

Stable-isotope relations ($\delta^{18}O$ & δD) of precipitation, the snowpack, and
permafrost, Mackenzie Delta area, NWT

by

Loucas Diamant-Boustead

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Loucas Diamant-Boustead

Abstract

Stable-isotope data from precipitation and the snowpack near Inuvik and from permafrost at Illisarvik, on the western Arctic coast, were used to assess the isotopic similarities between precipitation and permafrost in the western Canadian Arctic. The NOAA HYSPLIT model was used to determine the origin of 111 storms at Inuvik in 2015-18. The $\delta^{18}\text{O}$ values of these storms were consistent in winter (-25‰) and summer (-17.5‰) irrespective of origin, but northern storms came with lower air temperatures. The isotopic composition of the entire snowpack at the end of winter represented mean winter precipitation well. Samples of near-surface permafrost indicated a thaw unconformity due to the Holocene Thermal Maximum at 240-cm depth, with mixed Pleistocene and Holocene water above this level. The isotopic composition of ice wedges at Illisarvik was mostly higher in $\delta^{18}\text{O}$ than regional winter precipitation. These observations suggest wedge ice may not be entirely composed of winter precipitation.

Data Declaration

Various data sets are considered in this thesis. Some are unpublished and presented here, and others are already available. First, for analysis of synoptic conditions and their influence on $\delta^{18}\text{O}$ for precipitation, I have used data presented by Fritz et al. (2021, 2022). Fritz et al. (2022) used these data to calculate a co-isotope line for precipitation at Inuvik, but did not conduct any meteorological analyses. For consideration of $\delta^{18}\text{O}$ vs. Temperature at Inuvik, I have used data for December 1985 - February 1989 available on the Global Network of Isotopes in Precipitation (GNIP, <https://www.iaea.org/services/networks/gnip>) and for March 1989 - May 1995 which is yet unpublished. These GNIP data have been used by Bailey et al. (Fig. 2a, 2015) to characterize $\delta^{18}\text{O}$ for mean winter precipitation in the western Arctic and by Fritz et al. (2022) to create co-isotope plots for precipitation at Inuvik.

Related Research Output

This thesis is not an integrated thesis as defined by the Faculty of Graduate and Postdoctoral Affairs at Carleton University. However, portions of chapters 1 – 7 of the thesis are drawn from a refereed conference paper that was published at the time of thesis submission. The Co-Author Statement (p. 154) describes the contributions of each author towards that paper.

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1 Overview and Objectives

1.1 Introduction

This thesis examines the variation in stable-isotopic composition of water in a range of environments and physical features in the Mackenzie Delta area. The purpose is to assemble information on the ^{18}O concentration in precipitation, the snowpack, and permafrost in order to investigate the relations for this variable between them. The thesis first examines variation in the stable-isotopic composition of modern-day precipitation on a storm-by-storm time basis in the Mackenzie Delta area. Then it assesses the composition of winter precipitation and the snowpack. Finally, it investigates the stable-isotopic composition of near-surface permafrost at a site on Richards Island, in the outer delta area. The thesis explicitly considers the relations between stable water isotopes that are apparent in these environments.

The stable isotopes of precipitation have been extensively studied (e.g., Craig, 1961; Dansgaard, 1954, 1964), and a statistically significant linear relationship between the annual mean air temperature and the annual mean isotopic composition of precipitation is consistently reported (Dansgaard, 1964; Rozanski et al., 1993). However, short-term variations in precipitation isotopes at sub-seasonal time scales are influenced by synoptic-scale weather patterns and other environmental factors (Bailey et al., 2015; Akers et al., 2017), reducing the correlation of temperature and isotopes on a storm-by-storm basis. In ground ice, stable isotopes serve as valuable tools for interpreting origin and stratigraphy over millennial time scales (Burn et al., 1986; Lacelle and Vasil'chuk, 2013).

This thesis examines specific factors influencing the stable-isotopic composition of individual storm events in 2015-18 at Inuvik, NWT (Fritz et al., 2021) to enable

identification of specific influences on the isotopic composition of precipitation at the locality in addition to temperature. In addition, precipitation and late winter snow samples collected near Inuvik between 1985-95 and 1991-95, respectively, provide insight into the isotopic relations between winter precipitation and the snowpack. Finally, permafrost and hillslope ice-wedge samples from near the Illisarvik drained lake basin (Fig. 1.1), collected in August 2022, were analyzed for their stable-isotopic composition to further the knowledge of permafrost development in the region. This thesis does not address specifically the $\delta^{18}\text{O} : T$ relations in the Mackenzie Delta area.

1.2 Isotope Climatology

Naturally occurring waters contain stable isotopes of hydrogen (^1H and deuterium (^2H or D)) and oxygen (^{16}O , ^{17}O , and ^{18}O). There are three principal compositions for water: H_2^{16}O , HD^{16}O , and H_2^{18}O . Isotopic abundance is measured with reference to the composition of Vienna Standard Mean Ocean Water (VSMOW; Gonfiantini, 1978), which is the mean ratio of heavy and light isotopes found in the water throughout the world's oceans. The isotopic compositions are typically represented through δ - notation, where the quantity of heavy and light isotopes in a sample is normalized with the respective quantities in VSMOW. For $\delta^{18}\text{O}$ this is:

$$\delta^{18}\text{O} = \frac{\left(\frac{^{18}\text{O}}{^{16}\text{O}}\right)_{\text{sample}} - \left(\frac{^{18}\text{O}}{^{16}\text{O}}\right)_{\text{vsmow}}}{\left(\frac{^{18}\text{O}}{^{16}\text{O}}\right)_{\text{vsmow}}} \times 1000 \quad (1.1)$$

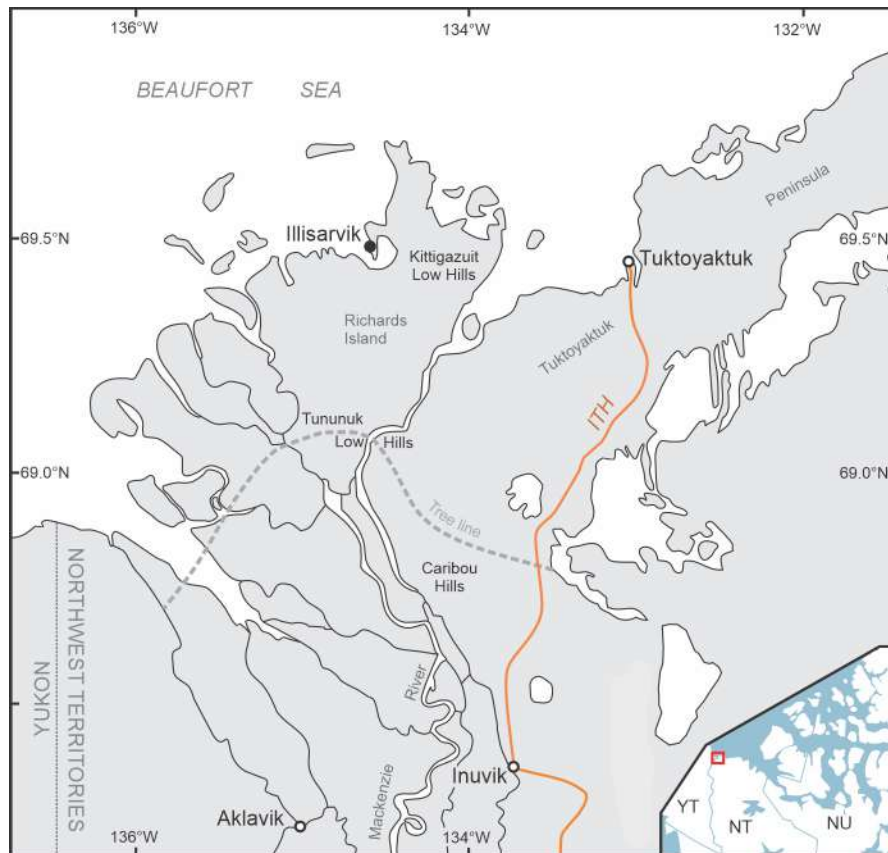


Figure 1.1: Mackenzie delta area, western Arctic coast, Canada. Illisarvik is on the west point of Richards Island. The ITH is the Inuvik-Tuktoyaktuk Highway. The location of the tree line was identified by Mackay (1963). Cartography by H. B. O'Neill.

where ^{18}O and ^{16}O represent the concentration for the isotope. δD is calculated similarly, but with $\text{D} / ^1\text{H}$. Air temperature during an event has been identified as an important environmental variable associated with the isotopic composition of precipitation (Jouzel and Merlivat, 1984; Jouzel et al., 1997), and a significant relation between these variables has been identified at an annual scale (Dansgaard, 1964). Porter et al. (eq. 6, 2016) and Kostrova et al. (Fig. 4a, 2020) reported $\delta^{18}\text{O} : T$ relations of $\delta^{18}\text{O} = 0.41T - 20.2$ from continental polar North America and $\delta^{18}\text{O} = 0.5T - 20.01$ from Siberia, where T is the mean annual air temperature and $\delta^{18}\text{O}$ is the mean annual ^{18}O concentration in precipitation. The initial relation between these variables was determined by Dansgaard (eq. 9, 1964) for Greenland and North Atlantic stations:

$$\delta^{18}\text{O} = 0.69T - 13.6 \quad (1.2)$$

Other factors such as the origin and trajectory of an air mass are also of significance to the isotopic composition of precipitation, and affect this relation at seasonal or meteorological time scales (Dansgaard, 1964; Bailey et al., 2015).

1.3 Ice Wedges

Ice wedges are V-shaped features of massive ice that form in permafrost just below the active layer (Leffingwell, 1915). They can be classified into three types based on surface characteristics where they form. Epigenetic ice wedges develop in flat topography that is neither aggrading nor degrading (Mackay, 1990; Liljedahl et al., 2016); syngenetic ice wedges develop in areas of colluvial or alluvial deposition such as the bottom of hillslopes or river deltas, and grow upwards with the aggradation of surface material and the top of permafrost (Mackay, 1990; Morse and Burn, 2013); anti-syngenetic ice wedges develop in areas of erosion and surface loss that is accompanied by lowering of the permafrost table (Mackay, 1990, 1995; Burn et al.,

2021). Each type has a distinct shape (Fig. 1.2). The large width of anti-syngenetic ice wedges at the top of permafrost facilitates the collection of samples representing a range of ages. Figure 1.2 shows that the oldest ice is found in relatively large quantities near the edges of the wedge.

Epigenetic ice wedges develop a distinct pair of ridges that run along the outer limits of the wedge. Between these is a trough or depression, which in well-established wedges will fill with water (Mackay, 1990; Burn, 2004). Since anti-syngenetic ice wedges develop at sites of erosion and slope movement, these ridges may not be apparent on hillslopes (Mackay, 1995; Burn et al., 2021). As anti-syngenetic ice wedges are near to the surface, and in dry ground are easily accessible from above, consistent sampling across the wedge is possible. In contrast, traditional sampling of ice wedges has been on exposed sections of permafrost (e.g., Opel et al., 2017; Campbell-Heaton et al., 2021).

Invoking the connection between precipitation isotopes and temperature, Meyer et al. (2015) and Opel et al. (2017) have suggested that ice-wedge ice indicates a general warming trend in Arctic regions during the latter portion of the Holocene, in contrast to the consensus of a cooling trend found with other temperature proxies up to the Industrial Revolution. The formation of ice wedges from snow melt provides a representation of winter precipitation (Mackay, 1974a), and when combined with dating techniques may provide a chronology of potential conditions during the formation of the wedge ice (e.g., Vasil'chuk et al., 2000, 2006, 2022). However, Michel (1990) noticed that the isotopic concentration in wedge ice does not match winter precipitation, as subsequently noted in other research (e.g., Lauriol et al., 1995; Holland et al., 2020; Campbell-Heaton et al., 2021) and discussed later in this thesis.

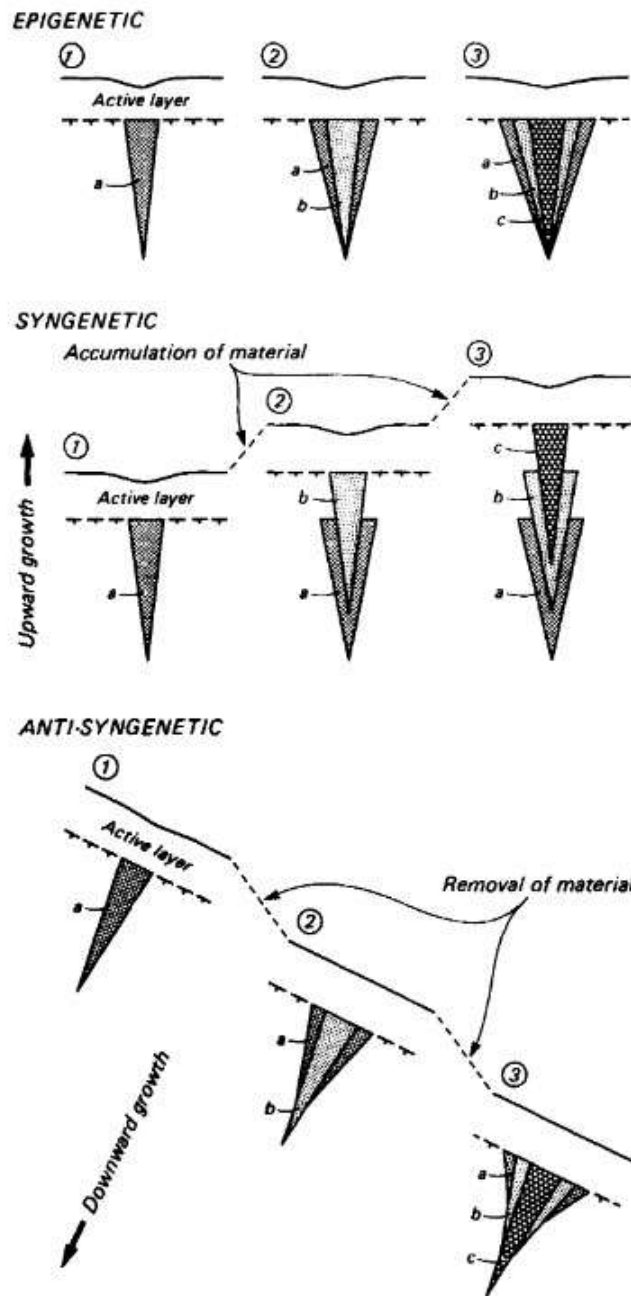


Figure 1.2: Diagram of the different types of ice wedge. 1, 2, 3 represent the stages of ice-wedge growth, and a, b, c represent successively younger ice within each wedge (Fig. 1, Mackay, 1990). Image reproduced with permission from John Wiley and Sons.

1.4 Research Objectives

The purposes of this thesis are: (1) to improve understanding of various influences on the isotopic composition of precipitation in the Mackenzie Delta area; and (2) to analyze the isotopic composition of hillslope ice wedges at Illisarvik to further the understanding of these features.

The specific objectives of this thesis are to investigate:

1. the stable-isotopic composition of precipitation at Inuvik, NWT;
2. the influence of the trajectory of an air mass carrying precipitation to Inuvik on its isotopic composition;
3. the $\delta^{18}\text{O} : T$ relation for precipitation at Inuvik on at seasonal and sub-seasonal time scales;
4. variation in the isotopic composition of the late winter snowpack with respect to winter precipitation;
5. the isotopic composition of near-surface permafrost at Illisarvik, NWT;
6. the isotopic variation within hillslope ice wedges at Illisarvik, NWT.

The research hypotheses are twofold: (1) that storm origin and air mass patterns will have a substantial effect on the isotopic composition of precipitation, in line with the findings of Bailey et al. (2015); and (2) the isotopic composition of the surface ice from hillslope ice wedges will adequately represent variation in winter conditions throughout the Holocene on Richards Island.

1.5 Research Methods

This thesis examines data gathered at or around Inuvik by various scientists. Isotopic concentrations in precipitation from individual storm events at Inuvik between

2015-18 were determined by Fritz et al. (2021). Here, the NOAA HYSPLIT back trajectory tool is used to determine air mass origin and path of these storm systems and to investigate the role of physiography and temperature on stable-isotope values in precipitation. C.R. Burn collected monthly aggregate precipitation samples from the Inuvik airport in 1985-95 and snowpack samples at the end of winter from various lake and forest environments near Inuvik in 1991-95. Both sets of samples were analysed for $\delta^{18}\text{O}$ composition and are used by this thesis to determine the variation in isotopic composition between late winter snowpack and winter precipitation at Inuvik. These snowpack data are not yet published.

Permafrost samples were collected in August 2022 at the Illisarvik drained-lake research site on Richards Island, NWT (Fig. 1.1). Two boreholes were drilled on the tundra to the west of the drained lake basin, with samples of permafrost collected to depths of 3 m. Stable-isotope profiles for six hillslope ice wedges were created from samples obtained by drilling through the trough and directly into the wedge. The profiles represent the surface of the wedge and were analysed in light of the context gained from the investigation of precipitation isotopes.

1.6 Thesis Structure

The thesis contains seven chapters. Chapter 2 reviews the literature surrounding isotope hydrology, and then proceeds to explore ice-wedge growth and paleoclimate reconstruction. A review of geologic and climate-related aspects of the western Canadian Arctic is presented in Chapter 3. Chapter 4 outlines the specific methodology used in the field and during analysis. Results of analysis and field work are presented in Chapter 5. These results are discussed in Chapter 6, which further addresses the primary objectives of this thesis. Final conclusions and suggestions for future research are made in Chapter 7.

2 Literature Review

In this chapter, background material is presented on (1) isotope climatology and fractionation, including the relations between $\delta^{18}O$ and δD as well as $\delta^{18}O$ and temperature, and (2) the state of permafrost, in order to interpret the stable-isotopic composition of precipitation and ice wedges of the western Canadian Arctic.

2.1 Isotope Hydrology

Isotopes represent atoms of the same species with varying amounts of neutrons in the nuclei. Isotopes occur in two categories: stable or unstable. Stable isotopes do not decay over time (Hoefs, 2015). This thesis considers the stable isotopes of hydrogen and oxygen in water.

Minute differences in chemical behaviour of an element due to variations in atomic mass are known as isotope effects (Hoefs, 2015). The energy of a molecule is restricted to certain discrete energy levels, the lowest being above the minimum of the potential energy curve of the molecule by $\frac{1}{2}h\nu$, where h is Planck's constant (6.63×10^{-34} Js) and ν is the frequency at which two atoms within a molecule vibrate with respect to each other (Hoefs, 2015). The vibration occurs at a fundamental frequency that is dependent on the mass of the constituent isotopes (Hoefs, 2015). Thus, different isotopes of varying species will have different zero-point energies: the heavier isotopes will have a lower zero-point energy than lighter isotopes due to their lower vibrational frequency. The bonds between molecules of lighter isotopes are weaker, and will react more quickly to phase change than those of the heavier isotopes (Hoefs, 2015). As a result, there are small changes in the physicochemical

properties of the molecules. Table 2.1 presents differences in physicochemical properties between H_2^{16}O and H_2^{18}O .

2.1.1 *Equilibrium Fractionation*

Isotope fractionation is the partitioning of molecules with differing isotopic composition between two phases (Hoefs, 2015). It can occur as an equilibrium or non-equilibrium (kinetic) process (Craig, 1961; Dansgaard, 1964). In equilibrium fractionation, the difference in isotopic composition between two phases follows from the difference in energy state between them. The result is a steady state concentration difference in the phases due to the physicochemical properties of the isotopes. Air mass formation above an ocean, for instance, occurs under equilibrium conditions since the ocean is considered as an infinite source with unchanging $\delta^{18}\text{O}$ and δD values (Dansgaard, 1964). VSMOW is the standard used to represent ocean water. The constant isotopic composition of surface water from turnover and the saturation and movement of the air mass supports the infinite source and equilibrium state of the vapour. Water will evaporate into the air mass until saturation is reached (100% humidity), at which point the vapour pressure will be greatest. H_2^{16}O exerts a higher vapour pressure than H_2^{18}O (e.g., Table 2.1). The higher energy of the H_2^{16}O molecule means that it evaporates preferentially to H_2^{18}O , and the vapour is depleted in H_2^{18}O relative to the source water.

When the ocean and air mass are in chemical equilibrium with each other, the isotopic composition of the two phases remains proportional to each other. Equilibrium fractionation (ϵ) above an ocean can then be calculated using:

$$\epsilon = (\alpha - 1) \cdot 1000 \quad (2.1)$$

Table 2.1: Physicochemical properties of the three abundant water species. Data from the CRC Handbook (Rumble, 2023); abundances from Dansgaard (1964).

Property	H ₂ ¹⁶ O	HD ¹⁶ O	H ₂ ¹⁸ O
Melting point @ 101.325 kPa (°C)	0	-0.56	-0.8
Boiling point @ 101.325 kPa (°C)	100	101.42	100.6
Vapour pressure @ 100°C (kPa)	101.325	101.156	101.278
Abundance (ppm)	997680	320	2000

where α is the coefficient of fractionation for each isotope. This value is governed by the isotopic exchange between the water vapour and corresponding atmosphere, and is dependent on temperature, which determines the saturation vapour pressure of the air mass (Urey, 1947; Dansgaard, 1953; Ross, 2017). α for HDO and H_2^{18}O at 25°C and under equilibrium conditions are approximately 1.08 and 1.009, respectively (Dansgaard, 1964). At 0°C, α is 1.0117 and 1.111 for D and ^{18}O , respectively (Majoube, 1970). This is due to the difference in vapour pressures between HDO and H_2^{18}O at different temperatures (Urey, 1947; Hoefs, 2015). At 25°C, water vapour should be depleted some 8‰ in δD and 0.9‰ in $\delta^{18}\text{O}$ relative to the original water source (Urey, 1947; Dansgaard, 1964). During equilibrium freezing at 0°C, α for D and ^{18}O is 1.018 and 1.003, respectively. The resulting ice will be enriched in δD by some 17‰ and in $\delta^{18}\text{O}$ by 3‰ (O'Neil, 1968). If freezing occurs very quickly, no fractionation may occur (O'Neil, 1968; Michel, 1990).

The Rayleigh Model

The concentration of ^{18}O increases relative to the concentration of ^{16}O as water moves from a higher energy state into a lower, and vice versa. As condensation occurs within an air mass, ^{18}O will preferentially condense onto the growing ice crystals. This results in the precipitation being relatively enriched (high $\delta^{18}\text{O}$ value) and the remaining vapour in the air mass relatively depleted in heavier isotopes (low $\delta^{18}\text{O}$ value; Fig. 2.1). As an air mass moves poleward over land from the source, it will slowly become depleted in ^{18}O and D in relation to VSMOW due to condensation and rainfall or snowfall events (Dansgaard, 1953). In consequence, precipitation in the Canadian Arctic that originates from the Pacific Ocean has negative values of $\delta^{18}\text{O}$ with respect to VSMOW (Dansgaard, 1964). Equilibrium fractionation that occurs in the atmosphere during evaporation can be described by the Rayleigh equation (Rayleigh, 1896):

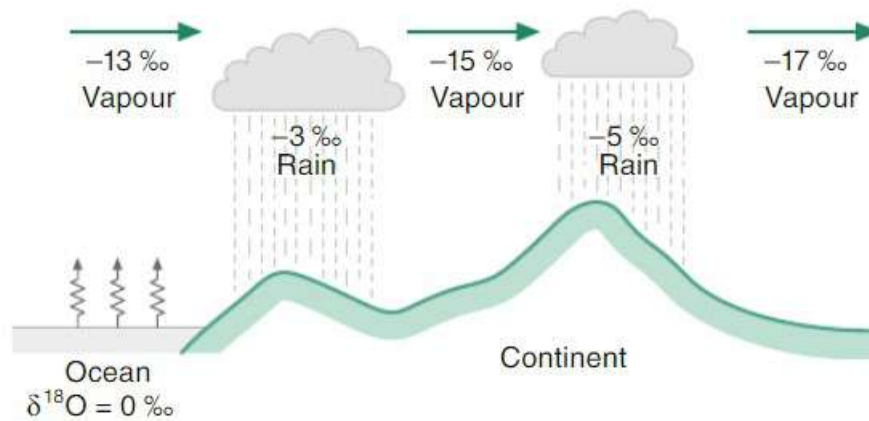


Figure 2.1: Schematic $\delta^{18}\text{O}$ fractionation of water in the atmosphere. From Hoefs (Fig. 3.15, 2015). Reproduced with permission from Springer Nature.

$$R_V = R_{V_o} f^{\alpha-1} \quad (2.2)$$

where R_V is the concentration ratio of the vapour, R_{V_o} is the concentration ratio of the diminishing reservoir, f is the proportion of the original substance in the reservoir, and α is the fractionation factor during formation of ice crystals at the cloud temperature. For the evaporation from a waterbody to vapour, this can be expressed as:

$$\frac{{}^{18}\text{O}}{{}^{16}\text{O}} = \left(\frac{{}^{18}\text{O}}{{}^{16}\text{O}}\right)_O f^{\alpha-1} \quad (2.3)$$

where $\frac{{}^{18}\text{O}}{{}^{16}\text{O}}$ represents the concentration of oxygen isotopes within the vapour, and $\left(\frac{{}^{18}\text{O}}{{}^{16}\text{O}}\right)_O$ is that of the waterbody. By using the abundances for ${}^{18}\text{O}$ and ${}^{16}\text{O}$ from Table 2.1 (2000 ppm and 997680 ppm, respectively), the fractionation factor determined earlier (1.009), and assuming 99% of the original waterbody remains:

$$R_V = \frac{2000}{997680} (0.99)^{(1.009-1)} \quad (2.4)$$

The proportion of ${}^{18}\text{O}$ in the air mass would be 0.001998, or 1998 ppm. This demonstrates the preferential evaporation of the lighter ${}^{16}\text{O}$ from a waterbody, as the vapour has a smaller concentration of ${}^{18}\text{O}$ than the water, thus the vapour is depleted in ${}^{18}\text{O}$.

During laboratory experiments, Jouzel and Merlivat (1984) showed that when temperatures drop below -20°C , isotopic values fail to be predicted by the Rayleigh model. This is due to the rapid formation of ice crystals through vapour deposition (Rogers and Yau, 1996). Thus, kinetic fractionation occurs and formation is no longer occurring under equilibrium conditions (Jouzel and Merlivat, 1984). According to the Bergeron-Findeisen process, the temperature at which precipitation forms in cold clouds is always near -20°C (Ross, 2017).

A fractionation coefficient (α_s) for the vapour-solid phase change was determined by Merlivat and Nief (1967), which depends only on temperature. Between 0°C and -40°C, the α_s :T relation can be represented by:

$$\alpha_s = -9.45(10^{-2}) + \frac{16.289}{T^2} \quad (2.5)$$

where T is the temperature during phase change (°C). The Rayleigh model operates under the assumption that the condensate is removed from the system immediately, as in the case of precipitation. So the isotopic composition of precipitation from a saturated air mass is initially dependent upon (1) the temperature of condensation, and (2) the isotopic composition of the initial water body (Dansgaard, 1964; Merlivat and Jouzel, 1979). Additionally, factors such as sea surface temperature, relative humidity, and wind speed can further influence the isotopic composition of an air mass since they influence α and the extent of equilibrium and kinetic fractionation (Merlivat and Jouzel, 1979; Gat, 1996; Jouzel et al., 2007; Benetti et al., 2014).

Kinetic Fractionation

Continuing with the water/water vapour example, kinetic fractionation occurs when there is an imbalance between evaporation and condensation in the water and vapour (Craig, 1961; Dansgaard, 1964; Hoefs, 2015). This process is generally assumed to occur during the formation of precipitation. α must account for the diffusion ratio (μ) between the light and heavy isotopes, thus the kinetic fractionation factor (k) becomes:

$$k = \mu\alpha \quad (2.6)$$

Assuming α for D and ^{18}O at 25°C: $k_D = \sqrt{19/18} \cdot 1.08 = 1.11$ for HD^{16}O , and $k_{18} = \sqrt{20/18} \cdot 1.009 = 1.063$ for H_2^{18}O . Where 19 and 20 are the molecular weights of HD^{16}O and H_2^{18}O , respectively, and 18 is the weight of H_2^{16}O . Kinetic fractionation

can be interpreted as the consequence of different isotopic components having varying rates of reaction, which results in an uneven partition of isotopes (Dansgaard, 1964; Clark and Fritz, 1997). Kinetic fractionation influences $\delta^{18}\text{O}$ more than δD (Dansgaard, 1964).

When an air mass is forced to gain altitude as it travels up a mountain slope, rapid cooling of the air mass results in condensation and the formation of precipitation (Holdsworth et al., 1991; Poage, 2001). There may be a substantial amount of kinetic fractionation within the air mass.

2.2 $\delta^{18}\text{O}$ & δD in Precipitation

The strong linear relation between $\delta^{18}\text{O}$ and δD is due to the proportional rates of fractionation of oxygen and hydrogen isotopes during evaporation and condensation. This is the result of corresponding differences in the vapour pressures of H_2O and HDO , and H_2^{16}O and H_2^{18}O (Craig, 1961). The global meteoric water line (GMWL) is a statistical relation originally derived by Craig (1961, Fig. 1) using the $\delta^{18}\text{O}$ and δD values of samples from water bodies around the world:

$$\delta D = 8\delta^{18}\text{O} + 10\text{‰} \quad (2.7)$$

where the slope (8) is from the proportional rates of fractionation of D (1.08) and ^{18}O (1.009), and the intercept (10) is an empirical constant (Dansgaard, 1964; Putman et al., 2019). The constant is a result of the average humidity above an ocean being 85%, which causes an enrichment of D within the air mass by 10‰ when compared to the ocean water (Kendall and Caldwell, 1998). Since 1960, the International Atomic Energy Agency (IAEA) has conducted monthly surveys of precipitation from stations around the world, providing a large database of isotope values for worldwide precipitation. Using some of these data, the GMWL based on the

long-term arithmetic means of each station has been reported as ($R^2 = 0.99$, $n = 206$; Rozanski et al., 1993, eq. 2):

$$\delta D = (8.17 \pm 0.06)\delta^{18}O + (10\text{‰} \pm 0.65) \quad (2.8)$$

2.2.1 *Regional Variation*

Regional variation in the $\delta^{18}O:\delta D$ relation was assessed by Rozanski et al. (1993), who noted that marine locations typically have a weaker correlation between the two isotopes. In one extreme, the island of St. Helena yielded an R^2 value of 0.158, whereas R^2 for Ottawa and Vienna, both continental stations, were greater than 0.96. This was assumed to reflect the varying contribution of air masses from different sources at St. Helena, and the possibility of fresh air masses at the first stage of the rain-out process (Rozanski et al., 1993). This effect becomes less influential at higher latitudes, as other factors such as seasonal temperature variations become more influential (Araguás-Araguás et al., 2000).

As an air mass proceeds northward (southward in the southern hemisphere), physiographic features may influence its isotopic composition. Rozanski et al. (1993) noted a drop in $\delta^{18}O$ values of 2‰ per 1000 km travelled for air masses in Europe. However, Holdsworth et al. (1991) reported that in the St. Elias Mountains, $\delta^{18}O$ values in precipitation decrease by 0.6‰ per 100 m of elevation gain. The change in $\delta^{18}O$ with elevation is caused by Rayleigh distillation as precipitation occurs where the air mass rises and cools rapidly.

Regional variation requires Local Meteoric Water Lines (LMWL) to be created for precipitation specific to an area, as with derivation of a modern LMWL for the Inuvik area (Fritz et al., 2022, Fig. 3a):

$$\delta D = 7.39\delta^{18}O - 6.7 \quad (2.9)$$

2.2.2 Deuterium Excess

The deuterium excess is calculated from the $\delta^{18}\text{O}$ and δD concentrations in water, and when used for precipitation can be a method of estimating the kinetic fractionation that has occurred (Dansgaard, 1964; Merlivat and Jouzel, 1979; Jouzel and Merlivat, 1984). The D-excess is the intercept of the $\delta^{18}\text{O}:\delta D$ relation, and is given by:

$$d = \delta D - 8\delta^{18}\text{O} \quad (2.10)$$

The D-excess effectively indicates the difference in $\delta^{18}\text{O}$ and δD of a sample with reference to the expected difference for a sample of VSMOW (Pfahl and Sodemann, 2014). Variations in D-excess indicate changes in fractionation factors because of diffusivity differences in the vapour phase, as well as mass differences, between ^{18}O and D. The average D-excess in global precipitation is near 10‰, but it is generally lower near the poles and higher near the tropics (Feng et al., 2013; Benetti et al., 2014; Dütsch et al., 2017).

Once an air mass crosses onto land, it loses the constant replenishment from the ocean and begins to deplete progressively following the Rayleigh model (eq. 2.1.1). Since the isotopic composition of newly evaporated vapour over an ocean is influenced by the sea surface temperature, relative humidity, and wind speed (Merlivat and Jouzel, 1979; Gat, 1996; Benetti et al., 2014), precipitation from an air mass may provide insight into the atmospheric conditions during formation. Further, D-excess is more indicative of the temperature at the source rather than the temperature during precipitation formation (Jouzel et al., 2007).

Any subsequent alterations of D-excess are indicative of fractionation occurring to an air mass through secondary processes such as recycled water vapour from plant transpiration or the open surface evaporation of large water bodies (Machavaram and Krishnamurthy, 1995). Depleted water vapour mixing and increased recycling

within the air mass will result in a higher D-excess in precipitation (Machavaram and Krishnamurthy, 1995; Masson-Delmotte et al., 2005).

2.2.3 *Snowpack Dynamics*

The snowpack accumulates winter precipitation (Taylor et al., 2001; Laudon et al., 2002; Ala-aho et al., 2017). The isotopic composition of a snowpack may be altered by various ice-water-vapour processes (Taylor et al., 2001, 2002; Earman et al., 2006). For example, sublimation results in the enrichment in heavier isotopes of the upper layers within the snow cover (Gat and Gonfiantini, 1981). Similarly, there is significant kinetic fractionation throughout the entire snowpack due to freeze/thaw cycles (O'Neil, 1968; Ala-aho et al., 2017). Further, various processes during the spring thaw result in the early season melt generally being more depleted, with the composition of runoff gradually becoming more enriched (Lauriol et al., 1995; Laudon et al., 2002; Earman et al., 2006). Therefore, early season melt may have a $\delta^{18}\text{O} : \delta D$ slope less than the GMWL, and the slope for late season melt may be near or greater than the GMWL (Lauriol et al., 1995).

Significant variation in snowpack composition and dynamics may occur due to the physiographic features within a region (Ala-aho et al., 2017, 2021). Ala-aho et al. (2021) showed that differences exist in the isotopic composition of snowpacks in taiga and tundra environments. They found that the spatial variability of isotopic composition in tundra snowpacks was much higher than in the taiga due to drifting and high sublimation rates.

2.2.4 *Use of Least-Squares Linear Regression*

Least-squares linear regression (LSL) has commonly been used to determine $\delta^{18}\text{O}:\delta D$ relations, for instance to create a LMWL (e.g., Putman et al., 2019; Fritz et al., 2022). In this method, all error is assigned to the variable plotted on

the Y-axis (Mark and Church, 1977). In fact, determinations of δD and $\delta^{18}O$ are both subject to error (Burn and Maxwell, 1993). In aggregate, a weighted error approach is best used to describe the $\delta^{18}O:\delta D$ relationship (Burn and Maxwell, 1993). The functional fit minimizes the distance between the points and the fitted line in both directions. In the context of $\delta^{18}O$ and δD , the analytical errors are typically assumed to be $\pm 0.2\text{‰}$ and $\pm 2\text{‰}$ respectively (Burn and Maxwell, 1993).

The slope (b_f) of the functional fit is (Mark and Church, 1977, eq. 9b):

$$b_f = \frac{(b_r^2 + R^2 - \lambda) + \sqrt{(\frac{b_r^2}{R^2 - \lambda})^2 + 4\lambda b_r^2}}{2b_r} \quad (2.11)$$

where b_r is the slope from the least-squares linear regression, R^2 is the coefficient of determination from the regression, and λ represents the ratio of error variances for $\delta^{18}O$ and δD , i.e., 10.

2.3 $\delta^{18}O$ - Temperature Relation

There is considerable literature regarding the relation between the isotopic composition of precipitation and temperature (e.g., Dansgaard, 1964; Jouzel et al., 1997; Fricke and O'Neil, 1999). Although the temperature of the air mass controls condensation, it is the interaction between the condensate and water vapour at the warmer temperatures of the cloud base that is thought to provide the basis of the relation between surface temperature and $\delta^{18}O$ (Gat, 1996; Fricke and O'Neil, 1999). The correlation between the two has been observed best in continental regions nearer to the poles, whereas in the tropics a stronger correlation is seen between the quantity of rainfall and $\delta^{18}O$ (Lawrence and White, 1991; Rozanski et al., 1993).

Dansgaard (1964) determined that there was a significant regional $\delta^{18}O : T$ relation between precipitation and annual mean temperatures at multiple stations throughout the North Atlantic and Greenland. He derived equation 1.2, which has

been used as a benchmark for the relation. Porter et al. (2016, eq. 6) determined a $\delta^{18}\text{O} : T$ relation using annual precipitation data from around Yukon:

$$\delta^{18}\text{O} = 0.41T - 20.2 \quad (2.12)$$

which has a slope lower than the $0.69 \text{ } \text{‰}^\circ\text{C}^{-1}$ calculated by Dansgaard (1964, eq. 9), or the relation at Ottawa ($0.49 \text{ } \text{‰}^\circ\text{C}^{-1}$; Rozanski et al., 1993, Table 2) or from Siberia ($0.50 \text{ } \text{‰}^\circ\text{C}^{-1}$; Kostrova et al., 2020, Fig. 4a).

2.3.1 *Air Mass Origin*

As the Rayleigh model shows, the isotopic composition of an air mass is influenced in part by the isotopic composition of the source waters and prior rainout events (Fig. 2.1). Koster et al. (1992) used a general circulation model (GCM) to simulate July precipitation in Antarctica and found that $\delta^{18}\text{O}$ varied consistently with the temperature of the evaporative source. The situation is more complex over Greenland, where Charles et al. (1994) used a GCM to estimate the isotopic composition of precipitation and found precipitation from the North Pacific had roughly 15‰ lower $\delta^{18}\text{O}$ than from the North Atlantic.

Charles et al. (1994) attributed the lower $\delta^{18}\text{O}$ from the Pacific to be due to the air mass having travelled along a longer, colder route before reaching the northern coast of Greenland. As an air mass travels, it replenishes water vapour in the atmosphere from various sources at the surface. Lakes or near shore ocean water offer a large source of replenishment (Machavaram and Krishnamurthy, 1995), while smaller scale sources such as the transpiration of plants and forests also contribute moisture to the air (Salati et al., 1979; Koster et al., 1993). Overall, however, an air mass generally becomes more depleted in heavy isotopes as distance from the source increases.

Specific to the region of interest for this thesis, Bailey et al. (2015) noted a difference of the $\delta^{18}\text{O}$ of precipitation in the western Arctic as a result of strong shifts in the Aleutian Low over the North Pacific. This suggests that origin has an impact on the isotopic composition of precipitation in the region.

2.4 Permafrost Terrain

Permafrost is defined as ground that remains at or below 0°C for at least two consecutive years (Harris et al., 1988). The active layer is the ground directly above permafrost that freezes and thaws annually (Muller, 1947; Burn, 1998). The depth of the active layer may vary annually and is dependent on numerous configurational factors including air temperature, soil type and water content, vegetation, aspect, and slope characteristics (Mackay and Burn, 2002a; French, 2007). The annual freezing of the active layer begins in the fall, primarily from the top downward, although freezing may also occur from the top of permafrost upwards (Mackay, 1984a).

A more comprehensive three-layer model of permafrost was introduced by Shur et al. (2005), to supplement the original two-layer model. The third layer is known as the *transition zone* and is located below the bottom of the active layer at the top of permafrost. The transition zone acts as a buffer protecting the permafrost below from thaw and is the primary location of downward water movement (Mackay, 1983a). The transition zone is ice rich and contains segregated ice lenses and ice veins (Shur et al., 2005). The uppermost portion of the transition zone is described as the *transient layer*, where thaw is more frequent than at greater depths. Deeper parts of the transition zone are protected from thaw in relatively warm years by the transient layer. Thus, deep thaw is controlled only by long term changes in climate or surface conditions (Shur et al., 2005).

2.4.1 Excess Ice

Excess ice is the volume of ice in the ground which exceeds the total pore volume of the soil when thawed (Harris et al., 1988). Excess ice forms supernatant water when a sample of ice-rich ground thaws. The excess ice content within a soil can be calculated with (eq. 1, Kokelj and Burn, 2003):

$$E_i = \frac{W_v * 1.09}{(W_v * 1.09) + S_v} * 100 \quad (2.13)$$

where W_v is the volume of supernatant water (cm^3) multiplied by 1.09 to account for expansion during freezing, and S_v is the volume of saturated sediment (cm^3) (Kokelj and Burn, 2003).

Ground ice with a gravimetric water content of at least 250% is considered massive ice (Mackay, 1989). There are two categories of massive ice: buried ice, commonly of glacial origin; and intrasedimental ice, which has formed through *in situ* freezing of water within sediments (Mackay, 1989; Mackay and Dallimore, 1992).

Consequences of excess ice

Excess ice is volume added into the ground, and as such causes an uplift of the ground surface. This is referred to as heave, and the subsequent thaw of excess ice results in the lowering of the surface, or subsidence (Mackay, 1970; Streletskiy et al., 2017). The yearly freeze-thaw cycle of the active layer is associated with differential heave and subsidence, which may cause damage to infrastructure as well as being an important driver of periglacial geomorphic processes. Substantial infrastructure and slope failures may result if seasonal air temperatures increase and thaw occurs rapidly within the transient layer.

Pool Ice

Ground water frozen *in situ* after a blockage in drainage is known as pool ice (Mackay, 1988, 1997; French, 2007). It is common in gullies alongside ice wedges where preferential erosion occurs. Obstructions to drainage occur as thaw progresses and masses of soil or slumped material plug gullies and cause a backup of standing water. This water may then freeze in subsequent winters (French, 2007). Pool ice is abundant in poorly drained tundra regions and was present in the near-surface of at least three of the seven ice wedges exposed during drainage of Lake Illisarvik (Mackay, 1997). Pool ice was adjacent to at least one of the six anti-syngenetic ice wedges sampled for this thesis.

2.5 Ice Wedges

Ice wedges are estimated to represent nearly 35% of all excess ice found in the top 3 m of earth materials on Richards Island (Pollard and French, 1980). Cracks in the ground occur when a rapid drop in temperature causes intense thermal contraction stresses that are greater than the tensile strength of the material (Lachenbruch, 1962). The crack may then be infilled with snowmelt during the spring that then freezes, blocking further infiltration. If the process repeats year after year, a distinct wedge of nearly pure ice forms (Fig. 2.2; Mackay, 1974a).

2.5.1 Thermal Contraction (or Expansion)

The most important condition required for cracking is a rapid temperature drop in already cold ground, which typically occurs in January or February in the western Canadian Arctic (Mackay, 1993; Fortier and Allard, 2005). Cracking is also dependent on the coefficient of thermal expansion (β) associated with the material of the ground. β is defined as the fractional change of length in response to change

in temperature (Andersland and Ladanyi, 2004). For ice, β varies with temperature $\theta(^{\circ}\text{C}) = 0 - T(^{\circ}\text{C})$ according to:

$$\beta(^{\circ}\text{C}^{-1}) = (54 - 0.18\theta)10^{-6} \quad (2.14)$$

At the air temperature that cracking occurs in permafrost (observed at -25°C to -40°C in the western Canadian Arctic; Mackay, 1993), this value is roughly $5 \times 10^{-5}\text{C}^{-1}$. Considering β is higher in ice than sand by a factor of about 5 (Andersland and Ladanyi, 2004), ice-rich soils develop thermal stresses sufficient for cracking at higher temperatures than dry soils with pore ice alone.

2.5.2 *Growth Over Time*

The three types of wedges (epigenetic, syngenetic, and anti-syngenetic) develop different surface morphology. Epigenetic wedges develop a complex polygonal network due to the consistent surface level and soil movement accompanying growth (Fig. 2.2) (Mackay, 2000; Mackay and Burn, 2002b). Syngenetic wedges create a similar surface morphology, but the effect is muted due to the narrow width and constant aggradation of the surface (Morse and Burn, 2013). Anti-syngenetic ice wedges, however, are developed at sites of erosion. Downslope movement of material fills in the trough (Burn et al., 2021). Thus, most anti-syngenetic ice wedges were unrecognized in the field until recently. They were initially identified by Mackay (1990), and subsequently examined by Mackay (1995) and Burn et al. (2021). Rapid warming in the Arctic regions has caused an acceleration in the degradation of anti-syngenetic ice wedges, which is occurring faster than the erosion of the ridges, causing broad troughs to become visible on many hillslopes (Burn et al., 2021).



Figure 2.2: An epigenetic ice wedge in early spring. A clear illustration showing a thermal contraction crack, and the deformation of stratigraphy along the edges of the ice wedge in the lighter soil. Photo reproduced with permission from C.R. Burn.

A rapid temperature drop must occur in the ground for cracking to occur (Lachenbruch, 1962). During the initial stages of ice-wedge growth this may be a yearly occurrence (Mackay and Burn, 2002b), but as vegetation and the ridges develop, more snow is trapped in the trough and the polygon, insulating the ground below. As a result, more extreme air temperature changes are required for the appropriate cracking conditions to occur. Mackay (1974a) noted that larger wedges (> 2 m wide) have a relatively low cracking frequency, related to snow accumulation and competition with other active wedges. In fact, Mackay (1992a) found little correlation between mean air temperature and cracking frequency for ice wedges in the western Canadian Arctic. Two sites on Garry Island within 1 km of each other showed little correlation in ice-wedge cracking occurrences (Mackay, 1993). In these cases, the wedges were exposed to the same variation in air temperatures, and thus should have cracked at similar times and rates. Mackay (1993) noted an average difference of 6°C in ground temperatures at a depth of 0.5 m beneath the trough and ridge of these wedges, indicating the impact that snow within the trough can have on the temperature regime of the wedge.

2.5.3 *Age and Activity*

If the width of an ice wedge directly controls the probability of cracking events as observed by Mackay (1974a), then the probability of cracking may lie on a normally distributed curve:

$$P = \frac{1}{\sigma\sqrt{2\pi}} e^{\frac{-(w-\bar{w})^2}{2\sigma^2}} \quad (2.15)$$

where w is the width of an ice wedge (m), $\bar{w}$ is the average ice wedge width within the region, and σ is the standard deviation of width. This represents the cracking probability under current conditions and cannot be used for past environments (Mackay, 1974a). Once the probability of cracking is known, one can estimate how

long it will take for an ice wedge to grow a certain amount:

$$T_{ij} = \frac{(w_j - w_i)}{\Delta w P_{ij}} \quad (2.16)$$

Where T_{ij} is years required for the wedge to increase in width from w_i to w_j , and P_{ij} is the mean crack probability for this interval (Mackay, 1974a). Δw represents the growth increment for ice wedges in the region and can vary from ≤ 1 millimeter to several millimeters (Mackay, 1974a). It is important to note that the young ice wedges of Illisarvik grew from 1 to 3 cm per year during the first 10 years of growth (Mackay and Burn, 2002b), so age estimation must consider the variation in growth rate as the ice wedge becomes larger.

Anti-syngenetic ice wedges

Mackay (1990) stated that the top of permafrost width (W) of an anti-syngenetic ice wedge can be represented by:

$$W = \frac{\Delta w}{2\Delta z}(L + \Delta z) \quad (2.17)$$

Where Δw is the average crack width (m) Δz is the annual increment of denudation and the annual deepening of the ice wedge at its apex (m), and L is the depth of the wedge (m). The key factors controlling Δz are erosional processes and for Δw the winter climate and thermal contraction coefficient. Thus, slope angle is key for controlling Δz , and snow depth and climate history are of importance for Δw (Burn et al., 2021). Mackay (1990) noted that the ratio between Δw and Δz is likely the critical variable controlling width. Assuming a mean annual crack width of 2 mm (Mackay, 1974a), annual denudation of 0.3 mm a^{-1} (as found on Garry Island; Mackay, 1981), and an ice wedge depth of 3 m, the top of permafrost width will be

10 m. This calculation also assumes that the thermal contraction crack penetrates the entire wedge.

The age of an anti-syngenetic ice wedge may be estimated by dividing the depth of the wedge by the annual denudation (Mackay, 1990). Assuming a similar wedge with a depth of 3 m and a constant annual denudation of 0.3 mm, the wedge is 10,000 years old.

3 Study Area

3.1 Introduction

This section provides an assessment of the present and historic climate of the western Canadian Arctic, and a review of the terrain subdivisions of Richards Island. The cryostratigraphy of the region is described to link the geologic and climatic history of the region with the modern environment.

3.2 Geographic Setting

Field work for this thesis took place at the Illisarvik drained lake basin on Richards Island (Fig. 1.1). The most recent deglaciation of the region is thought to have occurred no later than 14.3 ka BP (Rampton, 1988; Murton et al., 2007). All dates referenced are denoted as 'ka', and are representative of calibrated years before present. The study area lies in the Mackenzie Delta area east of the Western Cordillera (Fig. 3.1), which provides a barrier for air masses that approach the region from the south or west. About 1000 km to the south, the Pacific Coast is adjacent to St. Elias and Coast mountains, the tallest ranges in Canada, with elevations up to over 4000 m a.s.l (Fig. 3.1). To the north and east of these mountains is the Yukon Plateau, an area of forests and abundant lakes. If approaching from the south, Ogilvie and Mackenzie mountains are the final barrier before reaching the flat lowlands of Mackenzie Valley. From the west, Richardson Mountains represent the final range before reaching Inuvik (Fig. 3.1).



Figure 3.1: Elevation map of western North America. The study area is outlined by the red box (see also Fig. 1.1).

3.3 Climate

The distinct climates of Inuvik and Tuktoyaktuk are important for this thesis, and at both, the annual mean air temperature has risen since 1971 (Thienpont et al., 2013; Schetselaar et al., 2023).

Inuvik

Inuvik (68.36° N, 133.72° W; Fig. 1.1) is on the east side of Mackenzie Delta. The town is south of treeline (Fig. 1.1) and at an elevation of 15 m a.s.l. Beaufort Sea is about 100 km north of Inuvik, and the land to the east is the flat and rolling terrain of the Interior Plains. In general, air masses that arrive at Inuvik have either just made landfall from Beaufort Sea or travelled a great distance from southern or western moisture sources.

Data from 1991-2020 indicate that the mean annual temperature at Inuvik is -6.7°C, -3.2°C higher than in 1951-80 (Schetselaar et al., 2023). Inuvik has a monthly mean temperature below 0°C from October to April; January is the coldest month, averaging -24.4°C (ECCC, 2023). From June through August the mean temperature is above 10°C (Fig. 3.2). The rate of increase in annual mean air temperature since 1971 has been 0.77°C/decade (Schetselaar et al., 2023). Inuvik had a mean annual precipitation of 249 mm in 1991-2020, the majority of which falls between July and October (137 mm) (Schetselaar et al., 2023).

Tuktoyaktuk

Tuktoyaktuk (69.44° N, 133.03° W; Fig. 3.1) is a coastal community north of treeline. Air masses that reach Tuktoyaktuk are similar to those that arrive in Inuvik. Illisarvik is about 60 km west of Tuktoyaktuk, and has a very similar climate to that of the community (Burn and Zhang, 2010; Andersen et al., 2021).

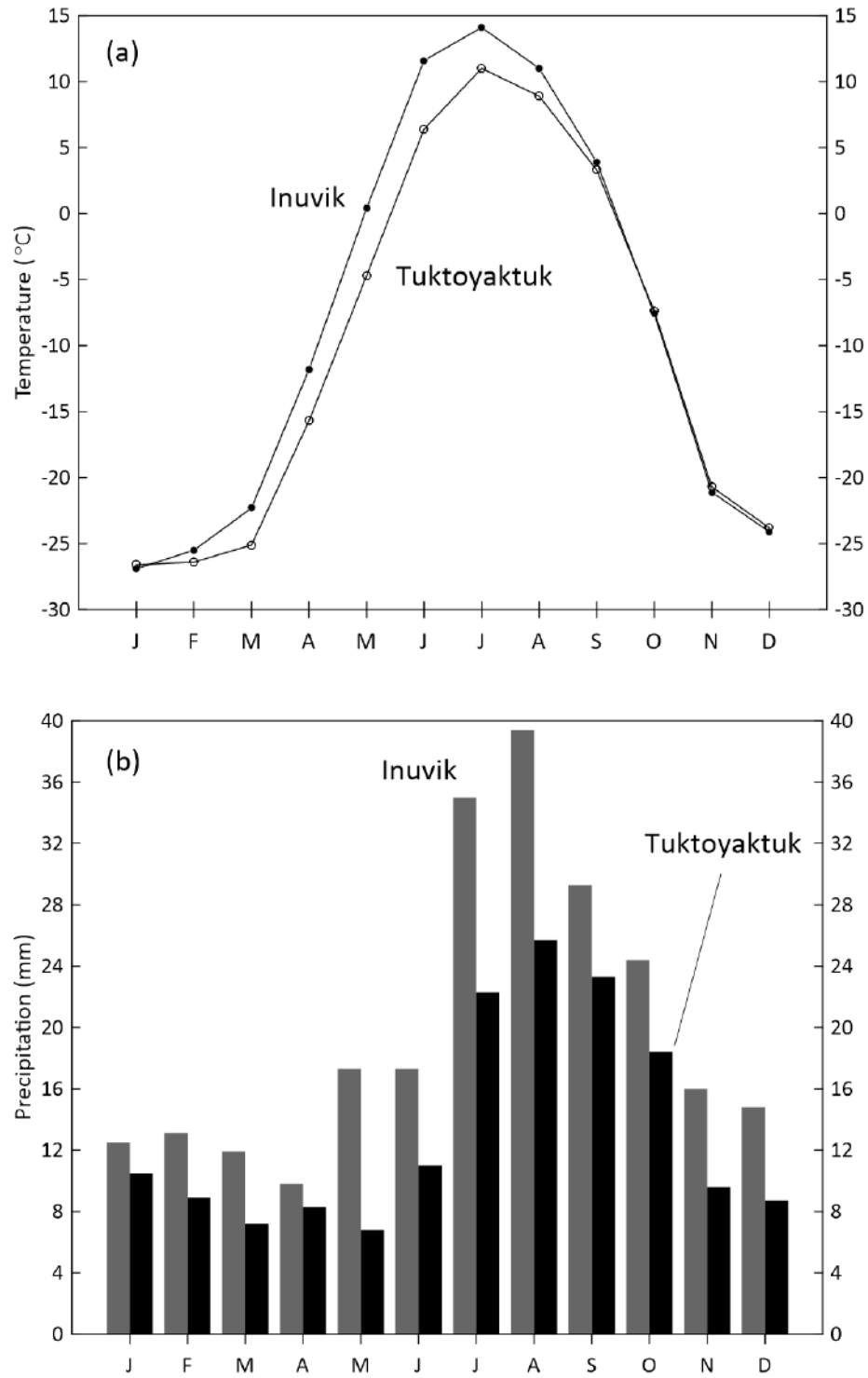


Figure 3.2: 30-year (1981-2010) monthly mean air temperature (a) and precipitation (b) for Inuvik and Tuktoyaktuk (ECCC, 2023).

The mean annual temperature from 1981-2010 was -10.1°C (ECCC, 2023). Both January and February have mean temperatures below -26°C , and the only month with a mean temperature higher than 10°C is July (Fig. 3.2) (ECCC, 2023). The mean temperature is above 0°C from June through September (ECCC, 2023). Precipitation at Tuktoyaktuk is low, with mean annual precipitation in 1981-2010 only 161 mm, primarily falling in late summer (ECCC, 2023). The climate at Tuktoyaktuk and on Richards Island is heavily influenced by the Arctic Ocean and sea ice presence throughout the spring and summer (Burn, 1997).

3.4 Environmental History During the Holocene

3.4.1 Holocene Thermal Maximum

Warming associated with a natural increase in solar insolation followed the recession of the Laurentide Ice Sheet (LIS) (Berger and Loutre, 1991). This period is known as the Holocene Thermal Maximum (HTM) and lasted in the Tuktoyaktuk Coastlands from ~ 10 ka BP to ~ 5 ka BP (Ritchie and Hare, 1971; Kaufman et al., 2004). These dates correlate well with the advance into and retreat of spruce forests from the Tuktoyaktuk Coastlands (MacDonald, 2000), with pollen evidence showing the warmest period began 9.5 ka BP (Ritchie et al., 1983; Burn, 1997). Various wood fragments collected by Mackay (1992b) indicate forest was present in the region as late as 6.5 ± 0.1 ka. Mean annual temperatures during this time were thought to be $1\text{-}3^{\circ}\text{C}$ higher than at present (Kaufman et al., 2004), with an annual increase of 200 growing degree-days (referenced to 5.5°C ; Ritchie, 1984).

The existence of subsea permafrost indicates a shoreline during the Pleistocene and Holocene significantly further north than today (Mackay, 1972a). A nick point at the edge of the continental shelf is present some 100 km from the current coastline, which was estimated to be formed about 14 ka BP (Pelletier, 1987). Between 60-70 km from the current shoreline is a terrace resembling a beach, where there was likely

a pause in sea level rise at 8.5 - 9 ka BP (Pelletier, 1987). Pelletier (1987) suggested that sea level has been continually rising since then, with levels still more than a meter below present as recently as 1 ka BP ago. Climate is heavily influenced by proximity to the sea, and it is possible that the study area experienced cooling as sea level rose and the coastline moved southward (Mackay, 1992b; Burn, 1997).

3.4.2 5 ka BP to Present

From the end of the HTM to the industrial era there has been general reduction in annual mean temperature (Wanner et al., 2008; Liu et al., 2014). In contrast, the retreat of ice sheets should have caused warming due to decreased albedo and solar reflection (Serreze and Barry, 2011). This has led to the term 'the Holocene temperature conundrum' (Liu et al., 2014). The interpretation of cooling since the HTM has been through paleoclimate reconstruction using proxies such as pollen and tree rings, which primarily represent annual or summer mean temperatures (Kaufman et al., 2004; Bradley, 2014; Liu et al., 2014).

Now that $\delta^{18}\text{O}$ and δD in wedge and glacial ice may be used as proxies for winter conditions, there has been increased understanding of how winter climate has changed during the later portion of the Holocene (e.g., Lecavalier et al., 2017; Holland et al., 2020; Vasil'chuk et al., 2020). Siberian ice wedges measured in two separate regions have both indicated warming of winter climate since 5 ka, at a rate that accelerated with industrialization (Meyer et al., 2015; Opel et al., 2017). These trends have also been seen in the ice wedges of the Tuktoyaktuk Coastlands (Holland et al., 2020).

3.4.3 Sea Ice

Sea ice has an important impact on precipitation and climate within a region (Noone and Simmonds, 2004; Kopec et al., 2016; Linderholm et al., 2018). Sea-ice coverage

and sea location have been noted as an influence on the climate of the study area (Burn, 1997). Proximity to sea ice has a significant impact on spring and summer temperatures at Tuktoyaktuk, where there are nearly 150 fewer growing degree-days in June than at Inuvik (Burn, 1997). This is a result of cold onshore winds blowing over the sea ice.

While the proximity of the ocean to the study area has an impact on the modern climate of the region, the coastline was further north during much of the Holocene (Mackay, 1972a; Polyak et al., 2010; Funder et al., 2011). Synoptic scale impacts on precipitation and climate patterns may result from changing sea ice (Noone and Simmonds, 2004; Kopec et al., 2016; Linderholm et al., 2018). In the Arctic, these include adjustments to the westerlies (Abell et al., 2021; Deng et al., 2022). This implies that the climate and precipitation regime of the study area, though more than 50 km from the coast during the mid-Holocene (Pelletier, 1987), would still have been influenced by variation in sea ice over the Beaufort Sea.

3.4.4 *Lake Illisarvik*

In August 1978, J.R. Mackay began an experiment to study permafrost growth *ab initio* through the drainage of Lake Illisarvik (Mackay, 1997). Draining of the lake has allowed for long-term observation of permafrost growth and development.

Just south of Lake Illisarvik was another lake, now referred to as Pingo Lake (Michel et al., 1989) (Fig. 3.3). The two lakes were originally one. Pingo Lake is thought to have drained about 2 ka, which resulted in a subsequent lowering of Lake Illisarvik (Michel et al., 1989). Extensive sampling of permafrost in boreholes around the area began shortly after drainage of Lake Illisarvik in 1978 (Michel, 1982). The study sites for this thesis were not covered by either lake, and thus the permafrost at the sites was not influenced by the lakes (Rampton, 1988; Murton, 2009).

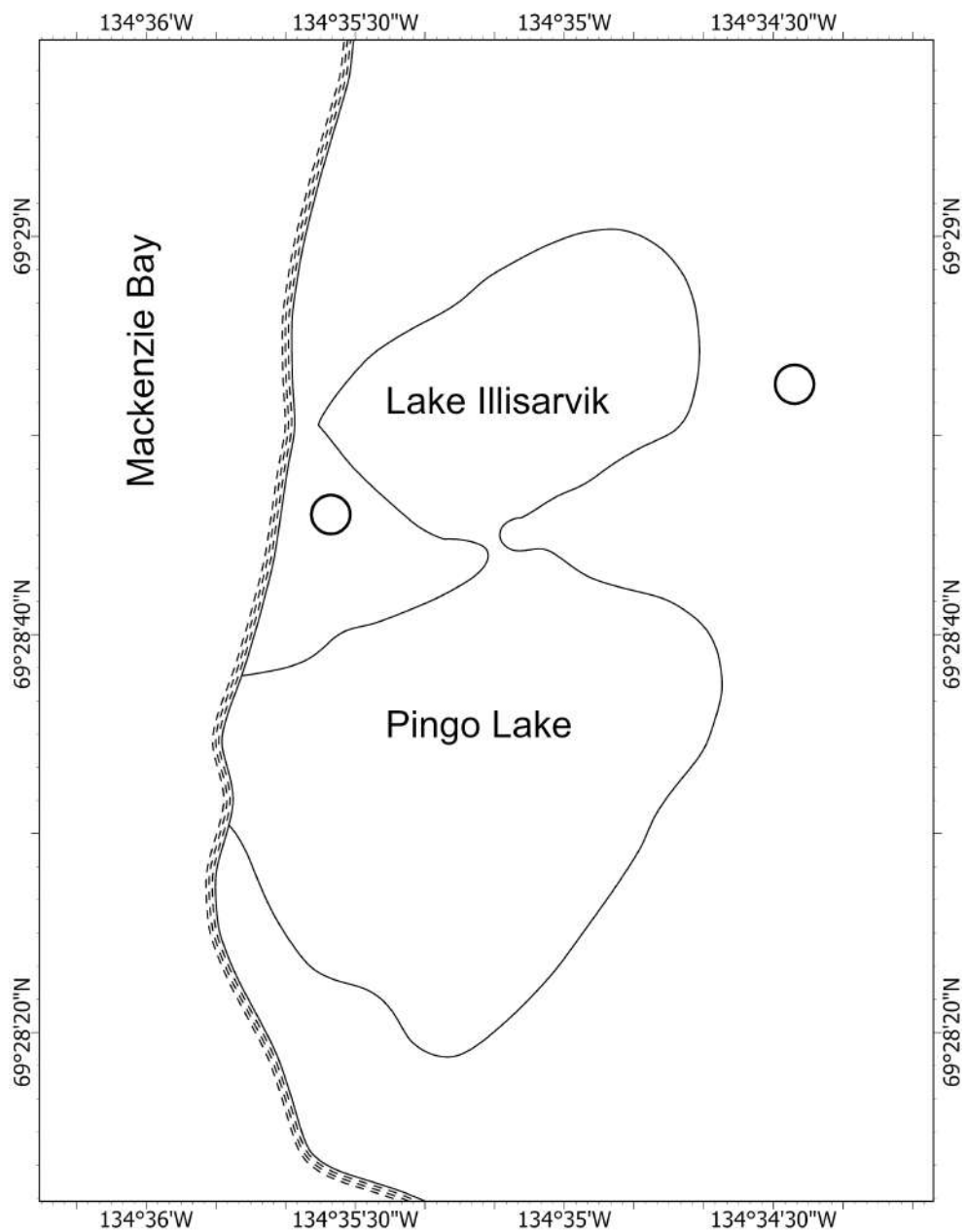


Figure 3.3: Map of Lake Illisarvik and Pingo Lake when joined shortly before the drainage of Pingo Lake. Circles indicate sampling sites for this thesis.

3.5 Physiographic Subdivisions

Richards Island (Fig. 1.1, 3.4) has three distinct physiographic subdivisions, which form a part of the broader Tuktoyaktuk Coastlands. The island is separated into Big Lake Delta Plain (BLD), Tununuk Low Hills (TLH), and Kittigazuit Low Hills (KLH) (Fig. 3.4; Rampton, 1988). Illisarvik is located within TLH, which are comprised of sediments ranging from clay to sandy gravel and a rolling topography (Mackay, 1963; Rampton, 1988).

3.5.1 Vegetation

Vegetation found within TLH varies greatly with topography, soil moisture, and nutrient conditions (Mackay, 1963). The uplands of the Tuktoyaktuk Coastlands are well drained and the shrub vegetation is primarily *Salix* spp. (low willows), *Betula glandulosa* (dwarf birch), and an assortment of herbs, lichens, and moss (Rampton, 1988). The slopes and valleys typically support willows and *Alnus viridis* subsp. *crispa* (alders) while the poorly drained areas are dominated by *Carix* spp. (sedges) (Mackay, 1963).

3.6 Geologic Foundation

The lithostratigraphy of Richards Island has been well documented (Mackay, 1963; Rampton, 1988). There are seven main units that were deposited during the Pleistocene, in ascending stratigraphic order, these are: the Kendall sediments, Hooper clay, Kidluit Formation (Fm), Kittigazuit Fm, Toker Point Member (Mb), Turnabout Mb, and a pebble-boulder lag that currently sits a few meters below the surface (Dallimore et al., 1997; Murton, 2009).

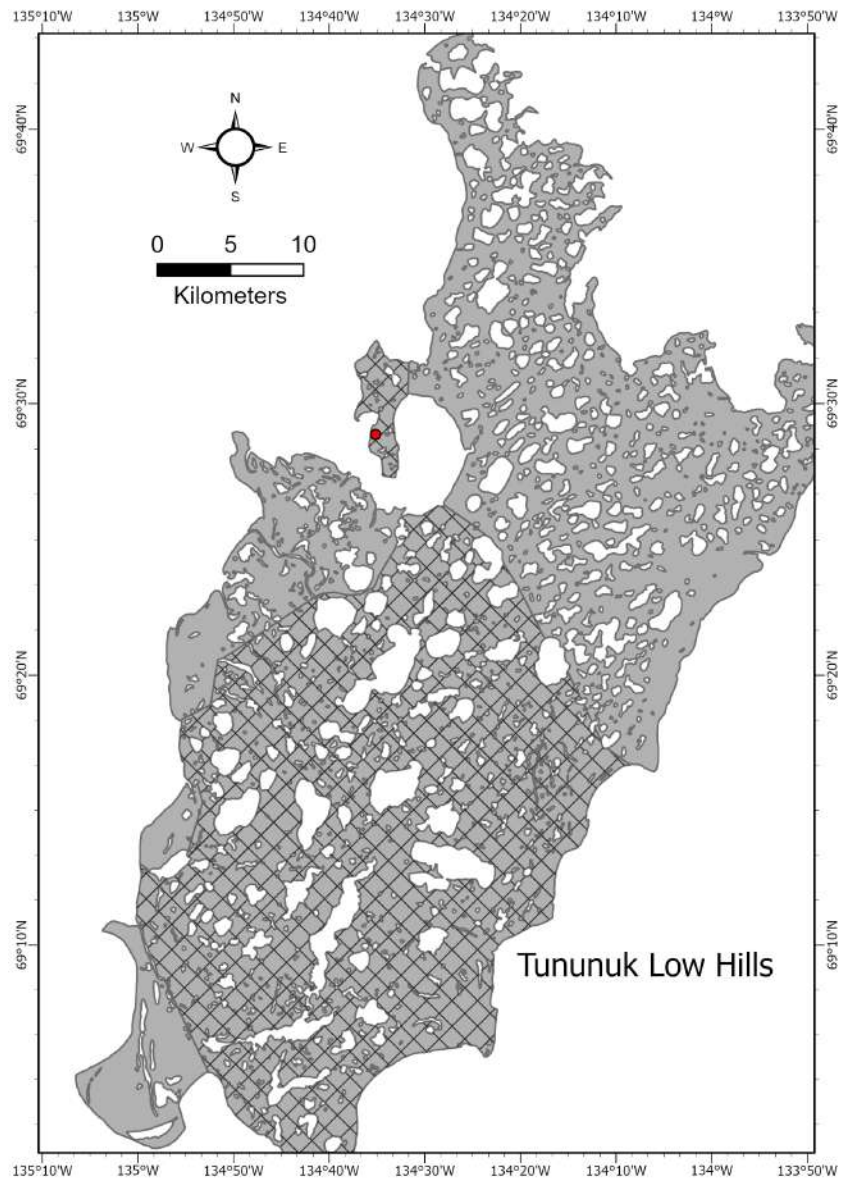


Figure 3.4: Richards Island, with the Tununuk Low Hills denoted by the cross-hatch pattern and Illisarvik by the red dot. The land to the west of TLH is the Big Lake Delta Plain, and to the north is the Kittigazuit Low Hills.

Near Illisarvik, the top 2 m of soil consists of a veneer of silty till and organic matter, which rests above the sands of the Kittigazuit Fm (Michel et al., 1989). Of interest to this thesis are the Kidluit and Kittigazuit Fms, as the latter is developed from the former.

3.6.1 *Units*

The Kidluit Fm is a well-sorted, stratified sand that was deposited by a braided river system before the culmination of Wisconsin glacialiation (Rampton, 1988; Murton et al., 2017). Dating of material from this Fm suggests an age range of 76 - 27 ka BP (Murton et al., 2017). $\delta^{18}\text{O}$ values from ground ice of -20.6‰ match those of modern day active-layer pore water and are consistent with warmer than current ground temperature conditions (Mackay and Matthews, 1983).

The transition from the fluvial deposition of the Kidluit Fm to the eolian deposition of the Kittigazuit Fm occurred due to a blockage of a prominent river system within the region at the Last Glacial Maximum approximately 25 ka BP (Lemmen et al., 1994). The dry landscape was windswept, and the Kittigazuit Fm was deposited, consisting primarily of eroded sand from the Kidluit Fm (Dallimore et al., 1997). Seed and moss fragments from the Kittigazuit Fm yield radiocarbon dates of 37.4 ± 0.81 ka BP and 33.71 ± 0.46 ka BP (Dallimore et al., 1997), while sand has provided luminescence dates between 25.6 ± 1.3 ka BP and 12.8 ± 0.8 ka BP (Bateman and Murton, 2006; Murton et al., 2007). The youngest material of the Kittigazuit Fm corresponds well with the expected deglaciation of the Tuktoyaktuk Coastlands (Murton et al., 2007). Upon drilling at Illisarvik, these sands were found to be near the surface, beneath a layer of till and organic material (Michel, 1982).

3.7 Cryostratigraphy

The cryostratigraphy and lithostratigraphy of a region can be linked through time; however, different geologic units may contain ground ice of a variety of origins and ages. This is largely due to configurational factors within a region, including glaciotectonic deformation and active-layer thawing events (Murton, 2009), as reflected here with the HTM. Warming caused the development of thermokarst lakes and a large thaw unconformity that marks the deepest active layer during the Holocene (Burn, 1997). Wood fragments taken from directly above the unconformity on Pelly and Garry islands suggest that the active layer was deepest 9 ka BP (Burn, 1997), again correlating well with the findings of Ritchie et al. (1983). A distinct isotopic difference exists between Pleistocene age permafrost and any aggradational permafrost that has formed since the HTM (Mackay, 1983b).

There are four prominent stratigraphic positions of ice wedges: (1) beneath the Kidluit-Kittigazuit Fm contact (Mackay and Matthews, 1983); (2) within the Kittigazuit Fm; (3) beneath the thaw unconformity (Mackay, 1975); and (4) between the thaw unconformity and the modern day active layer. The varying stratigraphic position of ice wedges implies that the process of thermal contraction cracking has been occurring since Kittigazuit Fm deposition.

Permafrost on Richards Island reaches depths greater than 400 m (Judge et al., 1987). Recent climate warming has caused the upper 120 m of the permafrost to warm (Burn and Zhang, 2009), resulting in a 2°C rise in average permafrost temperatures since the 1970s (Kokelj et al., 2017). Modern measured ground temperatures at Illisarvik range between -6 and -7°C (Burn and Kokelj, 2009). The top 10 m of permafrost on Richards Island is estimated to be 14.3% excess ice, predominantly ice wedges (Pollard and French, 1980). The mean thaw depth at several sites around Illisarvik increased from 42 to 50 cm between 1983 and 2008 (Burn and Kokelj, 2009). This trend has continued since 2010 (Burn et al., 2021).

There is, however, known regional variation in active-layer depth based upon extrinsic factors such as snow cover and soil moisture. A transect across a valley on Garry Island shows the sites along the dry hillslope have thaw depths 10-15 cm greater than those found in the moist valley bottoms (Mackay, 1974b; Burn and Kokelj, 2009).

3.7.1 Isotopic Values

There is considerable literature discussing the isotopic values found in the permafrost of the western Canadian Arctic. The most isotopically depleted ground ice is from the Pleistocene, typically found with $\delta^{18}\text{O}$ values $< -27\text{‰}$ (Mackay, 1983b). Wedge ice that has developed during the Holocene, above the thaw unconformity, is more enriched, with $\delta^{18}\text{O}$ values between -25‰ and -20‰ (Mackay, 1983b). Boreholes in the recently developed permafrost of the Illisarvik drained lake basin have presented $\delta^{18}\text{O}$ values greater than -18‰ , representative of lake water from prior to drainage (Michel, 1982). These isotopic ranges align well with the $\delta^{18}\text{O}$ values found in wedge ice from the Tuktoyaktuk Coastlands, as values determined within the region are between -29‰ and -20‰ (Table 3.1; Mackay, 1983b; Michel, 2011; Porter et al., 2016). A recent study by Holland et al. (2020) collected 782 samples from 4 ice wedges in the Tuktoyaktuk Coastlands with an average $\delta^{18}\text{O}$ of -24.1‰ (Table 3.1). Holocene ice wedges from Herschel Island have shown more enrichment than the Tuktoyaktuk Coastlands (Table 3.1; Wetterich et al., 2023), with recent sampling resulting in $\delta^{18}\text{O}$ values similar to the modern ice at Illisarvik (Michel, 1982).

Table 3.1: Mean $\delta^{18}O$ for Holocene or modern ice wedges near the study area. Location denoted by * is wedge ice from the Pleistocene.

Location	$\delta^{18}O(\text{‰})$	Studies
Illisarvik Basin	-22.5	Michel (1990)
North Point of Richards Island	-24.6	Holland et al. (2020)
Hooper Island	-23	Holland et al. (2020)
Pelly Island	-25.2	Holland et al. (2020)
Tuktoyaktuk area	-24	Mackay (1983b), Holland et al. (2020)
Herschel Island	-18.4	Wetterich et al. (2023)
Old Crow	-25.5	St-Jean et al. (2011)
Tuktoyaktuk area	-28	Mackay (1983b)

4 Methods

4.1 Introduction

This chapter describes the methods taken for data collection of precipitation at Inuvik between 2015-18 by Fritz et al. (2021), and the methodology of the HYSPLIT analysis for these data. Then, the collection methods of precipitation and snowpack samples from near Inuvik between 1985-95 by C.R. Burn are outlined. This is followed by a brief section on the sampling of permafrost near Illisarvik. The study sites at Illisarvik are assessed, with detailed geometry of the ice-wedge surfaces and troughs. The specific procedure for the sampling of these ice wedges and the permafrost boreholes is explained. Lastly, a brief review of the laboratory work is presented.

4.2 Precipitation

4.2.1 2015-2018

Fritz et al. (2021) collected 134 precipitation samples at the Aurora Research Institute (ARI; 191 Mackenzie Road, Inuvik, NT, Canada) between August 2015 and August 2018 (Table 4.1). Precipitation was collected during each event using a gauge constructed with a PVC funnel of 30 cm upper diameter, above a 10 cm diameter vertical PVC pipe. A 250 mL HDPE bottle directly below the funnel captured the precipitation. The gauge was emptied into 15- or 30-mL LDPE bottles and stored in a refrigerator at 4°C.

Table 4.1: Summary of precipitation events sampled by month and year (Fritz et al., 2022)

Month	Year				Monthly
	2015	2016	2017	2018	
Jan	-	1	3	1	5
Feb	-	0	5	2	7
Mar	-	1	2	1	4
Apr	-	5	1	2	8
May	-	0	2	6	8
Jun	-	3	2	3	8
Jul	-	4	3	7	14
Aug	4	11	8	14	37
Sep	12	6	7	-	25
Oct	4	1	1	-	6
Nov	2	4	0	-	6
Dec	2	4	0	-	6
Yearly	24	40	34	34	

Specific information regarding other storage considerations are presented by Fritz et al. (2022). Samples were analyzed for δD and $\delta^{18}O$ at the Alfred Wegener Institute in Potsdam, Germany. Fritz et al. (2021) posted the δD and $\delta^{18}O$ on the PANGAEA database (<https://doi.pangaea.de/10.1594/PANGAEA.935027>). Daily weather data from the Inuvik Airport, i.e., the amount of precipitation and the mean temperatures on the days of rainfall or snowfall, were downloaded for analysis with the isotopic values (ECCC, 2023).

4.2.2 NOAA *Hysplit*

The NOAA Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPPLIT) model is a tool developed in 1982 to trace and model the movement of atmospheric particles (Stein et al., 2015). It was originally an improvement on the method being used to trace radioactive particles from Soviet nuclear test sites.

Parameters

The HYSPPLIT model was run for samples that could be confidently assigned to a specific precipitation event. Single back-trajectories from Inuvik were generated with the ‘normal’ trajectory option. This method allows for back-trajectories to be calculated for 315-hours (13 days), leading to the likely source location of the air mass. The National Weather Service’s National Centers for Environmental Prediction (NCEP) Global Data Assimilation System (GDAS, 0.5° resolution, data from September 2007 – June 2019) was the chosen meteorological grid model. Back-trajectories were calculated at three different levels in the atmosphere (250, 500, and 1000 m above ground level). The trajectories in these levels all commonly originated from the same location, but if they did not, the location with the majority was chosen. Should all three levels indicate a different source location, the sample was excluded from consideration. This occurred in 19 cases.

The built in cluster analysis function was run on all 558 precipitation events at Inuvik airport with more than 0.1 mm of precipitation between August 1st, 2015, and August 31st, 2018. This analysis sorts the storm trajectories into 'clusters' based on origin. The cluster analysis utilized reanalysis data as the meteorological grid.

Data Considerations

The data collected by Fritz et al. (2021) were the basis for the precipitation analysis conducted to improve understanding of the isotopic composition of precipitation in the western Canadian Arctic. There are errors in the data set associated with typographical inconsistencies or dates that had no precipitation. These samples were excluded from this analysis as no storm could be assigned to the data with confidence. In addition, there were many days with precipitation that did not have a sample collected. It is unknown whether precipitation from these storms was mixed with samples collected after later storms. Fritz et al. (2022) state that the gauge was emptied after every rain or snowfall event, but, for example, there were 10 days with precipitation in the last 11 days of November 2015 and no samples were collected during this time.

Due to such considerations, only 111 of the 134 samples collected by Fritz et al. (2021) were used for this analysis. One sample was excluded by Fritz et al. (2022) due to an exceedingly low D-excess value; two samples were omitted due to what are assumed to be labelling errors; nineteen were excluded due to inconclusive results from the HYSPLIT model (3), or an inability to confidently identify the storm responsible for the precipitation within the gauge (16). As there was only one sample originating from the east, it was not considered due to lack of corroborating data. The 111 samples collected represent 47% of the precipitation during the sampling period (Table 4.2).

Table 4.2: Total precipitation (t, mm) and in the events sampled by Fritz et al. (2021) (R, mm). Percent indicates the proportion of precipitation represented in the samples collected by Fritz et al. (2021).

Period	n	R	t	Percent
Aug-Dec 2015	21	71.7	168.9	43%
Jan-Dec 2016	34	100.2	203.5	49%
Jan-Dec 2017	26	113.4	278.3	41%
Jan-Aug 2018	30	113.8	200.8	57%
Total	111	399.3	851.5	47%

4.2.3 1985-95

Precipitation at Inuvik Airport, 12 km SE of Inuvik, was collected monthly from December 1985 to March 1986 and again between December 1987 and May 1995, by staff of the Atmospheric Environment Service, Environment Canada. Precipitation received at the rain or snow gauge was stored daily in a sealed Nalgene bottle, thus mixing all monthly precipitation. At the end of each month, samples of this quantity were transferred into sealed vials and sent to the University of Waterloo Isotope Laboratory (Dec 1985 – Feb 1989) or the University of Western Ontario Department of Geology (Mar 1989 – May 1995) for stable isotope analysis. This process was completed monthly throughout the sampling period.

The data processed by the University of Waterloo were released by the laboratory to the Global Network of Isotopes in Precipitation (GNIP) and were available to Fritz et al. (2022) for comparative analysis of co-isotope ratios in precipitation at Inuvik in 1985-1989. The data generated at the University of Western Ontario are presented in the Appendix.

4.2.4 *Snowpack data*

Samples from the snowpack near Inuvik were taken in April between 1991 and 1995. The samples were collected by C.R. Burn and consisted of an entire column of the snowpack, ensuring full representation of its isotopic composition. The samples were collected in an aluminum Mt. Rose snow sampler, normally used to determine snow water equivalent. The contents of the sampling cylinder were emptied into a plastic bag and sealed in the field. Upon return to the ARI, the snow in the bags was thawed at room temperature and a 125 ml sample of the melt was sealed in a Nalgene bottle. This sample was then sent for $\delta^{18}\text{O}$ analysis.

Samples were collected from various lakes in the Mackenzie Delta and nearby forests repetitively over five years. This allows for the representation of windswept

environments similar to that of modern-day Richards Island as well as forested environments, much like that of mid-Holocene Richards Island.

4.3 Permafrost near Illisarvik

In August 2022, samples of near-surface permafrost were collected adjacent to the Illisarvik drained lake basin.

4.3.1 Sampling Procedure

Once the sampling location along each ice wedge was determined, a two-foot wide trench was dug across the site (Fig. 4.1). The trench was dug from ridge to ridge, and digging ceased once the frozen layer was reached to allow for accurate measurements of thaw depth.

Once digging was complete, an arbitrary 'middle' point was identified within the trench. From there, every 30 cm outwards to the left and right were marked and had depth recorded to the surface. These were the drill hole locations for each trench. In some cases, drilling revealed the trench was too short, so it was extended and new holes were marked. If a hole was drilled past the edge of the wedge, a new point was positioned half way between the last hole with ice and the first without.

Drilling was with a Stihl BT-360 gas powered auger using a 50-mm-diameter CRREL core barrel (Fig. 4.2). Each site typically required two intervals of drilling. The first was through the frozen soil at the base of the trench and roughly 10 cm into the wedge. The core was then removed, the frozen soil measured, and discarded. To ensure minimal cross-contamination, the core barrel was cleaned of all soil and ice, and any excess material within the hole was removed and discarded from the site.



Figure 4.1: The trench dug at a wedge after drilling was completed. The northern slope of the Illisarvik drained lake basin is visible in the background.



Figure 4.2: Drilling at T2 with field assistant Aidan Gill (right).

The second interval of drilling typically contained the final sample and was drilled to a depth of 25 cm in the wedge. The ice core was extracted and placed on a plastic sheet, and all soil was removed from the potential sample (Fig. 4.3). When deciding which ice to sample, priority was given to depth, then to integrity of the sample (large pieces were prioritized over small shards), and lastly pureness of the sample (soil particles within the ice). Samples were then double bagged using 120 ml Whirl-Pack sample bags and returned to camp, where they were melted and poured into 20 ml vials. Duplicate samples were collected if there was sufficient meltwater. Each vial was sealed with plastic tape to minimize evaporation or contamination.

All samples arrived at the laboratory without sign of damage or contamination. The samples were stored at 4°C in the Water and Ice Research Lab at Carleton University. Contamination during storage is a concern and can impact the isotopic composition of the samples. Spangenberg (2012) showed that, over 659 days, samples stored in bottles similar to those used in this thesis showed minimal changes in isotopic composition ($< 1.0\text{‰}$) and no trend over time. The samples used for this thesis were stored for less than 4 months (< 120 days), so any concern regarding contamination during storage is limited.

4.3.2 $\delta^{18}\text{O}$ and δD Analysis

Analysis of the samples took place in the Carleton University Water and Ice Research Lab (WIRL), utilizing a Los Gatos Research TL-WIA liquid water isotope analyzer (LGR analyzer) to determine $\delta^{18}\text{O}$ and δD values. Analysis began the week after returning from the field. Each sample was transferred from the original vial through a 0.45 μm filter into a new 20 mL vial.



Figure 4.3: Example of a sample after cleaning and removing any soil and excess ice but before bagging was completed.

A new syringe and filter were used for each sample. The samples were then transferred into a 5 mL vial using a pipette to be inserted into the LGR analyzer. The LGR analyzer was calibrated using a set of standardized samples with a known isotopic composition. The average sample precision for $\delta^{18}\text{O}$ and δD was 0.2‰ and 2‰, respectively, based on four measured injections per sample and a total of 111 samples (including duplicates). The D-excess for each sample of near-surface permafrost and wedge ice was calculated using eq. 2.10 ($\delta D - 8\delta^{18}\text{O}$).

4.3.3 *Permafrost Boreholes*

Two boreholes were drilled on the western side of the Illisarvik lake basin to sample the permafrost sediments in which the anti-syngenetic ice wedges had developed (Fig. 4.4). The frost table at borehole A was at 50 cm, and samples were collected to a depth of 290 cm. A total of 15 samples were collected. Borehole B had a frost table depth of 80 cm and had 20 samples collected to 320 cm below the ground surface. The isotopic concentrations of the samples were derived from the thawed excess ice in the sediments.

4.3.4 *Anti-Syngenetic Ice Wedges*

Six anti-syngenetic ice wedges were sampled at a depth of 25 cm at 30 cm intervals horizontally across each wedge. Two study locations were selected, one on the east side of the Illisarvik basin, and one on the west (Fig. 4.4). Three wedges in close proximity to each other were sampled at each location. The wedges were selected based on their relation to the slope and definition of the trough. At each location one wedge ran parallel with the contour, and two were perpendicular to the contour. On the western side of the basin, the two wedges perpendicular to the contour ran NW-SE (T3, T4) while the wedge running parallel was oriented NE-SW (T2).

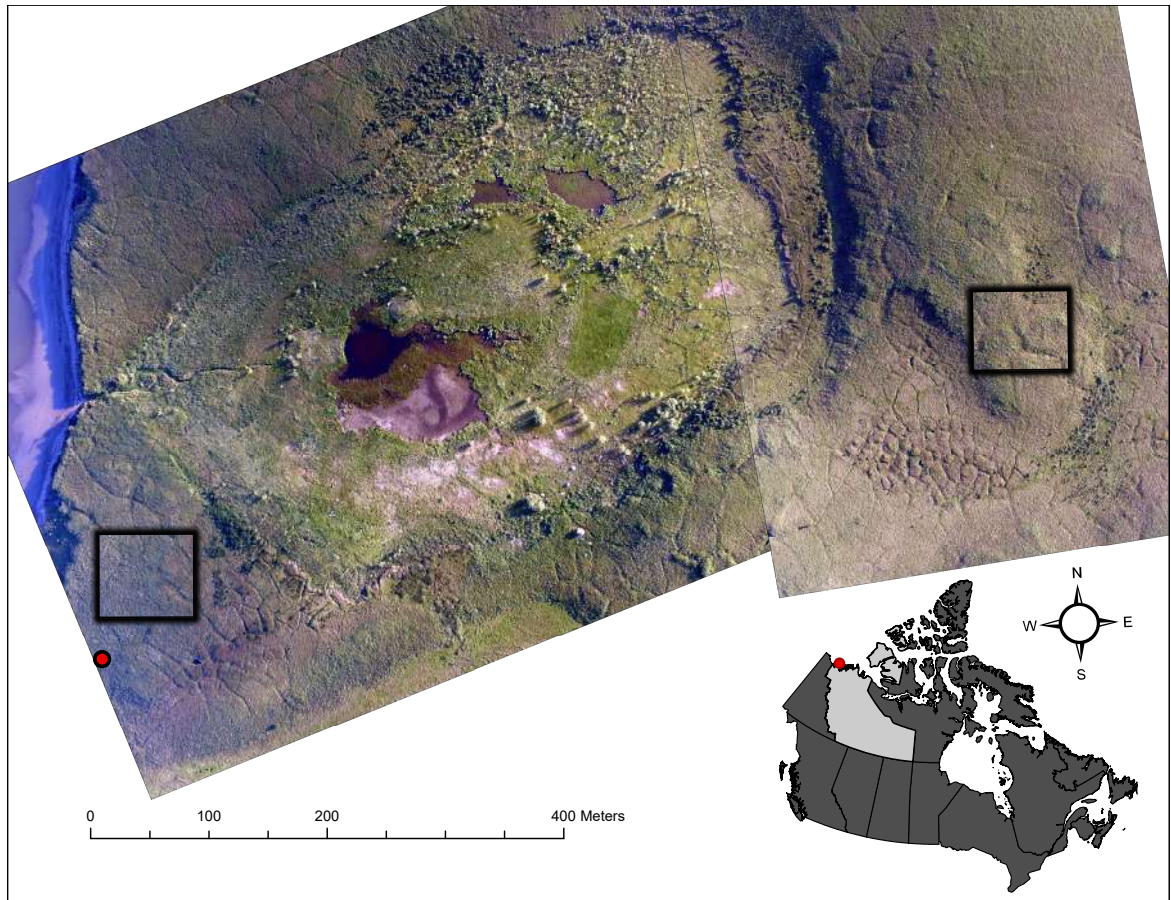


Figure 4.4: Aerial image of Illisarvik drained lake basin. The location of the sites on the eastern and western sides are illustrated by squares. The location of boreholes are denoted by the red dot adjacent to the western site. The image was taken by A. G. Lewkowicz using a UAV in 2016.

On the eastern side, one wedge ran NE-SW (E1) while the remaining two ran N-S (E2, E3). The changing aspect of the slope in this area meant that E1 and E3 ran perpendicular to the contours, while E2 ran parallel.

Elevation surveys were completed at each site to gather information on the ground surface, thaw depth, and ice-wedge surface. Ground surface elevations were determined with a Wild NA2 level with an optical micrometer and an invar stadia rod. The precision of the leveling is better than 1 cm (Mackay, 2000; Burn et al., 2021). Additional measurements were conducted during the drilling operations.

T2, T3, and T4 surveys were conducted relative to a common reference location, enabling direct comparisons among them. While E1 and E2 were executed using the same reference point, the topographical position of E3 was situated too far upslope to permit surveying from that reference point.

The soil directly above each wedge was primarily organic material with interspersed pockets of fine-grained mineral soil. However, at or near the edge of the wedges there were sharp transitions to fine-grained soil (Figure 4.5). The only exception to this was E2, where the downslope edge of the wedge was extended past this transition (Figure 4.5).

Throughout the remainder of this thesis when discussing the anti-syngenetic ice wedges that were studied, the term 'ridge' will be used to broadly define the small slope from the trough to the surrounding ground level. The traditional 'ridge' shape that is present with developed epigenetic wedges was not apparent at these anti-syngenetic wedges. As a result, the ground surface was level with the polygons adjacent to the trough at the top of the 'ridges'.



Figure 4.5: Downslope edge of E2. The hole on the right was at the edge of the wedge with the light colour indicating ice along the wall of the drill hole, while the hole on the left did not encounter wedge ice.

4.3.5 *Western Sites*

The wedges on the western side of the basin were located about 300 meters south of the gulley created by lake drainage in 1978 (Mackay, 1997) (Fig. 4.6). The three wedges were near the top of a hillslope facing SW and with a gentle gradient of $\sim 5^\circ$.

The western wedges were narrow with well-defined troughs and ridges. Basic measurements can be found in Table 4.3. On average, each ridge rose 20 cm from the edge of the trough, at a mean gradient of 22° . The vegetation in this area consists primarily of ground birch and other small shrubs, with none reaching above knee height. Sampling began on the western side of the basin with the wedge T1 (not included in this thesis) used for training in drilling. Thus, the western wedges begin with T2.

Notably along T2, the edge of the upslope (easternmost) portion of the wedge extended past the ridge, while the downslope side ended within the trough. This was the only wedge featuring this configuration. Within the trough, the surface of the ice wedge changed very little, only fluctuating within a range of 3 cm (Fig. 4.7). The upslope side of the ice wedge ended abruptly, with seemingly no tapering. Three samples at this edge were taken up to 33 cm below the top of the wedge but drilling 7.5 cm further upslope yielded no ice. In contrast, the final drill hole on the downslope side of the wedge only contained 13 cm of ice, indicating that the edge tapered off as expected (Fig. 4.7). The upslope and downslope ridges rose with a 23° and 17° gradient, respectively. T3 was just upslope of T2 and the two wedges connected downslope from the T3 trench (Figure 4.6). The southernmost ridge had a gradient of 13° while the northernmost rose at 26° . T4 was across the same wedge that has existing benchmarks further downslope used to survey ground deformation since 1996 (Burn et al., 2021). The trench at T4 was dug 200 m upslope to ensure minimal disturbance to the existing benchmarks.

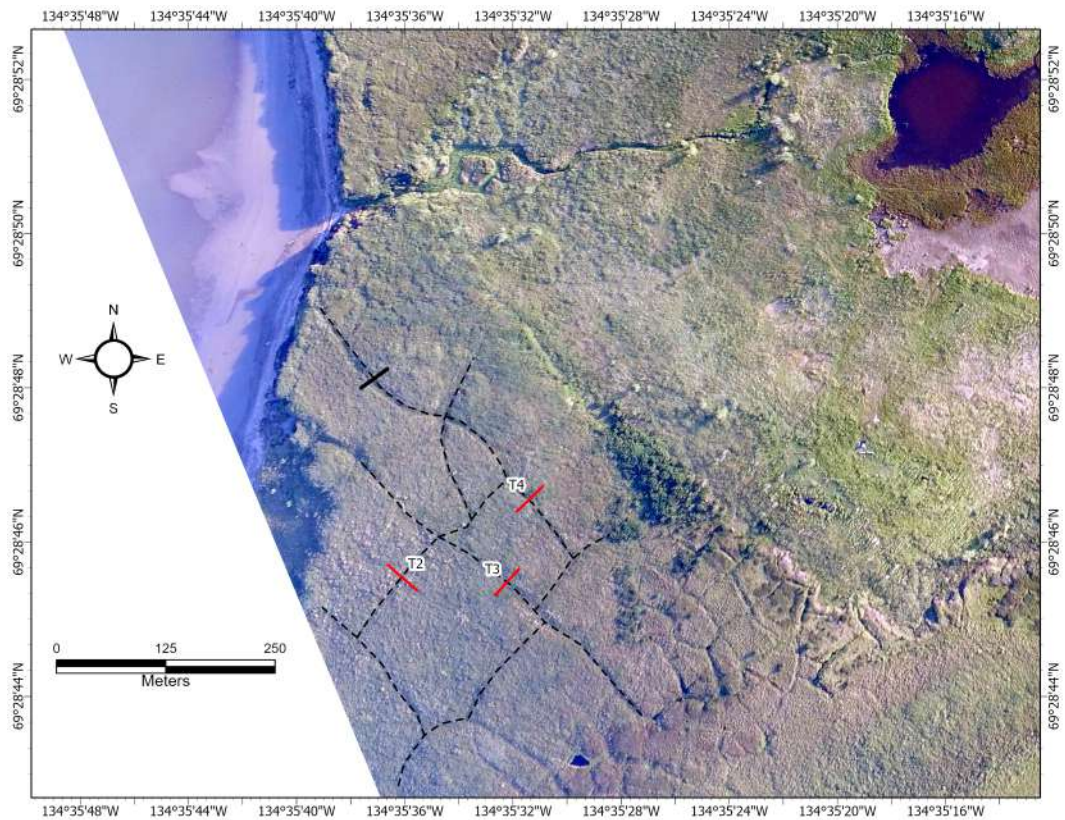


Figure 4.6: Location of trenches dug across the three ice wedges on the western side of the lake basin. Dotted black lines indicate the anti-syngenetic ice wedges in the area. Solid black line represents location of benchmarks B of Burn et al. (2021).

Table 4.3: Measurements and characteristics of the sampled hillslope ice wedge troughs. Orientation is denoted by + (perpendicular) or || (parallel). Trough depth represents the deepest point in the trough to the highest point on the ridge. Wedge depth represents the average depth to the top of the ice wedge. All values are averaged for the summary rows.

Wedge	Orientation	Width (cm)	Trough Depth (cm)	Wedge Depth (cm)
Western				
T2	+	172	28	49
T3		240	35	56
T4		330	40	57
Mean		247	34	56
Eastern				
E1		580	43	45
E2	+	615	67	53
E3		390	31	54
Mean		523	47	49

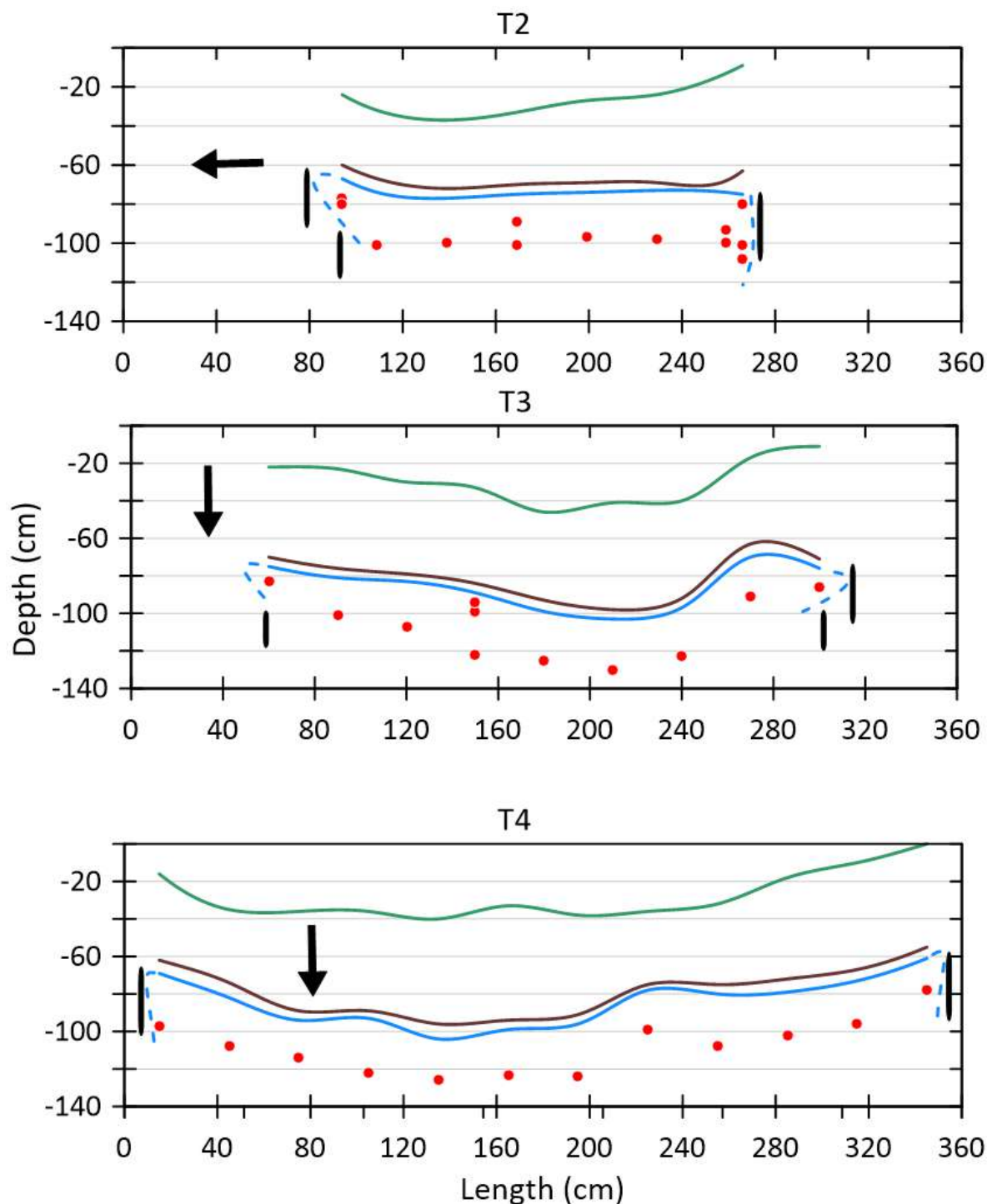


Figure 4.7: Survey profiles of the western ice wedges. The green line represents the ground surface, brown represents the thaw depth on the date of drilling, and blue represents the top of the ice wedge. Dashed blue line indicates the edge of the ice wedge, as estimated from drilling. Red dots represent locations of samples. Black lines represent drill holes that failed to encounter wedge ice. Black arrows indicate slope direction. The scales are equal for all three wedges, and the y-axis is in reference to the highest point on the ground surface that was surveyed.

4.3.6 *Eastern Sites*

The sites on the eastern side of the basin were located near the top of a hill (Fig. 4.8, 4.9) with a gradient of $\sim 5^\circ$. Vegetation on this side of the basin consisted of larger willows, dwarf birch, and alders.

In contrast to the western side of the basin, hummocks occurred within the polygons and troughs. The southernmost ridge of E1 was the steepest of all ridges surveyed, being 34° , while the northernmost ridge had a gradient of 22° . The northern ridge reached 10 cm higher than the highest point on the southern ridge. A hole was dug further downslope along E1 but filled with water immediately. This was a common observation with downslope holes. Both of the ridges along E2 rose at similar gradients, the upslope ridge at 21° and the downslope at 23° . Almost directly in the centre of the wedge there was a sudden rise in the frost table (Fig. 4.10) that was not apparent in the surface or ice-wedge profiles. E3 was located on a different hillslope than E1 and E2. Where E1 and E2 were on a hillslope facing NE, E3 was on a slope facing SE. Notably, this wedge had the largest difference in gradient between ridges, as the eastern ridge only rose at 8° while the western ridge rose at 28° .



Figure 4.8: E1 after refilling of the trough. Note the alders further downslope and small willows in the foreground.

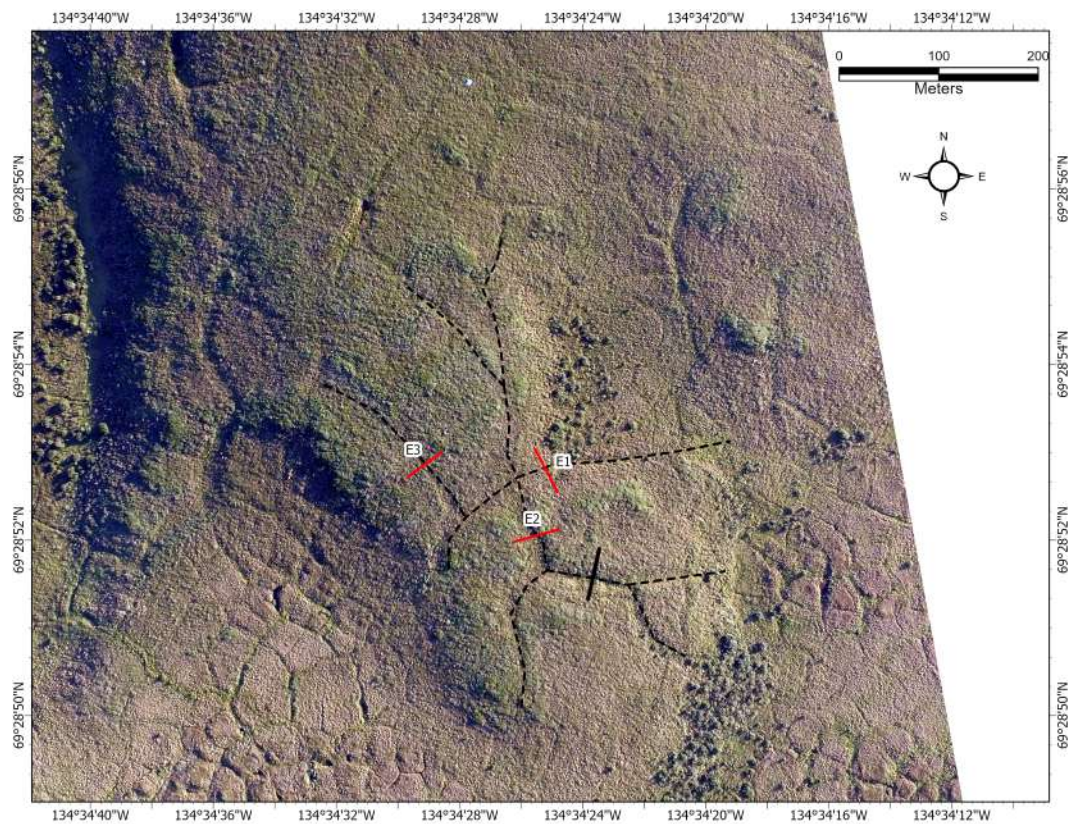


Figure 4.9: Location of trenches dug across the three ice wedges on the eastern side of the lake basin. Dotted black lines indicate the anti-syngenetic ice wedges in the area. Solid black line represents location of benchmarks A of Burn et al. (2021).

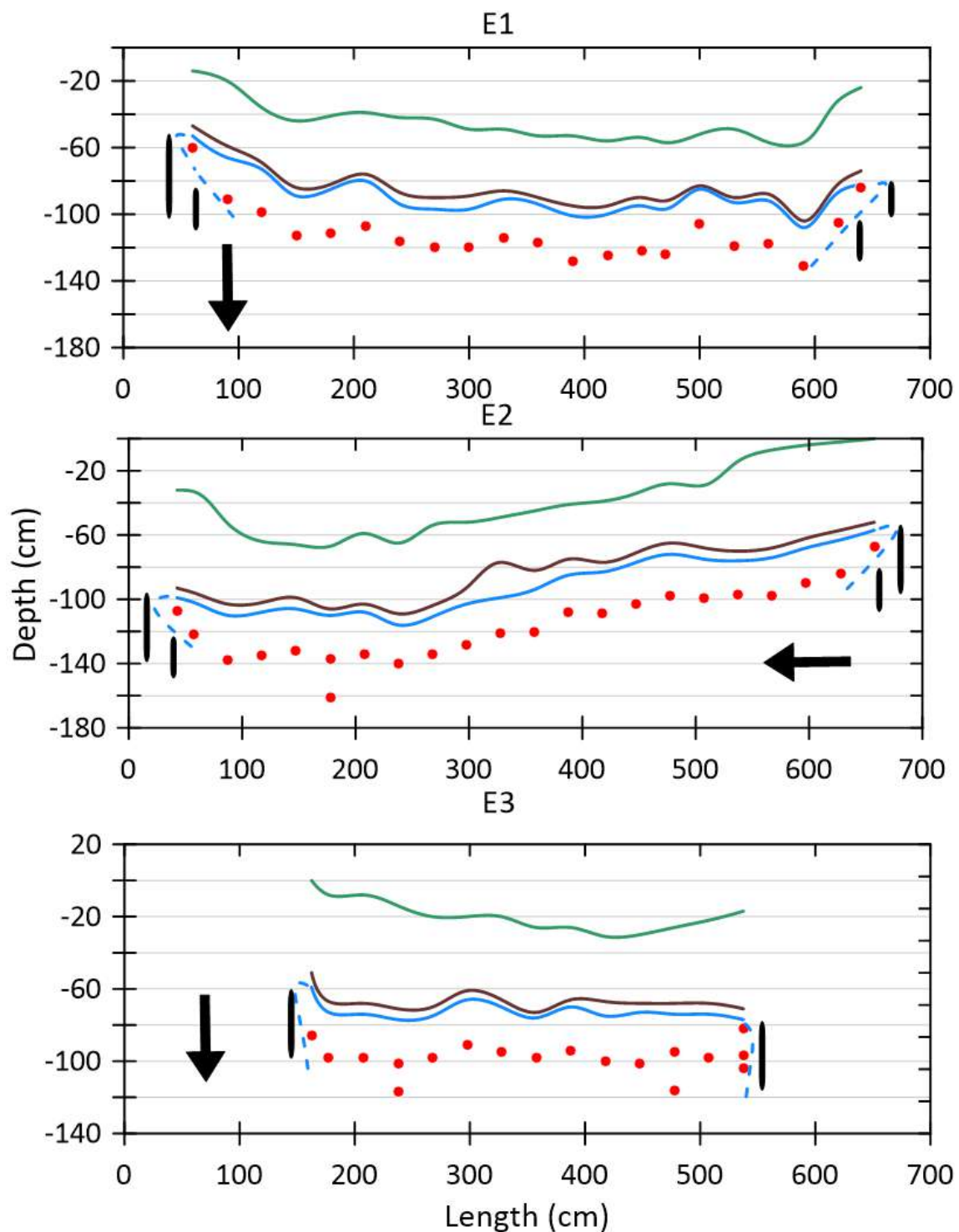


Figure 4.10: Cross-sectional profiles of the Eastern ice wedges. The green lines represent the ground surface, brown the thaw depth on the date of drilling, and blue the top of the ice wedge. The dashed blue lines indicate the edge of the ice wedge, as estimated from drilling. Red dots represent locations of samples. Black lines represent locations of drilling that failed to encounter wedge ice. Black arrows indicate slope direction. The scales are equal for E1 and E2, but E3 has a different y-axis.

5 Results

5.1 Introduction

The first section of this chapter examines the isotopic content of precipitation and snow pits from near Inuvik. The isotopic content of precipitation is assessed in relation to its origin and path prior to rain or snowfall. Then, the isotopic composition of forest- and lake-based snowpacks are analysed. Following this, samples of permafrost from terrain adjacent to the Illisarvik drained lake basin are examined. Two boreholes from the western edge of the basin are analysed to give better context to the six anti-syngenetic ice wedges sampled from two sides of the basin. This analysis shows the importance of understanding the original source of precipitation and the variation within the anti-syngenetic ice wedges near Illisarvik.

5.2 Precipitation at Inuvik

5.2.1 *Storm Origin and Isotopic Concentration*

The precipitation data collected by Fritz et al. (2021) were sorted into the categories of storms from the North, West or South depending on the HYSPLIT analyses. Storms from the North originated in the Arctic Ocean, from the South in the Pacific Ocean, and from the West in the Bering Sea, Sea of Okhotsk, or Chukchi Sea. The data were further classified into Winter (October – April) or Summer (May – September) storms. Table 5.1 presents the distribution of precipitation events by origin.

Table 5.1: Statistics for precipitation data at Inuvik. Total n is total number of events with more than 0.1 mm of precipitation in August 2015 - August 2018. Mean values of $\delta^{18}\text{O}$ (‰), D-excess (‰), and temperature (°C) in summer (May – September), and winter (October – April) are presented with the standard error of the mean (SE) for both $\delta^{18}\text{O}$ and T. * Data for $\delta^{18}\text{O}$ from Fritz et al. (2021) represent individual storm events in the study period. Temperature for day of event is given. ** Data for the monthly means of $\delta^{18}\text{O}$ in precipitation collected in 1985-95 (C.R. Burn).

	Sampled n	Total n	$\delta^{18}\text{O}$	D-excess	T	$\delta^{18}\text{O}$ (SE)	T (SE)
North*	74	264	-19.0	4.8	1.1	0.6	1.5
Winter	20	137	-24.3	8.1	-17.6	0.7	1.0
Summer	54	127	-16.7	3.6	5.8	0.5	0.9
South*	14	115	-21.0	5.6	0.9	1.3	2.5
Winter	6	84	-25.2	9.1	-7.1	1.0	2.2
Summer	8	31	-17.8	3.0	6.8	1.2	2.6
West*	23	59	-20.5	4.6	-1.4	0.8	3.0
Winter	11	36	-23.3	9.6	-14.4	1.0	2.5
Summer	12	23	-18.0	0.1	10.4	0.8	1.5
Total*	111	438	-19.6	5.0	-1.0	0.5	1.2
Winter	37	257	-25.1	8.7	-15.9	0.6	1.4
Summer	74	181	-17.5	3	6.7	0.3	0.6
1985-95**	91		-24.8	-	-9.5	0.6	1.7
Winter	57		-28.2	-	-20.8	0.6	1.4
Summer	34		-19.2	-	8.6	0.7	1.4

The proportions of events sampled by Fritz et al. (2021) were 66, 21, and 12% from the North, West, and South respectively. For all events in the same period with ≥ 0.1 mm of precipitation the proportions were 60, 26, and 13% (Table 5.1). The remaining 1% came from the East. The distribution of samples collected by Fritz et al. (2021) is representative of the broader population at Inuvik during the sampling period.

When separated by season and direction, there were only small differences between mean $\delta^{18}\text{O}$ concentrations of the samples (Table 5.1). Figure 5.1a shows a box-and-whisker plot of $\delta^{18}\text{O}$ values grouped by direction of origin and indicates the similarity of distributions within the seasonal groups. Student's t-tests for differences between these various groups (North, West, South) by season (all year, summer, winter) did not show any statistically significant ($\alpha = 0.05$) differences between the mean values of the groups.

The meteoric water lines for the precipitation at Inuvik between 2015-18, separated by direction and season, can be found in Table 5.2. There is a decrease in slope between the least-squares linear (LSL) and functional fits (FF). All relations were significant at $\alpha = 0.05$ and have R^2 values greater than 0.95. The slope of the functional fit for western storms year-round was statistically significantly different from northern and southern storms. Winter storms had a higher FF regression coefficient than summer storms, but the two seasons were not statistically different.

The difference between the calculated MWL for all samples and the LMWL calculated by Fritz et al. (2022) ($7.39\delta^{18}\text{O} - 6.7$) is due to the calculation methodology. Fritz et al. (2022) derived their MWL as a LSL of the monthly averages from their samples, whereas the values in Table 5.2 were calculated by using individual samples. The primary reason for this is inadequate data when sorted by season and direction to accurately represent monthly conditions. Holland et al. (2020) discuss the importance of using monthly, or even yearly aggregate precipitation samples to calculate a long-term MWL due to the noise created by individual storm events.

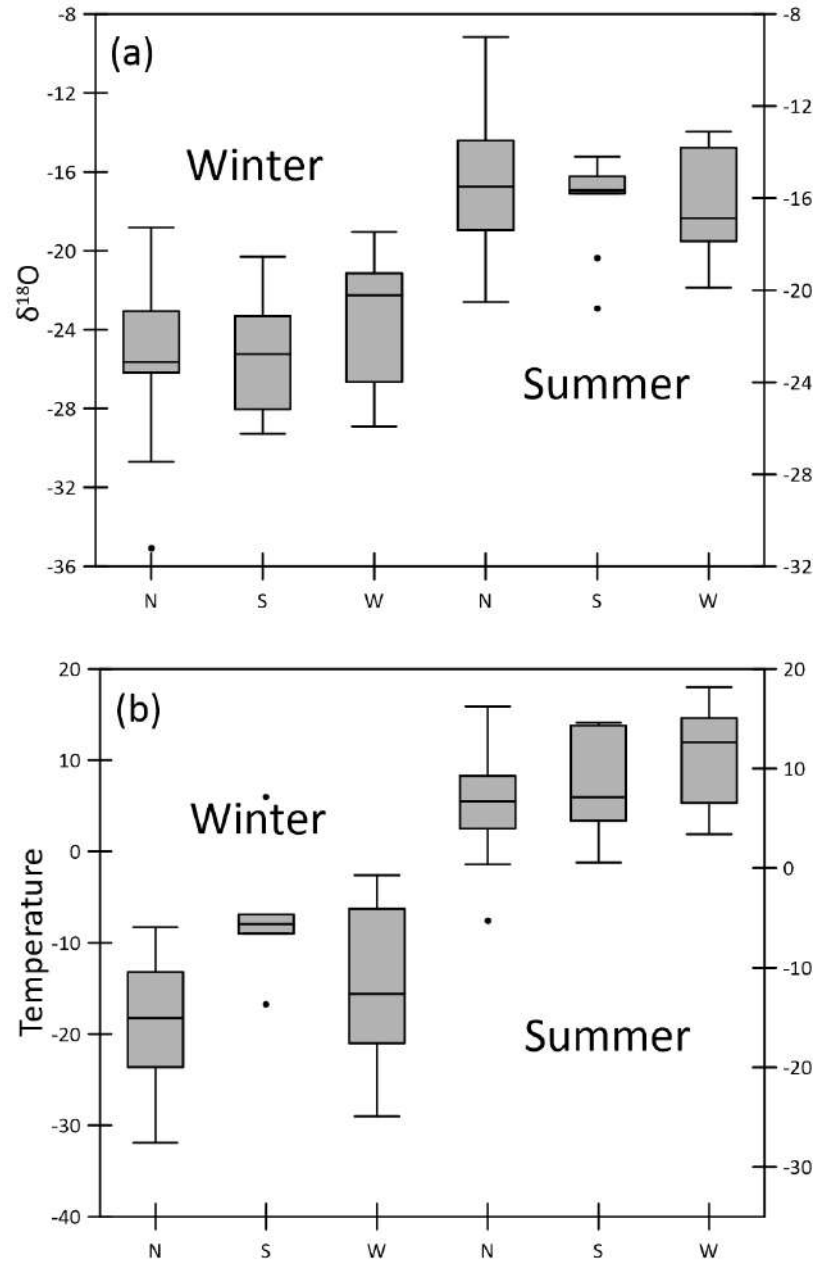


Figure 5.1: (a) Box-and-whisker plots of $\delta^{18}\text{O}$ grouped by direction of origin and season; (b) Mean temperature on day of precipitation grouped by origin and season.

Table 5.2: LSL and FF regression lines for δD on $\delta^{18}O$ from samples taken between 2015 and 2018 separated by direction and season. Data from Fritz et al. (2021).

Origin	LSL	FF
North	$7.31\delta^{18}O - 8.4$	$7.36\delta^{18}O - 7.32$
Winter	$7.29\delta^{18}O - 10$	$7.37\delta^{18}O - 7.87$
Summer	$6.94\delta^{18}O - 14$	$7.11\delta^{18}O - 11.21$
South	$7.29\delta^{18}O - 9.4$	$7.32\delta^{18}O - 8.75$
Winter	$7.42\delta^{18}O - 5.6$	$7.45\delta^{18}O - 4.92$
Summer	$7.53\delta^{18}O - 5.2$	$7.64\delta^{18}O - 3.29$
West	$6.70\delta^{18}O - 22.1$	$6.76\delta^{18}O - 20.73$
Winter	$7.40\delta^{18}O - 4.5$	$7.43\delta^{18}O - 3.8$
Summer	$6.71\delta^{18}O - 23.1$	$6.87\delta^{18}O - 20.2$
All Directions	$7.24\delta^{18}O - 10.1$	$7.29\delta^{18}O - 9.01$
Winter	$7.42\delta^{18}O - 5.6$	$7.47\delta^{18}O - 4.29$
Summer	$7.06\delta^{18}O - 13$	$7.24\delta^{18}O - 9.97$

The D-excess of samples ranged from -14.61‰ to 16.45 , similar to samples of precipitation in southeast Siberia (Kostrova et al., 2020). Mean D-excess values can be found in Table 5.1. The higher values during winter are characteristic of snow (Jouzel and Merlivat, 1984), and lower values during rainfall were to be expected (Stewart, 1975). There were no statistically significant ($\alpha = 0.05$) differences in D-excess between precipitation in each season when sorted by direction.

5.2.2 *Atmospheric Temperature and Stable Isotopic Concentration*

Storm by Storm

The distribution of temperature and $\delta^{18}\text{O}$ values for all storms can be seen in Fig. 5.2. Most storms (66%) originated from the North and were accompanied by the lowest mean daily temperatures during summer and winter but did not have the lowest $\delta^{18}\text{O}$ values in either season (Table 5.1). In fact, during the summer months, storms from the North had the highest $\delta^{18}\text{O}$ values.

Storms from the South represent 12% of the sample population, and their precipitation at Inuvik had relatively low $\delta^{18}\text{O}$ values (Table 5.1). In the winter, storms from the South had $\delta^{18}\text{O}$ concentrations 1‰ lower than northern storms but with, on average, mean daily temperatures 10°C higher (Table 5.1). The relationship was similar in the summer, but the temperature difference was 1°C . Western storms had the highest $\delta^{18}\text{O}$ values in winter, but the lowest in summer, despite having the warmest conditions in summer by nearly 4°C on average (Table 5.1).

Fig. 5.1b is a box-and-whisker plot illustrating the seasonal and directional distributions of temperature during sampled storm events. Considerable overlap exists between northern and western storms in the winter, and for all directions in the summer.

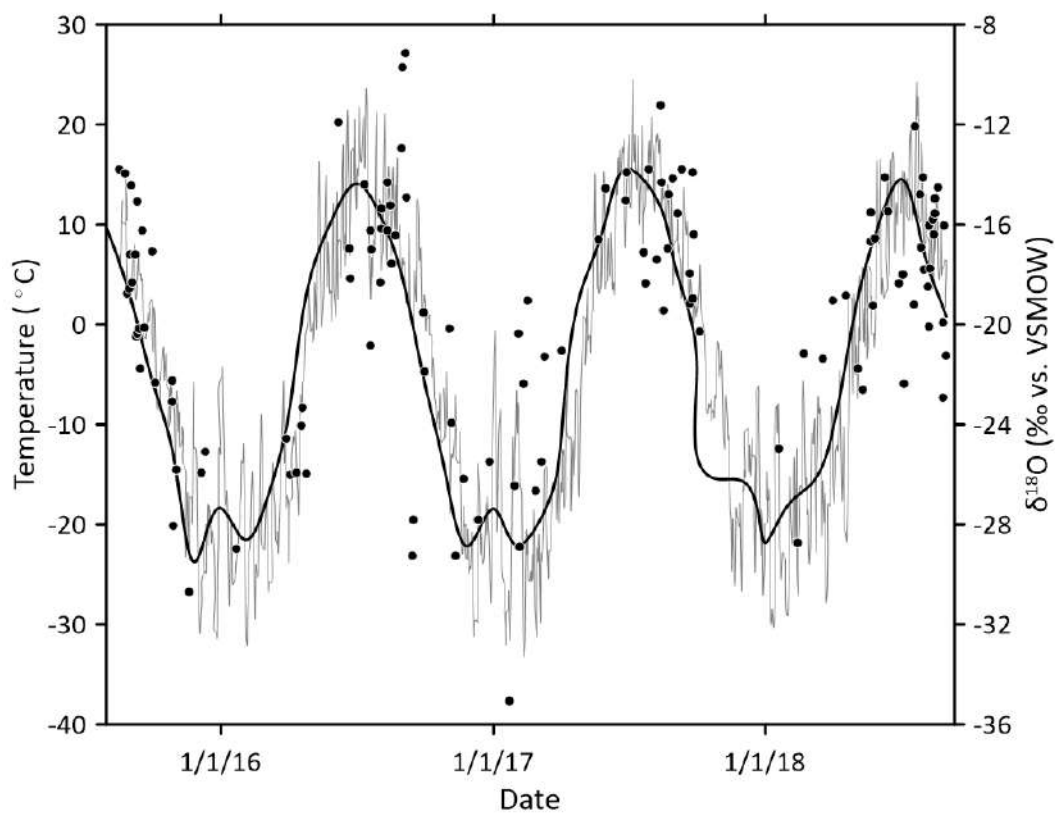


Figure 5.2: Time series of mean daily temperature (thin lines), mean monthly temperature (thick line), and $\delta^{18}\text{O}$ values during individual storm events in 2015-18 at Inuvik (Fritz et al., 2021).

A series of t-tests indicated a statistically significant difference ($\alpha = 0.05$) in winter storms between those from the South and those from the North and West, and similarly a difference between western storms and northerly and southerly storms in the summer. These differences suggest that the $\delta^{18}\text{O}:\text{T}$ relation varies by direction and season.

The relation between atmospheric temperature and stable-isotopic composition of precipitation can be seen in Fig. 5.3. The split between summer and winter storms is visually evident, as the two distinct groups were separated at 0°C . Table 5.3 contains the least-squares linear regression line fitted to each direction and season for these events. Least-squares linear regression can be used when deriving the $\delta^{18}\text{O}:\text{T}$ relation since measurements of T are precise. When applied to the entire dataset, there is a statistically significant slope with a coefficient of determination of 0.59. Similar R^2 values were calculated when sorted by direction (Table 5.3). There were relatively large variations in regression coefficients for these data, but they were not statistically different at $\alpha = 0.05$.

Seasonal Precipitation

Given the seasonality of ground-ice development (Mackay, 1983b, 1990), it is important to understand the seasonal differences within the isotopic composition of precipitation. Seasonal data from 2015-18 are presented in Fig. 5.4. While both lines were statistically significant at $\alpha = 0.05$, they present low coefficients of determination, which indicates that little variation in $\delta^{18}\text{O}$ is explained by T on a daily basis. By separating the storms by direction (Table 5.4) only northern and western storms in the summer have statistically significant slopes at $\alpha = 0.05$.

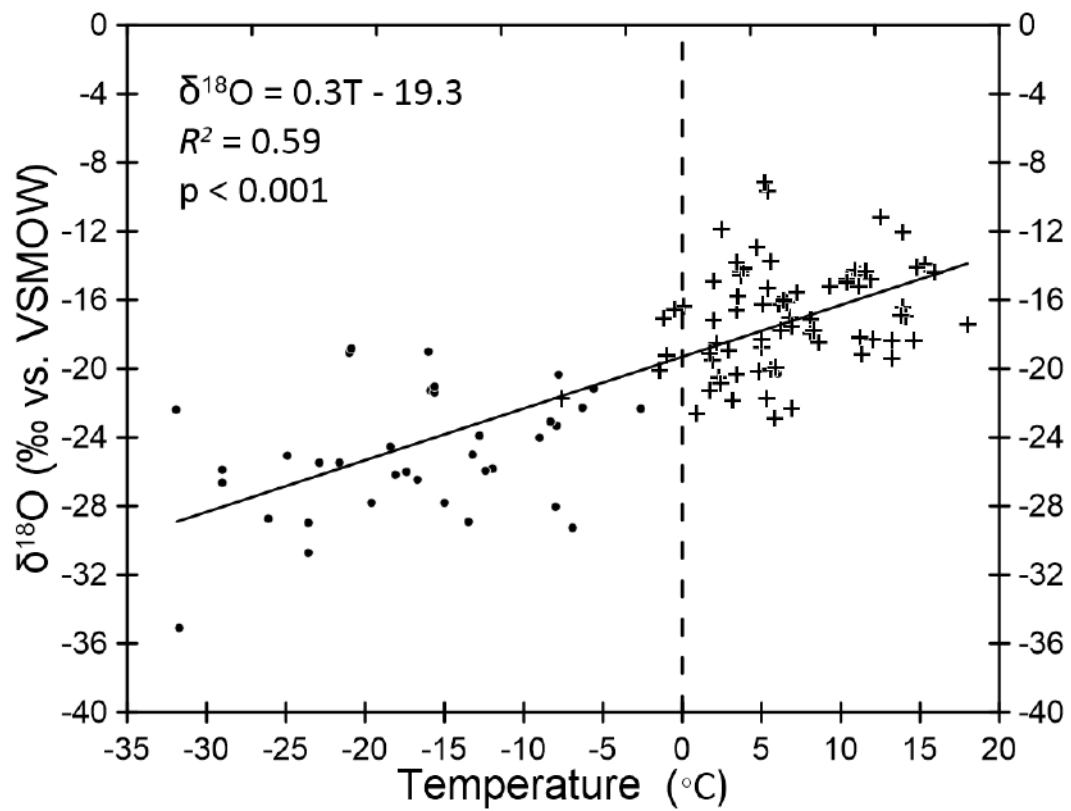


Figure 5.3: Temperature and $\delta^{18}\text{O}$ plotted for the storms from Fritz et al. (2021) analysed in this thesis. Circles represent winter storms, crosses are summer.

Table 5.3: Least-squares linear regression for $\delta^{18}\text{O}$ and temperature from samples taken between 2015 and 2018 separated by direction and season. Data for $\delta^{18}\text{O}$ from (Fritz et al., 2021).

Origin	n	LSL	R^2	p-value
North	74	0.33T - 18.7	0.64	<0.001
South	14	0.4T - 21.33	0.62	<0.001
West	23	0.22T - 20.21	0.60	<0.001
All Directions	111	0.3T - 19.33	0.59	<0.001
Summer	74	0.19T - 18.23	0.01	0.007
Winter	37	0.16T - 22.16	0.15	0.02

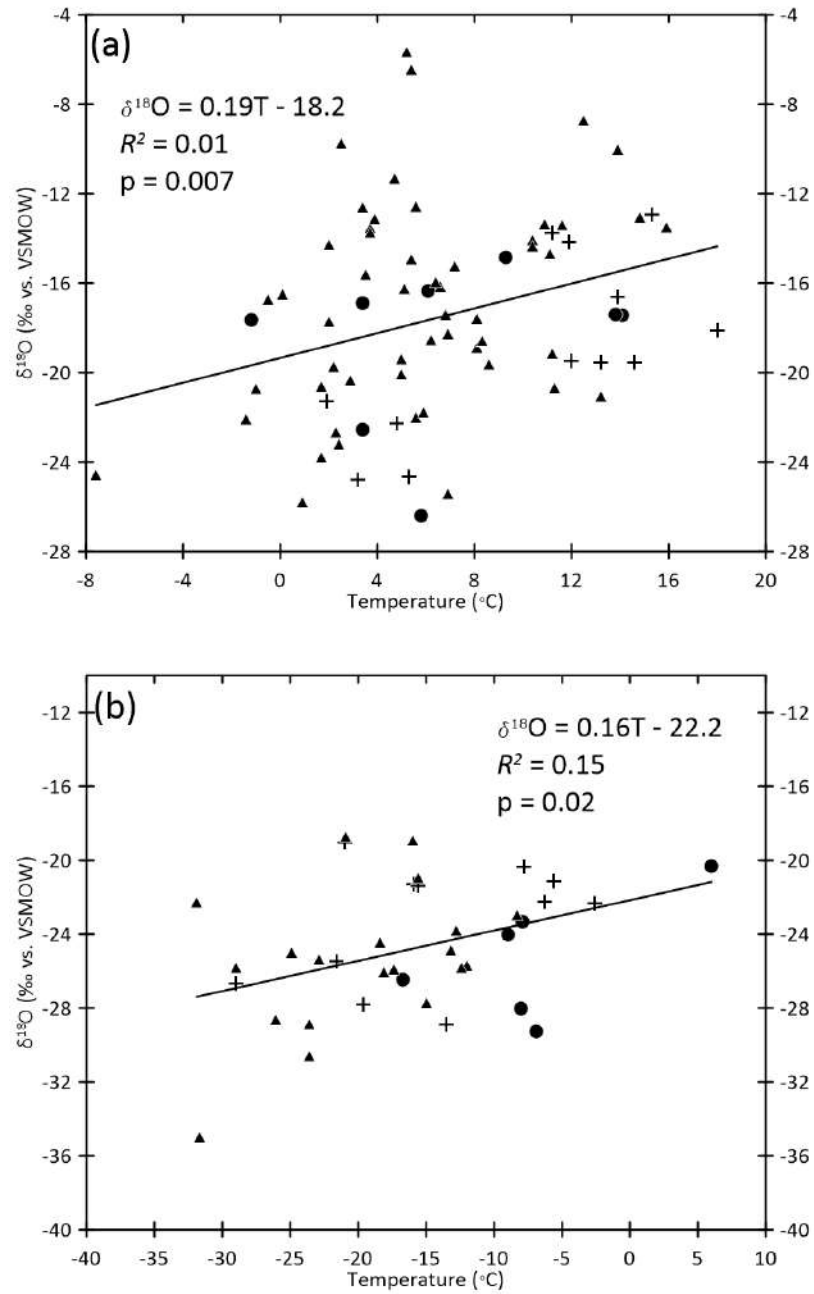


Figure 5.4: Scatterplots of temperature and $\delta^{18}\text{O}$ for (a) summer and (b) winter events in 2015–18, with northern (triangle), southern (circle), and western (+) origins. Each line has been fitted by least-squares linear regression.

Table 5.4: Least-squares linear regression lines for $\delta^{18}\text{O}$ on temperature from summer and winter samples taken in 2015–18 separated by direction. Data for $\delta^{18}\text{O}$ from Fritz et al. (2021).

Origin	n	LSL	R^2	p-value
Summer				
North	54	0.24T - 18.1	0.13	0.01
South	8	0.11T - 18.5	0.05	0.58
West	12	0.35T - 21.6	0.47	0.01
Winter				
North	20	0.22T - 21.2	0.15	0.09
South	6	0.28T - 23.3	0.37	0.2
West	11	0.16T - 21	0.16	0.22

There is substantial difference in winter temperatures and $\delta^{18}\text{O}$ concentrations between data from 1985-95 and 2015-18 (Table 5.1). However, these data cannot be compared statistically as collection methods differ. Weighted monthly values cannot accurately be calculated for the data from 2015-18 as precipitation was only collected from storms that represent 47% of the total amount during the sampling period. It is unknown at this stage whether the precipitation in the unrepresented storms is proportioned similarly to the sampled storms.

Monthly Aggregate Sampling

Year round least-squares linear regression lines for $\delta^{18}\text{O}:\text{T}$ relations in 1985-95 can be found in Table 5.5. The relatively high R^2 values compared to the storm-by-storm data may be a result of signal smoothing from the monthly aggregate collection method, given the temperatures were monthly means. The differences in summer and winter coefficients are illustrated in Fig. 5.5, as upon transitioning to summer the slope of the relation increases.

5.2.3 Physiographic Influences on Stable Isotopic Concentration

The impact of physiographic influences can be seen clearly when analyzing the origin of storms. The most notable influence was from mountainous regions, which, on average year-round, result in isotopic concentrations 2‰ lower than those that do not pass mountains (Table 5.6). There was, however, no significant difference ($\alpha = 0.05$) in the winter concentration of $\delta^{18}\text{O}$ between storms that travelled over the mountains and those that did not. In contrast, there was a significant difference ($\alpha = 0.05$) in the summer storms that crossed the mountains.

Table 5.5: Least-squares linear regression lines for $\delta^{18}O$ on temperature from all monthly aggregate samples collected in 1985-95 separated by season.

	n	LSL	R ²	p-value
All Data	91	0.31T - 21.9	0.71	<0.001
Summer	34	0.46T - 23	0.55	<0.001
Winter	57	0.26T - 23	0.28	<0.001

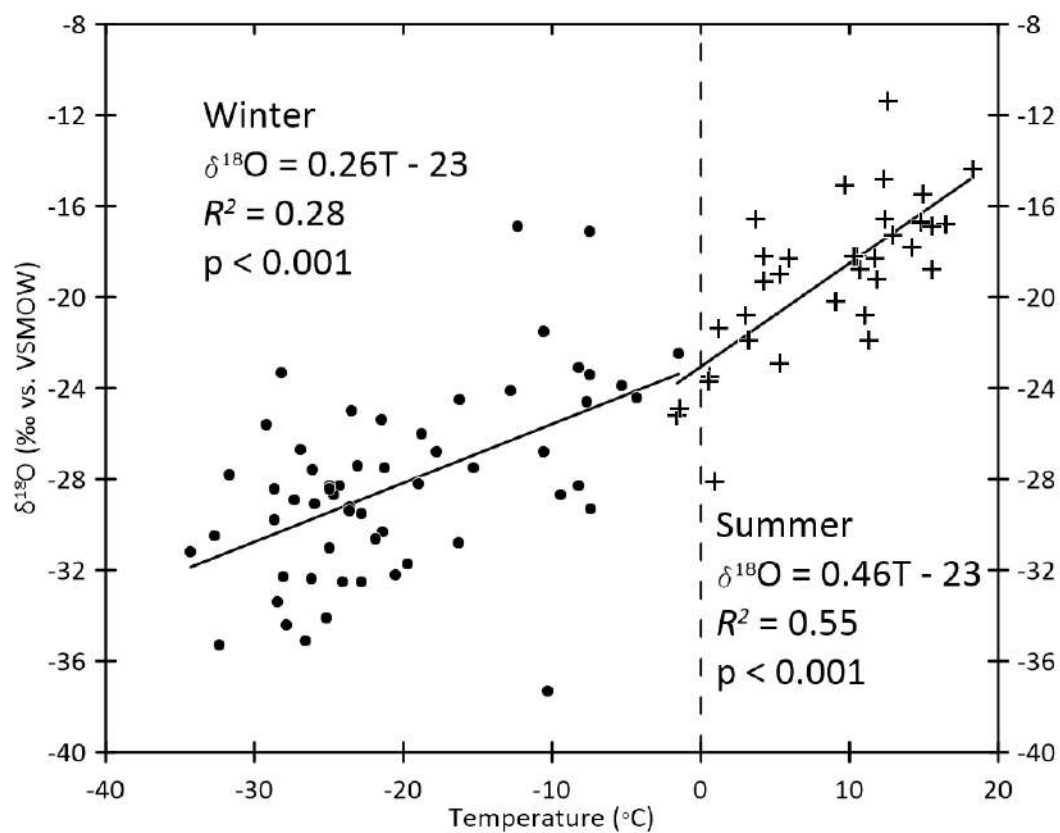


Figure 5.5: Scatterplot of temperature and $\delta^{18}\text{O}$ for monthly aggregate precipitation samples between 1985–95. Circles represent winter and crosses summer months.

Table 5.6: Mean values of $\delta^{18}\text{O}$ (‰), temperature ($^{\circ}\text{C}$), and D-excess (‰) of storms that passed over one or more mountain ranges and those that came directly from a water source. Data for $\delta^{18}\text{O}$ from Fritz et al. (2021). Also included is the standard error of the mean (SE) for both $\delta^{18}\text{O}$ and T.

	n	$\delta^{18}\text{O}$	T	D-excess	$\delta^{18}\text{O}$ (SE)	T (SE)
Winter	37					
Mountain	16	-25.44	-14.06	9.73	0.7	1.5
Direct	21	-24.12	-17.38	8.23	1.1	2.5
Summer	74					
Mountain	19	-17.55	9.34	0.39	0.4	0.7
Direct	55	-16.50	5.67	3.76	0.6	0.9
Yearly	111					
Mountain	35	-21.16	-1.36	4.66	5.7	14
Direct	76	-18.88	-0.68	4.99	4.6	11.8

Fig. 5.6 displays the influence of orographic features on the isotopic concentration of precipitation at Inuvik. The routes taken to Inuvik by two northern storms with precipitation occurring at a temperature of -32°C can be seen on Fig. 5.6a. One occurred on 23rd January 2017, and had an $\delta^{18}\text{O}$ value of -35.1‰ . This storm originated over the Arctic Ocean and travelled over the Brooks Range and Yukon mountains before reaching Inuvik. The other storm, 17 days later on 10th February 2017, consisted of precipitation with an $\delta^{18}\text{O}$ of -22.4‰ . It reached Inuvik directly from the Beaufort Sea. The contrast in $\delta^{18}\text{O}$ values between the storms may be attributed to the fractionation during condensation of the air mass above the mountains.

Similarly, the routes to Inuvik of two other storms, one northern and one southern, from the fall of 2015 are presented in Fig. 5.6b. The southern storm had a lower $\delta^{18}\text{O}$ value despite its warmer conditions. The northern storm approached Inuvik directly from the North while the southern storm travelled over various mountain ranges. This is the primary reason why westerly and southerly storms have more negative $\delta^{18}\text{O}$ concentrations despite being accompanied by higher air temperatures when compared with northerly storms.

5.3 Snowpack near Inuvik

Over the five-year period, a total of 41 lake snow pit and 37 forest snow pit samples were collected. Each year at least six samples were collected from each environment, except for the forest snow pits in 1993, when only two were sampled. The mean $\delta^{18}\text{O}$ for the lake and forest snow pits were -28.55‰ and -28.85‰ , respectively. A Student's t-test indicated that there is no significant difference ($\alpha = 0.05$) between the two groups. These data are presented in Table 5.7.

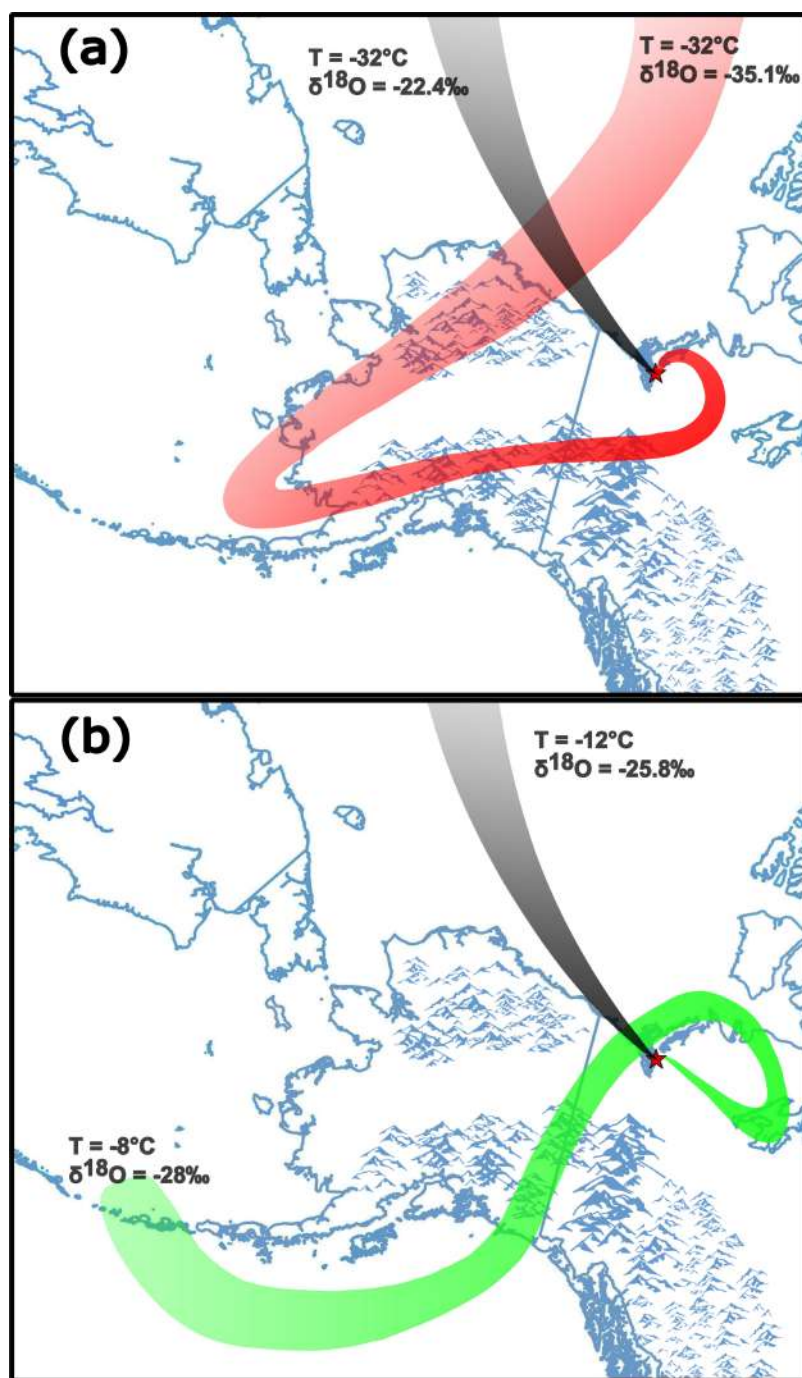


Figure 5.6: Illustration of storm trajectories derived from the HYSPLIT model. All trajectories show 13 days of travel. Red trajectory in (a) represents the path taken by the air mass responsible for an event on 23 January 2017; the black trajectory represents a path up to 10 February 2017. The black trajectory in (b) represents the path taken by the air mass responsible for an event on 29 October 2015; the green trajectory represents a path up to 2 November 2015.

Table 5.7: $\delta^{18}\text{O}$ (‰) of lake and forest snow pits in the Mackenzie Delta region, between 1991 and 1995. The bottom row displays the average values for the respective years. The overall mean of the data is -28.03‰

1991		1992		1993		1994		1995	
Lake	Forest	Lake	Forest	Lake	Forest	Lake	Forest	Lake	Forest
-29.5	-29.6	-27.5	-24.8	-26.7	-29.9	-29.8	-29.4	-28.4	-30.9
-31.1	-29.0	-27.3	-27.0	-29.1	-27.3	-29.2	-26.3	-32.4	-30.4
-32.2	-30.8	-26.3	-24.6	-30.1		-29.2	-28.3	-31.4	-30.3
-27.5	-28.6	-26.1	-29.4	-25.3		-24.8	-31.5	-29.0	-28.9
-30.5	-32.2	-27.0	-25.1	-24.6		-30.0	-28.8	-27.5	-29.0
-27.5	-29.8	-28.4	-26.6	-25.6		-27.9	-30.9	-30.7	-26.6
-26.8	-31.3		-25.8			-29.7	-29.4	-28.2	-30.6
	-34.6					-29.8	-28.6	-31.6	-31.4
						-33.9	-29.3	-23.7	-26.3
						-26.2			
-29.3	-30.74	-27.1	-26.18	-26.9	-28.15	-29.37	-28.87	-29.21	-29.34

An analysis of variation between the values for forest and lakes indicated that there is no statistically significant difference in the mean values of samples from the two environments ($\alpha = 0.05$). The minima ($\sim -34\text{‰}$), maxima ($\sim -24\text{‰}$) and mean and median values (between -28 and -29‰) were nearly identical.

Data from the snow pits align well with the weighted mean $\delta^{18}\text{O}$ value of winter precipitation collected between October and March 1990-95 at Inuvik (Fig. 5.7). Generally, the snowpack is more depleted than winter precipitation, with the mean difference between them being 0.9‰ . However, upon removing 1994 the mean difference is 0.5‰ . This suggests that the snowpack as a whole at Inuvik is representative of seasonal winter precipitation.

5.4 Boreholes near Illisarvik

Isotopic data from boreholes drilled outside the western edge of the Illisarvik drained lake are presented in Fig. 5.8. The base of the active layer can be seen at around 90 cm depth, with samples above having $\delta^{18}\text{O}$ values generally greater than -21‰ and those below being less than -23‰ (Fig. 5.8).

A co-isotope plot for boreholes A and B is presented in Fig. 5.9. Borehole A had a FF regression that was not statistically different from yearly precipitation at Inuvik (Table 5.2), however borehole B was significantly different.

5.4.1 Boreholes A & B

At borehole A, three samples were taken from the active layer (Fig. 5.8). Near the surface, the $\delta^{18}\text{O}$ was -18.4‰ . The other two samples from the active layer had $\delta^{18}\text{O}$ values above -21‰ . These values align well with those associated with active-layer ice (Mackay, 1983b).

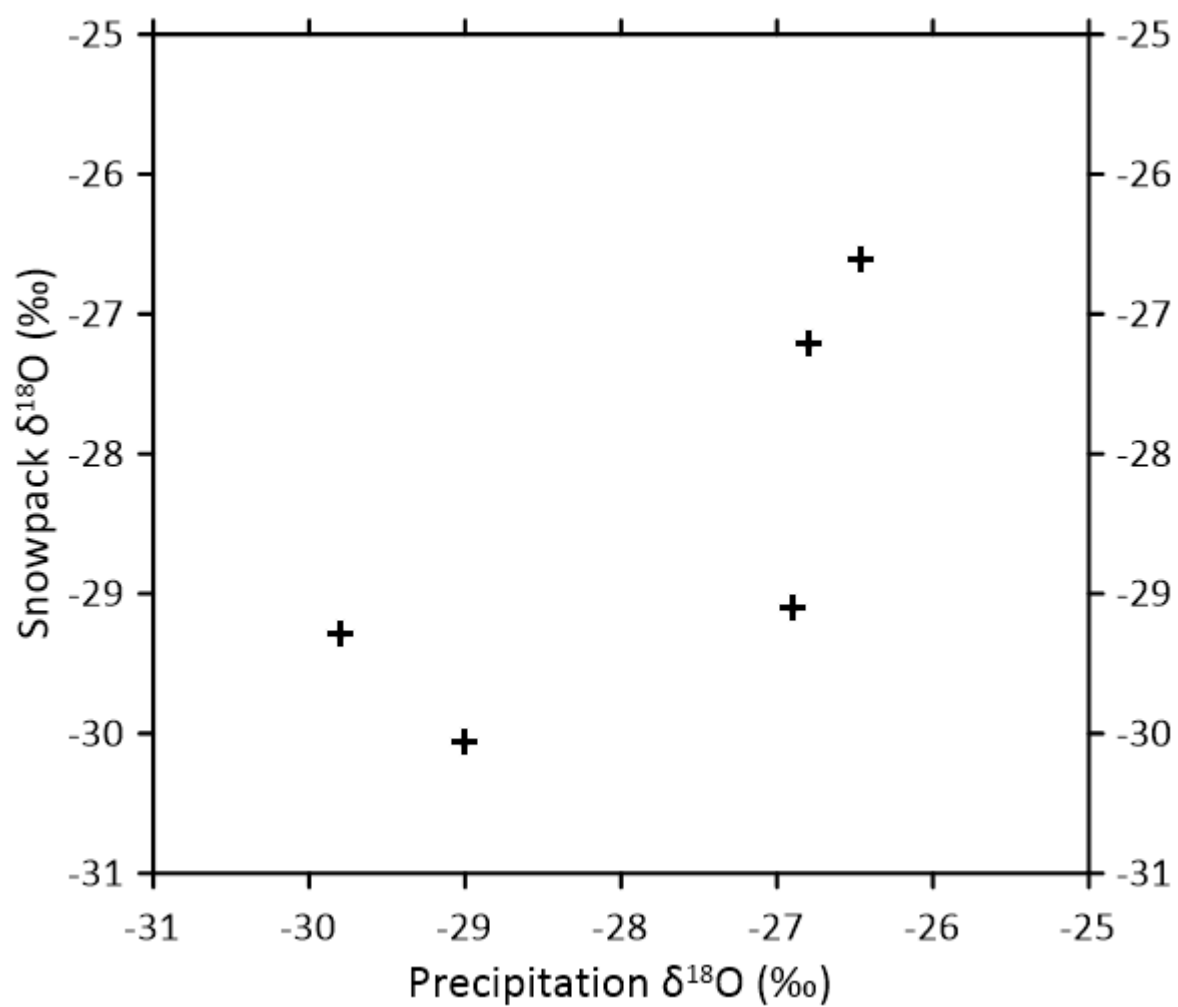


Figure 5.7: Each point represents the seasonal $\delta^{18}\text{O}$ for snow pits and precipitation near Inuvik. Snow pit samples were collected in April. Precipitation values are calculated as the weighted mean for precipitation from October to March.

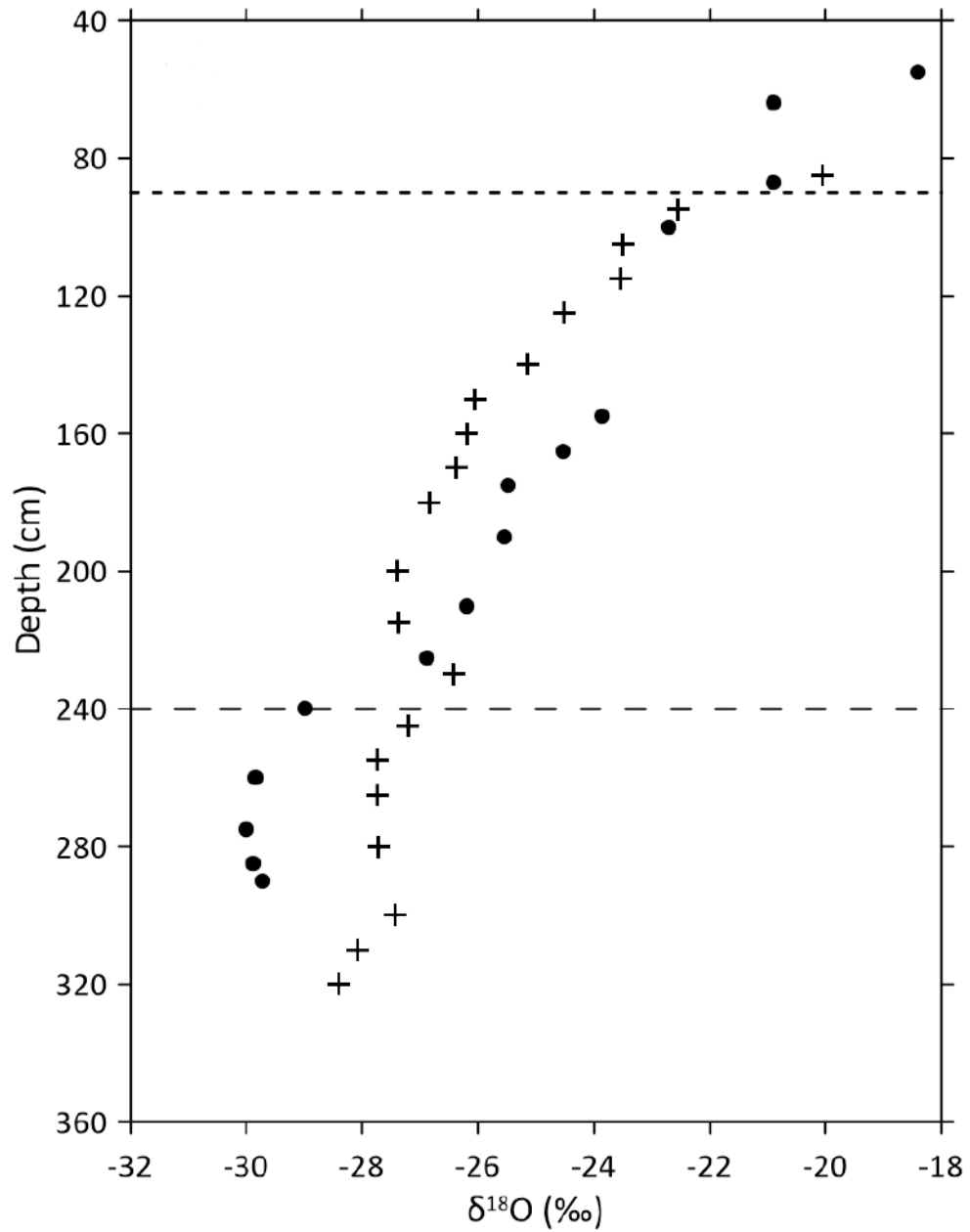


Figure 5.8: Plot of $\delta^{18}\text{O}$ with depth for boreholes A (dots) and B (crosses) in permafrost near Illisarvik in August, 2022. Short dashed line indicates estimated active layer base, long dashes indicate estimated thaw unconformity.

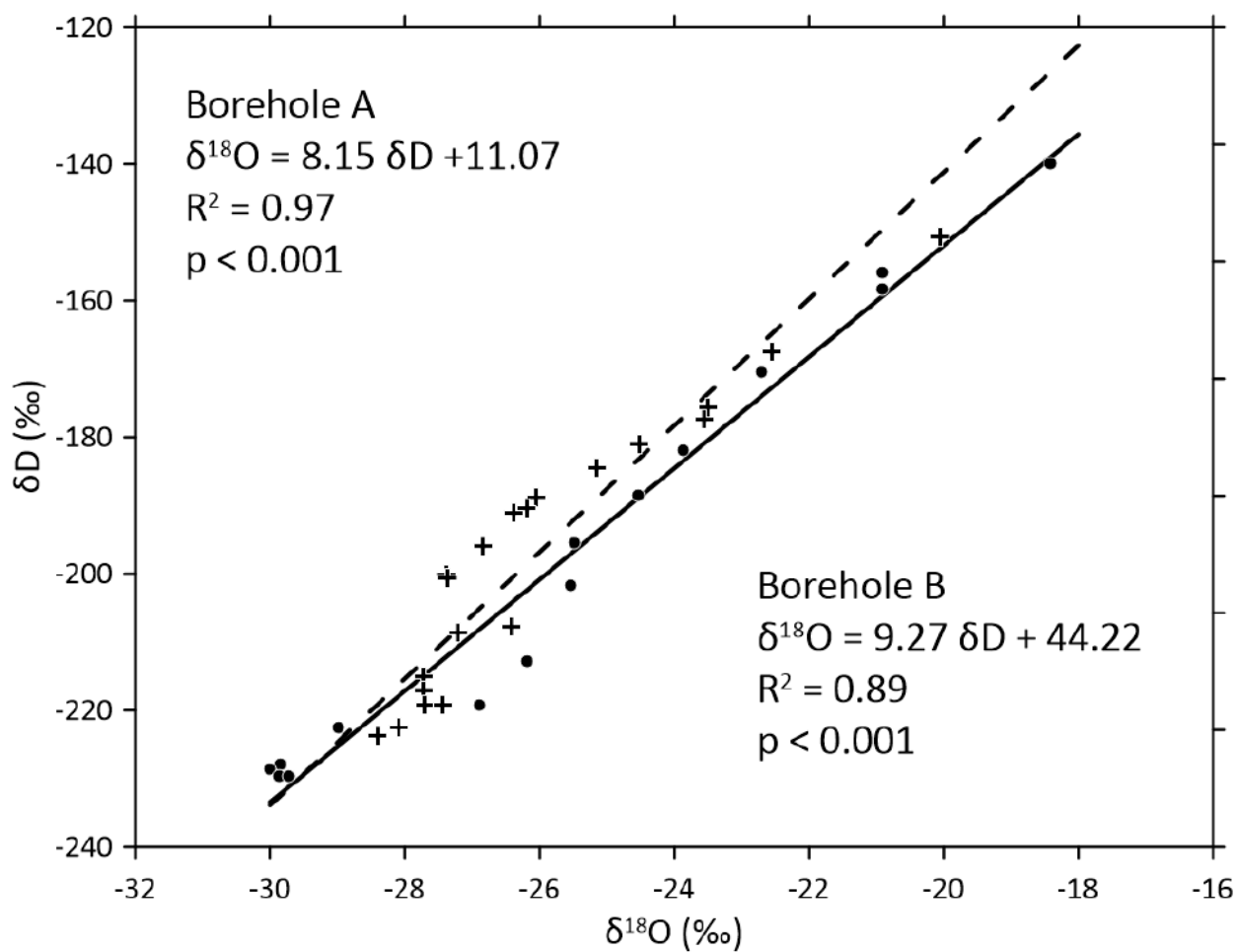


Figure 5.9: Co-isotope plots for boreholes A (dots) and B (crosses) from permafrost near Illisarvik. Equations and FF lines for A (solid line) and B (dashed line) are also presented.

$\delta^{18}\text{O}$ values just below the base of the active layer in borehole A were -22.7‰ . The next sample down had an $\delta^{18}\text{O}$ value of -23.9‰ , and all subsequent lower samples had progressively more negative $\delta^{18}\text{O}$, with the lowest value (-30‰) being at 275 cm depth.

The soil at borehole A was brown until 135 cm depth, below which it became grey, roughly 45 cm below base of the active layer. Ice lenses and other forms of excess ice were common, but there were some intervals in which there was no excess ice.

For borehole B, the frost table during sampling was at a depth of 73 cm. Between the frost table and base of the active layer was a $\delta^{18}\text{O}$ value of -20‰ , and the value at the base was -22.6‰ . Similar to borehole A, $\delta^{18}\text{O}$ within the permafrost decreased with depth, from -23.5‰ to -28.4‰ at 320 cm.

Borehole B was more depleted than borehole A to a depth of 230 cm, at which point borehole B became relatively enriched compared to borehole A for the remainder of the sampling (Fig. 5.8).

5.4.2 *Thaw Unconformity*

There is a well established thaw unconformity in the region due to the Holocene Thermal Maximum (HTM) (Burn et al., 1986; Burn, 1997). Mackay (1983b) associated $\delta^{18}\text{O}$ values in permafrost below -27‰ with a Pleistocene age. The thaw unconformity for these boreholes is assumed to be at a depth of 240 cm, because of the gradual increase in $\delta^{18}\text{O}$ and δD above 240 cm from more consistent Pleistocene values below (Fig. 5.8). This also coincides with a drastic shift in D-excess for both boreholes (Fig. 5.10).

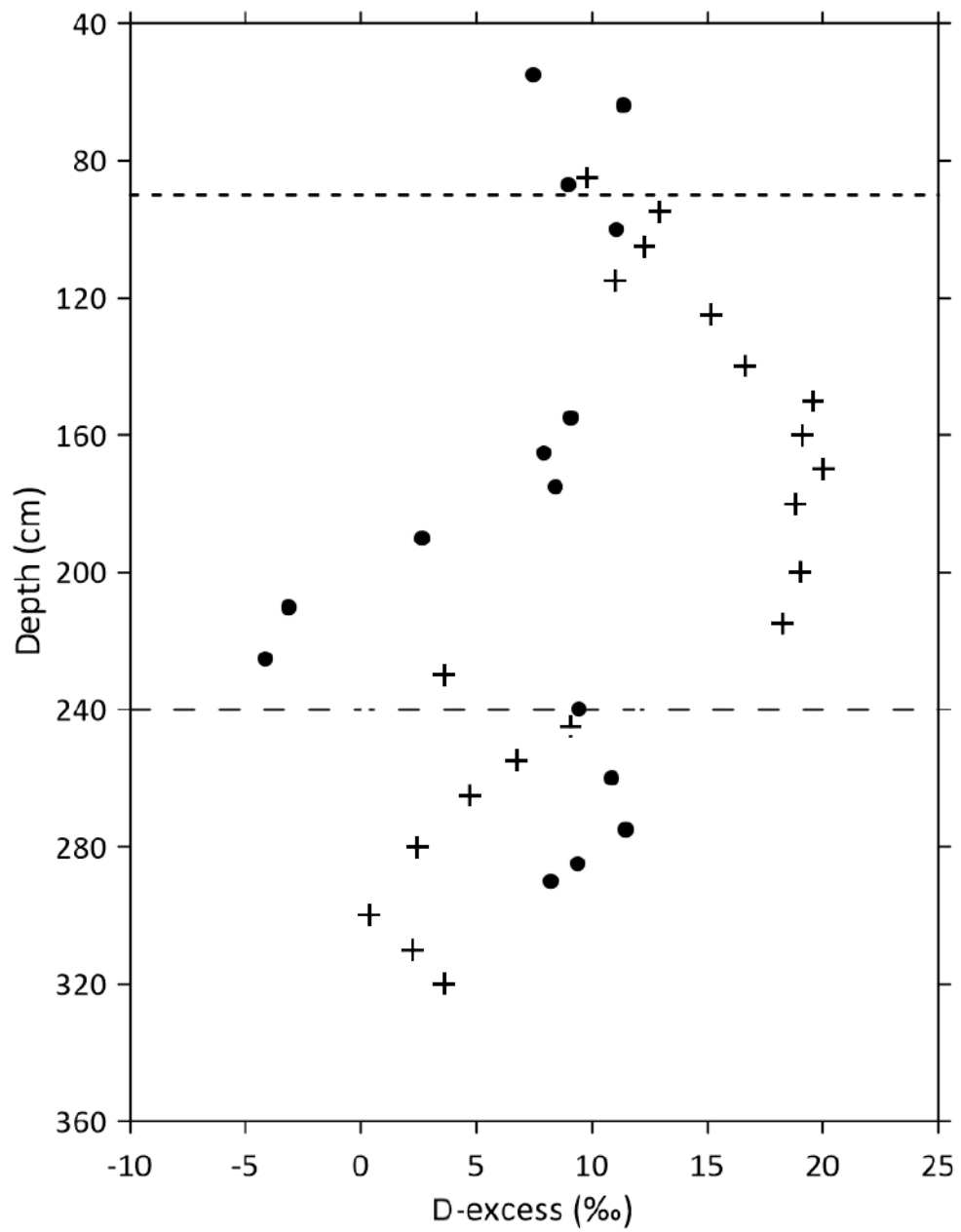


Figure 5.10: The D-excess at depth for the two boreholes drilled into permafrost near Illisarvik in August, 2022. Dotted line indicates suspected thaw unconformity.

5.5 Anti-Syngenetic Ice Wedges of Illisarvik

A total of 101 samples were analysed from six anti-syngenetic ice wedges sampled at Illisarvik. Sixty-five samples were collected from wedges on the eastern side of the basin and 36 from the West. Basic statistics for each wedge can be found in Table 5.8. The $\delta D:\delta^{18}O$ relations for all samples as well as each individual wedge are shown in Table 5.9. All slopes of the co-isotope plots were statistically significant ($\alpha = 0.05$), and while they varied for individual wedges, the $\delta D:\delta^{18}O$ least-squares linear relation for all samples is similar to the LMWL at Inuvik of Fritz et al. (2022) ($\delta D = 7.39\delta^{18}O - 6.7$).

A co-isotope plot of all wedges is shown in Fig. 5.11, which indicates the majority of samples have $\delta^{18}O$ values between -26‰ and -21‰ . These values are higher than those in the snow pits (Table 5.7) and winter precipitation near Inuvik between 1985-95 (-28.2‰ , Table 5.1). This is consistent with other studies examining the isotopic composition of ice wedges (Mackay, 1983b; Lauriol et al., 1995; Opel et al., 2017; Holland et al., 2020). The values from these anti-syngenetic ice wedges align well with winter precipitation at Inuvik between 2015-18 (-25.1‰ , Table 5.1). Generally, these $\delta^{18}O$ values match those found in ice wedges around the region (Michel, 2011; Holland et al., 2020; Porter and Opel, 2020).

T2

T2 is the western wedge situated parallel with the contours. There are two statistically similar groups of samples on T2 at the edges and near the middle of the wedge (Fig. 5.12). The wedge shows an expected distribution of isotopes across a wedge, as the two edges are composed of near-identical ice with $\delta^{18}O$ values below -24‰ . In the middle of the wedge there is a group of samples with similar values (-23.5‰), and a sample nearer the upslope side of the wedge that has a higher value (-22‰).

Table 5.8: Distribution of values for $\delta^{18}O$ (‰), δD (‰) and D-excess values of each anti-syngenetic ice wedge; w is the width (cm) and n is the number of samples.

Wedge	w	n	$\delta^{18}O_{min}$	$\delta^{18}O_{max}$	δD_{min}	δD_{max}	D-excess _{min}	D-excess _{max}
T2	172	13	-24.85	-21.99	-189.6	-161.8	4.91	15.69
T3	240	11	-24.04	-21.81	-181.7	-164.5	4.92	11.89
T4	330	12	-24.93	-21.05	-192.1	-158.7	5.90	14.47
E1	580	23	-26.73	-22.66	-200.5	-169.9	4.68	16.91
E2	615	24	-27.86	-22.48	-205	-175.1	4.82	17.92
E3	375	18	-29.4	-20.94	-227.6	-158.3	0.57	14.28

Table 5.9: LSL and FF for each wedge and all combined samples.

Wedge	Least-Squares Linear	Functional Fit	R^2
T2	$\delta D = 7.93\delta^{18}\text{O} + 8.41$	$\delta D = 9.03\delta^{18}\text{O} + 34.79$	0.75
T3	$\delta D = 6.83\delta^{18}\text{O} - 17.68$	$\delta D = 7.23\delta^{18}\text{O} - 8.60$	0.85
T4	$\delta D = 7.95\delta^{18}\text{O} + 8.8$	$\delta D = 8.26\delta^{18}\text{O} + 15.94$	0.91
E1	$\delta D = 6.38\delta^{18}\text{O} - 20.87$	$\delta D = 6.97\delta^{18}\text{O} - 14.50$	0.77
E2	$\delta D = 5.73\delta^{18}\text{O} - 45.4$	$\delta D = 5.95\delta^{18}\text{O} - 40.05$	0.87
E3	$\delta D = 8.25\delta^{18}\text{O} + 16.56$	$\delta D = 8.32\delta^{18}\text{O} + 18.21$	0.98
All	$\delta D = 7.48\delta^{18}\text{O} - 2.29$	$\delta D = 7.72\delta^{18}\text{O} - 3.31$	0.92

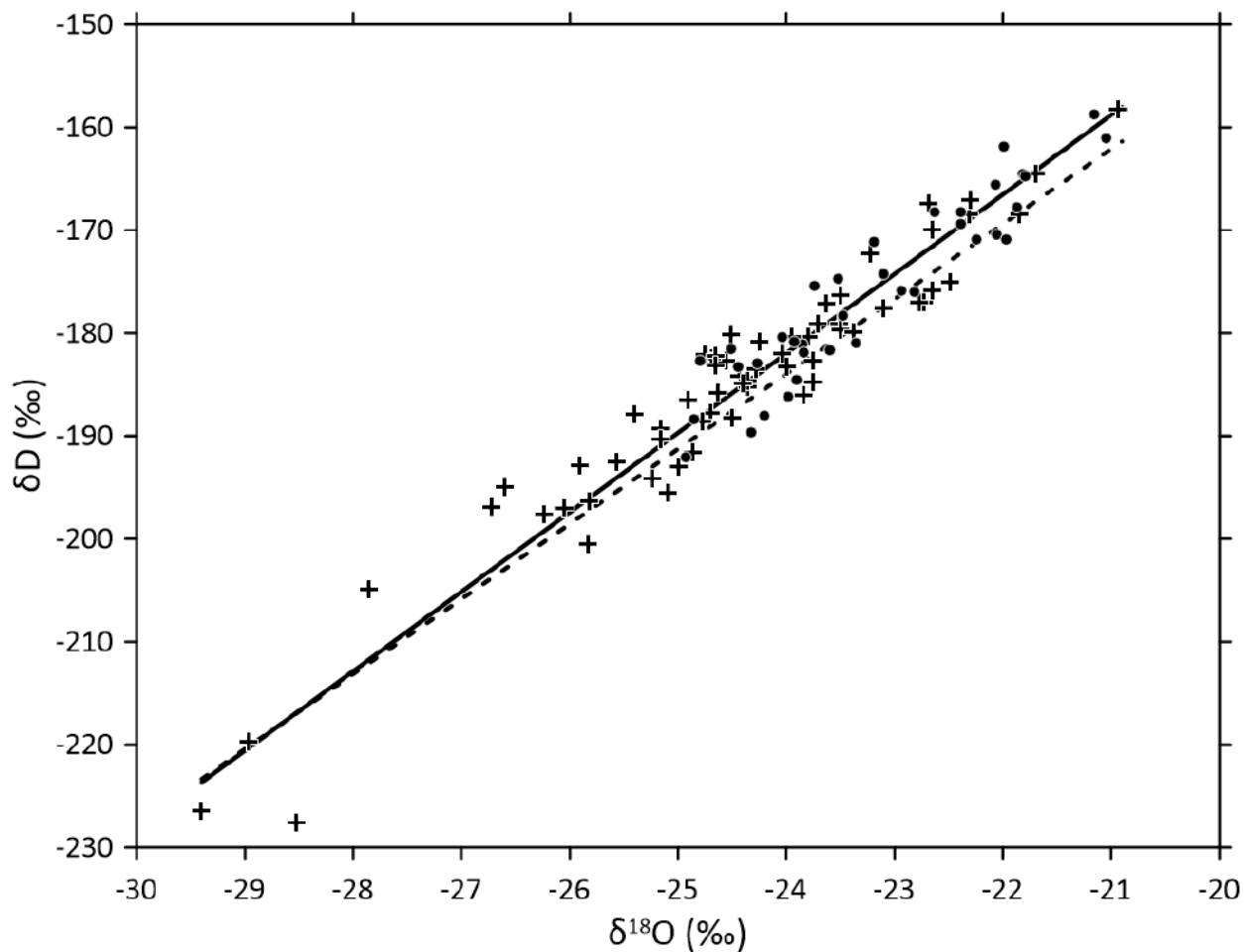


Figure 5.11: Co-isotope plots of $\delta^{18}\text{O}$ and δD of samples from anti-syngenetic wedge ice surrounding Illisarvik drained lake basin ($n = 101$). Crosses indicate eastern wedges and dots indicate western. Functional fit regression of the samples (solid line) is presented with the functional fit regression of samples from Inuvik precipitation between 2015-18 (dashed line).

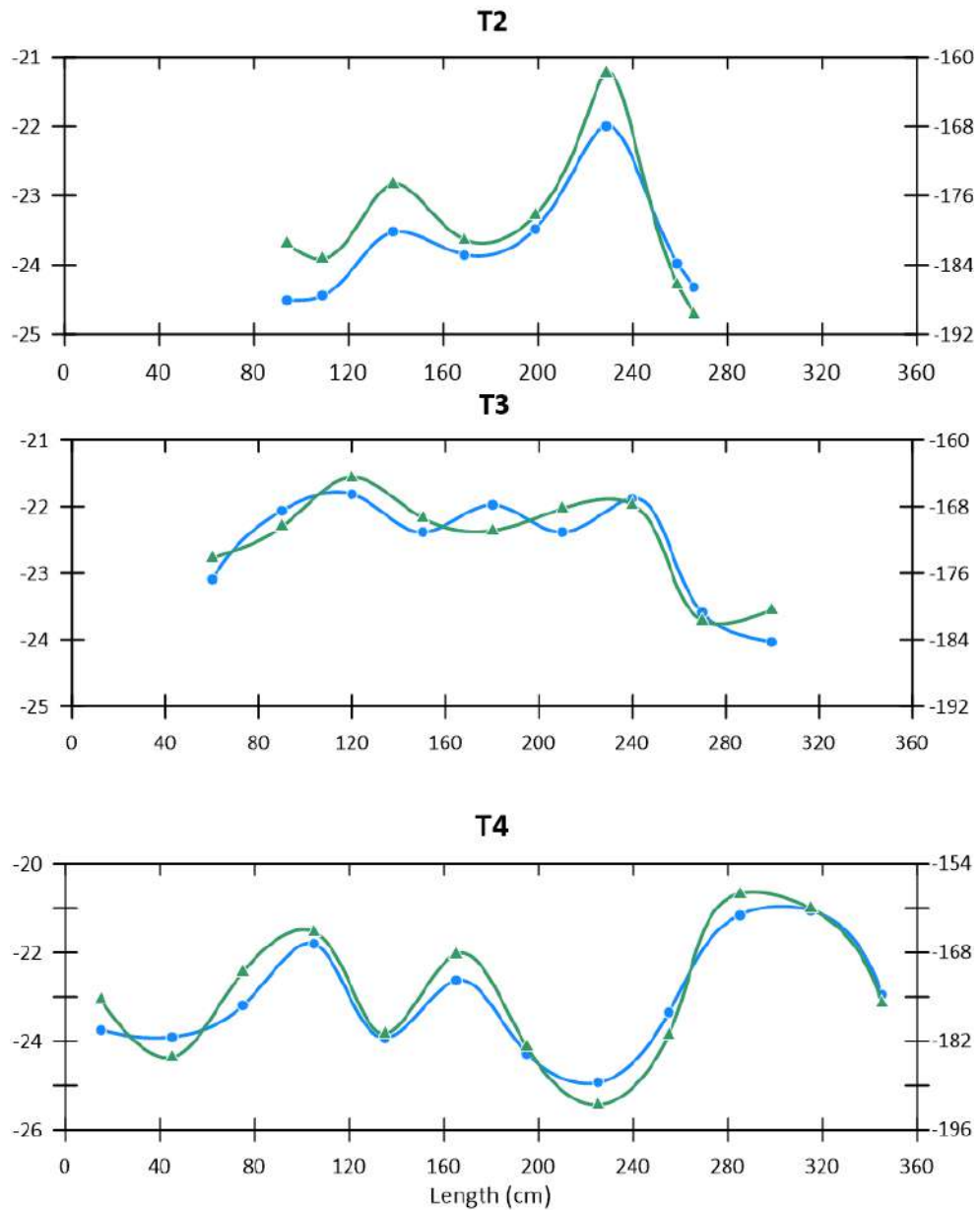


Figure 5.12: Isotope values at 25 cm depth across each western wedge. Blue circles and left axis present $\delta^{18}O$ (‰) values while green triangles and right axis present D (‰) values. The line connecting dots is for visualisation purposes and does not represent interpolation or values between samples.

The $\delta D:\delta^{18}O$ functional fit was significantly greater than the fit derived from precipitation at Inuvik and the other wedges sampled near Illisarvik (Table 5.9). Deuterium values along the wedge follow a similar pattern to that of $\delta^{18}O$ (Fig. 5.12). T2 has the highest maximum D-excess of the western wedges (Table 5.8 and Fig. 5.13), as well as the highest mean (10.1‰) but the lowest median (9.7‰). The D-excess of T2, and all other wedges, is similar to D-excess values in winter precipitation at Inuvik (Tables 5.1 & 5.8).

T3

Unlike T2, T3 does not have similar $\delta^{18}O$ near each edge and is situated perpendicular to the contours. There is a large group of similar samples through the middle of the wedge that spans almost its entirety. These samples have $\delta^{18}O$ values above -23‰ . The pattern is similar with deuterium across the wedge (Fig. 5.12). The consistent $\delta^{18}O$ and δD values in the middle of the wedge have variable D-excess. The lowest D-excess (4.9‰) and second highest (10.8‰) values were taken next to each other within the stretch of consistent isotope values (Fig. 5.13).

T4

T4 is perpendicular to the contours and has groups of statistically similar samples scattered across the wedge (Fig. 5.12). Further, this wedge has the largest range in both $\delta^{18}O$ and δD (Table 5.8). The highest isotope values along the wedge are found near, but not at, the right edge, and the lowest values are near the center. The D-excess across the wedge varies, although there was more variation near the left edge (Fig. 5.13).

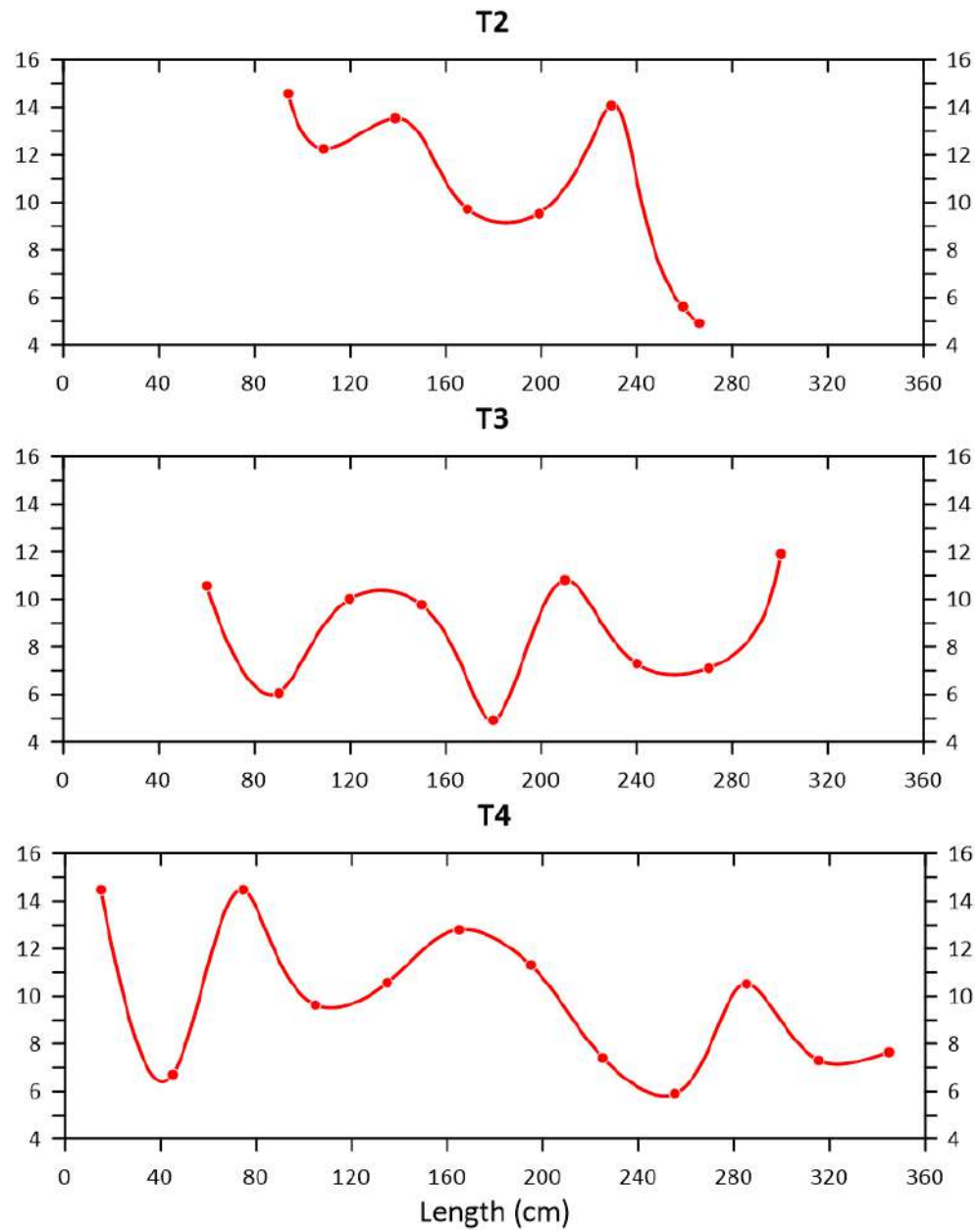


Figure 5.13: D-excess (%) values at 25 cm depth across each western wedge. Line connecting dots is for visualisation purposes and does not represent interpolation or values between samples.

E1

There was seemingly no pattern in $\delta^{18}\text{O}$ values across E1, however it contains the smallest $\delta^{18}\text{O}$ range of the eastern wedges (Fig. 5.14). E1 runs perpendicular to the contours. There are samples of similar isotopic composition on opposite sides of the wedge, and $\delta^{18}\text{O}$ values for the groups (-24‰ and -25‰) are similar to the groups found in the other eastern wedges. A section of pool ice occurred adjacent to E1, as determined from the dark and clear ice. It had a $\delta^{18}\text{O}$ of -18.7‰ , similar to active-layer ice from borehole A.

In contrast to all other wedges, the highest $\delta^{18}\text{O}$ values in E1 were found at the edges, and the lowest value was found near the middle of the wedge. The $\delta^{18}\text{O}$ value at the right edge of the wedge can likely be attributed to the close proximity of the sample to the wedge surface (2 cm). Deuterium values at the middle and right end of the wedge separate from the $\delta^{18}\text{O}$ values more so than in the other wedges (Fig. 5.14), and this corresponds with high D-excess values (Fig. 5.15). The D-excess was highest in the middle of the wedge and near the edges and a lower but consistent range in between the middle and edges (Fig. 5.15). E1 had the lowest internal D-excess range (12.2‰).

E2

E2 is the eastern wedge parallel with the contours. There were two groups of statistically similar ice on either edge of the wedge, both between -25‰ and -23‰ . Also similar to E1 was the decrease in $\delta^{18}\text{O}$ values near the middle of the wedge, which was adjacent to the highest values within the wedge. This should indicate colder ice, but it does not follow the expected growth pattern of an ice wedge. The lowest $\delta^{18}\text{O}$ value was present near the right edge, where the -27.8‰ value may be indicative of wedge ice that developed during the Pleistocene (Mackay, 1983b).

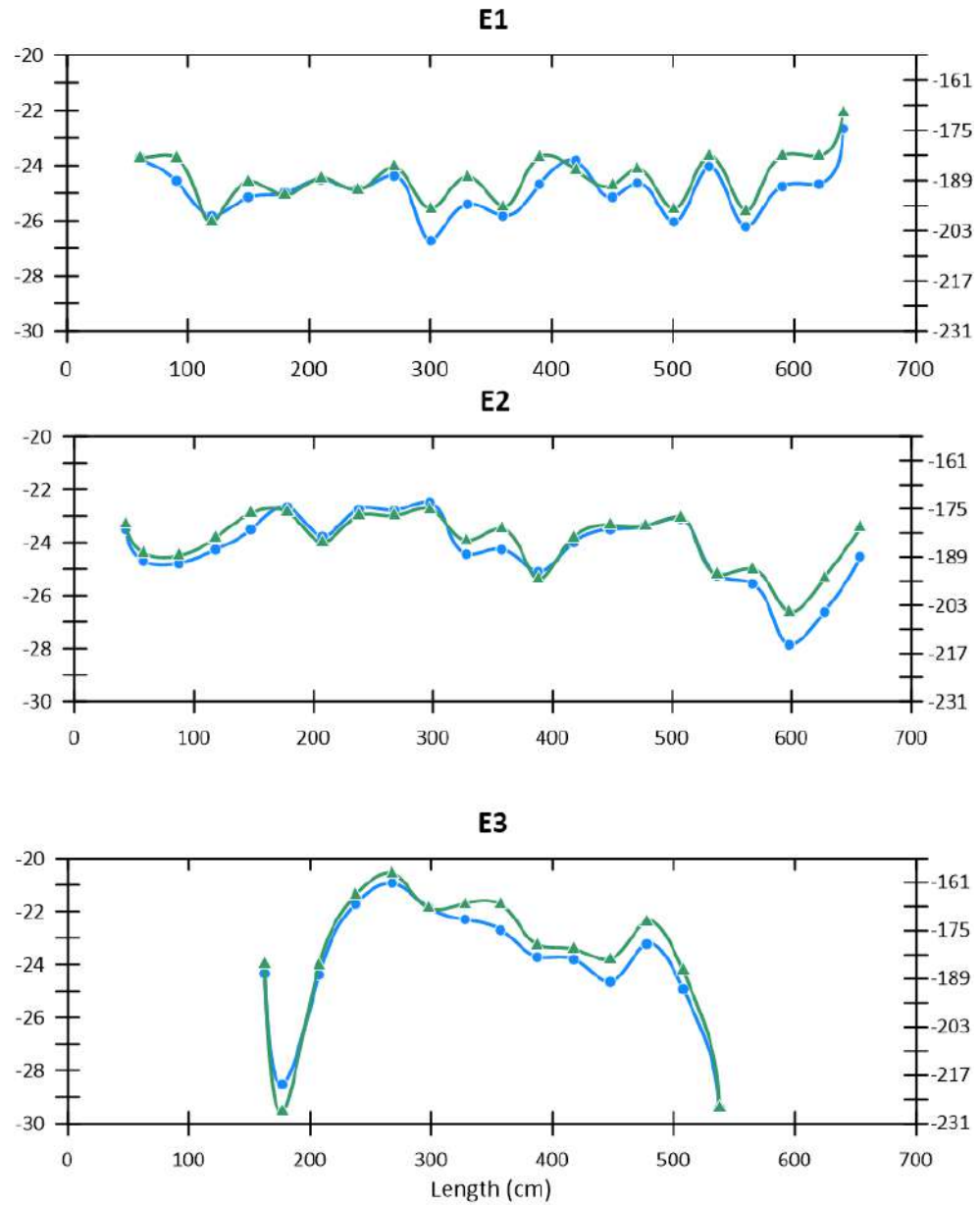


Figure 5.14: Isotope values at 25 cm depth across each eastern wedge. Blue circles and left axis present $\delta^{18}O$ (‰) values while green triangles and right axis present D (‰) values. Line connecting dots is for visualisation purposes and does not represent interpolation or values between samples.

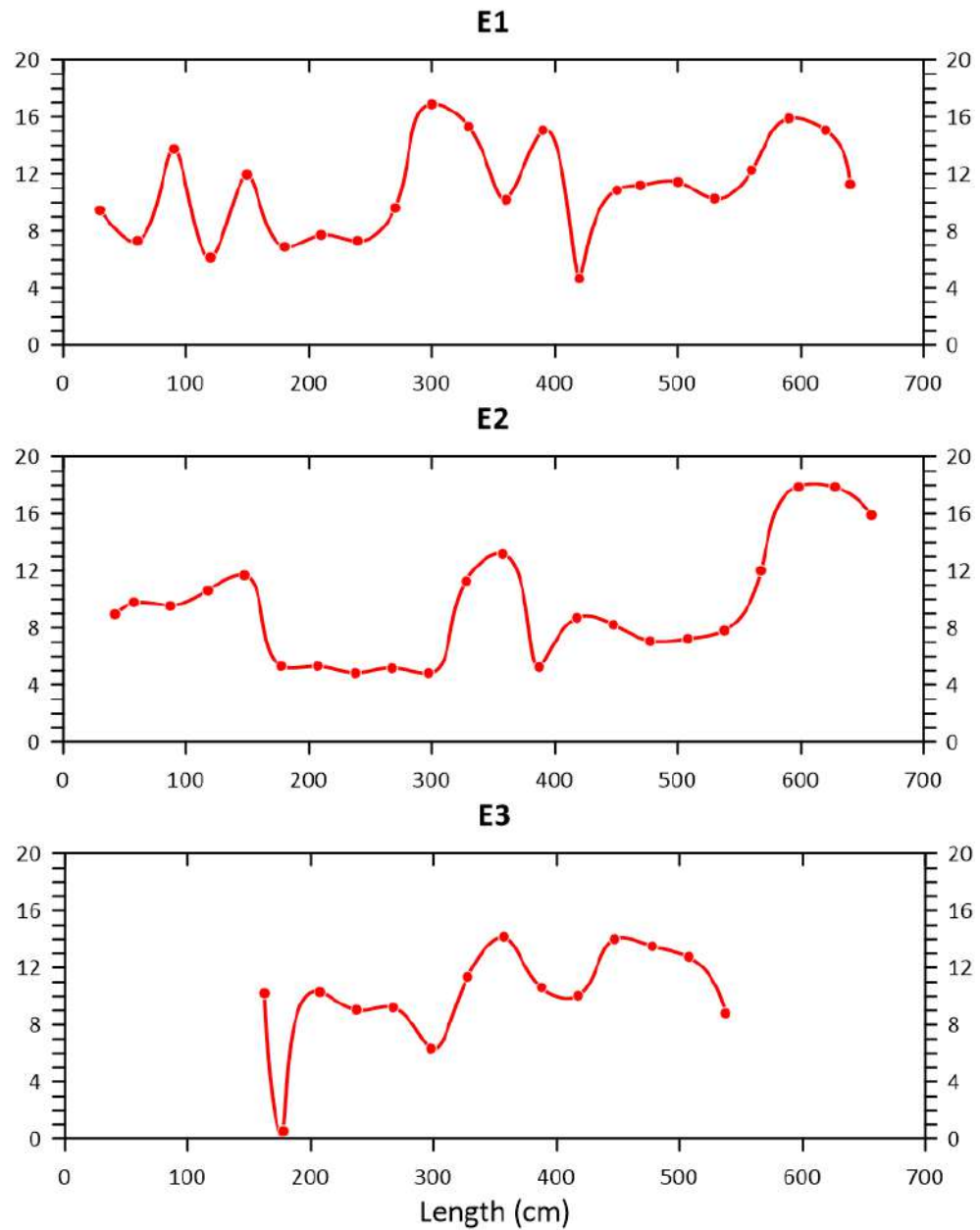


Figure 5.15: D-excess (‰) values at 25 cm depth across each eastern wedge. Line connecting dots is for visualisation purposes and does not represent interpolation or values between samples.

The samples to the right of this had higher $\delta^{18}\text{O}$ values and were taken as the wedge tapers (Fig. 4.10).

E2 presented the highest D-excess value of all the wedges sampled. There was a group of five low D-excess samples on the downslope side of the wedge (Fig. 5.15), which coincided with a consistent section of isotopically enriched ice. The $\delta^{18}\text{O}$ of the ice was statistically similar to the $\delta^{18}\text{O}$ of a section on the upslope side of the wedge which also exhibited consistently low D-excess values. There is a broadly similar pattern to E1 where high D-excess values are present near the middle and edges, with consistently lower values between the edge and middle (Fig. 5.15). E2 had the lowest slope for the $\delta D:\delta^{18}\text{O}$ relation (Table 5.9), suggesting substantial fractionation has occurred to the water prior to freezing.

E3

The final wedge sampled, E3, had the widest range for both $\delta^{18}\text{O}$ and δD , and apparent Pleistocene age ice at both edges. Three samples were taken from the right edge of E3 (Fig. 4.10), at 5 cm, 20 cm, and 27 cm. Respectively, their $\delta^{18}\text{O}$ values were -25.9‰ , -29‰ , and -29.4‰ . In fact, both edges contain ice with $\delta^{18}\text{O}$ below -28‰ , while the middle of the wedge contains consistently high ($> -24\text{‰}$) $\delta^{18}\text{O}$ values. The ice with values at or below -28‰ may be interpreted as wedge ice originally formed during the Pleistocene. The highest $\delta^{18}\text{O}$ of all the wedges sampled came from the middle of E3.

The lowest D-excess value from all sampled wedges was determined near the left edge, however it is an isolated sample (Fig. 5.15). This sample also coincides with the lowest $\delta^{18}\text{O}$ value (Fig. 5.14). Similar to the other eastern wedges, there are groups of consistent D-excess values across the wedge. E3 has a functional fit for the $\delta D:\delta^{18}\text{O}$ relation that is greater than precipitation at Inuvik, and is the highest of the eastern wedges.

5.5.1 *Eastern vs. Western Wedges*

The eastern wedges are substantially larger and likely older, because of their width and minimum $\delta^{18}\text{O}$ values (Table 5.8, 5.10). The mean $\delta^{18}\text{O}$ value of the eastern sites was -24.5‰ , while the minimum and maximum values were -29.4‰ and -20.9‰ , respectively. The minimum and maximum values from the western wedges were -24.9‰ and -21.1‰ , respectively. The western wedges had a mean $\delta^{18}\text{O}$ value of -23.2‰ , which according to a Mann-Whitney test was statistically significantly different from that of the eastern wedges ($p > 0.001$). The similar maximum values of the two groups indicate that Holocene growth has been occurring at both sets of ice wedges, whereas the quite different minima suggest some of the eastern wedges contain ice from a much colder period than the western wedges. Box-and-whisker plots in Fig. 5.16 illustrate the differences between the two groups.

Where the western wedges had the lowest $\delta^{18}\text{O}$ near their edges, the eastern wedges generally had sharp increases in $\delta^{18}\text{O}$. In E1, the values near the right edge of the wedge were the highest throughout, and it is possible that the ice received active layer melt during a deep thaw season. This is likely the case with E2 as well, as both wedges exhibit much narrower edges than the western wedges (Fig. 5.14).

The eastern wedges had a larger range in D-excess than the western wedges (17.4‰ and 10.8‰), which indicates that more fractionation occurred to the meltwater entering the eastern wedges. The range was skewed for the eastern wedges due to a very low minimum value at E3.

Table 5.10: Mean values for $\delta^{18}O$ and δD (‰ vs. VSMOW) and D-excess of anti-syngenetic ice wedges based on the side of basin (East and West) and relationship to the contours (Perpendicular and Parallel).

Wedge	Width (cm)	$\delta^{18}O$	δD	D-excess	n
Eastern	523	-24.5	-185.4	10.3	62
Western	247	-23.2	-175.9	9.6	36
Perpendicular	381	-23.9	-181.3	10.2	62
Parallel	394	-24.1	-182.9	9.8	36
All	385	-24	-181.9	-10.1	98

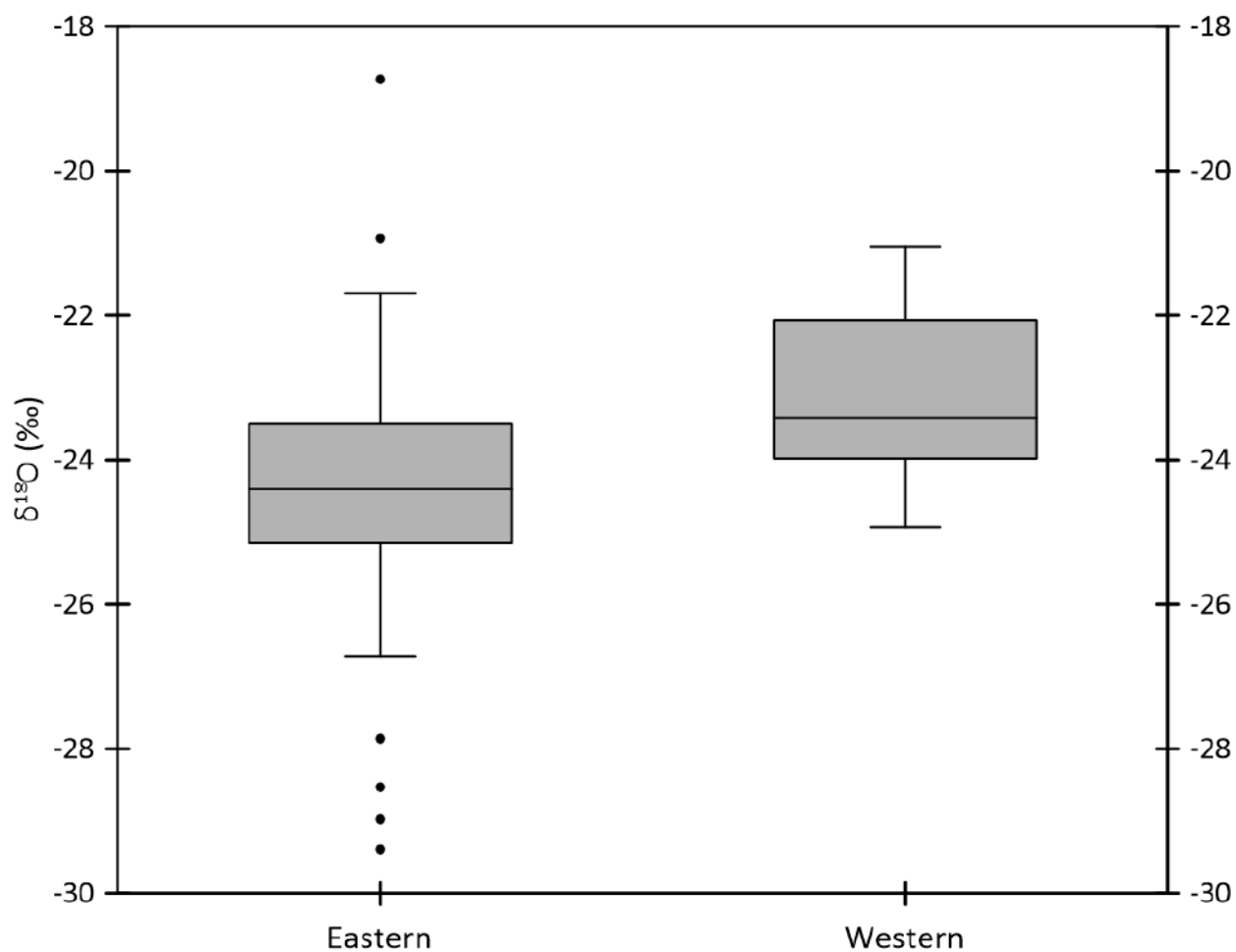


Figure 5.16: Boxplots of $\delta^{18}\text{O}$ between the eastern and western wedges. Note the similar maximum values, but the much lower values from the eastern wedges. The outliers represent points near the edges, and the outlier near -18‰ represents pool ice from adjacent the wedge.

There is a difference in slopes for the functional fit of $\delta D:\delta^{18}O$ between the two sides of the basin (Table 5.11). The eastern wedges have a $\delta D:\delta^{18}O$ similar to the $\delta D:\delta^{18}O$ relation seen in the total population of sampled wedge ice (7.72, Table 5.9). The $\delta D:\delta^{18}O$ from both eastern and western sites are greater than the functional fit from precipitation at Inuvik (7.29, Table 5.2).

5.5.2 *Parallel vs. Perpendicular to Slope*

The two wedges parallel with the contour, T2 and E2, contrast with each other. The upslope edge of T2 does not taper, instead going directly down (Fig. 4.7), whereas the edges of E2 taper as would be expected with an anti-syngenetic wedge (Fig. 4.10).

The isotopic differences between ice wedges that are parallel and perpendicular to the slope are similar to those seen between the eastern and western ice wedges. Perpendicular and parallel wedges have maximum $\delta^{18}O$ values of -20.9‰ and -22‰ , respectively. The minimum $\delta^{18}O$ values are -29.4‰ for perpendicular sites and -27.9‰ for those that were parallel. The mean $\delta^{18}O$ value is similar among wedges, at -23.9‰ for perpendicular sites and -24.1‰ for those parallel. According to an unpaired t-test, there is no significant difference between the two means ($p > 0.05$).

The slopes of the functional fits for the $\delta D:\delta^{18}O$ relations exhibit a significant difference between the perpendicular and parallel wedges (Table 5.11). There is a distinction between wedges parallel to the contours, with T2 displaying a slope of 9.03 and E2 measuring 5.95 (Table 5.9).

Table 5.11: Least-squares linear regression and functional fit of anti-syngenetic ice wedges based on the side of basin (Eastern and Western) and relationship to the contours (Perpendicular and Parallel).

Characteristic	LSL (R^2)	FF
Eastern	$\delta D = 7.47\delta^{18}\text{O} - 2.52$ (0.92)	$\delta D = 7.71\delta^{18}\text{O} + 3.18$
Western	$\delta D = 7.56\delta^{18}\text{O} - 0.63$ (0.88)	$\delta D = 7.95\delta^{18}\text{O} + 8.39$
Perpendicular	$\delta D = 7.78\delta^{18}\text{O} + 4.94$ (0.95)	$\delta D = 7.94\delta^{18}\text{O} + 8.70$
Parallel	$\delta D = 6.17\delta^{18}\text{O} - 4.19$ (0.82)	$\delta D = 6.56\delta^{18}\text{O} - 24.89$

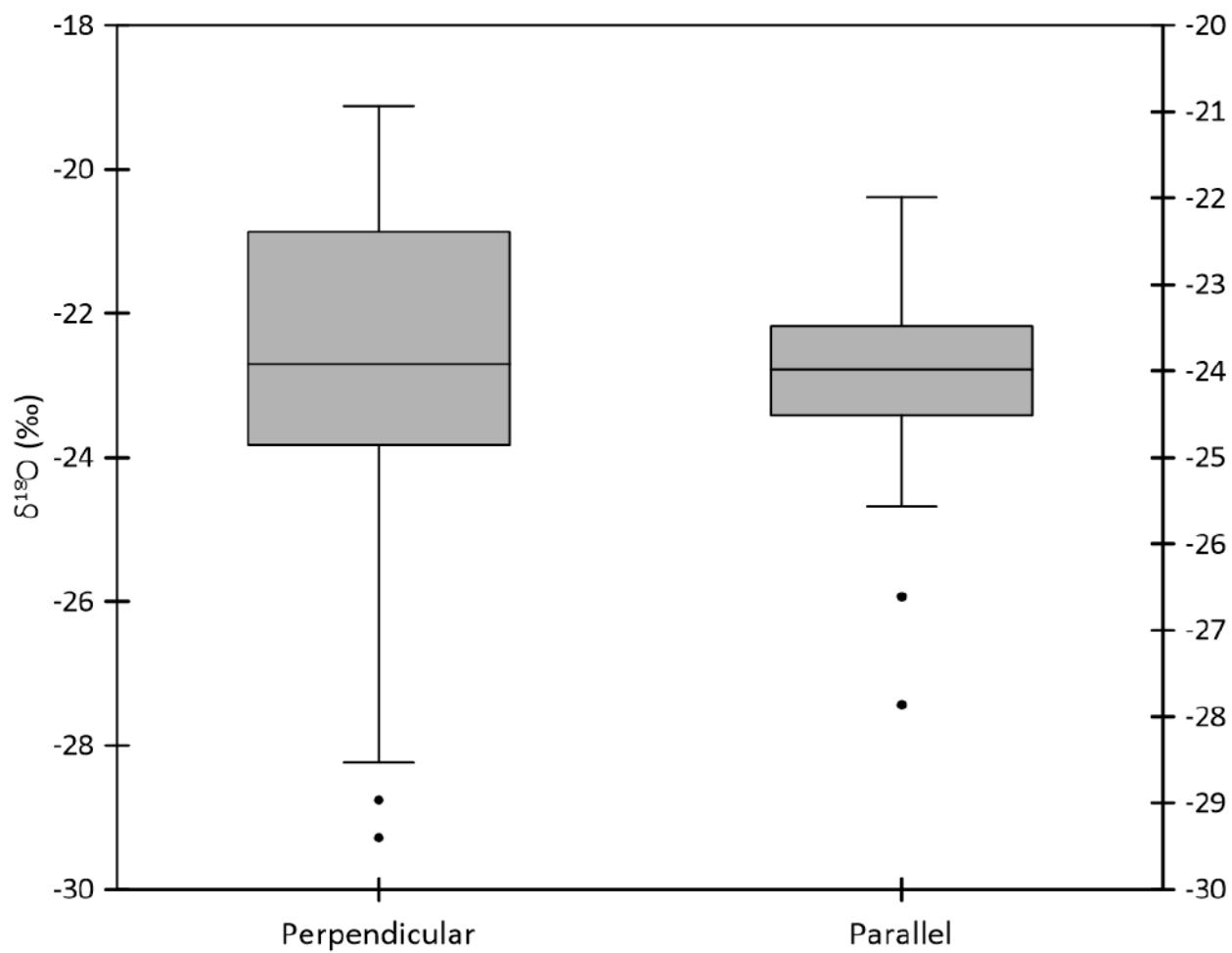


Figure 5.17: Boxplot of $\delta^{18}\text{O}$ between wedges that are perpendicular and parallel to the slope

5.6 Summary of Results

Analysis of the stable isotopes of precipitation at Inuvik indicated that there is no significant difference in $\delta^{18}\text{O}$ when considering the origin of the air mass. In the summer, however, storms from the South arrive with a significantly different temperature than western and northern storms. Similarly, western storms in the winter have a significantly different temperature than the other directions.

When assessing the $\delta^{18}\text{O}$:T relation, it is apparent there is minimal relation between the $\delta^{18}\text{O}$ and $\delta^{18}\text{T}$ when analyzing seasonal precipitation on a storm-by-storm basis. By analysing the trajectory of air masses prior to precipitation in Inuvik, it is clear that Pacific and Bering Sea air masses that cross the western Cordillera carry depleted precipitation to Inuvik in comparison with storms from the nearby Beaufort Sea.

The winter snowpack at Inuvik exhibits $\delta^{18}\text{O}$ values similar to winter precipitation, and there is no significant difference between forest and lake environments. $\delta^{18}\text{O}$ values from the winter snowpack were more depleted than those found in the anti-syngenetic ice wedges near Illisarvik.

Near-surface permafrost adjacent to Illisarvik indicates the location of the thaw unconformity on the western side of the basin. Below the unconformity, Pleistocene age ice is present as demonstrated by $\delta^{18}\text{O}$ values below -28‰ .

There is substantial variation in both characteristics and isotopic composition of anti-syngenetic ice wedges near Illisarvik. Wedges on the western side of the basin are narrower, with a higher minimum $\delta^{18}\text{O}$ value while eastern wedges may contain Pleistocene age ice.

6 Discussion

6.1 Introduction

This chapter reviews the primary objectives of this thesis with the results from Chapter 5. First, the isotopic composition of precipitation at Inuvik is addressed with special attention to storm origin and trajectory. Following this the $\delta^{18}\text{O}:T$ relation for precipitation at Inuvik is discussed. An analysis of the isotopic variation in near-surface permafrost and anti-syngenetic ice wedges at Illisarvik is then presented.

6.2 Stable Isotopes of Precipitation at Inuvik

Figure 5.6 and Table 5.6 illustrate the impact of orographic features on the isotopic composition of air masses that bring precipitation to Inuvik. Although the absolute difference in $\delta^{18}\text{O}$ concentrations between precipitation from air masses originating directly from the Arctic Ocean and those that traverse mountains is not significantly different at $\alpha = 0.05$, in summer there is a statistically significant difference in mean daily temperatures between these precipitation event groups (Table 5.6). The temperatures during winter precipitation are not significantly different at $\alpha = 0.05$ due to increased variation in the samples, as indicated by the standard error of the means (Table 5.6). This is despite the absolute difference in mean temperature between storms from different origins being similar to that of summer precipitation. Inuvik occupies a specific position, where the influence of fractionation during passage over the western Cordillera reduces $\delta^{18}\text{O}$ concentrations to values similar to those derived from the nearby cold Arctic Ocean.

The impact of the St. Elias Mountains on the isotopic composition of precipitation has been observed previously to result in a decrease of 0.6‰ in $\delta^{18}\text{O}$ for every 100 m of altitude gained (Holdsworth et al., 1991). Consequently, western and southern storms exhibit $\delta^{18}\text{O}$ concentrations similar to those of northern storms, despite being accompanied by warmer air temperatures. Air masses from the South and West traverse the Coast and St. Elias mountains upon landfall, and the Mackenzie Mountains further inland. Western storms approaching from Alaska typically pass over the Brooks Range and Richardson Mountains before reaching Inuvik. Occasionally, air masses from the West make landfall over eastern Siberia, traveling over the Chukotka Mountains as well.

As a result of these heavily fractionated air masses, Inuvik is situated so that storms from the North with colder temperatures are represented by a similar isotopic signal as found in warmer storms from the West and South. It is possible that substantial shifts in air mass circulation patterns could result in a change in regional temperature that would not be noticeable in the isotopic composition of precipitation.

6.2.1 Seasonality and D-excess

While meteoric water lines exhibit limited variation for each direction (Table 5.2), slight differences in derived slopes and D-excess are apparent for the different groups of storms. There is a distinct difference in D-excess for winter and summer precipitation (Table 5.1). Winter precipitation shows higher D-excess, likely due to low evaporation rates during formation and transit, contrasting with summer precipitation derived from increased evaporation in warmer conditions above the source. A similar interpretation is presented by Dansgaard (1964), Stewart (1975), and Jouzel and Merlivat (1984). The D-excess values for winter generally aligned with D-excess values for the anti-syngenetic ice wedges near Illisarvik (Tables 5.1 &

5.8).

The consistency of D-excess across air masses undergoing significant fractionation over mountains and those that do not, coupled with its clear distinction by season, strengthens its utility.

6.2.2 *Snowpack Changes*

Over several years, from 1991-1995, the snowpack near Inuvik in both lake and forest environments exhibited the characteristic isotopic signature of seasonal winter precipitation. There was no significant variation between these two environmental groups, and within-group variation was minimal, indicating similar isotopic conditions. Similar findings were reported by Ala-aho et al. (2021) in taiga snowpacks, although they observed considerable variation between tundra snowpacks at multiple sites in close proximity to each other. This suggests that the isotopic similarity observed in the sampled taiga environments cannot be assumed for those in the tundra.

Illisarvik, while currently in a tundra environment, was in a taiga environment as recently as 6.5 ± 0.1 ka (Mackay, 1992b). Consequently, it may be assumed that the close relationship between the snowpack and precipitation observed at Inuvik could have applied to the Illisarvik region when it was in a taiga environment. However, the recession of treeline since then may have led to significant isotopic variation in the snowpack. This should be confirmed through widespread snowpack sampling.

6.3 Permafrost Boreholes

Based on the observed shifts of $\delta^{18}\text{O}$ values in permafrost boreholes near Illisarvik (Fig. 5.8), the thaw unconformity formed during the Holocene Thermal Maximum (HTM) is identified at approximately 240 cm depth. Below the unconformity, the $\delta^{18}\text{O}$ values of $\leq -28\text{‰}$ are comparable with those reported in previous studies

indicative of Pleistocene permafrost in the region (Michel, 1982; Mackay, 1983b). The gradual increase in $\delta^{18}\text{O}$ values above the unconformity may be associated with aggradational permafrost, formed from a mixture of meteoric water and water in the paleo-active layer from thawed permafrost (Mackay, 1983a).

The D-excess values between 0 and 10‰ below the thaw unconformity suggest a predominantly meteoric water source, consistent with the D-excess values observed in Inuvik precipitation. However, above the unconformity, both boreholes exhibit a distinct change in D-excess. Borehole A shows negative D-excess values for the first 40 cm above the unconformity, returning to standard values around 10‰. In contrast, borehole B maintains consistent D-excess values at approximately 20‰ for nearly a meter above the unconformity before settling, similar to borehole A.

The divergence above the thaw unconformity from the consistent D-excess values below suggests that aggradational permafrost developed from water sources that had undergone significant fractionation after precipitation. This may have been a result of annual freeze-thaw cycles over millennia. The proximity of the two boreholes makes it difficult to assign a clear hypothesis as to why their D-excess values above the unconformity differ so greatly. Both boreholes exhibit similar patterns in $\delta^{18}\text{O}$ and δD despite different D-excess patterns, indicating that the isotopic signature remains similar despite the changes in D-excess.

6.4 Anti-Syngenetic Ice Wedges

Figures 4.7 and 4.10 illustrate substantial variations in the surface geometries of both the ground surface and ice wedges, along with their isotopic compositions at the eastern and western sites. The isotopic composition of the wedges falls within a similar range as that observed in the near-surface boreholes (−30‰ to −20‰).

Although the horizontal sampling was not as detailed as in other isotope studies of ice wedges (e.g., Opel et al., 2017; Holland et al., 2020), the lower resolution

presented here should still preserve general characteristics, though likely averaged over the ice of several veins from a series of cracking events.

6.4.1 *Individual Wedges*

Western Wedges

The three western wedges sampled exhibit distinct surface geometries (Fig. 6.1), isotopic profiles (Fig. 6.1 & 5.12), and D-excess profiles (Fig. 5.13). In T2 (parallel to contours) and T3 (perpendicular to contours), the lowest $\delta^{18}\text{O}$ values are near the edges, where the oldest ice is anticipated to be located (Mackay, 1990). However, T4 (perpendicular to contours) deviates from this pattern, with the lowest $\delta^{18}\text{O}$ values near the middle and the highest values near the edges. Potentially as a result of contamination near the edges of the wedge.

Despite little D-excess consistency across the wedges (Fig. 5.13), there is some uniformity in the ranges ($\sim 5\text{‰}$ - $\sim 15\text{‰}$). There is no consistent relationship between the D-excess and $\delta^{18}\text{O}$ values across the wedges, as sections with stable $\delta^{18}\text{O}$ values often show substantial D-excess variation (see T3, Fig. 5.12 & 5.13).

The $\delta^{18}\text{O}$ values within the western wedges are similar to $\delta^{18}\text{O}$ values in the paleo-active layer of the boreholes (Fig. 5.8). As previously discussed, ice in the aggradational permafrost originates from older water in the paleo-active layer, characterized by a lower $\delta^{18}\text{O}$ value, and summer meteoric water, with a higher $\delta^{18}\text{O}$ value (Mackay, 1983a). According to current understanding of ice-wedge development, wedge ice does not incorporate old active layer water, so the similarity in $\delta^{18}\text{O}$ values suggests that meteoric water from winter precipitation, of isotopic value between Holocene summer water and Pleistocene ice, served as a main source for the formation of wedge ice. However, the most enriched ice from all three wedges ($> -22\text{‰}$) corresponds well with the active-layer values from both permafrost boreholes, implying a relatively modern origin.

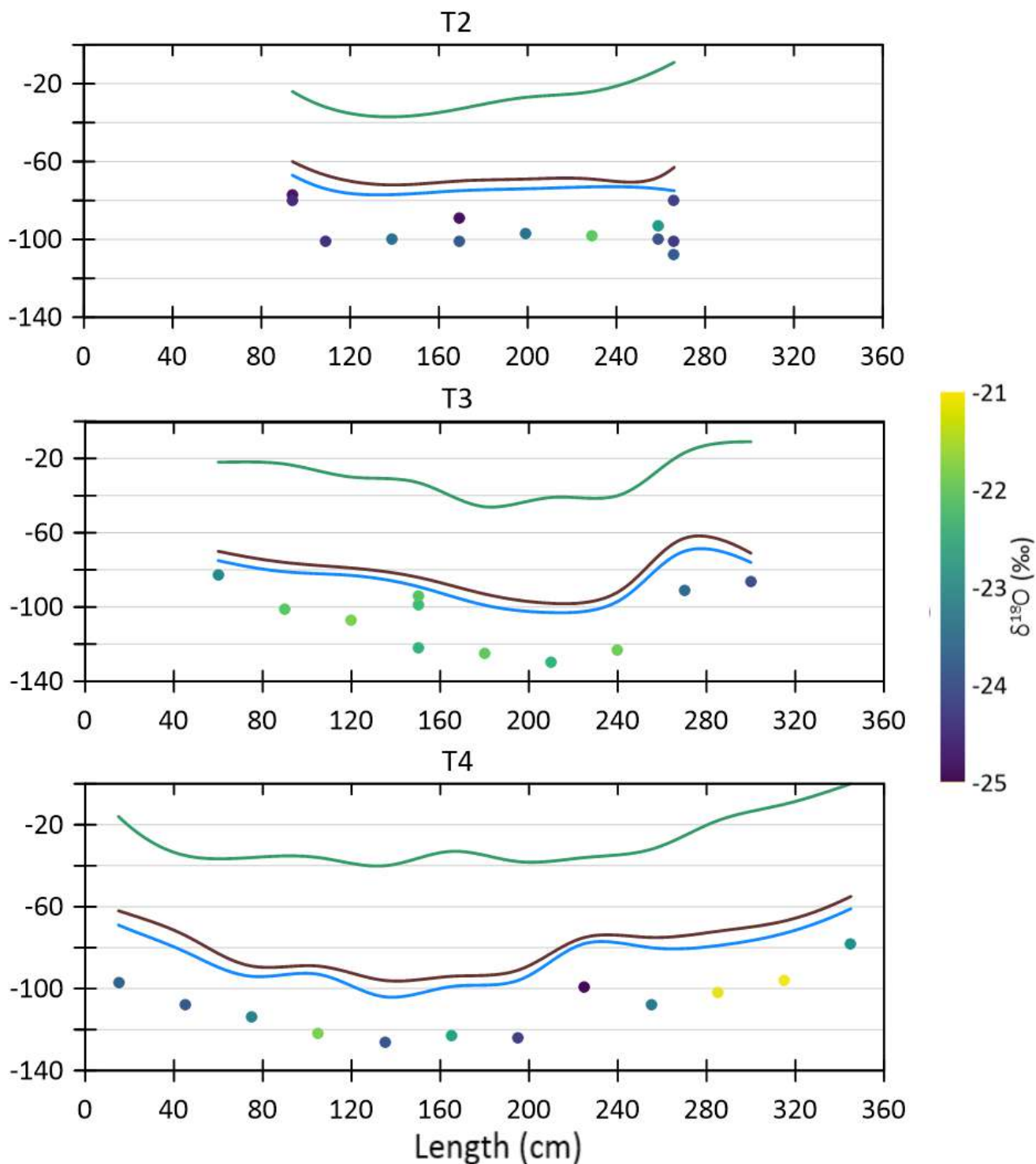


Figure 6.1: Cross-sectional profiles of the western ice wedges with $\delta^{18}\text{O}$ values of each sample. The green line (top) represents the ground surface, the brown line (middle) represents the thaw depth on the date of drilling, and the blue line (bottom) represents the top of the ice wedge. Each dot represents the location of a sample, and its colour denotes the $\delta^{18}\text{O}$ values. The scales are equal for all three wedges, and the y-axis is in reference to the highest ground surface point. For T3 and T4, the diagram is a cross-section of the trough. For T2, left is downslope.

The most distinctive geometry among the three western wedges is observed in T3 (Fig. 6.1). The depression in the surface level at the middle of the wedge mimics the trough geometry, and the rise of the right-most edge follows the ridge geometry (Fig. 6.1). While the depression may have formed due to melt flowing down the middle of the trough, no studies have explored the interaction between runoff and anti-syngenetic ice wedges.

Eastern Wedges

The surface geometry of the eastern sites exhibits greater variability than the western sites along with the ice-wedge surface geometries and the associated $\delta^{18}\text{O}$ profiles (Fig. 4.10 & 5.14).

Both E1 (perpendicular to contours) and E2 (parallel to contours) contain sections wider than one meter with a consistent isotopic signature, although both still have areas with considerable variation (Fig. 5.14).

Samples from these edges were taken a few centimeters into the wedge (Fig. 6.2), likely explaining their enriched values (Fig. 5.14 & 6.2), as further discussed in section 6.4.4. E3 (perpendicular to contours) is the only sampled wedge that contains what may definitively be assumed as Pleistocene ice, present near both edges (Fig. 5.14 & 6.2). Apart from these two samples, the middle portion of the wedge exhibits $\delta^{18}\text{O}$ values similar to the other two eastern wedges.

The $\delta^{18}\text{O}$ values in the eastern wedges, while generally lower than those on the western side of the lake basin, are also similar to values above the thaw unconformity within the near-surface permafrost (Fig. 5.8). The highest $\delta^{18}\text{O}$ values within the wedges are only slightly more depleted than that of water in the active layer of the boreholes.

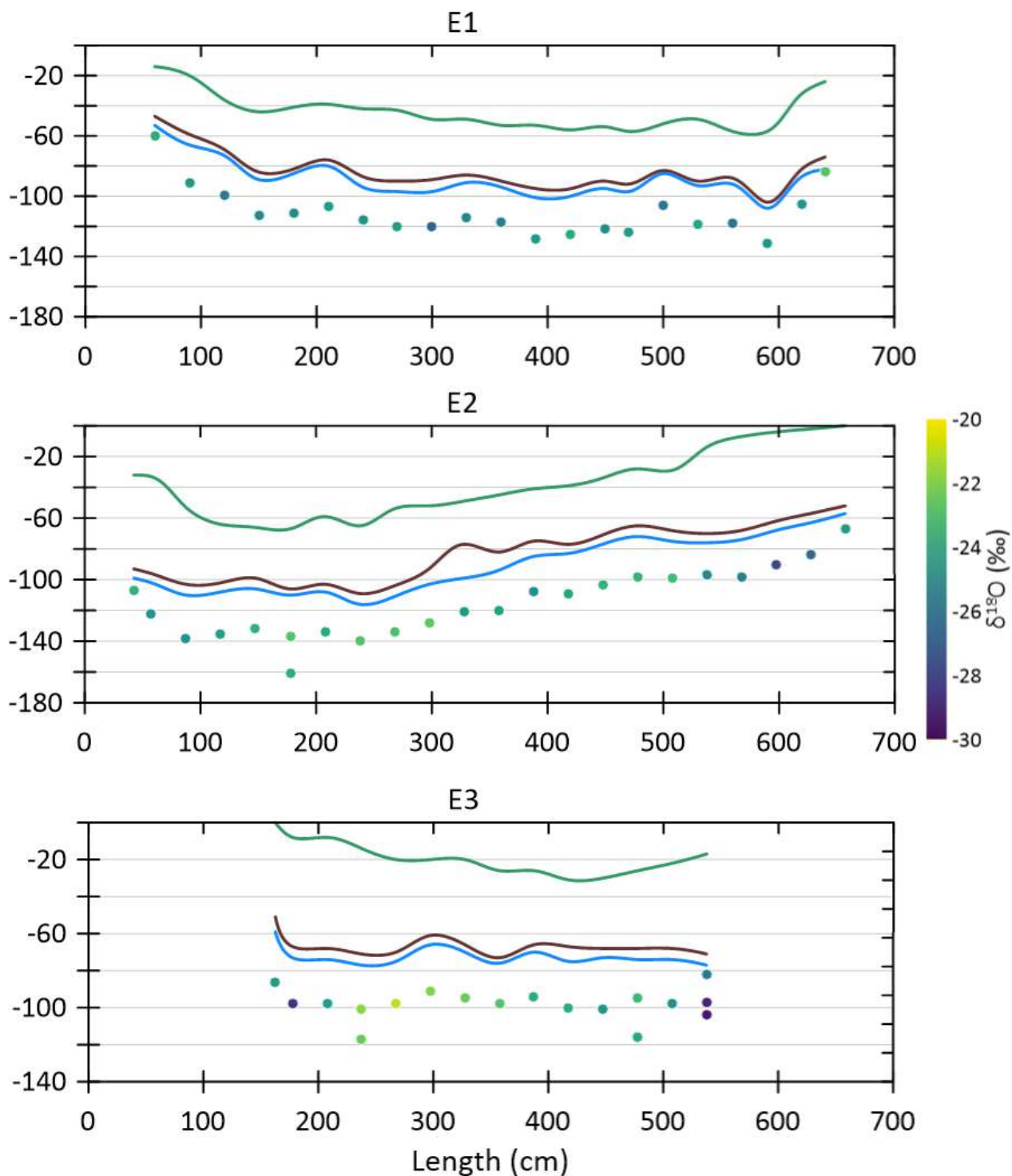


Figure 6.2: Cross-sectional profiles of the eastern ice wedges with $\delta^{18}O$ values of each sample. The green line (top) represents the ground surface, the brown line (middle) the thaw depth on the date of drilling, and the blue line (bottom) the top of the ice wedge. Each dot represents the location of a sample, and its colour denotes the $\delta^{18}O$ values. The scales are equal for E1 and E2, however E3 has an independent y-axis. For E1 and E3, the diagram is cross-section of the trough. For E2, left is downslope.

In contrast to the western wedges, the D-excess pattern across the eastern wedges shows some consistency (Fig. 5.15). All three wedges exhibit similar patterns in D-excess distribution. Near each edge, values are variable, with a section of relatively low values towards the middle of the wedge. The middle section has a relatively high D-excess (Fig. 5.15). Each of these sections is roughly 1 m across the wedge. Consistent D-excess values have been suggested as an indicator of a constant moisture source (Meyer et al., 2010; Wetterich et al., 2021). Moreover, the locations with consistent D-excess generally have consistent $\delta^{18}\text{O}$ values.

6.4.2 $\delta D:\delta^{18}\text{O}$ Relation

Lauriol et al. (1995) propose that a lower slope in most $\delta D:\delta^{18}\text{O}$ relations for epigenetic ice wedges may be due to late-season meltwater entering the crack. This meltwater, having undergone more evaporation than early-season runoff, leads to lower slopes. Generally, a lower slope in the $\delta D:\delta^{18}\text{O}$ relation suggests more fractionation has occurred under non-equilibrium conditions.

In the western Canadian Arctic, $\delta D:\delta^{18}\text{O}$ slopes for epigenetic ice wedges have typically been lower than the GMWL or LMWL (e.g., St-Jean et al., 2011; Holland et al., 2020). Holland et al. (2020) presented four wedges from the Tuktoyaktuk Coastlands all with $\delta D:\delta^{18}\text{O}$ slopes lower than the GMWL.

The regressions from wedge ice derived using a functional fit cannot be directly compared to the GMWL but may be compared to the functional fit of precipitation at Inuvik. Unlike examples from the literature, most of the anti-syngenetic ice wedges sampled near Illisarvik exhibit slopes equal to or greater than those observed in Inuvik's winter precipitation (7.47, Tables 5.1 & 6.1). When compared to winter precipitation, three wedges (T2, T4, E3) have substantially higher slopes, while T3 shows a nearly identical relationship (Table 6.1). After removing outliers, five wedges show functional fit slopes greater than precipitation (Table 6.1).

Table 6.1: Least-Squares Linear (LSL) and Functional Fit (FF) relations for each wedge and all combined samples when outliers have been removed. Also includes the regressions for winter precipitation at Inuvik between 2015 - 2018

Wedge	Least-Squares Linear (LSL)	Functional Fit (FF)
T2	$\delta D = 5.41\delta^{18}\text{O} - 52.44$	$\delta D = 7.48\delta^{18}\text{O} - 2.87$
T3	$\delta D = 7.17\delta^{18}\text{O} - 10.23$	$\delta D = 7.90\delta^{18}\text{O} + 6.07$
T4	$\delta D = 7.99\delta^{18}\text{O} + 10.14$	$\delta D = 8.47\delta^{18}\text{O} + 21.30$
E1	$\delta D = 7.04\delta^{18}\text{O} - 13.18$	$\delta D = 8.45\delta^{18}\text{O} + 21.88$
E2	$\delta D = 6.55\delta^{18}\text{O} - 25.76$	$\delta D = 7.14\delta^{18}\text{O} - 11.51$
E3	$\delta D = 8.32\delta^{18}\text{O} + 18.39$	$\delta D = 8.42\delta^{18}\text{O} + 20.96$
All Wedges	$\delta D = 7.70\delta^{18}\text{O} + 2.6$	$\delta D = 7.99\delta^{18}\text{O} + 9.66$
Winter Precipitation	$\delta D = 7.42\delta^{18}\text{O} - 5.6$	$\delta D = 7.48\delta^{18}\text{O} - 4.29$

These results contrast with previous studies of epigenetic wedges, suggesting that the infilling process in anti-syngenetic wedges may differ significantly.

Removing outliers also reduces variation between the wedges, indicating that they are being infilled by a similar moisture source undergoing comparable processes. None of the revised slopes are significantly different from one another. The changes are primarily due to the removal of extremely negative Pleistocene ice values and particularly high values associated with modern conditions. This may imply that the regressions found in Table 6.1 are representations of Holocene ice.

The two lowest slopes (Table 6.1) correspond to the two wedges oriented parallel to the contours. Although the regression coefficients are not significantly different from those of the other wedges, it is possible that orientation along the hillslope has an effect on the timing and nature of the infilling and freezing process.

6.4.3 Downslope Movement and Ice Wedge Depth

The position of anti-syngenetic ice wedges on hillslopes means that there is continuous movement of material above the wedges (Mackay, 1990; Burn et al., 2021). Previous studies, such as those on Pelly Island, have shown the impact of downslope movement on the depth to the thaw unconformity (Fig. 6.3) (Burn, 1997).

The sampled wedges in this thesis were all collected near the summits of their respective hillslopes due to the significant water flow occurring further downslope, placing the sampling location near the point of maximum erosion along the hillslope (Burn, 1997). The highly depleted ice near the edge of E3 likely formed before the HTM, possibly during the Pleistocene, as suggested by Mackay (1983b) for ice of similar isotopic composition. It is the only wedge ice sampled with such values. This ice would have remained unaffected during the HTM.

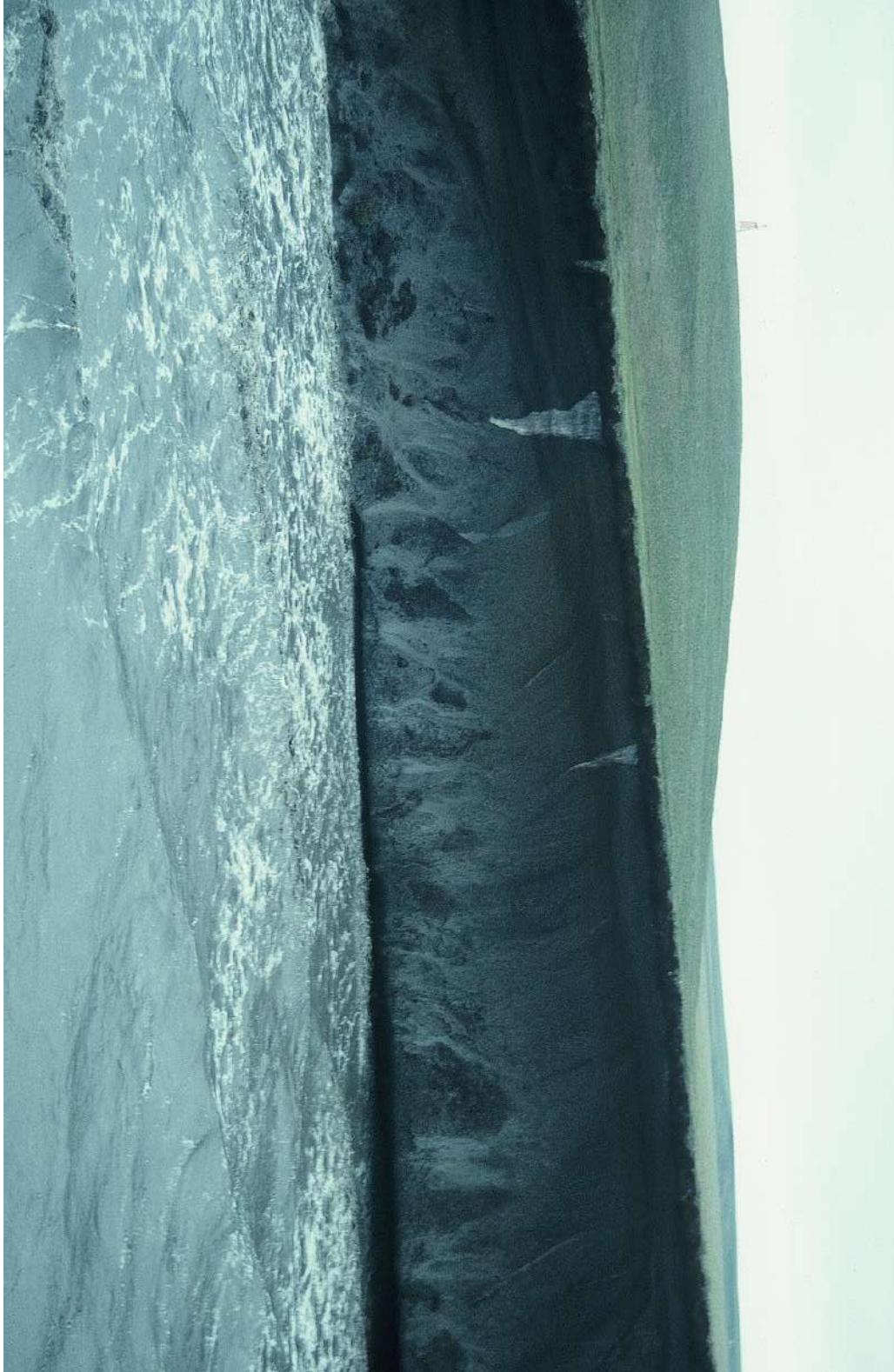


Figure 6.3: A coastal exposure on Pelly Island demonstrating the downslope movement of soil. The paleo-active layer is denoted by the darker soils near the surface and progressively gets deeper at the lower slopes. Photo reproduced with permission from C.R. Burn.

The presence of this ice at the top of the permafrost can be explained in three possible ways. First, E3 may be the only wedge sampled that began development during the Pleistocene. Second, the sampling resolution was coarse and any small amount of Pleistocene ice within a sample had its isotopic signal obscured by the majority of the sample being comprised primarily of Holocene aged ice. Third, the rate of denudation at E3 may have been less than the other wedges. Recalling Fig. 1.2 as proposed by Mackay (1990), this would suggest that all other sampled wedges have eroded past the oldest ice along their edges, whereas E3 contains a small portion of what would be its oldest ice.

6.4.4 Wedge-Ice Composition

The conventional understanding of wedge ice suggests its primary derivation from winter precipitation (Mackay, 1972b, 1983b, 1984b; Lauriol et al., 1995). St-Jean et al. (2011) proposed that during the Late Pleistocene in cold and dry environments, ice-wedge development may have occurred mainly through snow densification, while in the Holocene, development was through the freezing of liquid water. Additional research has indicated minimal fractionation during the freezing of water within the wedge (Michel, 1990, 2011), however, there is some evidence that progressive freezing may cause ice at lower depths to be more depleted (Campbell-Heaton et al., 2021). Though samples for this thesis were taken near the surface of the wedge, thus negating the progressive freezing, it is possible that the current surface is not the original surface of the wedge, as discussed in the previous section. This must be considered when analyzing the isotopic composition of anti-syngenetic ice wedges.

Considering that the aggregate snowpack at Inuvik exhibits similar $\delta^{18}\text{O}$ values as winter precipitation, and wedge ice is assumed to be derived from snowpack melt with little fractionation upon freezing in the wedge, it would be assumed that they should wedge ice and winter precipitation should consist of similar

isotopic composition. However, wedge ice is consistently reported with an isotopic composition that is significantly more enriched than winter precipitation (e.g., Mackay, 1983b; Holland et al., 2020; Wetterich et al., 2023).

Holland et al. (2020) propose post-deposition fractionation within the snowpack as the cause for this enrichment, but at Inuvik the aggregate snowpack composition was similar to winter precipitation. Furthermore, it is unlikely that the wedges described in the literature, and by this thesis, would have been formed during conditions where the isotopic composition of precipitation was similar to that of today. Therefore, perhaps the water entering the wedge is a mixture of winter and spring moisture. This hypothesis must be tested through further field investigations to fully understand the ice-wedge infilling processes.

Detailed stratigraphic sampling of the Illisarvik snowpack would be necessary to determine if different layers within the snowpack are suitably enriched to justify the values observed within wedge ice. Additionally, a modern ice wedge within the Illisarvik drained lake basin should be sampled (Mackay and Burn, 2002b). This wedge has developed entirely post-drainage, and a comparative analysis can be performed with the precipitation data gathered by C.R. Burn between 1985-95.

Ice Adjacent to Wedge

E1 and E2 show an increase in isotope values in the final sample at either end. These values may be attributed to a mixture with active layer water, since the samples were taken within a few centimeters of the surface. Conversely, samples near the surface from the western wedges generally exhibited the lowest values within the wedge.

An increase in the volume of excess ice was observed near the ice wedges of the eastern sites. The modern pool ice that was sampled from next to E1 had a $\delta^{18}\text{O}$ value of -18.7‰ , implying an origin from summer precipitation. This was determined as pool ice due to its clear nature, contrasting the cloudy appearance of

wedge ice. Given the amount of excess ice near the ice wedges, particularly near the edge of E1, and as noted by Mackay (1997) during the drainage of Illisarvik, it is possible that the ice assumed to represent the edge of the three eastern wedges may not originate from the wedge itself.

D-excess and Ground Ice Origin

The D-excess, a parameter with various applications in understanding groundwater dynamics (Clark and Fritz, 1997; Kendall and Caldwell, 1998; Machida et al., 2022), has not been consistently used for interpreting ground ice. While some studies have analyzed D-excess within ice wedges (e.g., Holland et al., 2020; Campbell-Heaton et al., 2021; Wetterich et al., 2021) and near-surface permafrost (e.g., Lacelle et al., 2014; Takano et al., 2019), there is a lack of uniformity in its application.

Some studies of near-surface permafrost have demonstrated consistent D-excess throughout the entire borehole (Lacelle et al., 2014; Takano et al., 2019). At Illisarvik, D-excess is consistent below the thaw unconformity, however, a noticeable shift in D-excess occurs above the unconformity (Fig. 5.10). This discrepancy may stem from substantial fractionation resulting from repeated freezing and thawing of water within the active layer. The aggradational ice, a mixture of meteoric water and older active-layer water (Mackay, 1983a), perhaps underwent unknown fractionation before becoming incorporated in permafrost.

The D-excess values in the top 150 cm of ground near Inuvik typically range between 0 and 10‰ (Lacelle et al., 2014), resembling values below the thaw unconformity at Illisarvik (Fig. 5.10). These values are consistent with a combination of winter and summer precipitation at Inuvik and the GMWL (Table 5.1). However, the distinct change in D-excess above the thaw unconformity, as well as the disparity between two closely situated boreholes, suggest D-excess may not be an effective method for determining moisture sources for permafrost. Figures 5.8 and 5.10 illustrate that the $\delta^{18}\text{O}$ trend within ground ice remains similar despite significant

differences in D-excess values.

The D-excess in wedge ice is variable (Campbell-Heaton et al., 2021; Wetterich et al., 2021), as also observed across anti-syngenetic ice wedges near Illisarvik. No discernible pattern or correlation exists between D-excess and $\delta^{18}\text{O}$ along the western wedges. The D-excess in ice wedges has typically been used to indicate shifting moisture sources (Meyer et al., 2010; Wetterich et al., 2021), but it is difficult to identify consistent moisture source patterns in the Illisarvik wedges. Despite having the same precipitation source, both western and eastern wedges lack consistent D-excess patterns. Therefore, relying on D-excess within these wedges alone cannot determine the moisture source.

7 Summary and Conclusions

7.1 Summary of Results

This thesis presents a unique analysis of precipitation isotopes at Inuvik and an isotopic analysis of anti-syngenetic ice wedges and adjacent permafrost from a site in western Arctic Canada. By utilizing the NOAA HYSPLIT back trajectory tool, the spatial origins of 111 storms at Inuvik were determined. The results from this analysis were combined with data on the winter snowpack at Inuvik and permafrost samples from near the Illisarvik drained lake basin. Anti-syngenetic ice-wedge samples provide geometrical observations and isotopic values to better understand the source of water entering the wedge.

7.1.1 *Precipitation*

The NOAA HYSPLIT analysis revealed that 60, 26, and 13 % of storms at Inuvik between August 2015 and August 2018 originated from the North, South, and West, respectively. The remaining 1% originated from the East. Northern storms developed over the Beaufort Sea and travelled a short distance before reaching Inuvik. Southern storms generally originated over the Pacific Ocean, while western air masses typically generated above the Bering or Chukchi Seas. Both southern and western storms travelled across mountains prior to reaching Inuvik but arrived in warmer air masses than those from the North.

While there was no statistical difference in $\delta^{18}\text{O}$ values between storms from the South or West and those from the North, there is substantial difference in associated temperature. This is a result of increased fractionation occurring in the southern and western air masses as they passed over the mountains, depleting the air mass of

the heavy isotopes and causing the isotopic values to be similar to those generated in colder environments. Seasonality in precipitation at Inuvik was shown through a substantial difference in D-excess, as summer precipitation had values near 3‰ and winter near 9‰.

7.1.2 Near-Surface Permafrost and Anti-Syngenetic Ice Wedges

The thaw unconformity identified on the western side of the Illisarvik drained lake basin was at a depth of approximately 240 cm, as demonstrated by Pleistocene $\delta^{18}\text{O}$ values in ice below this depth and a gradual increase in $\delta^{18}\text{O}$ values above it. The D-excess above the unconformity significantly differed from that below, a result attributed to the influence of multiple freeze-thaw cycles on the isotopic composition of the water within the soil.

A notable geometric distinction existed between the anti-syngenetic ice wedges on the western side of the Illisarvik drained lake basin and those on the eastern side. The western wedges exhibit a narrower profile (approximately 3 m) with relatively flat surfaces, while the eastern wedges are notably wider (approximately 6 m) and have uneven surfaces.

The isotopic values within the wedges suggest Holocene development, as $\delta^{18}\text{O}$ values were generally within the range identified by Mackay (1983b) with Holocene wedge ice. There were, however, $\delta^{18}\text{O}$ values near the edges of E3 that reached as low as -29‰ , suggesting the presence of Pleistocene-age ice near the edges of the eastern wedges.

7.2 Conclusions

There are seven conclusions that can be drawn from this analysis of stable isotopes in precipitation and anti-syngenetic ice wedges near Inuvik and Illisarvik:

1. In the western Canadian Arctic, the stable water isotopic composition of precipitation appears insensitive to different source areas for storm origin because fractionation during transport over mountains to the South or West induces a similar stable isotopic concentration as in precipitation coming from the Arctic Ocean.
2. There is distinct seasonality in the isotope signal represented by the D-excess of modern precipitation. The D-excess of winter precipitation is near 9‰ and of summer precipitation is near 3‰.
3. The winter snowpack near Inuvik represents seasonal winter precipitation at Inuvik. Further, snow pits in lake environments exhibit statistically similar isotope values to those in the forest, indicating uniformity of snowpack isotope values in the region.
4. Stable water isotopes from boreholes on hillslopes near Illisarvik have indicated a thaw unconformity near 240 cm.
5. The anti-syngenetic ice wedges on the western side of the Illisarvik drained lake basin have been developing since the HTM.
6. Some anti-syngenetic ice wedges on the eastern side of the Illisarvik drained lake basin may have been developing since the Pleistocene, as indicated by ice with $\delta^{18}\text{O} \leq -28\text{‰}$. The ice is limited to the edges and has almost been fully removed due to degradation of the wedges by hillslope movement.
7. Ice-wedge ice may not be comprised exclusively of winter precipitation, as it is consistently more enriched compared to winter precipitation and snowpack composition.

7.3 Research Implications and Future Areas of Study

Given this is the first study to assess the isotopic composition of ice wedges explicitly identified as anti-syngenetic, there are multiple pathways for future research that have arisen. First, more specific research into the water that infiltrates thermal contraction cracks must be completed. The theory of ice-wedge development has considered snowmelt as the source for wedge ice (Mackay, 1972b, 1974a). This thesis, as in other research, has found wedge ice to be less depleted in ^{18}O than modern winter precipitation in the area (e.g., Mackay, 1983b; Michel, 1990; Holland et al., 2020; Wetterich et al., 2023). Therefore a revision or accommodation to the theory of ice-wedge formation might be developed using direct experimentation to determine the source water for wedge ice. Investigation into the moisture that infiltrates the wedges is required to understand why they are substantially more enriched than winter precipitation.

There will need to be more research into slope dynamics, specifically above ice wedges, in order to fully understand the variation along these ice wedges. Detailed analysis of the surface using high resolution digital elevation models may provide information on how the surface is influenced by the wedges and may indicate differing water paths for each wedge. This would build upon the research by Chiasson-Poirier et al. (2020) and Deng et al. (2022) into thaw on hillslopes.

7.4 Study Limitations

This thesis has unfortunately been affected by various factors which delayed and interrupted field work. First, a COVID-19 diagnosis the day prior to transit to the field in July 2022 delayed arrival at Illisarvik until August, cutting field time in half. This resulted in significantly less samples than originally planned and a modification of the original sampling methodology. Field assistant Aidan Gill and C.R. Burn

were able to collect the borehole samples and prepare camp so upon arrival drilling into the anti-syngenetic ice wedges could commence immediately. Second, field work was planned for August 2023, however these plans were cancelled due to wildfires in the area. This second field season was planned to collect extra samples to provide more context along each wedge. A detailed geometry of the bottom of the ice wedge would have been gathered, giving significantly more information on the shape of the wedge.

References

- Abell, J. T., G. Winckler, R. F. Anderson, and T. D. Herbert (2021), Poleward and weakened westerlies during Pliocene warmth, *Nature*, 589(7840), 70–75, 75A–75L, doi:10.1038/s41586-020-03062-1.
- Akers, P. D., J. M. Welker, and G. A. Brook (2017), Reassessing the role of temperature in precipitation oxygen isotopes across the eastern and central United States through weekly precipitation-day data, *Water Resources Research*, 53(9), 7644–7661, doi:10.1002/2017WR020569, _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2017WR020569>.
- Ala-aho, P., D. Tetzlaff, J. P. McNamara, H. Laudon, P. Kormos, and C. Soulsby (2017), Modeling the isotopic evolution of snowpack and snowmelt: Testing a spatially distributed parsimonious approach, *Water Resources Research*, 53(7), 5813–5830, doi:10.1002/2017WR020650.
- Ala-aho, P., J. M. Welker, H. Bailey, S. Højlund Pedersen, B. Kopec, E. Klein, M. Mellat, K.-R. Mustonen, K. Noor, and H. Marttila (2021), Arctic snow isotope hydrology: A comparative snow-water vapor study, *Atmosphere*, 12(2), 150, doi:10.3390/atmos12020150.
- Andersen, T. S., P. A. Jardine, and C. R. Burn (2021), Long-Term (2000–2017) Response of Lake-Bottom Temperatures and Talik Configuration to Changes in Climate at Two Adjacent Tundra Lakes, Western Arctic Coast, Canada, pp. 1–13, in *Permafrost 2021*, American Society of Civil Engineers, Reston, VA, doi: 10.1061/9780784483589.001.

- Andersland, O., and B. Ladanyi (2004), *Frozen ground engineering*, 2 ed., American Society of Civil Engineers, Reston, VA.
- Araguás-Araguás, L., K. Froehlich, and K. Rozanski (2000), Deuterium and ^{18}O isotope composition of precipitation and atmospheric moisture, *Hydrological Processes*, 14(8), 1341–1355, doi:10.1002/1099-1085(20000615)14:8<1341::AID-HYP983>3.0.CO;2-Z.
- Bailey, H. L., D. S. Kaufman, A. C. G. Henderson, and M. J. Leng (2015), Synoptic scale controls on the $\delta^{18}\text{O}$ in precipitation across Beringia, *Geophysical Research Letters*, 42(11), 4608–4616, doi:10.1002/2015GL063983, _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2015GL063983>.
- Bateman, M. D., and J. B. Murton (2006), The chronostratigraphy of Late Pleistocene glacial and periglacial aeolian activity in the Tuktoyaktuk Coastlands, NWT, Canada, *Quaternary Science Reviews*, 25(19-20), 2552–2568, doi:10.1016/j.quascirev.2005.07.023.
- Benetti, M., G. Reverdin, C. Pierre, L. Merlivat, C. Risi, H. C. Steen-Larsen, and F. Vimeux (2014), Deuterium excess in marine water vapor: Dependency on relative humidity and surface wind speed during evaporation, *Journal of Geophysical Research: Atmospheres*, 119(2), 584–593, doi:10.1002/2013JD020535.
- Berger, A., and M. Loutre (1991), Insolation values for the climate of the last 10 million years, *Quaternary Science Reviews*, 10(4), 297–317, doi:10.1016/0277-3791(91)90033-Q.
- Bradley, R. S. (2014), *Paleoclimatology: Reconstructing climates of the Quaternary*, 3 ed., Elsevier Science & Technology, Saint Louis, USA.

- Burn, C. R. (1997), Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 34(7), 912–925, doi:10.1139/e17-076.
- Burn, C. R. (1998), The active layer: Two contrasting definitions, *Permafrost and Periglacial Processes*, 9, 411–416.
- Burn, C. R. (2004), A field perspective on modelling ‘single-ridge’ ice-wedge polygons, *Permafrost and Periglacial Processes*, 15(1), 59–65, doi:10.1002/ppp.475.
- Burn, C. R., and S. V. Kokelj (2009), The environment and permafrost of the Mackenzie Delta area, *Permafrost and Periglacial Processes*, 20(2), 83–105, doi:10.1002/ppp.655.
- Burn, C. R., and M. G. Maxwell (1993), Proper determination of the $\delta^{18}\text{O}$ – δD relationship for ice and water by least-squares cubic regression, *Canadian Journal of Earth Sciences*, 30(1), 109–112, doi:10.1139/e93-010.
- Burn, C. R., and Y. Zhang (2009), Permafrost and climate change at Herschel Island (Qikiqtaruq), Yukon Territory, Canada, *Journal of Geophysical Research: Earth Surface*, 114(F2), doi:10.1029/2008JF001087.
- Burn, C. R., and Y. Zhang (2010), Sensitivity of active-layer development to winter conditions north of treeline, Mackenzie delta area, western Arctic coast, pp. 1458–1465, in *Proceedings, 6th Canadian Permafrost Conference*, Canadian Geotechnical Society, Calgary, AB; Richmond, BC.
- Burn, C. R., F. A. Michel, and M. W. Smith (1986), Stratigraphic, isotopic, and mineralogical evidence for an early Holocene thaw unconformity at Mayo, Yukon Territory, *Canadian Journal of Earth Sciences*, 23(6), 794–803, doi:10.1139/e86-081, publisher: NRC Research Press.

- Burn, C. R., A. G. Lewkowicz, and M. A. Wilson (2021), Long-term field measurements of climate-induced thaw subsidence above ice wedges on hillslopes, western Arctic Canada, *Permafrost and Periglacial Processes*, 32(2), 261–276, doi:10.1002/ppp.2113.
- Campbell-Heaton, K., D. Lacelle, and D. Fisher (2021), Ice wedges as winter temperature proxy: Principles, limitations and noise in the $\delta^{18}\text{O}$ records (an example from high Arctic Canada), *Quaternary Science Reviews*, 269, 107–135, doi:10.1016/j.quascirev.2021.107135.
- Charles, C. D., D. Rind, J. Jouzel, R. D. Koster, and R. G. Fairbanks (1994), Glacial-interglacial changes in moisture sources for Greenland: Influences on the ice core record of climate, *Science* (Washington, D.C.); (United States), 263:5146, doi:10.1126/science.263.5146.508.
- Chiasson-Poirier, G., J. Franssen, M. J. Lafrenière, D. Fortier, and S. F. Lamoureux (2020), Seasonal evolution of active layer thaw depth and hillslope-stream connectivity in a permafrost watershed, *Water Resources Research*, 56(1), e2019WR025,828, doi:10.1029/2019WR025828.
- Clark, I. D., and P. Fritz (1997), *Environmental isotopes in hydrogeology*, CRC Press/Lewis Publishers, Boca Raton, FL.
- Craig, H. (1961), Isotopic variations in meteoric waters, *Science*, 133(3465), 1702–1703.
- Dallimore, S. R., S. A. Wolfe, J. V. Matthews Jr., and J.-S. Vincent (1997), Mid-Wisconsinan eolian deposits of the Kittigazuit Formation, Tuktoyaktuk Coastlands, Northwest Territories, Canada, *Canadian Journal of Earth Sciences*, 34(11), 1421–1441, doi:10.1139/e17-116.

- Dansgaard, W. (1953), The abundance of ^{18}O in atmospheric water and water vapour, *Tellus*, 5(4), 461–469, doi:10.1111/j.2153-3490.1953.tb01076.x.
- Dansgaard, W. (1954), The ^{18}O abundance in fresh water, *Geochimica et Cosmochimica Acta*, 6(5), 241–260, doi:10.1016/0016-7037(54)90003-4.
- Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, 16(4), 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x.
- Deng, K., C. Azorin-Molina, S. Yang, C. Hu, G. Zhang, L. Minola, and D. Chen (2022), Changes of Southern Hemisphere westerlies in the future warming climate, *Atmospheric Research*, 270, 106,040, doi:10.1016/j.atmosres.2022.106040.
- Dütsch, M., S. Pfahl, and H. Sodemann (2017), The impact of nonequilibrium and equilibrium fractionation on two different deuterium excess definitions, *Journal of Geophysical Research: Atmospheres*, 122(23), 12,732–12,746, doi:10.1002/2017JD027085.
- Earman, S., A. R. Campbell, F. M. Phillips, and B. D. Newman (2006), Isotopic exchange between snow and atmospheric water vapor: Estimation of the snowmelt component of groundwater recharge in the southwestern United States, *Journal of Geophysical Research: Atmospheres*, 111(D9), doi:10.1029/2005JD006470.
- ECCC (2023), Canadian Climate Normals 1981-2010 Station Data - Climate - Environment and Climate Change Canada.
- Feng, F., Z. Li, M. Zhang, S. Jin, and Z. Dong (2013), Deuterium and ^{18}O in precipitation and atmospheric moisture in the upper Urumqi River Basin, eastern Tianshan Mountains, *Environmental Earth Sciences*, 68(4), 1199–1209, doi:10.1007/s12665-012-1820-y.

- Fortier, D., and M. Allard (2005), Frost-cracking conditions, Bylot Island, eastern Canadian Arctic archipelago, *Permafrost and Periglacial Processes*, 16(2), 145–161, doi:10.1002/ppp.504.
- French, H. M. (2007), *The Periglacial Environment*, 3 ed., John Wiley and Sons.
- Fricke, H. C., and J. R. O’Neil (1999), The correlation between $^{18}\text{O}/^{16}\text{O}$ of meteoric water and surface temperature: its use in investigating terrestrial climate change over geologic time, *Earth and Planetary Science Letters*, 170(3), 181–196, doi: 10.1016/S0012-821X(99)00105-3.
- Fritz, M., H. Meyer, S. Wetterich, and J. McAlister (2021), Stable oxygen ($\delta^{18}\text{O}$) and hydrogen (δD) isotopes in precipitation from Inuvik in the western Canadian Arctic (2015–2018), *PANGAEA*, doi:10.1594/PANGAEA.935027.
- Fritz, M., S. Wetterich, J. McAlister, and H. Meyer (2022), A new local meteoric water line for Inuvik (NT, Canada), *Earth System Science Data*, 14(1), 57–63, doi:10.5194/essd-14-57-2022.
- Funder, S., H. Goosse, H. Jepsen, E. Kaas, K. H. Kjær, N. J. Korsgaard, N. K. Larsen, H. Linderson, A. Lyså, P. Möller, J. Olsen, and E. Willerslev (2011), A 10,000-year record of Arctic Ocean sea-ice variability—view from the Beach, *Science*, 333(6043), 747–750, doi:10.1126/science.1202760.
- Gat, J. R. (1996), Oxygen and Hydrogen isotopes in the hydrologic cycle, *Annual Review of Earth and Planetary Sciences*, 24, 39.
- Gat, J. R., and R. Gonfiantini (1981), Deuterium and ^{18}O in the water cycle, IAEA.
- Gonfiantini, R. (1978), Standards for stable isotope measurements in natural compounds, *Nature*, 271(5645), 534–536, doi:10.1038/271534a0, number: 5645 Publisher: Nature Publishing Group.

- Harris, S. A., H. M. French, J. A. Heginbottom, G. H. Johnston, B. Ladanyi, D. C. Sego, and R. O. van Everdingen (1988), Glossary of permafrost and related ground-ice terms, , National Research Council of Canada. Associate Committee on Geotechnical Research. Permafrost Subcommittee.
- Hoefs, J. (2015), *Stable isotope geochemistry*, 7 ed., Springer Berlin Heidelberg, New York, NY.
- Holdsworth, G., S. Fogarasi, and H. R. Krouse (1991), Variation of the stable isotopes of water with altitude in the Saint Elias Mountains of Canada, *Journal of Geophysical Research: Atmospheres*, 96(D4), 7483–7494, doi:10.1029/91JD00048.
- Holland, K. M., T. J. Porter, D. G. Froese, S. V. Kokelj, and C. A. Buchanan (2020), Ice-wedge evidence of Holocene winter warming in the Canadian Arctic, *Geophysical Research Letters*, 47(12), e2020GL087942, doi:10.1029/2020GL087942.
- Jouzel, J., and L. Merlivat (1984), Deuterium and ^{18}O in precipitation: Modeling of the isotopic effects during snow formation, *Journal of Geophysical Research: Atmospheres*, 89(D7), 11,749–11,757, doi:10.1029/JD089iD07p11749.
- Jouzel, J., R. B. Alley, K. M. Cuffey, W. Dansgaard, P. Grootes, G. Hoffmann, S. J. Johnsen, R. D. Koster, D. Peel, C. A. Shuman, M. Stievenard, M. Stuiver, and J. White (1997), Validity of the temperature reconstruction from water isotopes in ice cores, *Journal of Geophysical Research: Oceans*, 102(C12), 26,471–26,487, doi:10.1029/97JC01283.
- Jouzel, J., M. Stievenard, S. J. Johnsen, A. Landais, V. Masson-Delmotte, A. Sveinbjornsdottir, F. Vimeux, U. von Grafenstein, and J. W. C. White (2007), The GRIP deuterium-excess record, *Quaternary Science Reviews*, 26(1), 1–17, doi: 10.1016/j.quascirev.2006.07.015.

- Judge, A., B. Pelletier, and I. Norquay (1987), Permafrost base and distribution of gas hydrates, in *Marine science atlas of the Beaufort Sea: geology and geophysics*, Geological Survey of Canada; Map 39.
- Kaufman, D. S., T. A. Ager, N. J. Anderson, P. M. Anderson, J. T. Andrews, P. J. Bartlein, L. B. Brubaker, L. L. Coats, L. C. Cwynar, M. L. Duvall, A. S. Dyke, M. E. Edwards, W. R. Eisner, K. Gajewski, A. Geirsdóttir, F. S. Hu, A. E. Jennings, M. R. Kaplan, M. W. Kerwin, A. V. Lozhkin, G. M. MacDonald, G. H. Miller, C. J. Mock, W. W. Oswald, B. L. Otto-Bliesner, D. F. Porinchu, K. Rühland, J. P. Smol, E. J. Steig, and B. B. Wolfe (2004), Holocene thermal maximum in the western Arctic (0–180°W), *Quaternary Science Reviews*, 23(5), 529–560, doi:10.1016/j.quascirev.2003.09.007.
- Kendall, C., and E. Caldwell (1998), Fundamentals of isotope geochemistry, pp. 51-86, in *Isotope tracers in catchment hydrology*, Elsevier Science & Technology, Amsterdam.
- Kokelj, S. V., and C. R. Burn (2003), Ground ice and soluble cations in near-surface permafrost, Inuvik, Northwest Territories, Canada, *Permafrost and Periglacial Processes*, 14(3), 275–289, doi:10.1002/ppp.458.
- Kokelj, S. V., M. J. Palmer, T. C. Lantz, and C. R. Burn (2017), Ground temperatures and permafrost warming from forest to tundra, Tuktoyaktuk Coastlands and Anderson Plain, NWT, Canada, *Permafrost and Periglacial Processes*, 28(3), 543–551, doi:10.1002/ppp.1934.
- Kopec, B. G., X. Feng, F. A. Michel, and E. S. Posmentier (2016), Influence of sea ice on Arctic precipitation, *Proceedings of the National Academy of Sciences of the United States of America*, 113(1), 46–51, publisher: National Academy of Sciences.
- Koster, R. D., J. Jouzel, R. J. Suozzo, and G. L. Russell (1992), Origin of July Antarctic precipitation and its influence on deuterium content: a GCM analysis, *Climate Dynamics*, 7(4), 195–203, doi:10.1007/BF00206861.

- Koster, R. D., D. P. de Valpine, and J. Jouzel (1993), Continental water recycling and H_2^{18}O concentrations, *Geophysical Research Letters*, 20(20), 2215–2218, doi:10.1029/93GL01781.
- Kostrova, S. S., H. Meyer, F. Fernandoy, M. Werner, and P. E. Tarasov (2020), Moisture origin and stable isotope characteristics of precipitation in southeast Siberia, *Hydrological Processes*, 34(1), 51–67, doi:10.1002/hyp.13571.
- Lacelle, D., and Y. K. Vasil'chuk (2013), Recent Progress (2007–2012) in Permafrost Isotope Geochemistry, *Permafrost and Periglacial Processes*, 24(2), 138–145, doi:10.1002/ppp.1768, publisher: Wiley.
- Lacelle, D., M. Fontaine, A. P. Forest, and S. Kokelj (2014), High-resolution stable water isotopes as tracers of thaw unconformities in permafrost: A case study from western Arctic Canada, *Chemical Geology*, 368, 85–96, doi:10.1016/j.chemgeo.2014.01.005.
- Lachenbruch, A. (1962), *Mechanics of thermal contraction cracks and ice-wedge polygons*, 70 ed., Geological Society of America, Denver, CO.
- Laudon, H., H. F. Hemond, R. Krouse, and K. H. Bishop (2002), ^{18}O fractionation during snowmelt: Implications for spring flood hydrograph separation, *Water Resources Research*, 38(11), 40–1–40–10, doi:10.1029/2002WR001510.
- Lauriol, B., C. Duchesne, and I. D. Clark (1995), Systématique du remplissage en eau des fentes de gel: Les résultats d'une étude ^{18}O et Deutérium, *Permafrost and Periglacial Processes*, 6(1), 47–55, doi:10.1002/ppp.3430060106.
- Lawrence, J., and J. White (1991), The elusive climate signal in the isotopic composition of precipitation, *Stable isotope geochemistry: a tribute to Samuel Epstein*. Special Publication, The Geochemical Society, 3.

- Lecavalier, B. S., D. A. Fisher, G. A. Milne, B. M. Vinther, L. Tarasov, P. Huybrechts, D. Lacelle, B. Main, J. Zheng, J. Bourgeois, and A. S. Dyke (2017), High Arctic Holocene temperature record from the Agassiz ice cap and Greenland ice sheet evolution, *Proceedings of the National Academy of Sciences of the United States of America*, 114(23), 5952–5957.
- Leffingwell, E. d. K. (1915), Ground-ice wedges: The dominant form of ground-ice on the north coast of Alaska, *The Journal of Geology*, 23(7), 635–654.
- Lemmen, D. S., A. Duk-Rodkin, and J. M. Bednarski (1994), Late glacial drainage systems along the northwestern margin of the Laurentide Ice Sheet, *Quaternary Science Reviews*, 13(9-10), 805–828, doi:10.1016/0277-3791(94)90003-5.
- Liljedahl, A. K., J. Boike, R. P. Daanen, A. N. Fedorov, G. V. Frost, G. Grosse, L. D. Hinzman, Y. Iijima, J. C. Jorgenson, N. Matveyeva, M. Necsoiu, M. K. Raynolds, V. E. Romanovsky, J. Schulla, K. D. Tape, D. A. Walker, C. J. Wilson, H. Yabuki, and D. Zona (2016), Pan-Arctic ice-wedge degradation in warming permafrost and its influence on tundra hydrology, *Nature Geoscience*, 9(4), 312–318, doi: 10.1038/ngeo2674.
- Linderholm, H. W., M. Nicolle, P. Francus, K. Gajewski, S. Helama, A. Korhola, O. Solomina, Z. Yu, P. Zhang, W. J. D'Andrea, M. Debret, D. V. Divine, B. E. Gunnarson, N. J. Loader, N. Massei, K. Seftigen, E. K. Thomas, J. Werner, S. Andersson, A. Berntsson, T. P. Luoto, L. Nevalainen, S. Saarni, and M. Väliranta (2018), Arctic hydroclimate variability during the last 2000 years: current understanding and research challenges, *Climate of the Past*, 14(4), 473–514, doi:10.5194/cp-14-473-2018.
- Liu, Z., J. Zhu, Y. Rosenthal, X. Zhang, B. L. Otto-Bliesner, A. Timmermann, R. S. Smith, G. Lohmann, W. Zheng, and O. E. Timm (2014), The Holocene temperature conundrum, *Proceedings of the National Academy of Sciences*, 111(34), E3501–E3505, doi:10.1073/pnas.1407229111.

- MacDonald, G. M. (2000), Postglacial vegetation and climate, in Evidence from north-west Canada for an early Holocene Milankovitch thermal maximum; pp. 57-63,, L.D. Dyke and G.R. Brooks, Geological Survey of Canada, Bulletin 547.
- Machavaram, M. V., and R. V. Krishnamurthy (1995), Earth surface evaporative process: A case study from the Great Lakes region of the United States based on deuterium excess in precipitation, *Geochimica et Cosmochimica Acta*, 59(20), 4279–4283, doi:10.1016/0016-7037(95)00256-Y.
- Machida, I., M. Ono, T. Kamitani, and Y. Muranaka (2022), Applicability of d-excess and ^{17}O -excess as groundwater tracers for determination of recharge area, *Hydrogeology Journal*, 30(7), 2027–2041, doi:10.1007/s10040-022-02526-0.
- Mackay, J. R. (1963), The Mackenzie Delta area, N.W.T Department of Mines and Technical Surveys, Geographical Branch (Memior 8).
- Mackay, J. R. (1970), Disturbances to the tundra and forest tundra environment of the western Arctic, *Canadian Geotechnical Journal*, 7(4), 420–432, doi:10.1139/t70-054.
- Mackay, J. R. (1972a), Offshore permafrost and ground ice, Southern Beaufort Sea, Canada, *Canadian Journal of Earth Sciences*, 9(11), 1550–1561, doi:10.1139/e72-137.
- Mackay, J. R. (1972b), The world of underground ice, *Annals of the Association of American Geographers*, 62(1), 1–22.
- Mackay, J. R. (1974a), Ice-wedge cracks, Garry Island, Northwest Territories, *Canadian Journal of Earth Sciences*, 11(10), 1366–1383, doi:10.1139/e74-133.
- Mackay, J. R. (1974b), Seismic shot holes and ground temperatures, Mackenzie Delta area, Northwest Territories. Report of Activities 74-1A, Geological Survey of Canada,

- Mackay, J. R. (1975), Relict ice wedges, Pelly Island, Northwest Territories. Report of Activities 75-1A, Geological Survey of Canada,
- Mackay, J. R. (1981), Active layer slope movement in a continuous permafrost environment, Garry Island, Northwest Territories, Canada, *Canadian Journal of Earth Sciences*, 18(11), 1666–1680, doi:10.1139/e81-154.
- Mackay, J. R. (1983a), Downward water movement into frozen ground, western arctic coast, Canada, *Canadian Journal of Earth Sciences*, 20(1), 120–134, doi:10.1139/e83-012.
- Mackay, J. R. (1983b), Oxygen isotope variations in permafrost, Tuktoyaktuk Peninsula Area, Northwest Territories. Current Research 83-1B, Geological Survey of Canada,
- Mackay, J. R. (1984a), The frost heave of stones in the active layer above permafrost with downward and upward freezing, *Arctic and Alpine Research*, 16(4), 439–446, doi:10.2307/1550906.
- Mackay, J. R. (1984b), The direction of ice-wedge cracking in permafrost: downward or upward?, *Canadian Journal of Earth Sciences*, 21(5), 516–524, doi:10.1139/e84-056.
- Mackay, J. R. (1988), Catastrophic lake drainage, Tuktoyaktuk Peninsula area, District of Mackenzie. Current Research 88-1D, Geological Survey of Canada,
- Mackay, J. R. (1989), Massive ice: some field criteria for the identification of ice types, *Current Research, Geological Survey of Canada*, 89(1G), 5–11, doi:10.4095/127554.
- Mackay, J. R. (1990), Some observations on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges, *Permafrost and Periglacial Processes*, 1(1), 15–29, doi:10.1002/ppp.3430010104.

- Mackay, J. R. (1992a), The frequency of ice-wedge cracking (1967–1987) at Garry Island, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 29(2), 236–248, doi:10.1139/e92-022.
- Mackay, J. R. (1992b), Lake stability in an ice-rich permafrost environment: Examples from the western Arctic coast, N.H.R.I Symposium, Series 7, pp. 1-26, in *Aquatic ecosystems in semi-arid regions: Implications for resource management*, Environment Canada, Saskatoon, Sask., Canada.
- Mackay, J. R. (1993), Air temperature, snow cover, creep of frozen ground, and the time of ice-wedge cracking, western Arctic coast, *Canadian Journal of Earth Sciences*, 30(8), 1720–1729, doi:10.1139/e93-151.
- Mackay, J. R. (1995), Ice wedges on hillslopes and landform evolution in the late Quaternary, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 32(8), 1093–1105, doi:10.1139/e95-091.
- Mackay, J. R. (1997), A full-scale field experiment (1978–1995) on the growth of permafrost by means of lake drainage, western Arctic coast: a discussion of the method and some results, *Canadian Journal of Earth Sciences*, 34(1), 17–33, doi:10.1139/e17-002.
- Mackay, J. R. (2000), Thermally induced movements in ice-wedge polygons, western arctic coast: a long-term study, *Géographie physique et Quaternaire*, 54(1), 41–68, doi:10.7202/004846ar.
- Mackay, J. R., and C. R. Burn (2002a), The first 20 years (1978-1979 to 1998-1999) of active-layer development, Illisarvik experimental drained lake site, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 39(11), 1657–1674.

- Mackay, J. R., and C. R. Burn (2002b), The first 20 years (1978-1979 to 1998–1999) of ice-wedge growth at the Illisarvik experimental drained lake site, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 39(1), 95–111, doi:10.1139/e01-048.
- Mackay, J. R., and S. R. Dallimore (1992), Massive ice of the Tuktoyaktuk area, western Arctic coast, Canada, *Canadian Journal of Earth Sciences*, 29(6), 1235–1249, doi:10.1139/e92-099.
- Mackay, J. R., and J. V. Matthews (1983), Pleistocene ice and sand wedges, Hooper Island, Northwest Territories, *Canadian Journal of Earth Sciences*, 20(7), 1087–1097, doi:10.1139/e83-097.
- Majoube, M. (1970), Fractionation Factor of ^{18}O between Water Vapour and Ice, *Nature*, 226(5252), 1242–1242, doi:10.1038/2261242a0.
- Mark, D. M., and M. Church (1977), On the misuse of regression in earth science, *Journal of the International Association for Mathematical Geology*, 9(1), 63–75, doi:10.1007/BF02312496.
- Masson-Delmotte, V., J. Jouzel, A. Landais, M. Stievenard, S. J. Johnsen, J. W. C. White, M. Werner, A. Sveinbjornsdottir, and K. Fuhrer (2005), GRIP Deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin, *Science*, 309(5731), 118–121, doi:10.1126/science.1108575.
- Merlivat, L., and J. Jouzel (1979), Global climatic interpretation of the D- ^{18}O relationship for precipitation, *Journal of Geophysical Research: Oceans*, 84(C8), 5029–5033, doi:10.1029/JC084iC08p05029.
- Merlivat, L., and G. Nief (1967), Fractionnement isotopique lors des changements d'état solide-vapeur et liquide-vapeur de l'eau à des températures inférieures à 0°C , 19(1), 122, doi:10.3402/tellusa.v19i1.9756.

- Meyer, H., L. Schirrmeister, K. Yoshikawa, T. Opel, S. Wetterich, H.-W. Hubberten, and J. Brown (2010), Permafrost evidence for severe winter cooling during the Younger Dryas in northern Alaska, *Geophysical Research Letters*, 37(3), doi:10.1029/2009GL041013.
- Meyer, H., T. Opel, T. Laepple, A. Y. Dereviagin, K. Hoffmann, and M. Werner (2015), Long-term winter warming trend in the Siberian Arctic during the mid- to late Holocene, *Nature Geoscience*, 8(2), 122–125, doi:10.1038/ngeo2349.
- Michel, F. A. (1982), Isotope investigations of permafrost waters in northern Canada, Ph.D. thesis, University of Waterloo, Waterloo, Ontario.
- Michel, F. A. (1990), Isotopic composition of ice-wedge ice in northwestern Canada, vol. 54, pp. 5-9, in *Fifth Canadian Permafrost Conference*, Quebec City, Canada.
- Michel, F. A. (2011), Isotope characterisation of ground ice in northern Canada: Isotope characterisation of ground ice, *Permafrost and Periglacial Processes*, 22(1), 3–12, doi:10.1002/ppp.721.
- Michel, F. A., P. Fritz, and R. J. Drimmie (1989), Evidence of climatic change from oxygen and carbon isotope variations in sediments of a small arctic lake, Canada, *Journal of Quaternary Science*, 4(3), 201–209, doi:10.1002/jqs.3390040302.
- Morse, P. D., and C. R. Burn (2013), Field observations of syngenetic ice wedge polygons, outer Mackenzie Delta, western Arctic coast, Canada, *Journal of Geophysical Research: Earth Surface*, 118(3), 1320–1332, doi:10.1002/jgrf.20086.
- Muller, S. (1947), *Permafrost or Permanently Frozen Ground and Related Engineering Problems*, Edwards, Ann Arbor, MI.
- Murton, J. B. (2009), Stratigraphy and palaeoenvironments of Richards Island and the eastern Beaufort Continental Shelf during the last glacial-interglacial

- cycle: Beaufort Shelf Palaeoenvironments of the Last Glacial-interglacial Cycle, *Permafrost and Periglacial Processes*, 20(2), 107–125, doi:10.1002/ppp.647.
- Murton, J. B., M. Frechen, and D. Maddy (2007), Luminescence dating of mid- to Late Wisconsinan aeolian sand as a constraint on the last advance of the Laurentide Ice Sheet across the Tuktoyaktuk Coastlands, western Arctic Canada, *Canadian Journal of Earth Sciences*, 44(6), 857–869, doi:10.1139/e07-015.
- Murton, J. B., M. D. Bateman, A. M. Telka, R. Waller, C. Whiteman, and S. Kuzmina (2017), Early to mid Wisconsin fluvial deposits and Palaeoenvironment of the Kidluit Formation, Tuktoyaktuk Coastlands, Western Arctic Canada, *Permafrost and Periglacial Processes*, 28(3), 523–533, doi:10.1002/ppp.1946.
- Noone, D., and I. Simmonds (2004), Sea ice control of water isotope transport to Antarctica and implications for ice core interpretation, *Journal of Geophysical Research: Atmospheres*, 109(D7), doi:10.1029/2003JD004228.
- O'Neil, J. R. (1968), Hydrogen and oxygen isotope fractionation between ice and water, *The Journal of Physical Chemistry*, 72(10), 3683–3684, doi:10.1021/j100856a060.
- Opel, T., T. Laepple, H. Meyer, A. Y. Dereviagin, and S. Wetterich (2017), Northeast Siberian ice wedges confirm Arctic winter warming over the past two millennia, *The Holocene*, 27(11), 1789–1796, doi:10.1177/0959683617702229.
- Pelletier, B. R. (1987), Late Quaternary glaciation and sea level history, in *Marine science atlas of the Beaufort Sea: geology and geophysics*, Miscellaneous report / Geological Survey of Canada, Energy, Mines and Resources Canada: Canadian Government Publishing Centre, Ottawa, Canada.
- Pfahl, S., and H. Sodemann (2014), What controls deuterium excess in global precipitation?, *Climate of the Past*, 10(2), 771–771.

- Poage, M. A. (2001), Empirical relationships between elevation and the stable isotope composition of precipitation and surface waters: considerations for studies of paleoelevation change, *American Journal of Science*, 301(1), 1–15, doi:10.2475/ajs.301.1.1.
- Pollard, W. H., and H. M. French (1980), A first approximation of the volume of ground ice, Richards Island, Pleistocene Mackenzie delta, Northwest Territories, Canada, *Canadian Geotechnical Journal*, 17(4), 509–516, doi:10.1139/t80-059.
- Polyak, L., R. B. Alley, J. T. Andrews, J. Brigham-Grette, T. M. Cronin, D. A. Darby, A. S. Dyke, J. J. Fitzpatrick, S. Funder, M. Holland, A. E. Jennings, G. H. Miller, M. O'Regan, J. Savelle, M. Serreze, K. St. John, J. W. White, and E. Wolff (2010), History of sea ice in the Arctic, *Quaternary Science Reviews*, 29(15-16), 1757–1778, doi:10.1016/j.quascirev.2010.02.010.
- Porter, T. J., and T. Opel (2020), Recent advances in paleoclimatological studies of Arctic wedge- and pore-ice stable-water isotope records, *Permafrost and Periglacial Processes*, 31(3), 429–441, doi:10.1002/ppp.2052.
- Porter, T. J., D. G. Froese, S. J. Feakins, I. N. Bindeman, M. E. Mahony, B. G. Pautler, G.-J. Reichert, P. T. Sanborn, M. J. Simpson, and J. W. H. Weijers (2016), Multiple water isotope proxy reconstruction of extremely low last glacial temperatures in Eastern Beringia (Western Arctic), *Quaternary Science Reviews*, 137, 113–125, doi:10.1016/j.quascirev.2016.02.006.
- Putman, A. L., R. P. Fiorella, G. J. Bowen, and Z. Cai (2019), A global perspective on local meteoric water lines: Meta-analytic insight into fundamental controls and practical constraints, *Water Resources Research*, 55(8), 6896–6910, doi:10.1029/2019WR025181.
- Rampton, V. N. (1988), Quaternary geology of the Tuktoyaktuk Coastlands, Northwest Territories, *Canadian Journal of Earth Sciences*, 25, 423–437, doi:10.4095/126937.

- Rayleigh, J. W. (1896), Theoretical considerations respecting the separation of gases by diffusion and similar processes, *The London, Edinburgh, and Dublin Philosophical Magazine and Journal of Science*, 42(259), 493–498, doi:10.1080/14786449608620944.
- Ritchie, J., and F. Hare (1971), Late-quaternary vegetation and climate near the arctic tree line of northwestern North America, *Quaternary Research*, 1(3), 331–342, doi:10.1016/0033-5894(71)90069-X.
- Ritchie, J. C. (1984), *Past and present vegetation of the far northwest of Canada*, University of Toronto Press.
- Ritchie, J. C., L. C. Cwynar, and R. W. Spear (1983), Evidence from north-west Canada for an early Holocene Milankovitch thermal maximum, *Nature*, 305(5930), 126–128, doi:10.1038/305126a0.
- Rogers, R. R., and M. K. Yau (1996), *A short course in cloud physics*, International series in natural philosophy, 3 ed., Butterworth Heinemann, Woburn, Mass.
- Ross, S. L. (2017), *Weather and climate: an introduction*, 2 ed., Oxford University Press, Don Mills, Ontario, Canada.
- Rozanski, K., L. Araguás-Araguás, and R. Gonfiantini (1993), Isotopic patterns in modern global precipitation, pp. 1-36, in *Climate Change in Continental Isotopic Records*, American Geophysical Union (AGU).
- Rumble, J. E. (2023), *CRC Handbook of Chemistry and Physics*, 104 ed., CRC Press.
- Salati, E., A. Dall'Olio, E. Matsui, and J. R. Gat (1979), Recycling of water in the Amazon Basin: An isotopic study, *Water Resources Research*, 15(5), 1250–1258, doi:10.1029/WR015i005p01250.

- Schetselaar, A., T. Andersen, and C. R. Burn (2023), Performance of climate projections for Yukon and adjacent Northwest Territories, 1991 – 2020, *ARCTIC*, doi:10.14430/arctic77263.
- Serreze, M. C., and R. G. Barry (2011), Processes and impacts of Arctic amplification: A research synthesis, *Global and Planetary Change*, 77(1), 85–96, doi:10.1016/j.gloplacha.2011.03.004.
- Shur, Y., K. M. Hinkel, and F. E. Nelson (2005), The transient layer: implications for geocryology and climate-change science, *Permafrost and Periglacial Processes*, 16(1), 5–17, doi:10.1002/ppp.518.
- Spangenberg, J. E. (2012), Caution on the storage of waters and aqueous solutions in plastic containers for hydrogen and oxygen stable isotope analysis, *Rapid Communications in Mass Spectrometry*, 26(22), 2627–2636, doi:10.1002/rcm.6386.
- St-Jean, M., B. Lauriol, I. D. Clark, D. Lacelle, and C. Zdanowicz (2011), Investigation of ice-wedge infilling processes using stable oxygen and hydrogen isotopes, crystallography and occluded gases., *Permafrost and Periglacial Processes*, 22(1), 49–64, doi:10.1002/ppp.680.
- Stein, A. F., R. R. Draxler, G. D. Rolph, B. J. B. Stunder, M. D. Cohen, and F. Ngan (2015), NOAA's HYSPLIT atmospheric transport and dispersion modeling system, *Bulletin of the American Meteorological Society*, 96(12), 2059–2077, doi:10.1175/BAMS-D-14-00110.1.
- Stewart, M. K. (1975), Stable isotope fractionation due to evaporation and isotopic exchange of falling waterdrops: Applications to atmospheric processes and evaporation of lakes, *Journal of Geophysical Research*, 80(9), 1133–1146, doi:10.1029/JC080i009p01133.

- Streletskiy, D. A., N. I. Shiklomanov, J. D. Little, F. E. Nelson, J. Brown, K. E. Nyland, and A. E. Klene (2017), Thaw subsidence in undisturbed tundra landscapes, Barrow, Alaska, 1962–2015, *Permafrost and Periglacial Processes*, 28(3), 566–572, doi:10.1002/ppp.1918.
- Takano, S., A. Sugimoto, S. Tei, M. Liang, R. Shingubara, T. Morozumi, and T. C. Maximov (2019), Isotopic compositions of ground ice in near-surface permafrost in relation to vegetation and microtopography at the Taiga-Tundra boundary in the Indigirka River lowlands, northeastern Siberia, *PLoS ONE*, 14(10), e0223720–e0223720, doi:10.1371/journal.pone.0223720.
- Taylor, S., X. Feng, J. W. Kirchner, R. Osterhuber, B. Klaue, and C. E. Renshaw (2001), Isotopic evolution of a seasonal snowpack and its melt, *Water Resources Research*, 37(3), 759–769, doi:10.1029/2000WR900341.
- Taylor, S., X. Feng, M. Williams, and J. McNamara (2002), How isotopic fractionation of snowmelt affects hydrograph separation, *Hydrological Processes*, 16(18), 3683–3690, doi:10.1002/hyp.1232.
- Thienpont, J. R., K. M. Rühland, M. F. J. Pisaric, S. V. Kokelj, L. E. Kimpe, J. M. Blais, and J. P. Smol (2013), Biological responses to permafrost thaw slumping in Canadian Arctic lakes: *Aquatic biota response to permafrost thaw*, *Freshwater Biology*, 58(2), 337–353, doi:10.1111/fwb.12061.
- Urey, H. C. (1947), The thermodynamic properties of isotopic substances, *Journal of the Chemical Society (Resumed)*, (0), 562–581, doi:10.1039/JR9470000562.
- Vasil'chuk, Y. K., N. A. Budantseva, A. C. Vasil'chuk, and J. N. Chizhova (2020), Winter air temperature during the Holocene optimum in the north-eastern part of the east European plain based on ice wedge stable isotope records, *Permafrost and Periglacial Processes*, 31(2), 281–295, doi:10.1002/ppp.2043.

- Vasil'chuk, Y., J. Vasil'chuk, N. Budantseva, and A. Vasil'chuk (2022), MIS 3-2 paleo-winter reconstructions obtained from stable water isotope records of radiocarbon-dated ice wedges of the Batagay Ice Complex (Yana Upland, Eastern Siberia), *Radiocarbon*, 64(6), 1403–1417, doi:10.1017/RDC.2022.60.
- Vasil'chuk, Y. K., J. van der Plicht, H. Jungner, E. Sonninen, and A. C. Vasil'chuk (2000), First direct dating of Late Pleistocene ice-wedges by AMS, *Earth and Planetary Science Letters*, 179(2), 237–242, doi:10.1016/S0012-821X(00)00122-9.
- Vasil'chuk, Y. K., V. N. Zaitsev, and A. C. Vasil'chuk (2006), A ^{14}C -dating and oxygen-isotope diagram of a Holocene-reformed ice wedge on the Chara River (Transbaikal region), *Doklady Earth Sciences*, 407(1), 265–270, doi:10.1134/S1028334X06020267.
- Wanner, H., J. Beer, J. Bütikofer, T. J. Crowley, U. Cubasch, J. Flückiger, H. Goosse, M. Grosjean, F. Joos, J. O. Kaplan, M. Küttel, S. A. Müller, I. C. Prentice, O. Solomina, T. F. Stocker, P. Tarasov, M. Wagner, and M. Widmann (2008), Mid- to Late Holocene climate change: an overview, *Quaternary Science Reviews*, 27(19), 1791–1828, doi:10.1016/j.quascirev.2008.06.013.
- Wetterich, S., H. Meyer, M. Fritz, G. Mollenhauer, J. Rethemeyer, A. Kizyakov, L. Schirrmeister, and T. Opel (2021), Northeast Siberian Permafrost Ice-Wedge Stable Isotopes Depict Pronounced Last Glacial Maximum Winter Cooling, *Geophysical Research Letters*, 48(7), e2020GL092087, doi:10.1029/2020GL092087, _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020GL092087>.
- Wetterich, S., A. I. Kizyakov, T. Opel, H. Grotheer, G. Mollenhauer, and M. Fritz (2023), Ground-ice origin and age on Herschel Island (Qikiqtaruk), Yukon, Canada, *Quaternary Science Advances*, 10, 100,077, doi:10.1016/j.qsa.2023.100077.

Appendix & Statement of Co-Authorship

Table .1: Mean monthly temperature (T, °C), total monthly precipitation (P, mm), and $\delta^{18}\text{O}$ (‰) for aggregate samples of monthly precipitation collected at Inuvik Airport between March 1989 and May 1995

Date	T	P	$\delta^{18}\text{O}$	Date	T	P	$\delta^{18}\text{O}$
3/1989	-23.5	4.4	-25.0	4/1992	-15.3	28.1	-27.5
4/1989	-9.4	1.6	-28.7	5/1992	-1.6	16.4	-25.2
5/1989	0.5	18.7	-23.7	6/1992	11.8	17.1	-19.2
6/1989	12.3	20.2	-14.8	7/1992	14.2	26.6	-17.8
7/1989	16.5	63.2	-16.8	8/1992	11.7	29	-18.3
8/1989	15.5	32.8	-16.9	9/1992	-1.4	48.2	-24.9
9/1989	5.3	53.9	-19.0	10/1992	-10.6	26.8	-21.5
10/1989	-8.2	36.8	-28.3	11/1992	-19	14.8	-28.2
11/1989	-27.9	11	-34.4	12/1992	-24.7	19.1	-28.7
12/1989	-26.1	11.4	-27.6	1/1993	-22.8	27.5	-29.5
1/1990	-32.7	18.4	-30.5	2/1993	-21.3	23.2	-27.5
2/1990	-34.3	9	-31.2	3/1993	-18.8	6.6	-26.0
3/1990	-16.3	19.6	-30.8	4/1993	-1.5	5.6	-22.5
4/1990	-7.7	13.1	-24.6	5/1993	0.9	8.6	-28.1
6/1990	11.3	16	-21.9	6/1993	12.6	3.7	-11.4
7/1990	15.5	58	-18.8	8/1993	11	41.6	-20.8
8/1990	9.7	10.2	-15.1	9/1993	4.2	30.6	-18.2
9/1990	3.2	31.7	-21.9	10/1993	-4.3	27.3	-24.4
10/1990	-7.4	25.8	-29.3	11/1993	-17.8	13.9	-26.8
11/1990	-26.9	20.8	-26.7	12/1993	-23.1	19.6	-27.4
12/1990	-28.7	12.4	-28.4	2/1994	-27.3	21.2	-28.9
1/1991	-25	20	-28.3	3/1994	-21.9	26.1	-30.6
2/1991	-28.5	13.4	-33.4	4/1994	-12.3	14.8	-16.9
3/1991	-23.6	19.2	-29.2	5/1994	3	20	-20.8
4/1991	-8.2	6	-23.1	6/1994	10.3	31.9	-18.2
5/1991	5.9	20.6	-18.3	7/1994	18.3	26	-14.4
6/1991	10.5	33.4	-18.2	8/1994	14.9	44.5	-15.5
7/1991	12.4	16.7	-16.6	9/1994	1.2	31.7	-21.4
8/1991	9.1	73.2	-20.2	10/1994	-5.3	21.4	-23.9
9/1991	4.2	41.5	-19.3	11/1994	-22.8	34.7	-32.5
10/1991	-7.5	29.6	-23.4	12/1994	-25.2	19.7	-34.1
11/1991	-24.1	13.1	-32.5	1/1995	-23.6	8	-29.4
12/1991	-29.2	9.3	-25.6	2/1995	-26	23.2	-29.1
1/1992	-28.7	24	-29.8	3/1995	-25	23.2	-28.4
2/1992	-28.2	14.8	-23.3	4/1995	-7.5	3.3	-17.1
3/1992	-16.2	11.4	-24.5	5/1995	5.3	25.6	-22.9