Figure 4.28**). This can be attributed to the elevated concentrations of Ca^{2+}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-} found in the source water, Lake 5B.

In the previous sections, it was established that SRTS at Lake 5B increases the supply of Ca^{2+}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-} to runoff, subsequently increasing the supply of these ions to Lake 5B. Based on the work on Moquin (2011), Moquin et al. (2014), Thompson et al. (2008), and Thompson et al. (2012), we know that these charged particles bind with dissolved organic matter in the water column and fall to the bottom of the lake in a process called sedimentation. The results presented in this section indicate that Lake 5B is not a sink for these major ions, sedimentation does not remove all of these charged particles from the water column. Rather, these charged particles are transported from Lake 5B via the outflow channel to the adjacent Noell Lake. In other words, SRTS affected lakes are a source of major ions to downstream lakes.

Interestingly, the concentration of Cl^{-} in the outflow channel draining Lake 5B was not significantly different from that of the outflow channel draining Lake 5A ($p > 0.05$). This is counterintuitive, since the concentration of Cl^{-} in Lake 5B was higher than that of Lake 5A. This suggests that, in contrast with the other major ions, Lake 5B is a sink for Cl^{-} .

The mean concentrations of Ca^{2+}, Cl^{-}, Mg^{2+}, Na+, and SO_4^{2-} in the outflow channel draining Lake 5A and Lake 5B exhibited significant variability between the two seasons. The mean concentrations of these ions were significantly higher during the mid to late open-water period than during the spring melt and early open-water period. In contrast with the above, the concentrations of K^{+} in the two outflow channels did not exhibit significant variability between the two seasons. This is counterintuitive because the concentration of K^{+} in lake water did exhibit significant seasonal variability. For most major ions, the water in Lake 5A and Lake 5B became more concentrated during the summer months due to an increase in mineral weathering and concentration by evaporation. As a result, ion-rich lake water was delivered from Lake 5A and Lake 5B to their respective outflow channels and, in turn, ion-rich water was delivered to downstream Noell Lake.

It's important to note that there was a statistically significant interaction effect between Outflow Channel and Season for the dependent variables Ca^{2+}, Mg^{2+}, Na^{+}, and SO_4^{2-}. That is, the influence of SRTS on the concentrations of Ca^{2+}, Mg^{2+}, Na^{+}, and SO_4^{2-} in the outflow channel is dependent on Season. Based on the mean concentrations presented in **Figure 4.26** to **Figure 4.28**, the mean differences in Ca^{2+}, Mg^{2+}, Na^{+}, and SO_4^{2-} concentrations between the two seasons is greater for the 5B Outflow than for the 5A Outflow.

Table 4.10. An ANOVA table for the parameters measured in the outflow channels draining Lake 5A and Lake 5B in 2008 to 2010. The dependent parameters (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) were tested according to Outflow Channel (5A Outflow vs. 5B Outflow) and Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	Outflow Channel (5A Outflow vs. 5B Outflow)			Season (Spring Melt & Early Open-water vs. Late Open-water)			Outflow Channel * Season		
	F	df	P	F	df	P	F	df	P
Ca^{2+}	54.516	1.000	**0.000**	78.035	1.000	**0.000**	16.175	1.000	**0.002**
Cl^{-}	1.155	1.000	0.304	13.664	1.000	**0.003**	3.412	1.000	0.089
K^{+}	13.466	1.000	**0.003**	3.089	1.000	0.104	3.542	1.000	0.084
Mg^{2+}	47.627	1.000	**0.000**	64.234	1.000	**0.000**	15.101	1.000	**0.002**
Na^{+}	40.488	1.000	**0.000**	77.553	1.000	**0.000**	22.550	1.000	**0.000**
SO_4^{2-}	81.664	1.000	**0.000**	71.586	1.000	**0.000**	4.990	1.000	**0.045**

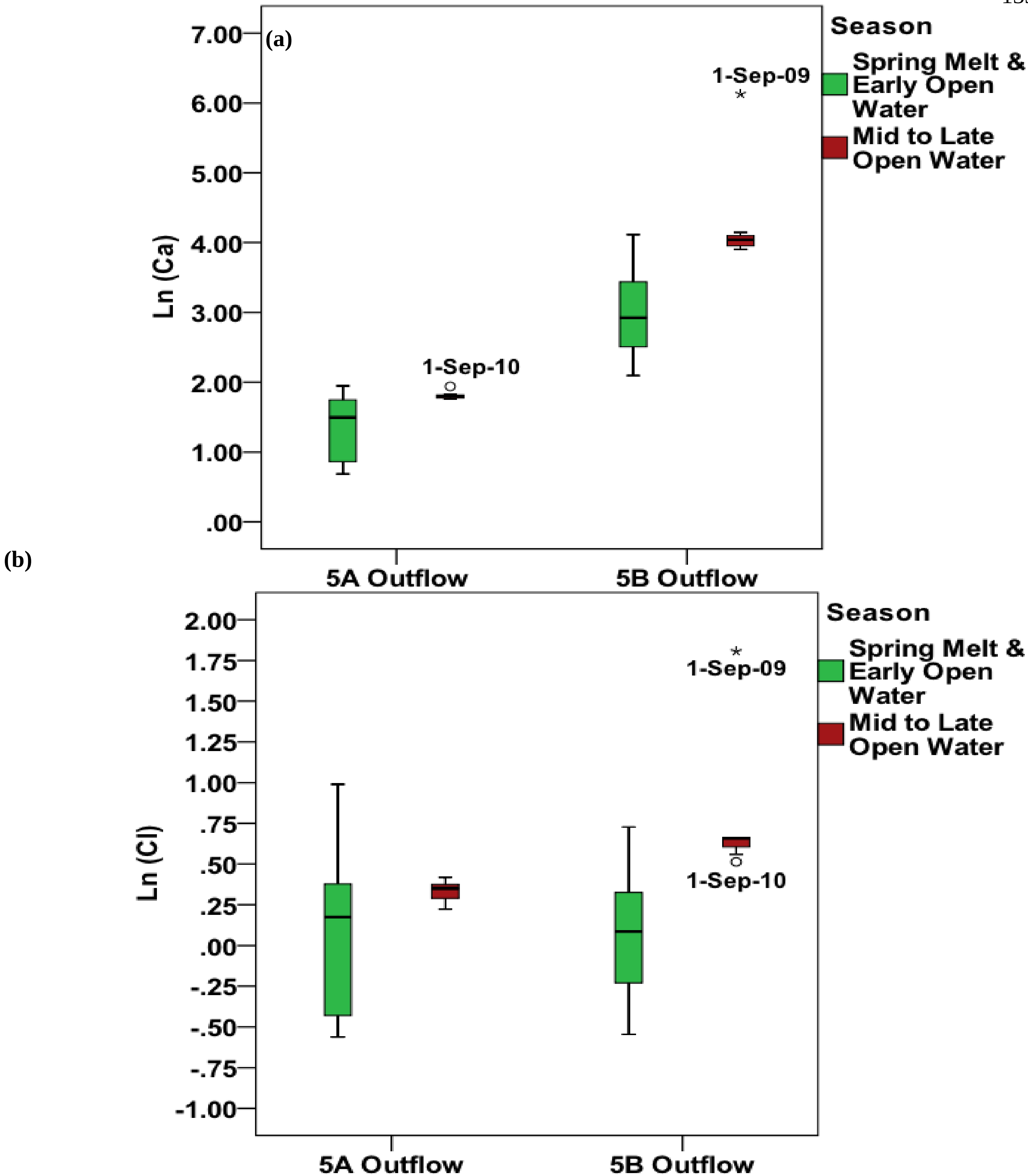


Figure 4.26. Box plots displaying the concentration of (a) Ca^{2+} and (b) Cl^{-} in the outflow channels draining Lake 5A and Lake 5B. The dependent parameters are grouped based on Lake (5A Outflow vs. 5B Outflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and the logarithmically transformed.

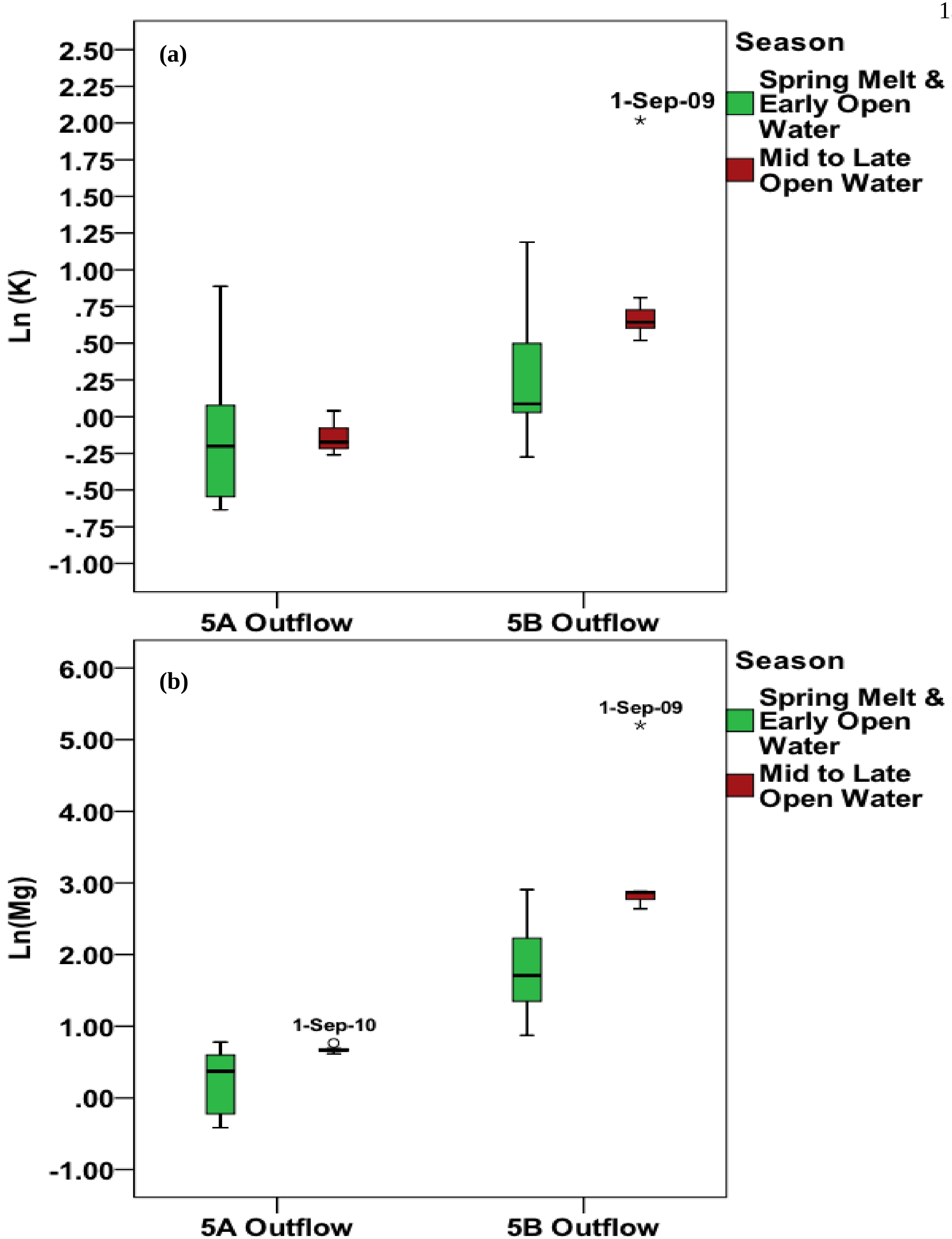


Figure 4.27. Box plots displaying the concentration of (a) K^+ and (b) Mg^{2+} in the outflow channels draining Lake 5A and Lake 5B. The dependent parameters are grouped based on Outflow Channel (5A Outflow vs. 5B Outflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically

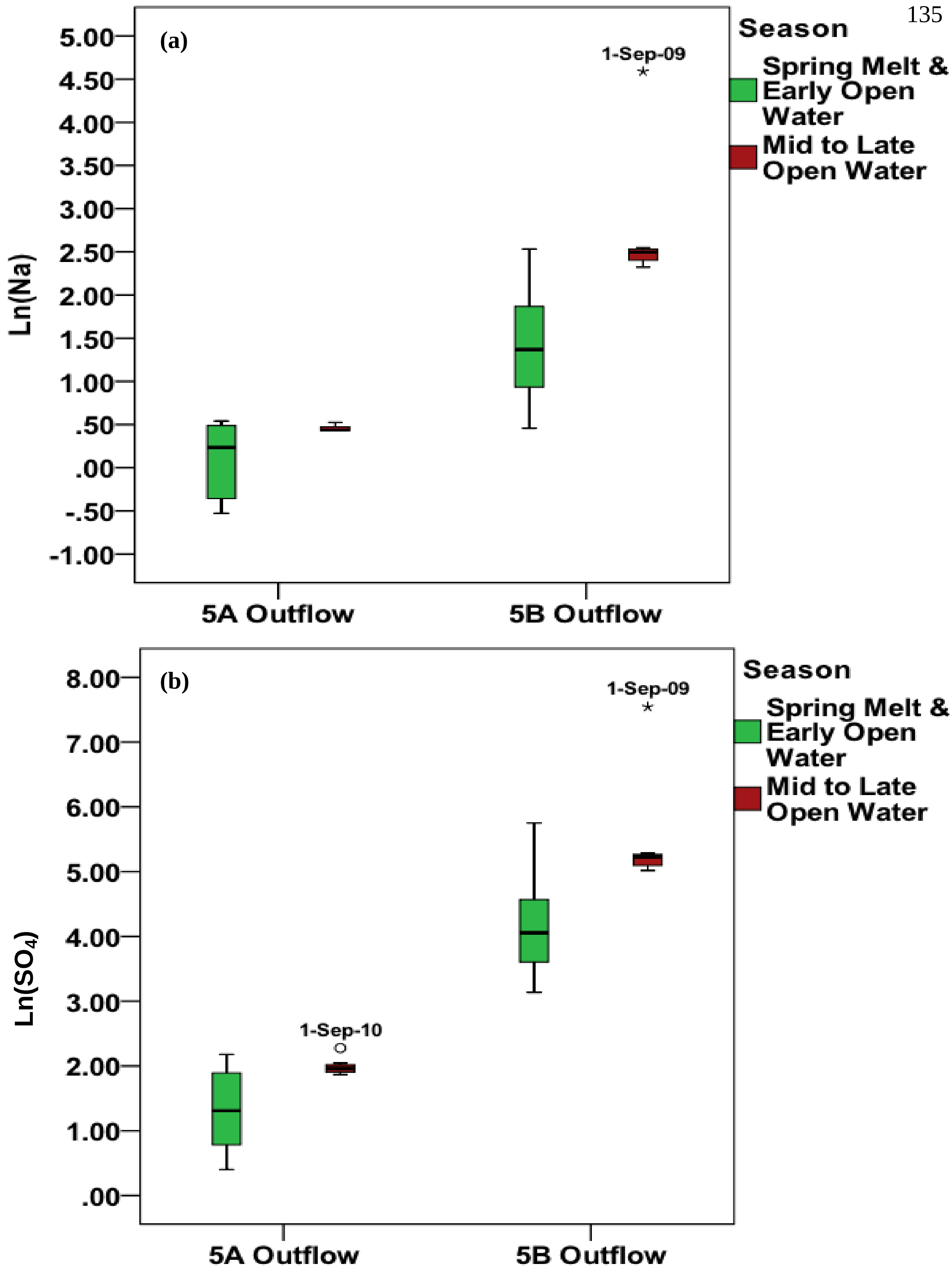


Figure 4.28. Box plots displaying the concentration of (a) Na^{+} and (b) SO_4^{2-} in the outflow channels draining Lake 5A and Lake 5B. The dependent parameters are grouped based on Lake (5A Outflow vs. 5B Outflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically

Nutrients

The mean concentrations of TP, TN, and TDN in the 5B outflow were not significantly different from that of the 5A Outflow (ANOVA, $p > 0.05$, **Table 4.11; Figure 4.30** and **Figure 4.31**). This is not entirely surprising, because the concentrations of TP, TN, and TDN in Lake 5B was not significantly different from those of Lake 5A.

Despite some apparent seasonal variability, the concentrations of TP, TN, and TDN in the 5A and 5B Outflows did not significantly vary between the two seasons. The concentrations of TP, TN, and TDN in the 5A Outflow was generally lower during open-water than during spring melt. This is similar to the general trends observed in the lake water collected at Lake 5A. Conversely, the concentration of TP, TN, and TDN in the 5B Outflow is higher during open-water than during spring melt. This is counterintuitive, since the concentration of TP in Lake 5B decreased during open-water. This suggests that the concentration of TP in the outflow may be more strongly driven by the adjacent landscape than by the lake water.

Table 4.11. An ANOVA table for the parameters measured in the outflow channels draining Lake 5A and Lake 5B in 2008 to 2010. The dependent parameters (TP, TN, and TDN) were tested according to Outflow Channel (5A Outflow vs. 5B Outflow) and Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). Significant results ($p < 0.05$) are bolded.

Parameter	Outflow Channel (5A Outflow vs. 5B Outflow)			Season (Spring Melt and Early-Open Water vs. Mid to Late Open-Water)			Outflow Channel * Season		
	F	df	P	F	df	P	F	df	P
TP	0.159	1.000	0.728	0.168	1.000	0.722	1.281	1.000	0.375
TN	0.273	1.000	0.653	0.005	1.000	0.952	2.662	1.000	0.244
TDN	0.022	1.000	0.895	1.088	1.000	0.406	0.658	1.000	0.503

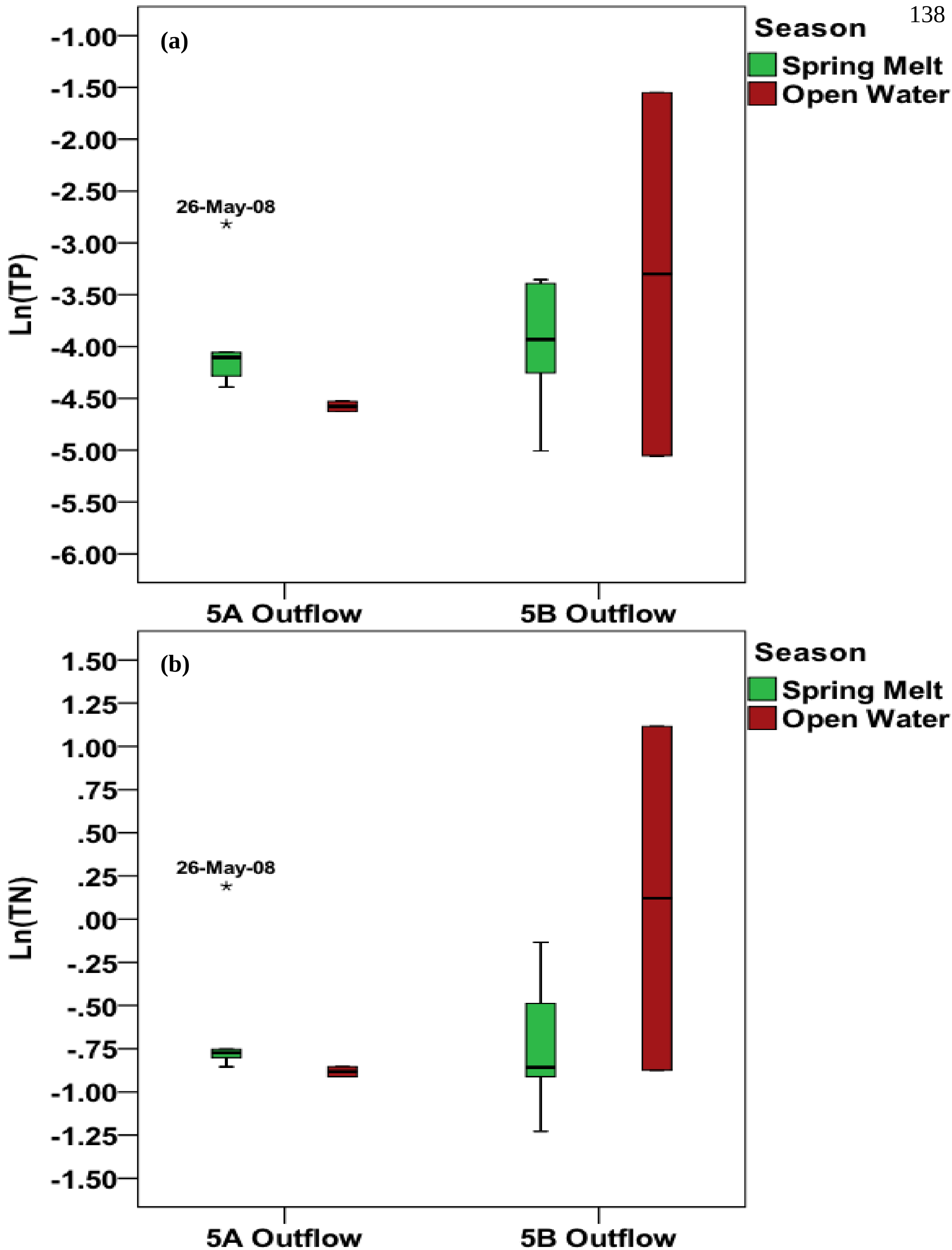


Figure 4.29. Box plots displaying the concentration of (a) TP and (b) TN in the outflow channels draining Lake 5A and Lake 5B. The dependent parameters are grouped based on Outflow Channel (5A Outflow vs. 5B Outflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

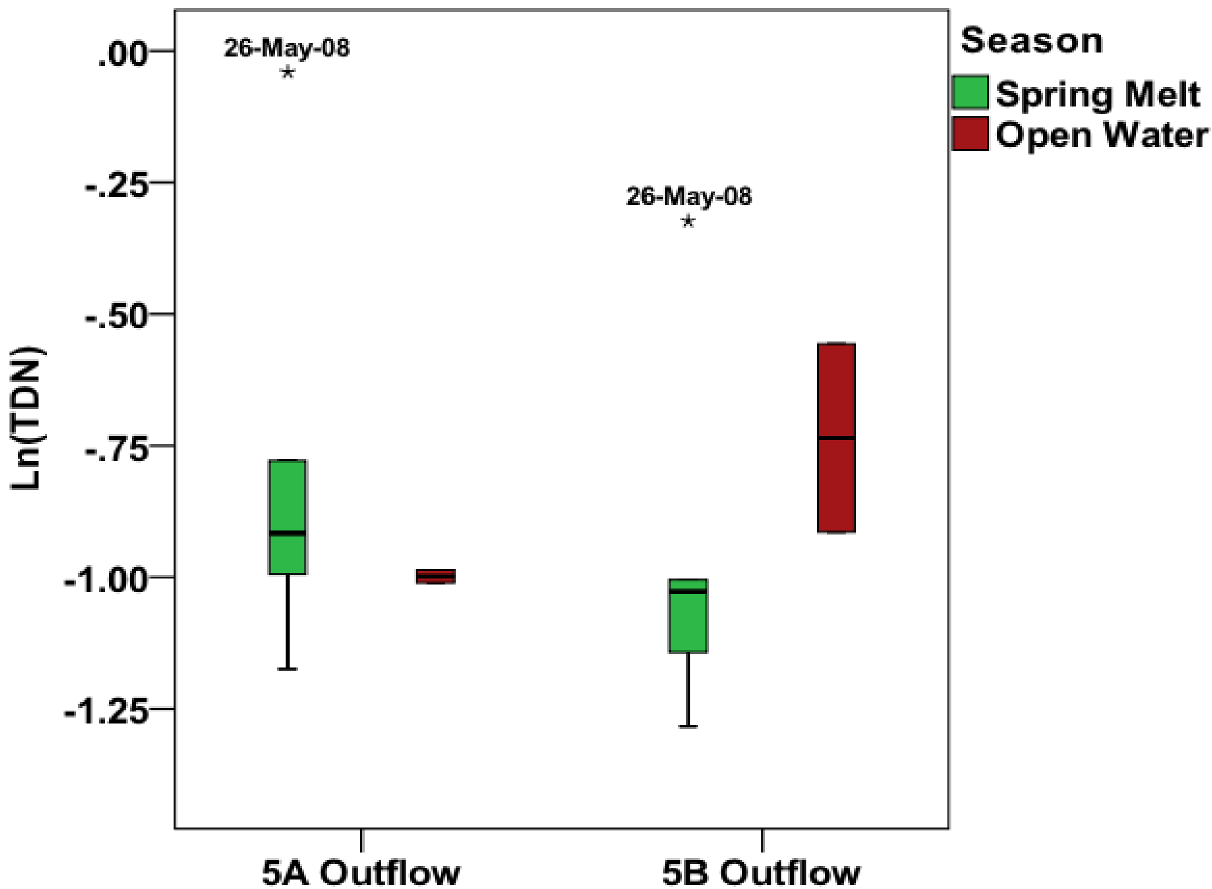


Figure 4.30. Box plots displaying the concentration of TDN in the outflow channels draining Lake 5A and Lake 5B. The dependent parameters are grouped based on Outflow Channel (5A Outflow vs. 5B Outflow) and categorized based on Season (Spring Melt & Early Open-water vs. Mid to Late Open-water). All data was measured in Mg.L^{-1} and then logarithmically transformed.

4.5.3.2 Regional Study Lakes

Major Ions

The geochemical data collected at the outflow channels draining the 9 regional lakes and the YaYa Subcatchment lake in 2009 and 2010 were pooled based on Lake Type (Unaffected vs. SRTS-affected). The mean concentrations of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} in the outflow channels draining SRTS-affected lakes was higher than that of the outflow channels draining unaffected lakes (T-Test, $p < 0.05$, **Table 4.12**, **Figure 4.32**). This is partially consistent with the results presented for Lake 5A and Lake 5B. SRTS increases the supply of some major ions to runoff, subsequently increasing the supply of major ions to SRTS-affected lakes. In turn, SRTS-affected lakes become a source of some major ions to the lakes that they drain into.

The mean concentrations of Cl^{-} and Na^{+} were not significantly different between the outflow channels draining SRTS-affected lakes and those of unaffected lakes. This is not surprising, given that the concentrations of Cl^{-} and Na^{+} in the lake water of SRTS-affected lakes were not significantly different from that of the unaffected lakes.

Overall, the results of this study suggest that small tundra lakes are not sinks for major ions, but sources of major ions for the downstream lakes their outflow channels drain into.

Table 4.12. A Independent Samples T-Test testing for the parameters measured in the outflow channels draining the 9 regional study lakes and the YaYa Subcatchment Lake in 2009 and 2010. The dependent parameters (Ca^{2+}, Cl^{-}, Mg^{2+}, Na^{+}, K^{+}, and SO_4^{2-}) were tested according to Outflow Channel (outflow channels draining SRTS-affected lakes and the outflow channels draining unaffected lakes). Significant results are bolded ($p < 0.05$).

Parameter	**Outflow (Unaffected vs. Affected)**		
	t	**df**	**P**
Ca^{2+}	-3.708	13.000	**0.003**
Cl^{-}	1.463	13.000	0.167
K^{+}	-3.858	13.000	**0.002**
Mg^{2+}	-3.075	13.000	**0.009**
Na^{+}	-0.207	13.000	0.839
SO_4^{2-}	-3.146	13.000	**0.008**

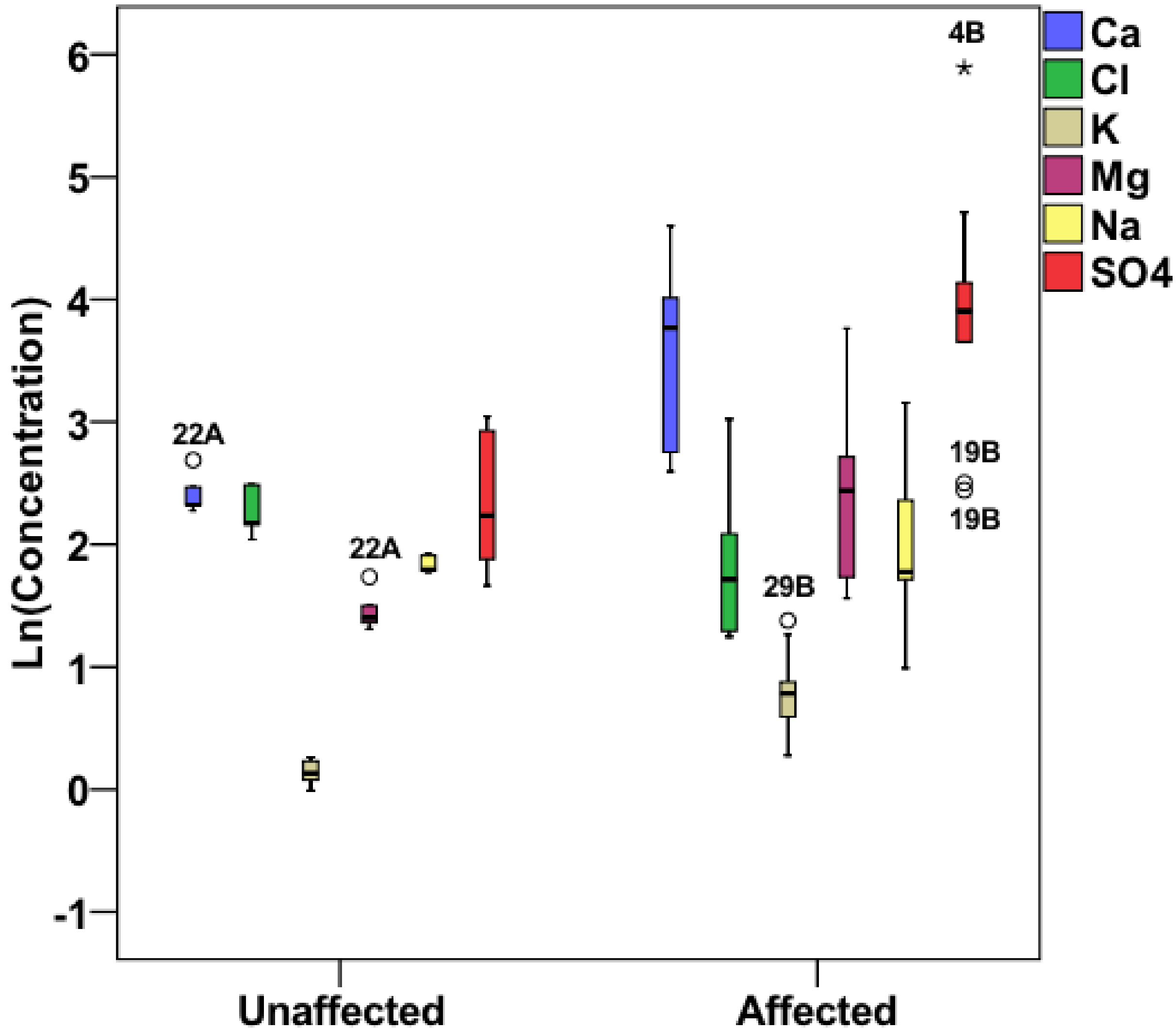


Figure 4.31. Box plots displaying the concentration of major ions (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) in the outflow channels draining the 9 regional study lakes. The dependent parameters are grouped based on Lake Type (Unaffected vs. Affected). All data was measured in $Mg.L^{-1}$ and then logarithmically transformed.

4.6 Summary and Conclusions

The overall goal of this chapter was to investigate the influence of the contributing catchment on the seasonal geochemistry of small tundra lakes affected by SRTS. The concentrations of major ions (Ca^{2+}, Cl^-, Mg^{2+}, K^+, Na^+, SO_4^{2-}) and TP in the two primary study lakes exhibited significant ($p < 0.05$) seasonal variability that was correlated with ice formation and ablation, spring snowmelt, summer rainfall, run-off, and evaporation. In contrast, the concentrations of TN and TDN did not exhibit significant seasonal variability ($p > 0.05$). The concentrations of major ions in the two primary study lakes typically increased over the winter months, due to ion exclusion during ice formation, reaching an annual peak in late-winter. Then, declined during spring snow and ice melt, due to dilution by melt water inputs. In the absence of summer rainfall events, the concentrations of major ions in the two primary study lakes increased over the open-water period, due to concentration by evaporation and the addition of ion rich runoff water. The concentration of Ca^{2+}, Cl^-, K^+, and Mg^{2+} decreased at points during the open-water period at Lake 5A, in response heavy rainfall events. Notably, the concentration of Ca^{2+}, Cl^-, K^+, and Mg^{2+} increased over the entire open water period at Lake 5B, due to the addition of ion-rich runoff from SRTS-affected terrain. This suggests that runoff generated by summer rainfall events could be a key factor driving the concentration of Ca^+, Cl^-, K^+, and Mg^{2+} in SRTS-affected lakes.

The concentration of TP in the two primary study lakes increased over the winter season, due to ion exclusion during ice formation. At the beginning of spring snowmelt, TP continued to increase, due to the mobilization of organic materials by surface run-off, but then declined over the rest of the snowmelt period due to the addition of relatively dilute runoff water. In the absence of heavy rainfall events, the concentration of TP in runoff to Lake 5A and Lake 5B increased over the open-water period. In contrast, the concentration of TP in lake water decreased over the open water period. This was attributed to dilution via rainfall inputs, higher rates of sedimentation, and increases in biological uptake.

The geochemistry of the two primary study lakes and the 10 regional study lakes was strongly driven by SRTS. Overall, SRTS-affected lakes had higher concentrations of Ca^{2+}, Mg^{2+}, K^+, and SO_4^{2-} than unaffected lakes due to the addition of ion-rich runoff from SRTS-affected terrain. Notably, the concentration of Ca^{2+}, Mg^{2+}, K^+, and SO_4^{2-} in slumpflow was significantly higher than that of run-off from unaffected soils, indicating that SRTS increases the supply of these ions to runoff and eventually to lakes. In contrast with the results presented for the 10 regional study lakes, the concentration of Cl^- and Na in Lake 5B was higher than that of Lake 5A due to the addition of Na and Cl^- rich run-off from the adjacent shoreline slump. This suggests that the effects of SRTS on the concentration of Na and Cl^- in small tundra lakes may vary between study lakes, depending on the surficial geology of the contributing lake catchment.

The effects of SRTS do not stop at the adjacent lake. The results of this study indicate that SRTS-affected lakes increase the supply of Ca^{2+}, Mg^{2+}, K^+, and SO_4^{2-} to the ephemeral outflow channels draining them. For the two primary study lakes and the 10 regional study lakes, the mean concentration of Ca^{2+}, Mg^{2+}, K^+, and SO_4^{2-} in the outflow channels draining SRTS-affected lakes was significantly higher than that of the outflow channels draining unaffected lakes.

Given that small tundra lakes in this region are often connected via their drainage channels, SRTS likely has a broader effect on freshwater systems in the study region than was anticipated by previous studies. For instance, the two primary study lakes drain into downstream Noell Lake, which eventually drains into the Husky Lakes and the Arctic Ocean. Increases in the flux of charged particles through the drainage channels of SRTS-affected lakes could affect the ecology of downstream lakes. As was noted earlier, increases in the concentration of charged particles within the water column leads to increases in the rate of sedimentation of dissolved organic matter (Thompson et al., 2008). This leads to decreases in both lake water colour and the attenuation of light through the water column. Furthermore, higher rates of sedimentation lead to lower nutrient concentrations within the water column and higher nutrient concentrations within the lake bottom sediments (Moquin et al., 2014; Thompson et al., 2012). Changes in key processes driving production, such as the availability of photosynthetically active radiation and the availability of nutrients, directly affects lake ecology.

SRTS may be an extreme form of permafrost degradation that only affects approximately 8% of the thousands of small tundra lakes located in the study region. Active layer thickening, associated with warmer ambient air and ground temperatures and the thawing of near surface permafrost, has and will likely continue to affect the majority of lakes in the study region. The results of this study indicate that near surface permafrost is a significant source of major ions to runoff pathways and in turn, to small tundra lakes. Furthermore, increases in the flux of charged particles, associated with the thawing of near surface permafrost, has significant affects on the ecology of freshwater systems.

Chapter 5: Summary, Conclusions, and Future Directions

5.1 Summary and Conclusions

The overall goal of this thesis was to further our understanding of the hydrological and geochemical linkages between the contributing landscape and small tundra lakes affected by shoreline retrogressive thaw slumping (SRTS). The results of this study demonstrate that the primary factors controlling the water balance and geochemistry of small tundra lakes are ice formation and ablation, the addition of snowmelt water in spring, snow and ice damming the outlet channel, evaporation, summer rainfall events, lake drainage, and shoreline retrogressive thaw slumping (SRTS).

The hydrology and geochemistry of the study lakes exhibited strong seasonal variability. The Lake Level (LL) of the two primary study lakes, Lake 5A and Lake 5B, increased slightly over the ice-covered months, due to the growing weight of the overlying snow. In addition to LL, the concentrations of major ions (Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, Na^{+}, and SO_4^{2-}) and TP also increased over the winter months, due to particle exclusion associated with ice formation. The mean concentration of major ions in Lake 5A and Lake 5B was significantly ($p < 0.05$) higher in late-winter than during the spring melt and open water periods. This was substantiated by the results presented for the 10 regional study lakes. In contrast with Lake 5A and Lake 5B, there was significant seasonal variability in the mean concentration of TDN in the 10 regional study lakes, whereby TDN was significantly higher during the ice covered months than during the spring melt and open-water periods. This was also attributed to the processes associated with ice formation.

The addition of snowmelt water from the contributing landscape in early spring led to a rapid rise in LL that was enhanced by snow and ice damming the outlet channel. Notably, the annual peak LL occurred during the early stages of spring snowmelt. During the later stages of spring snowmelt, lake water had carved a trench through the snow/ice dam in the outlet channel, initiating lake drainage and a rapid decline in LL. Meanwhile, the concentration of major ions in the two primary study lakes decreases substantially, due to the addition of relatively dilute runoff water. The mean concentrations of major ions in both landscape runoff and lake water was significantly lower during the spring snowmelt period than during the ice-covered and open-water periods. This was also substantiated by the results presented for the 10 regional study lakes. Also in contrast with Lake 5A and Lake 5B, the mean concentration of TDN in the 10 regional study lakes was significantly lower during the spring snowmelt period than during the ice covered and open-water periods. This was also attributed to dilution via spring snowmelt.

In contrast with the general trends observed in major ion concentrations, the concentration of TP in the two primary study lakes increased at the beginning of spring snowmelt, due to the addition of relatively nutrient rich runoff water from the contributing lake catchment. The initiation of spring snowmelt liberated nutrient-rich organic material from the contributing lake catchment, leading to an increase in the concentration of nutrients in runoff (Quinton and Pomeroy, 2006). As spring snowmelt progressed, the concentrations of TP in both landscape runoff and lake water decreased. With the progression of spring snowmelt, the soil organic layer likely became saturated, forcing runoff to move on top of the soil organic layer rather than through it (Quinton and Pomeroy, 2006). Overall, the mean concentration of TP in Lake 5A and Lake 5B was significantly ($p < 0.05$) higher during the spring melt period than during the ice covered and open-water periods. Notably, no significant ($p > 0.05$) seasonal variability in the concentration of TP in the 10 regional study lakes was observed.

Changes in arctic snow cover, associated with historical climate variability and change, has important implications for the hydrology and geochemistry of small tundra lakes. The Annual May 1st Snowpack in Tuktoyaktuk increased at a significant ($p < 0.05$) rate from 1958 to 2009. This is in line with recent climate projections, which indicated that projected climate warming will lead to an increase in snowcover depth and duration for coastal regions of the Arctic, such as Tuktoyaktuk (AMAP, 2012). A larger May 1st snowpack could lead to a greater rise in LL during spring snowmelt for small tundra lakes located at the northern end of the study region, which could lead to a higher peak LL and more dilution of lake water.

The results of this study indicate that SRTS increases the snow water equivalent (SWE) of small tundra lake catchments in the study region. In 2008 and 2009, the SWE of the SRTS-affected terrain at Lake 5B was 21% and 28% greater, respectively, than the adjacent unaffected terrain, leading to a 2% increase in snowmelt water inputs from the contributing catchment and a 2cm increase in the LL rise associated with spring snowmelt. These results indicate that SRTS could also lead to a greater rise in LL during spring snowmelt, a higher peak LL, and more dilution of lake water.

There was a significant decline in the timing of the spring freshet and the timing of ice-off in Tuktoyaktuk for the years 1958 to 2009. This is in line with other studies, which have indicated that climate warming has and will likely continue to lead to earlier spring snow and ice melt initiation for many arctic regions (AMAP, 2012; Bonsal and Prowse, 2003; Dibike et al., 2012). In contrast with Tuktoyaktuk, there was no statistically significant trend in the timing of spring snowmelt and the timing of ice-off in Inuvik for the same time period. This is in line with the work of Lesack et al. (2013), who did not observe any trends in the timing of the spring freshet at the East Branch of the Mackenzie River, at Inuvik. The effects of recent climate warming are generally amplified in arctic coastal regions, relative to inland regions, due to the overwhelming effects of declining sea ice extent on ambient air temperature (AMAP, 2012). This could be why changes in the timing of snow and ice melt observed in Tuktoyaktuk were not observed in Inuvik.

Summer rainfall events were an important source of water recharge for the two primary study lakes. In the absence of summer rainfall events, there was generally a decline in the LL of the two primary study lakes over the open-water period of the three study years due to lake drainage and evaporation. For instance, the LL of the two primary study lakes decreased over the entire open-water period in 2007, due to low summer rainfall inputs and high rates of evaporation. By comparison, summer rainfall events were an important source of water recharge to the study lakes in 2008 and 2009, which led to periods of LL rise in July, August, and September.

In addition to LL, this study revealed that summer rainfall events had a notable effect on lake water chemistry. The geochemistry of the study lakes was highly variable over the open-water period. In the absence of rainfall events, the concentration of major ions typically increased over the summer months. At Lake 5A, summer rainfall events led to a decrease in the concentration of Ca^{2+}, Cl^{-}, Na^{+}, and K^{+}, and an increase in the concentration of Mg^{2+} and SO_4. By comparison, summer rainfall events led to increases in all major ions in Lake 5B. These results indicate that summer rainfall events facilitate the leaching of major ions out of slump soils. Overall, the mean concentration of major ions in the two primary study lakes was significantly ($p < 0.05$) higher during the mid to late open-water period than during the spring melt and early open-water period. This result was substantiated by the results presented for the 10 regional study lakes. In contrast with the above, the concentration of TP in Lake 5A and Lake 5B decreased over the open-water period. This was attributed to increases in dilution via rainfall, sedimentation, and biological uptake. Overall, the mean concentration of TP in the two primary study lakes was significantly ($p < 0.05$) lower during the mid to late open-water period than during the spring melt and early open-water period. No significant ($p > 0.05$) seasonal variability in the concentration of TP in the 10 regional study lakes was observed.

The geochemistry of small tundra lakes in the study region was also driven by SRTS. The results of this study indicate that major ions leach out of SRTS-affected soils is surface runoff, leading to an increase in the ionic concentration of SRTS-affected lakes. For the two primary study lakes, Lake 5B had higher concentrations of all major ions than Lake 5A, due to the addition of ion rich runoff from SRTS-affected terrain. This was partially substantiated by the results presented from the 10 regional study lakes. For the 10 regional study lakes, SRTS-affected lakes had higher concentrations of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} than unaffected lakes, due to the addition of ion-rich runoff from SRTS-affected terrain. Similar to Lake 5A and Lake 5B, the mean concentrations of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} in slumpflow were significantly higher than runoff from unaffected terrain. This agrees with the postulation made by Kokelj et al. (2009), which was that major ions leach out of SRTS-affected soils over time. Notably, the concentration of TP, TDN, and TN was not significantly different in unaffected and SRTS-affected lakes.

The outflow from SRTS-affected lakes was a source of major ions to adjacent lakes. For the two primary study lakes, the mean concentrations of Ca^{2+}, Cl^{-}, K^{+}, Mg^{2+}, and SO_4^{2-} in the outlet channel draining Lake 5B were higher than that of Lake 5A. This was partially substantiated by the results presented for the 10 regional study lakes. For the 10 regional study lakes, the mean concentrations of Ca^{2+}, K^{+}, Mg^{2+}, and SO_4^{2-} in the outlet channels draining SRTS-affected study lakes were significantly higher than that of unaffected lakes. The outflow channels that drain small tundra lakes in the study region often flow into adjacent lakes. For instance, the two primary study lakes drain into downstream Noell lake. The results of this study indicate that SRTS-affected lakes could be a source of major ions to the downstream lakes that they drain into. As was noted in the previous chapter, increases in the flux of charged particles through the drainage channel of SRTS-affected lakes could affect the ecology of downstream lakes.

Overall, the results of this study further our understanding of how the water balance and geochemistry of small tundra lakes in the study region have been and will continue to be impacted by climate variability and change and SRTS.

5.2 Future Directions

Our understanding of the impacts of climate variability and change and SRTS on the water balance and geochemistry of small tundra lakes in northwestern Canada could be improved by targeted studies investigating the following:

- Detailed snow surveys taken at a range of SRTS-affected lake catchments could be used to examine the relative importance of slump size on the SWE of the contributing landscape.
- Detailed hydrologic measurements taken at the outflow channels would allow us to examine the relative importance of evaporation and lake drainage in driving the water balance of the study lakes, particularly during the spring snowmelt and open-water periods.
- More frequent, event-based sampling of landscape and lake water during the open water period would improve our understanding of how summer rainfall events affect the concentrations of major ions and nutrients in small tundra lakes. It would also be valuable in assessing the relative importance of summer rainfall, evaporation, and runoff in driving the geochemistry of small tundra lakes during the open-water period.
- Detailed water quality sampling of Noell Lake and surrounding lakes to examine the impact of how SRTS-affected lakes impact the geochemistry of downstream lakes.
- Sampling of active layer and permafrost water in addition to ephemeral rills would give us the opportunity to examine how much subsurface flow influences the geochemistry of the inflows leading to the study lakes.
- It would be interesting to explore how other types of disturbances (i.e., fire, construction) affect the geochemistry of runoff to small tundra lakes.

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Appendix A: Study Lakes

Table A.1. Physical characteristics of the 11 study lakes.

Study Lake	Lake Surface Area (LA) ha	Catchment Area (CA) ha	LA:CA	Maximum Depth m
5A	2.80	20.20	0.14	10.9
5B	3.29	25.01	0.13	10.9
22A	1.87	8.31	0.23	3.7
25A	4.88	28.48	0.17	5.6
30A	4.36	23.00	0.19	8.4
8B	6.49	32.71	0.20	2.8
16B	14.10	62.97	0.22	6.6
19B	6.11	28.11	0.22	7.0
22B	3.52	11.70	0.30	6.9
24B	7.56	40.26	0.19	2.9
29B	5.72	14.26	0.40	6.7
YaYa Sub	No Data	No Data	No Data	No Data

Appendix B: Infilling Missing Field Data

B.1 Lake Level

For each study lake, LL was pooled based on season (Winter, Spring, Early Summer, and Late Summer). Winter was defined as the period of time falling between October 1st and the onset of the spring freshet. Spring was defined as the period of time falling between the initiation of the spring freshet and the onset of the open-water period. Early summer was defined as the period of time falling between the onset of the open-water period and the end of July. Late summer was defined as August and September. Linear regression models were created, based on season, between LL at Lake 5A and LL at Lake 5B (**Table B.1** and **Table B.2**). In Winter, missing data was left blank. In Spring, early Summer, and late Summer, missing data was infilled using the using the appropriate linear regression equation.

Table B.1. Average daily water level relationships between Lake 5A and Lake 5B.

Year	Linear Regression Equation	R^2	N
Winter	5A = (0.59 x 5B) – 49.15	0.41	237
Spring	5A = (1.08 x 5B) – 16.87	0.59	24
Early - Summer	5A = (0.99 x 5B) – 5.02	0.95	79
Late-Summer	5A = (0.95 x 5B) + 0.99	0.99	126

Table B.2. Average daily water level relationships between Lake 5B and Lake 5A.

Year	Linear Regression Equation	R^2	N
Winter	5B = (0.70 x 5A) + 44.30	0.41	237
Spring	5B = (0.55 x 5A) + 63.905	0.59	24
Early-Summer	5B = (0.96 x 5A) + 11.22	0.95	79
Late-Summer	5B = (1.05 x 5A) + 0.04	0.99	126

B.2 Air Temperature

Lake 5A was used as the primary weather station. If average daily air temperature data was not available for Lake 5A, Lake 5B was used. After air temperature data from the two weather stations was combined, missing air temperature data accounted for approximately 14% of the study period (2007, 2008, and 2009). Missing data was infilled using a between-station regression-based approach, which has been found to be more accurate than within-station methods (Kemp et al, 1983; Degaetano et al, 1995). The average daily air temperature data was pooled based on month. On a monthly basis, linear regression models were created between the average daily air temperature at Lake 5A/Lake 5B and each of the two nearest weather stations (**Tables B.3** to **Table B.5**). The closest weather station with available data was always used.

Table B.3. Information for the climate stations used in data infilling.

Weather Station	Climate ID	Latitude	Longitude	Distance from Lake 5A (Km)
Lake 5A		68° 33’ 04.211” N	133° 38’ 23.387” W	0
Lake 5B		68° 32’ 15.323” N	133° 39’ 27.179” W	1.68
Trail Valley Creek	220N005	68° 44’ 46.800” N	133° 30’ 06.400” W	27.14
Inuvik Climate	2202578	68° 19’ 00.000” N	133° 31’ 00.000” W	29.96

Table B.4. Average daily temperature relationships between Trail Valley Creek (TVC) and Lake 5A.

Month	Linear Regression Equation	R^2	N
October	5A = (1.14 x TVC) + 0.07	0.95	92
November	5A = (0.98 x TVC) – 0.66	0.91	89
December	5A = (0.93 x TVC) – 2.21	0.92	90
January	5A = (1.06 x TVC) + 0.39	0.97	70
February	5A = (1.00 x TVC) – 1.03	0.96	34
March	5A = (0.93 x TVC) – 2.07	0.98	31
April	5A = (0.97 x TVC) – 0.05	0.98	30
May	5A = (0.98 x TVC) – 0.05	0.97	121
June	5A = (0.98 x TVC) + 0.78	0.92	119
July	5A = (0.93 x TVC) + 1.32	0.93	124
August	5A = (1.04 x TVC) – 0.33	0.95	123
September	5A = (0.98 x TVC) – 0.11	0.95	112
AVERAGE	5A = (0.99 x TVC) – 0.33	0.95	86

Table B.5. Average daily temperature relationships between Inuvik (INK) and Lake 5A.

Month	Linear Regression Equation	R^2	N
October	5A = (1.13 x INK) + 0.03	0.96	93
November	5A = (1.01 x INK) – 0.64	0.94	90
December	5A = (0.98 x INK) – 1.31	0.94	93
January	5A = (0.92 x INK) – 2.82	0.98	71
February	5A = (0.92 x INK) – 3.16	0.97	47
March	5A = (0.89 x INK) – 3.69	0.94	31
April	5A = (0.95 x INK) – 2.04	0.97	30
May	5A = (0.97 x INK) – 1.37	0.96	121
June	5A = (0.86 x INK) – 0.38	0.90	119
July	5A = (0.91 x INK) + 0.09	0.97	124
August	5A = (0.95 x INK) + 0.20	0.98	124
September	5A = (0.95 x INK) – 0.32	0.96	112
AVERAGE	5A = (0.95 x INK) – 1.28	0.96	88

B.3 Summer Rainfall

Missing rainfall data from Lake 5A was infilled using rainfall data from Lake 5B. After the total daily precipitation data from the two weather stations was combined, missing total daily rainfall was infilled using an anomaly-based approach, based on the methodology outlined by Goulding (2009). The average daily total precipitation, over the three year study period (e.g., 2007, 2008, and 2009), was determined for Lake 5A, Trail Valley Creek, and Inuvik. When a missing value was encountered, the departure of that day from the average daily value of the nearest weather station with available data was applied to the average daily value for Inuvik.

Appendix C: Inuvik Climate Data

C.1 Air Temperature

In 2009, Environment Canada published their First Generation Homogenized Temperature dataset (Vincent et al., 2009), which is housed in their Adjusted Historical Canadian Climate Data archive. Long-term temperature datasets, collected from weather stations across Canada, were corrected for changes in station location and measurement practices. Adjusted daily average, maximum, and minimum air temperature data was available for Inuvik, for the years 1958 to 2007, through the Adjusted Historical Canadian Climate Data archive (Climate ID: 2202578). Supplemental, unadjusted air temperature data was available for the years 1958 to 2006 (Inuvik A; Climate ID: 2202570) and 2003 to 2009(Inuvik Climate; Climate ID: 2202578), through Environment Canada’s Historical Climate Data Archive.

The Inuvik Climate station was the primary weather station. If air temperature data was not available for the Inuvik Climate station, data from the Inuvik A station was used. After the two Inuvik stations were combined, missing data accounted for approximately 1%, 1%, and 1% of the average, maximum, and minimum daily air temperature datasets, respectively. Missing data was infilled using a between-station regression-based approach, which has been found to be more accurate than within-station methods (Kemp et al, 1983; Degaetano et al, 1995). Average daily air temperature was pooled based on month. On a monthly basis, linear regression models were created between the average daily air temperature at Inuvik and each of the three nearest weather stations (**Table C.1** to **Table C.4**). For each missing day, the average daily air temperature at Trail Valley Creek was used for infilling. Even though Tuktoyaktuk is closer to Inuvik than Aklavik, there is a stronger relationship between the average daily air temperature in Inuvik and Aklavik than there is between Inuvik and Tuktoyaktuk. So, if average daily air temperature data was unavailable for Trail Valley Creek, data from Aklavik was used for infilling. The same methodology was used to infill the minimum daily and maximum daily air temperature datasets (**Table C.5** to **Table C.10**).

Table C.1. Information for climate stations used in data infilling.

Weather Station	Climate ID	Latitude	Longitude	Distance from Inuvik Climate (Km)
Inuvik Climate	2202578	68° 19’ 00.000” N	133° 31’ 00.000” W	0
Inuvik A	2202570	68° 18’ 15.000” N	133° 28’ 58.000” W	3.78
Trail Valley Creek	220N005	68° 44’ 46.800” N	133° 30’ 06.400” W	47.95
Tuktoyaktuk	2203910	69° 27’ 00.000” N	133° 00’ 00.000” W	137.39
Aklavik	2200100	68° 13’ 24.000” N	135° 00’ 21.000” W	154.78

Table C.2. Average daily temperature relationships between Trail Valley Creek (TVC) and Inuvik (INK).

Month	Linear Regression Equation	R^2	N
October	INK = (1.01 x TVC) + 0.17	0.94	340
November	INK = (0.97 x TVC) – 0.49	0.85	329
December	INK = (0.95 x TVC) – 1.15	0.90	338
January	INK = (1.07 x TVC) + 1.37	0.93	337
February	INK = (0.99 x TVC) – 0.19	0.92	264
March	INK = (1.00 x TVC) + 0.53	0.88	315
April	INK = (1.01 x TVC) + 1.73	0.92	257
May	INK = (0.93 x TVC) + 1.76	0.90	260
June	INK = (0.96 x TVC) + 3.15	0.83	229
July	INK = (0.98 x TVC) + 1.72	0.93	313
August	INK = (1.01 x TVC) + 0.86	0.93	340
September	INK = (1.00 x TVC) + 0.60	0.94	359
AVERAGE	INK = (0.99 x TVC) + 0.84	0.91	307

Table C.3. Average daily temperature relationships between Aklavik (AK) and Inuvik (INK).

Month	Linear Regression Equation	R^2	N
October	INK = (0.97 x AK) – 0.34	0.90	835
November	INK = (0.97 x AK) – 0.38	0.84	849
December	INK = (0.92 x AK) – 1.46	0.79	739
January	INK = (0.96 x AK) – 0.56	0.86	811
February	INK = (0.99 x AK) – 0.05	0.87	747
March	INK = (1.01 x AK) + 0.24	0.86	893
April	INK = (1.01 x AK) + 0.43	0.90	815
May	INK = (1.04 x AK) + 0.61	0.91	881
June	INK = (0.93 x AK) + 1.56	0.85	748
July	INK = (0.98 x AK) + 0.57	0.87	642
August	INK = (0.98 x AK) + 0.24	0.87	761
September	INK = (1.06 x AK) – 0.41	0.90	824
AVERAGE	INK = (0.99 x AK) + 0.04	0.87	795

Table C.4. Average daily temperature relationships between Tuktoyaktuk (TK) and Inuvik (INK).

Month	Linear Regression Equation	R^2	N
October	INK = (1.02 x TK) – 0.13	0.88	1540
November	INK = (1.04 x TK) – 0.20	0.84	1483
December	INK = (1.05 x TK) + 0.59	0.83	1543
January	INK = (1.09 x TK) + 2.26	0.86	1507
February	INK = (1.07 x TK) + 3.10	0.88	1367
March	INK = (1.04 x TK) + 3.59	0.84	1544
April	INK = (1.05 x TK) + 4.25	0.86	1506
May	INK = (1.02 x TK) + 4.44	0.83	1580
June	INK = (0.97 x TK) + 5.53	0.78	1522
July	INK = (0.87 x TK) + 4.62	0.80	1578
August	INK = (1.00 x TK) + 1.59	0.84	1581
September	INK = (1.10 x TK) + 0.37	0.87	1512
AVERAGE	INK = (1.03 x TK) + 2.50	0.84	1522

Table C.5. Maximum daily temperature relationships between Trail Valley Creek (TVC) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (1.05 x TVC) + 0.21	0.94	340
November	INK = (0.94 x TVC) - 0.90	0.87	329
December	INK = (0.92 x TVC) - 1.32	0.90	339
January	INK = (1.03 x TVC) + 0.34	0.92	337
February	INK = (0.98 x TVC) - 0.28	0.92	264
March	INK = (0.96 x TVC) - 0.03	0.83	316
April	INK = (0.97 x TVC) + 1.27	0.91	258
May	INK = (0.97 x TVC) + 1.97	0.88	260
June	INK = (0.86 x TVC) + 4.95	0.81	230
July	INK = (0.93 x TVC) + 2.55	0.92	313
August	INK = (0.99 x TVC) + 1.04	0.94	340
September	INK = (1.00 x TVC) + 0.62	0.94	359
AVERAGE	INK = (0.97 x TVC) + 0.87	0.90	307

Table C.6. Maximum daily temperature relationships between Aklavik (AK) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (0.95 x AK) + 0.22	0.89	836
November	INK = (0.89 x AK) - 1.45	0.79	860
December	INK = (0.85 x AK) - 2.57	0.77	758
January	INK = (0.86 x AK) - 2.61	0.78	816
February	INK = (0.93 x AK) - 1.15	0.85	756
March	INK = (0.95 x AK) - 1.08	0.85	900
April	INK = (0.97 x AK) - 0.21	0.89	836
May	INK = (1.00 x AK) + 0.50	0.85	874
June	INK = (0.88 x AK) + 3.62	0.80	775
July	INK = (0.94 x AK) + 2.12	0.85	686
August	INK = (0.94 x AK) + 1.66	0.84	790
September	INK = (0.99 x AK) + 0.65	0.88	835
AVERAGE	INK = (0.93 x AK) - 0.03	0.84	810

Table C.7. Maximum daily temperature relationships between Tuktoyaktuk (TK) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (1.04 x TK) + 0.55	0.86	1545
November	INK = (0.96 x TK) - 0.93	0.81	1489
December	INK = (0.98 x TK) - 0.35	0.82	1566
January	INK = (1.03 x TK) + 1.25	0.85	1509
February	INK = (1.00 x TK) + 1.68	0.85	1366
March	INK = (0.98 x TK) + 3.37	0.81	1545
April	INK = (1.04 x TK) + 4.94	0.85	1507
May	INK = (1.10 x TK) + 5.27	0.81	1580
June	INK = (0.82 x TK) + 8.36	0.74	1522
July	INK = (0.83 x TK) + 6.65	0.79	1579
August	INK = (0.98 x TK) + 3.31	0.82	1612
September	INK = (1.09 x TK) + 1.49	0.86	1513
AVERAGE	INK = (0.99 x TK) + 2.97	0.82	1528

Table C.8. Minimum daily temperature relationships between Trail Valley Creek (TVC) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (0.93 x TVC) - 0.53	0.86	341
November	INK = (0.96 x TVC) - 0.67	0.78	329
December	INK = (0.94 x TVC) - 1.71	0.85	338
January	INK = (0.84 x TVC) - 5.03	0.53	338
February	INK = (0.96 x TVC) - 1.30	0.83	264
March	INK = (0.99 x TVC) - 0.28	0.82	315
April	INK = (0.99 x TVC) + 1.65	0.88	257
May	INK = (0.84 x TVC) + 0.82	0.86	260
June	INK = (0.98 x TVC) + 2.70	0.73	229
July	INK = (0.97 x TVC) + 1.81	0.83	313
August	INK = (0.97 x TVC) + 1.18	0.85	340
September	INK = (0.94 x TVC) + 0.63	0.84	360
AVERAGE	INK = (0.94 x TVC) - 0.06	0.81	307

Table C.9. Minimum daily temperature relationships between Aklavik (AK) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (0.93 x AK) - 1.43	0.83	817
November	INK = (0.93 x AK) - 2.05	0.74	831
December	INK = (0.91 x AK) - 2.02	0.7	739
January	INK = (0.93 x AK) - 1.79	0.78	834
February	INK = (0.93 x AK) - 2.03	0.75	759
March	INK = (0.92 x AK) - 2.35	0.73	879
April	INK = (0.93 x AK) - 0.73	0.81	801
May	INK = (0.94 x AK) + 0.40	0.85	888
June	INK = (0.89 x AK) + 0.44	0.75	752
July	INK = (0.94 x AK) + 0.01	0.72	646
August	INK = (0.96 x AK) - 0.56	0.77	752
September	INK = (1.07 x AK) - 1.02	0.79	816
AVERAGE	INK = (0.94 x AK) - 1.09	0.77	793

Table C.10. Minimum daily temperature relationships between Tuktoyaktuk (TK) and Inuvik (INK).

Month	Equation	R^2	N
October	INK = (0.98 x TK) - 1.16	0.79	1546
November	INK = (1.04 x TK) - 0.56	0.75	1489
December	INK = (1.05 x TK) + 0.18	0.73	1550
January	INK = (1.11 x TK) + 2.18	0.78	1542
February	INK = (1.09 x TK) + 3.13	0.80	1407
March	INK = (1.05 x TK) + 2.76	0.73	1549
April	INK = (1.03 x TK) + 2.95	0.78	1507
May	INK = (0.88 x TK) + 2.54	0.74	1581
June	INK = (1.11 x TK) + 4.08	0.67	1526
July	INK = (0.84 x TK) + 3.30	0.62	1580
August	INK = (0.97 x TK) + 0.36	0.68	1581
September	INK = (0.67 x TK) + 0.62	0.72	1518
AVERAGE	INK = (0.99 x TK) + 1.70	0.73	1531

C.2 Precipitation

In 1999, Environment Canada published their First Generation Adjusted Precipitation dataset, also housed in the Adjusted Historical Canadian Climate Data archive (Mekis and Hogg, 1999). Long-term precipitation datasets, collected from weather stations across Canada, were corrected for gauge-specific under catch due to wind, evaporation, gauge-specific wetting losses, and trace precipitation. Adjusted total monthly precipitation was available for Inuvik, NT, for October 1957 to August 2003 (Inuvik A; Climate ID: 2202570) and September 2003 to November 2008 (Inuvik Climate; Climate ID: 2202578). Supplemental, unadjusted precipitation data was obtained for the years 1958 to 2006 (Inuvik A; Climate ID: 2202570) and 2003 to 2009 (Inuvik Climate; Climate ID: 2202578), from Environment Canada's National Climate Data Archive (Climate ID: 2202578).

Missing data accounted for approximately 11% of the adjusted total monthly precipitation data available for Inuvik. Missing months were infilled using unadjusted total monthly precipitation +/- a monthly correction (**Table C.11**). For each climate station (Inuvik A and Inuvik Climate), the adjusted total monthly precipitation data was pooled, based on month, and the unadjusted total monthly precipitation was pooled, based on month. The average unadjusted total monthly precipitation was subtracted from the average adjusted total monthly precipitation, and the resulting value was the monthly correction applied to the unadjusted total monthly precipitation values in infilling.

The Inuvik Climate station was the primary weather station. If total daily precipitation data was not available for the Inuvik Climate station, total daily precipitation data from the Inuvik A station was used. After the two weather stations were combined, missing data accounted for approximately 1% of the unadjusted total daily precipitation dataset. Unadjusted total daily precipitation was infilled using an anomaly-based approach, based on the methodology outlined by Goulding (2009). The average total daily precipitation, over the entire historical study period (e.g., 1958 to 2009), was determined for Inuvik, Trail Valley Creek, Aklavik, and Tuktoyaktuk. When a missing value was encountered, the departure of that day from the average daily value for one of three nearest weather stations was applied to the average daily value for

Inuvik. If data was available, Trail Valley Creek was always used first, Aklavik was used second, and Tuktoyaktuk was used third.

Table C.11. The monthly correction applied to unadjusted total monthly precipitation values in data infilling. The monthly correction is the average *adjusted* total monthly precipitation minus the average *unadjusted* total monthly precipitation.

Month	Average Total Monthly Precipitation (mm)					
	Inuvik A (October 1957 to August 2003)			Inuvik Climate (September 2003 to September 2007)		
	Adjusted	Unadjusted	Correction	Adjusted	Unadjusted	Correction
October	38.4	29.7	8.9	32.7	24.7	8.0
November	23.9	17.3	7.0	28.1	23.8	4.3
December	21.9	15.9	6.1	29.5	25.1	4.4
January	19.5	15.2	4.2	25.0	20.1	4.9
February	15.7	11.8	3.9	24.2	17.0	7.3
March	16.3	12.3	4.0	17.0	11.0	6.0
April	16.3	12.3	4.0	18.7	15.3	3.3
May	21.4	16.6	4.7	17.9	16.0	1.9
June	22.6	21.1	1.5	26.1	23.9	2.2
July	36.8	33.5	3.3	30.7	27.4	3.3
August	44.1	42.5	1.6	36.1	32.6	3.5
September	30.0	28.4	3.7	30.3	26.9	3.5

Appendix D: Tuktoyaktuk Climate Data

D.1 Air Temperature

Adjusted daily average, minimum, and maximum air temperature data was not available for Tuktoyaktuk, NT. Notably, in 2012, Environment Canada did release a Second Generation Homogenized Temperature dataset, which provides adjusted monthly average, minimum, and maximum temperature data for Tuktoyaktuk (Vincent et al., 2012). Unadjusted daily average, maximum, and minimum air temperature data was obtained from Environment Canada's National Climate Data Archive for the years 1958 to 1994 (Tuktoyaktuk; Climate ID: 2203910). Supplemental, unadjusted data was obtained for the years 1970 to 2007 (Tuktoyaktuk A; Climate ID: 2203912) and 1995 to 2009 (Tuktoyaktuk; Climate ID: 2203914) from Environment Canada's National Climate Archive.

Tuktoyaktuk was the primary weather station (Climate ID: 2203910). If daily air temperature data was not available for the Tuktoyaktuk station, daily air temperature data from the Tuktoyaktuk A station (Climate ID: 2203914) was used. If daily air temperature data was not available for the Tuktoyaktuk A station, daily air temperature data from the Tuktoyaktuk station (Climate ID: 2203912) was used. After the three Tuktoyaktuk-based stations were combined, missing data accounted for approximately 3%, 3%, and 3% of the average, maximum, and minimum daily air temperature datasets. Missing data was infilled using a between-station regression-based approach, which has been found to be more accurate than within-station methods (Kemp et al, 1983; Degaetano et al, 1995). The average daily air temperature for Tuktoyaktuk was pooled based on month. On a monthly basis, linear regression models were created between the average daily air temperature at Tuktoyaktuk and each of the three nearest weather stations (**Tables D.1 to Table D.4**). For each missing day, average daily air temperature from the closest weather station was always used first for infilling. The same methodology was used to infill the minimum daily and maximum daily air temperature datasets (**Tables D.5** to **D.10**).

Table D.1. Information for climate stations used in data infilling.

Weather Station	Climate ID	Latitude	Longitude	Distance from Inuvik Climate (Km)
Tuktoyaktuk	2203910	69° 27’ 00.000” N	133° 00’ 00.000” W	0
Tuktoyaktuk	2203912	69° 26’ 00.000” N	133° 01’ 00.000” W	2.55
Tuktoyaktuk A	2203914	69° 26’ 00.000” N	133° 01’ 35.000” W	3.33
Trail Valley Creek	220N005	68° 44’ 46.800” N	133° 30’ 06.400” W	94.34
Inuvik A	2202570	68° 18’ 15.000” N	133° 28’ 58.000” W	137.35
Inuvik Climate	2202578	68° 19’ 00.000” N	133° 31’ 00.000” W	137.39
Aklavik	2200100	68° 13’ 24.000” N	135° 00’ 21.000” W	158.91

Table D.2. Average daily temperature relationships between Trail Valley Creek (TVC) and Tuktoyaktuk (TK).

Month	Linear Regression Equation	R^2	N
October	TK = (0.92 x TVC) – 0.25	0.88	340
November	TK = (0.89 x TVC) – 1.65	0.80	329
December	TK = (0.90 x TVC) – 2.03	0.90	338
January	TK = (0.92 x TVC) – 3.18	0.90	337
February	TK = (0.91 x TVC) – 4.48	0.88	264
March	TK = (0.84 x TVC) – 6.14	0.79	315
April	TK = (0.91 x TVC) – 3.55	0.90	257
May	TK = (0.91 x TVC) – 2.03	0.91	260
June	TK = (0.85 x TVC) – 1.00	0.77	229
July	TK = (1.03 x TVC) – 1.70	0.91	313
August	TK = (0.88 x TVC) + 0.50	0.88	340
September	TK = (0.86 x TVC) + 0.53	0.91	360
AVERAGE	TK = (0.90 x TVC) – 2.08	0.87	307

Table D.3. Average daily temperature relationships between Inuvik (INK) and Tuktoyaktuk (TK).

Month	Linear Regression Equation	R^2	N
October	TK = (0.86 x INK) – 0.83	0.88	1540
November	TK = (0.81 x INK) – 3.04	0.84	1483
December	TK = (0.79 x INK) – 4.66	0.83	1543
January	TK = (0.79 x INK) – 5.58	0.86	1507
February	TK = (0.82 x INK) – 5.80	0.88	1367
March	TK = (0.80 x INK) – 7.08	0.84	1544
April	TK = (0.82 x INK) – 5.68	0.87	1506
May	TK = (0.81 x INK) – 4.34	0.83	1580
June	TK = (0.80 x INK) – 3.15	0.78	1522
July	TK = (0.93 x INK) – 2.13	0.80	1578
August	TK = (0.84 x INK) + 0.09	0.84	1581
September	TK = (0.79 x INK) + 0.11	0.87	1512
AVERAGE	TK = (0.82 x INK) - 3.51	0.84	1522

Table D.4. Average daily temperature relationships between Aklavik (AK) and Tuktoyaktuk (TK).

Month	Linear Regression Equation	R^2	N
October	TK = (0.88 x AK) – 0.84	0.87	839
November	TK = (0.70 x AK) – 5.69	0.68	802
December	TK = (0.78 x AK) – 4.65	0.72	693
January	TK = (0.81 x AK) – 4.67	0.81	741
February	TK = (0.85 x AK) – 4.91	0.79	669
March	TK = (0.84 x AK) – 6.13	0.79	827
April	TK = (0.87 x AK) – 4.94	0.83	776
May	TK = (0.86 x AK) – 3.96	0.81	852
June	TK = (0.79 x AK) – 2.30	0.73	721
July	TK = (0.95 x AK) – 2.18	0.77	621
August	TK = (0.85 x AK) – 0.14	0.78	740
September	TK = (0.86 x AK) – 0.34	0.85	809
AVERAGE	TK = (0.84 x AK) – 3.40	0.79	758

Table D.5. Maximum daily temperature relationships between Trail Valley Creek (TVC) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.84 x TVC) – 0.75	0.84	338
November	TK = (0.87 x TVC) – 1.66	0.74	329
December	TK = (0.93 x TVC) – 1.26	0.87	339
January	TK = (0.90 x TVC) – 3.21	0.87	337
February	TK = (0.88 x TVC) – 4.55	0.80	263
March	TK = (0.78 x TVC) – 7.16	0.66	316
April	TK = (0.83 x TVC) – 4.37	0.85	258
May	TK = (0.85 x TVC) – 2.44	0.86	260
June	TK = (0.92 x TVC) – 2.43	0.77	230
July	TK = (1.01 x TVC) – 2.71	0.87	313
August	TK = (0.91 x TVC) – 0.05	0.88	340
September	TK = (0.85 x TVC) – 0.04	0.90	360
AVERAGE	TK = (0.88 x TVC) – 2.55	0.83	307

Table D.6. Maximum daily temperature relationships between Inuvik (INK) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.83 x INK) – 1.14	0.86	1545
November	TK = (0.85 x INK) – 2.28	0.81	1489
December	TK = (0.84 x INK) – 3.48	0.82	1566
January	TK = (0.82 x INK) – 4.68	0.85	1509
February	TK = (0.86 x INK) – 4.92	0.85	1366
March	TK = (0.82 x INK) – 6.94	0.81	1545
April	TK = (0.82 x INK) – 5.77	0.85	1507
May	TK = (0.74 x INK) – 4.04	0.81	1580
June	TK = (0.90 x INK) – 4.65	0.73	1522
July	TK = (0.95 x INK) – 3.08	0.79	1579
August	TK = (0.84 x INK) – 0.51	0.82	1581
September	TK = (0.79 x INK) – 0.37	0.86	1513
AVERAGE	TK = (0.84 x INK) – 3.49	0.82	1525

Table D.7. Maximum daily temperature relationships between Aklavik (AK) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.81 x AK) – 0.97	0.84	837
November	TK = (0.81 x AK) – 2.58	0.72	810
December	TK = (0.78 x AK) – 4.12	0.72	725
January	TK = (0.75 x AK) – 5.73	0.76	745
February	TK = (0.86 x AK) – 4.69	0.78	678
March	TK = (0.79 x AK) – 7.70	0.71	835
April	TK = (0.80 x AK) – 5.96	0.77	792
May	TK = (0.75 x AK) – 3.93	0.72	844
June	TK = (0.86 x AK) – 2.68	0.68	745
July	TK = (0.94 x AK) – 2.11	0.74	675
August	TK = (0.82 x AK) + 0.28	0.72	763
September	TK = (0.81 x AK) – 0.17	0.81	819
AVERAGE	TK = (0.82 x AK) – 3.36	0.75	772

Table D.8. Minimum daily temperature relationships between Trail Valley Creek (TVC) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.94 x TVC) + 0.09	0.85	340
November	TK = (0.88 x TVC) – 2.48	0.74	329
December	TK = (0.87 x TVC) – 3.73	0.84	338
January	TK = (0.90 x TVC) – 4.22	0.84	338
February	TK = (0.88 x TVC) – 5.59	0.86	264
March	TK = (0.81 x TVC) – 7.57	0.79	315
April	TK = (0.94 x TVC) – 2.91	0.87	257
May	TK = (0.93 x TVC) – 1.35	0.90	260
June	TK = (0.68 x TVC) – 0.18	0.69	229
July	TK = (0.90 x TVC) + 0.59	0.78	313
August	TK = (0.75 x TVC) + 1.76	0.76	340
September	TK = (0.78 x TVC) + 1.13	0.80	360
AVERAGE	TK = (0.86 x TVC) – 2.04	0.81	307

Table D.9. Minimum daily temperature relationships between Inuvik (INK) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.81 x INK) – 1.22	0.79	1546
November	TK = (0.72 x INK) – 5.32	0.75	1489
December	TK = (0.70 x INK) – 7.64	0.73	1550
January	TK = (0.71 x INK) – 8.22	0.78	1542
February	TK = (0.74 x INK) – 8.48	0.80	1407
March	TK = (0.70 x INK) – 9.88	0.73	1549
April	TK = (0.76 x INK) – 6.76	0.78	1507
May	TK = (0.84 x INK) – 4.20	0.74	1581
June	TK = (0.60 x INK) – 2.05	0.67	1526
July	TK = (0.74 x INK) + 0.02	0.62	1580
August	TK = (0.70 x INK) + 1.50	0.68	1581
September	TK = (0.67 x INK) + 0.62	0.72	1518
AVERAGE	TK = (0.72 x INK) – 4.30	0.73	1531

Table D.10. Minimum daily temperature relationships between Aklavik (AK) and Tuktoyaktuk (TK).

Month	Equation	R^2	N
October	TK = (0.85 x AK) – 1.35	0.79	816
November	TK = (0.72 x AK) – 5.67	0.65	782
December	TK = (0.73 x AK) – 6.57	0.65	692
January	TK = (0.74 x AK) – 7.21	0.70	784
February	TK = (0.75 x AK) – 8.21	0.68	703
March	TK = (0.72 x AK) – 9.17	0.66	816
April	TK = (0.81 x AK) – 5.54	0.76	761
May	TK = (0.87 x AK) – 3.45	0.76	858
June	TK = (0.60 x AK) – 1.95	0.61	725
July	TK = (0.80 x AK) – 0.75	0.60	627
August	TK = (0.77 x AK) + 0.45	0.68	731
September	TK = (0.82 x AK) – 0.10	0.75	797
AVERAGE	TK = (0.77 x AK) – 4.13	0.69	758

D.2 Precipitation

In 1999, Environment Canada published their First Generation Adjusted Precipitation dataset, also housed in the Adjusted Historical Canadian Climate Data archive (Mekis and Hogg, 1999). Long-term precipitation datasets, collected from weather stations across Canada, were corrected for gauge-specific under catch due to wind, evaporation, gauge-specific wetting losses, and trace precipitation. Adjusted *total monthly* precipitation data was available for Tuktoyaktuk for the years 1958 to 1994 (Climate ID: 2203910), through Environment Canada's Adjusted Historical Climate Data archive. Supplemental, unadjusted *daily* precipitation data was obtained for the years 1958 to 1994 (Climate ID: 2203910), 1970 to 2007 (Climate ID: 2203912), and 1995 to 2009 (Climate ID: 2203914) from Environment Canada's National Climate Data Archive.

Missing data accounted for approximately 31% of the adjusted total monthly precipitation data available for Tuktoyaktuk. Missing months was infilled using unadjusted total monthly precipitation +/- a monthly correction (**Table D.11**). Adjusted total monthly precipitation data was pooled, based on month, and the unadjusted total monthly precipitation was pooled, based on month. The average unadjusted total monthly precipitation was subtracted from the average adjusted total monthly precipitation. The resulting value was the monthly correction applied to the unadjusted total monthly precipitation values in infilling.

The Tuktoyaktuk station (Climate ID: 2203910) was the primary weather station. If total daily precipitation data was not available for the Tuktoyaktuk station, total daily precipitation data from the other two Tuktoyaktuk stations was used. After the three weather stations were combined, missing adjusted precipitation data accounted for approximately 13% of the historical study period (e.g., 1958 to 2009). Missing data was infilled using an anomaly-based approach, which was based on the methodology outlined by Goulding (2009). The average daily total precipitation over the entire historical study period (e.g., 1958 to 2009) was determined for Tuktoyaktuk, Trail Valley Creek, Inuvik, and Aklavik. When a missing value was encountered, the departure of that day from the average daily value for the closest weather station with available data was applied to the average daily value for Tuktoyaktuk.

Table D.11. The monthly applied to unadjusted total monthly precipitation values in data infilling. The correction factor is the average *adjusted* total monthly precipitation minus the average *unadjusted* total monthly precipitation.

Month	Average Total Monthly Precipitation (mm) Tuktoyaktuk (Climate ID: 2203910) (October 1957 to June 1993)		
	Adjusted	Unadjusted	Correction
October	19.8	17.5	2.3
November	9.4	8.3	1.1
December	8.5	7.6	0.9
January	6.5	5.8	0.8
February	6.4	5.7	0.7
March	4.6	4.1	0.5
April	7.3	6.5	0.8
May	6.4	5.7	0.7
June	11.6	10.5	1.1
July	20.4	18.6	1.7
August	30.1	27.6	2.4
September	18.8	17.0	1.8

Appendix E: Validating Hydroclimatic Indices

A number of indices were used to evaluate the effects of historical climate variability and change on key hydroclimatic controls of the small tundra lake water balance. In this section, key hydroclimatic indices are validated using field data.

E.1 Air Temperature

Historical air temperature data collected at the Inuvik Climate station was used to develop a suite of key hydroclimatic indices (mean annual air temperature, spring freshet and ice-off timing, ice-on timing, and evaporation). The Inuvik Climate station was chosen because the average daily air temperature at Inuvik is the most representative of the average daily air temperature at Lake 5A. A linear regression analysis was used to model the relationship between the average daily air temperature at Lake 5A and the average daily air temperature at the three closest weather stations (Trail Valley Creek, Inuvik, and Tuktoyaktuk), over the four years with available data (2006, 2007, 2008, and 2009). The results presented in **Table E.1** indicate that, on average, 96% of the variation in average daily air temperature at Lake 5A can be explained by the average daily air temperature at Inuvik ($P < 0.05$). Notably, the average daily air temperature in Inuvik explains more of the variation in average daily air temperature at Lake 5A than Trail Valley Creek and Tuktoyaktuk. On average, 95% and 88% of the variation in average daily air temperature at Lake 5A can be explained by the average daily air temperature at Trail Valley Creek and Tuktoyaktuk, respectively ($P < 0.05$ and $P < 0.05$, respectively). This suggests that the ambient air temperature in Inuvik is a good indicator of how historical variability/change has affected the ambient air temperature of the two primary study lakes.

Table E.1. Linear regression models created for Lake 5A/5B and the three closest Environment Canada weather stations (Trail Valley Creek (TVC), Inuvik, and Tuktoyaktuk, NT) on a monthly basis.

Month	TVC			Inuvik			Tuktoyaktuk		
	N	R^2	P	N	R^2	P	N	R^2	P
January	70	0.97	0.000	71	0.98	0.000	71	0.90	0.000
February	34	0.96	0.000	47	0.97	0.000	47	0.92	0.000
March	31	0.98	0.000	31	0.94	0.000	31	0.89	0.000
April	30	0.98	0.000	30	0.97	0.000	30	0.93	0.000
May	121	0.97	0.000	121	0.96	0.000	121	0.89	0.000
June	119	0.92	0.000	119	0.90	0.000	119	0.84	0.000
July	124	0.93	0.000	124	0.97	0.000	124	0.87	0.000
August	123	0.95	0.000	124	0.98	0.000	124	0.85	0.000
September	112	0.95	0.000	112	0.96	0.000	112	0.88	0.000
October	92	0.95	0.000	93	0.96	0.000	93	0.93	0.000
November	89	0.91	0.000	90	0.94	0.000	90	0.81	0.000
December	90	0.92	0.000	93	0.94	0.000	93	0.83	0.000
AVERAGE	86	0.95	0.000	88	0.96	0.000	88	0.88	0.000

E.2 Precipitation

Historical adjusted total monthly precipitation data, collected at the Inuvik Climate station, was used to develop a suite of key hydroclimatic indices (total annual precipitation, annual snowpack index, annual rainfall index, and vertical water balance).

Annual Snowpack Index

The average weighted SWE on the contributing catchment at Lake 5A and Lake 5B was compared to the annual snowpack index for Inuvik (**Table E.2**). The mean average SWE of the contributing catchment at Lake 5A and Lake 5B does not appear to be positively or negatively correlated with the annual snowpack index for the Inuvik Airport. In 2007 and 2008, the two primary study lakes received 35.9mm to 48.3mm *more* snowfall than Inuvik, respectively. Conversely, in 2009, the two primary study lakes received 10.6mm *less* snowfall than Inuvik. This was expected, however, because there was a pre-freshet melting period in April 2009 that melted a substantial portion of the contributing snowpack at Lake 5A and Lake 5B. Overall, the annual snowpack index underestimates the SWE of the contributing catchment at Lake 5A and Lake 5B. Over the three study years, the Absolute Difference associated with using the annual snowpack Index (mm) for Inuvik, NT, to estimate the average SWE for the Lake Surface of Lake 5A and Lake 5B was 31.6mm.

Table E.2. The average SWE for the contributing catchment at Lake 5A and Lake 5B compared to the Annual Snowpack Index for Inuvik for the years with available data (2005, 2006, 2007, and 2008).

Year	Lake 5A Catchment SWE (mm)	Lake 5B Catchment SWE (mm)	Mean SWE (mm)	Inuvik Annual Snowpack Index (mm)	Absolute Difference (mm)
2007	136		136.0	100.1	35.9
2008	173	148	160.5	112.2	48.3
2009	108	150	129.0	139.6	10.6
Average	**139**	**149**	**141.8**	**117.3**	**31.6**

An Independent-Samples T-Test was used to test whether or not the average SWE for the catchment contributing to Lake 5A and Lake 5B was significantly different than the annual snowpack index for Inuvik for the years with available data (2007, 2008, and 2009) (**Table E.3**). The average SWE for the contributing lake catchment *is not* significantly different from the annual snowpack index for Inuvik ($t = -0.126$, $df = 6$, $p = 0.904$). This suggests that the annual snowpack index for Inuvik is a good indicator of how historical climate variability and change has affected the contributing snowpack at the two primary study lake catchments.

Table E.3. An Independent-Samples T-Test used to test whether or not the SWE of the contributing catchment at Lake 5A is significantly different from the annual snowpack index at Inuvik, for the three years with available data (2007, 2008, and 2009).

t	df	p
1.626	4	0.179

Annual Rainfall Index

The annual rainfall index for Lake 5A and Lake 5B was compared to the annual rainfall index for Inuvik for the years with available data (2006, 2007, 2008, and 2009) (**Table E.4**). There does not appear to be a positive or negative correlation between the annual rainfall index at Lake 5A and Lake 5B and that of Inuvik. In 2006, 2007, and 2008, the two primary study lakes received 30%, 43%, and 35% *less* rainfall than Inuvik, respectively. Conversely, in 2009, the two primary study lakes received 1% *less* rainfall than Inuvik, NT, respectively. On average, the absolute difference between the annual rainfall index for Inuvik to the annual rainfall index at the two primary study lakes was 27%.

An Independent Samples T-Test was used to test whether or not the annual rainfall index for Lake 5A and Lake 5B was significantly different than the annual rainfall index for Inuvik, for the four years with available data. The annual rainfall index for Lake 5A and Lake 5B was not significantly different than the annual rainfall index for Inuvik (**Table D.5**). This suggests that the annual rainfall index estimated for Inuvik, NT, is a good indicator of how historical climate variability and change has affected the summer rainfall received by the two primary study lakes.

Table E.4. The annual rainfall index at Lake 5A and Lake 5B was compared to the annual rainfall index at Inuvik, for the four years with available data (2006, 2007, 2008, and 2009).

Year	Time Period	Annual Rainfall Index Lake 5A and Lake 5B (mm)	Annual Rainfall Index Inuvik (mm)	Absolute difference (%)
2006	July 1st to August 31st	59.8	78.0	30
2007	July 1st to August 31st	42.4	60.4	43
2008	July 1st to September 30th	125.6	169.9	35
2009	June 1st to August 31st	115.1	113.8	1
Average		85.7	105.5	27

Table E.5. An Independent Samples T-Test used to test whether or not the annual rainfall index for Lake 5A/5B was significantly different than the annual rainfall index for Inuvik, for the four years with available data (2006, 2007, 2008, and 2009).

t	df	p
-0.626	6	0.554

E.3 Spring Freshet Timing

The actual date of the spring freshet at Lake 5A and Lake 5B, determined using the method outlined by Burn et al. (2008) using stream discharge data collected at nearby Trail Valley Creek, was used to validate two temperature-based methods of estimating the timing of the spring freshet, determined using the methods outlined by Pohl (Personal Communication, 2011) and Bonsal et al. (2003) using air temperature data collected at Lake 5A and Lake 5B (**Table E.6**). On average, the two temperature-based methods estimated spring freshet initiation within 5 days of the actual date.

Table E.6. The actual date of the spring freshet at Lake 5A and Lake 5B, determined using stream discharge data collected at nearby Trail Valley Creek, was used to validate two temperature-based methods of estimating the timing of the spring freshet, determined using air temperature data collected at Lake 5A and Lake 5B, for the three years with available data (2007, 2008, and 2009).

	Spring Freshet Initiation: Lake 5A and Lake 5B			**Absolute Difference (Days)**	
	Burn et al. (2008)	**Pohl (2011)**	**Bonsal et al. (2003)**	**Pohl (2011)**	**Bonsal et al. (2003)**
2007	May 26	May 23	May 29	3	3
2008	May 28	May 18	May 22	10	6
2009	May 18	May 20	May 24	2	6
Average	**May 24**	**May 20**	**May 25**	**5**	**5**

An Independent Samples T-Test was used to test whether or not the actual date of the spring freshet at Lake 5A and Lake 5B was significantly different than the dates estimated by the two temperature-based methods (**Table E.7**). The actual date of the spring freshet was not significantly different than the dates estimated using the methods outlined by Pohl (personal communication, 2011) and Bonsal et al. (2003). Both methods are good estimates of the timing of the spring freshet atthe two primary study lakes. The method outlined by Pohl (Personal Communication, 2011) was chosen.

Table E.7. An Independent Samples T-Test was used to test whether or not the Actual date of the spring freshet at Lake 5A and Lake 5B, determined using stream discharge data, was significantly different than the dates estimated by the two temperature-based methods of estimating the timing of the spring freshet, determined using air temperature data collected at Lake 5A and Lake 5B, for the three years with available data (2007, 2008, and 2009).

Method	**t**	**df**	**p**
1	1.049	4	0.139
2	-0.265	4	0.804

The timing of the spring freshet at Lake 5A and Lake 5B, was compared to the timing of the spring freshet at Inuvik (**Table E.8**). The timing of the spring freshet at Lake 5A and Lake 5B was estimated using the method outlined by Burn, 2008. The timing of the spring freshet at Inuvik was estimated using the method outlined by Pohl (Personal Communication, 2011) using air temperature data collected at the Inuvik Climate station. On average, there is a 7-day difference in the timing of the spring freshet in Inuvik and the timing of the spring freshet at Lake 5A and Lake 5B.

An Independent-Samples T-Test was performed, in order to test whether or not there was a significant difference in the timing of the spring freshet at Lake 5A and Lake 5B and the timing of the spring freshet at Inuvik (**Table E.9**). The Independent-Samples T-Test indicates that there was no significant difference in the timing of the spring freshet at Lake 5A and Lake 5B and the timing of the spring freshet at Inuvik. This suggests that the timing of the spring freshet in Inuvik is a good indicator of how historical variability/change has affected the timing of the spring freshet at the two primary study lakes.

Table E.8. The date of the spring freshet at Lake 5A/Lake 5B, estimated using stream discharge data, was compared to the date of the spring freshet at Inuvik, estimated using the method outlined by Pohl (Personal Communication, 2011), for the years with available data (1977 to 2009).

	Spring Freshet:		
Year	**Lake 5A and Lake 5B**	**Inuvik**	**Absolute Error (Days)**
1977	148	148	0
1978		149	
1979	141	139	2
1980		144	
1981	141	137	4
1982	137	143	6
1983	149	145	4
1984	138	135	3
1985	150	148	2
1986	150	149	1
1987	152	145	7
1988	147	142	5
1989	145	144	1
1990	132	141	9
1991	147	146	1
1992	143	160	17
1993	139	143	4
1994	124	150	26
1995	147	144	3
1996	138	157	19
1997	130	151	21
1998	141	129	12
1999	159	141	18
2000	156	156	0
2001	150	149	1
2002	148	148	0
2003	154	146	8
2004	145	148	3
2005		137	
2006		136	
2007	146	142	4
2008	149	142	7
2009	138	137	1
Average	**144**	**145**	**7**

Table E.9. An Independent-Samples T-Test was used to test whether or not there was a significant difference in the timing of the spring freshet at Lake 5A and 5B and the timing of the spring freshet at Inuvik, for the years with available data (1977 to 2009).

t	df	p
-0.165	60	0.870

E.4 Open-water Duration

The actual date of ice-off, determined using water temperature data obtained at Lake 5A and Lake 5B, was compared to the estimated date of ice-off, determined using Positive Degree Days at the Inuvik Climate station (**Table E.10**). On average, the estimated date of ice-off was within 3 days of the actual date.

Table E.10. The actual date ice-off at Lake 5A and Lake 5B, determined using field data, compared with the estimated date of ice-off at Inuvik, determined using PDD, for the four years with available data (2006, 2007, 2008, and 2009).

	Ice-Off (Actual)	PDD (°C)	Ice-Off (Estimated)	Absolute Error (Days)	Source
2006	June 15	244	June 14	1	Pohl et al. (2009)
2007	June 8	173	June 15	7	
2008	June 13	223	June 14	1	
2009	June 17	252	June 15	2	
Average	**June 14**	**223**	**June 15**	**3**	

An Independent Samples T-Test was used to test whether or not there was a significant difference between the actual date of ice-off, estimated using water temperature data, and the estimated date of ice-off, estimated using PDD. The estimated date of ice-off was not significantly different than the actual date of ice-off (**Table E.11**). This suggests that the timing of ice-off at Inuvik, NT, is a good indicator of how historical climate variability and change has affected the timing of ice-off at the two primary study lakes.

Table E.11. An Independent Samples T-Test was used to test whether or not there was a significant difference between the actual date of ice-off for Lake 5A and Lake 5B and the estimated date of ice-off for Inuvik, for the four years with available data (2006, 2007, 2008, and 2009).

t	**df**	**p**
-0.640	6	0.546

The date of ice-on at Lake 5A and Lake 5B was compared to the date of ice-on at Inuvik (**Table E.12**). In 2006 and 2008, ice-on in Inuvik, NT, occurred on the same day as ice-on at Lake 5A and Lake 5B. In 2007, ice-on occurred one day earlier than ice-on in Inuvik. On average, the absolute error associated with using air temperature data from Inuvik to estimate the timing of ice-off at the two primary study lakes is 1 day.

Table E.12. The date of ice-on at Lake 5A and Lake 5B compared with the date of ice on at Inuvik, for the four years with available data (2006, 2007, 2008, and 2009).

Year	**Ice-On**		**Absolute Error (Days)**
	Lake 5A and Lake 5B	**Inuvik**	
2006	October 18	October 18	0
2007	October 3	October 4	1
2008	October 5	October 5	0
Average	**October 8**	**October 9**	**1**

An Independent-Samples T-Test was used to test whether or not there was a significant difference in the timing of ice-on at Lake 5A and Lake 5B and the timing of ice-on at Inuvik, NT. The timing of ice-on at Lake 5A and Lake 5B was not significantly different from the timing of ice-on at Inuvik (**Table E.13**). This suggests that the timing of ice-on at Inuvik is a good indicator of how historical variability/change has affected the timing of ice-on at the two primary study lakes.

Table E.13. An Independent-Samples T-Test used to test whether or not there was a significant difference in the timing of ice-on at Lake 5A and Lake 5B and the timing of ice-on at Inuvik, NT, based on the three years with available data (2006, 2007, and 2008).

t	**df**	**p**
-0.053	4	0.961

E.5 Evaporation

Priestley-Taylor evaporation (E_{PT}), measured at Lake 5A and Lake 5B in 2007, 2008, and 2009 and at Trail Valley Creek in 2006, was compared to Hargreaves evaporation (E_{HG}), estimated using air temperature data collected at Inuvik. Overall, E_{HG} is a good estimate of E_{PT}. In 2006 and 2007, E_{HG} underestimated E_{PT} by 4mm and 1mm, respectively. In 2008 and 2009, E_{HG} over estimated E_{PT} by 1mm and 14mm. On average, the absolute difference between E_{HG} and E_{PT} was 5mm.

An Independent Samples T-Test was used to test whether or not there was a significant difference between E_{PT} and E_{HG}. E_{HG} was not significantly different from E_{PT} (**Table E.14**). This suggests that E_{HG}, estimated using historical temperature data collected in Inuvik, is a good indicator of how historical climate variability and change has affected E_{PT} from Lake 5A.

Table E.14. Hargreaves evaporation (E_{HG}), for Inuvik, was compared to Priestley Taylor evaporation (E_{PT}), for Lake 5A, using four years worth of available data (2006, 2007, 2008, and 2009).

Year	**Time Period**		**Location**	**E_{PT} (mm)**	**E_{HG} (mm)**	**Absolute Difference (%)**	**Source**
2006	15-Jun	31-Oct	Trail Valley Creek, NT	317	304	4	Pohl et al., 2009
2007	29-Jun	25-Aug	Lake 5A, NT	188	187	1	
2008	27-Jun	14-Sep	Lake 5A, NT	207	208	1	
2009	26-Jun	22-Sep	Lake 5A, NT	202	231	14	
Average				229	233	5	

Table E.15. An Independent Samples T-Test was used to test whether or not there was a significant difference between Priestley Taylor evaporation (E_{PT}), calculated for Lake 5A, and Hargreaves evaporation (E_{HG}), calculated for Inuvik, NT, based on four years worth of available data (2006, 2007, 2008, and 2009).

t	**df**	**p**
-0.102	6	0.922