

Université de Montréal

La glace de glacier enfouie dans le pergélisol de l'île Bylot:  
Origine, caractéristiques et impacts géomorphologiques

*Par*

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*Cette thèse intitulée*

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Origine, caractéristiques et impacts géomorphologiques**

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## Résumé

Au cours des dernières décennies, les observations de glace de glacier enfouie exposée dans les falaises côtières et les glissements de terrain causés par le dégel du pergélisol arctique démontrent que des quantités importantes de glace de glacier ont survécu à la déglaciation et sont toujours préservées dans le pergélisol. Le premier volet de cette étude visait à caractériser des expositions de glace massive observées à l'île Bylot (Nunavut), afin de connaître l'origine de la glace. Puisque la glace de glacier enfouie peut jouer un rôle important dans l'évolution des paysages périglaciaires, cette thèse s'intéresse également au rôle joué par la glace de glacier enfouie dans l'initiation et l'évolution de lacs de thermokarst.

Nos résultats démontrent que le pergélisol de l'île Bylot contient des restes de glace de glacier du Pléistocène qui ont survécu aux dernières déglaciations. Dans la vallée Qarlikturvik, des masses de glace intraglacière (dérivée du névé) sont associées à un courant de glace de l'inlandsis laurentidien qui recouvrait une partie de la plaine sud de l'île vers la fin du Pléistocène. Cette masse de glace formait une zone de convergence avec les glaciers alpins locaux s'écoulant de la calotte glaciaire centrée sur les monts Byam Martin. Sur l'un des plateaux bordant la vallée Qarlikturvik, une masse de glace de glacier enfouie est associée à la partie basale d'un glacier dont l'âge minimal est estimé à environ 0.77 Ma, mais pourrait être aussi vieux que 2.6 Ma. En raison de sa localisation, à environ 500 m d'altitude, la glace proviendrait vraisemblablement d'une avancée glaciaire régionale et pourrait être associée à la Glaciation de Baffin, soit la plus vieille avancée glaciaire régionale reconnue sur l'île Bylot.

De plus, cette glace représente le plus vieux reste de glacier connu en Amérique du Nord et l'une des premières indications de glaciations dans l'est de l'Arctique canadien. Puis, la persistance et la fonte tardive de ces épaisses couches de glace de glacier datant du Pléistocène ont eu des effets importants sur le paysage de l'île Bylot, notamment sur les lacs. En effet, nos résultats démontrent que l'initiation des lacs profonds (> 5 m) est liée à la fonte de la glace de glacier enfouie. Ces lacs de thermokarst glaciaire continueront d'évoluer dans un contexte

périglacière par la fonte de la glace intrasédimentaire (p. ex., glace de ségrégation) et des coins de glace formés ultérieurement dans les sédiments encaissants lors de l'aggradation du pergélisol suivant le retrait glaciaire. Alors qu'une grande partie des paysages arctiques est encore fortement déterminée par leur héritage glaciaire, la fonte de ces masses de glace aura un impact important sur la dynamique des géosystèmes et écosystèmes arctiques.

**Mots-clés :** Pergélisol, Glace de glacier, Arctique canadien, Glaciations, Cryostratigraphie, Géochimie, Paléomagnétisme, Pléistocène, Thermokarst.



## Abstract

Over the past decades, observations of buried glacier ice exposed in coastal bluffs and headwalls of retrogressive thaw slumps of the Arctic have indicated that considerable amounts of late Pleistocene glacier ice survived the deglaciation and are still preserved in permafrost. The first phase of this project aimed to characterize two exposures of massive ice observed on Bylot Island (Nunavut) to infer their origins. Since buried glacier ice can play a significant role in reshaping periglacial landscapes, this study also investigates the initiation and development of thermokarst lakes in a tundra valley in response to the melting of buried glacier ice.

Our results show that the permafrost of Bylot Island contains remnants of Pleistocene glacier ice that survived the past deglaciations. In the Qarlikturvik valley, bodies of englacial ice (*firn-derived*) originated from an ice stream flowing from the Laurentide Ice Sheet, which covered part of the southern plain of the island towards the end of the Pleistocene. These glacier ice bodies formed a convergence zone with local alpine glaciers flowing from the ice cap centred over the Byam Martin Mountains. On the edge of a flat plateau bordering the Qarlikturvik Valley, a buried glacier ice body is associated with the basal part of a glacier whose minimum age is estimated at 0.77 Ma, but could be as old as 2.6 Ma. Due to its location on a 500-m a.s.l. plateau, the ice likely originates from a regional glacial advance and could be associated with the Baffin Glaciation, which is the oldest known glaciation on Bylot Island. In addition, this buried glacier ice represents the oldest glacier ice preserved in ice-free Arctic landscapes, and the earliest evidence of a Pleistocene glaciation in the eastern Canadian Arctic Archipelago.

The persistence and delayed melting of these thick beds of buried Pleistocene glacier ice had wide-ranging effects on the landscape of Bylot Island. Our results suggest that the initiation of deeper thermokarst lakes (> 5 m) was triggered by the melting of buried glacier ice in our study area, while shallow thermokarst lakes were triggered from the melting of intrasedimental ice and ice wedges. These glacial thermokarst lakes will continue to evolve in a periglacial context through the melting of intrasedimental ice (e.g. segregation ice) and ice wedges subsequently formed in

the surrounding sediments during permafrost aggradation following the glacial retreat. As most of the glaciated Arctic landscapes are still strongly determined by their glacial legacy, the melting of these large ice bodies will have significant impacts on Arctic ecosystems and geosystems.

**Keywords** : Permafrost, Glacier ice, Canadian Arctic, Glaciations, Cryostratigraphy, Geochemistry, Paleomagnetism, Pleistocene, Thermokarst

# Table des matières

<b>Résumé.....</b>	<b>5</b>
<b>Abstract.....</b>	<b>7</b>
<b>Table des matières.....</b>	<b>9</b>
<b>Liste des tableaux .....</b>	<b>15</b>
<b>Liste des figures .....</b>	<b>17</b>
<b>Liste des sigles et abréviations .....</b>	<b>27</b>
<b>Remerciements.....</b>	<b>33</b>
<b>Chapitre 1 : Introduction générale.....</b>	<b>1</b>
<b>1.1 Glace massive et classification.....</b>	<b>1</b>
<b>1.2 Propriétés de la glace de glacier .....</b>	<b>3</b>
1.2.1 Glace intraglacière .....	3
1.2.2 Glace basale .....	5
1.2.3 Glace intrasedimentaire .....	7
1.2.4 Glace de congère .....	7
1.2.5 Glace de rivière, lac et mer .....	8
<b>1.3 Enfouissement de la glace de glacier .....</b>	<b>8</b>
1.3.1 Processus thermiques .....	9
1.3.2 Dépôts glaciaires .....	12
1.3.3 Melt-out till .....	13
1.3.4 Glaciotectonisme .....	14
1.3.5 Dépôts fluvioglaciaires .....	14
1.3.6 Mouvements de masse.....	15
<b>1.4 Méthodes de caractérisation de la glace.....</b>	<b>17</b>
1.4.1 Cryostratigraphie et cristallographie .....	18
1.4.2 Isotopes stables O et H .....	20
1.4.3 Composition ionique.....	23
1.4.4 Composition gazeuse .....	23

<b>1.5 Méthodes de datation de la glace</b> .....	<b>25</b>
1.5.1 Méthodes de datation radiométrique basée sur la désintégration radioactive .....	25
1.5.2 Méthodes de datation par luminescence .....	28
1.5.3 Datation d'exposition par isotopes cosmogéniques .....	30
1.5.4 Paléomagnétisme .....	31
1.5.5 Acides aminés .....	32
<b>1.6 Contexte géomorphologique et répartition géographique</b> .....	<b>33</b>
<b>1.7 Des traces d'anciennes glaciations</b> .....	<b>41</b>
<b>1.8 Structure et objectifs de la thèse</b> .....	<b>47</b>
<b><i>Chapitre 2 Origin, burial and preservation of late Pleistocene-age glacier ice in Arctic permafrost (Bylot Island, NU, Canada)</i></b> .....	<b>53</b>
<b>2.1 Abstract</b> .....	<b>54</b>
<b>2.2 Introduction</b> .....	<b>55</b>
<b>2.3 Regional settings</b> .....	<b>57</b>
<b>2.4 Material and methods</b> .....	<b>59</b>
<b>2.5 Results</b> .....	<b>64</b>
2.5.1 Cryostratigraphy and properties of the massive ice body .....	64
2.5.2 Cryostratigraphy and properties of the overlying sediments .....	70
<b>2.6 Discussion</b> .....	<b>72</b>
2.6.1 Origin of the massive ice.....	72
2.6.2 Burial and preservation of late Pleistocene glacier ice in permafrost .....	76
<b>2.7 Conclusion</b> .....	<b>78</b>
<b>2.8 Acknowledgements</b> .....	<b>79</b>
<b>Appendix A</b> .....	<b>80</b>
<b>Appendix B</b> .....	<b>80</b>
<b>Appendix C</b> .....	<b>81</b>

<b><i>Chapitre 3 Early Pleistocene glaciation of the eastern Canadian Arctic Archipelago: new evidence from buried glacier ice remnants, Bylot Island, Nunavut. ....</i></b>	<b>83</b>
<b>3.1 Abstract.....</b>	<b>84</b>
<b>3.2 Introduction .....</b>	<b>84</b>
<b>3.3 Site descriptions and Methods .....</b>	<b>86</b>
<b>3.4 Results .....</b>	<b>89</b>
3.4.1 Cryostratigraphy, $\delta D$ - $\delta^{18}O$ and major cations of the buried glacier ice.....	89
3.4.2 Cryostratigraphy, major cations and $\delta D$ - $\delta^{18}O$ of the sediment cover .....	94
3.4.3 Age determination .....	95
<b>3.5 Discussion .....</b>	<b>98</b>
3.5.1 Origin and burial of the ice .....	98
3.5.2 Age determination .....	100
<b>3.6 Conclusion.....</b>	<b>104</b>
<b>3.7 Acknowledgement.....</b>	<b>104</b>
<b>3.8 Supporting information.....</b>	<b>105</b>
S1. Field sampling and analyses.....	105
Field sampling .....	105
Ct-scan .....	105
Volumetric ice content (%) .....	106
Stable O-H isotopes and cation contents .....	107
Grain-size distribution .....	107
S2. Age determination .....	108
2.1 Radiocarbon dating.....	108
2.2 Paleomagnetic analyses.....	108
2.2 Pollens analysis .....	109
2.3 Diatoms analysis .....	110
<b><i>Chapitre 4 Contrasted geomorphological and limnological properties of thermokarst lakes formed in buried glacier ice and ice-wedge polygon terrain.....</i></b>	<b>121</b>
<b>4.1 Abstract.....</b>	<b>122</b>
<b>4.2 Introduction .....</b>	<b>123</b>

<b>4.3 Study area .....</b>	<b>125</b>
<b>4.4 Material and methods .....</b>	<b>128</b>
4.4.1 Landforms, surficial deposits and lake mapping.....	128
4.4.2 Distribution of lakes in the valley and the southern plain of Bylot Island .....	129
4.4.3 Lake morphology in the Qarlikturvik Valley.....	130
4.4.4 Stratigraphic profiles of lake bottom sediments .....	131
4.4.5 Water column profiles of temperature and dissolved oxygen .....	132
<b>4.5 Results .....</b>	<b>133</b>
4.5.1 Distribution of lakes in the valley and southern plain of Bylot Island .....	133
4.5.2 Lake morphology in the Qarlikturvik Valley.....	135
4.5.3 Stratigraphic profiles of lake bottom sediments .....	138
4.5.4 Water column profiles of temperature and dissolved oxygen .....	141
<b>4.6 Discussion .....</b>	<b>143</b>
4.6.1 Lake morphology and sediment stratigraphy of shallow thermokarst lakes formed in ice wedges .....	144
4.6.2 Lake morphology and sediment stratigraphy of deep glacial thermokarst lakes formed in buried glacier ice.....	145
4.6.3 Conceptual model of thermokarst lake development in ice-wedges polygon terrain and buried glacier ice .....	147
4.6.4 Implications on Arctic lakes ecosystem dynamics .....	153
<b>4.7 Conclusion.....</b>	<b>155</b>
<b>4.8 Acknowledgement.....</b>	<b>156</b>
<b>4.9 Supporting information.....</b>	<b>157</b>
S1 Bathymetric mapping through lake ice with ground-penetrating radar (GPR).....	157
S2 Computed tomography (CT) scanning of a sediment core collected in lakes K and G, Bylot Island, Nunavut, Canada .....	158
<b><i>Chapitre 5 : Conclusion générale et perspectives .....</i></b>	<b><i>164</i></b>
<b>5.1 Synthèse des résultats et avancées scientifiques.....</b>	<b>165</b>
5.1.1 Glace de glacier datant de la fin du Pléistocène .....	165
5.1.2 Glace de glacier datant du Pléistocène inférieur .....	168
5.1.3 Rôle géomorphologique de la glace de glacier .....	170
<b>5.2 Limitations et recommandations pour de futur travaux .....</b>	<b>172</b>

5.2.1 Limitations associées à l'échantillonnage .....	172
5.2.2 Critères diagnostiques pour l'identification de la glace de glacier .....	172
5.2.3 Datation de glace .....	173
5.2.4 Élargir la portée spatiale de l'étude des lacs de thermokarst glaciaires.....	175
5.2.3 Modélisation de la répartition spatiale de la glace de glacier enfouie .....	176
<b>5.3 Perspectives .....</b>	<b>177</b>
<b><i>Références bibliographiques .....</i></b>	<b>181</b>
<b><i>Annexe I. Buried remnants of the Laurentide Ice Sheet and connections to its surface elevation</i></b> <b>.....</b>	<b>227</b>

## Liste des tableaux

<b>Tableau 1.</b>	Densité de la neige, du névé et de la glace.....	8
<b>Tableau 2.</b>	Valeurs de ratios $O_2/Ar$ et $N_2/Ar$ de l'atmosphère et différents types de glace....	24
<b>Tableau 3.</b>	Glaciations locales et régionales documentées à l'île Bylot .....	46
<b>Tableau 4.</b>	Comparisons of the physical properties of different massive tabular ground ice found in the permafrost.....	73
<b>Tableau 5.</b>	Characteristics of lakes for which bathymetric data were collected in 2015.....	137



## Liste des figures

- Figure 1.** Classification génétique de la glace massive .....3
- Figure 2.** Exemples de mesures empiriques de la relation entre l'épaisseur des sédiments et la fonte de la glace .....10
- Figure 3.** Moraine médiane du glacier Matanuska, Alaska .....13
- Figure 4.** Exemples d'avalanches rocheuses sur des glaciers au Yukon et en Alaska .....16
- Figure 5.** a) Configuration des inlandsis au dernier maximum glaciaire selon Dalton et al., 2022. b) Carte de répartition des dépôts de surface couramment associés à la glace de glacier enfouie. Les flèches oranges indiquent la direction des courants de glace; c) Cartographie à grande échelle qui illustre une estimation de l'abondance de glace relique basées sur la modélisation d'O'Neill et al. (2019). 35
- Figure 6.** a) Formation des kettles; b) Photographie aérienne montrant une douzaine de dépressions en expansion se formant à la surface de la glace stagnante recouverte de débris dans la moraine médiane du glacier de Béring, forêt nationale de Chugach, en Alaska. La flèche blanche montre une exposition de glace de glacier enfouie dans la paroi d'un lac de kettle. La présence de glace combinée à la nature groupée des lacs suggèrent que la glace enfouie est continue entre les lacs et probablement présente dans l'ensemble de la zone délimitée. ....37
- Figure 7.** Configuration des calottes glaciaires dans l'hémisphère nord et localisation des expositions de glace de glacier enfouie rapportées dans la littérature : 1) île Barter, Alaska; 2) Lac Chapman; 3) île Herschel; 4) île Pullen; 5) Lac Ya-Ya; 6) Lac Eskimo; 7) Moraine Sand Hills, île de Banks; 8) Moraine Jesse, île Victoria; 9) Loch Point, île Victoria; 10) Péninsule de Wollaston, île Victoria; 11) Lac Ammalurtuq; 12) Moraine du lac Bluenose; 13) Lac Contwoyto; 14) Sud de la baie de Pelly; 15) Île Bylot; 16) Iqaluit; 17) Lac Hazen, île d'Ellesmere; 19) Péninsule de Taymir; 20) Ladyanaya Gora (aussi Ice Hills); 21) Baie de Baydaratskaya; 22) Cape Shpindler; 23) Lac Komovoye .....39
- Figure 8.** Localisation des expositions de glace de glacier enfouie rapportées dans la littérature en Antarctique : 1) Garwood valley; 2) University valley; 3) Taylor valley; 4) Ong valley; 5) Kenmar

valley; 6) Beacon valley; 7) Northern Foothills – Simpson Crags; 8) Northern Foothills – Boulder Clay Glacier; 9) Seymour Island; 10) Vega island. ....40

**Figure 9.** Carte des limites glaciaires rapportées par Klassen (1993), Dyke et Hopper (2001). Les flèches roses représentent la direction des courants glaciaires.....45

**Figure 10.** (a) Location of Bylot Island, (b) location of the Qarlikturvik Valley, (c) location of the study area within the Qarlikturvik Valley (valley of glaciers C-79 and C-93). Wedge ice, segregated ice and snow were also sampled near (< 1 km) the massive ice exposure. The thaw depths were measured with a steel probe at every 10 m along a 150 m transect. The orange area represents mounds of ice contact sediments, indicating past frontal glacier positions. The dashed red line indicates the marine limit (ca. 35 m). ....58

**Figure 11.** A photograph and a schematic cross section showing generalized stratigraphy of the massive ice exposure and the overlying sediments. The lower and lateral contacts of the massive ice have not been reached. The thaw depth measured in late July (2013) at the headwall reaches 55 cm. The red star indicates the sampling location of the organic material and the red dots show the sampling points for stable O-H isotope and hydrochemistry. ....61

**Figure 12.** Photographs (left) and CT scans (right) of the massive ice body. The CT scan images are not exactly the same position as the photographs. Unconsolidated sediments and rock (high-density minerals) appear light grey or white, respectively. Gases inclusions and water appear black, and other materials, such as ice, can have various shades of dark grey (a) Pure-ice facies. The dotted red line highlights one single crystal; (b) ice-poor sediment (sands and gravels) with suspended and crustal cryostructures forming a band in the massive ice (white fragments are rock clasts). It represents a close-up of the occasional thin bands of sediments cutting across the ice .....65

**Figure 13.** Comparison between the massive ice body and modern glacier ice (glacier C-93, Bylot Island). The first column shows unprocessed photographs of the ice taken under a plain light with surficial sediment inclusions highlighting the crystal boundaries. The dotted red line highlights one crystal. The second column shows thin sections of ice sample viewed under direct cross-polarized light. The third column shows the microstructure (crystal boundaries) extracted from the thin-section photograph. ....66

**Figure 14.** (a) A thin section of the massive ice viewed under cross-polarized light and a transverse cross section from a scan showing the gas inclusions within the ice. (air is black; ice is dark grey). Photos to the right show patterns of gas inclusions; (b) small (sub-millimetre to millimetre) spherical bubbles (vertical bars are from measuring ruler of microscope stage); (c) small disks up to 6 mm in diameter; (d) coalescent bubbles and small disks all flattened in the same direction.....67

**Figure 15.** Box plots comparing the distribution of ice crystal characteristics (area, long axis, circularity ratio) of horizontal thin sections (H) and vertical thin sections (V) obtained from massive ice samples. C-93 represents data obtained from a sample of modern glacier ice sampled from glacier C-93.....68

**Figure 16.** Box plots showing soluble cation concentration of the massive ice, ice wedge, glacier C-93 and intrasedimental ice sampled within the sediment layers covering the massive ice unit.....69

**Figure 17.** (a)  $\delta^{18}\text{O}$  and D-excess depth profiles including both the massive ice unit and the intrasedimental ice from the sediment cover; (b)  $\delta^{18}\text{O}$ - $\delta\text{D}$  diagram of the massive ice and other types of ground ice (ice wedge, intrasedimental ice), snow and modern glacier ice (C-93) sampled on Bylot Island. The solid red line represents the Global Meteoric Water Line (GMWL):  $d = \delta\text{D} - 8 \delta^{18}\text{O} + 10$ . Also shown are the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values of precipitation recorded at Pond Inlet (1990–1992) and Resolute Bay (1989–1993). LMWL is the local meteoric water line for Resolute Bay, Nunavut ( $\delta\text{D} = 7.8 \delta^{18}\text{O} + 5.33$ ) and Pond Inlet, Nunavut ( $\delta\text{D} = 8.0 \delta^{18}\text{O} + 10.55$ ).....69

**Figure 18.** Sedimentological data from the stratigraphic section. From left to right: a stratigraphic log showing the mean grain size of each unit; gravel, sand and mud percentages; and skewness and sorting.....71

**Figure 19.** Grain-size distribution curves of the sedimentary units (two samples per subunit). The gravel fraction was removed prior to analysis (appendix A1). .....80

**Figure 20.** Map showing the study area on Bylot Island at the northwestern tip of Baffin Island. The map at the left show the surficial geology of the valley and location of the sampling sites. The surficial geology map is based on field work observations, borehole data, interpretation of satellite imagery and digital elevation model and previous maps from Klassen (1993) and Allard

(1996). The yellow star indicates the location of the basal glacier ice exposure presented in this study. Wedge ice and segregated ice were also sampled near the basal glacier ice exposure (< 100 m). The blue triangle indicates the sampling location of C-93 basal glacier ice. The green, orange, and purple circles indicate the location of the stratigraphic sections sampled for paleomagnetic analyses. The altitude of each site is indicated in parenthesis. ....88

**Figure 21.** (a) A photograph showing the general stratigraphy of a well-exposed massive ice sequence ~10 m wide and ~ 5 m thick and the overlying sediments. The lower and lateral contacts of the massive ice have not been reached. The yellow squares show the sampling location of the three ice cores (core 1: 140 cm; core 2: 31 cm; core 3: 36 cm). The thaw depths were measured in late July (2013) with a steel probe at every 10 m along a 150-metre transect at the massive ice exposure, and depths ranged between 52 to 62 cm (mean:  $56.1 \pm 2.9$  cm); (b) Pebble- and cobble-sized clasts were abundant in the ice; (c) S-shaped folds similar to those observed in the basal ice of modern glaciers (see figure S3).....90

**Figure 22.** Log showing vertical variations in cryostructures, density, volumetric ice content (%),  $\delta^{18}\text{O}$  and D-excess isotopic composition of core 1 (140 cm).....91

**Figure 23.** Sedimentological data from the stratigraphic units (basal glacier ice and enclosed sediments). From left to right: a composite stratigraphic log showing the mean grain size of each unit; gravel, sand, and mud percentages; and skewness and sorting. ....92

**Figure 24.** (a) Very small (sub-mm) subspherical air bubbles concentrated and aligned along a crystal boundary, (b) Group of air bubbles in nearly pure ice, and flattened into plates roughly parallel to the sediment layers.....92

**Figure 25.**  $\delta^{18}\text{O}$ – $\delta\text{D}$  diagram of the massive ice and other types of ground ice (ice wedge, intrasedimental ice), snow and modern basal glacier ice (C-93) sampled on Bylot Island. The intrasedimental ice originates from the frozen sediment located above the buried glacier ice exposure. The solid red line represents the Global Meteoric Water Line (GMWL):  $d = \delta\text{D} - 8\delta^{18}\text{O} + 10$ . Also shown are the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values of precipitation recorded at Pond Inlet (1990–1992) and Resolute Bay (1989–1993). LMWL is the local meteoric water line for Resolute Bay, Nunavut ( $\delta\text{D} = 7.8 \delta^{18}\text{O} + 5.33$ ) and Pond Inlet, Nunavut ( $\delta\text{D} = 8.0 \delta^{18}\text{O} + 10.55$ ).....93

**Figure 26.** Lithostratigraphic logs, paleomagnetic data and polarity logs of three sedimentary sequences (A, B, C) located nearby the massive ice exposure. The upper unit of all three sections (A, B, C) consist of colluvial sediments with zones of obvious sediment disturbances that do not reliably record the geomagnetic field. As a result, no inclinations records are available for this unit. In section C, four samples (white circles) were rejected from the dataset, because their MAD values were above 15°, indicating unreliable directional data. Incl: Inclination; Decl: Declination; MAD: Maximum angular deviation; Black (white) intervals denote normal (reversed) polarity.....97

**Figure 27.** a) Location of Bylot Island, Nunavut, Canada and the study area in the Qarlikturvik valley (background: NRCan Landsat-7 orthorectified mosaic, 3 August 2010). The shaded area shows the southwestern plain of Bylot Island, b) Surficial geology of the valley and location of the sampling sites. The net pattern represents the polygonal patterned ground. The white arrows show the direction of paleo ice flow within and around Bylot. The purple dashed line indicates the glacial limit attained during Eclipse glaciation based on the inland extend of Eclipse glacial drift and moraine ridges.....127

**Figure 28.** a) Spatial point density of lake locations in the Qarlikturvik Valley, b) Spatial point density of lakes in the southern plain of Bylot Island. The blue dashed line shows former limits of local mountain glaciations. The purple dashed line shows the limit of the Laurentide Ice Sheet (LIS) as defined by Klassen (1993). The red stars indicate the locations of deep lakes (between 5 m and 21 m) studied in Coté et Pienitz (2010), c) Glacier terminus positions from 120 yr BP to present and d) Tasselled Cap transformation images obtained for Google Earth Engine. The accumulation and movement of sediments in the outwash plain and at the glacier front are represented by red and orange colours on the images (dry and unvegetated areas; TC brightness). Wetter areas, such as eroding cliff, lake shore or river channel are shown in blue (high TC wetness). Vegetated areas are distinguished by teal and yellow colours (TC greenness). .....134

**Figure 29.** Bathymetric maps of the 21 lakes surveyed in the Qarlikturvik valley (background: GeoEye, 2010). The yellow stars show the location of massive ice exposures. Purple triangles on lakes IWT1 and GT1 indicate sediment coring locations. Glacial thermokarst lakes (max. depth > 4 m): GT 1 to 8; Ice-wedge thermokarst lakes (max. depth < 4 m): IWT1 to 13. ....136

**Figure 30.** a) Aerial view of two exposures of buried glacier ice located nearby glacial thermokarst lake GT2, b) Map showing the distribution of active and stable thaw slumps nearby glacial thermokarst lake GT2 (background: GeoEye, 2010). The yellow stars show the location of massive ice exposures, c) We interpreted the massive ground-ice, exposed at the headwall of thaw slump, as buried glacier ice on the basis of cryostratigraphic, crystallographic and geochemical analyses (Coulombe et al., 2019), d) This massive ice exposure was not studied in detail since the ice had been buried again under a thick cover of slump material. However, the ice displays a very similar appearance to the first exposure located one hundred meters away (Photo courtesy of Denis Sarrazin).....138

**Figure 31.** General stratigraphy of cores sampled in lake GT1 (deep glacial thermokarst lake; max. depth = 12.2 m) and IWT1 (shallow ice wedge thermokarst lake, max. depth: 4.1 m). .....140

**Figure 32.** Early June, July and August temperature and dissolved oxygen (DO) profiles for ice wedge thermokarst lake IWT1 (a and d), and glacial thermokarst lakes GT1 (b and e) and GT2 (c and f). The grey rectangle represents the ice cover in early June) and the average maximum ice thickness is 2 m ± 20 cm (measured in 2015 and 2016). For lake GT2 (f), the dashed lines are simply connecting the two available data (from the mooring) but the shape is likely to follow the more detailed temperature profiles. The hatched rectangle indicates the lake bottom. Note that the dates are slightly different in lake GT2. Profiles from lakes IWT1 and GT1 in early June and late summer (August) are adapted from Preskienis et al. (2021; respectively corresponding to lakes BYL66 and BYL36), allowing to compare with lake GT2. ....143

**Figure 33.** Schematic diagram showing the sequence of formation of lakes in terrain underlain by relict glacier ice: a) Stage 0: Initial conditions, b) Stage 1: burial of glacier ice, c) Stage 2: initiation of buried glacier ice melting. The red arrows show the deepening of the active layer and d) Stage 3: lake inception and syngenetic ice growth. The black arrows indicate talik deepening in sub-lake sediments and lateral lake expansion. ....149

**Figure 34.** Sites de glace de glacier enfouie issue de l'inlandsis laurentidien (LIS) et son élévation lors du dernier maximum glaciaire. L'étendue de l'inlandsis (y compris les plates-formes glaciaires) au dernier maximum glaciaire est issue de Dyke et Prest (1987). L'élévation de la

surface du LIS est dérivée du modèle d'état stationnaire de Fisher et al. (1985) qui est basé sur les marges empiriques de la calotte glaciaire. Les élévations de surface sont en centaines de mètres au-dessus du niveau de la mer actuel (les erreurs sont de 5 à 7%). La ligne noire pointillée épaisse est la limite entre les lits déformables dans les régions des Prairies et des Grands Lacs et les lits durs pour les régions de l'intérieur et de l'est. Les lignes rouges en pointillés représentent l'écoulement glaciaire déduit et la zone source pour les quatre sites LIS enfouis. La topographie sous-jacente provient des données d'élévation numériques GTOPO30. Tous les détails sont disponibles dans Lacelle et al. (2018). .....167

## Figures in supporting information

**Figure S1.** Grain-size distribution curves of the sedimentary units (two samples per subunit) overlying the massive ice (buried basal glacier ice).....111

**Figure S2.** (a) Basal glacier ice exposure observed on glacier C-93 (Bylot, Nunavut), (b) Sample of basal glacier ice. The sediment-rich basal ice sampled at the margin of glacier C-93 presented an assemblage of lenticular and micro-suspended cryostructures. The matrix (sandy gravel) was very poorly sorted (sorting values: 6.8 and 9.5) and the grain-size distribution tended to approximately symmetrical ( $\sim 2$ ). (c) Transverse cross section from a scan showing the internal structure of the modern basal ice sample (glacier C-93; ice is dark grey; sediment is light grey, and clasts are white). The CT-scan image shows discrete layers of debris are separated by undulating layers of clear ice. (d) S-shaped folds observed in basal ice layers of the Matanuska glacier, Alaska. The basal ice layers are also characterized by a layered appearance, high debris content, the dominance of silt-sized sediment, and abundant clasts. No sedimentary data are available. (e) Recumbent fold observed in basal ice of the Matanuska glacier. The clear to milky white ice located above the sediment-rich ice is englacial ice.....112

**Figure S3.** Box plots showing soluble cation concentration of the buried basal glacier ice, basal glacier ice of glacier C-93 and intrasedimental ice sampled within the sediment layers covering the massive ice unit.....113

**Figure S4.** Logs and photographs of the stratigraphic sections. P1 represents the general stratigraphy of the buried basal glacier ice and the overlying sediments. Sections A, B and C represent three additional undisturbed stratigraphic sections, located nearby (< 220 m) and above the buried basal ice exposure, where sediment were collected for paleomagnetic and grain size analysis. For section C, no description of unit 1 is available.....115

**Figure S5.** Grain-size distribution curves of the sedimentary units (two samples per subunit) of sections A, B and C.....116

**Figure S6.** Percentage diagram presenting pollen and non-pollen palynomorph records from the buried basal ice exposure .....117

**Figure S7.** Ground penetrating radar profile of lakes K and L .....157

**Figure S8.** Pair Correlation Function results for the a) valley and the b) southern plain. The function  $g(r)$  is plotted against  $r$ , with increasing radii of analysis (in km) away from each lake. The black line shows observed  $g(r)$  values, the dashed red line indicates the mean of the inhomogeneous null model and the grey area  $r$  represents the simulation envelope (95% confidence interval), which is derived from the Monte-Carlo simulations of CSR. Values plotting with the grey envelope are not significantly different from a random spatial arrangement.....159

**Figure S9.** Box plots comparing the morphometric properties (area, perimeter, complexity, elongation index) of deep (>4 m) and shallow lakes (< 4 m) obtained from all the digitised lakes. The thick line marks the median value. The bottom and the top of the box correspond to the first and third quartiles, respectively. The whiskers show the range of observed values that are not within the first and third quartile but not further away than 1.5 times the interquartile range (IQR) from the hinges, and open circles represent outliers.....160

**Figure S10.** Burial of glacier ice at the margin of glacier C-93. The red dashed lines highlight areas of buried glacier ice near glacier C-93.....161



**Figure S11.** Satellite images of lakes. The red line highlights the shoreline and shape of the lakes.....163

## Liste des sigles et abréviations

(Les caractères italiques indiquent les termes en anglais)

A	<i>Area</i>
Al	Aluminium
A/m	<i>Ampere per metre</i>
AMS	<i>Accelerator mass spectrometry</i>
Ar	Argon
ARM	<i>Anhyseretic remanent magnetization</i>
a.s.l.	Above sea level
Be	Béryllium   <i>Beryllium</i>
C	Carbone   <i>Carbon</i>
C	<i>Circularity ratio</i>
Ca	Calcium
cal yr BP	<i>calibrated years before the present</i>
CH <sub>4</sub>	Méthane   <i>Methane</i>
ChRM	<i>Characteristic remanent magnetization</i>
Cl	Chlore   <i>Chloride</i>
CO <sup>2</sup>	Dioxyde de carbone   <i>Carbon dioxyde</i>
CSR	<i>Complete spatial randomness</i>
CT-scan	<i>Computed Tomography Scan</i>
D	Deutérium   <i>Deuterium</i>
<i>D<sub>L</sub></i>	<i>Shoreline development ratio</i>
DMG	Dernier maximum glaciaire
DMM	Droite des eaux météoriques mondiale
DO	<i>Dissolved oxygen</i>
DOC	<i>Dissolved orgavic carbon</i>
EPICA	<i>European Project for Ice Coring in Antarctica</i>

ER	<i>Elongation ratio</i>
Fe	Fer   <i>Iron</i>
GAD	<i>Geocentric axial dipole</i>
GHG	<i>Greenhouse gas</i>
GIC	<i>Gravimetric ice content</i>
GISP2	<i>Greenland Ice Sheet Project 2</i>
GMWL	<i>Global meteoric water line</i>
GPR	<i>Ground penetrating radar</i>
GRIP	<i>Greenland Ice Core Project</i>
H	Hydrogène   <i>Hydrogen</i>
He	Hélium   <i>Helium</i>
HU	<i>Hounsfield unit</i>
IRM	<i>Isothermal remanent magnetization</i>
K	Potassium
Ka	<i>Thousands of years</i>
Kr	Krypton
LGM	<i>Last glacial maximum</i>
LIA	<i>Little Ice Age</i>
LIS	<i>Laurentide ice sheet</i>
LMWL	<i>Local meteoric water line</i>
Ma	<i>Millions of years</i>
MAD	<i>Maximum angular variation</i>
Mg	Magnésium   <i>Magnesium</i>
MIS	<i>Marine isotopic stages</i>
Mn	Manganèse   <i>Manganese</i>
MPT	<i>Mid-Pleistocene transition</i>
Ms	Masse de sol sec
mT	Millitesla
Mw	masse de l'eau de l'échantillon

N	Azote   <i>Nitrogen</i>
Na	Sodium
NDWI	<i>Normalized Difference Water Index</i>
Ne	Néon   <i>Neon</i>
North GRIP	<i>North Greenland Ice Core Project</i>
NRM	<i>Natural remanent magnetization</i>
O	Oxygène   <i>Oxygen</i>
OSL	<i>Optical Stimulated Luminescence</i>
P	<i>Perimeter</i>
ppb	<i>One part per billion</i>
ppm	<i>One part per million</i>
POC	<i>Particulate organic carbon</i>
S	<i>Surface</i>
$t_{1/2}$	<i>demie-vie</i>
TC	<i>Tasseled Cap</i>
TL	<i>Thermoluminescence</i>
V	<i>Volume</i>
$V_i$	<i>Volume of ice</i>
$V_t$	<i>Total volume of the frozen sample</i>
VIC	<i>Volumetric ice content</i>
VSMOW	<i>Vienna Standard Mean Ocean Water</i>
U	Uranium
$\mu$	<i>linear attenuation coefficient of the material</i>
$\mu_w$	<i>linear attenuation coefficient of water</i>
$\psi$	<i>Sphericity index</i>

*À Elsie et Albert*

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# Chapitre 1 : Introduction générale

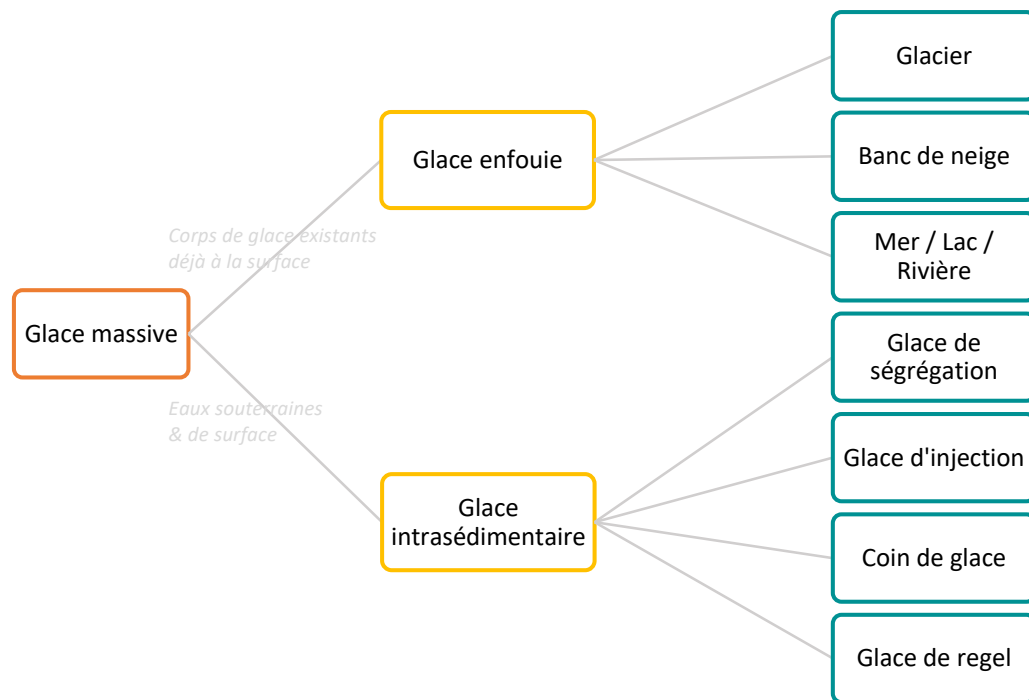
## 1.1 Glace massive et classification

Le pergélisol est un phénomène thermique dont la teneur en glace, la distribution et la forme varient selon la granulométrie du matériel, la géochimie, la teneur en eau et la vitesse d'engel. Cette variation a une grande incidence sur les propriétés et le comportement des matériaux du sol, sur les processus géomorphologiques et sur l'hydrologie de surface et souterraine, en plus de régir en grande partie l'aménagement du territoire et le développement. Plusieurs classifications génétiques des différents types de glace dans le sol furent élaborées jusqu'à ce jour, dont les premières proviennent de chercheurs russes (Shumskii, 1959; Solomatin, 1986; Shumskii et Vtyurin, 1966). Influencées par les travaux des chercheurs russes, des classifications nord-américaines ont par la suite été développées, dont celle proposée par Mackay (1972). Celle-ci distingue dix types de glace de sol en fonction de la source initiale de l'eau (atmosphère, surface ou sous-surface) et du principal processus de mouvement de l'eau souterraine ou de la migration de l'eau dans le sol. Les classifications nord-américaines se distinguaient alors de leurs homologues russes par l'absence d'une catégorie distincte dédiée à la glace enfouie, alors que ce type de glace était déjà étudié et reconnu en Sibérie et au Svalbard (Solomatin, 1977; Grigovyev et Karpov, 1982; Kaplanskaya et Tarnogradskiy, 1976). Mackay (1972) justifiait autrefois cette absence par la rareté de ce type de glace sur le territoire nord-américain. Or, plusieurs études menées principalement dans l'ouest de l'Arctique canadien, en Russie et au Svalbard vers la fin des années '80 ont par la suite adopté plus largement l'hypothèse de l'enfouissement de la glace de glacier pour expliquer la présence de corps massifs de glace préservés dans le pergélisol (Lorrain et Demeur, 1985; French et Harry, 1988; Dallimore and Wolfe, 1988; Kaplanskaya et Tarnogradskiy, 1986; Astakhov, 1986; Vaikmae et al., 1988; Vtyurin, 1989). Par conséquent, les classifications suivantes ont intégré la glace enfouie parmi les différents types de glace présents dans le pergélisol (p. ex. Pollard, 1990).

La glace massive forme aujourd'hui une catégorie distincte parmi l'ensemble des types de glace de sol. La glace est dite massive lorsque la teneur gravimétrique en eau excède 250% (pourcentage du poids final sec) ou une teneur en glace volumétrique supérieure à 95% (van Everdingen, 1988). On retrouve deux sous-catégories de glace massive, soit la glace intrasédimentaire et la glace enfouie (Figure 1; Pollard, 1990; Mackay, 1989). La glace intrasédimentaire est formée par le gel in situ de l'eau dans les sédiments à partir de processus de ségrégation ou d'injection. Quant à la glace enfouie, elle résulte de l'enfouissement de masses de neige ou de glace préexistantes et formées à la surface par des sédiments d'origine diverse (p. ex. colluvion, glaciaire et fluvioglaciaire). Alors que la ségrégation massive a longtemps été le seul mode de formation considéré pour la glace massive en Amérique du Nord, l'hypothèse de l'origine glaciaire de la glace souterraine suscite certainement un intérêt grandissant depuis quelques années et devient de plus en plus répandue et acceptée par la communauté scientifique.

Il existe plusieurs types de glace enfouie, soit la glace de glacier, de banc de neige, de rivière, de lac et de mer, mais la glace de glacier demeure la plus fréquemment rapportée dans la littérature sur le sujet (Dallimore et Wolfe, 1988; St-Onge et McMartin, 1999; Astakhov, 1986; Lacelle et al., 2007; Fritz et al., 2011; Belova, 2015; Lorrain and Demeur, 1985; Murton et al., 2005). Par ailleurs, une distinction est généralement établie entre la glace intraglacière (*englacial ice*) issue du métamorphisme de la neige et la glace basale (*basal ice*) formée par le regel de l'eau à la base d'un glacier. Notons également que la glace basale présente souvent des propriétés cryostratigraphiques très similaires à celles de la glace intrasédimentaire, ce qui rend la distinction génétique parfois difficile (Murton et al., 2005).





**Figure 1.** Classification génétique de la glace massive (modifiée de Pollard, 1990)

## 1.2 Propriétés de la glace de glacier

Il existe principalement deux types de glace de glacier, soit la glace intraglacière issue de la compaction du névé (*englacial ice*) et la glace basale (Harris et Murton, 2005; Lawson, 1979). Les faciès de glace de glacier, de même que les propriétés physiques et géochimiques, sont étudiés de façon détaillée et abondamment par les glaciologues (Rigsby, 1953; Langway, 1958; Allen et al., 1960; Gow, 1963; Thorsteinsson et al., 1997; Durand et al., 2006). Ces connaissances des propriétés de la glace de glacier contemporaine servent de point de comparaison pour identifier ce type de glace préservée dans le pergélisol.

### 1.2.1 Glace intraglacière

D'abord, la glace intraglacière (*englacial*) est formée par la compression progressive des cristaux de neige et de glace sous le poids des couches successives du névé et de la neige. La transformation de la neige en glace se produit lorsque la porosité et la perméabilité diminuent,

alors que la densité augmente (Herron et Langway, 1980; Paterson, 2001). Différents processus, mécaniques et thermodynamiques, sont impliqués dans le métamorphisme de la neige et du névé de glace de glacier, produisant à une grande variabilité de faciès et cristaux de glace. Un certain nombre de caractéristiques typiques peuvent être dégagées des études portant sur les propriétés physiques de la glace de glacier. La classification proposée par Lawson (1979) distingue deux types de glace intraglacière : *englacial diffused ice* et *englacial banded ice*. Le premier sous-faciès représente la glace typique de glacier formée par la densification de la neige et du névé, alors que le deuxième présente des bandes de sédiments grossiers et angulaires déposés à la surface du glacier dans la zone accumulation, puis enfouies sous les neiges hivernales et incorporées à la masse de glace.

La glace intraglacière est composée de cristaux équi-granulaires aux limites assez régulières et bien définies (c.-à.-d. euhédraux à subhédraux), dont la texture (taille et forme) varie selon sa position au sein de la masse glaciaire. La taille peut varier de quelques millimètres à une dizaine de centimètres et augmente de façon assez constante avec la profondeur et l'âge de la glace (Lliboutry, 1964; Herron et Langway, 1980; Sharp et al., 1994; Romanovsky et Cailleux, 1970; Paterson, 2001; Jennings et Hambrey, 2021). Quant à la forme, les cristaux sont typiquement arrondis dans le névé, puis évoluent vers des formes plus complexes à mesure que l'enfouissement des couches progresse. La glace composée de grains grossiers et riche en bulles d'air serait le type de glace le plus commun et abondant dans les glaciers. Tout comme la texture, l'orientation de l'axe-c des cristaux de glace (fabrique, axe optique) varie grandement selon sa position au sein du glacier et le type de dynamique glaciaire (Benn et Evans, 2010). Dans la zone d'accumulation, les couches de glace près de la surface sont formées de cristaux en pleine croissance dont l'orientation est aléatoire, alors que les couches profondes ont acquis deux ou quatre orientations préférentielles (Paterson, 2001). Dans les zones soumises à un fort cisaillement (c.-à.-d. langues glaciaires, zones de courant de glace, zone de séracs), la direction des axes optiques tend à s'orienter dans le sens du mouvement, alors que dans les zones centrales des dômes glaciaires et des inlandsis, on observe des orientations préférentielles verticales très marquées dans la direction de compression (glace immobile soumise à des contraintes de

compaction; Lliboutry, 1964; Durand, 2004; Rigsby, 1951; Gerbi et al., 2021; Fitzpatrick et al., 2014). Notons que l'origine glaciaire de la glace de sol est généralement exclue lorsque les orientations cristallines sont verticales, puisque celles-ci sont associées aux zones centrales des dômes glaciaires. En effet, on présume que la glace de glacier enfouie est habituellement issue des marges glaciaires et donc que l'orientation des cristaux est soit aléatoire ou orientée plus ou moins horizontalement dans la direction d'écoulement de la glace.

La glace de glacier est très souvent riche en bulles d'air héritées de l'air présent dans la neige au moment de sa déposition. Lors de la transformation du névé en glace, les pores jusqu'alors interconnectés deviennent désormais isolés les uns des autres, emprisonnant ainsi l'air dans les bulles isolées. Dans la glace nouvellement formée, 10% du volume de la glace est occupé par l'air, sous forme de bulles d'air (Gow, 1968). La majorité des bulles se retrouvent aux intersections entre les cristaux (joints de grains), bien que certaines d'entre elles peuvent se trouver à l'intérieur même d'un cristal (Rango et al., 2000; Lliboutry, 1964; Gow, 1968). Tout comme les cristaux de glace, la taille, la quantité et la forme des inclusions gazeuses dépendent essentiellement : 1) de la position de la glace au sein du glacier; 2) du type de glace; 3) du type de glacier et 4) du processus de formation de la glace (compaction, fonte-regel, fluage). Or, la quantité de bulles d'air est généralement plus élevée dans les couches superficielles d'un glacier et tend à diminuer graduellement avec la profondeur (Gow et Williamson, 1975). La forme des bulles est généralement sphérique, en particulier dans les couches de glace profondes. Cependant, des bulles d'air allongées de forme presque cylindrique ou aplatie (discoïdales) sont également observées abondamment dans la glace en marge des glaciers. Il a été rapporté que la direction d'allongement des bulles était, dans de nombreux cas, presque parallèle à la direction d'écoulement et à la foliation dans la glace de glacier (Lliboutry, 1964; Nakawo et Young, 1982; Gow, 1968; Hudleston, 2015; Fitzpatrick et al., 2014).

### **1.2.2 Glace basale**

Quant à la glace basale, elle est formée à la base d'un glacier ou d'une calotte glaciaire et présente des caractéristiques différentes de la glace intraglaciaire en raison des conditions et processus différents se produisant à l'interface entre le substrat rocheux et le glacier (Hubbard et

Sharp, 1995, 1989; Knight, 1997; Lawson, 1979). En général, les séquences de glace basale ont une teneur en débris beaucoup plus élevée que la glace intraglaciaire ainsi que des propriétés chimiques et isotopiques distinctes. Plusieurs classifications ont été proposées pour faciliter la description et l'interprétation des faciès de glace basale (Hubbard et al., 2009; Knight, 1997; Fortier et al., 2012; Lawson, 1979; Sharp et al., 1994; Sugden et al., 1987; Knight, 1987; Hubbard et Sharp, 1995; Christoffersen et al., 2006). Bien que ces classifications présentent des variabilités, elles distinguent généralement deux grandes catégories de faciès de glace basale : faciès uniforme (*uniform, dispersed, clotted*) et faciès stratifié (*layered, stratified, banded*). Le premier faciès est principalement caractérisé par une distribution uniforme de débris de taille argileuse à caillouteuse, alors que le second présente une alternance de bandes de glace pure et de bandes riches en sédiments d'épaisseur et d'étendue variables (Hubbard et al., 2009). La zone basale possède généralement des structures internes, à différentes échelles, telles que des failles, des plis, des boudinages ou d'autres structures de déformation plastique.

Ce type de glace est généralement formé de petits cristaux millimétriques et tubulaires, dont le grand axe est parallèle aux stratifications générales de glace et de débris et donc orienté dans le sens général de l'écoulement de la glace (Samyn et al., 2005; Hopkins et al., 2019). La taille des cristaux dépend essentiellement de la teneur en sédiment, alors que la glace pauvre en sédiments aura des cristaux plus gros (1-5 cm), tandis que la glace riche en sédiments aura de plus petits cristaux (Lawson, 1979; Sugden et al., 1987). La formation de cristaux plus larges peut aussi être attribuée à une deuxième phase de cristallisation (Larson et al., 2010; Davesne et al., 2022). Cette zone de glace riche en sédiments présente le plus souvent une très faible concentration de bulles d'air par rapport à la glace intraglaciaire. Lorsqu'il y en a, leur taille sera de l'ordre des millimètres et les bulles d'air sont généralement allongées dans le sens de la foliation générale locale (Hubbard et al., 2009; Samyn et al., 2005; Lawson, 1979; Sugden et al., 1987; Sharp et al., 1994). Cet alignement des bulles indique qu'un simple cisaillement engendre l'étirement et la rotation des bulles jusqu'à ce qu'elles soient parallèles aux directions locales d'écoulement (Samyn et al., 2005; Hooke and Hudlestone, 1978).

### **1.2.3 Glace intrasédimentaire**

En comparaison, la glace tabulaire intrasédimentaire formée dans le sol présente des propriétés cristallographiques différentes, bien que l'apparence générale puisse sembler similaire à celle de la glace de glacier. La ségrégation massive engendre généralement la formation de petits cristaux colonnaires aux limites irrégulières (c.-à-d. anhédraux). La taille des cristaux peut varier selon la vitesse de progression du front de gel et de la teneur en eau du sol. Un apport en eau continu combiné à un front de gel stationnaire permettra une plus grande croissance des cristaux (Pollard and Dallimore, 1988). Les cristaux présentent typiquement une légère orientation préférentielle verticale de l'axe-c parallèle à la direction du front de gel (French et Harry, 1990; Pollard, 1990; Shumskii, 1959; Gell, 1976; Mackay et Dallimore, 1992). Contrairement aux petites lentilles de glace, l'orientation verticale des cristaux de glace de ségrégation massive est généralement plus marquée. Toutefois, cette orientation préférentielle n'est pas toujours clairement définie et l'orientation des cristaux est parfois aléatoire, notamment lorsque le volume de sédiment dans la glace est élevé. Quant aux bulles d'air, elles sont très souvent allongées perpendiculairement au contact sol-glace, indiquant une direction du gel vers le bas (Mackay et Dallimore, 1992).

### **1.2.4 Glace de congère**

Quelques études ont rapporté les propriétés physiques de la neige compactée, enfouie et préservée dans le pergélisol (Fox, 2011; Lacelle et al., 2009; Shumskii, 1959; Pollard et Dallimore, 1988; Tikhonravova et al., 2021) ou dans les dépôts morainiques en marge d'un glacier (Østrem, 1963). Ce type de glace enfouie est typiquement formé de glace transparente à blanche avec de petits cristaux équigranulaires orientés aléatoirement, dont la surface atteint généralement quelques mm<sup>2</sup>. Elle est typiquement peu compactée et présente une densité plus faible que celle de la glace de glacier (Tableau 1). Elle comprend également une concentration élevée de petites (0.1–1 mm) bulles d'air sphériques (Tikhonravova et al., 2021). De plus, ce type de glace comprend des bandes de glace brun pâle presque horizontales qui contiennent des inclusions de sédiments fins et de fragments organiques en suspension dans la glace. Ces inclusions peuvent être déposées sur la neige par des processus de transport éolien ou gravitaire.

**Tableau 1.** Densité de la neige, du névé et de la glace (Van Vliet-Lanoë, 2005)

Type	Densité	Porosité (%)	Pression	Albédo (%)
Cristaux de neige fraîche	0.05 – 0.08	95	Croissante vers le bas	0.8 – 0.9
Neige granulaire	0.10 – 0.40	60 – 70		0.6 – 0.5
Névé	0.55	50		0.7 – 0.8
Glace de glacier	0.89 – 0.90	–		0.7 – 0.24
Glace pure	0.917	0		0.24

### **1.2.5 Glace de rivière, lac et mer**

Ces deux types de glace pourraient être enfouis sous des sédiments et préservés dans le pergélisol, mais aucune étude ne semble avoir rapporté ces types de glace à ce jour. Il faut se tourner vers des classifications et descriptions de glace contemporaine pour obtenir les caractéristiques principales de la glace de lac et de mer. Celles-ci comprennent généralement une superposition de deux couches horizontales de glace (Michel et Ramseier, 1971). La première couche (*primary ice*) est composée de cristaux de glace qui se forment à la surface et se rejoignent pour former une fine couche de glace lorsque l'eau à la surface du plan d'eau commence à geler. Dans des conditions calmes, le gel produit une fine couche (mm) de glace primaire où les axes optiques des cristaux de glace sont alignés verticalement. Lorsque la glace se forme dans des conditions plus agitées, la couche de glace primaire est formée de frasil où les cristaux sont orientés de manière aléatoire (Michel et Ramseier, 1971; Leppäranta, 2010). La croissance se poursuit en profondeur à partir de la couche primaire et la glace se compose principalement de longs cristaux disposés en colonnes (*secondary ice*; Michel et Ramseier, 1971; Leppäranta, 2010; Tikhonravova et al., 2021).

## **1.3 Enfouissement de la glace de glacier**

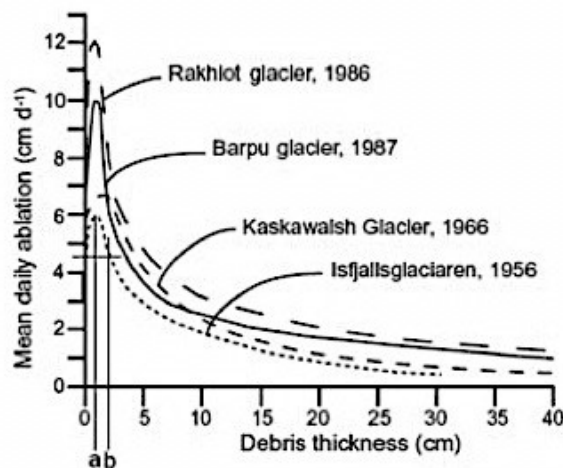
Un très large éventail de processus sédimentaires se produisent à la surface et en marge des glaciers. Les débris supraglaciaires peuvent être incorporés dans la glace de glacier à partir de diverses sources, selon le type de glacier et son environnement. L'accumulation de débris à la surface des glaciers est essentiellement déterminée par la topographie du bassin versant (Benn

et al., 2003). Les débris supraglaciaires sont généralement plus abondants sur les surfaces des glaciers alpins (vallée, cirque, piedmont) et des calottes glaciaires entourées de parois de vallée abruptes ou de nunataks qui peuvent fournir des sédiments par des processus gravitationnels (p. ex. éboulements, avalanches). Outre l'apport de débris, la pente du terrain au front des glaciers joue un rôle important sur le développement des marges glaciaire stagnante (*stagnant ice, dead-ice*; Benn et Evans, 2010). En raison des vitesses d'écoulement nulles ou très faibles, les zones de faible pente favorisent l'accumulation de glace morte, créant indirectement des conditions favorables à l'enfouissement. Les glaciers de piémont, qui débouchent d'une vallée et s'étalent en forme de spatule dans une plaine, sont souvent associés à ces zones de faible pente (Benn et Evans, 2010). Enfin, au-delà de l'enfouissement, le contexte climatique, la localisation et la colonisation végétale jouent un rôle déterminant dans la préservation à long terme de la glace. Les conditions froides des régions polaires et l'accroissement du pergélisol ont permis d'augmenter le potentiel de préservation des masses des glaciers enfouies, contrairement aux glaciers tempérés des plus basses latitudes.

### **1.3.1 Processus thermiques**

L'enfouissement de la glace se produit lorsqu'une couche de sédiments est déposée directement sur la glace, ce qui influence le bilan thermique de surface et le taux d'ablation de la glace. Alors qu'une mince couche de sédiments contribue à accélérer la fonte de la glace, une couche épaisse contribuera plutôt à isoler (Østrem, 1959; Nicholson et Benn, 2006). Au-delà d'une épaisseur critique ( $\sim 2$  cm), les débris protègent les glaciers de la fonte par effet isolant et le taux de fonte de la glace chute de façon exponentielle (Figure 2; Nicholson et Benn, 2006). Cette relation empirique entre l'épaisseur des dépôts supraglaciaires et le taux de fonte de la glace est nommée la «*courbe d'Østrem*» qui fut le premier à démontrer cette relation (Østrem, 1959). Cependant, cette valeur critique peut varier selon les conditions climatiques locales et les propriétés optiques et thermiques des sédiments (Nakawo et Rana, 1999). En milieu périglaciaire, la glace sera préservée dans le pergélisol tant et aussi longtemps que la profondeur de la couche active n'excèdera pas celle de l'épaisseur de la couche sédimentaire recouvrant la glace. La glace devient alors une composante à part entière du pergélisol (Harris et Murton, 2005). La

préservation à long terme de la glace dépend des conditions locales de terrain ainsi que des conditions climatiques (Moorman et Michel, 2000; Benn et Evans, 2010). Peu d'études sur la glace enfouie dans le pergélisol ont décrit les modes d'enfouissement de la glace et caractérisé de façon détaillée les dépôts encaissants et leur mode de déposition. Il faut se tourner vers la glaciologie et utiliser des analogues modernes pour mieux comprendre les mécanismes d'enfouissement et de préservation de la glace de glacier. Dans le cas de glaciers modernes, la glace enfouie est souvent associée à la «glace morte» (*dead ice*), formant des masses de glace stagnante qui ne subissent plus de mouvement glaciaire (Schomacker, 2008; Benn et Evans, 2010). La glace morte se présente essentiellement sous forme de glaciers couverts (*debris-covered glacier*), de moraines à cœur de glace et des masses de glaces d'anciens glaciers pléistocènes qui sont aujourd'hui préservés dans le pergélisol.



**Figure 2.** Exemples de mesures empiriques de la relation entre l'épaisseur des sédiments et la fonte de la glace. (Source: Nicholson et Benn, 2006)

Notons également que la présence de végétation contribue globalement à la préservation thermique et la stabilisation des masses de glace enfouies. Dans les environnements glaciaires, les taux de fonte de la glace sont parfois si lents que la surface peut demeurer stable pendant des périodes assez longues pour permettre le développement complet du couvert végétal (Krüger et Kjær, 2000). La surface des glaciers couverts de débris peut soutenir une grande biodiversité, avec plus de 40 espèces de plantes vasculaires, dont des arbres et des arbustes (Caccianiga et al., 2011). En Alaska, des forêts d'épinettes matures se sont développées sur la glace stagnante et



couverte de débris de certains glaciers (Fickert et al., 2022). La végétation va jouer un rôle important dans le bilan énergétique à la surface du sol et la préservation à long terme du pergélisol et la glace qui s’y trouve (Jorgenson et al., 2010; Anderson et al., 2019). Cependant, l’effet de la végétation sur la température du sol et l’épaisseur de la couche active est très complexe et variable selon les caractéristiques de la couche végétale. L’augmentation de la couverture et de la hauteur de la végétation atténue généralement le dégel du pergélisol en été. Or, dans les zones de végétation arbustive et arborescente, la végétation contribue à augmenter les températures annuelles en favorisant l’accumulation de neige (Porada et al., 2016; Soudzilovskaia et al., 2013; Heijmans et al., 2022; Fisher et al., 2016; Domine et al., 2016). Lors des premières phases de l’établissement du couvert végétal, la surface est d’abord colonisée par des espèces de bryophytes (c.-à-d. mousses) et les lichens. Lorsque sèches, la faible conductivité thermique des mousses leur confère une bonne capacité d’isolation en raison du volume d’air élevé entre les fibres et la faible conductivité thermique du carbone. Subséquemment, l’augmentation de la surface foliaire associée à l’établissement de nouvelles espèces végétales pourra contribuer à réduire le rayonnement au sol, ce qui réduira l’absorption de chaleur dans le sol pendant l’été et protégera le pergélisol (Marsh et al., 2010). Cependant, une croissance de la végétation à port érigé (p.ex. arbuste) pourra retenir davantage la neige, ce qui aura comme effet d’isoler le sol, réduire la perte de chaleur en hiver et ainsi influencer le bilan thermique annuel (Heijmans et al., 2022; Domine et al., 2016). Éventuellement, cela aura comme effet d’augmenter l’épaisseur de la couche active lors de la saison de dégel. Finalement, l’épaisseur de la couche organique du sol sous la couche de mousse joue un rôle tout aussi important dans la préservation du pergélisol et de la glace en raison de la faible conductivité thermique des horizons organiques par rapport à celle des sols minéraux (O’Donnell et al., 2009; Farouki, 1981). La conductivité thermique est étroitement liée à la densité du sol, sa teneur en eau ainsi que son état (gelé vs non-gelé; Andersland et Ladanyi, 2004). En raison de la porosité élevée des sols organiques, la conductivité thermique peut varier considérablement selon la teneur en eau et ainsi influencer la capacité isolante des horizons organiques, comme c’est le cas pour la mousse.

### **1.3.2 Dépôts glaciaires**

Les dépôts glaciaires réfèrent à l'ensemble des sédiments arrachés, transportés, déposés et incorporés à la masse glaciaire. Ils proviennent de débris du lit, des côtés ou de la surface du glacier, de la déformation de la roche mère et même de sédiments provenant des versants de la vallée et intégrés à la glace (Benn et Evans, 2010). L'accumulation de débris supraglaciaires sur les côtés, dans les crevasses et au front du glacier forme des moraines pouvant mener à l'enfouissement de larges portions de glacier (Figure 3). D'un point de vue sédimentologique, ce type de sédiment correspond à un diamicton caractérisé par un mélange de débris rocheux comprenant une matrice, dont la taille des grains est variable et généralement caractérisée par l'absence de stratification (Brodzikowski et van Loon, 1991). La chute de blocs et de débris rocheux le long de parois rocheuses affaiblies par la météorisation (p. ex. gel-dégel) demeure la source principale de débris supraglaciaires dans le cas des glaciers de vallées. Dans les zones non montagneuses, l'apport de débris supraglaciaires est souvent négligeable, à l'exception des glaciers se trouvant dans les zones d'activité volcanique active (p. ex. Islande). Dans les environnements périglaciaires, certaines études ayant rapporté la présence de glace de glacier enfouie ont suggéré qu'un dépôt de till serait à l'origine de l'enfouissement sans toutefois s'avancer sur le type de till (Lorrain et Demeur, 1985; Ingólfsson et Lokrantz, 2003; Fritz et al., 2011).



**Figure 3.** Moraine médiane du glacier Matanuska, Alaska (photo: Michel Sliger)

### **1.3.3 Melt-out till**

L'accumulation *in situ* de sédiments libérés par la fonte de la glace riche en sédiments représente un autre mécanisme d'enfouissement de la glace de glacier (French et Harry, 1988, 1990; Murton et al., 2005; Lawson, 1979; Boulton, 1970). Ce type de till (*melt-out till*) résulte de la fonte ou la sublimation de la glace sur place sans qu'il y ait eu de transport sédimentaire ou de déformation subséquente et forme ainsi une couche isolante de sédiments. La glace basale riche en sédiments possède le meilleur potentiel de préservation, puisque la fonte de cette glace produira rapidement une couche de débris isolante. Dans les conditions idéales, ce type de till conservera sa structure initiale suite à la fonte de la glace. Toutefois, plusieurs études affirment que cette forme est rare, puisqu'il arrive fréquemment que le till s'effondre lors de la fonte, suivie du lessivage du matériel par l'eau de fonte (Lawson, 1979; Paul et Eyles, 1990).

#### **1.3.4 Glaciotectonisme**

Quelques études ont rapporté des masses de glace de glacier enfouies sous des sédiments transportés par chevauchements tectoniques (Manley et al., 2001; Murton et al., 2005; Pollard, 1990; St-Onge et McMartin, 1999; Mackay, 1959). Ce processus engendre une déformation des dépôts qui sont poussés vers l'avant et le haut en réponse au cisaillement attribuable à des pressions exercées par le glacier en mouvement (Harris et Murton, 2005). Lors d'avancées glaciaires, la glace active peut chevaucher une portion de glace stagnante, pour ensuite l'enfouir sous une couche de till de fond suffisamment épaisse pour la préserver de la fonte. Certaines conditions sont favorables au chevauchement lors d'avancées glaciaires : 1) présence d'une unité peu résistante à la base (p. ex. glace morte) ; 2) un changement de topographie (p. ex. pente ascendante) qui génère une compression de la masse de glace et 3) la présence d'eau à la base et à l'intérieur même du pergélisol proglaciaire (Harris et Murton, 2005; Benn et Evans, 2010). Dans le cas de la troisième condition, une pression élevée de l'eau dans les pores du pergélisol sous-glaciaire facilite le chevauchement glaciotectonique en diminuant la résistance au cisaillement, car l'eau non gelée diminue la cohésion du sol en affaiblissant la force des liens entre les cristaux de glace et les particules de sol.

#### **1.3.5 Dépôts fluvioglaciaires**

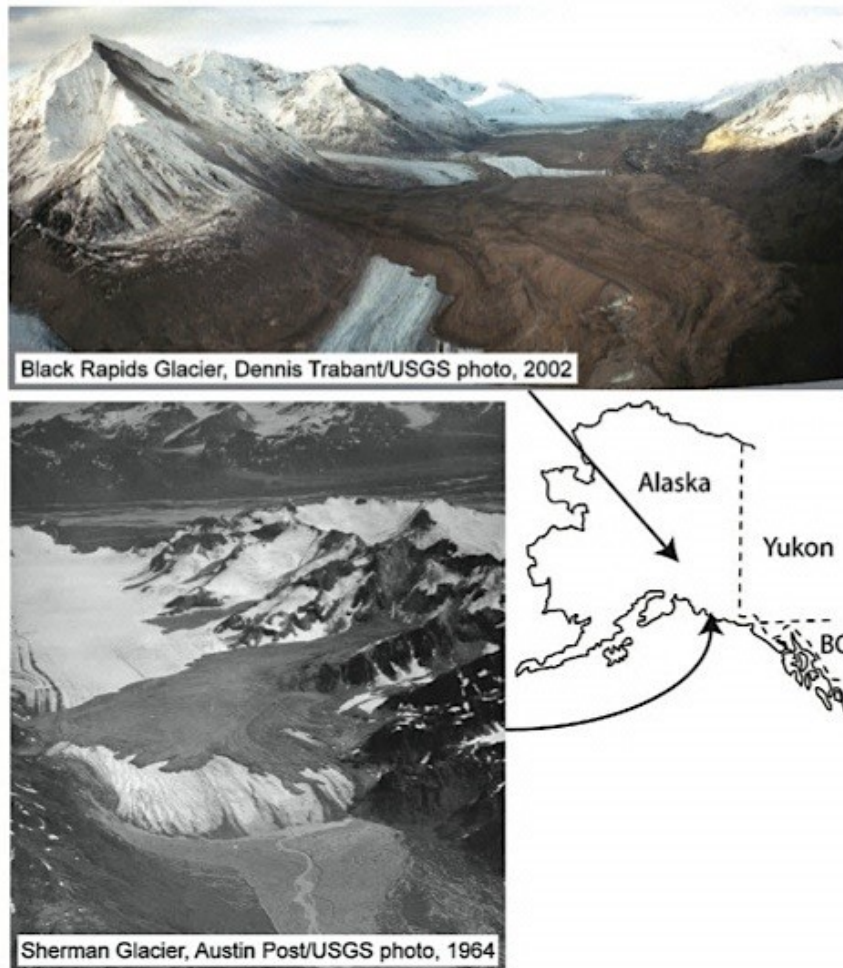
Plusieurs études ont observé de la glace de glacier enfouie sous des sables et graviers fluvioglaciaires (Hambrey, 1984; Kaplanskaya et Tarnogradskiy, 1986; Pollard et Dallimore, 1988). Les dépôts fluvioglaciaires se composent d'une part des dépôts juxtaglaciaires déposés par l'eau de fonte au contact du glacier et d'autre part des dépôts proglaciaires, transportés puis déposés à une certaine distance de la marge du glacier. Le drainage des eaux de fonte permet le transport de larges volumes de sédiments à la surface et à l'intérieur même du glacier. On retrouve, dans les glaciers un véritable réseau hydrologique formé de chenaux supra-, intra- et sous-glaciaires. Dans le cas de l'enfouissement de la glace de glacier, on s'intéressera davantage aux sédiments de contact glaciaire (c.-à-d. juxtaglaciaire), ceux-ci étant déposés par les eaux de fonte directement en contact avec la glace. Typiquement stratifiés, ils sont composés de sables et de graviers plus ou moins bien triés et de gros blocs (> 1 m) peuvent s'y trouver (Brodzikowski et van

Loon, 1991). Le caractère grossier du matériel combiné à la faible quantité de particules fines indique un mode de transport à forte énergie cinétique typique de la dynamique hydrologique juxtaglaciaire. Notons également que les dépôts deltaïques peuvent également contribuer à l'enfouissement de la glace de glacier bien que ce processus d'enfouissement n'ait été rapporté, jusqu'à ce jour, qu'à île Bylot en marge du glacier C-93 (Moorman et Michel, 2000, 2003). Plusieurs études ont d'ailleurs rapporté la présence de delta criblés de dépressions formés par la fonte de la glace de glacier enfouie sous des sédiments fluvioglaciaires en zone deltaïque (*kettled / pitted delta*; Lane et al., 2015; England, 1985; Occhietti, 2010; Benn et Evans, 2010; Dredge et Dyke, 2017).

### **1.3.6 Mouvements de masse**

Les mouvements de masse sont aussi un important processus géomorphologique menant à l'enfouissement de portions de glaciers (Figure 4). Il s'agit généralement de mouvements de masse catastrophiques et rapides (avalanches, coulées de débris, glissement de terrain) qui mobilisent une quantité très importante de sédiments issus des flancs de montagne bordant le glacier (Hewitt, 2009; Shulmeister et al., 2009; Reznichenko et al., 2011; Evans et Clague, 1999; Marangunic, 1972; Deline et al., 2022, 2015). Il s'agit certes d'évènements périodiques et peu fréquents, mais leurs impacts sur les glaciers sont généralement importants en raison de la masse sédimentaire impliquée (Larsen et al., 2005; Marangunic, 1972; Owens, 1992; Deline et al., 2015). Si les débris ne sont pas intégrés dans la glace, ils peuvent être remobilisés vers la moraine frontale par le mouvement glaciaire (Shulmeister et al., 2009; Reznichenko et al., 2011). Le retrait d'un glacier entraîne un relâchement des tensions internes dans la roche des versants adjacents et jouer un rôle majeur dans la dégradation et la rupture des versants rocheux. Aussi, plusieurs études ont démontré que la dégradation du pergélisol de montagne peut contribuer aux instabilités de versants et augmenter l'apport de sédiments vers la surface des glaciers (Kääb et al., 2005; Haeberli et al., 1997; Harris et al., 2001; Allen et al., 2009). Dans les zones montagneuses, les avalanches de neige et de glace peuvent également représenter un processus important à l'origine du transport de sédiments à la surface des glaciers et celles-ci seraient la source de sédiments supraglaciaires la plus importante (Benn et Evans, 2010). De plus, les

mouvements de masses se produisant à la surface des glaciers couverts contribuent également à la remobilisation des dépôts supraglaciaires. Les dépôts supraglaciaires issus de mouvements de masse importants présentent des caractéristiques sédimentologiques qui permettent de distinguer ces événements des moraines conventionnelles (Spedding et Evans, 2002). Les fragments rocheux sont typiquement anguleux et très fragmentés et la surface du dépôt est composée de sédiments plus grossiers que les couches inférieures.



**Figure 4.** Exemples d'avalanches rocheuses sur des glaciers au Yukon et en Alaska (Shugar et Clague, 2011)

## 1.4 Méthodes de caractérisation de la glace

Bien que certains types de glace soient faciles à identifier sur le terrain (p.ex. coins de glace, pingo), il est souvent difficile d'établir une distinction claire entre la glace massive intrasédimentaire et enfouie à partir de simples observations géomorphologiques de terrain, d'autant plus que ces différents types de glace massive coexistent très souvent à proximité l'un de l'autre (Fritz et al., 2011; Pollard, 1990). De plus, la glace de glacier peut se présenter sous différentes formes (glace intraglaciaire, la glace basale, la glace de regel formée dans les crevasses et tunnels) et cette diversité dans les faciès de glace constitue un défi supplémentaire pour l'identification de cette glace dans le pergélisol. L'identification est particulièrement difficile dans le cas de la glace basale. Puisque la glace basale est issue du processus de ségrégation de glace, elle présente souvent des faciès très similaires à ceux de la glace intrasédimentaire, formant des lentilles ou de couches de quelques millimètres à quelques centimètres d'épaisseur.

Il n'existe aucune méthode qui permette d'évaluer avec certitude l'origine de certaines formes de glace massive, mais il est possible de combiner plusieurs analyses physiques, géochimiques et géomorphologiques pour déterminer l'origine de la glace étudiée. Dans le cas de la glace massive d'origine glaciaire, les propriétés cryostratigraphiques, isotopiques et géochimiques peuvent être comparées à celles observées sur des glaciers contemporains pour renforcer ou rejeter l'hypothèse de l'origine glaciaire. En effet, les caractéristiques physiques et chimiques de la glace de glacier sont étudiées depuis longtemps par les glaciologues (Rigsby, 1953; Langway, 1958; Allen et al., 1960; Alley et al., 1997; Thorsteinsson et al., 1997; Durand, 2004; Hudleston, 2015; Jennings et Hambrey, 2021). Il existe principalement quatre méthodes pour étudier la glace souterraine dans le but de déterminer son origine : 1) cryostratigraphie et cristallographie; 2) isotopes stables de l'oxygène et de l'hydrogène; 3) ions et 4) composition gazeuse. Finalement, l'étude du contexte géomorphologique permet souvent d'obtenir des indices supplémentaires pour déterminer l'origine de la glace. Pour étudier les propriétés de la glace et des sédiments encaissants, des travaux d'excavation doivent d'abord être réalisés afin (1) d'évaluer sommairement le volume et l'étendue du corps de glace enfoui sous les dépôts de surface, (2) décrire et photographier les différentes unités cryostratigraphiques de la glace et des

sédiments sus-jacents; puis (3) échantillonner la glace et les sédiments afin de les analyser en laboratoire.

#### **1.4.1 Cryostratigraphie et cristallographie**

L'approche cryostratigraphique vise à décrire les différentes unités observées dans le pergélisol dans le but d'identifier les processus à l'origine de la formation d'un sol gelé, de retracer son histoire, mais aussi estimer les propriétés et le comportement du pergélisol à une échelle plus grande (French et Shur, 2010; Murton et French, 1994; Gilbert et al., 2016; Stephani et al., 2010). Cette approche comprend l'étude des cryostructures, du volume de glace, des cristaux de glace, des bulles d'air, des discontinuités et structures de déformation observées dans un sol gelé. Les cryostructures décrivent la forme, le volume et la distribution de la glace dans les sédiments gelés. Différents types de cryostructures sont associées à différents niveaux de développement et de teneur en glace (French et Shur, 2010). Les cryostructures de porosité et les veines de glace sont peu développées et indiquent un sol pauvre en glace, alors que les cryostructures lenticulaires, réticulaires, suspendues et stratifiées sont bien développées et sont typiquement associées à des sols riches en glace. En laboratoire, les propriétés cryostratigraphiques des échantillons sont analysées qualitativement et quantitativement par l'observation des échantillons en laboratoire et des images obtenues à partir de tomographie assistée par ordinateur (CT-scan ; Calmels et Allard, 2008; Calmels et al., 2010; Dillon et al., 2008; Verpaelst et al., 2017; Lapalme et al., 2017). La quantité de glace peut être déterminée en laboratoire à partir de la mesure des teneurs gravimétriques et volumétriques en glace. La teneur volumétrique en glace (*volumetric ice content*, VIC) peut être mesurée par déplacement d'eau dans un cylindre gradué en immergeant les échantillons scellés sous vide dans l'eau. Le volume est calculé de la façon suivante :

$$VIC (\%) = \frac{V_i}{V_t} \quad (1)$$

$V_t$  représente le volume total de l'échantillon gelé ( $\text{cm}^3$ ) et  $V_i$  représente le volume de la glace ( $\text{cm}^3$ ) estimé à partir de la perte de masse en utilisant la masse volumique théorique de la glace ( $0,9175 \text{ g cm}^{-3}$ ). Quant à la teneur gravimétrique en eau (*gravimetric water content*, GIC), elle se



défini par le pourcentage d'eau contenue dans l'échantillon rapportée au poids du sol sec (ASTM Standard D2216, 2019). Elle est donc calculée en divisant la masse de l'eau dans l'échantillon ( $M_w$ ) sur la masse de sol sec ( $M_s$ ) selon l'équation suivante :

$$GIC (\%) = \frac{M_w}{M_s} \times 100 \quad (2)$$

À l'échelle des cristaux de glace, une analyse détaillée de l'arrangement, la texture (taille et forme), la fabrique (c.-à-d. orientation de l'axe-c), les caractéristiques de l'interface intercrystallin permet d'obtenir des indices quant aux conditions et processus qui prévalaient au moment du gel, notamment la vitesse d'engel et la direction du front de gel (Pollard, 1990; Shumskii, 1959; Langway, 1958; Thorsteinsson et al., 1997; Tison et Hubbard, 2000; Hudleston, 2015). L'analyse cristallographique comprend l'étude de la forme, la taille, l'orientation et la concentration des bulles d'air et des sédiments emprisonnés à l'interface intercrystallin de la glace. La structure cristalline de la glace échantillonnée est typiquement analysée via l'observation de lames minces de glace en lumière polarisée (Langway, 1958; Hubbard et Glasser, 2005). Cette technique est basée sur une propriété optique particulière des cristaux : la biréfringence. Une lame mince est donc placée entre deux filtres polarisants, soit 1) le polariseur et 2) l'analyseur. Alors qu'un rayon lumineux incident traverse les cristaux, l'orientation (anisotropie optique) de chaque cristal lui confère une couleur spécifique. Hubbard et Glasser (2005) et Sinha (1977) décrivent les étapes de préparation des lames minces.

De plus, la nature du contact entre le corps de glace massive et les couches sédimentaires sus- et sous-jacentes est aussi un indicateur utilisé pour déterminer l'origine de la glace étudiée (Pollard, 1990; Mackay, 1989). Un contact graduel est caractérisé par une zone de transition entre la glace et la couche sédimentaire sus-jacente qui varie de quelques centimètres à plusieurs mètres de hauteur. Ce type de contact est typiquement associé à la progression lente du front de gel vers le bas menant à la formation de la glace de ségrégation massive et la glace d'injection. Un contact net (ou discordant) indique plutôt une surface d'érosion et ce type de contact est typiquement associé à la glace enfouie.

### 1.4.2 Isotopes stables O et H

Complémentaire à l'analyse cryostratigraphique, la composition isotopique de la glace, en particulier celle des isotopes stables ( $^{18}\text{O}/^{16}\text{O}$  et D/H), permet d'obtenir des informations sur la composition de la source d'eau, sur la température et les conditions au moment de sa formation (Michel, 2011; Lacelle et Vasil'chuk, 2013; Hubbard et Glasser, 2005). Elle permet ainsi de distinguer différents types de glace massive. Cette méthode repose sur l'analyse des fractionnements isotopiques associés aux différents changements de phase (p. ex. congélation, condensation, évaporation) dans le cycle hydrologique et l'écart à la droite des eaux météoriques. Lors de ces changements de phase, il y a un fractionnement isotopique qui s'opère, modifiant la composition isotopique initiale. À partir de lois physiques bien connues qui régissent la variabilité de ces rapports, il est possible d'interpréter les différences dans les rapports des isotopes stables comme le résultat de certains processus et des conditions dans lesquelles ils se sont produits (p.ex. température, humidité). Par exemple, la température est l'un des principaux mécanismes contrôlant la répartition entre deux isotopes d'un élément (Dansgaard, 1964). Cette thermodépendance permet de relier la composition isotopique des eaux météoriques à des paramètres géographiques ou paléogéographiques, tels que la latitude, l'altitude, la saison et la période climatique (p. ex. eau ou glace héritée des périodes glaciaires). Règle générale, plus le signal isotopique  $\delta^{18}\text{O}$  ( $^{18}\text{O}/^{16}\text{O}$ ) est négatif, plus les températures sont froides. Au fur et à mesure que l'eau s'évapore des océans et se déplace vers les pôles, des pluies successives entraînent un appauvrissement plus important des isotopes  $^{18}\text{O}$  et D qui sont plus lourds. On observe ainsi un appauvrissement des valeurs de  $d^{18}\text{O}$  par rapport à la composition isotopique initiale avec l'augmentation de la latitude et l'altitude. La mesure des rapports isotopiques se fait par spectrométrie de masse et les compositions isotopiques sont exprimées en notation delta par rapport au standard VSMOW. La composition isotopique moyenne de l'océan est choisie comme standard de référence pour ces isotopes avec  $\delta\text{VSMOW} = 0\text{‰}$  par définition. Les valeurs delta sont définies comme suit :

$$\delta_{ech/ref} = 1000 * \left( \frac{R_{\text{échantillon}} - R_{\text{VSMOW}}}{R_{\text{VSMOW}}} \right) \quad (3)$$

où  $R_{\text{échantillon}}$  et  $R_{\text{VSMOW}}$  sont les rapports isotopiques de l'eau ( $^{18}\text{O}/^{16}\text{O}$  ou D/H) des échantillons et VSMOW, respectivement. Le rapport isotopique  $R$  correspond à la concentration de l'espèce

isotopique rare ( $R = {}^{18}\text{O} / {}^{16}\text{O}$ ). Des valeurs négatives signifient que l'échantillon est plus pauvre que la valeur moyenne des eaux océaniques (VSMOW), alors que les valeurs positives indiquent le contraire.

La composition isotopique des précipitations ( $\delta^{18}\text{O}$  et  $\delta\text{D}$ ) montre une relation linéaire entre les deux isotopes stables à l'échelle planétaire, qui s'aligne sur la Droite météorique mondiale, DMM (*Global Meteoric Water Line*; GMWL). La relation générale de la droite de l'eau météorique est définie selon l'expression suivante :  $\delta\text{D} = 8 \delta^{18}\text{O} + 10$  (Craig, 1961). À l'échelle globale, la valeur de la pente ( $s=8$ ) représente le coefficient de fractionnement à l'équilibre pour  ${}^{18}\text{O}$  et D, ce qui signifie que l'appauvrissement ou l'enrichissement en deutérium est toujours supérieur à celui de l'oxygène 18 d'un facteur de 8. Cette différence s'explique par une différence entre les coefficients de fractionnements de l'oxygène et du deutérium au moment du changement de phase. L'ordonnée à l'origine marque l'excès en deutérium ( $d = \delta^2\text{H} - 8\delta^{18}\text{O}$ ; Dansgaard, 1964), ce qui reflète l'équilibre entre la composition isotopique moyenne de la vapeur et sa précipitation. Ce paramètre reflète les conditions climatiques à la surface de la région océanique lors de l'évaporation initiale des masses de vapeur d'eau et permet donc de distinguer l'origine des précipitations. À l'échelle régionale et selon les périodes (p.ex. saisons), la pente et l'ordonnée de cette droite présentent de légères variations par rapport à la GMWL en raison de la source d'humidité locale et des conditions de transport, ce qui entraîne une ligne d'eau météorique locale distincte (*Local Meteoric Water Line*; LMWL; Gat, 2000). Dans les régions polaires, la valeur de la pente est généralement plus faible que celle du GMWL selon les sites. Cette droite météorique peut aider à l'identification des types de glace souterraine puisque les valeurs isotopiques de la glace issue des précipitations (c.-à-d. glace de glacier ou de neige enfouie) se rassemblent le long d'une droite qui reflète les précipitations locales (Craig, 1961; Souchez and Jouzel, 1984; Michel, 2011; Lacelle et Vasil'chuk, 2013). Ainsi, la droite des eaux obtenue a généralement une pente très proche de la LMWL ou la GMWL.

*LMWL Inuvik, NWT, Canada:  $\delta\text{D} = 7.39 \delta^{18}\text{O} - 6.70 \text{‰}$  (Fritz et al., 2022)*

*LMWL Resolute Bay, Nunavut, Canada:  $\delta\text{D} = 7.80 \delta^{18}\text{O} - 5.33 \text{‰}$  (IAEA/WMO, 2016)*

Lors du gel de l'eau dans le sol, sous des conditions d'équilibre de l'eau, la pente de la relation  $^{18}\text{O}$ -D de la glace intrasédimentaire sera quant à elle plus faible que celle des eaux météoriques dues à des phénomènes de fractionnement isotopique qui se produisent pendant le gel (Souchez et Jouzel, 1984; Suzuoki et Kimura, 1973; Jouzel et Souchez, 1982; O'Neil, 1968; Michel et Fritz, 1982). Le gel lent de l'eau engendre un fractionnement isotopique lors du changement de phase, lors duquel les isotopes lourds ( $^{18}\text{O}$  et D) sont préférentiellement incorporés à la glace. La glace est enrichie par ces deux isotopes plus lourds d'environ 3‰ et 20‰, respectivement. Au fur et à mesure que le gel se poursuit, la glace s'appauvrit progressivement à mesure que le  $\delta^{18}\text{O}$  et  $\delta\text{D}$  dans l'eau résiduelle diminuent. Le gel lent de l'eau interstitielle et le front stationnaire associés aux lentilles de glace produit une droite de régression linéaire dont la pente est comprise entre 3 et 7, tout dépendant de la composition isotopique initiale de l'eau (Jouzel et Souchez, 1982; Lacelle et Vasil'chuk, 2013). Or, lors d'un gel rapide, le temps n'est pas suffisamment long pour permettre un fractionnement important et donc, il n'y a pas de différence entre la signature isotopique de la glace et celle de la source d'eau (Michel, 2011).

En somme, l'origine de la glace peut être déterminée en calculant la pente de régression entre  $\delta^{18}\text{O}$  et  $\delta\text{D}$  et en comparant celle-ci à la pente de la GMWL ou la LMWL d'un site proche du site d'étude. Les pentes de la droite de régression des échantillons de glace de glacier intraglacière enfouie devraient s'approcher de celle des eaux météoriques dont les valeurs avoisinent 7 et 8. La glace intradimentaire formée par le gel de l'eau dans le sol sous des conditions à l'équilibre devrait théoriquement produire une droite de régression avec une pente inférieure à 7, tout dépendant de la composition isotopique initiale de l'eau et des conditions de gel (Jouzel et Souchez, 1982). Cependant, une valeur de pente plus faible pourrait également indiquer une origine glaciaire pour la glace formée par le regel sous-glaciaire (glace basale; Jouzel et Souchez, 1982; Souchez et Jouzel, 1984). Dans ce cas, les isotopes stables de l'eau ne permettent pas de dissocier la glace basale enfouie de la glace intrasédimentaire.

### **1.4.3 Composition ionique**

L'analyse de la composition ionique de la glace peut également fournir également des indices quant à son origine (Lacelle et Vasil'chuk, 2013; Fritz et al., 2011). La concentration des principaux cations et anions contenus dans la glace peut être déterminée par spectrométrie d'émission optique avec plasma induit par haute fréquence. Les valeurs sont exprimées en milligrammes par litre. Une mesure de la conductivité de l'eau permet également une mesure indirecte de la concentration ionique, puisque la conductivité dépend de la nature des ions présents dans la solution et de leurs concentrations. Généralement, la glace de glacier enfouie présente de faibles concentrations en ions dissous, de l'ordre de parties par milliard (ppb) à de faibles valeurs de parties par millions (ppm), telles que mesurées sur la glace provenant de glaciers contemporains (Lacelle et Vasil'chuk, 2013). Quant à la glace intrasédimentaire, elle présente des concentrations ioniques plutôt élevées en raison des substances minérales présentes sous forme d'ions dissous contenues dans l'eau souterraine. Des échantillons prélevés dans les coins de glace peuvent également présenter de faibles concentrations ioniques, ce qui concorde avec l'origine atmosphérique de l'eau et une interaction limitée avec la géologie locale (Robinson and Pollard, 1998).

### **1.4.4 Composition gazeuse**

L'analyse de l'abondance relative des gaz atmosphériques piégés dans la glace permet aussi de distinguer la glace d'origine glaciaire de la glace intrasédimentaire formée sur place dans le sol. Cette approche repose sur le principe suivant lequel les inclusions gazeuses de la glace de glacier devraient refléter la composition gazeuse de l'atmosphère au moment de la formation de la glace. D'une part, la concentration de CO<sub>2</sub> dans les bulles d'air présentes dans la glace intrasédimentaire serait de 10 à 100 fois supérieure à celle de l'atmosphère et de l'air emprisonné dans la glace des glaciers modernes, mais similaire à celle du CO<sub>2</sub> du sol (Lacelle et al., 2004; Moorman, 1998). Cette différence significative s'explique essentiellement par l'activité microbienne dans les sols, ce qui cause des teneurs plus élevées dans la glace formée dans le sol (Lacelle et al., 2004). Le CO<sub>2</sub> de l'air piégé dans la glace est prélevé à l'aide d'une méthode

d'extraction sèche qui consiste à broyer la glace sous vide dans un cylindre en acier inoxydable, puis séparé de l'air libéré.

D'autre part, l'analyse de l'abondance relative de l'argon (Ar), l'azote (N<sub>2</sub>) et l'oxygène (O<sub>2</sub>), qui forment quelques-uns des principaux gaz atmosphériques, permettent aussi de distinguer la glace de glacier et la glace intrasédimentaire (Cardyn et al., 2007; Fox, 2010; Lacelle et al., 2007). Cette méthode est basée sur la connaissance de la solubilité spécifique des gaz dans l'eau qui est définie par la constante de Henry. Les ratios O<sub>2</sub>/Ar et N<sub>2</sub>/Ar mesurés dans la glace formée par le gel de l'eau dans le sol différent de ceux mesurés dans l'atmosphère en raison de la dissolution de ces gaz dans l'eau (Tableau 2). Les valeurs pour de nombreux gaz sont publiées dans la littérature. Cette différence de solubilité permet de distinguer la glace formée par le gel de l'eau dans le sol de celle d'origine météorique. L'air piégé dans la glace de glacier devrait refléter la signature atmosphérique moderne de ces gaz, car ceux-ci n'ont pas été dissous lors du processus de densification de la neige et du névé. Les ratios O<sub>2</sub>/Ar et N<sub>2</sub>/Ar de l'atmosphère sont de 22.42 et 83.60, respectivement (Cardyn et al., 2007). Il est admis qu'une fois l'air emprisonné dans la glace, le rapport gazeux est stable, mais en réalité, il peut varier légèrement lorsqu'il se produit un fractionnement gravitaire menant à une diminution des ratios O<sub>2</sub>/Ar et N<sub>2</sub>/Ar ou lorsqu'il y se trouve des microfissures dans la glace (Cardyn et al., 2007).

**Tableau 2.** Valeurs de ratios O<sub>2</sub>/Ar et N<sub>2</sub>/Ar de l'atmosphère et différents types de glace (Cardyn et al., 2007; Lacelle et al., 2007)

Type de glace	O <sub>2</sub> /Ar	N <sub>2</sub> /Ar	Location
Atmosphère	22.42	83.60	
Glace de glacier	22.54 ± 0.48	87.04 ± 2.81	Agassiz Ice Cap, NU
Glace de glacier	18.87 ± 1.03	78.42 ± 2.23	Barnes Ice Cap, NU
Aufeis (glace)	18.79 ± 1.49	68.45 ± 10.66	Northern Yukon, YT
Glace de glacier enfouie	16.81 ± 0.86	81.40 ± 2.51	Chapman Lake, YT
Glace de ségrégation/intrusive massive de la fin du Pléistocène	11.42	75.92	Aklavik Plateau, NWT
Glace de ségrégation massive de la fin de l'Holocène (palse)	3.00 ± 0.17	56.55 ± 2.49	Nunavik, QC

## 1.5 Méthodes de datation de la glace

Il existe de nombreuses méthodes absolues et relatives, généralement combinées, pour dater la glace de glacier et les dépôts glaciogènes (Alexanderson et al., 2014; Hubbard et Glasser, 2005; Briner, 2011). Les séquences stratigraphiques, de même que l'obtention de datations absolues, constituent des outils importants qui permettent non seulement de situer un événement dans le temps, mais aussi de préciser le cadre chronologique régional et d'établir des corrélations avec d'autres enregistrements sédimentaires. Dans le cas de la glace de glacier, il plusieurs méthodes de datation peuvent être utilisées, telles que le comptage des couches annuelles, la modélisation de l'écoulement de glace ou la reconnaissance d'horizon repère (p.ex. tephra, rejets radioactifs) dont l'âge est déjà bien connu. Or, ces méthodes ne peuvent évidemment pas être utilisées dans le cas de la glace de glacier enfouie en raison de l'absence de longues séquences continues de glace. Les méthodes de datation peuvent être généralement réparties en trois groupes : 1) qualitatives et comparative; 2) radiométriques et 3) paléodosimétriques. D'une part, les méthodes de datation qualitatives et comparatives comprennent notamment la dendrochronologie, la stratigraphie, les sédiments de varves et les corrélations de la composition isotopique ( $\delta^{18}\text{O}$ ) dans les carottes de glace et les carottes sédimentaires. D'autre part, les méthodes de datation radiométriques reposent sur le phénomène de désintégration radioactive d'isotopes, comme le radiocarbone ou l'uranium-thorium. Enfin, les méthodes paléodosimétriques (p.ex. luminescence) sont basées sur la quantification des dommages créés dans la matière cristalline par les rayonnements ionisants, comme les méthodes de la luminescence. De façon générale, des avancées récentes en matière de datation permettent d'analyse de petits volumes d'échantillons avec grande précision (Alexanderson et al., 2014). Quelques-unes des méthodes les plus communément utilisées pour dater la glace et les sédiments sont présentées plus bas.

### ***1.5.1 Méthodes de datation radiométrique basée sur la désintégration radioactive***

Ces méthodes reposent sur une loi de décroissance exponentielle radioactive d'isotopes d'un élément chimique. La demi-vie ( $t_{1/2}$ ) correspond au temps nécessaire pour que la moitié des noyaux d'un isotope radioactif se désintègrent. En connaissant le rapport qui existe dans la nature

entre deux isotopes stables et radioactifs, ainsi que la constante de désintégration (demi-vie), il est possible de calculer l'âge de certains matériaux. La datation au radiocarbone (ou datation au carbone 14) est certainement l'une des méthodes de datation absolue les plus utilisées. Après la mort d'un organisme, le carbone 14 commence à se désintégrer lentement à un rythme connu de 5730 ans alors que les formes stables du carbone ( $^{12}\text{C}$ ,  $^{13}\text{C}$ ) demeurent inchangées. Le temps écoulé depuis la mort de l'organisme est calculé en comparant le rapport entre les quantités de  $^{12}\text{C}$  et de  $^{14}\text{C}$  restantes dans un échantillon avec celui d'un échantillon standard de référence. Cette méthode permet de dater les carbonates ainsi que la matière organique (p.ex. charbon de bois, ossements, macro-restes végétaux, tourbe, coquillages) comprise dans la glace ou les sédiments encaissants. Néanmoins, la matière organique est souvent limitée dans les géosystèmes glaciaires de haute altitude et de haute latitude, ce qui limite souvent les possibilités de datation au radiocarbone (Balco, 2011).

Il demeure généralement difficile d'utiliser la méthode du radiocarbone pour dater directement la glace, mais plusieurs études ont daté au  $^{14}\text{C}$  le  $\text{CO}_2$  de l'air emprisonné dans la glace au moment de sa formation (Andree et al., 1984; van de Wal et al., 1990; Wilson et Donahue, 1989; Oeschger et al., 1967). De plus, des techniques de microradiocarbone s'appuyant sur l'analyse du carbone organique particulaire (*particulate organic carbon*; POC) ou du carbone organique dissous (*dissolved organic carbon*; DOC) extrait de la glace sont maintenant de plus en plus utilisées pour obtenir l'âge absolu de la glace de glacier (Fang et al., 2021; Uglietti et al., 2016; Drosch et al., 2007; Hoffmann et al., 2018; Jenk et al., 2006). Le carbone organique est émis directement sous forme solide (c.-à-d., aérosol primaire), puis déposé et intégré à la glace. L'origine et les processus d'émission de ces aérosols sont très diversifiés : mobilisation des débris végétaux, du pollen, des micro-organismes, des spores, de la fraction organique du sol. Il provient aussi de la combustion incomplète lors de la combustion de la biomasse (p.ex. incendies), des combustibles fossiles et des processus pétrogéniques (p.ex. minéraux, graphite). Des aérosols organiques secondaires sont quant à eux issus des VOC émis par des sources marines (Uglietti et al., 2016). Cependant, cette mesure est souvent reconnue comme étant très difficile en raison des faibles des quantités présentes dans les glaces polaires et du risque élevé de contamination



lors de l'échantillonnage, du stockage et du traitement des échantillons et même par l'air ambiant (Fang et al., 2019; Legrand et al., 2013). De plus, la production in situ à haute altitude de  $^{14}\text{C}$  cosmogénique dans la glace par rayonnement cosmique génère des âges plus jeunes que l'âge réel de la glace (Lal et al., 1990; Petrenko et al., 2016). Cependant, la quantité estimée de  $^{14}\text{C}$  produit in situ dans la fraction de DOC serait inférieure à l'incertitude analytique pour la plupart des échantillons (Fang et al., 2021; van der Kemp et al., 2000). Somme toute, la méthode de datation au carbone-14 demeure très précise, mais elle ne permet pas de remonter plus loin que 50 000 ou 60 000 ans et la contamination par du «vieux carbone» peut générer d'importantes incertitudes.

Au-delà de la période couverte par le radiocarbone, il faut se tourner vers d'autres méthodes de datation radiométrique pour remonter plus loin dans le temps. Différents isotopes cosmogéniques « atmosphériques », dont le béryllium-10 ( $^{10}\text{Be}$ ;  $t_{1/2} \sim 1.4 \text{ Ma}$ ), le chlore-36 ( $^{36}\text{Cl}$ ;  $t_{1/2} = 0.3 \text{ Ma}$ ), l'aluminium-26 ( $^{26}\text{Al}$ ;  $t_{1/2} = \sim 0.72 \text{ Ma}$ ) et le krypton-81 ( $^{81}\text{Kr}$ ;  $t_{1/2} = 0.230 \text{ ka}$ ) et l'argon ( $^{39}\text{Ar}$ ;  $t_{1/2} = 269 \text{ y}$ ), ont été utilisés pour la datation des glaces polaires plus anciennes (Nishiizumi et al., 1983; Eichler et al., 2000; Auer et al., 2009; Willerslev et al., 2007; Buizert et al., 2014; Yau et al., 2016; Goldstein et al., 2004; Feng et al., 2019; Fireman, 1986). Le ratio  $^{36}\text{Cl}/\text{Cl}$  a notamment été employée pour dater la glace de pergélisol (coin de glace, glace de ségrégation; (Murton et al., 2021; Gilichinsky et al., 2007; Blinov et al., 2009). En raison de leurs longues périodes, la gamme temporelle de ces isotopes s'étend quelques milliers à quelques millions d'années (p. ex. :  $^{36}\text{Cl} = 3 \text{ Ma}$ ,  $^{81}\text{Kr} = 1.5 \text{ Ma}$ ). Ces isotopes sont produits par interaction entre le rayonnement cosmique et les atomes atmosphériques au cours de réactions de spallation nucléaire dans la haute atmosphère, durant laquelle le noyau éclate sous l'intensité de l'impact (Lal et Peters, 1967). Par exemple, le  $^{10}\text{Be}$  est produit dans l'atmosphère par la spallation de l'oxygène et de l'azote. Ces isotopes sont déposés à la surface des glaciers et calottes glaciaires par les précipitations humides, ou sous la forme de dépôt sec (Lorius, 1990). Le potentiel de ces traceurs isotopiques est reconnu depuis longtemps, mais ces méthodes ont été longtemps limitées par la nécessité d'analyser de gros volumes d'échantillon, des incertitudes importantes sur l'âge et des complications liées à la chimie atmosphérique. Par exemple, la précision

(incertitude) de la datation par les séries de l'uranium est de l'ordre de 16–300 ka (Aciego et al., 2011), alors que celle de la datation par  $^{40}\text{Ar}$  est de  $\geq 180$  ka (Bender et al., 2008). Selon Buizert (2014), la datation absolue de la glace par le krypton serait plus précise pour les échantillons âgés d'environ 1 Ma (incertitude : 100 ka) à 1.5 Ma (incertitude : 300 ka). Au-delà de 1.5 Ma, la datation par l'argon-40 devient toutefois plus précise (Buizert et al., 2014).

Cependant, des progrès technologiques récents permettent désormais d'obtenir un âge fiable avec de plus petits volumes d'échantillons. Par exemple, une équipe de scientifiques a utilisé les isotopes du krypton-81 pour dater des échantillons de glace prélevés en Antarctique, dont l'âge fut estimé à 120 000 ans (Buizert et al., 2014). Le krypton est produit par les rayons cosmiques bombardant la Terre, puis stocké dans des bulles d'air emprisonnées dans les glaces polaires. La comparaison du ratio d'isotopes stable ( $^{83}\text{Kr}$ ) et radioactif ( $^{81}\text{Kr}$ ) permet ainsi d'obtenir l'âge de la glace. Une autre équipe a quant à elle utilisé une méthode de datation basée sur le déséquilibre radioactif au sein de la série de l'uranium. Ils ont obtenu des âges compris entre 80 ka et 870 ka à partir des mesures de la désintégration de  $^{238}\text{U}$  à  $^{234}\text{U}$  de la poussière emprisonnée dans la glace de la carotte EPICA Dome C en Antarctique (Aciego et al., 2011). Il s'agit certes d'une technique prometteuse, mais le degré d'incertitude demeure assez grand, de l'ordre de 16 ka à 300 ka. Au Groenland, la datation par  $^{40}\text{Ar}/^{39}\text{Ar}$  de l'air piégé dans la glace basale des sites de forage GRIP et Dye-3 a permis d'obtenir des âges de  $970 \pm 140$  ka et  $400 \text{ ka} \pm 170 \text{ ka}$ , respectivement (Yau et al., 2016).

### **1.5.2 Méthodes de datation par luminescence**

La datation par luminescence désigne un groupe de méthodes qui mesurent le temps écoulé depuis la dernière exposition de certains minéraux à la lumière du soleil ou à un chauffage suffisant, soit le temps écoulé depuis l'enfouissement du sédiment (Aitken, 1997; Hubbard et Glasser, 2005; Fuchs et Owen, 2008; Lamothe, 2016). Les minéraux des roches et des sédiments ont la capacité d'accumuler de l'énergie provenant du rayonnement cosmique et de la radioactivité ambiante résultant de la décroissance radioactive de l'uranium, du thorium et du potassium contenus naturellement dans les minéraux. Ces irradiations provoqueront des

ionisations qui vont séparer les électrons de son noyau, dont une partie d'entre eux seront ensuite attirés et piégés dans les défauts du réseau cristallin. Il est possible de connaître le nombre total d'électrons piégés par stimulation optique (*Optical Stimulated Luminescence*; OSL) ou par chauffage (*thermoluminescence*; TL) des grains de minéraux extraits d'un échantillon. Sous l'effet d'une chauffe ou d'une stimulation optique, les électrons sont libérés sous forme de lumière émise (luminescence) au moment de l'analyse en laboratoire. La lumière émise suivant la stimulation est proportionnelle à la quantité d'électrons piégés. Le principe de la datation par luminescence repose donc sur la remise à zéro (c.-à-d., horloge dosimétrique) du signal de luminescence de ces minéraux lorsqu'ils sont exposés à la lumière du soleil durant le transport et la sédimentation. Cette méthode permet de mesurer directement l'âge du dépôt, soit le temps écoulé depuis l'enfouissement du sédiment. Les minéraux les plus utilisés pour la datation sont le quartz et les feldspaths en raison de leur abondance et de leur résistance aux différents processus d'altération. La portée chronologique de ces méthodes s'étend bien au-delà de la période couverte par le radiocarbone et dépend des propriétés de luminescence uniques des minéraux examinés. En règle générale, le quartz peut être daté jusqu'à environ 100 à 400 ka et les feldspath de 200 à 300 ka (Fuchs et Owen, 2008; Wallinga et Cunningham, 2013). Cependant, la qualité et la précision des données obtenues dépendent de la mise à zéro complète avant ou pendant la sédimentation (Lamothe, 1996; Gemmell, 1988). Dans le cas où le signal de luminescence de ces minéraux n'est pas complètement remis à zéro, l'âge obtenu sera surestimé. Cette méthode de datation a été utilisée fréquemment pour dater des archives glaciaires (Henriksen et al., 2003; Mangerud et al., 2001; Dubé-Loubert et al., 2013; Murton et al., 2007; Forman et al., 2007; Sugden et al., 1995; Dalton et al., 2022). Il faut toutefois faire preuve de prudence lors de l'utilisation de cette technique de datation pour la glace basale et les dépôts sous-glaciaires, car une grande partie des sédiments transportés à la base des glaciers n'ont pas été exposés à la lumière du soleil pendant le transport et ne sont donc pas entièrement blanchis (mis à zéro) avant le dépôt (Gemmell, 1988). Cependant, quelques rares études ont également mesuré la luminescence sur les grains de quartz et de feldspath extraits de la glace basale riche en sédiments de la calotte glaciaire du Groenland (DYE-3, GRIP) et d'un glacier situé sur l'île d'Ellesmere (Willerslev et al., 2007) et d'une petite calotte glaciaire du Tibet (Zhang et al., 2018).

### **1.5.3 Datation d'exposition par isotopes cosmogéniques**

Les isotopes cosmogéniques (ou nucléides cosmogéniques) sont également utilisés pour mesurer le temps d'exposition ou d'enfouissement des roches et sédiments aux rayons cosmiques (Gosse et Phillips, 2001; Lal, 1988; Schaefer et al., 2022; Bierman, 1994). Les isotopes cosmogéniques sont produits in situ à la surface de certaines roches et sédiments en réponse au rayonnement cosmique, puis préservés dans la structure cristalline des minéraux. La concentration de nucléides cosmogéniques dans une roche est proportionnelle à la durée pendant laquelle elle a été exposée à la surface. La limite analytique de la datation par exposition cosmogénique permet de couvrir des centaines de milliers à des millions d'années tout dépendant de l'isotope utilisé. Les isotopes les plus utilisées sont les suivants :  $^3\text{He}$ ,  $^{10}\text{Be}$ ,  $^{14}\text{C}$ ,  $^{21}\text{Ne}$ ,  $^{26}\text{Al}$ ,  $^{36}\text{Cl}$ . Cette méthode permet, entre autres, de déterminer la chronologie du retrait glaciaire, l'âge des dépôts (p.ex., moraines, deltas, paléorivages) et des surfaces de substrat rocheux érodées par les glaciers, ou encore de quantifier les taux d'érosion (Ivy-Ochs et Kober, 2008; Young et al., 2021; Briner et al., 2006; Crump et al., 2020; Corbett et al., 2016; Refsnider et al., 2013; Christ et al., 2020; Balco, 2011). En Antarctique, des chercheurs ont analysé les concentrations d'isotopes cosmogéniques de débris supraglaciaires pour dater la glace de glacier enfouie (Mackay and Marchant, 2016; Bergelin et al., 2022; Bibby et al., 2016). Les âges obtenus varient de 12 à ~220 ka dans les zones d'écoulement actif et > 1.6 Ma dans les zones de glace stagnante ou d'écoulement très lent (Mackay and Marchant, 2016). Toujours en Antarctique, l'âge de la glace a été estimé à 4.3-5.1 Ma à partir de l'analyse des nucléides cosmogéniques  $^{10}\text{Be}$ ,  $^{21}\text{Ne}$  et  $^{26}\text{Al}$  (Bergelin et al., 2022). La qualité des données obtenues dépend toutefois de différents processus se produisant après la déposition initiale des roches et sédiments, notamment le mouvement (p. ex. le long d'une pente), l'enfouissement et l'exhumation des débris rocheux ou encore la croissance de la végétation à la surface de ces débris. Ces processus peuvent réduire la quantité de radiation reçue par les minéraux, causant ainsi une sous-estimation de l'exposition (Gosse et Phillips, 2001; Mackay et Marchant, 2016). De plus, Corbett et al., (2021) ont démontré que des processus d'érosion intense se produisant sous la calotte glaciaire du Groenland ont altéré de façon importante le signal d'exposition (c.-à-d. faibles concentrations de  $^{10}\text{Be}$ ).

#### **1.5.4 Paléomagnétisme**

L'âge des sédiments encaissants et, indirectement l'âge de la glace, peut être estimé à partir de l'analyse du signal paléomagnétique. Cette méthode repose sur le principe de la rémanence thermomagnétique, selon laquelle certains minéraux magnétiques présents à l'intérieur des sédiments s'alignent en fonction de la direction du champ magnétique terrestre et de son intensité au moment de la déposition (Stoner et St-Onge, 2007; Tauxe, 1993). L'enregistrement du signal se fait dans les premiers centimètres à l'interface eau-sédiment et on appelle cet intervalle de quelques centimètres, la « *lock-in depth* » (Tauxe, 1993). Cette méthode permet de déterminer l'âge relatif d'une séquence sédimentaire en corrélant celle-ci avec l'échelle de temps de polarité géomagnétique (*Geomagnetic Polarity Time Scale* ; GPTS Opdyke et Channell, 1996; Singer, 2014; Ogg, 2020) et d'autres enregistrements paléomagnétiques bien datés dans la région d'étude. L'enregistrement du vecteur magnétique (inclinaison, déclinaison et paléointensité) est synchrone à l'échelle mondiale et indépendant du contexte environnemental, c'est-à-dire que des sédiments marins, lacustres ou fluvioglaciaires enregistreront le même signal paléomagnétique (Stoner et St-Onge, 2007). L'objectif de cette analyse consiste donc à identifier les potentiels d'inversion magnétiques enregistrés dans les couches sédimentaires. Il est généralement admis que l'analyse du signal paléomagnétique n'est possible qu'à partir de sédiments dont la fraction granulométrique est fine (Stoner et St-Onge, 2007). Dans les régions arctiques, cette méthode de datation s'avère particulièrement intéressante, car l'amplitude des variations d'inclinaison et de déclinaison est typiquement plus grande près du pôle Nord, ce qui facilite l'identification de marqueurs chronostratigraphiques (Lisé-Pronovost et al., 2009). De plus, les dépôts glaciaires de l'Arctique peuvent fournir des enregistrements paléomagnétiques de qualité en raison de leur faible teneur en matière organique. Ils contiennent généralement une excellente source de magnétite et de titanomagnétites, ceux-ci étant d'excellents minéraux magnétiques capables d'enregistrer l'intensité et l'orientation du champ magnétique à une époque donnée (Stoner et St-Onge, 2007; Darby, 2003). Quelques études réalisées dans l'ouest de l'arctique canadien et en Alaska ont permis d'associer des dépôts terrestres à des épisodes glaciaires du Pliocène et Pléistocène inférieur (Barendregt et Vincent, 1990; Vincent et al., 1984; Duk-Rodkin et al., 2010; Barendregt et al., 1996; Wang et

Evans, 1997; Kaufman et al., 1991; Huston et al., 1990). À l'île Bylot, des analyses paléomagnétiques ont été réalisées pour estimer l'âge de la forêt fossile formée durant un interglaciaire à la transition Pliocène-Pléistocène vers 2,4–2,8 Ma (Piroux, 2005; Guertin-Pasquier, 2012). En Sibérie, la géochronologie des carottes de sédiments du lac El'gygytgyn a été en partie déterminée à partir du paléomagnétisme (Melles et al., 2012; Haltia et Nowaczyk, 2014). La magnétostratigraphie à haute résolution de ces séquences sédimentaires comprend plusieurs changements de polarité et permet de remonter dans le temps jusqu'à environ 3.6 Ma.

### **1.5.5 Acides aminés**

La datation par racémisation des acides aminés est une technique de datation relative qui repose sur le principe selon lequel les protéines contenues dans les os et les coquilles sont affectées par des réactions chimiques qui dépendent en partie du temps (Demarchi et al., 2011; Miller et al., 2013). Les acides aminés forment les unités de base des protéines. Elles sont composées de petites molécules organiques qui comprennent un groupe acide carboxylique et un groupe aminé. Une petite quantité d'acides aminés est piégée dans les ossements de vertébrés et les coquilles d'invertébrés (mollusques, foraminifères) au moment de leur formation. Dans les organismes vivants, les acides aminés sont uniquement lévogyres (L). À la mort d'un organisme, ils vont progressivement se transformer en acides aminés dextrogyres (D) (Miller et al., 2013; Demarchi et al., 2011; Miller et Brigham-Grette, 1989; Hare et Abelson, 1968). Au bout d'un certain temps, un équilibre s'établit entre les deux formes (D et L). La conversion des acides aminés se nomme la racémisation ou épimérisation. Le rapport entre la quantité d'énantiomères D- et L- (valeur D/L) indiquera donc le temps écoulé depuis la mort d'un mollusque. En raison du caractère progressif de la racémisation, les spécimens les plus dégradés sont également les plus vieux. Tout dépendant des acides aminés, la racémisation est très lente, d'où son application comme méthode de datation. La fenêtre de temps couverte par cette méthode s'étend de quelques milliers à plusieurs millions d'années. En raison de leur racémisation lente, certains acides aminés spécifiques (isoleucine, leucine, valine, acide glutamique) sont préférés à d'autres acides aminés à des fins de datation (Miller et al., 2013). La vitesse de racémisation des acides aminés dépend du type d'acide aminé et des conditions environnementales telles que la

température, dans lesquelles les échantillons carbonatés sont conservés. La vitesse de racémisation augmente notamment avec la température. La racémisation est plus longue (~ 1–2 Ma) dans les régions arctiques que dans les régions tropicales (Miller et al., 2013). Il est donc nécessaire de comparer les âges relatifs de fossiles d'une même région. Pour utiliser cette méthode, les échantillons doivent provenir d'un système fermé permettant la préservation des acides aminés, c'est-à-dire qu'il ne doit pas y avoir de perte ou d'enrichissement en acides aminés par des processus de lessivage, attaques bactériennes ou encore par contamination par d'autres acides aminés (Miller et al., 2013). La préservation des acides aminés dans les couches de sédiments marins est généralement optimale dans en eaux profondes, en raison des températures relativement stables en comparaison avec les conditions terrestres variables (Kaufman et al., 2008). Puisque la racémisation dépend de certaines conditions environnementales, cette méthode est surtout utilisée pour corréler des dépôts soumis à des conditions de température similaires, plutôt que de déterminer un âge précis (Miller et Brigham-Grette, 1989; Miller et Andrews, 1985). Cette technique permet entre autre de préciser la chronologie et géométrie des glaciations de même qu'estimer la durée des épisodes glaciaires à un site donné (Klassen, 1993; England, 1999; Lemmen et England, 1992; Fickert et al., 2022; Briner et al., 2009; Dalton et al., 2016, 2022; Barendregt et al., 2021).

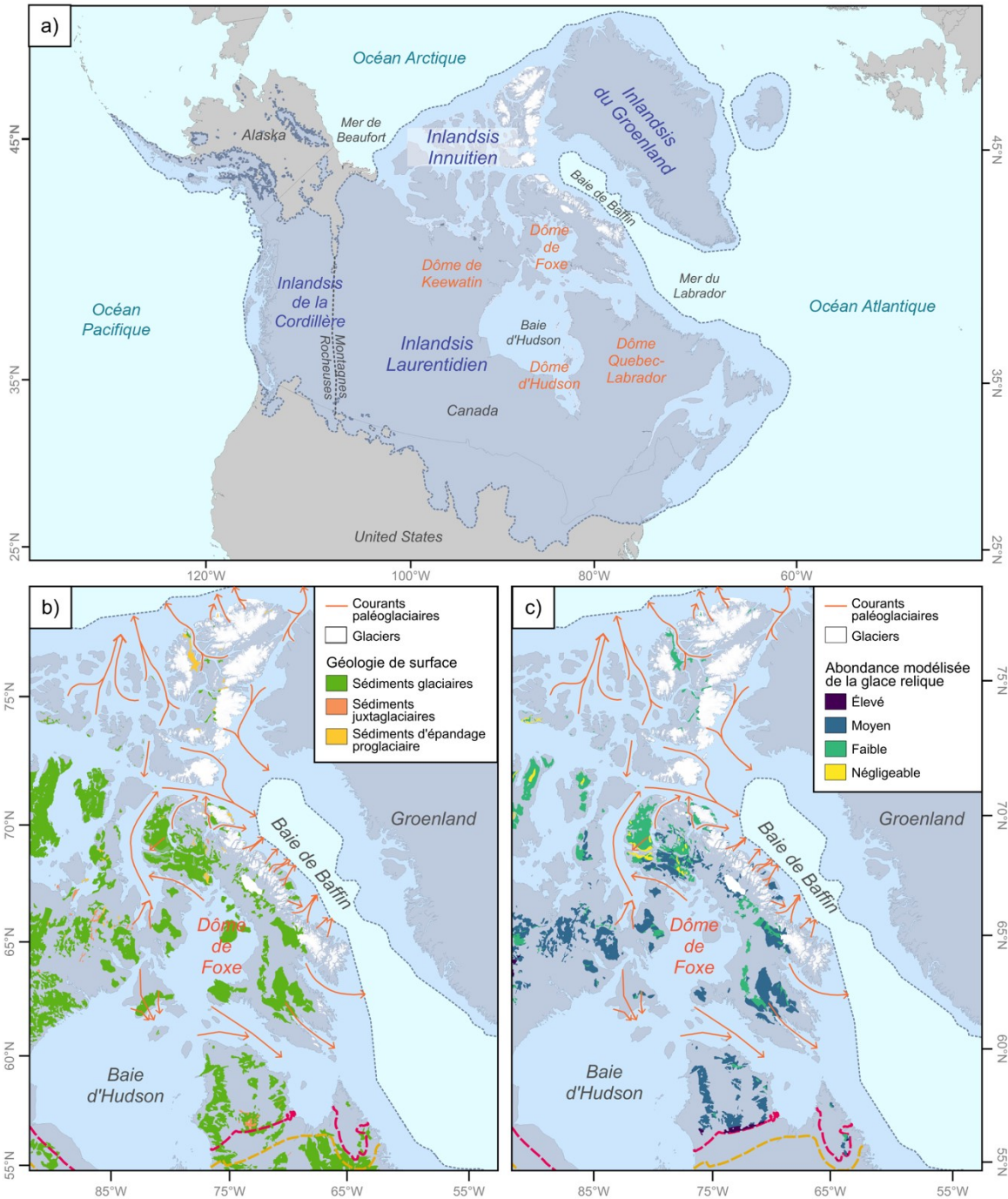
## 1.6 Contexte géomorphologique et répartition géographique

Au-delà des propriétés physiques et chimiques de la glace, le contexte géomorphologique fournit des indices importants quant à la présence de la glace de glacier enfouie. La répartition spatiale de la glace de glacier enfouie suit généralement un patron qui est étroitement lié à la répartition spatiale et l'épaisseur des dépôts meubles d'origine glaciaire ou fluvioglaciaire, aux formes glaciaires ainsi qu'aux limites glaciaires actuelles et passée (Kokelj et al., 2017a; Lakeman et England, 2012; Dyke et Savelle, 2000; Benn et Evans, 2010). O'Neill et al. (2019) ont d'ailleurs présenté une cartographie à l'échelle nationale de la répartition de la *glace relique* préservée dans le pergélisol (Figure 5c). Le dépôts glaciaires typiquement associés à la glace de glacier enfouie incluent notamment les complexes morainiques (Tm), le till en couverture continue (Tb) et le till bosselé (Th, *hummocky moraine*; Figure 5b). Cependant, les unités de till en couverture mince et

discontinue (Tv; épaisseur < 2m) ne devraient pas être considérées dans l'analyse de la distribution spatiale de la glace de glacier enfouie puisque ce type de dépôt est typiquement trop mince et discontinu pour favoriser la préservation de la glace. De plus, l'étude des dépôts supraglaciaires sur les glaciers alpins actuels démontrent que les dépôts de versants (c.-à-d. colluvions) contribuent massivement à l'enfouissement de portions de glaciers. Il est également admis que le potentiel de préservation de la glace enfouie devrait être faible ou nul dans les zones de submersion marine et lacustres suivant la déglaciation. Par conséquent, les surfaces marines et lacustres postglaciaires de grande superficie ne devraient pas contenir de masses de glacier enfouie.

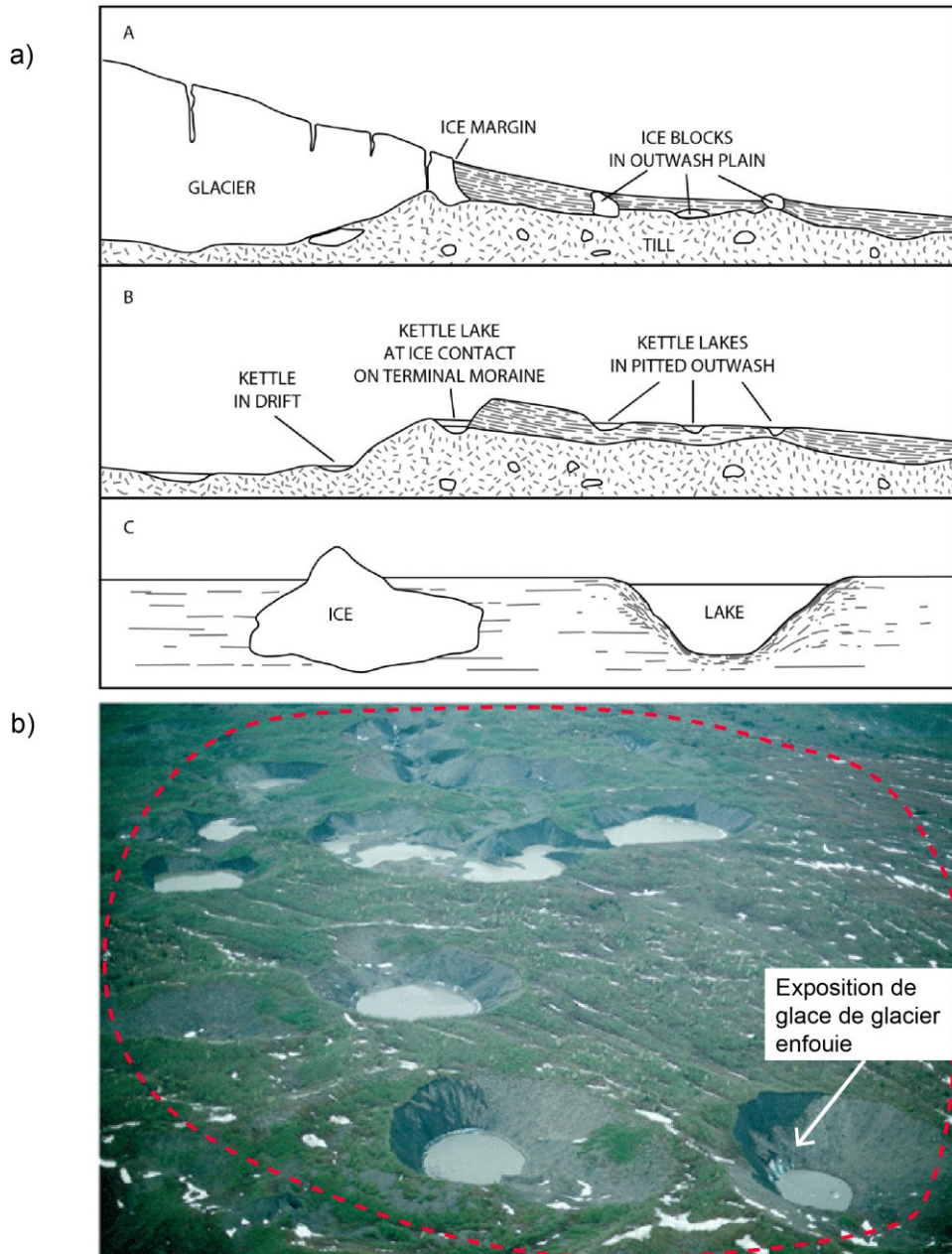
La présence de formes glaciaires peut être un indice important de la présence de glacier enfouie. Les plus communes sont les moraines à cœur de glace (*ice-cored moraines*) qui forment des crêtes et des collines de matériel très grossier ayant un noyau de glace de glacier (Lukas, 2011; Østrem, 1964; Souchez, 1971; Tonkin et al., 2016; Brook et Paine, 2012; Ewertowski et Tomczyk, 2015; Houmark-Nielsen et al., 1994). La glace de glacier enfouie est plus rarement associée à des formes sous-glaciaires, telles que les eskers et drumlins, mais pourrait être plus commune que ce qui est rapporté dans la littérature (Schomacker et Kjær, 2007; Schomacker et al., 2006; Ewertowski et al., 2019; Chandler et al., 2020; Moorman, 2005). La fonte sur place, partielle ou complète, de la glace enfouie crée un paysage irrégulier formant un ensemble de bosses et de creux de matériel morainique et fluvioglaciaire. Plusieurs termes désignent ces moraines bosselées dépourvues d'éléments linéaires et héritées, la plupart du temps, par la fonte de la glace stagnante : moraine de décrépitude, moraine bosselée (*hummocky moraine*), moraine de glace morte (*dead-ice moraine or terrain*), moraine de désintégration (*disintegration moraine*), moraine de stagnation (*stagnation moraine*; (Eyles et al., 1999; Benn et Evans, 2010).





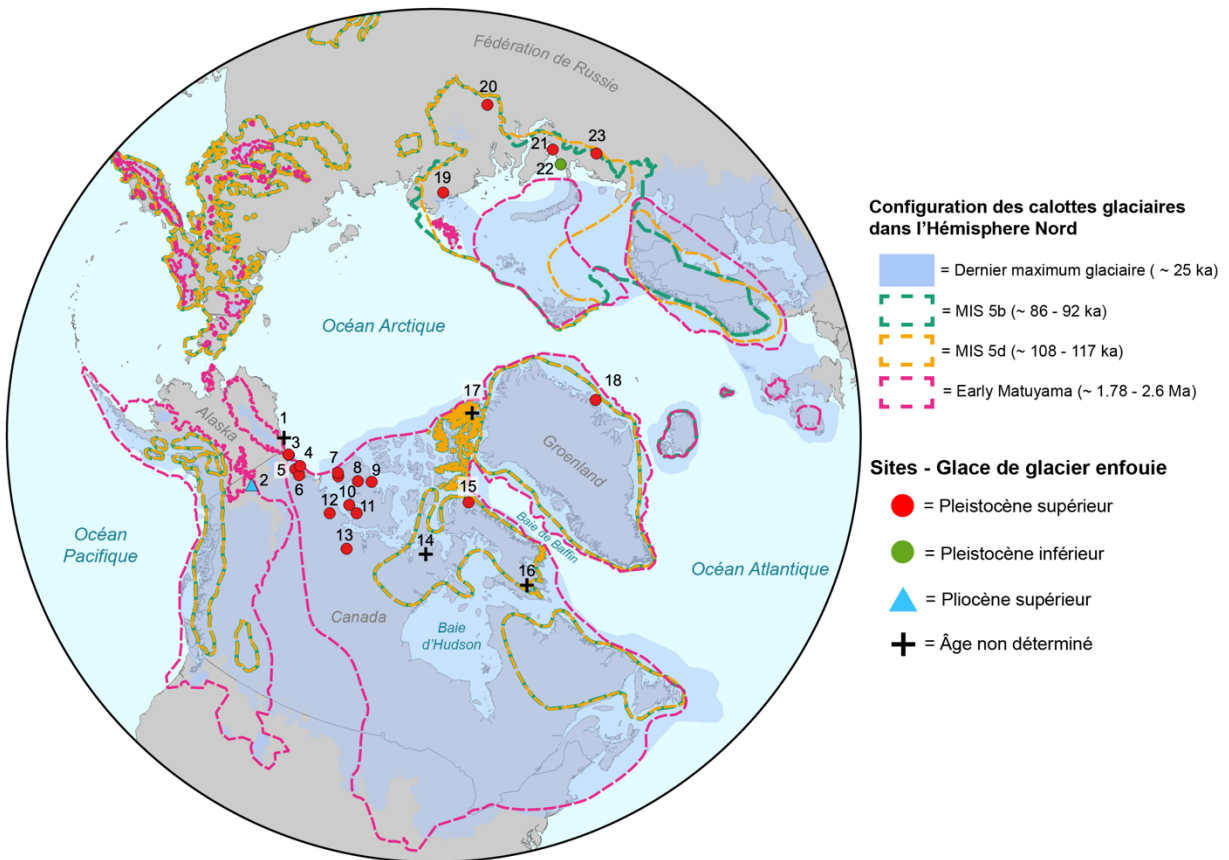
**Figure 5.** a) Configuration des inlandsis au dernier maximum glaciaire selon Dalton et al., 2022. B) Carte de répartition des dépôts de surface couramment associés à la glace de glacier enfouie. Les flèches oranges indiquent la direction des courants de glace (De Angelis et Kleman, 2007; Margold et al., 2015); c) Cartographie à grande échelle qui illustre une estimation de l'abondance de glace relique basées sur la modélisation d'O'Neill et al. (2019).

Les kames et kettles représentent aussi des formes glaciaires fréquemment associés à la glace de glacier enfouie (Benn et Evans, 2010; Bennett et Glasser, 2009). Formées par l'accumulation de matériel d'origine fluvioglaciaire directement sur la glace, les kames prennent souvent la forme de monticules ou de terrasses, isolées ou regroupées en marge des glaciers ou calottes glaciaires. Quant aux kettles, ils forment des dépressions dans une plaine d'épandage ou un autre dépôt glaciaire suite à la fonte sur place d'un gros bloc de glace enfoui sous des sédiments (Figure 6; Benn et Evans, 2010). Au fur et à mesure que la glace enfouie fond, la dépression s'élargit. La topographie en kames et kettles correspond à une zone où les deux formes coexistent pour former un paysage caractérisé par une alternance de reliefs positifs (kames) et négatifs (kettles; Benn et Evans, 2010). Les définitions de kettle mettent souvent l'accent sur la fonte de blocs ou culots de glace isolés et séparés de la langue glaciaire active (Fay, 2002; Bennett et Glasser, 2009; Götz et al., 2018). Toutefois, lorsque ces dépressions sont groupées, des observations sur le terrain suggèrent la présence de masse de glacier étendue et continue (c.-à-d. glace tabulaire) enfouies sous une couverture de sédiments glaciogènes. Enfin, la présence de formes thermokarstiques, telles que des glissements de terrain, des lacs de thermokarst glaciaire, peut également indiquer la présence de glace de glacier enfouie. Bien que la plupart de ces formes géomorphologiques soient héritées de la fonte de la glace de glacier enfouie, la fonte peut être partielle et des blocs de glace peuvent persister sous les dépôts.



**Figure 6.** a) Formation des kettles (source : [https://gq.mines.gouv.qc.ca/lexique-stratigraphique/quaternaire/kame-et-kettle\\_en/](https://gq.mines.gouv.qc.ca/lexique-stratigraphique/quaternaire/kame-et-kettle_en/)) ; b) Photographie aérienne montrant une douzaine de dépressions en expansion se formant à la surface de la glace stagnante recouverte de débris dans la moraine médiane du glacier de Béring, forêt nationale de Chugach, en Alaska. (source : <https://pubs.usgs.gov/of/2004/1216/jkl/jkl.html>). La flèche blanche montre une exposition de glace de glacier enfouie dans la paroi d'un lac de kettle. La présence de glace combinée à la nature groupée des lacs suggèrent que la glace enfouie est continue entre les lacs et probablement présente dans l'ensemble de la zone délimitée par la lignée pointillée rouge.

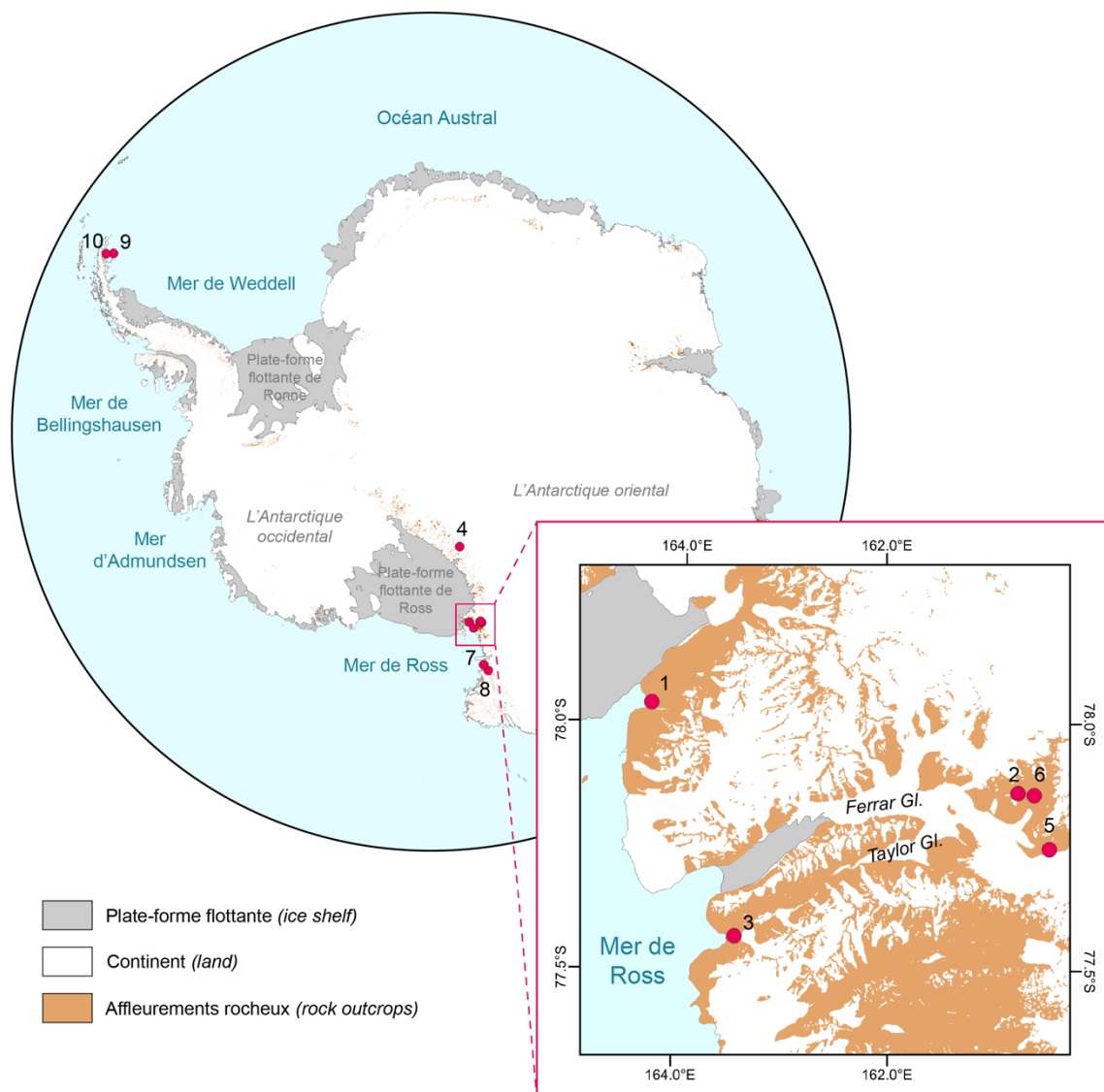
Dans l'Arctique canadien, la majorité des études portant sur ce type de glace a été réalisée dans l'Ouest et au centre de l'archipel arctique, notamment le long de la côte du delta du Mackenzie et la péninsule de Tuktoyaktuk (Figure 7; Dallimore et Wolfe, 1988; Pollard, 1990; Fritz et al., 2011; Murton et al., 2005; French et Harry, 1990; Murton et al., 2004). Dans l'est de l'Arctique, quelques rares observations ont été rapportées au sud de la baie de Pelly (Ozier, 2011; thèse PhD), à Iqaluit (Allard et al., non publié) et au nord du lac Hazen (Pachen et al., non publié), mais celles-ci n'ont pas fait l'objet d'études. Cette région est donc largement sous-représentée, malgré l'héritage glaciaire important de cette région de l'Arctique canadien. En Sibérie, les expositions de glace de glacier enfouie rapportées dans la littérature sont concentrées le long de la mer de Kara, notamment sur les péninsules de Yamal, Yugorski et Gydan ainsi dans la vallée et l'estuaire de la rivière Yenisei (Belova, 2015; Astakhov et al., 1996; Astakhov et Svendsen, 2002; Astakhov et Isayeva, 1988; Grigovyev et Karpov, 1982; Henriksen et al., 2003). En Antarctique, la glace de glacier semble également largement répandue dans les secteurs alpins et les vallées sèches de McMurdo au sein de la chaîne de montagne Quartermain (Figure 8; (Swanger, 2017; Lacelle et al., 2011; Gardner et al., 2022; Sugden et al., 1995; Bibby et al., 2016; Péwé, 1960). C'est dans cette région que les plus vieilles masses glaciaires enfouies ont été observées et dont l'âge a été estimé entre 1.5 et 8 Ma (Bibby et al., 2016; Sugden et al., 1995). En zone glaciaire et proglaciaire, de nombreuses études rapportent la présence de moraines à cœur de glace en marge des glaciers contemporains et calottes glaciaires (Ewertowski et Tomczyk, 2015; Østrem, 1964; Souchez, 1971; Tonkin et al., 2016; Sletten et al., 2001; Lukas et al., 2005; Lukas, 2011; Schomacker et Kjær, 2007; Johnson, 1971). Ces environnements représentent un analogue moderne de l'enfouissement et la préservation des masses de glaciers.



**Figure 7.** Configuration des calottes glaciaires dans l'hémisphère nord (Batchelor et al., 2019) et localisation des expositions de glace de glacier enfouie rapportées dans la littérature :

1) île Barter, Alaska (Kanevskiy et al., 2008); 2) Lac Chapman (Lacelle et al., 2007); 3) île Herschel (Fritz et al., 2011; Pollard, 1990); 4) île Pullen (Murton et al., 2005); 5) Lac Ya-Ya (Dallimore et Wolfe, 1988); 6) Lac Eskimo (French et Harry, 1990); 7) Moraine Sand Hills, île de Banks (Worsley, 1999; French et Harry, 1988); 8) Moraine Jesse, île Victoria; (Lakeman et England, 2012); 9) Loch Point, île Victoria (Lorrain et Demeur, 1985); 10) Péninsule de Wollaston, île Victoria (Dyke et Savelle, 2000); 11) Lac Ammalurtuq (Sharpe, 1992); 12) Moraine du lac Bluenose (St-Onge et McMartin, 1999); 13) Lac Contwoyto (Wolfe, 1998); 14) Sud de la baie de Pelly (Ozier, 2011); 15) Île Bylot (Coulombe et al., 2019; Moorman et Michel, 2003); 16) Iqaluit (Allard et al., non publié); 17) Lac Hazen, île d'Ellesmere (Z. Panchen, non publié); 19) Péninsule de Taymir (Alexanderson et al., 2002); 20) Ladyanaya Gora (aussi Ice Hills; Solomatin, 1977; Vaikmae et al., 1988; Astakhov et Isayeva, 1988; Astakhov, 1986; Grigovyev et Karpov, 1982); 21) Baie de Baydaratskaya (Belova, 2015); 22) Cape Shpindler (Ingólfsson et Lokrantz, 2003); 23) Lac Komovoye (Henriksen et al., 2003). Source des données géospatiales (fond de carte) : NOAA National Centers for Environmental Information (2014)





**Figure 8.** Localisation des expositions de glace de glacier enfouie rapportées dans la littérature en Antarctique : 1) Garwood valley (Levy et al., 2013; Stuiver et al., 1981a; Pollard et al., 2002); 2) University valley (Pollard et al., 2012; Lacelle et al., 2011; Lapalme et al., 2017a); 3) Taylor valley (Péwé, 1960; Gardner et al., 2022; Hall et al., 2000; Levy et al., 2018); 4) Ong valley (Bibby et al., 2016; Bergelin et al., 2022); 5) Kenmar valley (Swanger, 2017); 6) Beacon valley (Sugden et al., 1995; Marchant et al., 2002; Kowalewski et al., 2006); 7) Northern Foothills – Simpson Crags (Guglielmin et French, 2004); 8) Northern Foothills – Boulder Clay Glacier (Guglielmin et French, 2004); 9) Île Seymour (Ermolin et al., 2004); 10) Île Vega (Ermolin et al., 2004). Source des données géospatiales (fond de carte) : (Scientific Committee on Antarctic Research, 2022)

## 1.7 Des traces d'anciennes glaciations

Depuis quelques millions d'années, le climat oscille entre des périodes glaciaires, durant lesquelles des inlandsis se sont développés sur l'Amérique du Nord et l'Eurasie et des périodes interglaciaires avec un climat proche du climat actuel ou plus chaud. Le refroidissement progressif du climat ayant mené à une augmentation draconienne du volume des glaces polaires et l'implantation des cycles glaciaires-interglaciaires dans l'hémisphère Nord marque le début du Quaternaire et du Pléistocène (~2,588 Ma; Ravelo et al., 2004; Walker et al., 2018). Des reconstructions des calottes de glace présentes dans l'hémisphère nord ont permis de documenter un minimum de vingt glaciations majeures au cours des 2,6 derniers millions d'années (Ehlers et al., 2018). Les périodes glaciaires et interglaciaires alternent d'abord selon un cycle de 41 ka, puis le rythme des glaciations change progressivement à 100 ka lors de la transition du Pléistocène moyen (~ 800 ka, MPT; Ehlers et Gibbard, 2003). Les glaciations les plus étendues remontent à 790–928 ka (MIS 24-21 à aujourd'hui) dans l'hémisphère nord et sont associées à des cycles glaciaires-interglaciaires de 100 ka (Head et Gibbard, 2015; Hughes et Gibbard, 2018; Batchelor et al., 2019). Les avancées scientifiques récentes dans l'analyse et la datation des dépôts sédimentaires et des carottes de glace démontrent qu'il existe un asynchronisme dans le comportement des calottes glaciaires à l'échelle mondiale, notamment dans la chronologie des maxima glaciaires (Hughes et al., 2013; Böse et al., 2012; Larsen et al., 2013; Hughes et al., 2020). Cette asynchronisme est aussi observé entre les calottes continentales et locales (c.-à-d., glaciers alpins; (Gillespie et Molnar, 1995; Zech et al., 2003). Ces différences sont attribuées en partie aux différents patrons de répartition des températures et des précipitations influencés par les facteurs géographiques régionaux (masses d'eau, relief, types de surface).

Ces glaciations successives ont laissé des traces dans le paysage sous forme de dépôts, formes glaciaires ou encore de masses de glace enfouie et préservée sous des dépôts glaciogènes. Un modèle récent de l'étendue et la configuration des paléo-inlandsis ayant recouvert l'ensemble de l'hémisphère Nord durant le Quaternaire a été proposé Batchelor et al. (2019). Les configurations de ces calottes continentales lors des périodes glaciaires les plus récentes (DGM, MIS 5b et 5d), de même que lors d'une période glaciaire ayant eu lieu au début du Pléistocène

sont présentées à la figure 5. Pour l'Amérique de Nord, un modèle de l'évolution des inlandsis depuis 115 ka jusqu'au dernier maximum glaciaire (~25 ka) a été proposé par Dalton et al. (2022) pour l'Amérique du Nord. Lors du dernier maximum glaciaire, l'inlandsis laurentidien a recouvert environ la moitié de l'Amérique du Nord, ce qui lui confère le statut du plus grand inlandsis dans l'hémisphère nord. Selon le modèle actuel, cet inlandsis est le résultat de l'agglomération de plusieurs dômes d'accumulation et de dispersion glaciaire au cours du Wisconsinien supérieur (figure 7a). Quatre dômes majeurs coalescents (Keewatin, Foxe, Québec-Labrador et Hudson) ont été inventoriés et ceux-ci auraient atteint leur extension maximale à des moments différents (Dyke et Prest, 1987; Dyke et al., 1982; Ives et Andrews, 1963; Shilts et al., 1979; Hillaire-Marcel et al., 1980; Tyrrell, 1898).

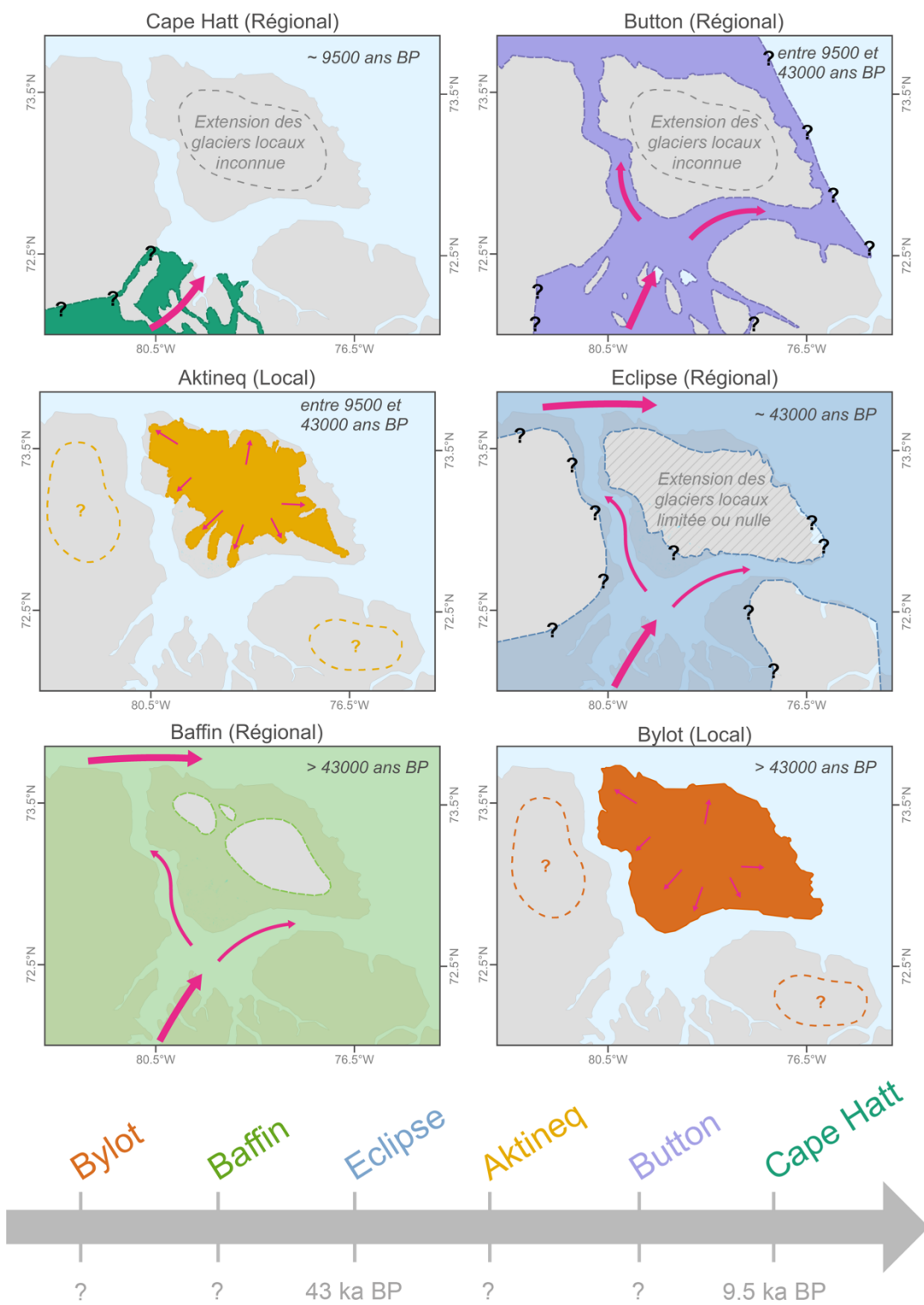
Cependant, plusieurs interrogations subsistent concernant l'étendue maximale et l'évolution des calottes glaciaires avant le dernier maximum glaciaire (DMG) puisque les dépôts les plus anciens sont remobilisés par les glaciations plus récentes (Stokes et al., 2015; Batchelor et al., 2019). Les traces de ces anciennes glaciations sont ainsi très fragmentées, à la fois dans l'espace et dans le temps. Des archives terrestres de ces premières glaciations pléistocènes ont été étudiées à quelques endroits dans l'ouest de l'Arctique canadien (Duk-Rodkin et al., 2004, 2010; Evans et al., 2021; Froese et al., 2000; Barendregt et al., 2010; Duk-Rodkin et Barendregt, 2011; Vincent et al., 1984; Vincent, 1990) tandis que des enregistrements de ces premières avancées glaciaires sont encore plus rares ou mal datés dans le centre et l'est de l'Arctique canadien (Brigham, 1983; Refsnider et al., 2013; Fyles et al., 1998). Dans l'est de l'île de Baffin, l'étude des séquences complexes de sédiments marins et glaciaires des formations de *Clyde Foreland* et *Qivitu* ont permis de détecter plusieurs avancées glaciaires ayant précédées le dernier maximum glaciaire (Refsnider et al., 2013; Miller et Andrews, 1985; Feyling-Hanssen, 1985; Nelson, 1982). Les plus vieilles unités de la formation de *Clyde Foreland* correspondent à des avancées glaciaires de l'inlandsis laurentidien sur l'île de Baffin vers le début de Pléistocène ou vers la fin du Pliocène. En Sibérie, la séquence sédimentaire du lac El'gygytgyn est sans doute la plus longue période temporelle continue (3.6 Ma) pour les régions continentales qui furent englacées (Andreev et al., 2013; Zhao et al., 2022; Wennrich et al., 2016).



Dans l'hémisphère Nord, les restants de glaciers, issus de glaciations ayant précédé le dernier maximum glaciaire, sont aussi très rares, voire inexistants. Les plus vieilles glaces de glacier connues proviennent de la calotte du Groenland. En effet, les carottes de glace du forage profond NorthGRIP au Groenland ont permis de reconstituer l'histoire du climat au cours des derniers 123 ka (North Greenland Ice Core Project members, 2004). Dans le centre du Groenland, la glace basale riche en sédiments à la base des carottes de glace du Greenland Ice Core Project (GRIP) et Dye-3 sont aussi anciennes que 950-970 ka et 400 ka, respectivement (Yau et al., 2016; Willerslev et al., 2007). D'autres études précédentes ont estimé que la glace basale située de 3022 à 3029 m de profondeur dans la carotte de glace GRIP, pourrait être aussi vieille que 2,4 Ma et serait associée au tout début de la construction de la calotte glaciaire du Groenland (Souchez et al., 1994, 2006). Les reconstructions stratigraphiques et la datation  $\delta^{40}\text{Ar}$  réalisées à partir de la carotte de glace du Greenland Ice Sheet Project 2 (GISP2) suggèrent que la glace basale la plus profonde aurait plus de 237 ka (Suwa et al., 2006). Dans les environnements périglaciaires, la glace de glacier enfouie est typiquement associée à la dernière avancée des inlandsis laurentidien, eurasiatique et fennoscandien (Coulombe et al., 2019; Astakhov et Svendsen, 2002; French et Harry, 1988; Ingólfsson et Lokrantz, 2003).

Bien que les connaissances de l'histoire glaciaire dans l'Arctique soient limitées pour le Pléistocène inférieur et moyen, il en est autrement pour la dernière glaciation (wisconsinienne). Les reconstructions de la configuration et l'évolution de l'inlandsis laurentidien et de la déglaciation depuis le DGM (~26,5 ka) dans l'Arctique sont beaucoup plus abondantes dans la littérature (Svendsen et al., 2004; Dyke, 2004; Dyke et al., 2002; Margold et al., 2015; Nielsen et Rasmussen, 2018; Lønne et Lyså, 2005; Stokes et al., 2012; Simon et al., 2014; Tarasov et al., 2012; England et al., 2006). Dans l'est de l'archipel arctique canadien, quelques études se sont intéressées à la géologie quaternaire de l'île Bylot et du nord de l'île de Baffin (Klassen, 1993; Klassen et Fisher, 1988; Margold et al., 2015; Dyke, 2000; Dyke et Hooper, 2001; Hodgson et Haselton, 1974). D'après les travaux de Klassen (1993), plusieurs glaciations locales et régionales ont façonné le paysage de l'île Bylot au cours du Quaternaire récent (Figure 9). Au total, sept

épisodes glaciaires distincts, entrecoupés de courts épisodes de réchauffement, ont été identifiés à partir de l'étude des dépôts de surface et du modelé glaciaire (Tableau 3 ; Klassen, 1993). Quatre épisodes glaciaires furent associés à des calottes glaciaires régionales basées sur l'île de Baffin, alors que trois épisodes furent associés à des glaciers de montagne locaux. La glaciation de Baffin demeure la plus ancienne glaciation majeure rapportée jusqu'à ce jour, mais son âge demeure inconnu. Vieille de 43 000 ans, la glaciation d'Éclipse représente quant à elle la plus récente des périodes glaciaires majeures de la glaciologie quaternaire de l'île Bylot. Klassen (1993) a utilisé cette dernière glaciation majeure connue, soit celle d'Éclipse, comme événement pivot autour duquel la chronologie des glaciations de l'île Bylot est reconstituée. La datation de cette glaciation régionale fut réalisée essentiellement à partir de la datation au radiocarbone et de la racémisation des acides aminés de coquillages remaniés et prélevés dans les dépôts glaciaires (till).



**Figure 9.** Carte des limites glaciaires rapportées par Klassen (1993) et Dyke et Hopper (2001). Les flèches roses représentent la directions des courants

**Tableau 3.** Glaciations locales et régionales documentées à l'île Bylot, de la plus récente à la plus vieille (Klassen, 1993)

Périodes	Type de glaciation	Description
Néoglaciaire	Locale	<ul style="list-style-type: none"> <li>• Maximum néoglaciaire : ~ 120 ans</li> </ul>
Cape Hatt	Régionale	<ul style="list-style-type: none"> <li>• Corrélation avec le dernier maximum glaciaire</li> <li>• Wisconsinien tardif ou du début de l'Holocène</li> <li>• La glace s'écoulait vers le Nord à partir de l'île de Baffin.</li> <li>• Corrélation avec le <i>Baffinland stade</i> (Andrews et Ives, 1978)</li> <li>• ~9500 ans BP</li> <li>• Acides aminés : 0.10–0.15</li> </ul>
Button	Régionale	<ul style="list-style-type: none"> <li>• Un ou plusieurs épisodes glaciaires distincts d'une calotte glaciaire issue de l'île de Baffin.</li> <li>• Évidences stratigraphiques limitées.</li> <li>• Extension maximale des glaces limitée marges périphériques de l'île et probablement confinés aux chenaux entourant l'île.</li> <li>• Pas d'âge précis, mais se situe entre 9500 et 43 000 BP</li> <li>• Acides aminés : 0.15–0.20</li> </ul>
Aktineq	Locale	<ul style="list-style-type: none"> <li>• Avancée majeure de certains glaciers de montagne locaux durant une phase de déglaciation au cours de la glaciation d'Eclipse.</li> <li>• Pas d'âge précis, mais se situe entre 9500 et 43 000 BP</li> </ul>
Eclipse	Régionale	<ul style="list-style-type: none"> <li>• Se produit tout juste avant le début du Wisconsinien</li> <li>• Glace étendue dans les basses terres du sud de l'île Bylot jusqu'à environ 300 à 375 m au-dessus du niveau de la mer, 10 à 20 km à l'intérieur des terres au sud de l'île.</li> <li>• Courants de glace dans Lancaster Sound, Navy Board Inlet et Eclipse Sound</li> <li>• &gt; 43 000 ans BP</li> <li>• Acides aminés : 0.25–0.30</li> </ul>

Baffin	Régionale	<ul style="list-style-type: none"> <li>• Représente un ou plusieurs épisodes glaciaires distincts</li> <li>• Occupe la majeure partie de l'île</li> <li>• Pourrait peut-être précédé l'Interglaciaire de Salmon River</li> <li>• Épisode reconnu par la présence et la distribution d'erratiques à l'intérieur des terres et au-dessus des limites glaciaires de la glaciation d'Eclipse et par des dépôts glaciaires d'âge pré-éclipse dans quelques sections stratigraphiques.</li> <li>• Erratiques : Quartz sandstone (Adams Sound Formation, Baffin Island)</li> <li>• Pas d'âge précis, mais plus vieux que 43 000 ans BP</li> <li>• Acides aminés : 0.6–0.8</li> </ul>
Bylot	Locale	<ul style="list-style-type: none"> <li>• Représente un ou plusieurs épisodes glaciaires distincts</li> <li>• La calotte glaciaire et les glaciers de vallées ont couvert la majeure partie de l'île</li> <li>• Pas d'âge précis, mais plus vieux que 43 000 ans BP</li> <li>• Acides aminés &gt; 0.8</li> </ul>

## 1.8 Structure et objectifs de la thèse

Rédigée dans le cadre du programme de doctorat en géographie, cette thèse est présentée en cinq chapitres, dont trois sous forme d'articles scientifiques. Le premier chapitre consiste en une mise en contexte qui expose les particularités de la glace de glacier enfouie et son potentiel en tant qu'archive environnementale, les modes d'enfouissement de la glace de glacier ainsi que les méthodes principales utilisées pour l'étude de cette glace. Les trois chapitres suivants constituent le cœur de l'étude, soit les articles rédigés en anglais et qui ont été soit acceptés, soumis, ou qui seront soumis prochainement pour publication dans des revues scientifiques avec évaluation par des pairs. Le dernier chapitre fait état des principales conclusions des trois articles scientifiques, de la portée de la thèse ainsi que des perspectives de recherche.

Sur la base des connaissances actuelles sur la glace de glacier enfouie dans les milieux glaciaires et périglaciaire, de la localisation et de l'apparence générale des expositions de glace massive, nous avons émis l'hypothèse suivante : **Le pergélisol de l'île Bylot contient des masses de glacier enfouie issues d'anciennes glaciations Pléistocènes et cette glace ce a influencé l'évolution géomorphologique des vallées glaciaire et des basses-terres de l'île.** Ainsi, les deux objectifs principaux de cette thèse de doctorat sont de déterminer (1) l'origine et le contexte d'enfouissement deux expositions de glace massive préservée dans le pergélisol de l'île Bylot (Nunavut) et (2) le rôle joué par ce type de glace sur l'évolution géomorphologique de l'aire d'étude, particulièrement sur la formation des lacs. Les objectifs spécifiques et résultats des différents chapitres sont détaillés ici-bas.

D'abord, la glace de glacier enfouie dans le pergélisol coexiste très souvent avec différents types de glace intrasédimentaire formés in situ dans le pergélisol, bien que leur genèse soit très différente. Comment distinguer la glace de glacier enfouie des autres types de glace qu'il est possible de trouver dans le pergélisol de l'Arctique canadien? Quelles sont les caractéristiques typiques de la glace de glacier enfouie dans le pergélisol? À quel moment la glace s'est-elle formée et quelles informations paléoclimatiques pouvons-nous soutirer de cette glace? Les deux premiers chapitres de cette thèse se penchent sur ces questions en étudiant les propriétés physicochimiques et l'âge des expositions de glace massive découvertes à l'île Bylot. La localisation, le contexte géomorphologique et l'âge des sédiments localisés au-dessus de la glace suggéraient initialement que les deux sites étudiés présentent des expositions de glace de glacier enfouie qui seraient associées à deux épisodes glaciaires distincts, soit du début du Pléistocène et du dernier maximum glaciaire. Le **Chapitre 2** de cette thèse se penche sur l'une des expositions de glace de glacier enfouie d'origine intraglaciaire située dans une vallée glaciaire de l'île Bylot. Les objectifs spécifiques de ce chapitre étaient d'abord (1) de caractériser les propriétés géocryologiques et géochimiques de la glace et de les comparer à d'autres types de glace présents dans l'aire d'étude afin de déterminer son origine. (2) Ce chapitre explore ensuite les modes de déposition ayant mené à l'enfouissement de la glace et sa préservation dans le pergélisol. L'exposition de glace massive et les sédiments sus-jacents ont été caractérisés selon l'approche

cryostratigraphique permettant une description détaillée des cryofaciès et cryostructures, de la granulométrie des sédiments, de la structure cristalline de la glace (forme, taille, orientation), des patrons de bulles d'air et de sédiments dans la glace ainsi que la composition isotopique (O-H) et cationique. L'âge de la glace a été estimé par datation au radiocarbone et paléomagnétique des sédiments recouvrant la glace enfouie. Les résultats ont confirmé la présence de glace de glacier typique formée par compression de la neige (*englacial ice*), puis enfouie par des sédiments fluvio-glaciaires de contact et des colluvions. La composition d<sup>18</sup>O de la glace était nettement inférieur à celle de la glace de glacier contemporaine et suggère que la glace provient d'un courant glaciaire de l'inlandsis Laurentidien qui a débordé sur les basses-terres de l'île Bylot vers la fin du Pléistocène.

Le **Chapitre 3** porte sur la deuxième exposition de glace basale enfouie découverte en bordure du plateau (alt. ~500 m) bordant la vallée Qarlikturvik. De façon similaire au chapitre 2, l'objectif principal de ce troisième chapitre visait également à (1) caractériser les propriétés géocryologiques et géochimiques de la glace afin de déterminer l'origine et (2) déterminer l'âge approximatif de la glace. Pour répondre à ces objectifs, la glace fut d'abord caractérisée à l'aide de l'approche cryostratigraphique. Puis, l'âge relatif de la glace a été estimé à partir de datations radiocarbone et du signal paléomagnétique des sédiments situés au-dessus de l'unité de glace de glacier. Enfin, des analyses complémentaires, telles que des analyses polliniques et l'identification des diatomées préservés dans la glace basale permettent de préciser l'âge de la glace de même que d'investiguer un lien potentiel entre la glace et l'unité de forêt fossile sous-jacente afin d'améliorer notre compréhension de la succession des événements. La glace présentait des caractéristiques très différentes de la glace présentée dans le chapitre 2. La glace a été interprétée comme étant de la glace de glacier d'origine basale en raison de la teneur élevée en sédiments et de la présence de cryofaciès très similaires à ceux observés dans les couches de glace basales de glaciers contemporains. L'analyse paléomagnétique des sédiments situés stratigraphiquement au-dessus de la glace démontre une succession de polarité positive-inverse-positive, donnant un âge minimum de 0.773 (c.-à-d. limite Buhnes-Matuyama) Ma. La glace de glacier enfouie est donc assignée au chron de Matuyama, mais pourrait correspondre à un

épisode glaciaire s'étant produit vers la fin du chron normal de Gauss (2.595–3.032 Ma). Ainsi, la glace basale enfouie serait associée à un épisode glaciaire du Pléistocène inférieur (2.58–0.77 Ma).

Le **Chapitre 4** s'intéresse ensuite l'évolution des paysages périglaciaires de l'île Bylot en relation avec la glace de glacier enfouie. Ce chapitre tente de répondre à la question suivante : Quel est le rôle joué par ce type de glace dans l'initiation et l'évolution de systèmes aquatiques thermokarstiques dans la vallée Qarlikturvik. Ce chapitre avait d'abord pour objectif d'examiner l'origine de 21 lacs situés dans cette vallée, puis de développer un modèle conceptuel de la création et de l'évolution des lacs initiés par la fonte retardée de la glace de glacier enfouie dans le pergélisol. La présence de glace de glacier enfouie combinée aux connaissances des environnements lacustres accumulées depuis de nombreuses années par différentes équipes de recherche au même site d'étude ont permis d'avancer l'hypothèse suivante : les lacs de thermokarst ont des origines différentes et présentent des différences dans leurs conditions morphologiques et limnologiques en raison des différentes conditions de glace souterraine (glace enfouie vs glace intrasédimentaire). Pour investiguer l'origine des lacs de thermokarst, nous avons étudié une zone riche en lacs à l'aide de carottes de sédiments lacustres, de données bathymétriques détaillées, de données de télédétection et d'observations d'expositions de glace de glacier enfouie.

Finalement, cette thèse comporte également un article en annexe, qui a été publié par le professeur Denis Lacelle et dont je suis co-auteure. Cet article accompagne le deuxième chapitre présenté dans cette thèse. L'**Annexe I** combine l'étude de la composition isotopique ( $\delta^{18}\text{O}$ ) glace datant de la fin du Pléistocène ainsi que la relation théorique entre  $\delta^{18}\text{O}$  et l'altitude pour estimer l'altitude de la surface glaciaire d'origine lors du dernier maximum glaciaire (LGM). Cette étude contribue à améliorer notre compréhension de la configuration de l'inlandsis laurentidien à la fin du Pléistocène (Lacelle et al., 2018). Cet article utilise des données isotopiques de six sites, incluant les calottes glaciaires de Barnes et de Penny, de même que la glace de glacier enfouie



dans le pergélisol de l'île Bylot. Ainsi, en collaboration avec le professeur Lacelle, j'ai contribué à cet article par l'échantillonnage et l'analyse des données isotopiques pour le site de l'île Bylot.

En leur qualité de directeurs de recherche dans le cadre de ce doctorat, les professeurs Daniel Fortier (Université de Montréal) et Denis Lacelle (Université d'Ottawa) ont supervisé la conception et la planification du projet de recherche, ainsi que la rédaction des articles. J'ai effectué la collecte d'échantillons sur le terrain avec la participation des personnes suivantes : Michel Paquette, Paschale Noël-Bégin, Etienne Godin, Audrey Veillette, Jonathan Lasnier, Naïm Perreault et Sabine Veuille. J'ai préparé les échantillons et effectué l'ensemble des analyses en laboratoire. Les analyses des isotopes stables et cations ont été réalisées avec l'aide de Denis Lacelle (Université d'Ottawa). Les analyses et interprétations paléomagnétiques ont été réalisées en collaboration avec Guillaume St-Onge et avec l'aide de Quentin Duboc et Quentin Beauvais (Institut des sciences de la mer). Les profils géoradar des lacs étudiés dans le chapitre 4 ont été effectués par Vilmantas Preskienis (Institut national de la recherche scientifique) et Audrey Veillette (Université de Montréal), alors que l'interprétation des profils a été réalisée par Michel Paquette (Université de Montréal). Les carottes de sédiments présentées dans le troisième chapitre ont été échantillonnées et analysées par Frédéric Bouchard (Université de Sherbrooke), Audrey Veillette (Université de Montréal) et Vilmantas Preskienis (Institut national de la recherche scientifique). Les analyses et interprétation des diatomées présentés dans le chapitre 4 ont été réalisées en collaboration avec le professeur Reinhard Pienitz (Université Laval). Les analyses des profils de température et d'oxygène dissous exposés dans le chapitre 4 ont été effectuées en collaboration avec la professeure Isabelle Laurion (Institut national de la recherche scientifique) avec l'aide de Vilmantas Preskienis (Institut national de la recherche scientifique). Je suis première auteure et rédactrice de tous les chapitres contenus dans cette thèse.

## Articles scientifiques présentés dans la thèse :

**Chapitre 2:** *Origin, burial and preservation of late Pleistocene-age glacier ice in Arctic permafrost (Bylot Island, NU, Canada).*

Auteurs: Coulombe, S., Fortier, D., Lacelle, D., Kanevskiy, M., & Shur, Y.

*The Cryosphere*, 13(1), 97–111. doi:10.5194/tc-13-97-2019

L'article a été accepté pour publication (novembre 2018, *The Cryosphere*).

**Chapitre 3:** *Early Pleistocene Glaciation of the Eastern Canadian Arctic Archipelago: new evidence from Matuyama Age Glacier Ice Remnants, Bylot Island, Nunavut.*

Auteurs: Coulombe, S., Fortier, D., Lacelle, D., Guertin-Pasquier, A. et St-Onge, G.

L'article est en préparation et sera soumis à la revue *Geophysical Research Letters*

### Chapitre 4:

*Contrasted geomorphological and limnological properties of thermokarst lakes formed in buried glacier ice and ice-wedge polygon terrain.*

Auteurs: Coulombe, S., Fortier, D., Bouchard, F., Paquette, M., Charbonneau, S., Lacelle, D., Laurion, I. et Pienitz, R.

*The Cryosphere*, 16(7), 2837-2857. doi: 10.5194/tc-16-2837-2022

L'article a été accepté pour publication (juin 2022, *The Cryosphere*).

**Annexe I:** *Buried remnants of the Laurentide Ice Sheet and connections to its surface elevation*

Auteurs: Lacelle, D., Fisher, D. A., Coulombe, S., Fortier, D., Frappier, R.

*Scientific Reports*, 8(1), 13286. doi: 10.1038/s41598-018-31166-2

L'article a été accepté pour publication (août 2018, *Scientific Reports*)

.

## Chapitre 2

### Origin, burial and preservation of late Pleistocene-age glacier ice in Arctic permafrost (Bylot Island, NU, Canada)

Stephanie Coulombe<sup>1,2,3</sup>, Daniel Fortier<sup>2,3,5</sup>, Denis Lacelle<sup>4</sup>, Mikhail Kanevskiy<sup>5</sup>, and Yuri Shur<sup>5,6</sup>

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*Minor edits to the published text have been made here, following suggestions by this thesis' reviewers.*

## 1 **2.1 Abstract**

2 Over the past decades, observations of buried glacier ice exposed in coastal bluffs and  
3 headwalls of retrogressive thaw slumps of the Arctic have indicated that considerable amounts  
4 of late Pleistocene glacier ice survived the deglaciation and are still preserved in permafrost. Relict  
5 glacier ice and intrasedimental ice often coexist and look alike but their genesis is strikingly  
6 different. This paper presents a detailed description and infer the origin of a massive ice body  
7 preserved in the permafrost of Bylot Island (Nunavut). The massive ice exposure and core samples  
8 were described according to the cryostratigraphic approach, combining the analysis of  
9 permafrost cryofacies and cryostructures, ice crystallography, stable O-H isotopes and cation  
10 contents. The ice was clear to whitish in appearance with large crystals (cm) and small gas  
11 inclusions (mm) at crystal intersections, similar to observations of englacial ice facies commonly  
12 found on contemporary glaciers and ice sheets. However, the  $\delta^{18}\text{O}$  composition ( $-34.0 \pm 0.4 \text{ ‰}$ )  
13 of the massive ice was markedly lower than contemporary glacier ice and was consistent with the  
14 late Pleistocene age ice in the Barnes Ice Cap. This ice predates the aggradation of the surrounding  
15 permafrost and can be used as an archive to infer paleo-environmental conditions at the study  
16 site. As most of the glaciated Arctic landscapes are still strongly determined by their glacial legacy,  
17 the melting of these large ice bodies could lead to extensive slope failures and settlement of the  
18 ground surface, with a significant impact on permafrost geosystem landscape dynamics,  
19 terrestrial and aquatic ecosystems and infrastructure.

20

## 1 **2.2 Introduction**

2           In the Arctic, extensive areas of ridged and hummocky moraines are underlain by buried  
3 glacier ice (Dyke and Savelle, 2000; Kokelj et al., 2017a; Solomatin, 1986). Massive-ice bodies  
4 within these landscapes are important indicators of past glacial, hydrologic and hydrogeologic  
5 conditions and are used to reconstruct regional paleo-environments and paleo-climates (Lacelle  
6 et al., 2018; Jorgenson and Shur, 2008; Murton et al., 2005). Areas with buried glacier ice are also  
7 becoming increasingly vulnerable to climate warming (Kokelj et al., 2017a). Glacier ice is the most  
8 common type of buried ice in permafrost and its occurrence was reported in Russia (Astakhov,  
9 1986; Belova et al., 2008; Ingólfsson and Lokrantz, 2003; Kaplanskaya and Tarnogradskiy, 1986;  
10 Solomatin, 1986), the Canadian Arctic (Dallimore and Wolfe, 1988; French and Harry, 1990),  
11 Alaska (Jorgenson and Shur, 2008; Kanevskiy et al., 2013) and Antarctica (Sugden et al., 1995;  
12 Swanger, 2017). Buried glacier ice has been commonly observed in the proglacial zone of  
13 contemporary glaciers and can be preserved in formerly glaciated areas (the paraglacial to  
14 periglacial zones) within large moraine belts, hummocky till and glaciofluvial deposits (Everest  
15 and Bradwell, 2003; Tonkin et al., 2016). In the Canadian Arctic, buried glacier ice has been mainly  
16 reported in the western and central regions, especially in the Mackenzie Delta region (French and  
17 Harry, 1990), Tuktoyaktuk Coastlands (Murton et al., 2005), Herschel Island (Fritz et al., 2011),  
18 Yukon (Lacelle et al., 2007), Banks Island (Lakeman and England, 2012) and Victoria Island (Dyke  
19 and Savelle, 2000; Lorrain and Demeur, 1985). While the permafrost of the eastern Canadian  
20 Arctic is expected to contain remnants of Pleistocene ice sheets or glaciers, very few have been  
21 reported so far.

22  
23           Distinguishing between buried glacier ice and other types of massive ground ice in the  
24 permafrost is usually based on cryostratigraphy combined with detailed studies of physical,  
25 geochemical and isotopic properties of the ice that may also include analyses of occluded gases  
26 (Cardyn et al., 2007; Fritz et al., 2011; Ingólfsson and Lokrantz, 2003). When classifying glacier ice  
27 facies, a distinction is made between englacial and basal ice facies (Fortier et al., 2012; Lawson,  
28 1979). The englacial (firn-derived) ice facies is formed by the gradual snow compaction and  
29 recrystallization, a process called firnification, and has a low debris content (Benn and Evans,

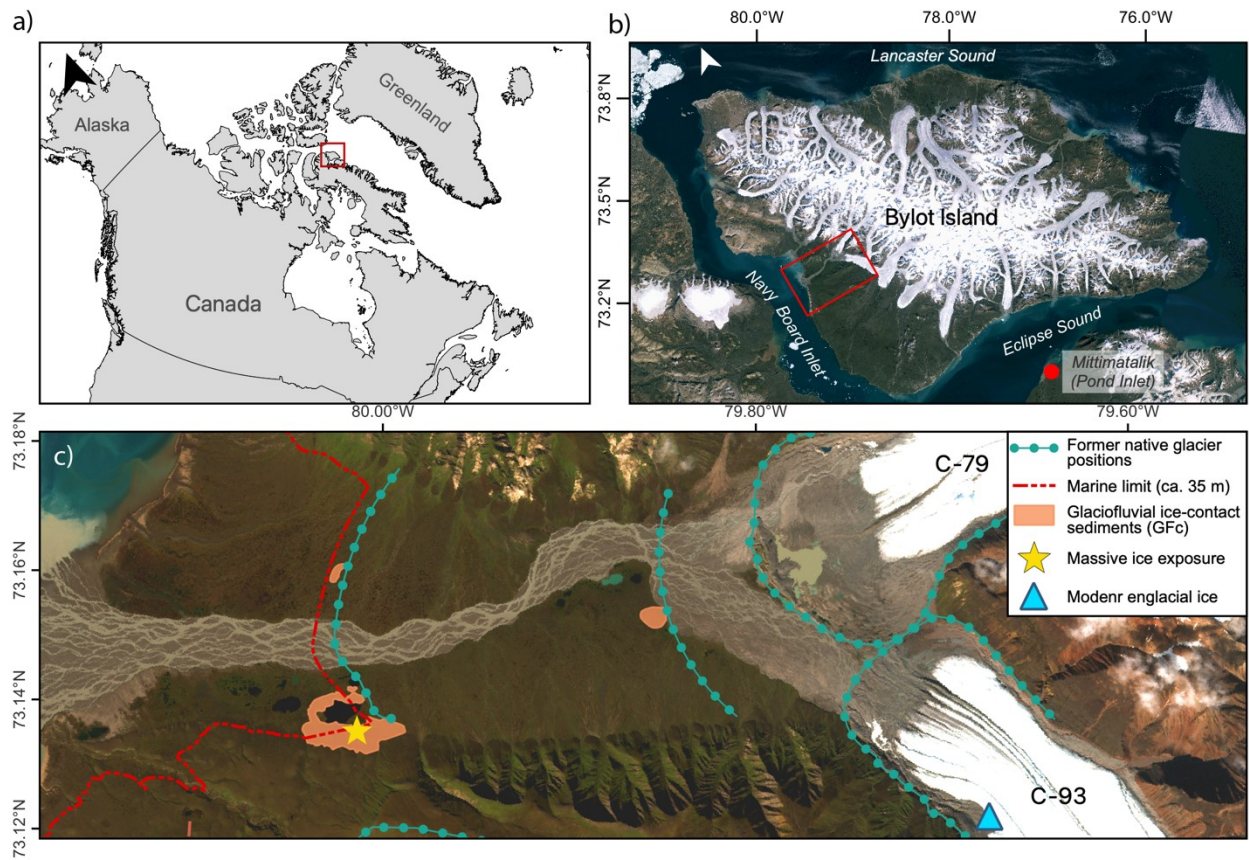
1 2010). Basal ice has distinctive physical and chemical characteristics and has a much higher debris  
2 content than the overlying englacial ice as a result of subglacial processes operating at the glacier  
3 bed: regelation, glaciohydraulic supercooling accretion, net basal adfreezing, incorporation of ice  
4 and sediments by overriding ice during glacier advance, glacio-tectonics, ice lens aggradation and  
5 downward propagation of cold temperature in the sediment at the glacier bed (congelation)  
6 (Alley et al., 1998; Fortier et al., 2012; Hubbard and Sharp, 1995; Knight, 1987; Lawson, 1979;  
7 Sharp et al., 1994). Both types of glacier ice may experience burial but basal ice is probably the  
8 most common form of buried glacier ice according to reports from various circumpolar  
9 permafrost regions (Belova et al., 2008; Fritz et al., 2011; Murton et al., 2005; St-Onge and  
10 McMartin, 1999). The process of burial of glacier ice has been described by Shur (1988), Solomatin  
11 (1986), Harris and Murton (2005) and citations therein. Burial of glacier ice may occur because of  
12 (1) accumulation of fluvial, lacustrine, aeolian or slope sediments on top of the ice; (2)  
13 glaciotectionic processes or (3) formation of insulating blanket of supraglacial melt-out till. With  
14 the latter scenario, sediment-rich basal ice has a greater potential to persist in a buried state than  
15 englacial ice with little debris. Buried glacier ice remains stable for a long period of time only if  
16 the soil temperature is below freezing, and the active-layer thickness does not exceed a depth to  
17 the massive ice body (Shur, 1988).

18  
19 In this study, we describe the occurrence of massive ice preserved in the permafrost of  
20 the Qarlikturvik Valley, south-western Bylot Island (NU, Arctic Canada). We investigated the  
21 physical and geochemical properties of a recently exposed body of massive ice and compared  
22 them with those of other ice types in the region (snow, glacier ice, wedge ice, segregated ice) to  
23 infer its origin. A cryostratigraphic approach was used to delineate cryostratigraphic units on the  
24 basis of their cryostructures, physical properties and thaw unconformities (French and Shur,  
25 2010; Gilbert et al., 2016), combined with crystallography and geochemical analyses of the  
26 different ground-ice types. The origin of the massive ice is discussed.

27

## 1 **2.3 Regional settings**

2           With ice-capped summits dominating the central highlands and glaciated valleys that  
3 extend near the coast, the mountainous central section forms a striking contrast with the  
4 relatively flat coastal lowlands. The Byam Martin Mountains (~ 1400 m a.s.l.) consist primarily of  
5 Archean–Aphebian crystalline igneous rocks and Proterozoic metasedimentary and metamorphic  
6 rocks (Jackson and Davidson, 1975). Klassen (1993) suggested that alpine glaciers, larger but  
7 similar in size to those present today, occupied Bylot Island during the late Wisconsinan. At the  
8 Last Glacial Maximum (LGM), ice streams of the Laurentide Ice Sheet (LIS) flowed in adjacent  
9 marine channels (Lancaster Sound, Navy Board Inlet and Eclipse Sound) and reached Bylot Island  
10 (De Angelis and Kleman, 2007). The study area is situated in the Qarlikturvik Valley (73° 09' N, 79°  
11 57' W, 29 m a.s.l.) on south-west Bylot Island at about 80 km north-west of Mittimatalik (Pond  
12 Inlet; Figure 10). This valley was eroded through a Cretaceous–Tertiary sequence of poorly  
13 consolidated sandstone and shale (Jackson and Davidson, 1975).



1  
 2 **Figure 10.** (a) Location of Bylot Island, (b) location of the Qarlikturvik Valley, (c) location of the  
 3 study area within the Qarlikturvik Valley (valley of glaciers C-79 and C-93). Wedge ice, segregated  
 4 ice and snow were also sampled near (< 1 km) the massive ice exposure. The thaw depths were  
 5 measured with a steel probe at every 10 m along a 150 m transect. The orange area represents  
 6 mounds of ice contact sediments, indicating past frontal glacier positions. The dashed red line  
 7 indicates the marine limit (ca. 35 m).  
 8

9 The Qarlikturvik Valley is a typical U-shaped glacial valley with surface sediments reflecting  
 10 the complex history of the valley: presence of unconsolidated glacial, colluvial, alluvial, marine,  
 11 aeolian and organic deposits dating back to the Late Pleistocene and Holocene (Allard, 1996;  
 12 Fortier and Allard, 2004). The valley comprises low-lying ice-wedge polygon terraces bordering a  
 13 proglacial braided river running in a glaciofluvial outwash plain and forming a delta in Navy Board  
 14 Inlet. During the Holocene glaciers C-93 and C-79 retreated up-valley and today they are located  
 15 about 14 km from the coast. Following glacial retreat, the valley became partially submerged  
 16 between 11 335 to 6020 cal yr BP as a result of a marine transgression (marine limit: ca. 35 m;



1 (Allard, 1996). Alternating layers of peat and aeolian sands and silts (~ 2–3 m) cover the  
2 glaciofluvial terraces of the valley, where an extensive network of syngenetic ice-wedge polygons  
3 has developed after 6000 cal yr BP (Fortier and Allard, 2004). Mounds of reworked till and ice-  
4 contact-stratified sediments mark a former position of the glacier in the valley. The glacier  
5 probably advanced into shallow marine water (Allard, 1996), and therefore the massive ice  
6 exposure was not submerged and covered by marine sediments.

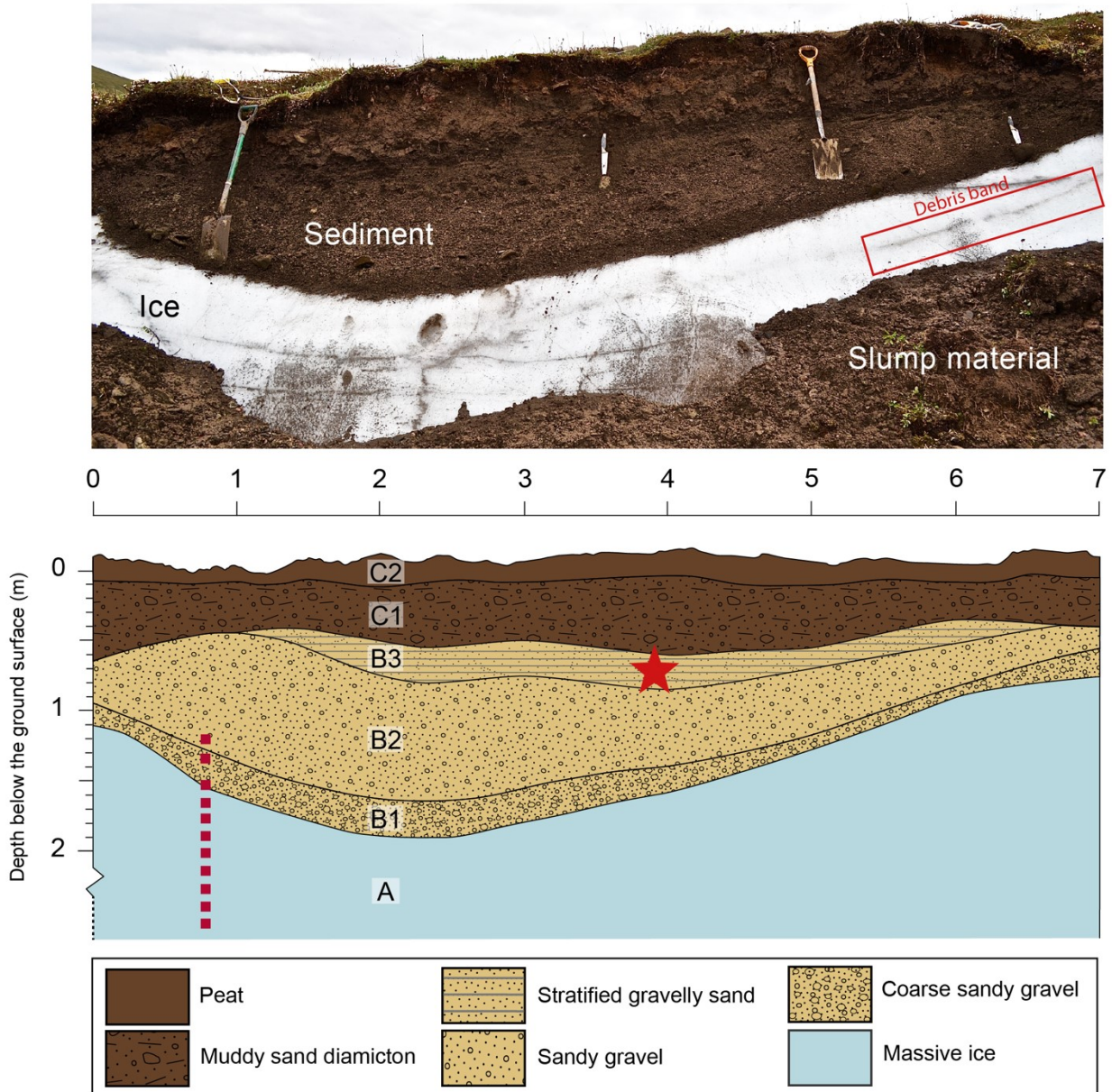
7  
8 Bylot Island belongs to the Arctic Cordillera and the northern Arctic terrestrial ecozones.  
9 The mean annual air temperature between 1971 and 2000 in Pond Inlet (NU, Canada) was  $-15.1$   
10  $\pm 5.1$  °C, increasing slightly to  $-14.6 \pm 4.9$  °C between the 1981 and 2010 (Environment Canada,  
11 2015). No significant trends in precipitation have been observed over the last decades, with a  
12 mean annual precipitation of 189 mm, much of it falling as snow. Thawing and freezing indices  
13 averaged (1981–2010) 473 degree-days above 0°C and 5736.1 degree-days below 0°C respectively  
14 (Environment Canada, 2015). Vegetation in the valley is typical of Arctic tundra environments and  
15 is largely determined by soil moisture and slope (Duclos et al., 2006). Wetlands occur in low-lying  
16 areas, commonly with low-centred polygons, and are typically dominated by grasses and sedges.  
17 Mesic tundra is found in areas characterized by better-drained soils (i.e. uplands and hillslopes).  
18 The climatic and vegetation conditions determine the presence of continuous permafrost in the  
19 ice-free areas of Bylot Island. Locally, permafrost thickness > 400 m has been detected from  
20 thermal measurements on the nearby Somerset and Devon islands (Smith and Burgess, 2002).  
21 Active-layer thickness varies from ~ 1 m in drained unvegetated sands and gravels to ~ 0.3–0.7  
22 m in peaty and silty soils (Allard et al., 2016).

## 24 **2.4 Material and methods**

25 A large body of massive ice was found exposed beneath ~ 1.7 m of sediments along the  
26 headwall of a thaw slump (29 m a.s.l.) in the Qarlikturvik Valley, ~ 10 km from the terminus of C-  
27 93 and C-79 glaciers (Figure 10c). Since a large amount of slump material covered the ice, the  
28 exposure was cleaned and enlarged to allow a better description and sampling. The excavated  
29 section along the headwall of the slump attained a height exceeding 7 m and extended laterally

1 over 10 m. The lower and lateral contacts between the ice body and the surrounding sediments  
2 could not be delineated. The exposure was subdivided into three units from the bottom: (a)  
3 massive ice, (b) sand and gravel and (c) muddy sand diamicton overlaid by peat (Figure 11). The  
4 thaw depths were measured with a steel probe at every 10 m along a 150 m transect that started  
5 at the upper headwall of the thaw slumps. Samples from the massive ice (unit A) were collected  
6 using an axe and a portable core drill equipped with an 8 cm diamond carbide core barrel.  
7 Between 2011 and 2013, ice cores were extracted every 10 cm or less from depths ranging from  
8 1 to 3 m below the surface (Figure 11) without reaching the bottom of the ice body. For  
9 comparison, modern glacier ice (englacial ice) was sampled from freshly collapsed ice blocks at  
10 the margin of glacier C-93 located a few kilometres up the valley (Figure 10c). Wedge ice,  
11 segregated ice and snow were also sampled for geochemical analysis at the site located near (< 1  
12 km) the massive ice exposure. Samples (n = 5) from the sediments (units B and C) overlying the  
13 massive ice were collected and characterized. All samples were kept frozen, shipped and stored  
14 in our laboratory for further analysis.

15



1  
 2 **Figure 11.** A photograph and a schematic cross section showing generalized stratigraphy  
 3 of the massive ice exposure and the overlying sediments. The lower and lateral contacts of the  
 4 massive ice have not been reached. The thaw depth measured in late July (2013) at the headwall  
 5 reaches 55 cm. The red star indicates the sampling location of the organic material and the red  
 6 dots show the sampling points for stable O-H isotope and hydrochemistry.

7  
 8 A cryostratigraphic approach was used to describe the massive-ice body and the overlying  
 9 sediments (Fortier et al., 2012; French and Shur, 2010; Gilbert et al., 2016; Murton and French,  
 10 1994). Cryostratigraphic units were delineated based on cryostructures and cryofacies. Cryofacies

1 are bodies of frozen sediment that are visually distinct from adjacent frozen sediments based on  
2 their cryostructures, volumetric ice content, ice-crystal size and sediment texture. Cryostructures  
3 refer to the shape, amount and distribution of ice within the frozen sediment. Gas inclusions  
4 visible within the ice and the deformation structures in the ice and sediments were also described  
5 (Hambrey and Lawson, 2000). To further investigate the cryostratigraphic characteristics of the  
6 massive ice body, all samples were observed under X-ray-computed tomography (CT) scanning  
7 (Siemens SOMATOM Sensation 64), a technique that allows visualizing and reconstructing the  
8 internal structure (2-D and 3-D) of permafrost samples at < 1 mm resolution (Calmels et al., 2010;  
9 Dillon et al., 2008; Fortier et al., 2012). The complete CT-scanning procedure used in this study is  
10 presented in Appendix A. Crystallographic analysis of the massive ice and modern glacier ice was  
11 conducted to measure their crystal size and shape (surface area, long axis and circularity ratio) as  
12 these parameters contain information about the conditions under which the ice was formed  
13 (French and Shur, 2010). The crystalline structure was investigated through thin sections of ice  
14 mounted on a glass slide under cross-polarized light. Thin sections were made by cutting the ice  
15 sample vertically and/or horizontally into thin slices (thickness: 0.2 to 0.4mm). C-axis orientations  
16 of the crystals have not been measured since the horizontal orientation of the ice samples was  
17 not preserved following the sampling. Fiji image analysis software was used to measure the area,  
18 long axis and circularity ratio of each crystal (Schindelin et al., 2012). Differences in the crystal  
19 shape (area, long axis, circularity ratio) of horizontal and vertical thin sections were tested with  
20 the Mann–Whitney U test using R, which is a programming language software for data analysis  
21 and graphics (R Development Core Team, 2016).

22  
23 The massive ice body, along with glacier C-93 ice, ice wedges, intrasedimental ice and  
24 snow, were analyzed for their geochemical and isotopic ( $\delta^{18}\text{O}$ ,  $\delta\text{D}$ ) composition, an approach that  
25 can shed light into the origin of ground ice (Lacelle and Vasil'chuk, 2013). Sampling of the massive  
26 ice body was done at 10 cm vertical intervals or less, depending on the unit change; prior to  
27 sampling, at least 10 cm of the ice surface was removed with an ice axe. In 2013, all samples  
28 ( $n = 80$ ) were melted, filtered (0.45  $\mu\text{m}$  diameter filter) and acidified in the laboratory in sealed  
29 polyethylene bottles prior to analyses. The concentration of major cations in the ice and snow

1 (Al<sub>tot</sub>, Ca<sup>2+</sup>, Fe<sub>tot</sub>, K<sup>+</sup>, Mg<sup>2+</sup>, Mn<sub>tot</sub>, Na<sup>+</sup>) was determined by inductively coupled plasma optical  
2 emission spectrometry (Vista Pro ICP-OES) at the University of Ottawa. Solutes are expressed in  
3 milligrams per litre and analytical reproducibility was ±1 %. The stable isotope ratios of oxygen (<sup>18</sup>  
4 O/<sup>16</sup> O) and hydrogen (D/H) were determined using a Los Gatos Research high-precision liquid  
5 water analyzer coupled to a CTC LC-PAL autosampler. The results are presented using the δ-  
6 notation (δ<sup>18</sup>O and δD), where δ represents the parts per thousand differences for <sup>18</sup>O/<sup>16</sup>O or D/H  
7 in a sample with respect to Vienna Standard Mean Ocean Water (VSMOW). Analytical  
8 reproducibility for δ<sup>18</sup>O and δD was ± 0.3 ‰ and ± 1 ‰, respectively.

9  
10 The origin of the sediment overlying the massive ice was inferred from the sedimentary  
11 facies, particle-size distribution, the clasts and sand-size quartz grain morphoscopy and the  
12 geochemical and isotopic (δD-δ<sup>18</sup>O) composition of pore water. Particle-size distributions were  
13 determined by dry sieving at 1/2φ intervals (size ranges -12 to 4φ). The hydrometer method was  
14 used to determine the distribution of the finer particles smaller than 4φ (ASTM Standard D422,  
15 2007). Descriptive statistics (mean grain size, sorting, skewness) and Folk and Ward sediment  
16 classes were determined using the RYSGRAN package for R (Gilbert et al., 2014; R Development  
17 Core Team, 2021). Fifty in situ pebble- to cobble-sized clasts were randomly collected from the  
18 uppermost unit (Unit C). These clasts were analyzed for shape, roundness and lithology using  
19 techniques described by Benn (2004). Morphoscopic analyses of small quartz grains (0.5–1.0 mm)  
20 were conducted using a binocular microscope (Cailleux and Tricart, 1963). Additionally, the  
21 ground ice in the sediments was analyzed for soluble ions (major cations) and δD-δ<sup>18</sup>O  
22 composition following the method described above. A fragment of poorly decomposed peat  
23 sampled in unit B was radiocarbon dated by accelerator mass spectroscopy (AMS) (ULA-6505,  
24 Centre for Northern Studies, Université Laval). Radiocarbon age was calculated as -8033ln (F14  
25 C) and reported in <sup>14</sup>C yr BP (BP=AD1950) (Stuiver and Polach, 1977) and then corrected to  
26 calendar years (cal yr BP) using Calib 7.10 and the IntCal13 calibration curve (Reimer et al., 2013;  
27 Stuiver et al., 2021).

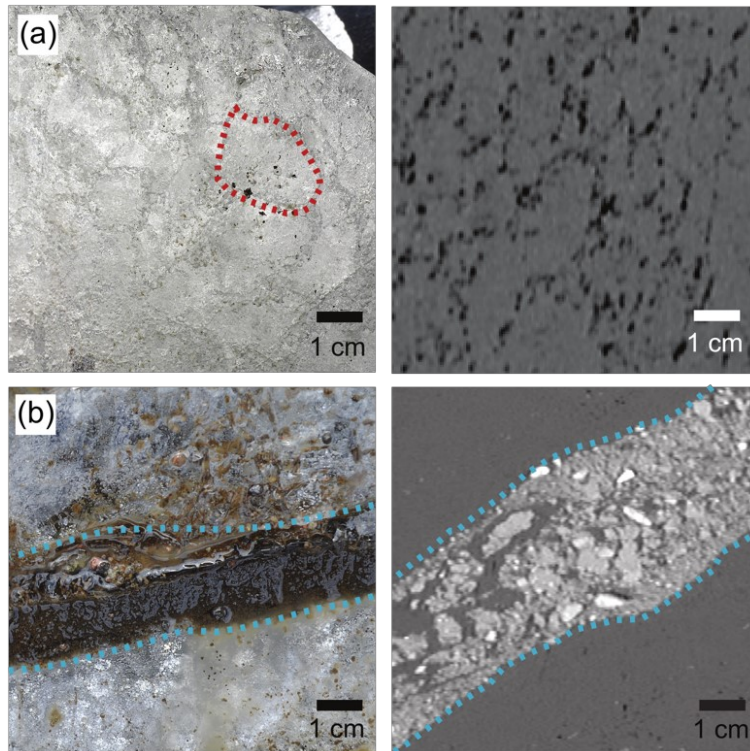
28

## 1 **2.5 Results**

### 2 ***2.5.1 Cryostratigraphy and properties of the massive ice body***

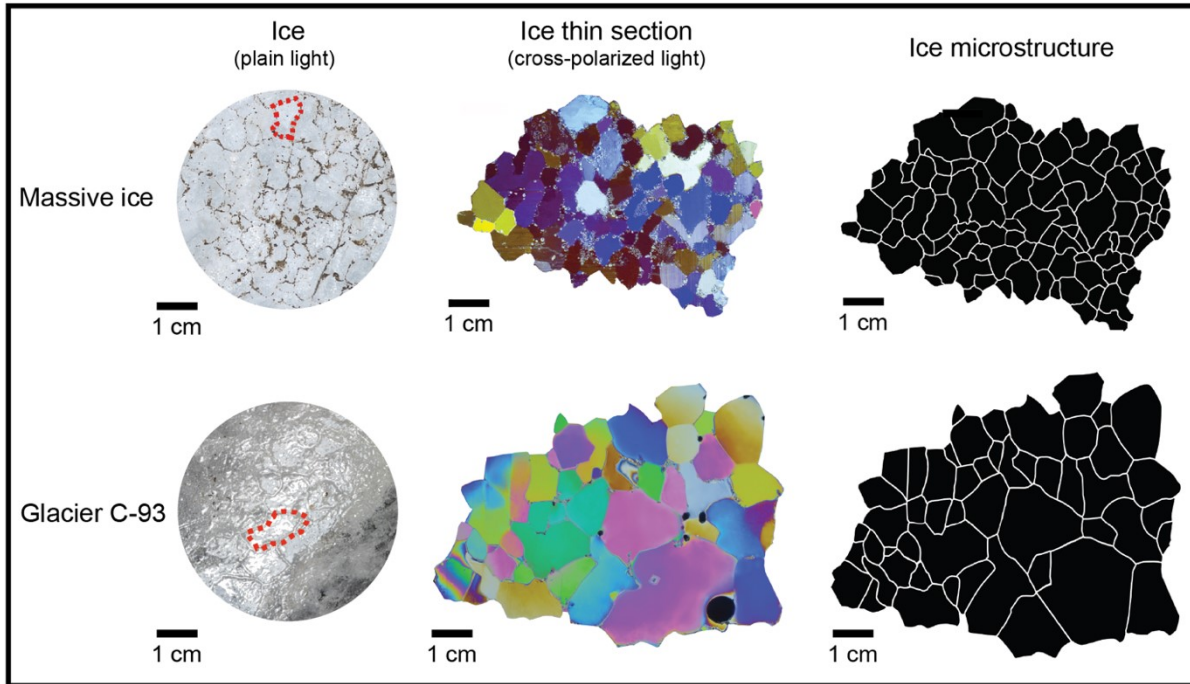
3           The exposed massive ice body (unit A) was > 10 m thick and had a clear to milky white  
4 appearance due to its high bubble content (Figure 12a). With a volumetric ice content near 100  
5 %, it refers to the “*pure ice*” cryofacies described by Murton and French (1994). Occasional thin  
6 bands of sediments (sands and gravels) with suspended and crustal cryostructures were cutting  
7 across the ice (Figures 11 and 12b). These discrete and planar bands were < 2 cm thick, subparallel  
8 to one another and showed a dip direction (21 to 31°) down the ice body in the south-eastern  
9 direction. Crystallographic analysis of thin sections of the massive ice under cross-polarized light  
10 showed that the crystals had mostly bluish colours, suggesting that the ice crystals had similar  
11 orientations (Figures 13 and 14a). Coarse-grained ice crystals characterized the massive ice body:  
12 long-axis average of  $7.97 \pm 2.97$  mm (3.13–16.58 mm); average surface area of  $34.9 \pm 25$  mm<sup>2</sup> (5.8  
13 and 153.5 mm<sup>2</sup>); and average circularity ratio of  $0.65 \pm 0.09$ , indicating the crystals were relatively  
14 rounded (Figure 15). No significant differences in the shape properties (surface area, long axis,  
15 circularity ratio) were observed between the horizontal and vertical thin sections (Mann–  
16 Whitney–Wilcoxon test,  $p > 0.05$ ), indicating that the ice crystals were nearly equiaxial. The glacier  
17 C-93 ice displayed ice crystals with varying colours (Figure 13). The ice crystals were larger  
18 (average surface area of  $125.29 \pm 148$  mm<sup>2</sup>) than those of the massive ice body (Figures 13 and  
19 15). Glacier C-93 ice consists of relatively rounded crystals as their mean circularity ratio averaged  
20  $0.65 \pm 0.01$ .

21



1  
 2 **Figure 12.** Photographs (left) and CT scans (right) of the massive ice body. The CT scan images are  
 3 not exactly the same position as the photographs. Unconsolidated sediments and rock (high-  
 4 density minerals) appear light grey or white, respectively. Gases inclusions and water appear black,  
 5 and other materials, such as ice, can have various shades of dark grey (a) Pure-ice facies. The  
 6 dotted red line highlights one single crystal; (b) ice-poor sediment (sands and gravels) with  
 7 suspended and crustal cryostructures forming a band in the massive ice (white fragments are rock  
 8 clasts). It represents a close-up of the occasional thin bands of sediments cutting across the ice.  
 9



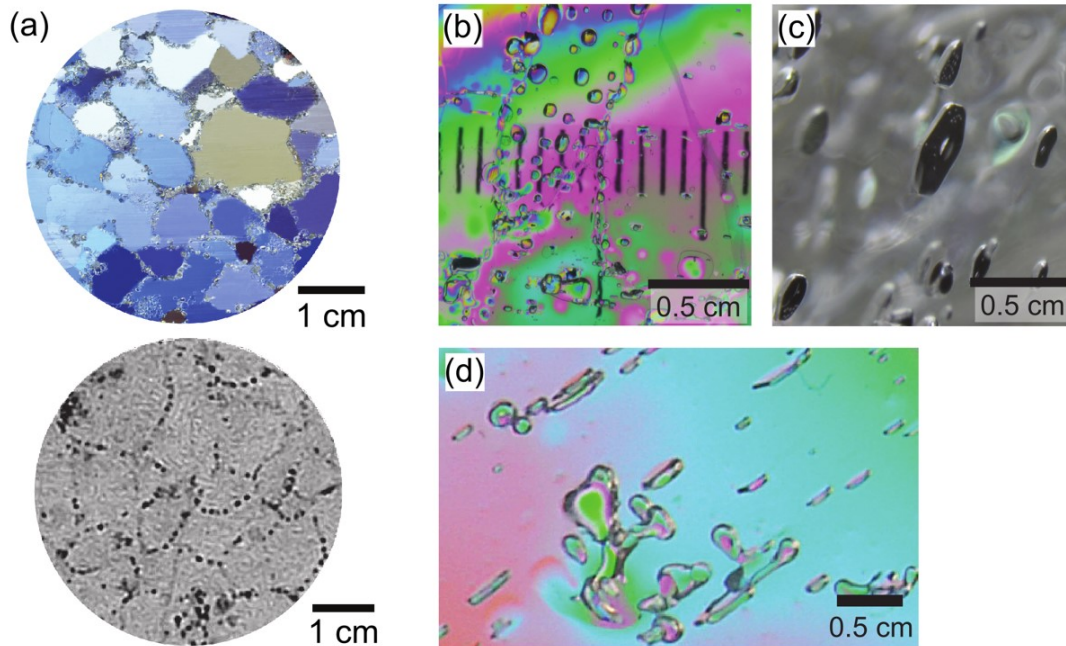


1  
 2 **Figure 13.** Comparison between the massive ice body and modern glacier ice (glacier C-93, Bylot  
 3 Island). The first column shows unprocessed photographs of the ice taken under a plain light with  
 4 surficial sediment inclusions highlighting the crystal boundaries. The dotted red line highlights one  
 5 crystal. The second column shows thin sections of ice sample viewed under direct cross-polarized  
 6 light. The third column shows the microstructure (crystal boundaries) extracted from the thin-  
 7 section photograph.

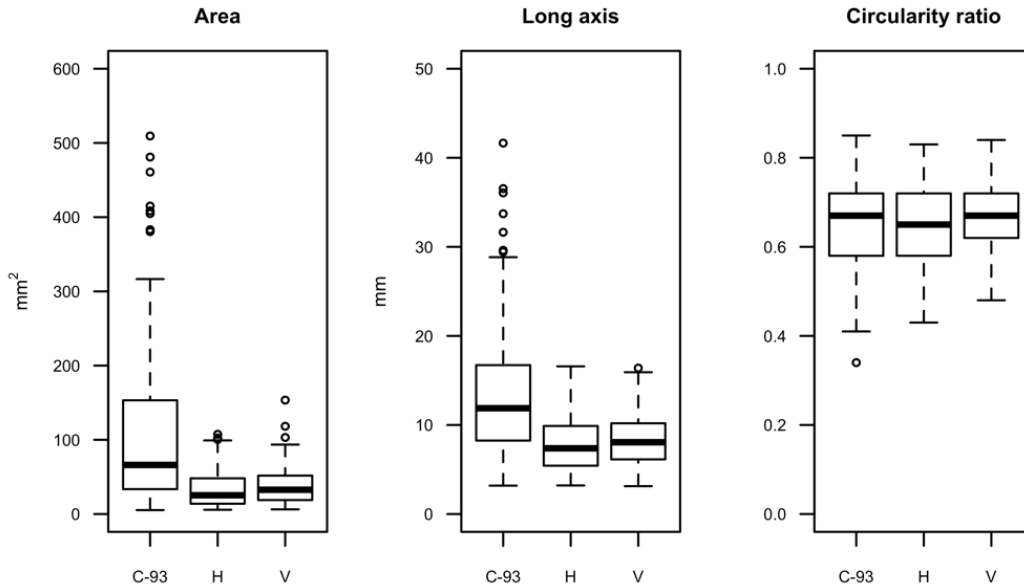
8  
 9 The total volumetric content of gas inclusions varied from 2 % to 10 % for both the massive  
 10 ice and glacier C-93 ice, with bubbles being mostly located along grain boundaries (Figure 14a).  
 11 The massive ice body contained three types of gas inclusions: (A) spherical bubbles, (B) flattened  
 12 disks and (C) clusters of deformed and coalescent bubbles (Figure 14b, c, d). The tiny gas bubbles  
 13 had a mean surface area of  $0.85 \pm 1.04 \text{ mm}^2$  with circularity ratio averaging  $0.89 \pm 0.17$ . Gas  
 14 bubbles observed in glacier C-93 were mostly spherical and small, with a few clusters of deformed  
 15 coalescent bubbles. The gas bubbles had a mean surface area of  $0.13 \pm 0.41 \text{ mm}^2$  and an average  
 16 circularity ratio of  $0.89 \pm 0.18$ .

17





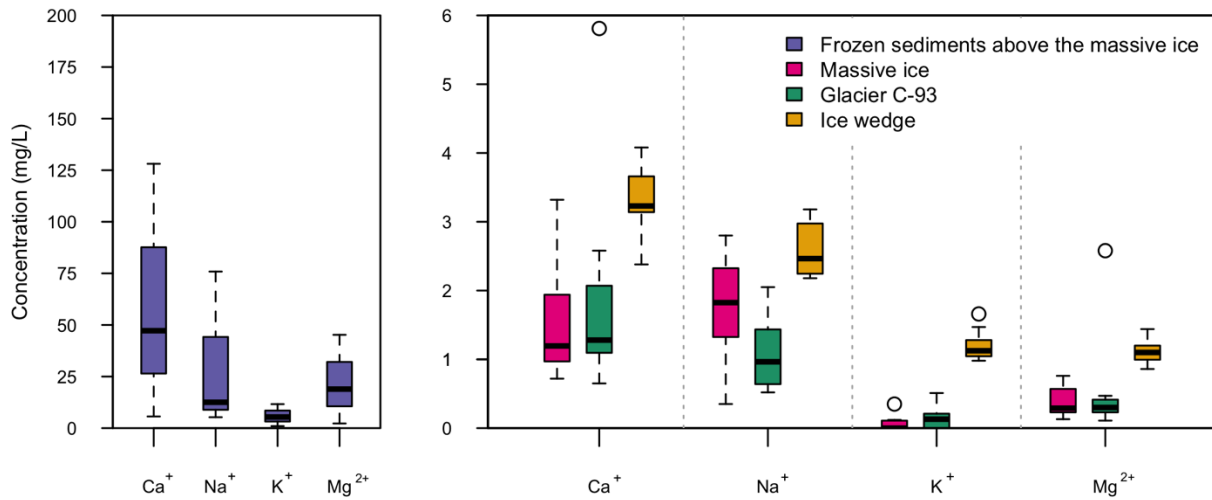
1  
 2 **Figure 14.** (a) A thin section of the massive ice viewed under cross-polarized light and a transverse  
 3 cross section from a scan showing the gas inclusions within the ice. (air is black; ice is dark grey).  
 4 Photos to the right show patterns of gas inclusions; (b) small (sub-millimetre to millimetre)  
 5 spherical bubbles (vertical bars are from measuring ruler of microscope stage); (c) small disks up to  
 6 6 mm in diameter; (d) coalescent bubbles and small disks all flattened in the same direction.  
 7



1  
 2 **Figure 15.** Box plots comparing the distribution of ice crystal characteristics (area, long axis,  
 3 circularity ratio) of horizontal thin sections (H) and vertical thin sections (V) obtained from massive  
 4 ice samples. C-93 represents data obtained from a sample of modern glacier ice sampled from  
 5 glacier C-93.  
 6

7 Major cations in the massive ice body (i.e.  $\text{Ca}^{2+}$ ,  $\text{Na}^+$   $\text{Mg}^{2+}$  and  $\text{K}^+$ ) all occurred in low  
 8 concentrations (mean  $< 1.76 \text{ mg L}^{-1}$ ; Figure 16). The cation concentrations of glacier C-93 ice were  
 9 very similar to those of the massive ice, whereas the ice wedge had slightly higher cation  
 10 concentrations, with average concentrations ranging from  $1.11 \pm 0.18 \text{ mg L}^{-1}$  for  $\text{Mg}^{2+}$ , to  
 11  $3.32 \pm 0.44 \text{ mg L}^{-1}$  for  $\text{Ca}^{2+}$ . The  $\delta\text{D}-\delta^{18}\text{O}$  composition of the massive ice, along with those from  
 12 glacier C-93 and other types of ground ice present in the valley are shown in Figure 17b. The  $\delta^{18}\text{O}$   
 13 composition of the massive ice had a narrow range (average  $\delta^{18}\text{O}$ :  $-34.0 \pm 0.4 \text{ ‰}$ ) with D-excess  
 14 ( $d = \delta\text{D} - 8\delta^{18}\text{O}$ ) averaging  $6.6 \pm 2.5 \text{ ‰}$ . The  $\delta^{18}\text{O}$  composition of the massive ice was much lower  
 15 than that of the snow (average  $\delta^{18}\text{O}$ :  $-30.4 \pm 1.8 \text{ ‰}$ ), ice-wedge ice (average  $\delta^{18}\text{O}$ :  $-25.6 \pm 0.95$   
 16  $\text{‰}$ ) and glacier C-93 ice (average  $\delta^{18}\text{O}$ :  $-25.0 \pm 3.1 \text{ ‰}$ ). The D-excess of snow, ice wedge, glacier  
 17 C-93 and ground ice (unit B) samples averages  $8.9 \pm 3.4 \text{ ‰}$ ;  $9.3 \pm 7.1 \text{ ‰}$ ,  $5.2 \pm 5.8 \text{ ‰}$  and  $-52.4 \pm$   
 18  $31.4 \text{ ‰}$  respectively. In a  $\delta\text{D}-\delta^{18}\text{O}$  diagram, the samples from the massive ice, glacier C-93, snow  
 19 and ice wedges were distributed along linear regression slope values of 6.0 ( $R^2 = 0.44$ ), 7.1 ( $R^2 =$   
 20  $0.95$ ), 6.6 ( $R^2 = 0.96$ ) and 5.2 ( $R^2 = 0.36$ ) respectively.

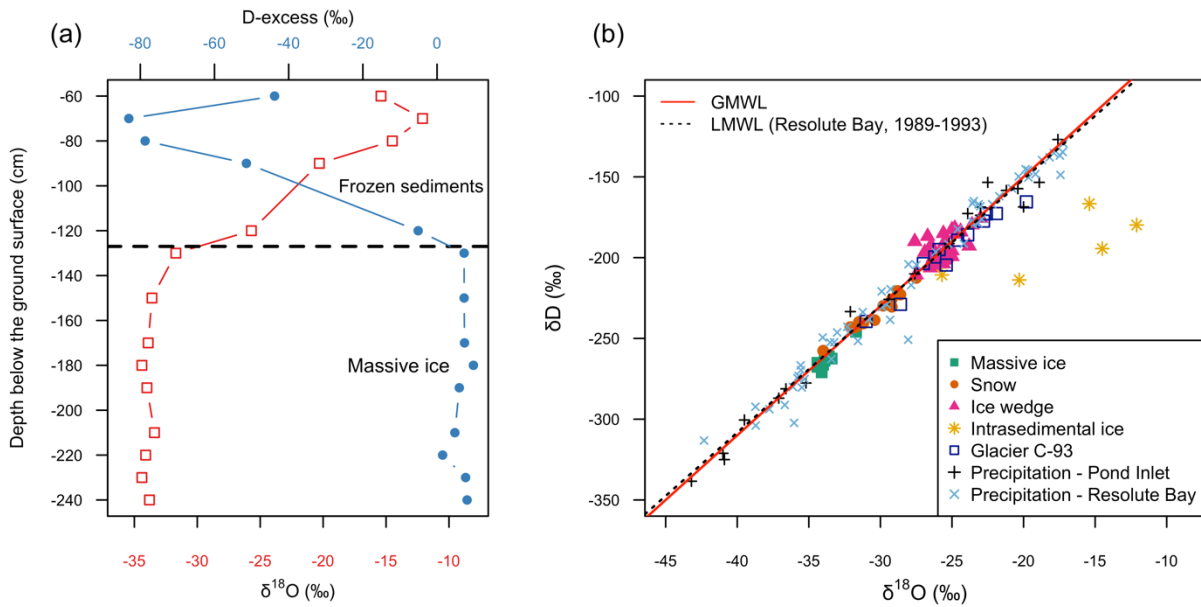
1



2

3 **Figure 16.** Box plots showing soluble cation concentration of the massive ice, ice wedge, glacier  
 4 C-93 and intrasedimental ice sampled within the sediment layers covering the massive ice unit.

5



6

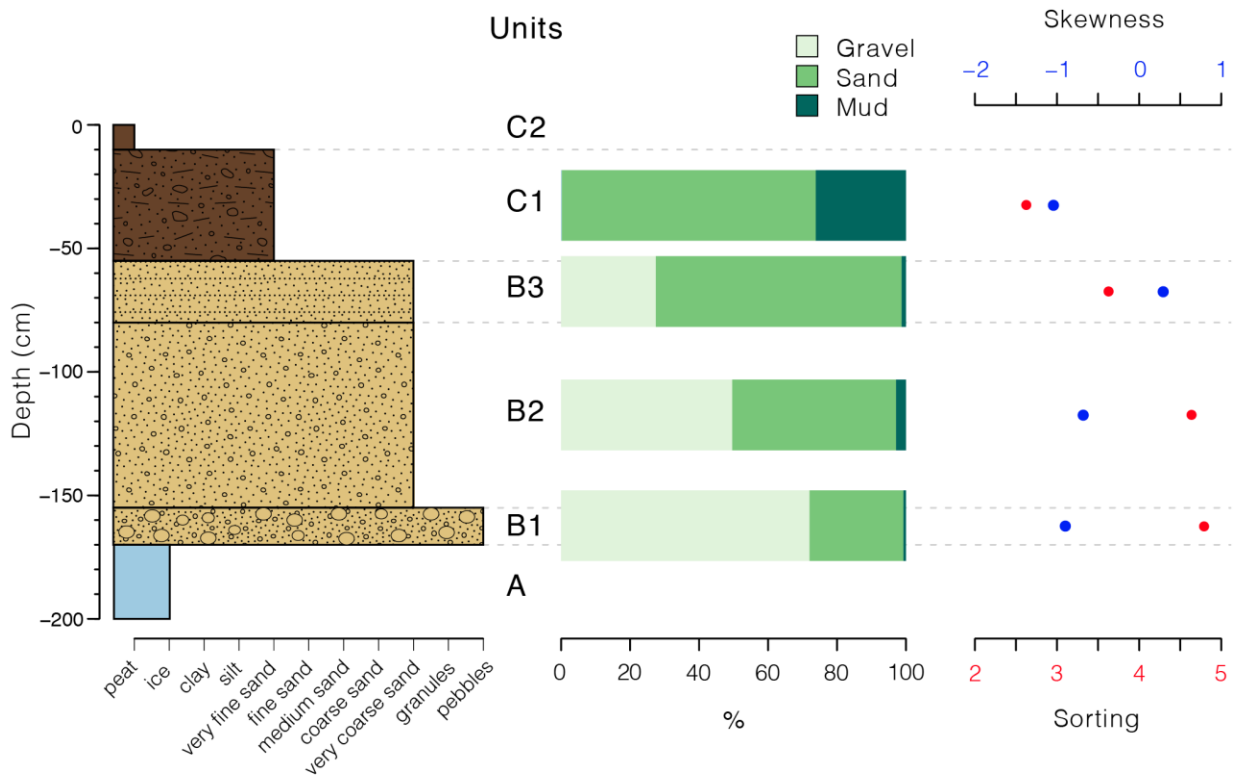
7 **Figure 17.** (a)  $\delta^{18}\text{O}$  and D-excess depth profiles including both the massive ice unit and the  
 8 intrasedimental ice from the sediment cover; (b)  $\delta^{18}\text{O}$ - $\delta\text{D}$  diagram of the massive ice and other  
 9 types of ground ice (ice wedge, intrasedimental ice), snow and modern glacier ice (C-93) sampled  
 10 on Bylot Island. The solid red line represents the Global Meteoric Water Line (GMWL):  
 11  $d = \delta\text{D} - 8 \delta^{18}\text{O} + 10$  (Dansgaard, 1964). Also shown are the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values of precipitation  
 12 recorded at Pond Inlet (1990–1992) and Resolute Bay (1989–1993; IAEA/WMO, 2016). LMWL is the

1 local meteoric water line for Resolute Bay, Nunavut ( $\delta D = 7.8 \delta^{18}O + 5.33$ ; IAEA/WMO, 2016) and  
2 Pond Inlet, Nunavut ( $\delta D = 8.0 \delta^{18}O + 10.55$ ; IAEA/WMO, 2016).  
3

#### 4 **2.5.2 Cryostratigraphy and properties of the overlying sediments**

5 The massive ice body was covered by  $\sim 1.7$  m of sediments. The thaw depths measured  
6 in late July (2013) ranges between 19 and 55 cm (mean:  $30 \pm 9$  cm). A sharp, subplanar and  
7 unconformable contact separated the ice from the overlying sediments (units B and C) along the  
8 exposed section (Figure 11). Unit B ( $\sim 115$  cm thick) directly overlaid the massive ice body and  
9 has ice-poor sediment cryofacies with a structureless cryostructure, essentially made of pore ice.  
10 Unit B is subdivided into three subunits: B<sub>1</sub> is coarse sandy gravel, B<sub>2</sub> is sandy gravel, and B<sub>3</sub> is  
11 stratified gravelly sand. All subunits were texturally similar, consisting mostly of sands and gravels,  
12 with silt- and clay-sized particles constituting  $< 3$  % of the sediment (Figure 18). The grain-size  
13 fraction  $> 32$  mm was not included in profile distribution, but abundant cobbles were observed  
14 within the bottom subunits. Sediments were poorly sorted to very poorly sorted (sorting values  
15 ranging between 3 and 5) and were coarser at the base and finer near the top (Figure 18). Subunits  
16 B<sub>1</sub> and B<sub>2</sub> were both characterized by unstratified sandy gravel that became finer towards the  
17 uppermost subunit B<sub>3</sub>, which consisted of stratified gravelly sand with thin subhorizontal laminae  
18 ( $< 1$  cm). Morphoscopic analysis of subunit B<sub>3</sub> showed that quartz grains were mostly angular  
19 (70%) to subangular (smooth and polished, glazing grains: 25%). Rounded and frosted grains  
20 represented  $< 5$  % of the total sand fraction. Plants fragments located at the base of subunit B<sub>3</sub>  
21 were dated to  $885 \pm 15$  <sup>14</sup>C yr BP (1164 cal yr BP;  $1\sigma$  range: 1058–1204). The sand and gravel  
22 sequence of unit B was sharply overlain by  $\sim 55$  cm of an unstratified (massive) thawed diamicton,  
23 with abundant pebble- to cobble-sized clasts (unit C; Figures 11 and 18). The matrix (sandy silt)  
24 was very poorly sorted and the grain-size distribution tended to be fine skewed. Clasts had a wide  
25 range of shapes, with predominance in the subangular and subrounded classes (76 %). Only 24 %  
26 of all clasts were rounded or angular. Gneiss was the dominant cobble type with an average of 60  
27 % in the matrix, whereas sedimentary and igneous rocks averaged 36 % and 4 %. A thin,  
28 continuous and irregular layer of dark fibrous peat with roots (subunit C2) overlaid the diamicton.  
29

1 Supernatant water samples (n = 5) from the ice-poor sediments in unit B showed markedly  
 2 higher  $\delta^{18}\text{O}$  and cation values compared to the underlying massive-ice body (Figure 16 and 17a).  
 3 In unit B, the  $\delta^{18}\text{O}$  values became progressively higher, from  $-25.7\text{‰}$  to  $-15.4\text{‰}$ , as we moved  
 4 upward from the ice-sediment contact towards the surface (Figure 17a). The cations content in  
 5 the ice-poor sediment was 1 to 2 orders of magnitude higher than in the massive ice and also  
 6 shifted to higher concentrations towards the surface (Figure 16).



7  
 8 **Figure 18.** Sedimentological data from the stratigraphic section. From left to right: a stratigraphic  
 9 log showing the mean grain size of each unit; gravel, sand and mud percentages; and skewness and  
 10 sorting.  
 11

## 1 **2.6 Discussion**

### 2 ***2.6.1 Origin of the massive ice***

3           The cryostratigraphic and crystallographic properties of the massive ice along with its  
4 isotopic and geochemical composition indicate that the exposed ice consists of relict Pleistocene  
5 englacial ice buried and preserved in the permafrost of Bylot Island. Table 4 summarizes the  
6 cryostratigraphic and crystallographic properties of different types of tabular massive ice  
7 described in the literature: buried glacier ice (basal and englacial), massive segregated-intrusive  
8 ice and buried snowbanks. The appearance and structure of the buried massive ice body are  
9 similar to those of englacial ice typically observed at the margin of glaciers, ice caps or ice sheets.  
10 The buried massive ice body has a whitish appearance owing to its high concentration of air  
11 bubbles. Coarse-bubbly ice is the most abundant type (90%–95 %) of englacial ice found in glaciers  
12 (Allen et al., 1960). Our results also show that the cross-sectional area of the crystals of the buried  
13 massive ice is smaller than that of neighbouring C-93 glacier ice, but there is no significant  
14 difference in their circularity ratio (Mann–Whitney–Wilcoxon test,  $p=0.89$ ). However, the  
15 difference in ice crystal size is not unforeseen since the ones from glacier ice can show variations  
16 of the order of a few millimetres to several centimetres in diameter (Gow, 1963; Thorsteinsson  
17 et al., 1997). Patterns of preferred crystal orientation combined with the occurrence of  
18 deformation features in the form of debris bands suggest that the ice has been subjected to long-  
19 continued shear stress caused by the motion of a glacier (Benn and Evans, 2010; Lawson, 1979).  
20 The debris bands cross-cutting the buried glacier ice are comparable to those observed in the  
21 terminus zone of Stagnation Glacier on Bylot Island, where basal sediments were transported to  
22 the glacier surface through shear planes (Moorman and Michel, 2000).

23

**Tableau 4.** Comparisons of the physical properties of different massive tabular ground ice found in the permafrost.

	Ice type	General appearance	Texture	Fabric	Gas inclusions	Reference (location, country)
<b>Intrasedimental ice</b>	<b>Massive segregated-intrusive ice</b>	<input type="checkbox"/> Clear <input type="checkbox"/> Sediment-free to sediment-poor	<input type="checkbox"/> Variable crystal size, but mostly medium to large grains (cm-scale) <input type="checkbox"/> Slightly elongated crystals <sup>1</sup> <input type="checkbox"/> Ahnedral to subhedral	<input type="checkbox"/> Preferred near-vertical oriented c-axis <sup>1</sup>	<input type="checkbox"/> Bubble-poor to bubble-rich <input type="checkbox"/> Small (mm- to cm-scale) <input type="checkbox"/> Elongated and tubular bubbles <sup>1</sup> <input type="checkbox"/> Trains of small spherical bubbles <sup>1</sup> <input type="checkbox"/> Inter- and intra-crystalline	Pollard, 1990 (Hershel Island, Canada) Yoshikawa, 1993 (Spitsbergen, Norway) Harry et al., 1988 (Yukon Coastal Plain, Canada) Gell, 1976 (Mackenzie Delta, Canada) Dallimore and Wolfe, 1998 (Richards Island, Canada) Mackay and Dallimore, 1992 (Tuktoyaktuk, Canada) Mackay, 1989 (no specific site)
	<b>Buried glacier ice (englacial)</b>	<input type="checkbox"/> Clear to milky white <input type="checkbox"/> Deformation structures <sup>2</sup> <input type="checkbox"/> Foliations of bubble-poor and bubble-rich ice	<input type="checkbox"/> Wide range of crystal sizes (sub-mm to tens of cm) <input type="checkbox"/> Mostly large grains (cm-scale) <input type="checkbox"/> Equigranular <input type="checkbox"/> Interlocked crystal boundaries	<input type="checkbox"/> Random or preferred c-axis orientation <sup>3</sup>	<input type="checkbox"/> Bubble-poor to bubble-rich <input type="checkbox"/> Located at crystal junctions <input type="checkbox"/> Bubbles truncated at the ice-sediment contact	Pollard, 1990 (Hershel Island, Canada) Gow, 1963 (Ross Ice Shield, Antarctica) Tison and Hubbard, 2000 (Glacier de Tsanfleuron, Switzerland)
	<b>Buried glacier ice (basal)</b>	<input type="checkbox"/> High sediment content <input type="checkbox"/> Debris laminations <input type="checkbox"/> Deformation structures <sup>2</sup> <input type="checkbox"/> Suspended pebbles and cobbles	<input type="checkbox"/> Small grains (mm-scale) <input type="checkbox"/> Ahnedral to subhedral	<input type="checkbox"/> Weak to strongly oriented c-axis	<input type="checkbox"/> Bubble-free to bubble-poor <input type="checkbox"/> Small ( $\approx$ m to mm-scale) <input type="checkbox"/> Flattened bubbles <input type="checkbox"/> Bubbles truncated at the ice-sediment contact <input type="checkbox"/> Located at crystal junctions	Sharp et al., 1994 (Variegated Glacier, U.S.A.) Sugden et al., 1995 (Beacon Valley, Antarctica) Murton et al., 2005 (Tuktoyaktuk Coastlands, Canada) Ingólfsson and Lokrantz, 2003 (Yugorski Peninsula, Russia)
<b>Buried snowbank</b>	<input type="checkbox"/> Milky white to light brown <input type="checkbox"/> Loosely compacted structure <input type="checkbox"/> Bands of nearly horizontal pale brown ice <input type="checkbox"/> Organic inclusions (e.g. twigs and leaf fragments)	<input type="checkbox"/> Small grains (mm-scale) <input type="checkbox"/> Crystal size area usually reaches a few mm <sup>2</sup> <input type="checkbox"/> Subhedral to euhedral <input type="checkbox"/> Equigranular	<input type="checkbox"/> Random c-axis orientation	<input type="checkbox"/> Bubble-rich <input type="checkbox"/> Layers of bubbles and dispersed in the ice <input type="checkbox"/> Small spherical bubbles ( $\approx$ m- to mm-scale) <input type="checkbox"/> Elongated and tubular bubbles <input type="checkbox"/> Vertically oriented bubbles	Lacelle et al., 2009 (Central Yukon, Canada) Pollard and Dallimore, 1988 (Yukon Coastal Plain, Canada) Ostrem, 1963 (Barnes Ice Cap, Canada; Kebnekaise, Sweden; Jotunheimen, Norway)	

<sup>1</sup> Parallel to the heat flow direction, indicating that freezing is downward.<sup>2</sup> Occurrences of debris bands, boudinage and pinch-and-swell structures, folding, thrust-faulting).<sup>3</sup> Depending on the position of the ice within the glacier or ice sheet at the time it becomes buried.

1 Cations and stable water isotopes measured in the buried massive ice also support its  
2 glacial origin. The low cation content in the buried massive ice is statistically similar to that of the  
3 ice of glacier C-93. Although a slope of 6.0 was calculated between  $\delta D$ - $\delta^{18}O$  of the buried ice,  
4 lower than the Global Meteoric Water Line (slope=8; Craig, 1961), this was due to the small range  
5 of the data ( $\delta^{18}O$ :  $-34.4$  ‰ to  $-33.4$  ‰), which prevented the calculation of a reliable regression  
6 slope. The  $\delta D$ - $\delta^{18}O$  values of the buried massive ice are similar to those of Pleistocene-age ice  
7 from the Barnes Ice Cap on the adjacent Baffin Island (median  $\delta^{18}O$ :  $-33.1$  ‰; Zdanowicz et al.,  
8 2002). This is supported by the average D-excess of  $6.6 \pm 2.5$  ‰ for the buried massive ice, which  
9 is within the range of the values of Barnes Ice Cap (Lacelle et al., 2018) and glacier C-93 ( $5.2 \pm 5.8$   
10 ‰) on Bylot Island.

11  
12 The average  $\delta^{18}O$  value of the buried massive ice ( $-34.0 \pm 0.4$  ‰) is much lower than that  
13 of the other ice types sampled in the study area (Fig. 13b). The  $\delta^{18}O$  values of the buried massive  
14 ice are also lower than in the ice of Penny Ice Cap from the last glacial period  
15 ( $\delta^{18}O \sim -31.3 \pm 1.1$  ‰; Fisher et al., 1998), but in the range of the late Pleistocene ice of Barnes  
16 Ice Cap (Zdanowicz et al., 2002). The late Pleistocene  $\delta^{18}O$  values on Barnes Ice Cap were  
17 6 ‰–10 ‰ lower than the expected  $\delta^{18}O$  [modern – last glacial maximum] and interpreted to be  
18 the product of two factors: (1) temperature effect – cold climate during the last glacial period that  
19 caused a  $\delta^{18}O$  shift of 6 ‰–7 ‰ at that latitude (Fisher and Alt, 1985) and  
20 (2) elevation effect – ice sourced from higher elevation on Foxe Dome ( $\sim 2200$ – $2400$  m), which  
21 caused an additional depletion in  $\delta^{18}O$  of local precipitation (Hooke and Clausen, 1982; Zdanowicz  
22 et al., 2002). Klassen (1993) suggested that the alpine glaciers on Bylot Island were larger but did  
23 not change much in elevation during the late Pleistocene; however, the  $\delta^{18}O$  values of our buried  
24 ice require ice sourced from elevations in the 2150–2350 m range and as such was likely sourced  
25 from the Foxe Dome (Lacelle et al., 2018). Mega-scale glacial lineations and streamlined  
26 landforms were mapped on the floor of Navy Board Inlet, Eclipse Sound and Lancaster Sound by  
27 De Angelis and Kleman (2007) and interpreted as a product of LIS ice streams. We proposed that  
28 the study area was most likely an area of confluence of LIS ice and local alpine glaciers during the



1 LGM as an ice stream moved through Navy Board Inlet and onto south-western Bylot as proposed  
2 by Dyke and Hooper (2001).

3

#### 4 **2.6.2 Burial and preservation of late Pleistocene glacier ice in permafrost**

5 The texture (sand and gravel), stratification and poorly sorted nature of the sediments  
6 (unit B) directly overlying the buried glacier ice suggest that glaciofluvial ice-contact sediments  
7 were deposited in direct contact with or at least in close proximity of glacier ice by surface glacial  
8 meltwater streams flowing on the surface of the glacier. Similar sediment characteristics were  
9 obtained by Fortier and Allard (2004) for glaciofluvial sands (angular grains: 75 %) located a few  
10 kilometres away from the study site. The abundance of angular grains (70 %) in unit B indicates  
11 little abrasion, which is consistent with transport by glaciofluvial water whereby grains were  
12 carried over short distances allowing little wear. Furthermore, the sharp and unconformable  
13 contact between the buried glacier ice and the overlying sediment suggests that thermal erosion  
14 caused by sediment-laden water flow affected the top of the ice of unit A. Our interpretation is  
15 similar to other studies that reported the burial of glacier ice by glaciofluvial sedimentation  
16 (Dallimore and Wolfe, 1988; Kaplanskaya and Tarnogradskiy, 1986; Ingólfsson and Lokrantz, 2003;  
17 Belova et al., 2008). The uppermost unit (C) is a diamicton, which has undergone reworking by  
18 non-glacial processes such as gravitational mass wasting (e.g. slumping and solifluction). In ice-  
19 contact environments, the sediment cover is subject to several cycles of subsidence and  
20 redeposition as the ice undergoes progressive and partial melting (Schomacker, 2008). The shape  
21 of the clasts found within this unit provides evidence that it has experienced active glacial  
22 transport, which is also supported by the occurrence of erratic clasts (gneiss) derived from distal  
23 bedrock located several kilometres from the study site. To summarize, large stagnant ice blocks  
24 could have been covered by glaciogenic sediments at or near the ice margins during the advance  
25 and stagnation of a glacier within the Qarlikturvik Valley. The burial of the ice occurred as  
26 meltwater streams deposited sediments in direct contact with glacier ice followed by the  
27 reworking and redeposition of supraglacial sediments, which formed a surficial cover of mud and  
28 sand, later affected by cryoturbations as is indicated by the incorporation of organic material  
29 dated at  $885 \pm 15$   $^{14}\text{C}$  yr BP (1164 cal yr BP;  $1\sigma$  range: 1058–1204).

1 A clear discontinuity in the  $\delta^{18}\text{O}$  and cations profiles is observed at the ice–sediment  
2 contact and in the overlying units (Figure 16 and 17a). Strong contrasting profiles between buried  
3 ice and the overlying sediment are related to different sources of water and formation history.  
4 The supernatant water from the ground ice in the overlying sediments (units B and C) has much  
5 higher average  $\delta^{18}\text{O}$  values of  $-17.6 \pm 5.4$  ‰. These values are comparable to  $\delta^{18}\text{O}$  values  
6 obtained from precipitation at Pond Inlet between 1990 and 1992 ( $-23.8$  ‰; IAEA/WMO, 2016)  
7 and modern segregation ice which generally has  $\delta^{18}\text{O}$  values of  $-18$  ‰ to  $-22$  ‰ (Michel, 2011).  
8 The average concentrations of  $\text{Ca}^{2+}$ ,  $\text{Na}^+$   $\text{Mg}^{2+}$  and  $\text{K}^+$  in units B and C are 10–86 times greater than  
9 those in the buried glacier ice. The substantial increase observed in the cation content from the  
10 sediment layers can be attributed to the great amount of mineral dissolution before the water  
11 freezes (Lacelle and Vasil'chuk, 2013).

12  
13 The formation of these ice-cored landforms follows a typical sequence of events where  
14 the ice is first buried under a sufficient amount of sediments exceeding the active-layer thickness  
15 (Lukas et al., 2005; Benn and Evans, 2010). Beyond the active margins of many glaciers and ice  
16 sheets, there are broad areas of glacial deposits, till and outwash underlain by glacier ice.  
17 Gravitational and glaciofluvial processes are often identified as the most important processes of  
18 sediment deposition and reworking that lead to the burial of englacial ice (Benn and Evans, 2010).  
19 In basal zones of glaciers and ice sheets, in situ melting of stagnant debris-rich basal ice produces  
20 a supraglacial sediment cover (i.e. melt-out till) that accumulates on the ice, which inhibits its  
21 ablation. Both stagnant englacial and basal ice becomes progressively isolated from the upper  
22 active-flowing ice. These ice-cored landforms adjust to non-glacial conditions and their evolution  
23 is strongly linked with climate-driven processes (e.g. active-layer deepening due to warmer  
24 atmospheric temperature or active-layer detachment slides following heavy rains). In the Arctic,  
25 large blocks of buried late Pleistocene glacier ice were left undisturbed for several millennia owing  
26 to cold and dry climatic conditions that favoured permafrost aggradation following deglaciation.  
27 Such ice-cored landforms inherited from the partial melting of buried glacier ice are common in  
28 formerly glaciated permafrost regions across northern Canada and Siberia but their spatial  
29 distribution remains poorly known (Belova et al., 2008; Ingólfsson and Lokrantz, 2003; Kokelj et

1 al., 2017a; Dyke and Savelle, 2000; Lakeman and England, 2012). In this study, the deposition of  
2 a cover of coarse and well-drained sediment exceeding the average active-layer thickness of the  
3 area has probably been the most important factor limiting the melting of the ice. Furthermore,  
4 plant colonization and the development of a continuous vegetation cover with organic  
5 accumulation have changed the thermal properties of the ground, which reduced heat flow from  
6 the atmosphere to the permafrost during the summer and favoured heat loss during winter. This  
7 resulted in a thinner active layer (the permafrost table moved upward) and ground ice  
8 aggradation. The preservation potential of buried glacier ice on a millennial timescale following  
9 the glacial retreat depends on the complex interactions between climate, geomorphology and  
10 the physical properties of the sediment cover. Between 1999 and 2016, ground thermal regime  
11 monitoring in an intact low-centred polygon near the study site showed maximum active-layer  
12 depths varying between 0.3 and 0.7 m, while the buried ice is located > 1 m below the ground  
13 surface. While the properties of the sediment cover had positive feedback on the long-term  
14 preservation of the buried ice, thaw slumping activity, and not atmospheric temperature per say,  
15 was a fundamental driver of its degradation by exposing the ice and accelerating its melting.

16

## 17 **2.7 Conclusion**

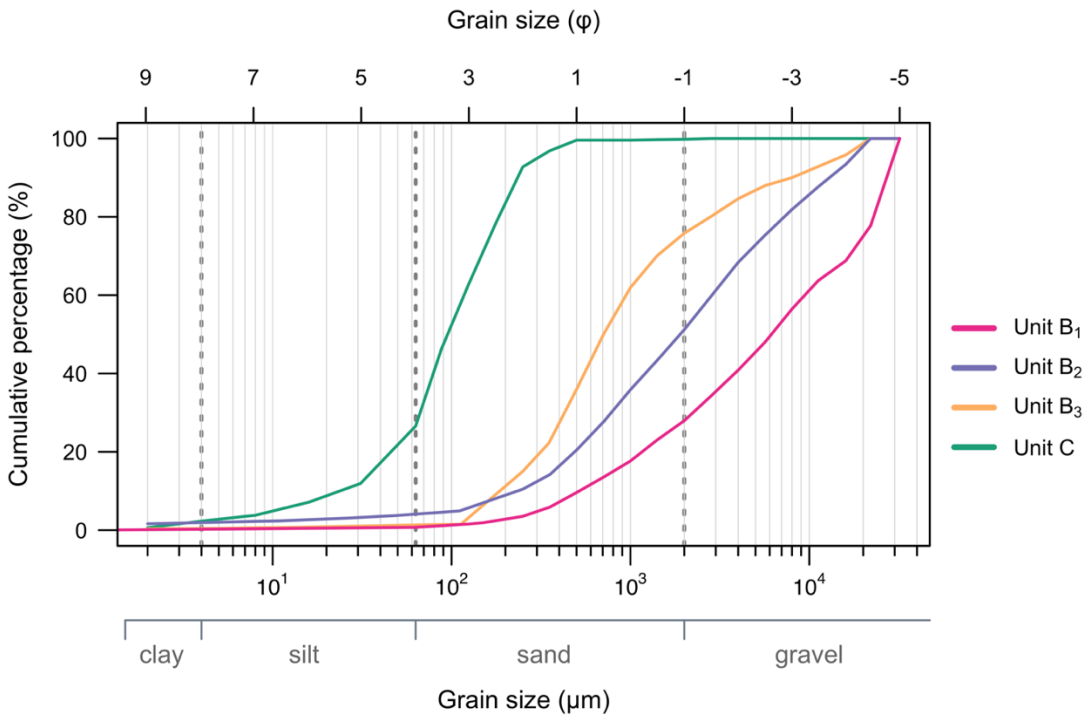
18 This study confirms that the permafrost of Bylot Island contains remnants of Pleistocene  
19 glacier ice that survived the last deglaciation. Geomorphic and cryostratigraphic observations  
20 along with the crystallographic properties of the massive ice suggest its englacial origin. Evidence  
21 that support the englacial origin of the massive ice are (1) sharp and unconformable upper contact  
22 between the ice and the overlying glaciofluvial sediments; (2) clear to whitish ice, with large  
23 crystals; (3) bubble-rich ice, with small gas inclusions (~ mm) mainly located at crystal junctions;  
24 (4) occasional debris bands of sand and fine gravel cross-cutting older debris-free ice;  
25 (5) geochemical similarities with contemporary glacier ice. The geochemical data show a clear  
26 discontinuity at the buried ice–sediment contact as well as the very low cation content similar to  
27 that of modern glacier ice. An origin of massive ice from a Pleistocene glaciation is suggested  
28 based on the isotope data: the buried englacial ice is strongly depleted in heavy isotopes, similarly  
29 to regional Pleistocene glacier ice. Glacier-derived permafrost contains ice that predates the

1 aggradation of the permafrost and represents unique environmental archives to reconstruct  
2 paleoenvironmental conditions at the study site. Although stable isotopes cannot yield  
3 information on absolute ages, these data show that the glacier ice body originated from an ice  
4 stream flowing from the Foxe Dome of the Laurentide Ice Sheet, which was subsequently buried  
5 and preserved in the permafrost of Bylot Island for thousands of years. The sedimentological data  
6 and interpretations presented in this study demonstrate that the first phase in the burial of the  
7 ice involved glaciofluvial deposition directly on the ice at the glacier front, which was followed by  
8 mass wasting and plant colonization. Knowledge regarding the occurrence, origin and  
9 preservation of buried glacier ice is of a great interest due to its potential impacts on the  
10 landscape stability upon melting. Since Bylot Island has experienced several periods of native and  
11 foreign Pleistocene glaciations and based on our findings, we propose that buried glacier ice  
12 should occur elsewhere on the island and probably on adjacent Baffin Island. In a context of  
13 climate change, active-layer deepening and increased activity of slope processes, such as active-  
14 layer detachment slides, thaw slumping and thermo-erosional gullyng, will very likely expose  
15 buried glacier ice and initiate major landscape changes, with cascade effects on the ecosystems.  
16

## 17 **2.8 Acknowledgements**

18 We are grateful to Etienne Godin, Katryne Larrivée, Naïm Perreault, Audrey Veillette,  
19 Jonathan Lasnier, Gaëlle Crête, Michel Paquette and Sabine Veuille, for their help in the field and  
20 in the laboratory. We also thank the team of Gilles Gauthier (U. Laval) and the staff of the Sirmilik  
21 National Park for logistical support and access to Bylot Island. This project was funded by  
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26 Adaptation to Permafrost in Transition” (ADAPT). Finally, this paper has benefited from  
27 constructive comments from Michael Fritz and one anonymous reviewer. We would also like to  
28 thank the handling editor, Julia Boike.

## 1 Appendix A



2  
3 **Figure 19.** Grain-size distribution curves of the sedimentary units (two samples per subunit). The  
4 gravel fraction was removed prior to analysis (appendix A1).  
5

## 6 Appendix B

7 To further investigate the cryostratigraphic characteristics of the ice, all samples were  
8 observed under X-ray-computed tomography (CT) scanning (Siemens SOMATOM Sensation 64).  
9 This technique relies on the calculation of the linear attenuation coefficient that measured the  
10 density of an object passed through an X-ray beam at different angles. A CT scan produces cross-  
11 sectional images (usually 512 by 512 pixels matrix) of an object where each pixel of the image is  
12 assigned an X-ray attenuation value ( $\mu$ ), also called a CT number. CT numbers are then  
13 standardized using the Hounsfield scale, where the radiodensity of water ( $\mu_{\text{water}}$ ) is arbitrarily  
14 defined as 0 HU (Hounsfield units) according to Eq. (1). Different shades of grey are assigned  
15 specific CT numbers to create the displayed image using a specific image processing software (Fiji)  
16 dedicated to DICOM (Digital Imaging and Communications in Medicine) images. In permafrost

1 samples, unconsolidated sediments and rock (high-density minerals) appear white or light grey,  
2 as the attenuation of these materials is very high. Gases inclusions and water appear black, and  
3 other materials, such as ice, can have various shades of grey depending on their density.

$$4 \quad HU \text{ value} = \frac{\mu - \mu_{water}}{\mu_{water}} \times 1000$$

5 This tool helps to refine cryostratigraphic characterization of permafrost cores as it can reveal  
6 characteristics otherwise difficult or even impossible to observe with the naked eye. It allows  
7 visualization and characterization of the internal components of the frozen sample, such as the  
8 ice, grain-size variations and gas inclusions. From a quantitative perspective, it has been used to  
9 segment images into regions of ice, gas and sediment in order to quantify the volumetric content  
10 of the sample scanned (Calmels et al., 2010; Dillon et al., 2008).

11

## 12 **Appendix C**

13 The typical method of measuring ice texture and fabric re-quires that an ice core or block  
14 is sliced into 1cm thick vertical and horizontal sections with a band saw and then thinned to the  
15 desired thickness (< 0.5 mm) following the standard microtoming procedure. Three photographs  
16 were taken of each sample at different angles (0°, 45°, 90°) under cross-polarized light to show  
17 the crystal boundaries as clearly as possible. The analysis is based mostly on qualitative  
18 observations, although statistics on crystal dimensions and geometry were obtained from a digital  
19 analysis of thin-section photographs. It should be noted that the values obtained from these  
20 parameters do not represent exact values and give limited information since a thin section is only  
21 a two-dimensional representation of the crystals (sectioning effect). A digital image analysis using  
22 ImageJ software allowed us to delineate and measure three geometrical parameters of each  
23 crystal. The area (mm<sup>2</sup>), the long axis (mm) and the circularity ratio determine the crystal texture  
24 (size and shape). The long axis is a measurement of the primary axis of the best-fit ellipse to an  
25 ice crystal. The circularity ratio (C) is a dimensionless parameter used to quantitatively describe  
26 the shape or outline form of a crystal. It is a function of the perimeter P and the area A:

$$27 \quad C = \frac{4\pi A}{p^2}$$

1 As  $C$  approaches 1, crystals become more circular, and as it approaches 0, crystals become very  
2 narrow relative to their length. Therefore, the circularity ratio is a measure of the deviation of a  
3 crystal cross-section shape from a perfect circle.

4

# Chapitre 3

## Early Pleistocene glaciation of the eastern Canadian Arctic Archipelago: new evidence from buried glacier ice remnants, Bylot Island, Nunavut.

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*In prep.*



### 1 **3.1 Abstract**

2 Massive ice was found beneath glacial sediments on a 500-m a.s.l. plateau on Bylot  
3 Island. The ice was interpreted as a remnant of basal glacier ice and presented cryostratigraphic  
4 and isotopic properties similar to those of basal ice studied of contemporary glaciers. The age of  
5 remnants of basal glacier ice buried in the permafrost was estimated through combined  
6 paleomagnetic, radiocarbon, diatoms and pollens data. The vertical succession of normal-  
7 reversed-normal polarities recorded by the sediments overlying the ice suggests that the age of  
8 the buried glacier ice must be at least 0.77 Ma or possibly older than 2.5 Ma. As a result, it  
9 represents the oldest glacier ice preserved in ice-free Arctic landscapes, and an earliest evidence  
10 of a Pleistocene glaciation in the eastern Canadian Arctic Archipelago.

11

### 12 **3.2 Introduction**

13 In the Arctic, buried glacier ice is widespread near the margin of formerly glaciated  
14 landscape (Astakhov, 2013; Coulombe et al., 2019; Dyke and Savelle, 2000; Everest and Bradwell,  
15 2003; Kokelj et al., 2017a; Lacelle et al., 2007; Lakeman and England, 2012). It can be preserved  
16 for thousands to millions of years in permafrost environments and survive interglacial climate  
17 conditions if the ice is buried deeper than the depth of seasonal thawing of the active layer  
18 (Astakhov, 1986; Belova et al., 2008; Bibby et al., 2016; Coulombe et al., 2019; Kaplanskaya and  
19 Tarnogradskiy, 1986; Lacelle et al., 2007; Murton et al., 2005; Swanger, 2017). These remnants of  
20 buried glacier ice preserved in the permafrost are unique sources of paleoglaciological,  
21 paleoclimatological and paleogeographical information whose potential remains largely  
22 unexploited. It provides complementary data on (i) the timing and configuration of glaciers and  
23 paleo-ice sheets, (ii) the sub-glacial conditions and processes beneath paleo-ice sheets, (iii)  
24 deglaciation history and (iv) the physical and geochemical properties of these glaciers and ice  
25 sheets (Astakhov, 1986; Kaplanskaya and Tarnogradskiy, 1986; Knight, 1997; Lacelle et al., 2018;  
26 Waller et al., 2009). In North America, most studies have focused on constraining the extent,  
27 retreat pattern, geometry and ice flow regime of the Laurentide Ice Sheet (LIS) and local glaciers  
28 since the Last Glacial Maximum (LGM; 27.5–23.3 ka; Brouard and Lajeunesse, 2017; Dyke et al.,

1 2002; Hughes and Gibbard, 2015; Lamoureux and England, 2000; Margold et al., 2015; Tarasov et  
2 al., 2012). Yet comparatively little is known about the configuration and evolution of paleo-ice  
3 sheets prior to the LGM because the scarcity of reliable pre-MIS2 (i.e. pre-LGM) terrestrial  
4 records, which tends to be obliterated or removed by successive glaciations (Stokes et al., 2015).  
5 There is potentially a minimum of twenty periods of major ice-sheet build-up during the  
6 Pleistocene (last 2.6 million years) in the Northern Hemisphere, with the most extensive  
7 glaciations (6) extending back to 790–928 ka (MIS 21-24; Batchelor et al., 2019; Ehlers and  
8 Gibbard, 2003; Walker et al., 2018). Terrestrial evidences of the early Pleistocene glaciations are  
9 documented at a few sites across the western Canadian Arctic and Alaska (Duk-Rodkin et al.,  
10 2004, 2010; Evans et al., 2021; Barendregt et al., 2010; Vincent, 1990; Froese et al., 2000;  
11 Barendregt et al., 1996; Mathews et al., 1989; Huston et al., 1990) while records of pre-LGM  
12 glacial advances in the central and eastern Canadian Arctic are either scarce or poorly dated  
13 (Brigham, 1983; Refsnider et al., 2013; Fyles et al., 1998; Miller et al., 2022). Most of the evidence  
14 of glacier expansion in the region over the past million years have been derived from marine  
15 sediments in central Baffin Bay (e.g. Andrews et al., 1988; Li et al., 2011; Simon et al., 2012) or  
16 from the Greenland Ice Sheet (Yau et al., 2016; Souchez et al., 2006; Christ et al., 2021; Bierman  
17 et al., 2014). On Bylot Island, in the eastern Canadian Arctic Archipelago, at least five distinct  
18 episodes of local and regional glaciation extending back to 43 ka BP were recognized prior to the  
19 Neoglacial maximum based on the mapping of glacial deposits onshore and dating of raised  
20 shoreline landforms (Dyke, 2000; Dyke and Hooper, 2001; Klassen, 1993). The widespread  
21 occurrence of erratics (e.g. quartz sandstone) inland and above Eclipse glacial limits (100–500 m  
22 a.s.l., > 43 ka) and foreign drift of pre-Eclipse age in stratigraphic sections allowed the  
23 identification of two older glacial episodes (Baffin and Bylot glaciations) that covered much of the  
24 island, but their absolute age remains unknown (Hodgson and Haselton, 1974; Klassen, 1993;  
25 Klassen and Fisher, 1988). Thus, knowledge of glacial history in this region is very limited for the  
26 Early and Middle Pleistocene. In this study, we report the discovery of glacier ice and related  
27 sediments buried in the permafrost of Bylot Island in the Eastern Arctic Archipelago with the goals  
28 of determining the origin and age of these ice bodies. Our hypotheses were that the buried glacier  
29 ice has a basal origin and originates from a glacial event predating the last major regional

1 glaciation. To test the above hypotheses, we used sedimentological, cryostratigraphic, and  
2 geochemical techniques together with radiocarbon, amino acids, and paleomagnetic dating.

3

### 4 **3.3 Site descriptions and Methods**

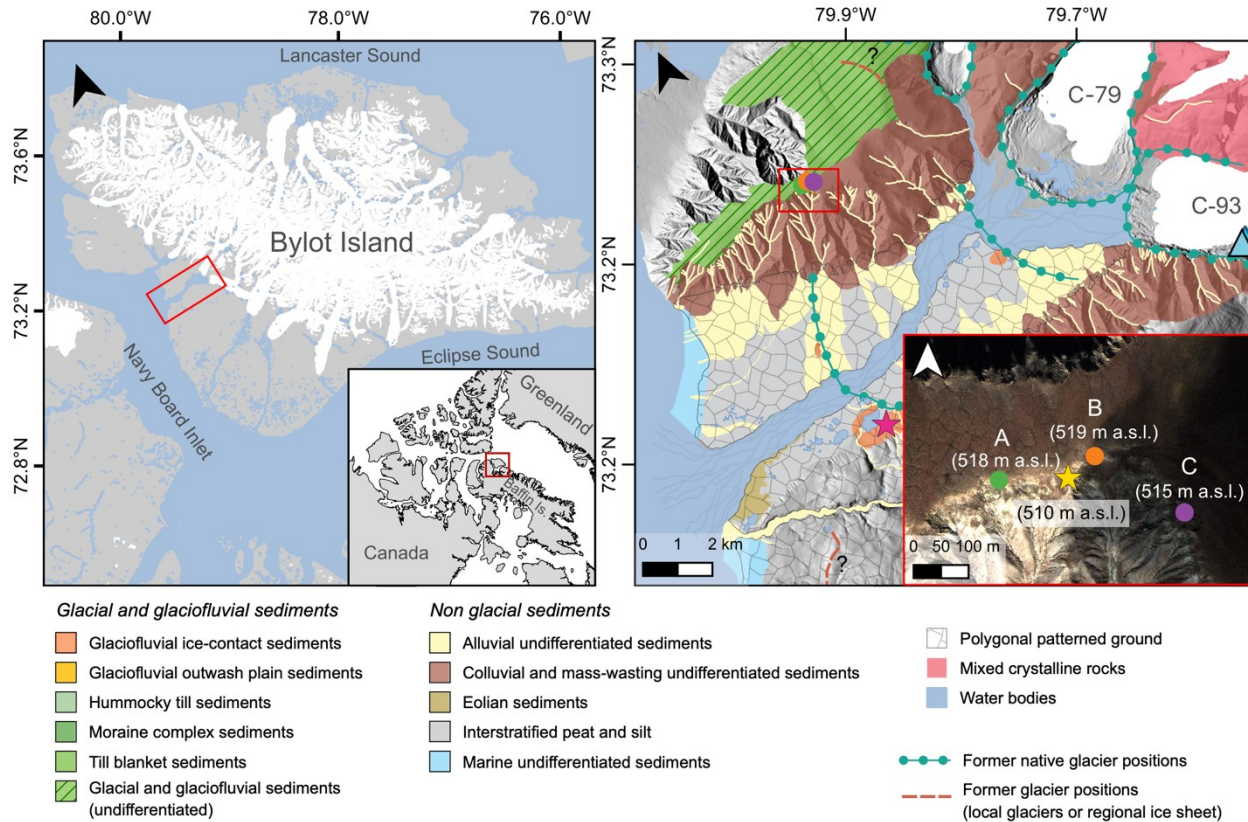
5 Located in the Eastern Canadian Arctic archipelago, Bylot Island is bounded by Lancaster  
6 Sound to the north, Eclipse Sound to the south and Navy Board Inlet to the southwest, which has  
7 been major pathways for ice streams of the LIS (Figure 16; Margold et al., 2015; Stokes et al.,  
8 2016). A local ice cap is centred on the Byam Martin Mountains range (1500–1900 m a.s.l.),  
9 flowing outward as valley glaciers and piedmont lobes towards the lowlands and the coast  
10 (Dowdeswell et al., 2007; Klassen, 1993). The southwestern part of Bylot Island is characterized  
11 by weathered upland plateaus (~ 500 m a.s.l.) intersected by glacially eroded valleys forming  
12 lowlands dominated by glacial, glaciofluvial, alluvial and marine sediments, which have been  
13 transformed by paraglacial and periglacial processes following deglaciation (Allard, 1996;  
14 Coulombe et al., 2022; Klassen, 1993). The surficial deposit on the plateaus typically includes  
15 glacial tills and erratic boulders scattered throughout the surface up to the highest points. These  
16 deposits may be associated with an extensive foreign glaciation named Baffin glaciation that  
17 occurred during a time bracketed by shells characterized by amino acids ratios of 0.8 and 0.6  
18 (D-alloisoleucine to L-isoleucine; Klassen, 1993). In the Arctic, cold temperatures throughout  
19 much of the Holocene have maintained climate-driven, continuous permafrost of hundreds of  
20 meters thick (200–400 m; Moorman and Michel, 2003; Shur and Jorgenson, 2007) which allowed  
21 the preservation of late Pleistocene-age glacier ice in the permafrost.

22

23 The study site is located on the edge of a flat plateau bordering the Qarlikturvik Valley,  
24 where a body of massive ice was found exposed within the headwall of a retrogressive thaw  
25 slump (73°13'N, 79°58'W, 510 m a.s.l.; Figure 20). Based on preliminary field investigations of the  
26 ice characteristics, the massive ice presented striking similarities with cryostructures and  
27 deformations of the debris-rich basal-ice layers observed at the base of contemporary glaciers  
28 and ice sheets (Hubbard et al., 2009; Knight, 1997; Lawson, 1979). In 2011, three shallow ice cores  
29 (140 cm, 31 cm, 36 cm) were extracted from the buried glacier ice (i.e. massive ice), and samples

1 were collected from the sediment layers overlying the ice. For comparison, modern basal ice was  
2 sampled from a fresh exposure at the margin of glacier C-93 (Figure 16). The upper portion of an  
3 ice wedge, which is located on the plateau within 100 m of the buried glacier ice exposure, was  
4 sampled horizontally to compare and show differences in ice facies and isotopic signature ( $^{18}\text{O}$   
5 and D). The buried glacier ice exposure, ice cores, sediment cover, and ice samples from  
6 contemporary basal glacier ice were described using cryostratigraphic techniques, including  
7 characterizing the cryofacies and cryostructures, and determining the volumetric ice content,  
8 grain-size distribution, and major soluble cations, and the  $\delta\text{D}$ – $\delta^{18}\text{O}$  composition of the ice (French  
9 and Shur, 2010; Gilbert et al., 2016; Murton and French, 1994; Lacelle and Vasil'chuk, 2013). A  
10 detailed description of the sampling and analysis techniques is given in supporting information  
11 S1.

12



1

2 **Figure 20.** Map showing the study area on Bylot Island at the northwestern tip of Baffin Island. The

3 map at the left show the surficial geology of the valley and location of the sampling sites. The

4 surficial geology map is based on field work observations, borehole data, interpretation of satellite

5 imagery and digital elevation model and previous maps from Klassen (1993) and Allard (1996). The

6 yellow star indicates the location of the basal glacier ice exposure presented in this study. Wedge

7 ice and segregated ice were also sampled near the basal glacier ice exposure (< 100 m). The blue

8 triangle indicates the sampling location of C-93 basal glacier ice. The green, orange, and purple

9 circles indicate the location of the stratigraphic sections sampled for paleomagnetic analyses. The

10 altitude of each site is indicated in parenthesis.

11

12

13 To determine the relative age of the massive ice, discrete sediment samples were

14 collected for grain size and paleomagnetic analysis (supporting information S2). Sediments

15 directly covering the massive ice were not suitable for paleomagnetic analysis due to surface

16 disturbances (i.e. faults, slumping material) hence samples were obtained from three undisturbed

17 stratigraphic sections (A, B, C) located nearby (between 60 and 220 m) and stratigraphically above

1 the massive ice exposure (Figure 16). The uppermost units of the plateau consists of a laterally-  
2 continuous glaciofluvial sequence capped by a diamicton. Radiocarbon analysis of marine shells  
3 collected in the sediments overlying the buried glacier ice and the dissolved organic carbon (DOC)  
4 fraction extracted from the buried ice allowed to constrain the age of the buried glacier ice and  
5 reinforce the paleomagnetic analysis (supporting information S2). Diatoms and pollen  
6 assemblages from the ice have also been analyzed to provide complementary information of the  
7 age of buried basal ice by investigating its stratigraphical relationship with the late Pliocene–early  
8 Pleistocene fossil forest deposit, which are dated to around 2.4–2.8 Ma (Piroux, 2005; Guertin-  
9 Pasquier, 2012; Csank et al., 2013)

10

## 11 **3.4 Results**

### 12 ***3.4.1 Cryostratigraphy, $\delta D$ - $\delta^{18}O$ and major cations of the buried glacier ice***

13 The massive ice (unit 1) is stratified with alternating layers of debris-rich ice and debris-  
14 poor ice, which has been interpreted as buried glacier ice of basal origin (Figure 21). The contact  
15 between the ice and the overlying material was sharp, continuous, and unconformable. The lower  
16 and lateral contacts of the massive ice were not exposed; hence it could not be delimited. Pebble-  
17 and cobble-sized clasts were abundant in the ice (Figure 21b). Ice layers showed internal  
18 deformations in some locations, such as isoclinal and S-shaped folds (Figure 21c). The ice contains  
19 erratics of crystalline rocks, mostly gneiss with a minor component of granite. The  
20 cryostratigraphic analysis of a frozen core extracted down to 140 cm depth in the massive ice unit  
21 allowed to identify three cryostratigraphic subunits based on their cryostructures (Figure 22 and  
22 Table S1). All subunits were texturally similar, consisting of very poorly sorted gravelly muddy  
23 sand of similar composition (Figures 23 and S1). The ice was bubble-poor (< 7 % in volume), and  
24 the bubbles were typically small and sub-spherical (average volume and sphericity:  $0.37 \pm 5.7$   
25  $\text{mm}^3$ ;  $0.67 \pm 0.1$ ). Throughout the ice cores, ice-coating ( $\sim 1$  mm thick) was occasionally observed  
26 around coarse fragments, forming crustal cryostructures. The bottom subunit 1a (460–470 cm;  
27 VIC of 16%) is characterized by an assemblage of poorly developed reticulate cryostructures and  
28 visible interstitial ice (i.e. pore ice). The upper contact with subunit 1b was sharp. Subunit 1b

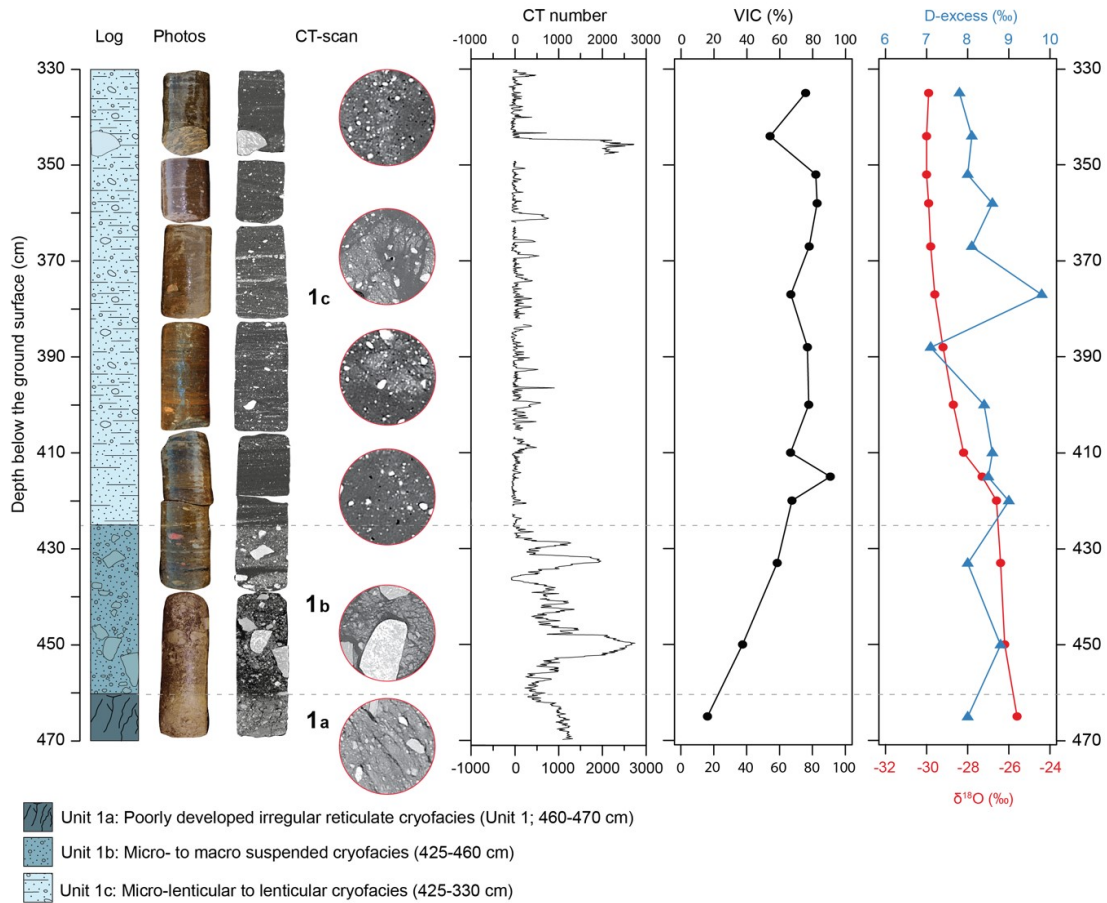


1 (425–460 cm) is dominated by micro- to macro-suspended cryostructures (Fortier et al., 2012).  
2 Volumetric ice content varied from 37 to 58%. Sediment occurs mainly as silty sand, pebble- and  
3 cobble-sized angular to subangular clasts (<5 cm) suspended in clear ice. Subunit 1c (330–425 cm)  
4 displayed alternating layers of sediment-rich to sediment-poor ice (VIC 54 to 91%) with lenticular  
5 to micro-suspended cryostructures. Light-coloured thin laminae (1–5 mm thick) of sand and silt  
6 as well as subangular to subrounded pebbles, cobbles and small boulders (diameters up to 5 cm)  
7 were observed throughout the unit from the CT-scan images. In layers of nearly pure ice, this  
8 subunit also contains occasional clouds of flattened bubbles (~ 1–2 mm in diameter) roughly  
9 parallel to the sediment layers (Figure 24). The  $\delta^{18}\text{O}$  values of the buried glacier ice from the three  
10 cores ranged between -30.0‰ and -22.5‰ (average =  $-27.7 \pm 2.1\text{‰}$ ; Figure 22). The samples were  
11 distributed along a regression slope of 4.6 ( $r^2 = 0.79$ ) and plotted below the local meteoric water  
12 line (LMWL Resolute Bay) of 7.8 (Figure 25). The D-excess values ranged between -21.5‰ and  
13 9.8‰ (average =  $3.7 \pm 8.8\text{‰}$ ; Figure 22). The average concentration of cations varied between  
14  $2.4 \pm 0.7 \text{ mgL}^{-1}$  for  $\text{K}^+$  to  $33.3 \pm 15.2 \text{ mgL}^{-1}$  for  $\text{Na}^+$  (Figure S2;  $\text{Na}^+ > \text{Ca}^{2+} > \text{Mg}^{2+} > \text{S}^+ > \text{K}^+$ ).



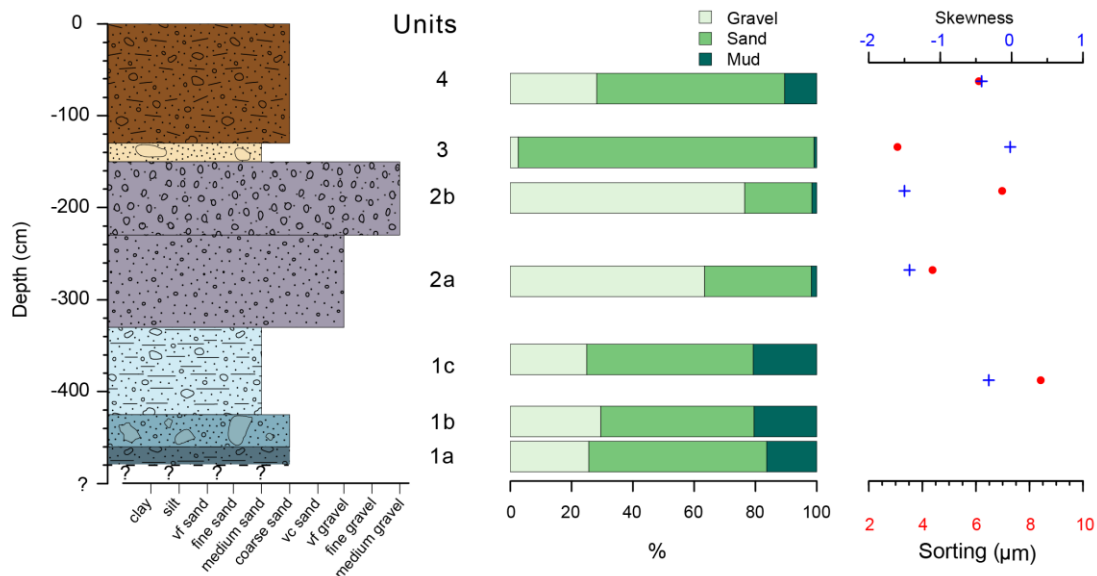
15  
16 **Figure 21.**(a) A photograph showing the general stratigraphy of a well-exposed massive ice  
17 sequence ~10 m wide and ~ 5 m thick and the overlying sediments. The lower and lateral contacts  
18 of the massive ice have not been reached. The yellow squares show the sampling location of the  
19 three ice cores (core 1: 140 cm; core 2: 31 cm; core 3: 36 cm). The thaw depths were measured in

1 late July (2013) with a steel probe at every 10 m along a 150-metre transect at the massive ice  
 2 exposure, and depths ranged between 52 to 62 cm (mean:  $56.1 \pm 2.9$  cm); (b) Pebble- and cobble-  
 3 sized clasts were abundant in the ice;  
 4 (c) S-shaped folds similar to those observed in the basal ice  
 5 of modern glaciers (see figure S3).  
 6



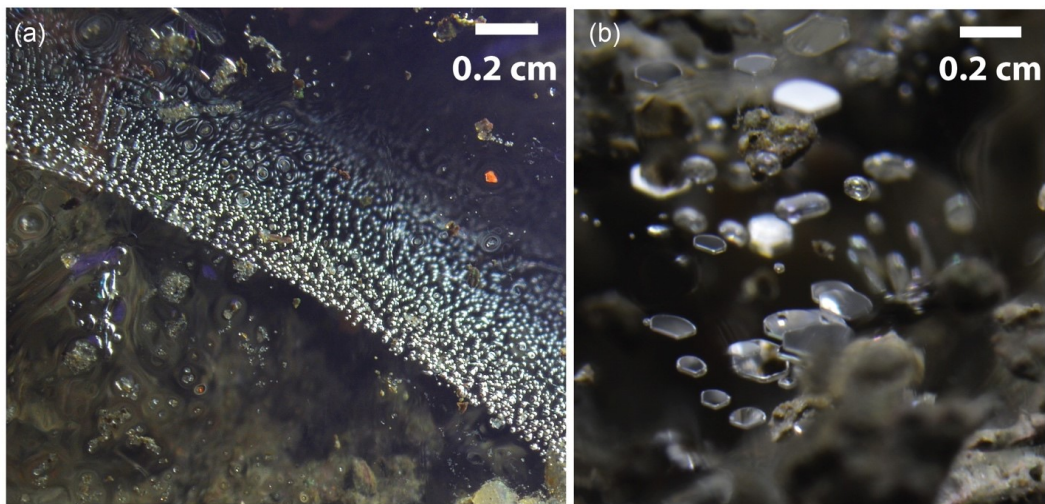
7  
 8 **Figure 22.** Log showing vertical variations in cryostructures, density, volumetric ice content (%),  
 9  $\delta^{18}\text{O}$  and D-excess isotopic composition of core 1 (140 cm).  
 10  
 11  
 12



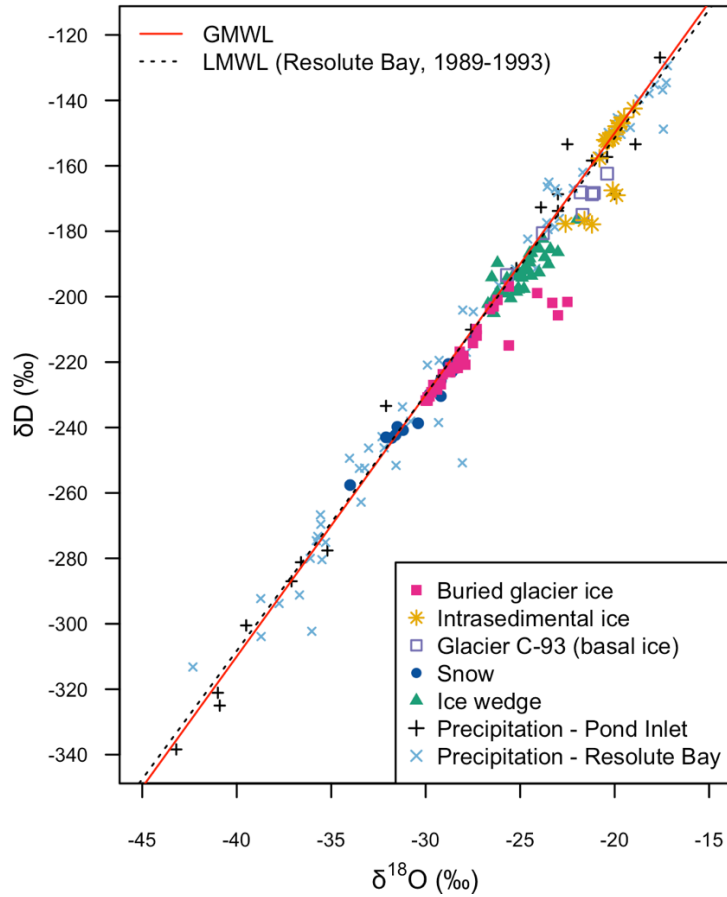


1  
 2 **Figure 23.** Sedimentological data from the stratigraphic units (basal glacier ice and enclosed  
 3 sediments). From left to right: a composite stratigraphic log showing the mean grain size of each  
 4 unit; gravel, sand, and mud percentages; and skewness and sorting.

5  
 6  
 7



8  
 9 **Figure 24.** (a) Very small (sub-mm) subspherical air bubbles concentrated and aligned along a  
 10 crystal boundary, (b) Group of air bubbles in nearly pure ice, and flattened into plates roughly  
 11 parallel to the sediment layers.



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**Figure 25.**  $\delta^{18}\text{O}$ – $\delta\text{D}$  diagram of the massive ice and other types of ground ice (ice wedge, intrasedimental ice), snow and modern basal glacier ice (C-93) sampled on Bylot Island. The intrasedimental ice originates from the frozen sediment located above the buried glacier ice exposure. The solid red line represents the Global Meteoric Water Line (GMWL):  $d = \delta\text{D} - 8\delta^{18}\text{O} + 10$  (Dansgaard, 1964). Also shown are the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  values of precipitation recorded at Pond Inlet (1990–1992) and Resolute Bay (1989–1993; IAEA/WMO, 2016). LMWL is the local meteoric water line for Resolute Bay, Nunavut ( $\delta\text{D} = 7.8\delta^{18}\text{O} + 5.33$ ; IAEA/WMO, 2016) and Pond Inlet, Nunavut ( $\delta\text{D} = 8.0\delta^{18}\text{O} + 10.55$ ; IAEA/WMO, 2016).

For comparison, the sediment-rich basal ice sampled at the margin of glacier C-93 presented an assemblage of lenticular and micro-suspended cryostructure (Figure S2). The ice is characterized by horizontal to subhorizontal layers of poorly sorted debris ranging from silt-sized particles to cobbles and occasional boulders. These discrete layers of debris are separated by undulating 1–10 mm thick, bubble-poor (~5% in volume) layers of clear ice. The larger clasts found

1 within the ice consist mainly of crystalline rocks and appears to have been derived from igneous  
2 and metamorphic rocks of the highlands. Air bubbles were small and subspherical (average  
3 volume and sphericity:  $0.29 \pm 16.4 \text{ mm}^3$ ;  $0.7 \pm 0.1$ ). The modern basal ice had  $\delta^{18}\text{O}$  values ranging  
4 from  $-25.7\text{‰}$  to  $-20.4\text{‰}$  (average:  $-22.2 \pm 1.9 \text{‰}$ ) and a D-excess values ranging from  $-1.3\text{‰}$  to  
5  $12.5\text{‰}$  (average:  $4.1 \pm 5.4 \text{‰}$ ). Modern basal glacier ice composition plot with a regression slope  
6 of 5.5 ( $r^2 = 0.95$ ). The lowest soluble cation concentration was measured in modern basal ice  
7 sampled at glacier C-93, with average concentrations ranging from  $2.5 \pm 0.3 \text{ mgL}^{-1}$  for  $\text{Mg}^{2+}$  to  
8  $16.6 \pm 3.7 \text{ mgL}^{-1}$  for  $\text{S}^+$  (Figure S3;  $\text{S}^+ > \text{Ca}^{2+} > \text{Na}^+ > \text{K}^+ > \text{Mg}^{2+}$ ).

9

#### 10 **3.4.2 Cryostratigraphy, major cations and $\delta\text{D}$ - $\delta^{18}\text{O}$ of the sediment cover**

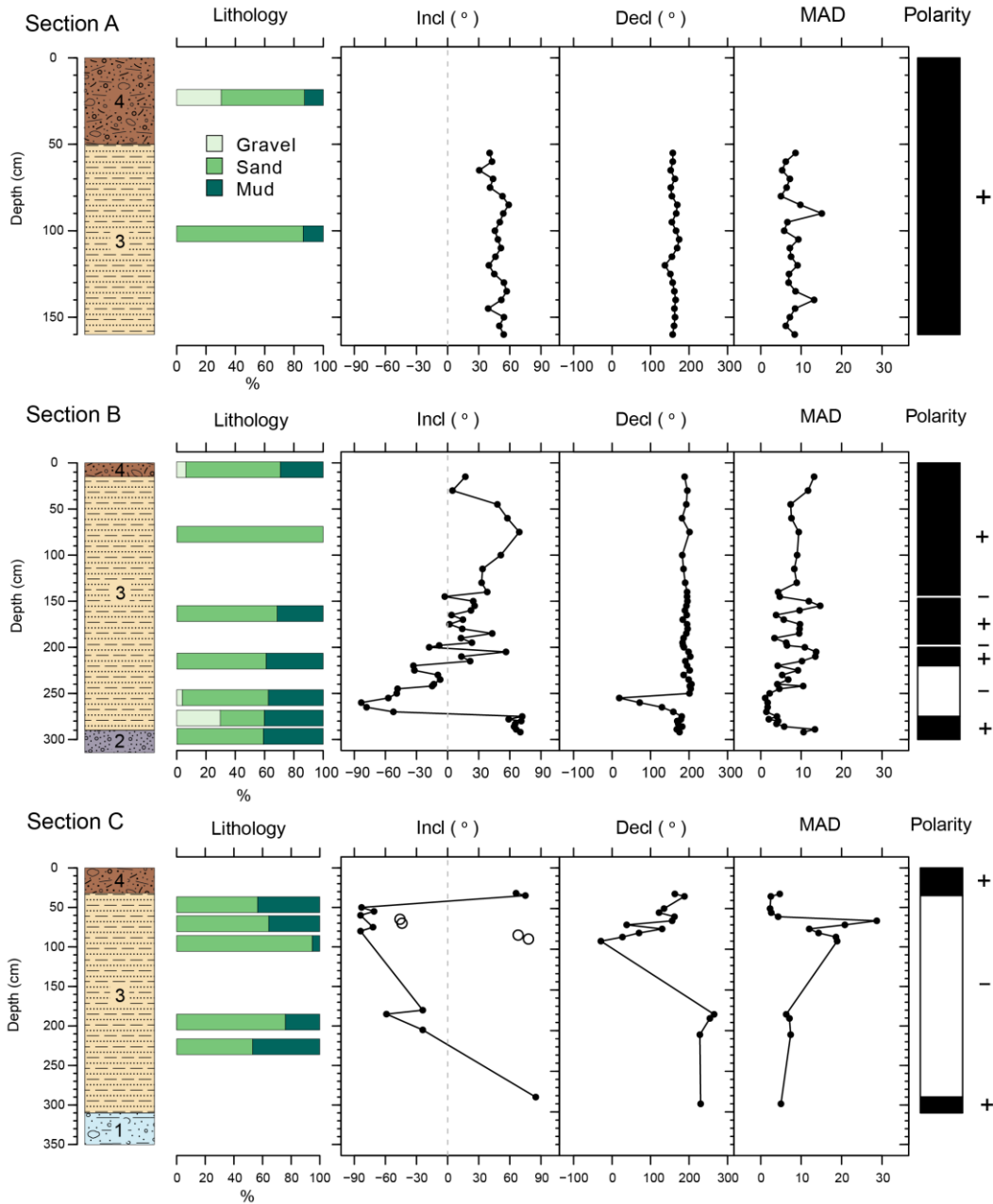
11 Massive ice on the plateau is consistently overlain by 300–400 cm of pebble- and cobble-  
12 rich deposit supported by a sand and gravel matrix (Figure 23). Most clasts are derived from  
13 crystalline rocks. We identified three units based on visual analysis, sedimentological  
14 characteristics and transition between units were generally sharp. The frozen sediment above the  
15 ice (units 2 and 3) presented an ice-poor sediment cryofacies with a structureless cryostructure,  
16 essentially made of visible interstitial ice (i.e., pore ice). Unit 2 (150–330 cm) directly overlaid the  
17 massive ice body (unit 1) and comprises a poorly to very poorly sorted, sandy gravel matrix which  
18 becomes coarser upward (subunit 2b). Pebble- and cobble-sized clasts are also abundant within  
19 unit 2. No sedimentary structures were observed within this unit. Unit 3 (130–150 cm) consists of  
20 poorly sorted, horizontally stratified sand with occasional pebble- to cobble-sized clasts and shell  
21 fragments. Unit 4 (0–130 cm) is a thawed, unstratified cobble-rich diamicton, with a very poorly  
22 sorted gravelly muddy sand matrix and rootlets in the upper portion. A thin, continuous, and  
23 irregular organic layer with rootlets overlaid this unit. The frozen sediments located above the  
24 buried glacier ice had the highest  $\delta^{18}\text{O}$  values, ranging between  $-22.6\text{‰}$  and  $-19.0 \text{‰}$  (average =  
25  $-20.3 \pm 0.8\text{‰}$ ; Figure 25). The D-excess values ranged between  $-9.8\text{‰}$  and  $11.9\text{‰}$  (average =  $6.8$   
26  $\pm 7.1\text{‰}$ ). These samples were distributed on a linear fit described by a much higher slope ( $\delta\text{D} =$   
27  $11.6 \delta^{18}\text{O} - 79.1$ ;  $r^2 = 0.65$ ) when compared to the buried basal glacier ice samples. Average  
28 concentrations of the major cations varied between  $6.4 \pm 3.7 \text{ mgL}^{-1}$  for  $\text{K}^+$  to  $64.7 \pm 68.7 \text{ mgL}^{-1}$  for  
29  $\text{S}^+$  (Figure S3;  $\text{S}^+ > \text{Ca}^{2+} > \text{Na}^+ > \text{Mg}^{2+} > \text{K}^+$ ).

### 1 **3.4.3 Age determination**

2 Radiocarbon dating of shell fragments sampled in the sediments (unit 3) deposited directly  
3 on the ice yielded two dates of  $46960 \pm 770$   $^{14}\text{C}$  yr BP (49 508 cal yr BP;  $1\sigma$  range: 48259–50392)  
4 and  $45200 \pm 1200$   $^{14}\text{C}$  yr BP (47558 cal yr BP;  $1\sigma$  range: 46133–48620). In addition to the  
5 radiocarbon dating, lithostratigraphic logs of three sedimentary sequences and associated  
6 paleomagnetic data are presented in Figure 26. These three sites revealed a similar stratigraphy,  
7 including ice-contact sediments at the base, overlain by glaciofluvial outwash sand and gravel,  
8 and capped by a colluvial diamicton. Section C also comprised a unit of fine-grained eolian  
9 sediments. More detailed descriptions of the stratigraphic sections are provided in Supporting  
10 Information (Figure S4 and S5). The intensity of the natural remanent magnetization (NRM) in the  
11 sediment ranged between  $1.5$  and  $16.7 \times 10^{-5}$  A/m (average =  $3.8 \times 10^{-5}$  A/m). The IRM/SIRM  
12 values were close to 1, which suggests that a low coercivity mineral such as magnetite is the main  
13 magnetic carrier for the remanent magnetization. Maximum angular deviation (MAD) values  
14 were consistently below  $10^\circ$  and average  $\sim 7^\circ$  throughout the sampled sequences, indicating well-  
15 defined directions (Figure 26). Some MAD values were slightly high (between  $10^\circ$  and  $15^\circ$ ) but are  
16 still within the range of values typically considered to indicate a well-resolved magnetization,  
17 while values higher than  $15^\circ$  were excluded from the interpretation as higher MAD values result  
18 in greater directional uncertainty. In section A, inclination values ranged between  $31^\circ$  and  $59^\circ$   
19 ( $n=22$ ), indicating a normal polarity signal. In section B, the polarity was normal in the uppermost  
20 215 cm, with inclinations ranging between ( $1.8^\circ$  to  $69^\circ$ ). However, two short intervals of weak  
21 negative inclinations (3 samples;  $-2.8$ ,  $-8.1$  and  $-17.8^\circ$ ) were documented in this section at 145 cm  
22 and 198 cm. Further down the sequence (220–270 cm), twelve samples exhibited a distinct  
23 reverse polarity signal ( $-7^\circ$  to  $-83^\circ$ ). The bottommost sediment in this section showed a normal  
24 polarity signal ( $59^\circ$  to  $72^\circ$ ). Section C presented a similar vertical paleomagnetic record. The  
25 polarity was normal in the uppermost 50 cm. Further down the sequence (50–97 cm), twelve  
26 sediment samples exhibited a distinct reverse polarity signal ranging between  $-24^\circ$  and  $-83^\circ$ . Only  
27 one sample at the base of this section showed a normal polarity signal, thus they cannot be  
28 correlated to a particular normal period due to lack of samples.

29

1            In total, 29 pollen taxa were identified within the seven buried basal ice samples (Figure  
2 S6). The pollen assemblages displayed little stratigraphic change throughout the subunits (1a, 1b  
3 and 1c). Among the identified taxa, *Pinus* (23–57%), *Picea mariana* (1–18%), *Betula* (up to 12%),  
4 *Larix* (up to 6%), *Populus* (up to 2.5%), and *Abies balsamea* (up to 1.6%) are the main arboreal  
5 pollen types; Ericacea (15–38%, *Juniperus* (up to 4%), *Alnus* (up to 1.6%) and *Salix* (up to 1.4%)  
6 are the most common shrub pollen types; Poaceae (1–10%), Cyperaceae (1.5–4.3%),  
7 Polygonaceae (up to 4%), Caryophyllaceae (up to 2%), *Draba sp.* (up to 2%). et *Thypha latifolia* (up  
8 to 1.6%) are the most dominant herb pollen types. The remainder of the arboreal, shrub and  
9 herbaceous counts (< 1% grains) were dominated by *Tsuga*, *Juglans*, *Epilobium*, Orchidaceae,  
10 *Hedrisarum alpinum*, *Arctostaphylos uva-ursi*, *Thalictrum*, *Drosera rotundifolia*, *Saxifraga*, and  
11 Ranunculacea. Of the total pollen count, about 1–8 % pollen grains remain unidentified. In  
12 addition, most of the *Betula* pollen grains encountered are considered to represent the tree  
13 *Betula papyrifera* rather than shrub forms (i.e., *B. glandulosa*) of this taxon due to the smaller size  
14 of the grains. Unlike pollen grains, no diatoms were found in the buried basal ice samples.



1  
 2 **Figure 26.** Lithostratigraphic logs, paleomagnetic data and polarity logs of three sedimentary  
 3 sequences (A, B and C) located nearby the massive ice exposure. The upper unit of all three  
 4 sections (A, B and C) consist of colluvial sediments with zones of obvious sediment disturbances  
 5 that do not reliably record the geomagnetic field. As a result, no inclinations records are available  
 6 for this unit. In section C, four samples (white circles) were rejected from the dataset, because  
 7 their MAD values were above 15°, indicating unreliable directional data. Incl: Inclination; Decl:  
 8 Declination; MAD: Maximum angular deviation; Black (white) intervals denote normal (reversed)  
 9 polarity.

## 1 **3.5 Discussion**

### 2 ***3.5.1 Origin and burial of the ice***

3           The cryostratigraphic properties of the massive ice along with its isotopic and cation  
4 composition indicate that the exposed ice consists of basal glacier ice buried beneath ice-contact  
5 glaciofluvial sands and gravels preserved within permafrost following ice retreat. Modern basal  
6 ice typically has distinctive physical and chemical characteristics, and typically has a higher debris  
7 content than the overlying cleaner glacier ice as the result of various subglacial processes  
8 operating at or near the glacier bed, such as melting, refreezing, debris entrainment, and  
9 subsequent deformation (Hubbard et al., 2009; Knight, 1997; Waller et al., 2000). Although  
10 modern basal ice cryofacies can vary greatly due to variations of the conditions at the ice-bed  
11 interface, the studied buried ice present striking similarities with basal ice exposed along the  
12 margins of contemporary glaciers and ice sheets, with alternating layers of debris-rich and debris-  
13 poor ice of variable thickness (cm to dm), small ice crystal and a low bubble content (Figure S3;  
14 Christoffersen et al., 2010; Hubbard and Sharp, 1995; Lawson, 1979; Zdanowicz et al., 1996).  
15 According to the basal ice classification of Hubbard et al. (2009), this layered buried basal ice can  
16 be classified as banded (cm to dm) to laminated (mm) dispersed. In contemporary glaciers, the  
17 occurrence of flattened bubbles and small-amplitude folds, from gentle waves to S-shaped folds,  
18 is indicative of shear deformation within the basal ice caused by the motion of a glacier (example  
19 from Matanuska Glacier: Figure S3; Astakhov et al., 1996; Hubbard and Sharp, 1995; Lawson,  
20 1979; Sharp et al., 1994; Waller et al., 2000). Similar deformation structures have also been  
21 observed in buried basal glacier ice preserved in the permafrost in the western Canadian Arctic  
22 and northern Russia (Belova et al., 2008; Fritz et al., 2011; Ingólfsson and Lokrantz, 2003; Murton  
23 et al., 2005). These deformations support the glacial origin, as it remains poorly explained by the  
24 ice-segregation process (Belova et al., 2008). As a result, we attributed the initial formation of the  
25 ice to net basal adfreezing followed by flow-related, glacio-tectonic deformation (Christoffersen  
26 and Tulaczyk, 2003; Herron and Langway, 1979; Hubbard and Sharp, 1995; Lorrain et al., 1981).  
27 During the process of basal adfreezing, repeated cycles of freezing will produce stratified basal  
28 ice as the freezing front migrates into saturated subglacial sediment or meltwater at the glacier  
29 sole (Hubbard, 1991; Weertman, 1961). Together with the cryostructures, the complete absence

1 of diatoms within the ice provides further evidence of its basal origin. While diatom valves are  
2 typically well preserved in the sediments of most permafrost environments (Smol and Stoermer,  
3 2010), the conditions at the base of a glacier or ice sheet likely precluded their preservation.

4  
5 Stable water isotopes measured in the buried massive ice also support the basal ice origin.  
6 The  $\delta^{18}\text{O}$  composition of the buried glacier ice (average  $\delta^{18}\text{O}$ :  $-27.7 \pm 2.1$  ‰) was lower than that  
7 of the intrasedimental ice within the overlying sediment cover (average  $\delta^{18}\text{O}$ :  $-20.3 \pm 0.8$  ‰) and  
8 ice-wedge ice (average  $\delta^{18}\text{O}$ :  $-24.9 \pm 1.1$  ‰) located nearby, indicating a different water source,  
9 climatic conditions and formation history. These  $\delta^{18}\text{O}$  values are consistent with modern  
10 segregation ice ( $-18$  to  $-22$  ‰) and modern ice wedges ( $-22$  to  $-25$  ‰) measured in the Canadian  
11 Arctic (Michel, 2011). Because the isotopic composition of the buried glacier ice is more negative  
12 than younger ice formed in situ in the ground, this suggests that the water contributing to ice  
13 formation originated in a colder environment under glacial conditions. The  $\text{dD}-\text{d}^{18}\text{O}$  regression  
14 slope of 4.6 of the buried basal glacier ice is lower than the slope of the local meteoric water line  
15 in the region (Pond Inlet: 8; Resolute Bay: 7.8), and the slope of meteorically derived glacier ice  
16 sampled in the accumulation zone (englacial ice) of glacier B-7 (7.8; Zdanowicz et al., 1996), while  
17 it was comparable to those of the thick basal sequences exposed at the base of glacier B-7,  
18 Aktineq and C-93 on Bylot Island (4.4, 4.9, 5.5; Jouzel and Souchez, 1982; Souchez et al., 1988;  
19 Zdanowicz et al., 1996). This observed heavier isotope composition is consistent with isotope  
20 fractionation due to refreezing of water at the glacier bed that shifts the  $\delta$  values towards less  
21 negative values and will also decrease the slope of the  $\delta^{18}\text{O}-\delta\text{D}$  co-isotopic relation (Jouzel and  
22 Souchez, 1982; Knight, 1989; Lorrain and Demeur, 1985).

23  
24 The texture, dominated by sand and gravel, and the poorly sorted nature of the sediments  
25 (unit 2; sorting: 4-7) suggest that the basal ice sequence has been primarily covered by  
26 glaciofluvial sediments deposited by glacial meltwater streams flowing directly over the ice at the  
27 glacier terminus that covered the plateaus of Bylot Island. Surface meltwater streams created a  
28 sharp and unconformable contact between the ice body and surrounding sediments (i.e., thaw  
29 unconformity), which is typically associated with buried glacier ice (Belova et al., 2008; French



1 and Harry, 1990; Mackay, 1989; Waller et al., 2009; Worsley, 1999). Differences in sediment  
2 properties between the basal ice unit and the overlying sediments suggest that sediments were  
3 likely not released by in situ melting of debris-rich basal ice. The ice-contact glaciofluvial deposits  
4 is overlain by a diamicton, which is associated with slope sedimentation predominantly resulting  
5 from gravitational downhill movement (e.g., slumping and solifluction) of a glacial till. Basal ice is  
6 the most common form of buried ice reported in the literature as sediment-rich basal ice has a  
7 greater potential to persist in a buried state with the formation of insulating blanket of  
8 supraglacial melt-out till (Belova et al., 2008; Fritz et al., 2011; Murton et al., 2005; St-Onge and  
9 McMartin, 1999).

10

### 11 **3.5.2 Age determination**

12 Of the three stratigraphic sections, only sections B and C exhibited both normal and  
13 reversed magnetic polarities. Some inclination records are consistent with the current expected  
14 geocentric axial dipole (GAD) field inclination for the study site ( $81^\circ$  at  $73^\circ$  N), while other values  
15 are lower, which may be due to the sampling procedure, sediment structures (e.g., slightly tilted  
16 laminations) or fluctuations of the magnetic field over time. Paleomagnetic measurements  
17 obtained from section A indicated a normal polarity signal, further suggesting that the entire  
18 record was deposited during the Brunhes chron (0–0.773 Ma). In sections B and C, the polarity  
19 was normal in the uppermost unit, which also indicates a Bruhnes age for most of the sediments.  
20 The occurrence of two short intervals of weak negative inclinations in section B might be  
21 associated with geomagnetic excursions since no deformation was observed in these intervals,  
22 nor changes in sedimentology, grain size or mineralogy. Polarity change at 220 cm and 60 cm is  
23 attributed to the Bruhnes-Matuyama (B/M) boundary dated at ca. 0.773 Ma (MIS 19; Singer,  
24 2014). Further down the sedimentary sequence, the samples exhibited a distinct reverse polarity,  
25 which can be linked to the Matuyama chron (0.773 to 2.61 Ma; Singer, 2014).

26

27 The bottommost sediment in sections B and C showed a normal polarity signal, which  
28 suggests three possible interpretations. (1) The normal polarity could represent deposition during  
29 the Jaramillo normal subchron (1.071–0.990 Ma), assuming continuous deposition throughout

1 the sections. (2) The normal polarity could also correspond to older normal subchron, (i.e., Cobb  
2 Mountain, Olduvai, Reunion) since stratigraphic gaps are common in glacial sediment  
3 sequences. Such interpretation assumes limited or no deposition nor erosion after initial  
4 deposition. (3) Given its stratigraphic position above a fossil forest deposit, the normal polarity at  
5 the base of sections B and C could be tentatively assigned to the late Gauss normal chron (2.595–  
6 3.032 Ma). Pollen and macrofossil evidence combined with the lack of tree species present in  
7 pollen records from the Arctic after 2.2 Ma indicate that an open boreal forest covered the island  
8 during an interglacial period around the Gauss/Matuyama boundary between 2.4–2.8 Ma (Piroux,  
9 2005; Guertin-Pasquier, 2012; Csank et al., 2013). A Gauss interpretation is supported by the  
10 presence of boreal trees and shrub taxa preserved in the buried basal ice that were also observed  
11 in the fossil forest deposit (Piroux, 2005; Guertin-Pasquier, 2012). Fossil pollen grains from the  
12 underlying Plio-Pleistocene deposit were likely incorporated into the basal ice layer when glacier  
13 ice covered the plateau. The palynological data suggest that the overlying basal ice may be  
14 temporally contiguous with the lower fossil forest deposit, therefore supporting a late Gauss age  
15 for the buried basal glacier ice. This correlation assumes that the Jaramillo, Cobb Mountain,  
16 Olduvai and Réunion normal subchrons were not recorded in the stratigraphic sequence. This  
17 scenario implies that the buried basal ice derives from a glacial event associated with the onset  
18 of extensive northern hemisphere glaciation. To summarize, paleomagnetic and stratigraphic  
19 data suggest that the basal glacier ice could be at least older than 0.77 Ma (e.g.  
20 Bruhnes–Matuyama reversal) or older than 2.6 Ma. Radiocarbon dating of marine shells (>  
21 47.5–49.5 ka) sampled in the upper part of the stratigraphic section suggest that these were  
22 deposited by a subsequent glacial advance on the plateau, thereby suggesting a hiatus in the  
23 sedimentary sequence. Similarly, glacially transported shell fragments were also observed by  
24 Klassen (1993) in a morainic ridge on the same plateau.

25  
26 Under these scenarios, this basal glacier ice body remained largely undisturbed for  
27 hundreds of thousands of years due to the cold climatic conditions that allowed permafrost  
28 aggradation following glacial retreat and throughout the Pleistocene. The minimum age for the  
29 ice suggests that it has survived at least 11 interglacial in the last 800 ka, including marine isotopic

1 stages 5e (last interglacial; 128–116 ka) and 11c (~400 ka), which both stand out as considerably  
2 longer and warmer than the present interglacial (MIS 1; Past Interglacials Working Group of  
3 PAGES, 2016). The rate of thaw and resilience of permafrost is influenced by ice content  
4 (Jorgenson et al., 2010; Woo, 1986; Lee et al., 2014). Permafrost with high ice content, such as  
5 buried glacier ice, is expected to thaw more slowly than permafrost with low ice content due to  
6 differences in latent heat required for thawing. The high volume of ice associated with buried  
7 glacier ice contributes to increase the thermal resistance of the ice, and thereby its long term  
8 preservation. This is illustrated by the occurrence of relict ice wedges of Middle Pleistocene age  
9 that were reported from the central Yukon (>700 ka; Froese et al., 2008) and Siberia (> 650 ka;  
10 Murton et al., 2021). As a result, the persistence of relict ground-ice and very old glacier ice within  
11 both the discontinuous and continuous permafrost zones emphasize the resilience of ice-rich  
12 permafrost to past warm intervals. This suggests that ice-rich permafrost may be more stable and  
13 resistant to the projected warming than previously thought, especially under minimally erosive  
14 conditions (Froese et al., 2008; Murton et al., 2021; Jorgenson et al., 2010; Grosse et al., 2011)

15  
16 The preservation potential of the buried ice is increased by the coarse nature and the  
17 thickness (~3 m) of the deposits that overlies the ice, which is far thicker than the modern active-  
18 layer thickness measured on Bylot Island (0.3–0.7 m in the valley bottom; Allard et al., 2020).  
19 Under certain conditions, undisturbed deposits predating the LGM can be preserved beneath  
20 minimally erosive, cold-based ice sheets or glaciers and can survive extensive and multiple glacial  
21 advances (Christ et al., 2021; Evans et al., 2021; Franke et al., 2021; Stokes et al., 2015; Beel et al.,  
22 2016; England, 1987; Dyke, 1993). The persistence of these remnants of basal glacier ice and  
23 glacial deposits through multiple glacial cycles (~8-10 in the last 1 Ma; Batchelor et al., 2019;  
24 Marshall et al., 2002) implies that (1) all successive glacial episodes must have had frozen bed  
25 conditions to allow for subglacial preservation of older glacial deposits or (2) the plateaus  
26 remained largely ice-free during the subsequent glacial events on Bylot Island. It also indicates  
27 that there has been little to no thermal and mechanical erosion that would have exposed the  
28 buried glacier ice and accelerates its melting (Kokelj et al., 2017b). According to Klassen (1993),  
29 there is little morphological evidence of glacial erosion on Bylot Island, suggesting the ice sheet

1 would have been generally protective of bedrock and older landforms. The nonerosive character  
2 of foreign ice sheets on Bylot Island is also supported by the occurrences of tors in sandstone  
3 bedrock in the southwestern part of the island (Klassen, 1993). Additionally, several studies have  
4 suggested that large areas of high mountains and coastal uplands on Baffin Island remained ice-  
5 free for at least the last glaciation because of their advanced weathering degree and lack of glacial  
6 erosional or depositional features (Ives, 1962; Margreth et al., 2016; Miller et al., 2002; Steig et  
7 al., 1998; Corbett et al., 2016; Miller et al., 2022), suggesting that some of the highest points of  
8 the plateaus of Bylot Island may also have been above the limit of actively eroding glacial ice  
9 during glacial advances post-dating the burial of the basal glacier ice.

10  
11 Very few studies have reported relict glacier ice predating the Wisconsin glaciation (MIS  
12 2-4) in the Canadian Arctic (Lacelle et al., 2007) and the Russian Arctic (Basilyan et al., 2008;  
13 Ingólfsson and Lokrantz, 2003). Basal ice from Dye-3 and GRIP cores in Greenland were estimated  
14 to be at least  $400 \pm 170$  ka and  $970 \pm 140$  ka, respectively (Yau et al., 2016). Although, these  
15 remnants of Middle or Early Pleistocene ice sheet are very rare, several sites across the Arctic  
16 record early glaciations spanning the Matuyama reversed chron (2.61–0.773 Ma), especially in  
17 the western Canadian Arctic (up to seven; Barendregt et al., 2010), Iceland (5 to 13; Geirsdóttir,  
18 2011), and Greenland (up to eight; Knutz et al., 2019). Based on the paleomagnetic data, the  
19 buried glacier ice presented in this study is possibly the oldest dated glacier ice reported in the  
20 Northern Hemisphere and represents the earliest evidence of glaciation in the Eastern Canadian  
21 Arctic Archipelago. As a result, our findings suggest that warm-based glacier ice occupied the  
22 plateau during an early Pleistocene glaciation and left an extensive cover of glacial sediments  
23 and isolated blocks of debris-rich ice (*dead-ice*) that have become separated from the active ice.  
24 Due to its location on a 500-m plateau, the buried basal ice may be associated with the regional  
25 *Baffin glaciation*, which is the oldest and most extensive glaciation reported on Bylot Island by  
26 Klassen (1993). During this glacial episode, regional ice sheets inundated much of the island and  
27 could have flowed northward across the mountain range (Klassen, 1993). The distribution and  
28 elevation of rare foreign erratics (quartz sandstone and carbonate) glacially derived from  
29 northern Baffin Island (e.g. Adams Sound Formation; Blackadar and Davison, 1968; Klassen, 1993)

1 provides evidences for extended ice cover prior to Eclipse glaciation; however, no age constraints  
2 on these erratics were given. These erratics were observed in a morainic ridge located on the  
3 plateau, about 3.7 km from the study site, at 580 m a.s.l. (~ 70 m above the basal ice exposure),  
4 but also on other plateaus (500–700 m a.s.l.) on southern and northeastern Bylot as well as on  
5 many peaks of the Byam Martin Mountain.

6

### 7 **3.6 Conclusion**

8         Based on the results presented above, the massive ground-ice exposed at the headwall of  
9 thaw slump found on the edge of a 500-m plateau on Bylot Island was interpreted as buried basal  
10 glacier ice. The ice is similar to basal ice facies formed at the base of contemporary glaciers and  
11 ice sheets in terms of cryofacies, cryostructures, grain size properties, ionic geochemistry, and  
12 oxygen isotopic composition. Our findings suggest that glacier ice overrode the plateau during an  
13 early Pleistocene glaciation in the eastern Arctic. This glacier left large stagnant ice block that  
14 became progressively isolated from the upper active-flowing ice and buried under a thick cover  
15 of glacial sediments. The persistence of buried glacier ice indicates stability and resilience of  
16 these ice-cored terrain to past warmer climate, and possibly future climate warming.  
17 Interpretation of our paleomagnetic data suggests that the ice has minimum age of 0.773 Ma or  
18 could be older than 2.595 Ma, thereby indicating that it has been preserved since at least the  
19 early Pleistocene. It is possibly correlative of the basal glacier ice at the bottom of the GRIP ice  
20 cores on Greenland and potentially the oldest known dated glacier ice dated in the Northern  
21 Hemisphere.

22

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9 Sirmilik National Park.

10

## 11 **3.8 Supporting information**

### 12 ***S1. Field sampling and analyses***

#### 13 *Field sampling*

14 We described the massive ice exposure (buried basal glacier ice) and overlying sediments  
15 based on lithology, texture, structure, clast size and shape, and the nature of contacts. Unit  
16 boundaries were defined at contacts indicating depositional hiatuses or major changes in material  
17 properties. Three cores were collected using an earth auger equipped with an 8 cm diamond  
18 carbide core barrel. We also sampled the overlying sediment at 10 cm vertical intervals or less  
19 from the surface to a depth of ~ 2 m to obtain the isotopic and cation composition of the sediment  
20 cover. Prior to sampling, at least 10-20 cm of ice and sediment were removed to obtain samples  
21 unaffected by surficial thawing and contamination resulting from the movement of surface  
22 meltwater. All samples were transported frozen and intact to the laboratory for further analysis.

#### 23 *Ct-scan*

24 Computed Tomography (CT) scans on the cores were obtained at the *Institut national de*  
25 *la recherche scientifique, Centre Eau-Terre-Environnement* (INRS-ETE, Quebec City, QC, Canada)  
26 to visualize internal structures (air, ice, sediment) of the frozen cores, and to further investigate  
27 the cryostratigraphic characteristics of the ice (Calmels and Allard, 2008; Calmels et al., 2010;

1 Dillon et al., 2008). CT-scan images of frozen cores were acquired using a SIEMENS® Somaton  
2 Volume Access dual CT scanner (voxel dimensions of 0.205 in the x, y axes and 0.6 mm in the z  
3 axe). Each pixel (picture element) in an image records the linear attenuation coefficient of the  
4 material,  $\mu$ , relative to that of water,  $\mu_w$  (Equation 1). This is known as the CT number, expressed  
5 in Hounsfield Units (HU), and is the linear transformation. The HU value of water is 0 and the HU  
6 value of air is  $-1000$ . Higher density and higher atomic numbers result in greater X-ray  
7 attenuation. In the case of permafrost, black pixels represent air, dark grey pixels ice and white  
8 pixels unconsolidated sediments and rock (high density minerals). Ice can have various shades of  
9 grey depending on its density.

10

$$11 \quad HU = \frac{(\mu - \mu_w)}{\mu_w} \quad (1)$$

12 Raw data from CT scanning was processed using the ImageJ FIJI software platform, which is a  
13 simple and open-source software to conduct image analysis (Schindelin et al., 2012). Illustrations  
14 of radiographs shown in the manuscript correspond to coronal images. Segmentation was  
15 performed using the threshold adjustment function in FIJI, and multiple stack slices were checked  
16 for quality control. The first interval, from  $-1024$  HU to  $-150$  HU, characterised gas-filled pixels.  
17 3D analysis using 3D ImageJ Suite plugin was used to measure quantitative parameters related to  
18 the size (volume,  $\text{mm}^3$ ) and shape of the air bubbles (sphericity index; Ollion et al., 2013). The  
19 sphericity index ( $\psi$ ) is dimensionless parameter measuring the degree to which a particle  
20 approaches the shape of a sphere. Sphericity values range from 1 (perfect sphere  $\psi=1$ ) to 0  
21 (elongated shape).

22

$$23 \quad \psi = 36\pi \frac{V^2}{S^3} \quad (2)$$

#### 24 *Volumetric ice content (%)*

25 Following CT scan imaging, the ice cores were subsampled at about 10 cm intervals using  
26 a circular rock saw and thawed in sealed plastic bags to measure the volumetric ice content (VIC)  
27 of the massive ice cores. The VIC was measured by water displacement method (immersion of  
28 vacuum-sealed samples in water) and calculated using

1 
$$VIC (\%) = \frac{V_i}{V_t} \quad (3)$$

2 where  $V_t$  is the total volume of the frozen sample ( $\text{cm}^3$ ) and  $V_i$  is the volume of ice ( $\text{cm}^3$ ) estimated  
3 from weight loss after drying using the theoretical density of ice ( $0.9175 \text{ g cm}^{-3}$ ). The supernatant  
4 water was immediately transferred in sealed polyethylene bottles airtight bottles for isotopic and  
5 geochemical analysis.

6

### 7 *Stable O-H isotopes and cation contents*

8 The supernatant water extracted from the ice cores and soil samples above the massive  
9 ice were analyzed for stable water isotopes ( $\text{d}^{18}\text{O}$ ,  $\text{dD}$ ) and cation concentrations at the University  
10 of Ottawa Stable Isotope Laboratory (Ottawa, ON, Canada) to discriminate between different  
11 water sources and detect geochemical discontinuities (Lacelle and Vasil'chuk, 2013; Michel,  
12 2011). In 2013, all samples were thawed, filtered ( $0.45 \mu\text{m}$  diameter filter) and acidified in the  
13 laboratory in sealed polyethylene bottles prior to analyses. Stable water isotopes analyses were  
14 carried out with a Los Gatos Research high-precision liquid water analyzer coupled to a CTC LC-  
15 PAL autosampler, with an analytical reproducibility for  $\delta^{18}\text{O}$  and  $\delta\text{D}$  is  $\pm 0.3\text{‰}$  and  $\pm 1\text{‰}$ ,  
16 respectively. The results are expressed in delta per mil notation (d, ‰) relative to the Vienna  
17 Standard Mean Ocean Water (VSMOW) standard. Cation concentration in the ice and snow ( $\text{Al}_{\text{tot}}$ ,  
18  $\text{Ca}^{2+}$ ,  $\text{Fe}_{\text{tot}}$ ,  $\text{K}^+$ ,  $\text{Mg}^{2+}$ ,  $\text{Mn}_{\text{tot}}$ ,  $\text{Na}^+$ ) were measured by inductively coupled plasma optical emission  
19 spectrometry (Vista Pro ICP-OES) at the University of Ottawa. Solutes are expressed in milligrams  
20 per litre and analytical reproducibility was  $\pm 1 \%$ .

21

### 22 *Grain-size distribution*

23 Detailed sedimentological characteristics of the sediments covering the ice were  
24 documented in the field, unit depths, boundaries, and sedimentary structures. The dominant clast  
25 lithology was determined visually in the field. Particle-size distributions of sediment cover and  
26 buried basal ice were determined by dry sieving at  $1/2\phi$  intervals (size ranges  $-12$  to  $4\phi$ ). The  
27 hydrometer method was used to determine the distribution of the finer particles smaller than  $4\phi$   
28 (ASTM Standard D422, 2007). Grain size distribution and statistical parameters were calculated



1 using the G2Sd package in the open-source software R (version 4.0.5; Fournier et al., 2012; R  
2 Development Core Team, 2021).

3

## 4 **S2. Age determination**

### 5 *2.1 Radiocarbon dating*

6 Two shell fragments from the overlying sediment (unit 3) were sampled and radiocarbon-  
7 dated by accelerator mass spectrometry (AMS) at the Keck-Carbon Cycle AMS Facility (University  
8 of California, Irvine, CA, USA). Calibrated ages (cal yr BP) were calculated using “CALIB 8.2”  
9 (Stuiver et al., 2021; Reimer et al., 2020; IntCal20 dataset).

10

### 11 *2.2 Paleomagnetic analyses*

12 Three undisturbed stratigraphic sections (A, B, C) adjacent to the buried glacier ice  
13 exposure were selected for detailed paleomagnetic studies. Paleomagnetism is a relative dating  
14 technique based on the principle that changes in the Earth's magnetic field are preserved in rocks  
15 and sediments (e.g., Stoner and St-Onge, 2007). Prior to collection, we cleaned the sediment  
16 exposure to a vertical face and inserted cubic plastic boxes horizontally into the sediment at 2 cm  
17 intervals or less. No samples were collected from the uppermost unit, which is interpreted as a  
18 glacial diamicton reworked by paraglacial slope processes. Paleomagnetic measurements were  
19 carried out at the *Institut des sciences de la mer de Rimouski* (ISMER, Rimouski, QC, Canada) with  
20 a 2G-Enterprises™ SRM-755 cryogenic magnetometer and pulse magnetizer module (for  
21 Isothermal Remanent Magnetization, IRM). The natural remnant magnetization (NRM) was  
22 measured using stepwise alternating field (AF) demagnetization steps from 0 to 80 mT (5 mT  
23 increments) until the samples were almost fully demagnetized. Inclination and declination of the  
24 characteristic remanent magnetization (ChRM) were determined by principal component analysis  
25 (PCA; Kirschvink, 1980) using an Excel spreadsheet (Mazaud, 2005) with AF demagnetization steps  
26 from 5 to 50 mT (9 steps). Polarities were assigned to each sample based on mean directions. The  
27 maximum angular deviation (MAD) and median destructive field (MDF) values were also  
28 computed and used as indicators of the quality of the palaeomagnetic data. Mean directions are

1 considered unreliable where MAD values are  $\geq 15^\circ$  (Stoner and St-Onge, 2007). The median  
2 destructive field (MDF) is a magnetic mineralogy indicator depending on the nature of the  
3 magnetic minerals and on the magnetic grain size (Dankers, 1981). When the magnetic  
4 mineralogy is uniform (e.g., magnetite-dominated mineralogy), the MDF reveals information on  
5 the magnetic grain size. An anhysteretic remanent magnetization (ARM) was acquired in a 100 mT  
6 AF with a direct current (DC) bias field of 0.05 mT. Additionally, an isothermal remanent  
7 magnetization (IRM) and a saturated isothermal remanent magnetization (SIRM) were acquired  
8 applying a DC field of 0.3 and 0.95 T. After each step, the ARM, IRM, and SIRM were demagnetized  
9 and measured with the same AF demagnetization steps as for NRM.

10

## 11 *2.2 Pollens analysis*

12 Pollen extraction was performed at Biogeography Laboratory (Université de Montréal,  
13 Montreal, QC, Canada) on seven subsamples obtained from the 140-cm ice core sampled from  
14 the buried basal ice exposure. Standard pollen processing procedures were then followed (Faegri  
15 and Iversen, 1989). Each sample was sieved (15 and 225  $\mu\text{m}$ ) and subjected to successive chemical  
16 treatments with HCl (removal of carbonates), KOH (deflocculates the sample and breaks down  
17 organics), HF (removal of clays) and acetolysis solution (removal of cellulose). Samples also  
18 underwent heavy-liquid separation using sodium polytungstate (SPT) to assist in the removal of  
19 heavy minerals. *Lycopodium* spore tablets were added to allow for the calculation of total pollen  
20 and spore concentrations following Stockmarr (1971). The residue was then mounted in glycerin  
21 and analyzed under light microscope at 400x magnification. The microscopic analyses revealed  
22 sufficiently high pollen concentration and sufficiently good preservation of pollen grains to allow  
23 the counting of 190 to 260 pollen grains per sample. Pollen was identified using the Biogeography  
24 Laboratory (University of Montreal) slide reference collection, as well as published atlases  
25 (McAndrews et al., 1973; Moore and Webb, 1980). In total, 35 levels were counted, with at least  
26 190 pollen and spores enumerated for most subsamples. Taxa percentages were calculated based  
27 on setting the total sum of pollen as equal to 100%. Indeterminable pollens includes damaged  
28 pollens and unidentifiable pollens (i.e., did not match any of the reference collections). Results of

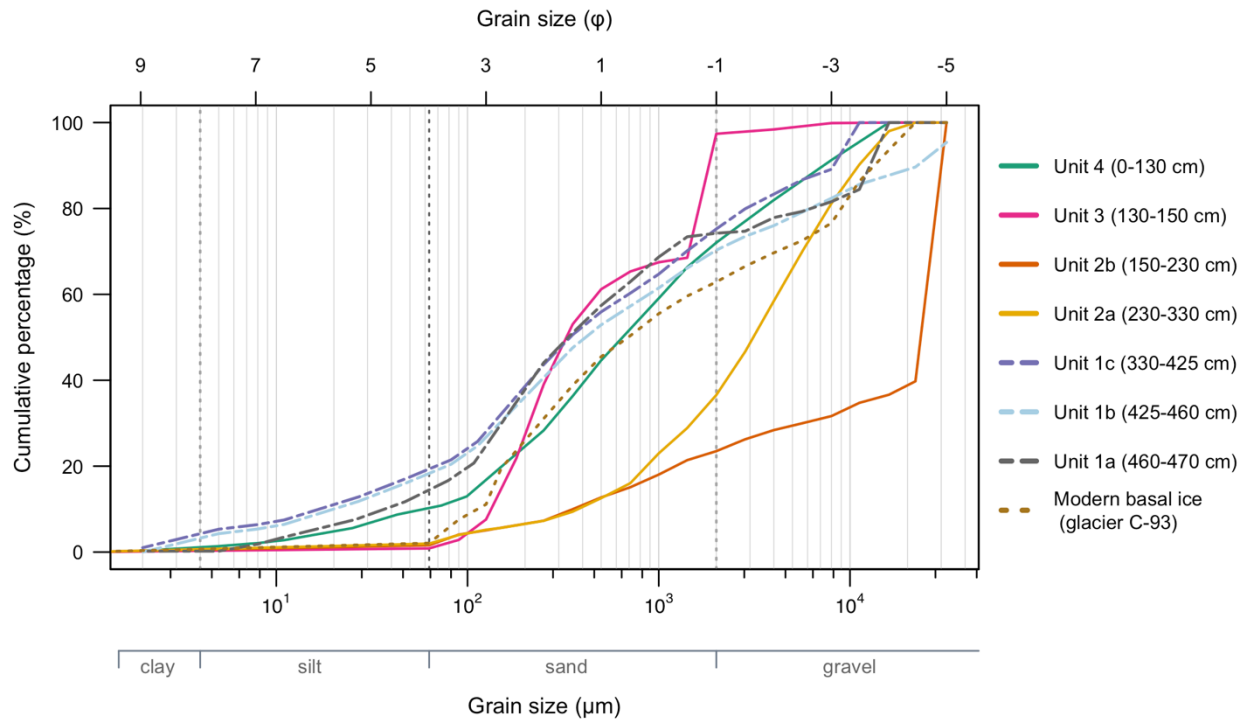
1 pollen analyses are displayed in a diagram produced with the *rioja* (v. 0.9-26) R package (Juggins,  
2 2015).

3

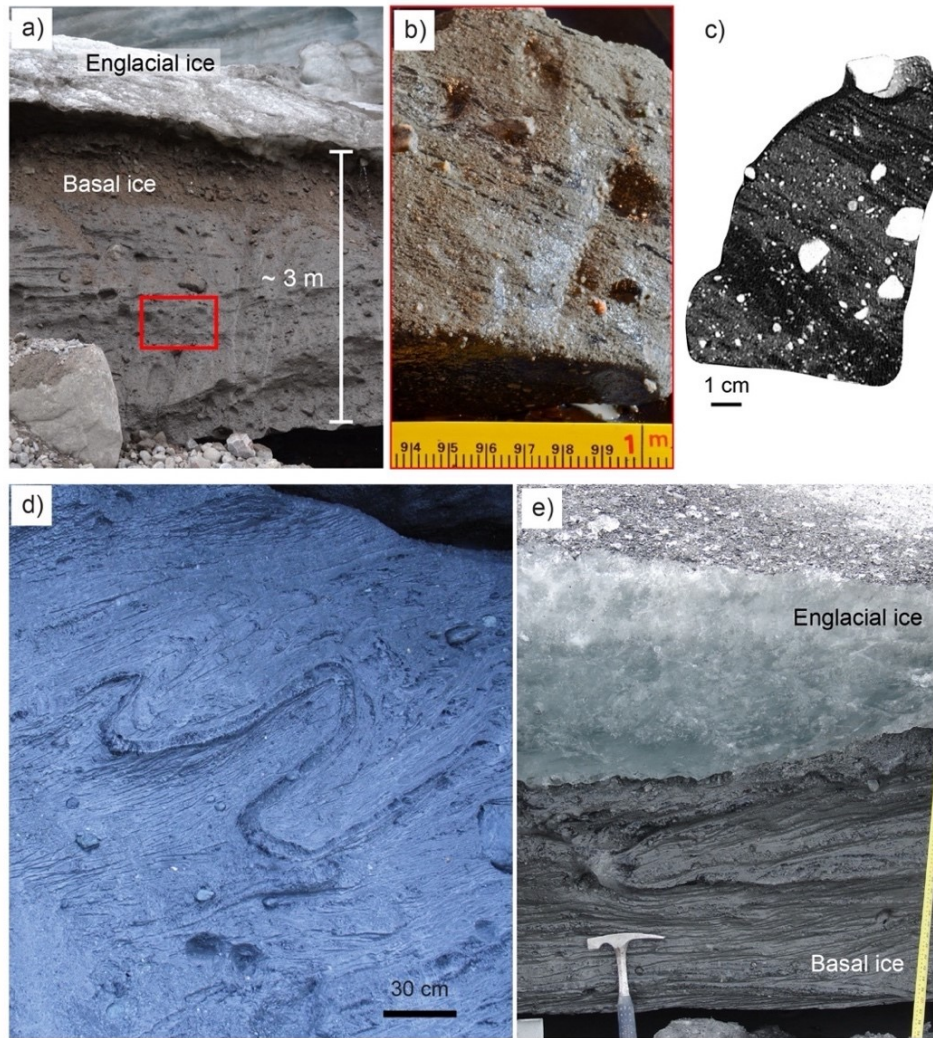
#### 4 *2.3 Diatoms analysis*

5 Sediment samples (n=7) were prepared for diatom taxonomic identification using  
6 standard procedures in the Aquatic Paleoecology Laboratory (Laval University, Canada). The  
7 cleaned diatom samples were dried on glass cover slips and mounted in Naphrax® mounting  
8 medium. Diatoms were identified and counted at 1000X magnification using a Leica DMRB  
9 microscope.

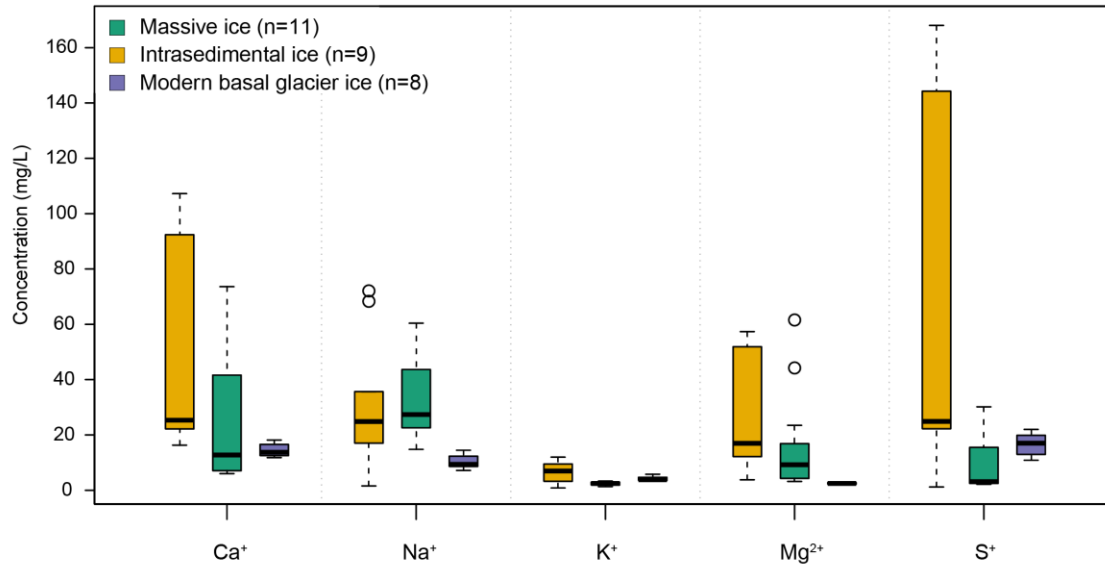
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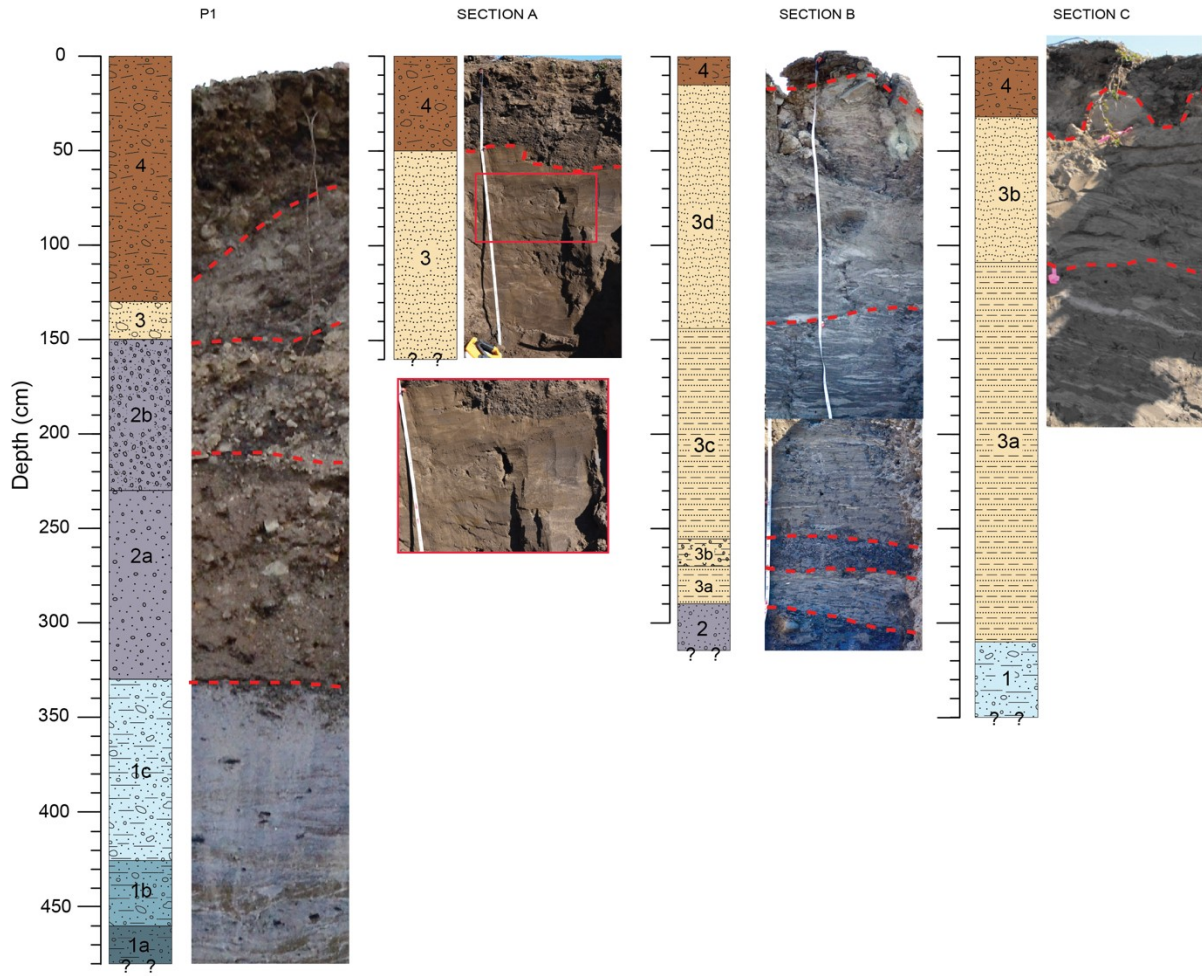
1  
 2 **Figure S1.** Grain-size distribution curves of the sedimentary units (two samples per subunit)  
 3 overlying the massive ice (buried basal glacier ice).  
 4



1  
2 **Figure S2.** (a) Basal glacier ice exposure observed on glacier C-93 (Bylot, Nunavut), (b) Sample of  
3 basal glacier ice. The sediment-rich basal ice sampled at the margin of glacier C-93 presented an  
4 assemblage of lenticular and micro-suspended cryostructures. The matrix (sandy gravel) was very  
5 poorly sorted (sorting values: 6.8 and 9.5) and the grain-size distribution tended to approximately  
6 symmetrical ( $\sim 2$ ). (c) Transverse cross section from a scan showing the internal structure of the  
7 modern basal ice sample (glacier C-93; ice is dark grey; sediment is light grey, and clasts are white).  
8 The CT-scan image shows discrete layers of debris are separated by undulating layers of clear ice.  
9 (d) S-shaped folds observed in basal ice layers of the Matanuska glacier, Alaska. The basal ice layers  
10 are also characterized by a layered appearance, high debris content, the dominance of silt-sized  
11 sediment, and abundant clasts. No sedimentary data are available. (e) Recumbent fold observed in  
12 basal ice of the Matanuska glacier. The clear to milky white ice located above the sediment-rich ice  
13 is englacial ice.  
14



1  
 2 **Figure S3.** Box plots showing soluble cation concentration of the buried basal glacier ice, basal  
 3 glacier ice of glacier C-93 and intrasedimental ice sampled within the sediment layers covering the  
 4 massive ice unit.  
 5



1



## Massive ice exposure



Unit 4: Colluvial sediments (0 - 130 cm, gravelly muddy sand; contains numerous subrounded to subangular cobbles and small boulders; very poorly sorted; 28.0 % gravel, 61.4 % sand, 10.5 % mud)



Unit 3: Glaciofluvial sediments (130 - 150 cm; the matrix consists of faintly stratified gravelly sand with occasional subrounded to subangular cobbles; poorly sorted; upper contact is well defined; contains shell fragments; 2.6% gravel, 96.5% sand, 0.9 % mud)



Unit 2: Ice-contact sediments (150 - 330 cm; sandy gravel; poorly to very poorly sorted; contains numerous subrounded to subangular cobbles and small boulders; gradational contact with overlying unit)



Subunit 2b: 150 - 230 cm; 76.5% gravel, 21.9% sand, 1.6 % mud.



Subunit 2a: 230 - 330 cm; 63.4% gravel, 34.6% sand, 2.0 % mud.



Unit 1: Massive ice (330 - ? cm; gravelly muddy sand; very poorly sorted; upper contact is well defined)



Subunit 1c: 330 - 425 cm; 25.0% gravel, 54.3% sand, 20.7 % mud.



Subunit 1b: 425 - 460 cm; 29.7% gravel, 49.9% sand, 20.5 % mud.



Subunit 1a: 460 - ? cm; 25.8% gravel, 57.7% sand, 16.4 % mud.

## Section A



Unit 4 : Colluvial sediments (0 - 50 cm, muddy sandy gravel; contains occasional subrounded to subangular pebbles and small boulders; very poorly sorted; 30.4 % gravel, 56.7 % sand, 12.9 % mud)



Unit 3: Glaciofluvial sediments (50 - ? cm; slightly gravelly muddy sand containing faint, discontinuous and wavy laminations; moderately sorted; upper contact is well defined; 0% gravel, 86.3 % sand, 13.7 % mud)

## Section B



Unit 4: Colluvial sediments (0 -15 cm, gravelly muddy sand; contains numerous subrounded to subangular pebbles and boulders; very poorly sorted; 6.3 % gravel, 64.4 % sand, 29.3 % mud)



Unit 3: Glaciofluvial sediments (15 - 290 cm; muddy sand; poorly to moderately sorted)



Subunit 3d: 15 - 144 cm; slightly gravelly sand with faint, discontinuous and wavy laminations; upper contact is well defined; 0.1 % gravel, 99.9 % sand, 0 % mud.



Subunit 3c: 144 - 255 cm; fine to medium sand laminated and interbedded with dark grey gravelly muddy sand; upper contact is gradational; 1.3 % gravel, 62.7 sand, 35.9% mud.



Subunit 3b: 255 - 270 cm; massive gravelly mud; faintly stratified; upper contact is well defined; 29.8% gravel; 30.1% sand; 40.1% mud.



Subunit 3a: 270 - 290 cm; very fine sand laminated and interbedded with dark grey muddy sand; upper contact is well defined; 0.1% gravel; 59.1% sand; 40.8% mud.



Unit 2: Ice-contact sediments (290 - ? cm; gravelly sand; poorly to very poorly sorted; contains occasional subrounded to subangular cobbles and small boulders; upper contact is well defined; 24.1% gravel, 75.9% sand, 0% mud)

## Section C



Unit 4: Colluvial sediments (0 - 32 cm; sand and gravel; contains occasional subrounded pebbles and small boulders no sedimentological data)



Unit 3: Glaciofluvial sediments (32 - 310 cm, muddy sand)



Subunit 3b: 32 - 109 cm; Reworked and wind-blown muddy sand; faintly laminated; poorly to moderately sorted; upper contact is well defined; 0% gravel, 56.6% sand, 43.4% mud.



Subunit 3a: 109 - 310 cm; Fine sand laminated and interbedded with dark grey muddy sand; poorly to moderately well sorted; gradational upper contact with overlying unit; grading downward into muddy sand; 0% gravel, 94.8 to 53.1% sand; 5.2 to 46.9% mud.



Unit 1: Massive ice (290 - ? cm; no sedimentological data)

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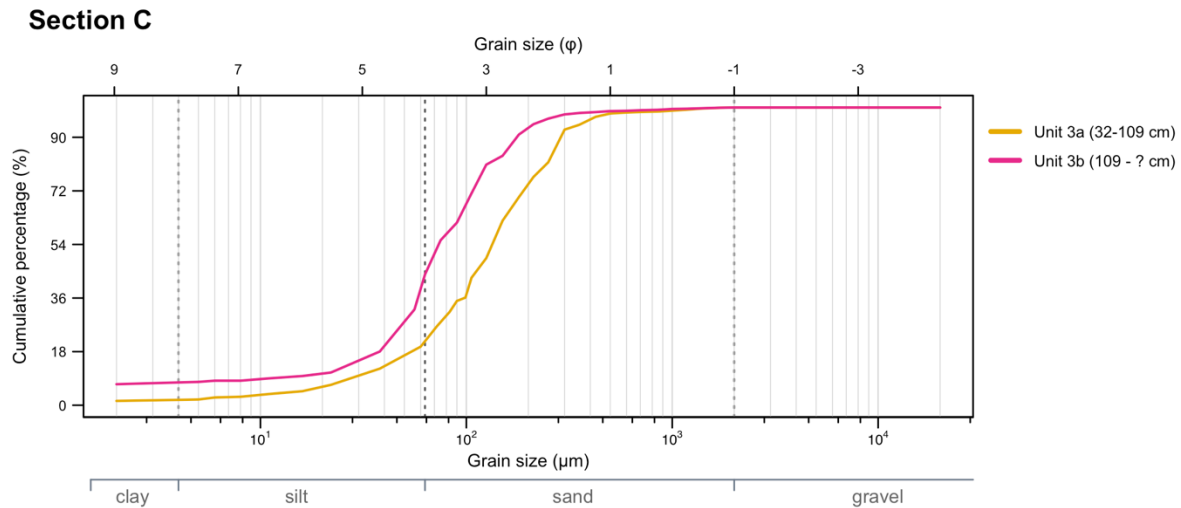
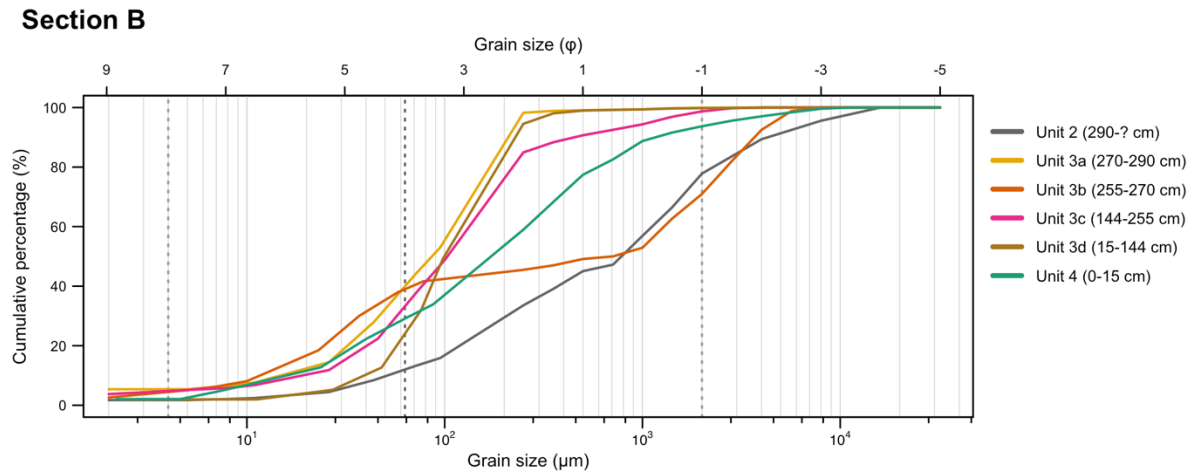
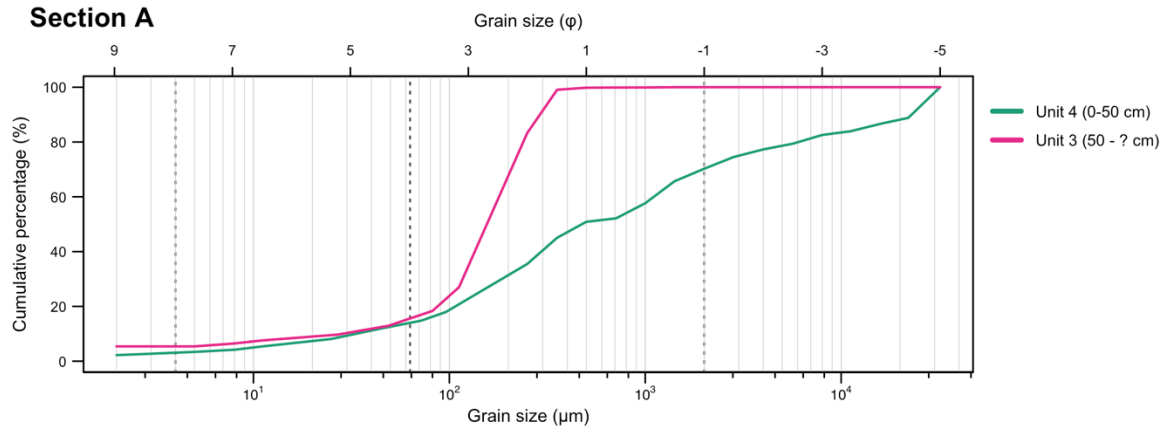
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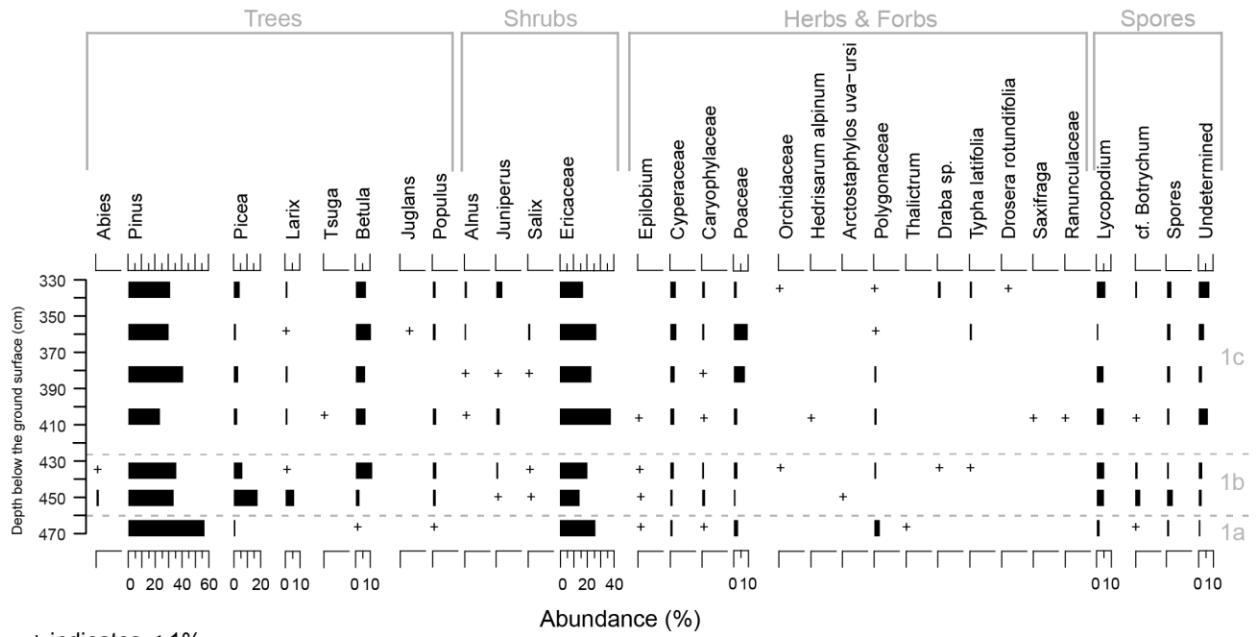
**Figure S4.** Logs and photographs of the stratigraphic sections. P1 represents the general stratigraphy of the buried basal glacier ice and the overlying sediments. Sections A, B and C represent three additional undisturbed stratigraphic sections, located nearby (< 220 m) and above the buried basal ice exposure, where sediment were collected for paleomagnetic and grain size analysis. For section C, no description of unit 1 is available.





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**Figure S5.** Grain-size distribution curves of the sedimentary units (two samples per subunit) of sections A, B and C.



1

+ indicates < 1%

2

**Figure S6.** Percentage diagram presenting pollen and non-pollen palynomorph records from the buried basal ice exposure

3

4

1 **Table S1.** Macrorests (M) and pollen taxa (P) obtained from the fossil forest deposit located on  
 2 Bylot Island (Piroux, 2005; Guertin-Pasquier, 2012) and the buried basal ice unit. Several  
 3 herbaceous taxa (n=11) were identified for the first time in the buried basal ice unit. The  
 4 abbreviation 'cf' indicates uncertainty in providing the species name.

Taxon	Fossil forest deposit				Buried basal ice
	2005		2012		2022
	M	P	M	P	P
<i>Picea glauca</i>	X			X	
<i>Picea mariana</i>	X			X	X
<i>Picea</i> sp.		X	X	X	
<i>Pinus strobus</i>				X	
<i>Pinus</i> cf. <i>banksiana</i>				X	
<i>Pinus</i> sp.	X	X		X	X
<i>Betula</i> cf. <i>papyrifera</i>	X				X
<i>Betula</i> cf. <i>glandulosa</i>	X				
<i>Betula</i> sp.		X		X	
<i>Larix</i> sp.	X	X		X	X
Cupressaceae		X		X	
<i>Thuja/Juniperus</i>				X	X
<i>Abies</i> sp.				X	
<i>Abies balsamea</i>					X
<i>Populus</i> sp.		X		X	X
<i>Fraxinus nigra</i>				X	
<i>Acer</i> cf. <i>rubrum</i>				X	
<i>Carya</i> cf. <i>ovata</i>				X	
<i>Quercus</i> sp.				X	
<i>Ostrya/Carpinus</i>				X	
<i>Salix</i> sp.		X		X	X
<i>Juglans</i> sp.					X
<i>Alnus crispa (viridis)</i>				X	
<i>Alnus incana (rugosa)</i>				X	
<i>Alnus</i> sp.		X			X
<i>Corylus</i> cf. <i>cornuta</i>				X	
<i>Myrica arctogale</i>	X			(M. cf. g)	
Ericaceae		X		X	X
<i>Empetrum</i> cf. <i>nigrum</i>			X		
<i>Andromeda polifolia</i>	X				
<i>Chamaedaphne calyculata</i>	X				
<i>Artemisia</i> sp.		X		X	
Chenopodiaceae				X	
Rosaceae	X			X	
<i>Potentilla</i> sp.		X			
Ranunculaceae				X	X
Cyperaceae		X	X	X	X
<i>Carex</i> spp.	X				
Poaceae		X		X	X

<i>Menyanthes trifoliata</i>	X	X		X	
Saxifragaceae		X			X
<i>Epilobium</i> sp.		X			X
<i>Oxyria digyna</i>		X		X	
Caryophyllaceae					X
Orchidaceae					X
<i>Hedrisarum alpinum</i>					X
<i>Arctostaphylos uva-ursi</i>					X
Polygonaceae					X
<i>Draba</i> sp.					X
<i>Typha latifolia</i>					X
<i>Drosera rotundifolia</i>					X
<i>Lycopodium</i> cf. <i>annotinum</i>				X	X
<i>Lycopodium</i> cf. <i>clavatum</i>				X	
<i>Lycopodium</i> cf. <i>selago</i>					X
<i>Thalictrum</i>				X	X
<i>Potamogeton/Triglochin</i>				X	
Monolete spores				X	X
Trilete spores				X	X
<i>Osmunda</i> cf. <i>regalis</i>	X				
<i>Hippuris</i> cf. <i>tetraphylla</i>	X				
<i>Hippuris</i> cf. <i>vulgaris</i>			X	X	
<i>Aracites globosa</i>	X		X		
<i>Drepanocladus</i> sp.			X		
<i>Sphagnum</i> sp.	X	X	X	X	
cf. <i>Botrychum</i>					X
<b>Total taxa</b>	16	19	7	39	
<b>TOTAL</b>		32		42	30

1

**Table S2.** Summary of the cryostratigraphic and isotopic properties of the buried basal glacier ice and modern basal glacier ice sampled at glacier C-93 (Bylot Island).

Units	Depth (cm)	Cryostructure	VIC (%)	Stable water isotopes (‰)				Air bubbles		Grain size distribution (%)								
				$\delta D$	$\delta^{18}O$	D-excess	n	Sphericity	Volume (mm <sup>3</sup> )	Gravel (%)	Sand (%)	Mud (%)	Textural group	Mean	Verbal Mean	Sorting	Verbal Sorting	
Buried glacier ice	1c	330-425	La, Su, Cr	74.4	-218.0 ± 10.9	-27.7 ± 2.1	3.7 ± 8.8	34	0.66 ± 0.1	0.39 ± 5.2	25.0	54.3	20.7	Gravelly muddy sand	382.1	Medium sand	8.4	Very poorly sorted
	1b	425-460	Su, Cr	47.9	202	-26.3	8.4	2	0.66 ± 0.1	0.29 ± 5.0	29.7	49.9	20.5	Gravelly muddy sand	528.1	Coarse sand	11.8	Very poorly sorted
	1a	460-470	Ri, Cr	16.0	-196.9	-25.6	8	1	0.65 ± 0.1	0.19 ± 0.42	25.8	57.7	16.4	Gravelly muddy sand	490.3	Medium sand	8.0	Very poorly sorted
Basal glacier ice (C-93)	-	-	La, Su, Cr	53.8	-173.8 ± 10.4	-22.2 ± 1.9	4.1 ± 5.5	7	0.71 ± 0.1	0.29 ± 16.4	29.15	61.6	9.28	Sandy gravel	526.8	Coarse sand	9.6	Very poorly sorted

**Table S3.** Stable isotope ( $d18O$ ,  $dD$  and D-excess) mean, standard deviation, minimum and maximum values, as well as slopes and linear regression coefficients of the  $dD$ - $d18O$  relation for the different ice types sampled.

Sites	$\delta^{18}O$ (‰ V-SMOW)				$\delta D$ (‰ V-SMOW)				D-excess				Slope	Intersect	R <sup>2</sup>	n
	mean	std.dev	min	max	mean	std.dev	min	max	mean	std.dev	min	max				
Buried basal glacier ice	-27.7	2.1	-30.0	-22.5	-218.0	10.9	-231.8	-196.9	3.7	8.8	-21.5	9.8	4.6	-91.4	0.79	34
Intrasedimental ice	-20.3	0.8	-22.6	-19.0	-155.1	11.0	-177.9	-142.5	6.8	7.1	-9.8	11.9	11.6	79.1	0.65	22
Modern basal glacier ice (C-93)	-22.2	1.9	-25.7	-20.4	-173.8	10.4	-193.4	-162.4	4.1	5.4	-1.3	12.5	5.5	-52.3	0.95	7
Snow	-30.4	1.8	-34.0	-27.5	-234.4	12.1	-257.6	-212.6	8.9	3.4	3.0	14.2	6.6	-33.0	0.96	13
Ice wedge (Horizontal transect)	-24.9	1.1	-26.7	-22.0	-192.9	6.3	-205.0	-176.5	6.4	4.7	-2.7	19.8	4.9	-70.0	0.69	38

## **Chapitre 4**

# **Contrasted geomorphological and limnological properties of thermokarst lakes formed in buried glacier ice and ice-wedge polygon terrain**

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## 4.1 Abstract

In formerly glaciated permafrost regions, extensive areas are still underlain by a considerable amount of glacier ice buried by glacial sediments. It is expected that large parts of glacier ice buried in the permafrost will melt in the near future, although the intensity and timing will depend on local terrain conditions and the magnitude and rate of future climate trends in different Arctic regions. The impact of these ice bodies on landscape evolution remains uncertain since the extent and volume of undisturbed relict glacier ice are unknown. These remnants of glacier ice buried and preserved in the permafrost contribute to the high spatial variability in ground ice condition of these landscapes, leading to the formation of lakes with diverse origin, morphometric and limnological properties. This study focuses on thermokarst lake initiation and development in response to varying ground ice conditions in a glacial valley, on Bylot Island (Nunavut). We studied a lake-rich area using lake-sediment cores, detailed bathymetric data, remotely sensed data and observations of buried glacier ice exposures. Our results suggest that initiation of thermokarst lakes in the valley was either triggered from the melting of buried glacier ice or intrasedimental ice and ice wedges. Over time, all lakes enlarged through thermal and mechanical shoreline erosion, as well as vertically through thaw consolidation and subsidence. Some of them coalesced with neighbouring water bodies to develop larger lakes. These glacial thermokarst lakes formed in buried glacier ice now evolve as “classic” thermokarst lakes that expand in area and volume as a result of the melting of intrasedimental ground ice in the surrounding material and the underlying glaciofluvial and till material. It is expected that the deepening of thaw bulbs (taliks) and the enlargement of Arctic lakes in response to global warming will reach undisturbed buried glacier ice where it is still present, which in turn will substantially alter lake bathymetry, geochemistry and greenhouse gas emissions from Arctic lowlands.

## 4.2 Introduction

Arctic landscapes underlain by ice-rich permafrost are highly vulnerable to climate change and permafrost degradation (Segal et al., 2016; Rudy et al., 2017; Lewkowitz and Way, 2019; Nitzbon et al., 2020; Douglas et al., 2021). These ice-rich permafrost landscapes are experiencing thermokarst, through the thawing of near-surface ice-rich permafrost and/or the melting of ice wedges or massive ice, which may result in land subsidence and ponding (Kokelj and Jorgenson, 2013; Farquharson et al., 2019; Liljedahl et al., 2016; Edwards et al., 2016; Jorgenson and Osterkamp, 2005). In flat-lying terrains, thermokarst processes often result in the formation of numerous wetlands, ponds and lakes. This creates or modifies existing 'limnoscapes' (lake-rich landscapes) through thermal and mechanical erosional processes as well as thaw consolidation and subsidence beneath waterbodies (Bouchard et al., 2020; Grosse et al., 2013; Shur et al., 2012; Plug and West, 2009). The formation and growth of these lacustrine ecosystems have important effects on the hydrology, ecology, biogeochemistry and geomorphology of affected landscapes (Vonk et al., 2015). Shoreline erosion may affect key biogeochemical processes within these lakes, such as the burial of organic matter in sediments, and its degradation and release as greenhouse gases (GHG; CO<sup>2</sup> and CH<sub>4</sub>) to the atmosphere (Matveev et al., 2016; Vonk et al., 2015; Heslop et al., 2020). For example, the synthesis study by Wik et al. (2016) showed that lakes and ponds north of ~50°N are large methane emitters (notably glacial/postglacial lakes due to their larger areal extent), equivalent to roughly two-thirds of the inverse model calculation of all natural methane sources in the region. Lake basin morphology also influence GHG flux patterns during the open-water season by affecting the mixing regime (Hughes-Allen et al., 2021; Prėskienis et al., 2021).

The extent to which permafrost degradation occurs is dependent on the distribution and abundance of ground ice. In formerly glaciated permafrost regions, extensive areas still contain a considerable amount of glacial ice buried beneath glacial sediments (Belova, 2015; Coulombe et al., 2019; French and Harry, 1990; Ingólfsson and Lokrantz, 2003; Kanevskiy et al., 2013; Swanger, 2017; Dyke and Savelle, 2000; Lakeman and England, 2012). Remnants of buried glacier ice remain stable as long as the ground temperature is below freezing, and the active layer



thickness (i.e. depth of annual thawing) does not exceed the depth to the massive ice body (Shur, 1988). Ice-cored moraines landscapes may lose their buried ice cores thousands to millions of years after the major glacial retreat (Bibby et al., 2016; Coulombe et al., 2019; Lacelle et al., 2007; Swanger, 2017). The persistence of thick beds of buried Pleistocene glacier ice in contemporary permafrost environments indicates that deglaciation is still incomplete (Astakhov and Isayeva, 1988; Everest and Bradwell, 2003; Kaplanskaya and Tarnogradskiy, 1986; Lenz et al., 2013). The broad distribution and the substantial amount of ground ice in glaciated permafrost landscapes make it highly vulnerable to disturbances, such as thermokarst, under the ongoing climate warming (Kokelj et al., 2017a; Segal et al., 2016). As such, some of these landscapes are now entering a second phase of landscape evolution (Astakhov and Isayeva, 1988; Everest and Bradwell, 2003). For example, on hillslopes, the thawing of permafrost terrain underlain by remnants of glacial ice triggered mass wasting processes, such as retrogressive thaw slump and active layer detachment slides (Rudy et al., 2017; Kokelj et al., 2017a). In flat or very gently sloping terrain, formation and evolution of ponds and lakes are typically associated with the melting of intrasedimental ice, such as ice wedges and segregation ice (Grosse et al., 2013; Bouchard et al., 2017). These lakes tend to be shallow, with deeper central pools (~ 2–5 m) and shallow littoral shelves (~ 1 m), or shallow flat-bottomed basins (Grosse et al., 2013; Hinkel et al., 2012; Bouchard et al., 2020; Jorgenson and Osterkamp, 2005). It is generally recognised that numerous Arctic lakes were formed during deglaciation in depressions left by in-situ melting of stagnant blocks of glacier ice (also named kettle lakes or postglacial lakes). However, very few studies have linked lake inception to the thawing of sediments containing glacier ice that had been buried and preserved in permafrost for decades to millennia (Henriksen et al., 2003; Worsley, 1999; Astakhov and Isayeva, 1988). As a result, there is little information on the spatial distribution and abundance and evolution on these glacial thermokarst lakes in modern paraglacial and periglacial environments.

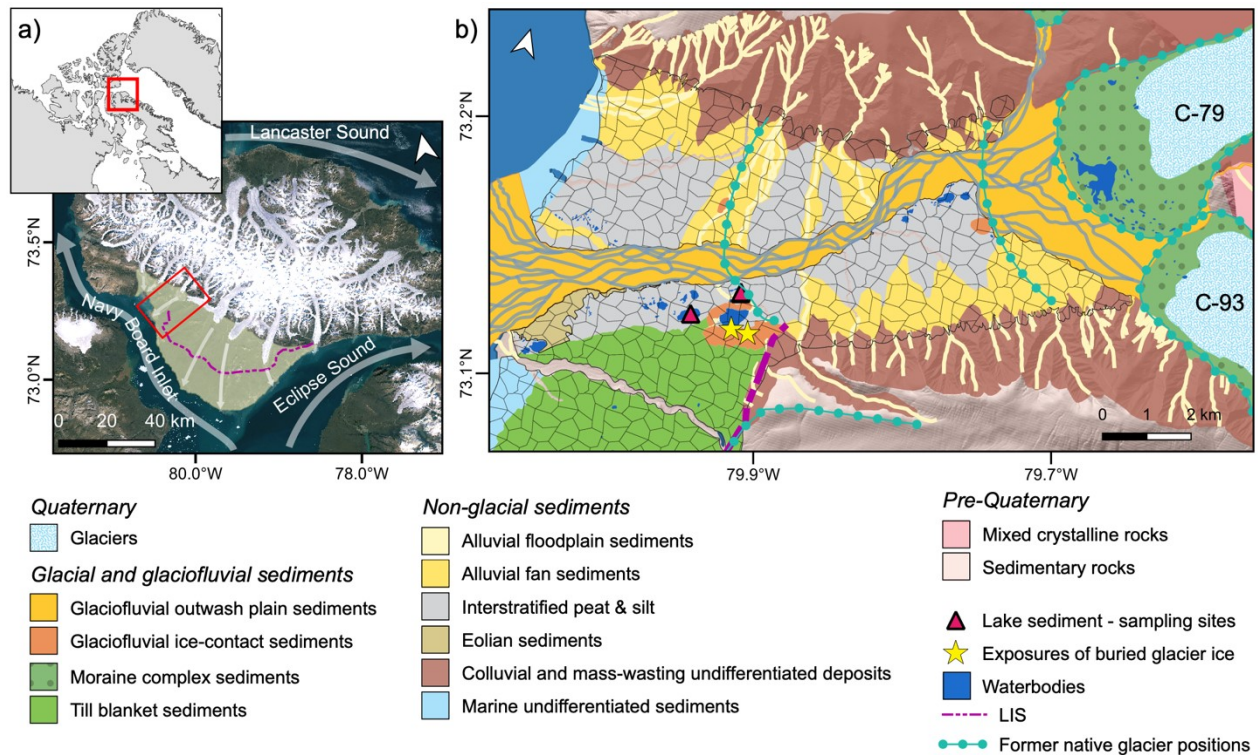
The Quaternary geology of the eastern Canadian Arctic records several glaciations by ice sheets and local mountain glaciers, which means that the landscape stores vast amounts of buried glacial ice, and there is potential for significant postglacial landscape change associated with the

ablation of this buried ice. The resulting landscape can be covered with a large number of thermokarst lakes of diverse origin that impact their physical and limnological properties. This study builds on the findings of Coulombe et al. (2019) conducted on Bylot Island (Nunavut), where blocks of stagnant ice became separated from an ice stream flowing from the Foxe Dome of the Laurentide Ice Sheet and subsequently buried by aggradation of glaciofluvial sands and gravels at the margins of the receding glacier. Subsequent neoglacial cooling resulted in widespread permafrost aggradation and preservation of this glacial ice. Here, we investigate the inception and evolution of twenty-one lakes from the lower reach of a glacial valley on Bylot Island, which presents heterogeneous permafrost ground ice conditions. We hypothesised that thermokarst lakes have different origins and exhibit differences in their morphological and limnological conditions as well as future sensitivity to change. In the Qarlikturvik Valley, remnants of buried glacier ice in lowlands slowly melted during the Holocene, which created deep depressions that formed glacial thermokarst lakes, while the thawing of an ice- and organic-rich polygonal terrace created shallow thermokarst lakes. The specific objectives were therefore (1) to compare the morphological and limnological properties of these two types of thermokarst lakes; (2) to examine the link between the spatial pattern of lakes and past glacier positions in the Qarlikturvik Valley and broader southern plain, and (3) to develop a conceptual model of lake inception and evolution, with a focus on lakes formed by the delayed melting of buried glacier ice.

### **4.3 Study area**

The study area is in the Qarlikturvik Valley (73°09' N, 79°57' W) on the southwest plain of Bylot Island, in the Canadian Arctic Archipelago (Figure 27a). The landscape was glaciated on several occasions in the late Quaternary by both local mountain glaciers and the Laurentide Ice Sheet (LIS; Klassen, 1993). The study area was most likely a confluence zone between LIS ice and local alpine glaciers with glacier ice flowing out of major valleys but Laurentide ice flowing into the southern plain and up the valleys (Dyke and Hooper, 2001; Lacelle et al., 2018). The maximum extent of the LIS is outlined by the Eclipse moraine, a major moraine system across the outer coastal mountains of Bylot Island and parts of adjacent Baffin Island (Klassen and Fisher, 1988). Today, Bylot Island remains 40% glacierized as numerous valleys and piedmont glaciers still flow

from the local ice cap and terminate in lowlands underlain by sedimentary rock of Cretaceous-Tertiary ages (Dowdeswell et al., 2007). The Qarlikturvik Valley is one of the many U-shaped glacial valleys with ice-rich sediments dating back to the Late Pleistocene and Holocene, which are highly susceptible to thermokarst (Fortier and Allard, 2004; Bouchard et al., 2020). The valley contains abundant and diverse water bodies, including a proglacial river, lakes, trough and polygon ponds, small streams, and thermos-erosion gullies (Godin et al., 2014; Muster et al., 2017; Prèskienis et al., 2021). With glaciers ending within the continuous permafrost zone, this lake-rich valley represents a typical glaciated valley geosystem that incorporates numerous depositional environments associated with ice-marginal, proglacial, paraglacial and periglacial processes, which makes it an ideal location to study ice types and thermokarst lake development under varying ground ice and terrain conditions. In the Qarlikturvik Valley, mounds of reworked till and ice-contact stratified sediments mark former positions of the glacier margins (Figure 27b). The earliest postglacial radiocarbon date from marine shells retrieved from marine clays is 11,331 cal yr BP (IntCal20), suggesting that the valley was partially ice-free by this time (Allard, 1996). About 2–3 meters of ice-rich Quaternary silt and sand derived from aeolian deposition, interstratified with peat, overlies ice-poor glaciofluvial outwash deposits (Figure 27b; Fortier and Allard, 2004). Syngenetic ice wedge growth has created extensive polygonal patterned ground. Thermokarst is an active landscape change mechanism operating in the valley, as demonstrated by the abundance of lakes, thermo-erosional gullies and thaw slumps within the study area (Bouchard et al., 2020; Godin et al., 2014; Fortier et al., 2007). Previous work in the area have examined various aspects of thermokarst lake dynamics such as GHG exchanges (Bouchard et al., 2015; Prèskienis et al., 2021), photochemical and microbial decomposition of organic matter (Laurion et al., 2021), microbial diversity (Negandhi et al., 2014), nutrient inputs from the goose colony (Côté et al., 2010) and methylmercury (MacMillan et al., 2015), as well as lake development in syngenetic ice-wedge polygon terrain (Bouchard et al., 2020).



**Figure 27.** a) Location of Bylot Island, Nunavut, Canada and the study area in the Qarlikturvik valley (background: NRCan Landsat-7 orthorectified mosaic, 3 August 2010). The shaded area shows the southwestern plain of Bylot Island, b) Surficial geology of the valley and location of the sampling sites. The net pattern represents the polygonal patterned ground. The white arrows show the direction of paleo ice flow within and around Bylot (Margold et al., 2015). The purple dashed line indicates the glacial limit attained during Eclipse glaciation based on the inland extend of Eclipse glacial drift and moraine ridges.

The mean annual air temperature at Pond Inlet for the 1981–2010 normal is  $-14.6^{\circ}\text{C}$ , which is  $0.5^{\circ}\text{C}$  higher than the previous 1971–2000 record (Environment Canada, 2021). The mean annual precipitation for the 1981–2010 period was  $189\text{ mm yr}^{-1}$ , with rainfall representing 91 mm. Bylot Island is located within the continuous permafrost zone. Permafrost thickness was estimated to be at least 200–400 m based on shallow ground temperature measurements on the island (Moorman, 2003). On average, the active-layer thickness varies between 0.3 and 0.7 m in peaty and silty soils, to more than 1 m in drained unvegetated sands and gravels (Allard et al., 2020). Thawing and freezing indices averaged (1981–2010 period) 473 degree-days above  $0^{\circ}\text{C}$  and 5736 degree-days below  $0^{\circ}\text{C}$ , respectively (Environment Canada, 2021).

## 4.4 Material and methods

Two spatial scales were used to investigate the role of buried glacier ice in the formation and evolution of thermokarst lakes. First, we focused on the Qarlikturvik Valley ( $\sim 75 \text{ km}^2$ ), where buried Pleistocene glacier ice has been found in permafrost (Fig. 23; Coulombe et al., 2019). We examined the morphology and conducted bathymetric surveys of 21 lakes and analysed lake sediment cores from two of these lakes to infer probable lake origin. We also analysed water column profiles of temperature and dissolved oxygen of these same lakes. The studied lakes are among the largest in the valley and most of them are close to former glacier positions. Then we examined the spatial distribution of lakes on the broader coastal plain of Bylot Island ( $\sim 122 \text{ km}^2$ ) to link the extension of former local and regional glaciations to lake distribution.

### 4.4.1 Landforms, surficial deposits and lake mapping

We used contemporary high-resolution GeoEye satellite imagery (2010, pixel = 0.5 m), WorldView-1 (2010, pixel = 0.5 m) and ArcticDEM data (pixel = 2 m) to map lakes, Quaternary surficial deposits and landforms in the Qarlikturvik Valley. The surficial geology map is based on field work observations, borehole data, interpretation of satellite imagery and digital elevation model and previous maps from Klassen (1993) and Allard (1996). We used field- and remote-based data to map glacier frontal positions of glacier C-79 and C-93 to investigate the formation of new lakes in the valley at the termini of these glaciers over the past 60 years: 1) historical aerial photos (1961, 1982; National Air Photo Library) 2) GeoEye satellite imagery (2010, pixel = 0.5 m); 3) Sentinel-2 (2016, 2020, pixel = 10 m) and 4) field measurements using a real-time kinematic (RTK) global positioning system (July 2011; Trimble R8). The positions refer to the contact between the ice and moraine material. A Sentinel-2 image mosaic (2016, pixel = 10 m) of the southern plain of Bylot Island served as the basis for mapping the water bodies outside the valley (Copernicus, 2016). We also used the Google Earth Engine Timelapse dataset (2000-2019) to visually assess terrain change and sediment accumulation at the glacier terminus based on Tasseled cap (TC) trend analysis of a Landsat image stack (Fraser et al., 2012; Gorelick et al., 2017; Nitze and Grosse, 2016). The tasseled cap transformation reduces the Landsat reflectance bands to three orthogonal indices called brightness, greenness and wetness (Crist and Cicone, 1984). Data

processing and analyses were performed using QGIS (v.3.16; QGIS Development Team, 2021). To extract all water bodies, we used the reflectance properties of water in the Green and NIR bands (McFeeters, 1996). Because water bodies have high 'Normalized Difference Water Index' (NDWI) values, a simple thresholding technique was used to isolate most water features. Lake shorelines were extracted as vector data and converted to polygon topology. Lakes smaller than 1000 m<sup>2</sup> were automatically removed from the analysis to exclude polygon ponds and collapsed ice-wedge troughs filled with water.

#### ***4.4.2 Distribution of lakes in the valley and the southern plain of Bylot Island***

We examined the spatial distribution of lakes to examine possible association with past glaciers positions in the Qarlikturvik Valley and the broader southern plain on the island. This can provide additional evidence on the glacial origin of lakes because these ice-marginal zones often comprise discrete bodies of glacier ice left behind by a retreating glacier and buried underneath sediment. To map the density of lakes in the Qarlikturvik Valley and the broader southern plain of Bylot Island, a kernel density estimation was performed using the 'spatstat' package in R (v. 3.5.3; Baddeley et al., 2019; R Development Core Team, 2021). Input for kernel density came from lake centroids obtained from the vector polygon, which were calculated automatically in R as the geometric center of the lakes. We defined the extent as all areas of the Qarlikturvik Valley and the broader southern coastal plain of Bylot Island, excluding the bedrock outcrops, slopes ( $> 5^\circ$ ), glaciers and outwash plains. To analyse lake spatial patterns, we also performed a clustering analysis using the inhomogeneous pairwise correlation function with 100 Monte Carlo simulations and 95% confidence interval, which accounts for spatial inhomogeneity in lake locations (quadra test;  $p = 0.001$ ). This function considers the intensity (density) of the observed points by simulating completely spatially random point patterns based on the average intensity in the observed point pattern. This technique allows distinguishing between dispersed ( $R > 1$ ) and clustered ( $R < 1$ ) spatial patterns by comparing the observed point patterns against the expectation for a randomly distributed sample population (CSR model), which assumes that the objects can be distributed anywhere in the region of interest. A high spatial clustering suggests

that the spatial distribution of lakes is dependent on an external variable which we interpreted as the probable presence of patches of buried glacier ice.

#### **4.4.3 Lake morphology in the Qarlikturvik Valley**

Detailed bathymetric data were collected for 21 lakes across the valley using a Humminbird 859XD Sonar with a built-in global positioning system. Lake bathymetric surveys were conducted with an inflatable boat when the lake was free of ice (June to August). Geographic location and water depth were recorded each second along transect lines that were spaced at approximately 5 to 25 m (depending on lake size) to entirely cover the lake. Some uncontrolled conditions have degraded the accuracy of the survey, such as the presence of littoral vegetation and waves, especially in shallow lakes. Depth and location data were imported into QGIS for visualization and additional processing. Initial processing included the removal of spurious data points (outliers) such as single-point depths located substantially above or below the general depth of lake-bottom. We used a spline algorithm to generate an interpolated surface from the individual depths. Ground penetrating radar (GPR) surveys with 50 Hz antenna were conducted across frozen lakes to investigate the lake-bottom morphology (see Supplementary material S1 for further details). For each lake, we also calculated the area, perimeter, elongation ratio (long axis/short axis), and shoreline development or  $D_L$  from the digitised shoreline polygons to compare lake metrics and determine if they can be used to discriminate between thermokarst lakes formed in ice-wedge polygon terrain and thermokarst lakes formed by the melting of buried glacier ice. For comparison, the morphological attributes of glacial thermokarst lakes formed in proglacial outwash deposits in front of glaciers C-79, C-93 and C-67 were also calculated. Very few studies have examined glacial thermokarst lake morphology, but studies on kettle lakes report enclosed and steep-sided depressions, roughly circular and inverse-conical (Fay, 2002; Gorokhovich et al., 2009; Borsellino et al., 2017). The shoreline development ratio ( $D_L$ ) is a standard measure of the complexity of the shoreline, which is the ratio of the length of the shoreline of a lake (i.e. perimeter) to the circumference of a circle of area equal to that of the lake (Equation 1; Hutchinson, 1957).

$$D_L = \frac{Perimeter}{2\sqrt{Area*\pi}} \quad (1)$$

$D_L$  for a perfect circle is 1.0, and its value increases ( $\gg 1$ ) as the shape of the lake surface deviates from that of a circle, indicating the shoreline is more dendritic or irregular. Glacial thermokarst lakes should have low complexity values ( $\sim 1$ ) whereas thermokarst lakes expanding laterally in ice-wedge polygon terrain should be more irregular and have values  $> 1$ . A highly indented shoreline may also indicate coalescent lakes formed by shoreline expansion. An elongation ratio (ER) of 1 indicates a circular object with increasing ER values for more elongate forms. Correlation between shoreline morphology variables and basin morphometry (maximum depth) were tested using the non-parametric Kendall tau rank correlation for non-normally distributed data. All statistical tests were run in the open-source software R (QGIS Development Team, 2021).

#### ***4.4.4 Stratigraphic profiles of lake bottom sediments***

We selected two nearby lakes (IWT1 and GT1) exhibiting different morphometry to compare the stratigraphic profiles of lake bottom sediment. According to the bathymetric surveys, lake GT1 is the deepest in the valley (max. depth = 12.2 m), and it lies directly next to an ice-contact deposits mound. We also sampled lake IWT1 (max. depth = 4.1 m) as lake bottom imagery revealed submerged ice-wedge polygons ( $\sim 1$  m depth) and degraded ice-wedge troughs, which confirmed that this lake is evolving through the melting of permafrost intrasedimental ice and ice wedges (see video supplement in Bouchard et al., 2020). Two sediment cores of 109 cm and 114 cm were collected in spring 2015 from lakes IWT1 and GT1, respectively, through a 2-m thick ice cover using a 7-cm diameter handheld percussion corer (Aquatic Research Instruments), sealed, and returned to the laboratory where they were stored in the dark at 4°C. Coring occasionally caused minor deformations to the sediments owing to friction and pressure along coring tubes. Both cores were observed under X-ray computed tomography (CT), allowing to visualize and reconstruct the internal structure (2D and 3D) of the cores. Details on the CT scanning procedure are provided in supplementary material S6. Facies were identified based on visual inspection and physical properties, including sedimentary structures, grain size, colour, and density. Percentage dry weight was determined for all samples (drying overnight at 105°C). Organic matter content was determined by weight loss (loss-on-ignition, LOI), following a combustion of dried samples at 550°C for 4 h (Heiri et al., 2001). Sediment grain size was



measured in triplicates using a Malvern Mastersizer 2000 and Hydro2000G liquid handling unit. Bulk sediment and fossil plant fragments were radiocarbon-dated by accelerator mass spectrometry (AMS) at Keck Carbon Cycle AMS Facility (University of California, Irvine, CA, USA). Calibrated ages (cal yr BP) were calculated using “CALIB 8.2” (IntCal20 dataset; Stuiver et al., 2021; Reimer et al., 2020). In the case of lake IWT1, facies are described in more details based on other proxies, such as organic content and fossil diatoms (Bouchard et al., 2020; see Figs. 4 and 5). Finally, diatom assemblages were investigated in sediment sections from lake GT1 for comparison purposes with those of lake IWT1, presented in Bouchard et al., (2020). Sediment samples (n=16) were prepared for diatom taxonomic identification using standard procedures in the Aquatic Paleoecology Laboratory (Laval University, Canada). The cleaned diatom samples were dried on glass cover slips and mounted in Naphrax® mounting medium. Diatoms were identified and counted at 1000X magnification using a Leica DMRB microscope.

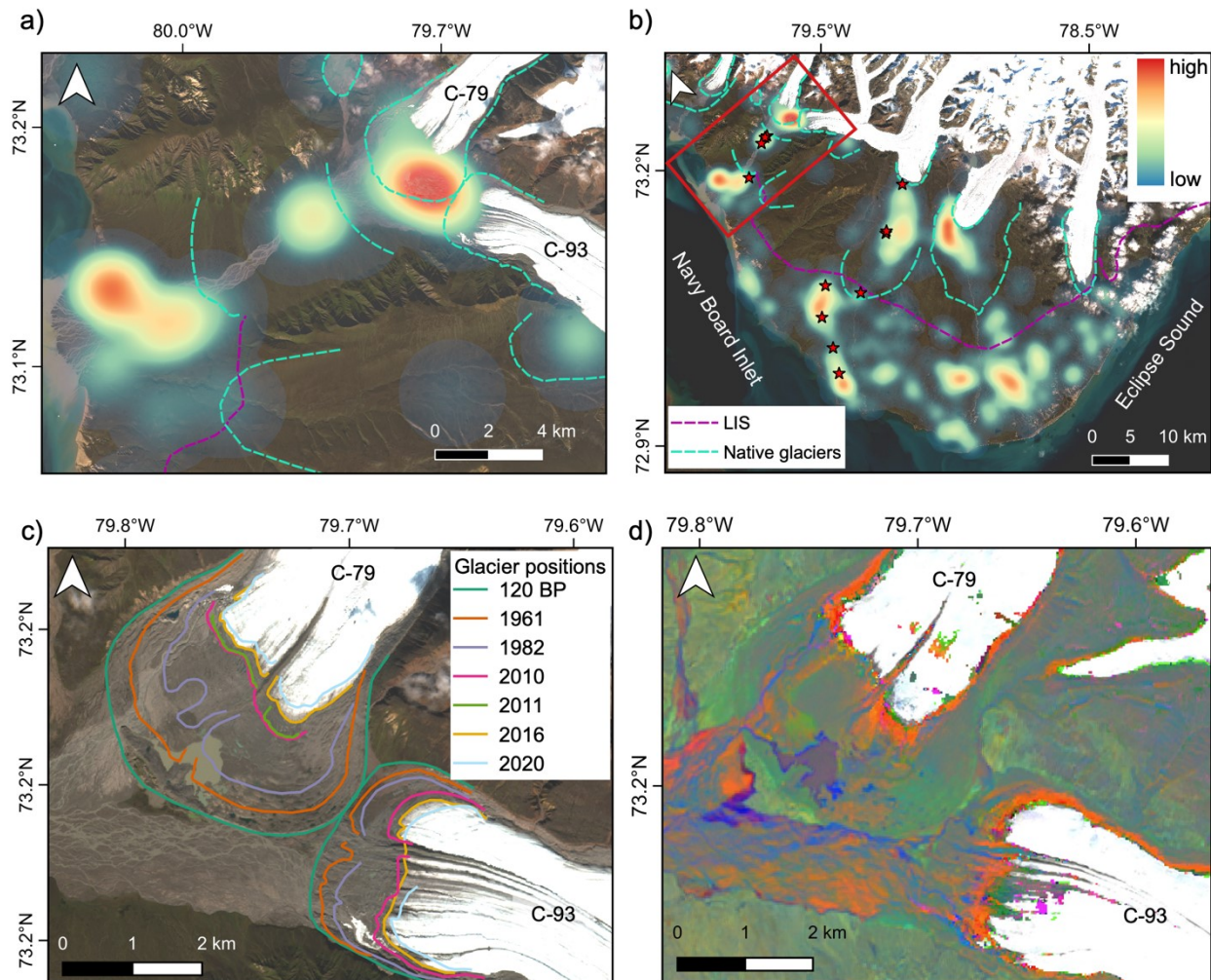
#### ***4.4.5 Water column profiles of temperature and dissolved oxygen***

We profiled the water column of lakes IWT1, GT1, and lake GT2 in late winter under the ice cover (early June 2015) and during the ice-free period (July and August 2015) to examine differences in water temperature and dissolved oxygen (DO) between lake types. Discrete profiles were measured manually from lakes IWT1 and GT1 with a ProODO profiler (YSI Inc.), while submersible temperature loggers (Vemco Minilog-II-T installed at 2, 4, 6, 8 and 10 m depth) and DO loggers (PME MiniDOT; 2 and 9 m depth) were installed in lake GT2. The loggers recorded annual cycles of stratification (1h frequency), from which profiles were selected to match the discrete profiles obtained from the other two lakes. Sensor specifications can be found in Prėskienis et al. (2021).

## 4.5 Results

### *4.5.1 Distribution of lakes in the valley and southern plain of Bylot Island*

Using remote sensing classification, we detected 845 lakes larger than 1000 m<sup>2</sup> within the study area (total lake area reaching 14 km<sup>2</sup> over ~ 1700 km<sup>2</sup>, or 0.8% of the area), of which 189 lakes (totalling 1.6 km<sup>2</sup>) are in the Qarlikturvik Valley (~ 122 km<sup>2</sup>; 1.3% of the area). The spatial distribution of the lakes showed a significant aggregation pattern in both Qarlikturvik Valley and the southern coastal plain of Bylot Island (Figure 28). Patterns of distribution emerge in the valley with higher densities, 10 to 85 lakes per km<sup>2</sup>, detected nearby mounds of ice-contact deposits or in areas of unvegetated moraine in front of glaciers C-79 and C-93 (Figure 28a). A third group of lakes is also observed on the plateau bordering glacier C-93. According to the point pattern analyses, the lakes in the Qarlikturvik Valley show significant clustering in short distance (far above the 95% confidence envelope;  $r < 0.85$  km) and a regular distribution further away ( $r > 0.85$  km; Figure S8). On the southern plain of Bylot Island, the highest densities occur directly in front of contemporary glaciers and within the extent of local mountain glaciations and LIS, with up to 40 lakes per km<sup>2</sup> (Figure 28b). The observed points (lake centroids) show considerable clustering at smaller distances (< 3.3 km) but show regularity beyond ~ 4 km (Fig. S7). Because the highest densities were observed in association with past glacier and ice sheet margins (LIS), we also analysed the formation of lakes in front of glaciers C-93 and C-79 since their last major advances during the Little Ice Age (LIA;  $120 \pm 80$  <sup>14</sup>C yr BP; Klassen, 1993). From LIA to 2020, 383 new glacial thermokarst lakes developed after of glacier retreat (~2 km; Figure 28c). In addition, the TC trend analysis revealed sediment accumulation at the front of the receding glaciers between 2000 and 2019 as represented by red colours (drier and unvegetated areas) on TC images (Figure 28d). This shows the active burial of glacier ice at the front of glaciers C93 and C79, thereby providing a modern analogue for the past burial of ice when the glacier was several km further down-valley.



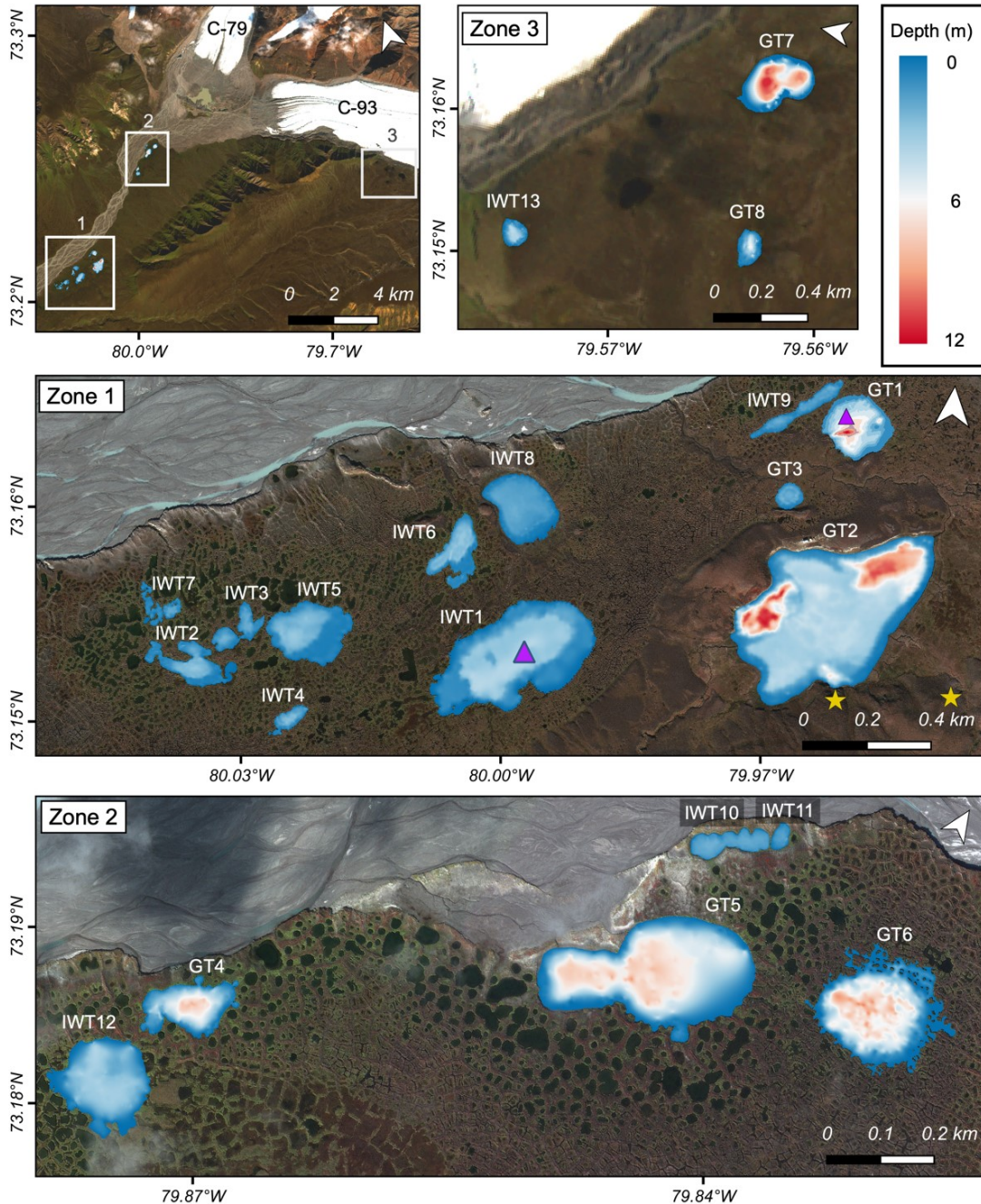
**Figure 28.** a) Spatial point density of lake locations in the Qarlikturvik Valley, b) Spatial point density of lakes in the southern plain of Bylot Island. The blue dashed line shows former limits of local mountain glaciations. The purple dashed line shows the limit of the Laurentide Ice Sheet (LIS) as defined by Klassen (1993). The red stars indicate the locations of deep lakes (between 5 m and 21 m) studied in Coté et Pienitz (2010). (Background: Sentinel-2 (ESA) image courtesy of the Copernicus Open Access Hub), c) Glacier terminus positions from 120 yr BP to present and d) Tasseled Cap transformation images obtained for Google Earth Engine (Gorelick et al., 2017) The accumulation and movement of sediments in the outwash plain and at the glacier front are represented by red and orange colours on the images (dry and unvegetated areas; TC brightness). Wetter areas, such as eroding cliff, lake shore or river channel are shown in blue (high TC wetness). Vegetated areas are distinguished by teal and yellow colours (TC greenness).

#### **4.5.2 Lake morphology in the Qarlikturvik Valley**

In the valley, we identified two groups of lakes according to their depth range and lake-floor morphometry (Figure 29). Table 5 summarizes the characteristics of the lakes ( $n=21$ ) for which bathymetric data were collected in 2015. The first group of lakes (GT1 to GT8; deep) stands out by their greater depths and sizes, and in some cases, the presence of multiple sub-basins. The maximum measured depths recorded in these lakes range from 5.9 to 15.4 m. Most of these lakes are characterised by a relatively deep central lake basin surrounded by shallower areas, ranging between 0.5 and 1.0 m (mean depth =  $0.6 \pm 0.4$  m; Table 5). Three lakes (GT2, GT5, GT7) have two or three steep-sided and confined sub-basins that are surrounded by a relatively shallow marginal platform. The GPR profiles indicated that these deeper lakes usually have smoother microtopography at the lake bottoms, whereas lake GT2 also exhibits an irregular lake floor micromorphology in the shallowest areas (Figure S7), with submerged polygon-patterned ground and degraded ice wedge under frost crack troughs (Bouchard et al., 2020). The bathymetric map of lake GT2 also revealed a deeper depression that is aligned with a lakeside thaw slump exposing buried glacier ice (Figures 29 and 30). The bottoms of lakes GT1 and GT2 are, respectively, 5.5 m and 5 m below the current sea level. In the valley (zones 1 and 2), the deeper lakes are located near mounds of stratified ice-contact glaciofluvial deposits. The second group of lakes (IWT1 to IWT13; shallow) showed markedly different characteristics (Figure 29). At the lake scale, these shallow water bodies have relatively flat and homogeneous beds with a deeper central basin surrounded by shallower nearshore zones ( $< 2$  m deep). The lake floor is irregular at a finer scale (microtopography), which is attributed to submerged polygons (see the video supplement in Bouchard et al., 2020). These lakes have maximum depths ranging between 1 and 4 m, with mean depth reaching  $1.4 \pm 0.7$  m across their platforms (Figure 29). They are characterised by irregular shorelines, which generally follow the deep troughs caused by the melting of ice wedges from their tops. Despite the above-stated differences, lakes from both subgroups present similar shapes and shoreline characteristics as their morphometric properties (area, perimeter, elongation ratio, complexity) were not significantly different (Mann–Whitney–Wilcoxon test,  $p > 0.05$ ; Figure S9). In addition, mean or maximum depths did not show any significant correlation with the other morphometric variables ( $p > 0.1$ ). In addition, glacial thermokarst lakes ( $n=490$ )



located near the front of glaciers C-79, C-93 and C-67 have an average shoreline development index of  $0.9 \pm 0.1$  and average elongation ratio of  $1.7 \pm 0.5$ , indicating the shorelines are relatively regular and are mostly oval-shaped.



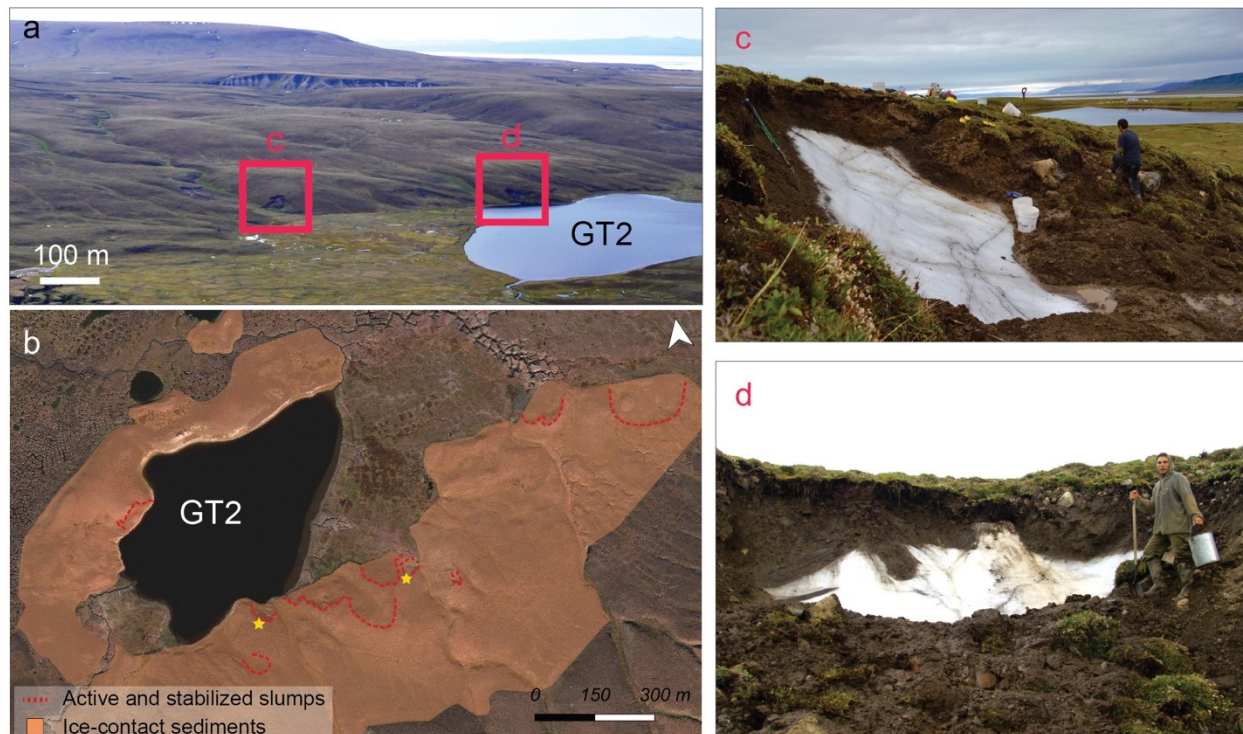
**Figure 29.** Bathymetric maps of the 21 lakes surveyed in the Qarlikturvik valley (background: GeoEye, 2010). The yellow stars show the location of massive ice exposures. Purple triangles on lakes IWT1 and GT1 indicate sediment coring locations. Glacial thermokarst lakes (max. depth > 4 m): GT 1 to 8; Ice-wedge thermokarst lakes (max. depth < 4 m): IWT1 to 13.

**Tableau 5.** Characteristics of lakes for which bathymetric data were collected in 2015.

Zone	Name	Latitude	Longitude	Surface Elevation (m a.s.l.)	Max. depth (m)	Depth - Basin (m)		Depth - Platform (m)		Area (m <sup>2</sup> )	Perimeter (m)	Shoreline development index	Other names	Elongation ratio	Distance to glacier margin (m)
						Mean	Std. dev.	Mean	Std. dev.						
1	IWT1	73.153	-79.999	8.4	4.1	2.9	0.8	1.3	0.4	10076.6	694.9	2	BYL66	1.2	464
	IWT2	73.152	-80.030	7.1	3.2	2.0	0.6	1.1	0.3	19367.8	1111.2	2.3		1.5	1444
	IWT3	73.153	-80.026	7.0	2.4	1.5	0.4	0.9	0.2	30968.2	709.7	1.1		0.4	1280
	IWT4	73.151	-80.021	8.8	2.8	1.9	0.5	1.2	0.3	115306.2	1689	1.4		1.2	1197
	IWT5	73.153	-80.019	7.0	3.3	2.2	0.6	0.9	0.2	37549.1	926.7	1.3		0.6	1064
	IWT6	73.156	-80.006	7.7	4.3	2.3	0.6	1.7	0.2	18379.7	873.3	1.8		1	560
	IWT7	73.153	-80.034	7.1	2.9	1.5	0.4	1.2	0.3	6470.4	952.6	3.3		0.8	1505
	IWT8	73.157	-80.000	5.7	2.5	1.5	0.3	0.8	0.2	34167	790.5	1.2		0.8	329
	IWT9	73.160	-79.973	8.3	3	1.5	0.3	1.1	0.2	11501.6	790.1	2.1	BYL123	3.8	586
	GT1	73.160	-79.968	7.9	12.2	5.4	1.5	3.6	1.0	5226.1	447.4	1.7	BYL36	0.7	491
	GT2	73.155	-79.969	8.2	11.7	4.7	2.0	2.2	0.6	209426.6	1991.9	1.2	BYL37	1.1	514
	GT3	73.158	-79.974	6.8	6.1	1.8	0.3	1.0	0.2	5264.5	272.9	1.1		0.7	491
	2	GT4	73.185	-79.870	15.1	8.4	4.2	1.8	1.6	0.4	12963.4	695.3	1.7		1.1
GT5		73.190	-79.848	15.1	9.4	4.7	2.1	0.8	0.2	59699.7	1190	1.4		1.1	868
GT6		73.191	-79.836	18.3	9.8	5.0	2.1	2.4	0.3	35566.3	2673.8	4		0.8	441
IWT10		73.192	-79.849	16.9	2.3	1.5	0.5	0.5	0.1	5354.2	403.7	1.6		1.7	768
IWT11		73.193	-79.847	17.0	1.8	1.4	0.3	0.8	0.2	1292.5	139.8	1.1		1	674
IWT12		73.183	-79.872	14.2	3.9	2.6	0.8	0.9	0.2	22894.5	817.8	1.5		0.7	1873
3	GT7	73.144	-79.531	354.0	15.4	6.9	4.0	1.0	0.3	54464.4	944.1	1.1		1	256
	GT8	73.144	-79.554	346.5	5.9	2.6	1.4	0.9	0.3	12754.9	446.4	1.1		0.6	829
	IWT13	73.153	-79.558	359.4	4.3	2.6	1.0	1.2	0.3	9757.7	367.4	1		0.8	380

\*A platform is any relatively flat or gently sloping surface, bounded by a steeper descending slope on the opposite side (towards the center of the lake).

\*\*A basin represents the deeper part of the lake, bounded by the platform



**Figure 30.** a) Aerial view of two exposures of buried glacier ice located nearby glacial thermokarst lake GT2, b) Map showing the distribution of active and stable thaw slumps nearby glacial thermokarst lake GT2 (background: GeoEye, 2010). The yellow stars show the location of massive ice exposures, c) We interpreted the massive ground-ice, exposed at the headwall of thaw slump, as buried glacier ice on the basis of cryostratigraphic, crystallographic and geochemical analyses (Coulombe et al., 2019), d) This massive ice exposure was not studied in detail since the ice had been buried again under a thick cover of slump material. However, the ice displays a very similar appearance to the first exposure located one hundred meters away (Photo courtesy of Denis Sarrazin).

#### **4.5.3 Stratigraphic profiles of lake bottom sediments**

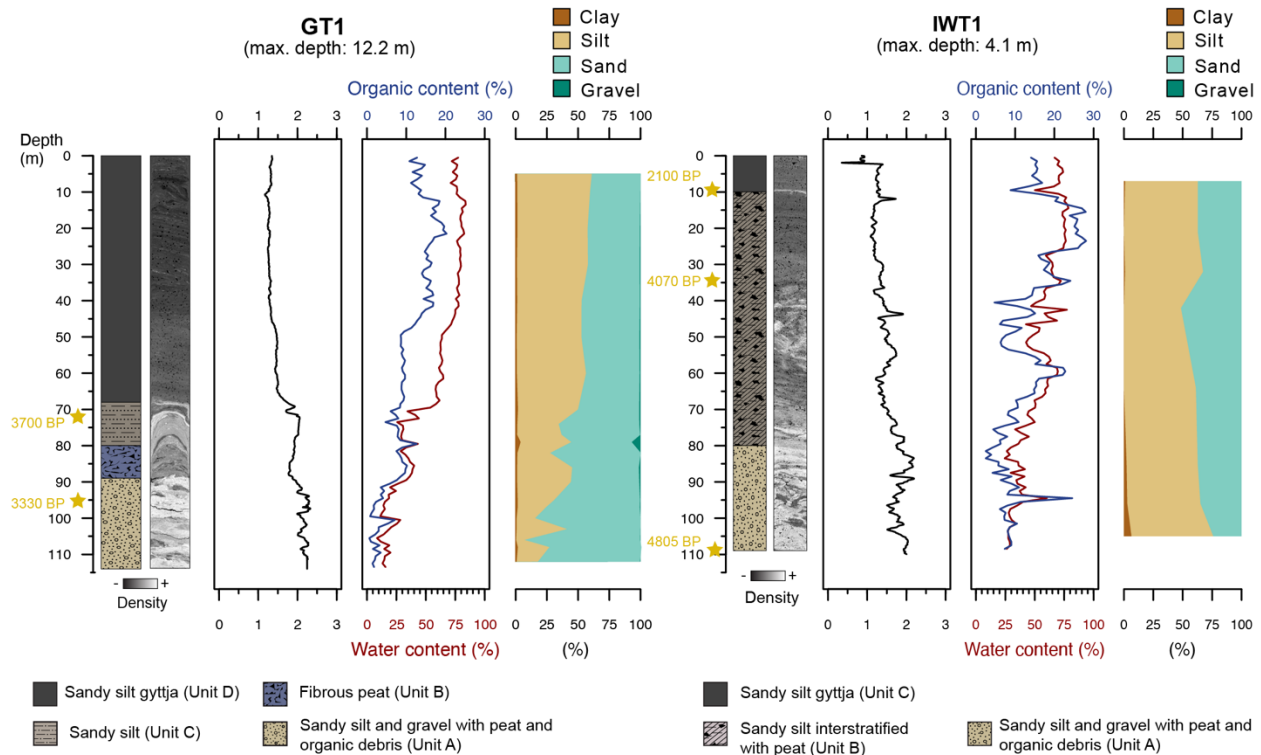
Four distinct lithofacies or units, labelled from core bottom to top, were identified in lake GT1 (max. depth: 12.2 m) based on visual analysis of CT-scan images and field description: A) sandy silt and gravel with interspersed peat/organic debris (114-89 cm); B) sandy silt fibrous peat (89-80 cm); C) laminated to massive sandy silt (80-68 cm); D) sandy silt gyttja (organic lacustrine mud; 68-0 cm; Figure 31). The lower unit (A) is a weakly stratified black and yellowish-brown fine sediments (coarse silt and fine sand) and gravel with scattered organic material, which was dated



near its top (95 cm) at 3531 cal yr BP (3330 <sup>14</sup>C BP; Figure 31). The coring operation did not reach the bottom of unit A, so its total thickness is unknown. Compared to other units, it has a higher mean density (2.0 to 2.5 g cm<sup>-3</sup>), typical of dominantly mineral material. Unit B consists of dry and fibrous organic-rich material (peat) with fine sand and silt. Unit C consists of laminated to massive sandy silt containing very sparse and fine gravels, which is massive in its uppermost 5 cm. This unit also displays sharp lower and upper contacts, and includes some deformation structures, caused by the coring operation (layers bended downward near the coring tube walls). Bulk sediment near the top of this unit (73 cm) has been dated at 4036 cal yr BP (3700 <sup>14</sup>C BP; Figure 31). The uppermost unit (D) is composed of laminated gyttja that grades upwards into soft and loose gyttja. The CT-scan image along with the LOI profile show that the mineral input steadily decreases towards the middle of unit B. Light-coloured thin laminae of silt (0.3 and 0.9 cm) are common in the upper part of the sequence. The upper section of this unit (50–0 cm depth) is less compact compared to deeper sediments (> 50 cm depth) and it has a high water content that becomes dryer towards the bottom of the unit (gravimetric water content decreases from about 80% at the surface to 15% at the bottom; Figure 31). A similar trend is also observed for the organic content as it decreases from ~15% at the surface to near 1% in the lower portions of the core (Figure 31). Among the 16 levels analysed for fossil diatoms, only 8 contained identifiable diatom taxa. These species were restricted to units A and B only, whereas upper units (C, D) contained only dispersed fragments that could not be identified. In the 4 levels ranging from 90–90.5 to 74–74.5 cm (= levels 5 – 8), clastic debris was abundant but also diatoms, The majority of these were fragmented, but there were also a few remarkably well-preserved and intact specimens in the assemblage belonging to the genera *Eunotia* and *Cymbella*. *Eunotia* species are known to be adapted and associated with mosses (bryophytes) in peat and humid environments, sometimes exposed to the air, rich in *Sphagnum* mosses. They indicate circumneutral to acidic, oligotrophic (nutrient-poor) and shallow environments. *Cymbella* species (e.g., *C. cistula*) live as epiphytes on the stems and leaves of freshwater aquatic plants, while all other identified specimens of less abundant genera belonging to *Pinnularia* (e.g., *P. maior*, *P. balfouriana* = aerophilic species), *Caloneis*, *Navicula* (*N. pseudoscutiformis*), *Achnanthes*, *Tabellaria* and *Cocconeis* (*C. placentula* = species which attaches to rocky surfaces) reflect a cold shallow



freshwater environment, circumneutral to slightly acidic, oligotrophic to ultraoligotrophic, and with a rocky/sandy substrates carpeted with mosses and a few aquatic plants. These are exclusively elongate (= pennate) and benthic diatoms living on substrates, and the absence of deep water centric (planktonic) species reflects a very shallow water body with frequently alternating wet and dry conditions.



**Figure 31.** General stratigraphy of cores sampled in lake GT1 (deep glacial thermokarst lake; max. depth = 12.2 m) and IWT1 (shallow ice wedge thermokarst lake, max. depth: 4.1 m).

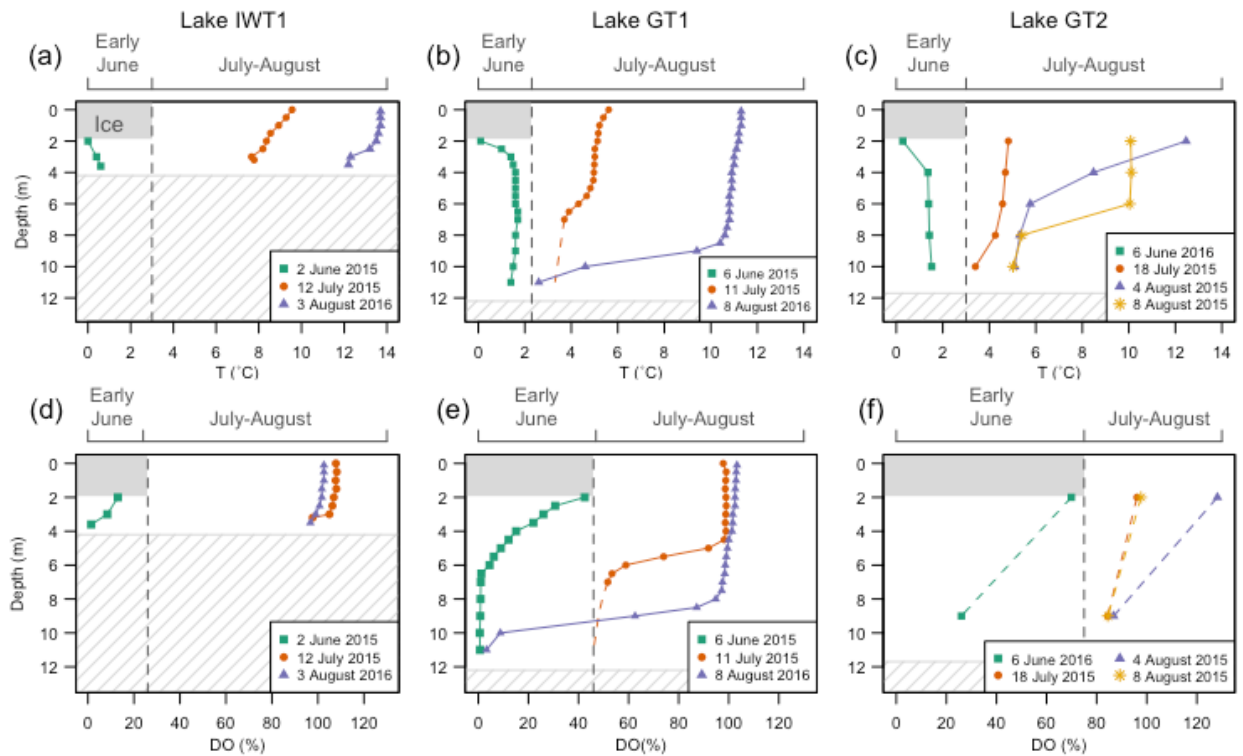
Three lithofacies were identified in lake IWT1 (max. depth: 4.1 m): A) organic-poor sandy silt (109-80 cm); B) organic-rich sandy silt interstratified with peat (80-10 cm); C) sandy silt gyttja (organic lacustrine mud; 10–0 cm; Figure 31). More details on lake sediment stratigraphy of lake IWT1 (named *Gull Lake*) are available in Bouchard et al. (2020). Unit A is composed of sand and gravel with scattered, cm-scale peat and organic debris. The lower section of this unit has a higher density ( $\sim 2.0 \text{ g cm}^{-3}$ ) and a low organic content (mean:  $7.7\% \pm 3.7$ ) compared to the upper sections of the core. The bottom deposit of unit A contains organic matter older than  $4805 \pm 15$

$^{14}\text{C}$  yr BP) (5507 cal yr BP;  $1\sigma$  range: 5584–5586), based on dating of a wood fragment at a depth of 108 cm (Bouchard et al., 2020). Unit B consists of medium to dark brown peat, dated to  $4070 \pm 45$   $^{14}\text{C}$  yr BP (35 cm; 4567 cal yr BP;  $1\sigma$  range: 4444–4789), and interbedded with mm- to cm-thick silt and sand laminations, and with gradational upper and lower boundaries. Throughout the unit, sand-silt layers (aeolian) are roughly interbedded with layers of organic detritus as recorded by shifts in organic matter contents. Towards the upper boundary, unit B progressively grades into dark brown gyttja (unit C). The upper part of unit C is faintly stratified, as the organic-rich material becomes regularly interspersed with silty material. These silty laminae (0.3 to 0.9 cm) are visually distinguishable by their light-grey colour on the CT-scan image and their higher density. The top sediments have the highest water ( $70.6\% \pm 6.1$ ) and organic matter contents ( $20.3\% \pm 5.2$ ) compared to the bottom units. Contrary to lake GT1 the boundaries between the units are diffuse. The bottom of this unit (bulk sample collected at 10 cm) yielded an age of around  $2100 \pm 20$   $^{14}\text{C}$  yr BP (2061 cal yr BP;  $1\sigma$  range: 2004–2101). Fossil diatom assemblages in the 3 units reflect changes in the hydro-climatic conditions and available substrates (both terrestrial and aquatic) in the past. Taxa in Unit A show a poor diversity and are generally associated with cold, organic-poor and mostly alkaline (pH  $\sim 8$ ) waters, typical of Arctic streams; taxa in Unit B show much higher diversity and reflect permafrost peatland environmental conditions (i.e. shallow tundra ponds in organic-rich ice-wedge polygon terrains); and taxa in Unit C are dominated by strictly aquatic (both benthic) species generally living in organic-rich, high-nutrient deeper waters (Bouchard et al., 2020; see Figure. 5 and accompanying section 4.2.2).

#### **4.5.4 Water column profiles of temperature and dissolved oxygen**

Profiles done in early June under the ice cover of lakes GT1 and GT2 (deep; group 1) and lake IWT1 (shallow; group 2) showed an inverse thermal stratification with bottom water temperature reaching  $0.6^\circ\text{C}$  (lake IWT1),  $1.4^\circ\text{C}$  (lake GT1) and  $2.0^\circ\text{C}$  (lake GT2; Figure 32). DO was much lower in lake IWT1 (13% saturation below the ice cover, decreasing to  $\sim 1\%$  near sediment at 3.6 m depth in 2015) than in lake GT1 (43% below the ice,  $<1\%$  below 6 m depth in 2015) and lake GT2 (no data in spring 2015; 70% below the ice, 26% at 9 m depth in 2016, but  $< 2\%$  by mid-May in 2019; no data in spring 2015). Quickly after the ice cover melted at the beginning of July,

the water column became weakly stratified in all three lakes, and warmer in lake IWT1 (above 8°C at the surface, as compared to ~5°C at the surface of the deeper lakes). By then, the water column already showed signs of DO depletion in the deeper lakes (GT1, GT2) but the summer stratification period was short, lasting for about a month in the larger lake GT2. At the bottom of lake GT2, oxygen depletion occurred as soon as stratification established from the beginning of July and decreased down to 84% of saturation by mid-August 2015 (65% in 2018), until the autumnal turnover increased saturation level up again (unpubl. data). While only weak hypoxia was encountered at the bottom of lake GT2 in late summer, anoxia was reached in the hypolimnion of lake GT1 (Figure 32). On its margin, lake IWT1 was generally well-mixed during the open-water period, but presented weakly stratified periods during warm and calm days (e.g. on 3 August 2016). Early August profiles indicate that the entire water column was above 13°C in the shallow thermokarst lake IWT1, while the surface of lake GT1 and GT2 was slightly colder (respectively ~10°C and 11°C in 2016).



**Figure 32.** Early June, July and August temperature and dissolved oxygen (DO) profiles for ice wedge thermokarst lake IWT1 (a and d), and glacial thermokarst lakes GT1 (b and e) and GT2 (c and f). The grey rectangle represents the ice cover in early June) and the average maximum ice thickness is  $2\text{ m} \pm 20\text{ cm}$  (measured in 2015 and 2016; Preskienis et al., 2021). For lake GT2 (f), the dashed lines are simply connecting the two available data (from the mooring) but the shape is likely to follow the more detailed temperature profiles. The hatched rectangle indicates the lake bottom. Note that the dates are slightly different in lake GT2. Profiles from lakes IWT1 and GT1 in early June and late summer (August) are adapted from Preskienis et al. (2021; respectively corresponding to lakes BYL66 and BYL36), allowing to compare with lake GT2.

## 4.6 Discussion

The ice-marginal permafrost environment in the Qarlikturvik Valley is highly heterogeneous, as ground ice types and content can vary and coexist over short distances, leading to significant small-scale differences in lake types, in their morphological and limnological conditions, as well as their vulnerability to climate drivers and disturbances. The bathymetric data

revealed the coexistence of two types of lakes with different morphological characteristics. We also found that different sedimentary facies were present in the cores collected from each group, suggesting different origins and evolutionary conditions.

#### **4.6.1 Lake morphology and sediment stratigraphy of shallow thermokarst lakes formed in ice wedges**

Sixty-two percent of the lakes in Qarlikturvik Valley are shallow (~2-4 m) and relatively flat at their bottom, with a central deeper pool. This group of shallow lakes displays maximum depths very similar to those in 'classic' thermokarst lakes (~1-4 m deep) that developed in segregation ice and ice wedge polygonal terrain, excluding lakes formed in Yedoma-type permafrost (Hinkel et al., 2012; Kanevskiy et al., 2014). Their depth is controlled by the depth of syngenetic ice wedges, and by the amount and distribution of ground ice in the substrate (Grosse et al., 2013). Their maximum depths are also in accordance with the thickness of the peaty silt sequence (~2-3 m) forming the surrounding material, which developed during the Late Holocene (Fortier et al., 2006). Subsequent thermokarst evolution in those basins is not likely to result in substantial subsidence of the lake or basin floor, which can be inferred from the moderate to low ice content of the lowermost glaciofluvial stratigraphic unit (A) of lake IWT1. The lake has been slowly expanding in the frozen silt-peat terrace, and thawing has reached the underlying glaciofluvial sand (Bouchard et al., 2020). The intermediate unit (B) includes a layer (~35-55 cm) of convoluted horizons, which is absent in lake GT1, and likely originates from collapsed bank material in response to thermo-mechanical erosion processes or disturbed horizon due to lake bottom subsidence after ground ice melting. The sediment profile from lake IWT1 is very similar to those found in lakes initiated by the degradation of ice-wedge and intrasedimental ice. These lakes typically have a transitional organic-rich layer containing peat derived from permafrost thawing and subsidence, underlying a layer of laminated organic-rich lacustrine mud (Biskaborn et al., 2013; Bouchard et al., 2017; Farquharson et al., 2016; Murton, 1996). Such an interpretation is further supported by the fossil diatom record investigated in lake IWT1 (named *Gull Lake*; Bouchard et al., 2020), showing a few species typical of cold, oligotrophic and organic-poor (e.g. glaciofluvial) streams in the bottom section (unit A), then dominated by diatom species typical of

moss/peat substrates in the middle section (Unit B), while showing lacustrine conditions with more diverse habitats (benthic/tychoplanktonic taxa) in the upper section (Unit C).

#### ***4.6.2 Lake morphology and sediment stratigraphy of deep glacial thermokarst lakes formed in buried glacier ice***

The other lakes (38%) stand out by their notably deeper basins (~ 5–12 m), and in some cases the presence of multiple sub-basins (e.g., lake GT2). Owing to the size of the lake depressions and their location adjacent to mounds of ice-contact deposits, these deeper lakes were primarily formed by the melting of buried glacier ice. This interpretation is supported by the presence of two exposures of glacier ice revealed by lakeside slumps, which also indicates that the shoreline of lake GT2 is still ice-cored by glacier ice. Côté et al. (2010) have also reported the coexistence of shallow lakes and deeper lakes in similar depositional environments of the Qarlikturvik and the adjacent valley of glacier C-93, where nearly half the lakes had depths greater than 5 m (Figure 28b; depth range: 5-21 m). The cryostratigraphic context of the Qarlikturvik valley is not conducive to the formation of deep depressions. The uppermost unit consists of interstratified peat and silt (thickness ~ 2–4 m) with high volumetric ice content ( $74.6 \pm 10.6\%$ ; Veillette et al., unpublished data), while the underlying unit is glaciofluvial sand and gravel, which typically have low excess-ice contents and minimal expected settlement. Given the thickness of the silt and peat sequence and low ice content of the glaciofluvial sands, the amount of intrasedimental ice, especially segregated ice in excess of the porosity, is not sufficient to create lakes with depths reaching up to 12 m, even if all the intrasedimental ice melted and the resulting water drained out of the soil porosity. In addition, the deepest sections of some of these lakes, for example lakes GT1 and GT2, are ~5 m below current sea level, indicating burial in a glaciomarine/glaciofluvial environment followed by isostatic uplift.

The glacial origin of these deep lakes is further corroborated by the analysis of the sediment core collected in lake GT1 (12.2 m), which differs significantly from the one obtained in lake IWT1 (Figure 31). In lake GT1, four depositional stages were inferred from the sediment profiles. The inception of lake GT1 began with the collapse of supraglacial material during melt-

out of stagnant glacier ice, which resulted in re-sedimentation of sand and gravel from glaciofluvial and mass movement deposits into a forming basin. Inclusions of fibrous peat in unit A and the ~10 cm darker layer (Unit B) of organic debris and inorganic material (mostly silt and sand) were probably derived from surficial vegetation on upland surfaces washed in as the lake basin developed. This interpretation is further confirmed by the presence of scattered, although identifiable, diatom species typical of this depositional environment (e.g., *Eunotia*, *Cymbella*, *Pinnularia*). This basin was then filled by sandy mud (Unit C) deposited by the combined action of meltwater streams and aeolian activity. The most prominent features of this core from lake GT1 are the sharp boundaries of units B and C, indicating a shift in the depositional conditions, which do not reflect gradual deposition within a stable lake floor (Henriksen et al., 2003). Like the upper core section in lake IWT1, the upper unit D includes recent lacustrine sediments composed of organic-rich mud (gyttja). This material was deposited within deeper, calmer waters where fine material can settle. The laminations reflect variability in minerogenic inputs and are likely related to terrestrial runoff or aeolian activity. Similar stages of sedimentation were identified from glacial thermokarst lake basins, including one lake with multiple sub-basins (up to > 14 m deep) in the continuous zone of permafrost in northern European Russia, where buried glacier ice has survived for ca. 80 000 years from the last glaciation (Henriksen et al., 2003) , or in older glacial lakes in northern USA (Yansa et al., 2020). Reversal of ages in the core suggests that older organic matter was also washed into the lake during the mid-Holocene, causing abnormally old dates in basal core sediments, a common dating problem in high-latitude lakes (Bouchard et al., 2017; Wolfe et al., 2004). Furthermore, the absence of identifiable diatom taxa (i.e. presence of scarce fragments only) in the upper section of the core (specifically Unit A) is puzzling. The relatively high pH (> 8) in this lake, especially during the spring bloom (pH ~ 10), could help explain the poor preservation of diatom valves (Ryves et al., 2006).

Moreover, our spatial analysis demonstrated that lake distribution is strongly linked to the maximum and recessional positions of local mountain glaciers and the LIS in both Qarlikturvik Valley and the southern plain of Bylot Island. This is particularly evident in the valley, as shown by the three well-defined lake clusters (Figure 28a). For most of the southern lowland coastal plain,

moderate to high point densities were also encountered within the extent of the LIS. There is also a notable increase in lake density close to the contemporary ice margin, which we interpret to be the result of a relatively recent and continuous deglaciation process. We found that, even after accounting for landscape heterogeneity (i.e. high slope gradients, bedrock exposures), the lakes are still far more clustered when compared to a random spatial distribution. As a result, we propose that the clustering reveal patterns caused by the presence of patches of buried glacier ice. This provides additional evidence for supporting the glacial origin of these lakes. The presence of deep lakes and numerous thaw slumps in Qarlikturvik Valley indicates the delayed melting of several bodies of buried glacier ice as compared to the Holocene glacier retreat. The ice-free zones of Bylot Island are therefore still strongly influenced by its glacial legacy given the presence of late Pleistocene-age glacier ice buried in the permafrost in Qarlikturvik Valley (Coulombe et al., 2019) and in other valleys and coastal plains of the island (Klassen, 1993; Moorman and Michel, 2000). These ice-cored landforms have been adjusting to non-glacial conditions and their evolution is strongly linked with geomorphological processes and local terrain conditions and stability.

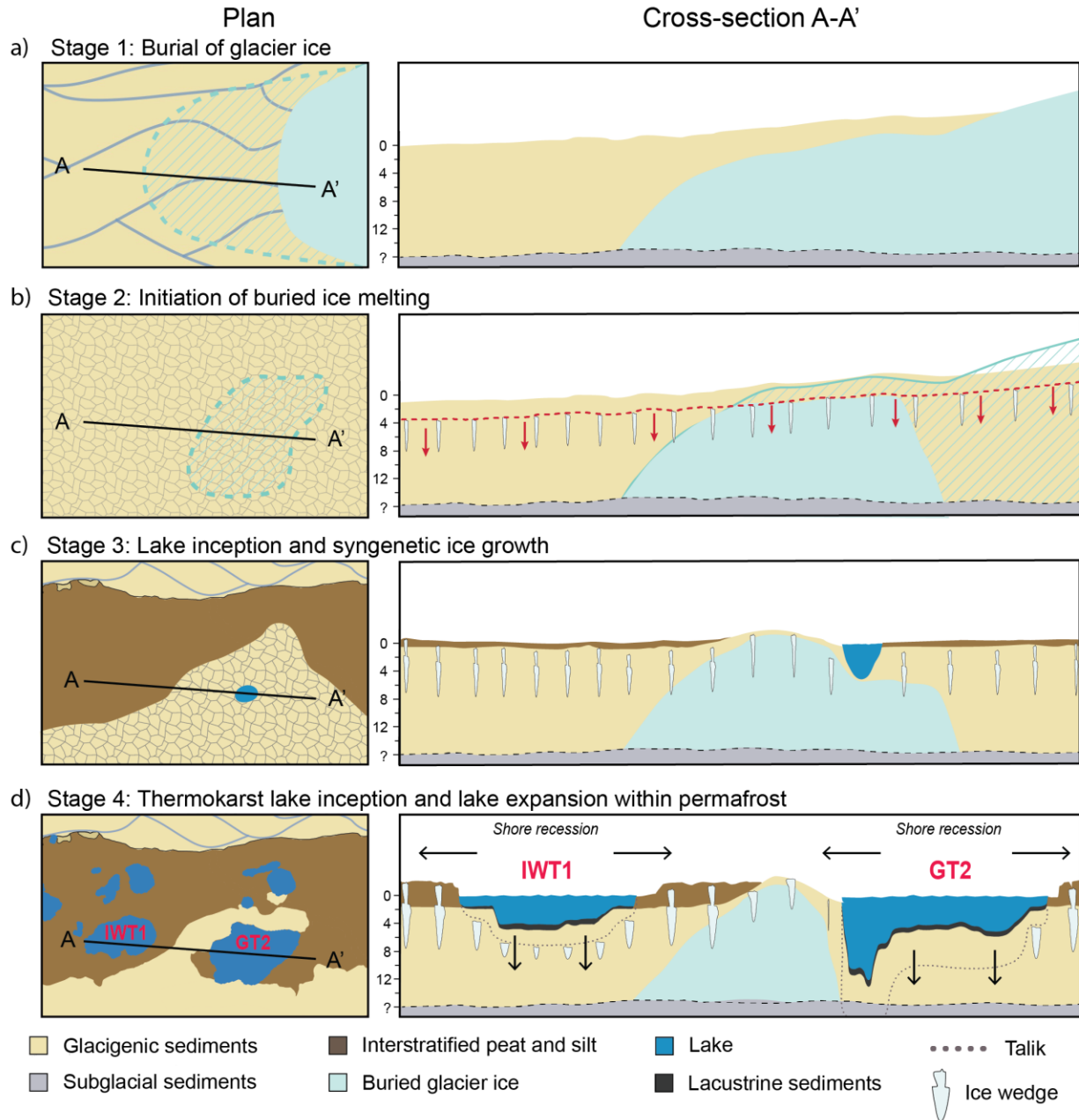
#### ***4.6.3 Conceptual model of thermokarst lake development in ice-wedges polygon terrain and buried glacier ice***

Several studies have described the stages of thermokarst lake development and thaw lake cycle in permafrost environments, such as in the Yedoma (Morgenstern et al., 2011; Shur et al., 2012), lacustrine environments with ice wedges aggradation and degradation in the basins (Billings and Peterson, 1980; Jorgenson and Shur, 2007), and ice-rich cryogenic mounds (Calmels et al., 2008), or ice-wedge and intrasedimental ice (Czudek and Demek, 1970; Jorgenson and Osterkamp, 2005). In a previous study, Bouchard et al. (2020) presented a four-stage conceptual model for lake IWT1 (named *Gull Lake*) that describes thermokarst inception and evolution in syngenetic ice-wedge polygon terrain during the Holocene. Based on this model, lake IWT1 developed in a pre-existing topographic depression (~1-2 m) that collected snow and meltwater (stage 0, initial conditions). The first phase of thermokarst started at around 2100 BP in response to active layer deepening and ice wedge melting, which initiated the development of small and



shallow ponds over the degrading ice wedges (stage 1). Thermokarst ponds started to coalesce with neighbouring water bodies over and at the edge of ice-wedge polygons to form a small lake (stage 2). Over time, this lake expanded in the ice-rich polygon terrace because of surface permafrost degradation via lateral thermal erosion and vertical thaw settlement and consolidation in the ice-rich silt-peat terrace, and eventually in the underlying glaciofluvial sediments (stage 3). The last stage suggests a possible long-term future scenario where the lake disappears through the gradual gyttja accumulation and lake infilling or lake drainage, which can sometimes be catastrophic (Bouchard et al., 2020 and citations therein). The conversion of these aquatic ecosystems to terrestrial or wetland ecosystems is usually followed by a reactivation of old ice wedge networks or growth of pingos as permafrost aggrades in unfrozen drained lake deposits once exposed to cold temperatures, which eventually begin a new phase of the thaw-lake cycle (Billings and Peterson, 1980; Mackay and Burn, 2002; Jorgenson and Shur, 2007).

Based on the geomorphology of the deeper lakes and lake sediment profiles of lake GT1, we also developed a four-stage model of glacial thermokarst lake formation and evolution in the specific stratigraphic context of buried glacier ice within the study area (Figure 33).



**Figure 33.** Schematic diagram showing the sequence of formation of lakes in terrain underlain by relict glacier ice: a) Stage 0: Initial conditions, b) Stage 1: burial of glacier ice, c) Stage 2: initiation of buried glacier ice melting. The red arrows show the deepening of the active layer and d) Stage 3: lake inception and syngenetic ice growth. The black arrows indicate talik deepening in sub-lake sediments and lateral lake expansion.

**Stage 0: Initial conditions:**

During the Last Glacial Maximum, the LIS and local ice caps covered much of the Qarlikturvik Valley, and many outlet glaciers were channelled through major valleys of Bylot Island and terminated on the lowlands.

**Stage 1: Burial of glacier ice:**

Beyond the active margins of local glaciers or the LIS, wide areas of glacier ice were likely buried *in situ* by glacial sediments transported and reworked on top of an active or stagnant glacier margin by mass movement and meltwater (Figure 33a). The burial of glacial ice can still be seen today at the margins of many glaciers on Bylot Island. The TC brightness index exhibits a strong positive trend at the glacier margins, indicating dryer and unvegetated surfaces. This corresponds to sediment accumulation onto the glacier surface and represents a modern analogue of the burial of glacier ice. The brightness trend correlates well with the active burial of ice observed at numerous locations at the margins of glaciers C-93 and C-79 (Figure S10). Progressively, stagnant ice blocks became isolated from the upper active flowing ice. On Bylot Island, bodies of glacier ice were preserved at various places in the outwash plain, in mounds of ice-contact stratified drift, and moraines (Coulombe et al., 2019; Klassen, 1993). Interpretations of the sedimentary sequence overlying the buried ice studied in the valley indicated that the burial of the ice involved glaciofluvial deposition directly on the ice, which was followed by plant colonization. This situation can occur during or after postglacial isostatic uplift. In some cases, glaciofluvial sands and gravels were also covered by colluvial sediments as debris were transferred away from topographic highs by mass movements and meltwater (Coulombe et al., 2019). Preservation of the ice for several millennia was possible because the sediment cover became sufficiently stable and reached or exceeded the active layer thickness, but also because the neoglacial climatic conditions during the second half of the Holocene were conducive to syngenetic permafrost aggradation following glacier retreat (Fortier et al., 2006; Fortier and Allard, 2004; Bouchard et al., 2020).

## **Stage 2: Initiation of buried glacier ice melting:**

Melting of the upper glacier ice and formation of a depression begins when a deepening active layer reaches the glacier ice (Figure 33b). Local factors such as topography, thickness of the active layer, snow accumulation and water pooling in pre-existing depressions, as well as thermal properties of soil all play a role in initiating ice melting. We suggest three scenarios of initial pond inception following the burial of glacier ice: (1) ponds may have formed quickly in the proglacial environment or later during deglaciation as water pooled in these pre-existing topographic depressions caused by the uneven ablation of the glacier surface (i.e., differential melting of dead ice); (2) the ice was first buried under a thin cover of glacial sediment, which was close to the active layer thickness. The sediment cover was thick enough to slow down the melting rate of underlying ice without completely preserving it, which allowed a gradual melting of the ice creating small depressions; (3) the ice was buried under thick sediment cover, acting as a barrier to heat transfer, and preserving the ice in the long term. However, unusual warm and wet conditions have periodically caused the active layer to deepen considerably, initiating the melting of the underlying ice and creating new depressions. Following the burial of the ice, the uneven ablation of the glacier surface produced an irregular topography of ridges and mounds. Since buried glacier ice is still present in the study area, thaw slump activity is thought to have been a fundamental driver of its degradation by exposing the ice and accelerating its melting (Coulombe et al., 2019).

## **Stage 3: Lake inception and syngenetic ice growth**

Episodes of warming sufficient to cause degradation of the existing permafrost and the buried glacier ice in the valley triggered thermokarst lake initiation (Figure 33c). Summer-melt layers from the Agassiz Ice Cap and Greenland ice sheet provide robust records of warmer events during the Holocene (~7-8 ka; Fisher et al., 1995; Westhoff et al., 2021). These warmer periods likely initiated or accelerated the ice melt when (1) a thin layer of sediments covered the ice, or (2) topographic depressions allowed the accumulation of snow and water, hence overall warmer conditions, which further accelerated the melting of the buried glacier ice. This resulted in subsidence of the terrain surface, deeper snow accumulation in winter and ponding of surface

water during the warm season, which began to thaw the underlying permafrost. During the first phase of lake development, relict glacier ice can serve as a focal point for the onset of accelerated thermokarst degradation. If exposed, the ice core then undergoes accelerated wastage through the effects of solar radiation or becomes buried again under the slumping material until a new thermal balance can be reached. We cannot deduce absolute timing for the inception of lake GT1 since no reliable basal dates are available. However, we suggest that lake inception of these deeper glacial lakes occurred sometime during the mid-Holocene and preceded that of the shallowest lakes formed by the thawing of ice wedges, which were  $^{14}\text{C}$  dated at around 2100 yr BP (Bouchard et al., 2020).

#### **Stage 4: Thermokarst lake inception and lake expansion within permafrost.**

Once a lake gets deeper than the maximum thickness of the winter ice cover ( $\sim 2 \text{ m} \pm 20 \text{ cm}$  in the valley as measured by Prėskienis et al., 2021), it will continue to grow laterally (thermo-mechanical erosion) and vertically (subsidence) by thermokarst processes each year (Figure 33d). The water sensitive index TC-wetness exhibits a moderate to strong positive trend for many lakes in the valley, driven by the gradual erosion lake shores containing ice-rich permafrost. The rate of expansion depends on the local climatic conditions, ground-ice content and lakebed temperature. In cases where buried glacier ice remains present beneath the lakebed, the ice will slowly continue to melt, causing lake bottom subsidence. Further ground ice melting and the resulting thaw slumps contribute to lake expansion, as shown by the head scarps located close to the shoreline of lake GT2 (Figure 30b). Other studies have shown that thaw slumping is an important mechanism of lake expansion (Hinkel et al., 2012; Plug and West, 2009; Kokelj et al., 2009). Our results indicate that these glacial thermokarst lakes also evolved at a later stage as ‘classic’ thermokarst lakes that are now slowly expanding in area and volume, because of the melting of intrasedimental ground ice and ice wedges in the frozen silt-peat terrace and in the underlying glaciofluvial and till material. The shorelines of glacial lakes are expected to be very smooth and roughly circular or oval-shaped, as shown by the morphological analysis of glacial lakes located at the termini of glaciers C-93, C-79 and C-67. However, most lakes studied here display slightly irregular shorelines (Figure S11). In the Qarlikturvik Valley, the shoreline morphology of the

deeper glacial thermokarst lakes is very similar to the other thermokarst lakes, indicating that all lakes are now laterally expanding in the polygon terrace by thermal and mechanical erosion. Thermokarst is an active landscape change mechanism currently operating in the valley and on the island in general (Bouchard et al., 2020; Godin et al., 2014; Fortier et al., 2007). Today, the lakes can expand by thermal subsidence and different shoreline erosional processes including: (1) the development of thermo-erosional niches; 2) the mechanical erosion caused by lake ice pushing against the shore, and 3) the incorporation of adjacent polygonal ponds into the lake (Jones et al., 2011). Eventually, the lakes may cease expanding in the event of partial surface/subsurface drainage through various permafrost degradation processes. Complete drainage of these glacial thermokarst lakes remains impossible due to their great depth which is below the base level of streams and river and even below sea level in some instances. This situation allows these lakes to persist over time, unlike shallow lakes that have developed in segregation ice and ice-wedge polygons, which are susceptible to complete drainage (Mackay, 1992) and to return to terrestrial conditions. This shows the interplay of climatic (external) and local landscape (internal) processes in the inception and evolution of thermokarst lakes in general, including the ones developed through melting of buried glacier ice.

#### ***4.6.4 Implications on Arctic lakes ecosystem dynamics***

Lake morphometry, specifically depth, plays an important role in regulating lake water temperature and associated biogeochemistry. It influences the mixing regime and the number of thermal overturn events per year during the open-water period (i.e., if the lake is monomictic, dimictic or polymictic, the latter being more common for Arctic lakes; Rautio et al., 2011). This is intrinsically linked to water column aeration and light regime, thus exerting a strong control on respiration and primary production (Vincent, 2010). Results indicate that the three studied lakes can be considered as cold polymictic (or potentially dimictic depending on the year for lake GT1, although a mooring would be needed to validate this). Among the three lakes studied for their limnological characteristics, bottom water of shallow lake IWT1 was the coldest by late winter but the warmest by late summer, a pattern directly linked to its mixing regime where meteorological conditions are more likely to influence bottom water temperature and talik

formation. We also found that lake morphology influences dissolved oxygen. Lake IWT1 showed the lowest oxygen concentration at the end of the winter ( $< 2 \text{ mg L}^{-1}$  or  $< 13\%$  saturation in 2015), likely linked to its large sediment area to water volume ratio and its higher organic content at the lake bottom (submerged peat polygons, as opposed to less organic sediments in the deeper lakes GT1 and GT2) leading to a faster depletion of oxygen. In addition of controlling GHG cycling, this can be a significant limiting factor for overwintering fish populations (Leppi et al., 2016). For the deeper lakes (GT1, GT2), the difference in water column stability controlled bottom oxygen saturation levels during the open-water period, which decreased well below 60% in lake GT1 (reaching anoxia just above sediment), while it always remained above this level in lake IWT1. The stronger gradient in lake GT1 is likely related to its smaller size (smaller fetch) and greater depth. Climate change may therefore not only affect water temperature, mixing regime and oxygen availability through warming and summer lengthening, but also through effects on the evolution of lake morphology from the melting of buried glacier ice.

These differences in the mixing regime and oxygen availability, controlled by lake morphology (size and especially depth), exert a strong control on the timing (seasonal differences) and magnitude of GHG emissions ( $\text{CH}_4$ ,  $\text{CO}_2$ , and their relative proportion) from the water column to the atmosphere (Hughes-Allen et al., 2021; Prėskienis et al., 2021; Bouchard et al., 2015; Matveev et al., 2016). Previous studies showed that lake IWT1 generally maintained high GHG fluxes during the open-water period as compared to the deeper lakes GT1 and GT2 (named *kettle lakes* Prėskienis et al., 2021). Once more, the combination of warmer temperatures and higher organic content of lake IWT1 likely explains its higher GHG emissions (BYL66, Prėskienis et al. 2021; see Fig. 3 and 4, and Table 4), averaging  $27.1 \text{ mmol CO}_2 \text{ equivalent m}^{-2} \text{ d}^{-1}$ , as compared to  $10.8 \text{ mmol CO}_2 \text{ equivalent m}^{-2} \text{ d}^{-1}$  from lake GT1 (mainly caused by differences in  $\text{CH}_4$  ebullition; not assessed in lake GT2). Considering that most GHG are emitted from lake sediment (Bastviken et al., 2004), it is important to underline that the largest sediment area of a lake is in contact with epilimnetic (shallow) waters, and therefore not only bottom water temperature in the deepest pelagic section of a lake needs to be assessed. Moreover, glacial thermokarst lakes subjected to partial drainage will maintain year-round GHG emissions whereas thermokarst lakes formed in

polygonal terrain and subjected to complete drainage will have a totally different GHG emissions regime (terrestrial GHG emission during the warm season when active layer soils are thawing). Due to the importance and diversity of lakes across the circumpolar Arctic, a better knowledge of their bathymetry and landscape variability is necessary to upscale local biogeochemical assessments to regional or continental scales. The future melting of buried ice, widespread in certain regions of the Arctic but overlooked, will form new lakes that will present different features than classic thermokarst lakes, notably in terms of water temperature, mixing regime, oxygen availability, GHG production, and GHG ages.

## 4.7 Conclusion

Spatial variability in ground ice conditions is an important factor driving lake inception, evolution and distribution on Bylot Island. This study confirms that glacierized permafrost terrain containing various types of ground ice, including buried glacier ice, can influence the spatial distribution of lakes, lake bathymetry, limnological properties as well as lake bottom morphology and sediment stratigraphy. The origin and growth of numerous thermokarst lakes in the Qarlikturvik Valley, Bylot Island, has been examined using bathymetric and field surveys, high-resolution remotely sensed imagery, and lake sediment analysis. Slightly more than half of the twenty-one studied lakes tend to be shallow (~2-3 m), while the other lakes stand out by their notably larger depths (~9-12 m). The stratigraphic analysis of two lake sediment cores revealed two distinct basin types in terms of sediment accumulation, although more work is needed to confirm this difference by collecting sediment cores from a larger set of lakes in the Qarlikturvik Valley, and in other glacierised Arctic tundra settings. These dissimilarities indicate that these lakes have a different origin and evolution throughout the Holocene as well as distinct depositional history and sedimentological signature. These results suggest that the melting of ice wedges and intrasedimental initiated the formation of the shallow lakes (< 5 m), while the melting of buried glacier ice has triggered the inception of the deeper lakes (> 5 m, up to 12 m) in the study area. The glacial origin of deeper thermokarst lakes is supported by the past and current presence of buried glacier ice as well as numerous stable and active thaw slumps in the study area. In addition, the shallow and deeper lakes coexist within the same depositional environment,



indicating that these lakes have been subjected to the same environmental and climatic conditions, and therefore notable depth difference must be related to different ground ice volume or time spanned since inception. Moreover, analysis of lake morphometry and distribution revealed that lakes are more densely distributed near the most recent ice positions. This suggests a relationship between the formation of lakes and the deglaciation patterns in both Qarlikturvik Valley and the broader southern plain of Bylot Island. Given future climate projections, it is likely that Arctic lowlands with glacier ice buried in permafrost will change dynamically because of surface permafrost degradation and melting of relict glacial ice. It is expected that the deepening of the active layer and talik development, as well as the enlargement of Arctic lakes in response to global warming, will reach undisturbed buried glacier ice. This will create new aquatic ecosystems and strongly modify existing ones through the lateral expansion of lakes caused by wind- and circulation-driven erosion, thaw slumping and thaw subsidence along lake margins. In turn, this will likely have pervasive effects on geomorphological, hydrological, and ecological processes of affected landscapes, including the high-latitude and global carbon budgets and oxythermal quality of fish habitats.

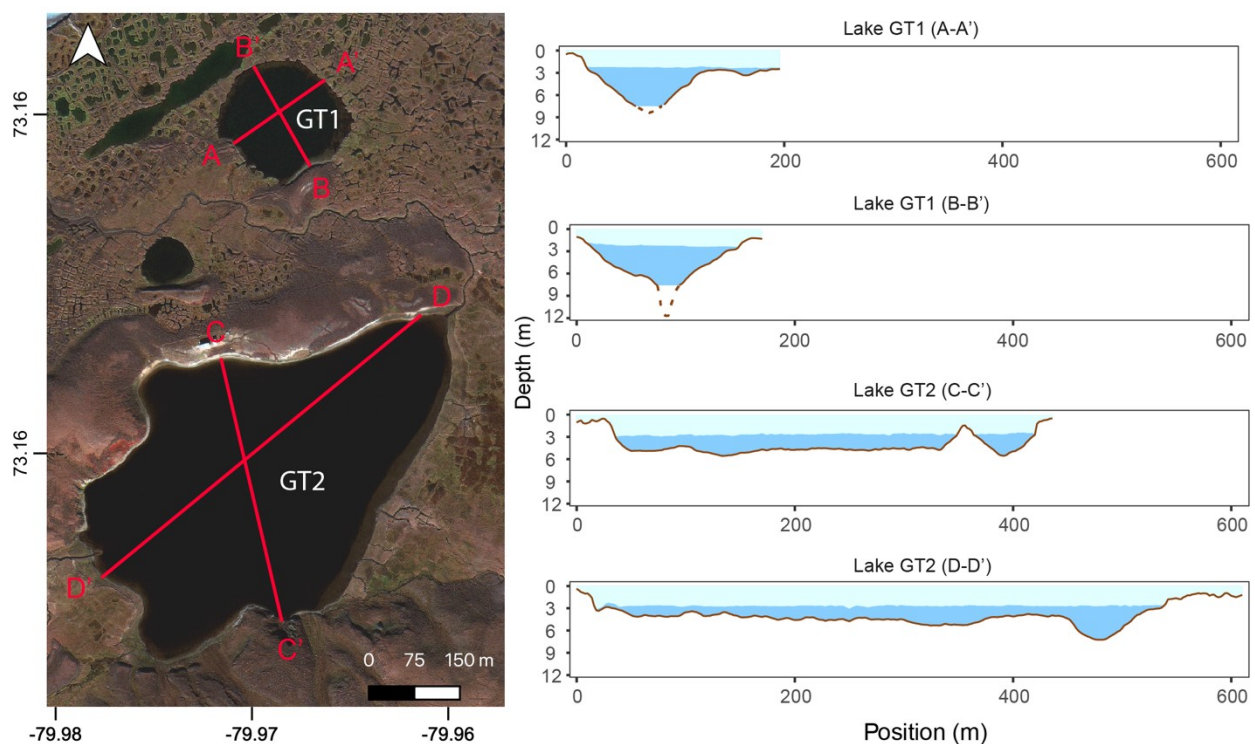
## **4.8 Acknowledgement**

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## 4.9 Supporting information

### *S1 Bathymetric mapping through lake ice with ground-penetrating radar (GPR)*

The difference between the dielectric properties of ice, water, and the sediments under lake bottoms makes GPR an effective method for determining lake depth profiles in the field (Moorman, 2002; Paquette et al., 2015). In this study, we used a pulseEKKO PRO controller manufactured by Sensors and Software coupled with 100 megahertz (MHz) antennas to obtain depth profiles from lake K and lake L. GPR profiles were calibrated and correlated with water depth measurements. Profiles were post-processed using Sensor and Software EKKO Project Version 5 proprietary software. Post-processing included time-zero correction and integration of GPS data, topography, and horizontal filtering to improve visualization of horizontal reflectors.



**Figure S7.** Ground penetrating radar profile of lakes K and L (background: GeoEye, 2010)

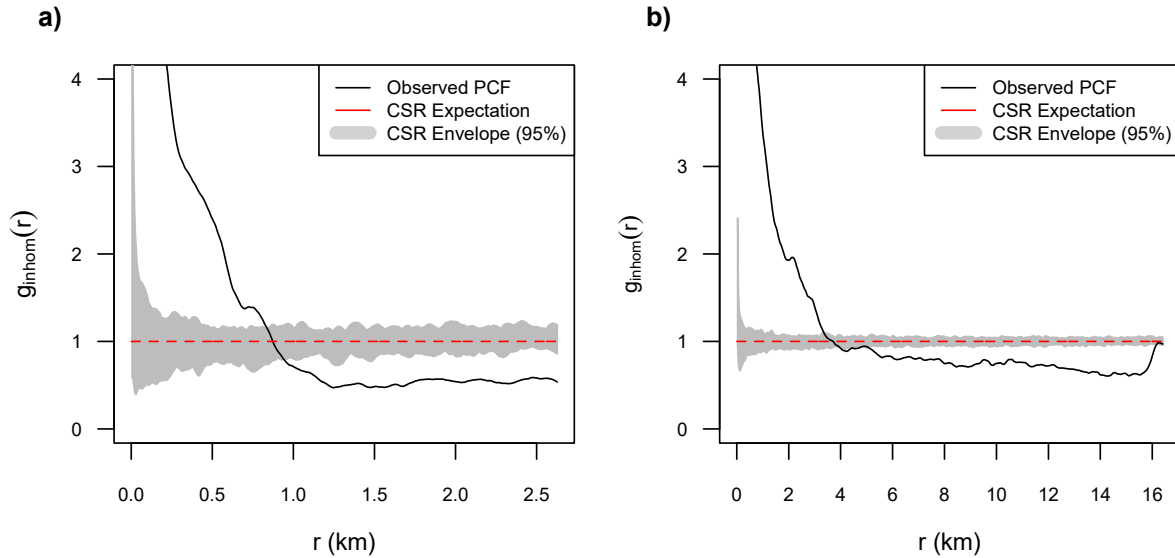
## ***S2 Computed tomography (CT) scanning of a sediment core collected in lakes K and G, Bylot Island, Nunavut, Canada***

To further investigate the cryostratigraphic characteristics of the ice and lake sediments, all samples were observed under X-ray computed tomography (CT) scanning (Siemens SOMATOM Sensation 64) at INRS- ETE (Quebec City, Canada), as in Calmels and Allard (2004). This technique relies on the calculation of the linear attenuation coefficient that measured the density of an object passed through an X-ray beam at different angles (Boespflug et al., 1994). A CT-scan produces cross-sectional images (usually 512 by 512 pixels matrix) of an object where each pixel of the image is assigned an X-ray attenuation value ( $\mu$ ), also called a CT number. CT numbers are then standardized using the Hounsfield scale, where the radiodensity of water ( $\mu_{\text{water}}$ ) is arbitrarily defined as 0 HU (Hounsfield units) according to Equation 1 (Hounsfield, 1973). Different shades of gray are assigned specific CT numbers to create the displayed image using a specific image processing software (Fiji) dedicated to DICOM (Digital Imaging and Communications in Medicine) images. In permafrost samples, unconsolidated sediments and rock (high density minerals) appear white, as the attenuation of these materials is very high. Gas inclusions and water appear black and other materials, such as ice, can have various shades of grey depending on their density.

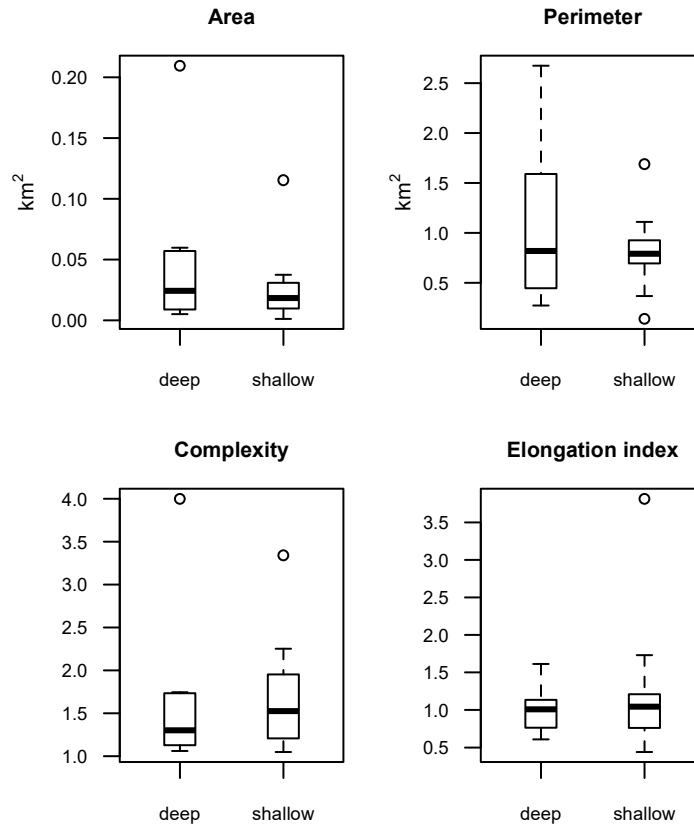
$$\text{HU value} = \frac{\mu - \mu_{\text{water}}}{\mu_{\text{water}}} \times 1000 \quad (1)$$

This tool helps refine cryostratigraphic characterization of permafrost cores as it can reveal characteristics otherwise difficult or even impossible to observe with the naked eye. It allows visualization and characterization of the internal components and structures of the frozen sample, such as ice, grain-size variations, layer orientation in space, and gas inclusions. From a quantitative perspective, it has been used to segment images into regions of ice, gas and sediment in order to quantify the volumetric content of the scanned sample (Calmels et al., 2010b; Dillon et al., 2008b). Images were processed with FIJI to map X-ray attenuation coefficients on longitudinal images and to visualize sedimentary structures. The resulting images are displayed in greyscale, with darker grey representing a lower X-ray attenuation. Greyscale values are

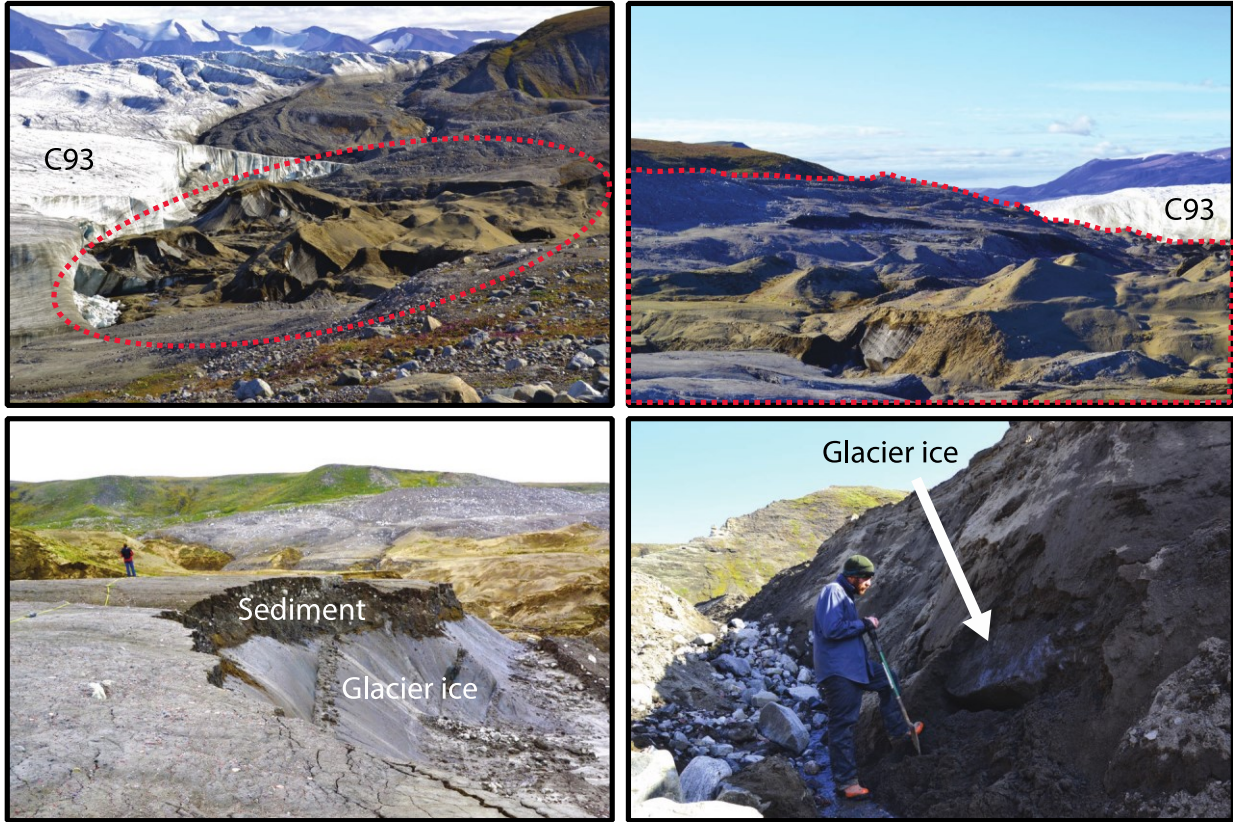
expressed as CT numbers, which are complex units related to the mineralogy, organic matter content, grain size, and bulk density.



**Figure S8.** Pair Correlation Function results for the a) valley and the b) southern plain. The function  $g_{inhom}(r)$  is plotted against  $r$ , with increasing radii of analysis (in km) away from each lake. The black line shows observed  $g(r)$  values, the dashed red line indicates the mean of the inhomogeneous null model and the grey area represents the simulation envelope (95% confidence interval), which is derived from the Monte-Carlo simulations of CSR. Values plotting within the grey envelope are not significantly different from a random spatial arrangement.

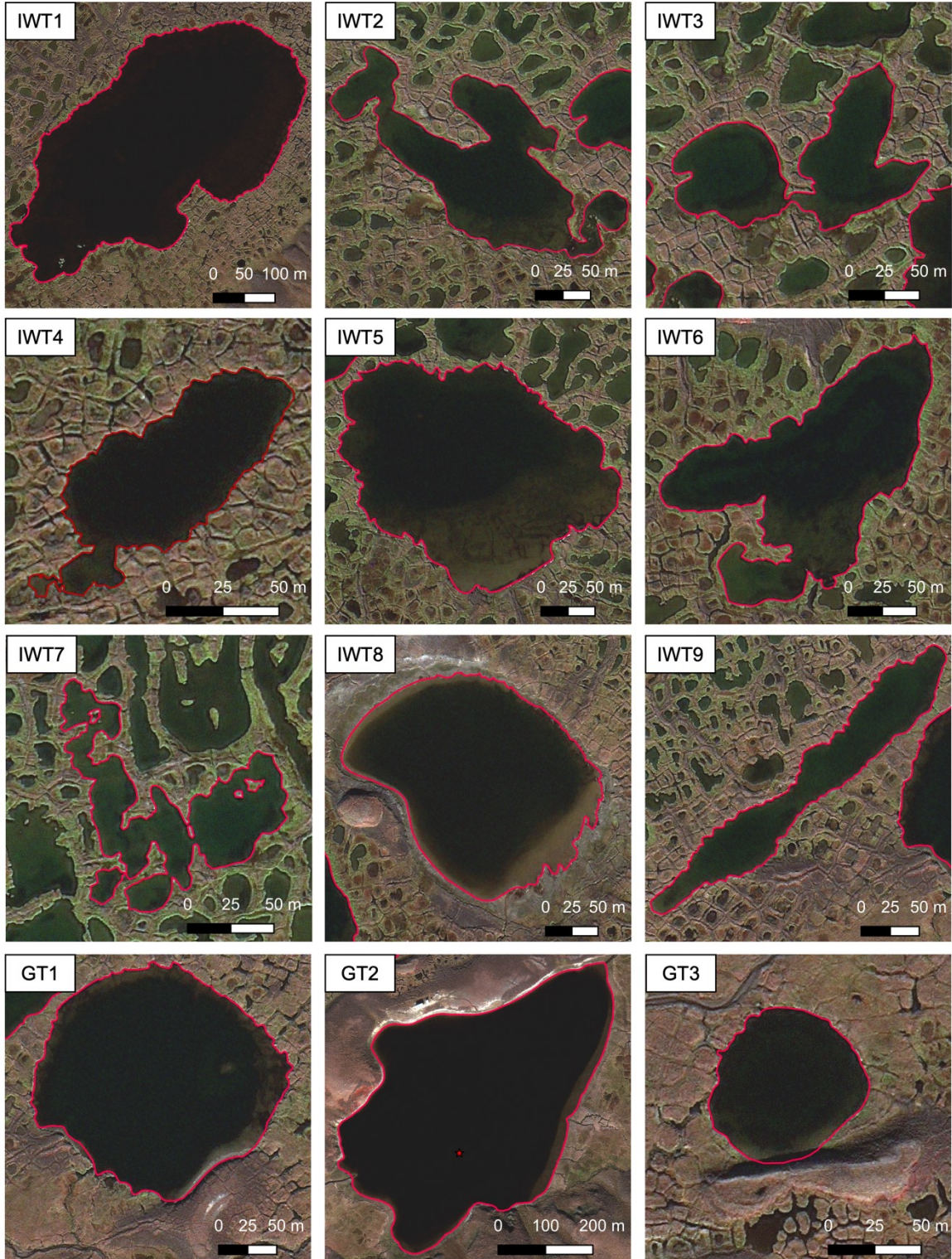


**Figure S9.** Box plots comparing the morphometric properties (area, perimeter, complexity, elongation index) of deep (>4 m) and shallow lakes (< 4 m) obtained from all the digitised lakes. The thick line marks the median value. The bottom and the top of the box correspond to the first and third quartiles, respectively. The whiskers show the range of observed values that are not within the first and third quartile but not further away than 1.5 times the interquartile range (IQR) from the hinges, and open circles represent outliers.

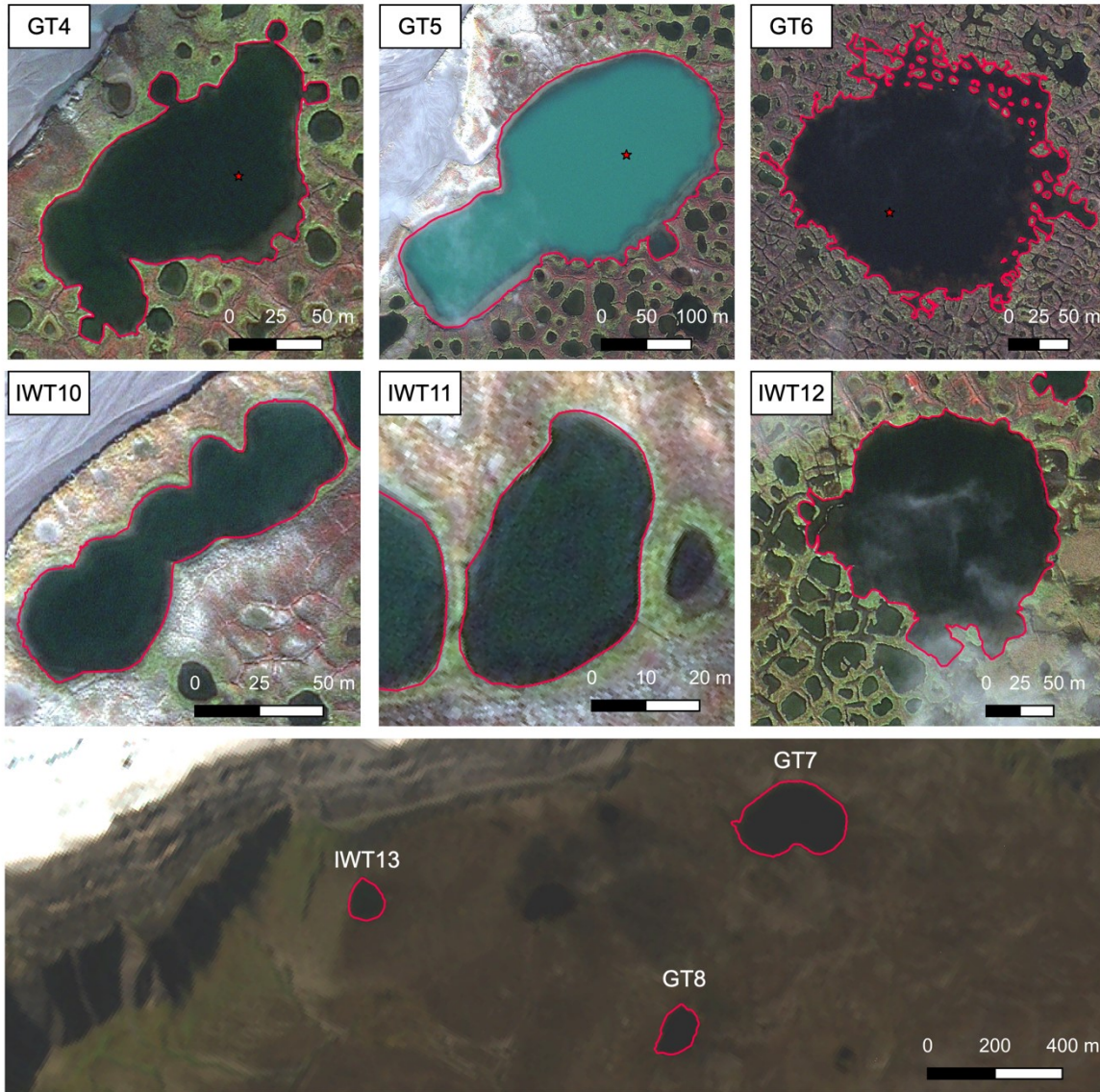


**Figure S10.** Burial of glacier ice at the margin of glacier C-93. The red dashed lines highlight areas of buried glacier ice near glacier C-93.









**Figure S11.** Satellite images of lakes (background: GeoEye, 2010). The red line highlights the shoreline and shape of the lakes.



## Chapitre 5 : Conclusion générale et perspectives

La glace de glacier enfouie est une composante importante des environnements périglaciaires façonnés par une succession d'avancées et de reculs glaciaires durant le Quaternaire. Depuis environ deux décennies, l'augmentation des observations de glace de glacier enfouie dans l'ensemble des régions arctiques tend à démontrer que ce type de glace est plus abondant et répandu qu'initialement estimé (Fritz et al., 2011; Coulombe et al., 2019; Wolfe et al., 2021; Rudy et al., 2017; Kokelj et al., 2017a; Belova, 2015; Lakeman and England, 2012; Lacelle et al., 2007; O'Neill et al., 2019; Murton et al., 2005). Il demeure néanmoins difficile d'estimer la quantité de glace de glacier enfouie dans le pergélisol en raison de son caractère invisible. Lorsque visible, les observations directes de glace de glacier enfouie demeurent sporadiques. Malgré cela, il est maintenant reconnu que la glace de glacier enfouie aura une incidence déterminante sur l'évolution géomorphologique des environnements arctiques en raison de sa distribution et du volume de glace impliqué, en particulier dans un contexte de réchauffement du climat arctique (Kokelj et al., 2017a; Rudy et al., 2017; Coulombe et al., 2022; Segal et al., 2016; Senderak et al., 2017; Lenz et al., 2013). De plus, la glace de glacier enfouie suscite, depuis quelques années, un intérêt grandissant dans différentes sphères de la recherche nordique, car la fonte de ces noyaux de glace engendre une multitude d'effets en cascade sur différentes composantes des écosystèmes arctiques (Rudy et al., 2017; Coulombe et al., 2022; Lenz et al., 2013; Prèskienis et al., 2021; Brosius et al., 2021). Au-delà de son impact sur l'évolution du paysage, ce type de glace constitue aussi une source importante d'information pour l'étude des dynamiques glaciaires, des changements environnementaux et des reconstitutions paléogéographiques, tant à l'échelle locale que régionale (Lacelle et al., 2018; Coulombe et al., 2019; Lacelle et al., 2007; Harris et Murton, 2005; Murton et al., 2005; Basilyan et al., 2008). Dans ce contexte, cette thèse s'intéresse à ces différents aspects touchant la glace de glacier enfouie. Ce projet fait également le pont entre les domaines glaciaire et périglaciaire.

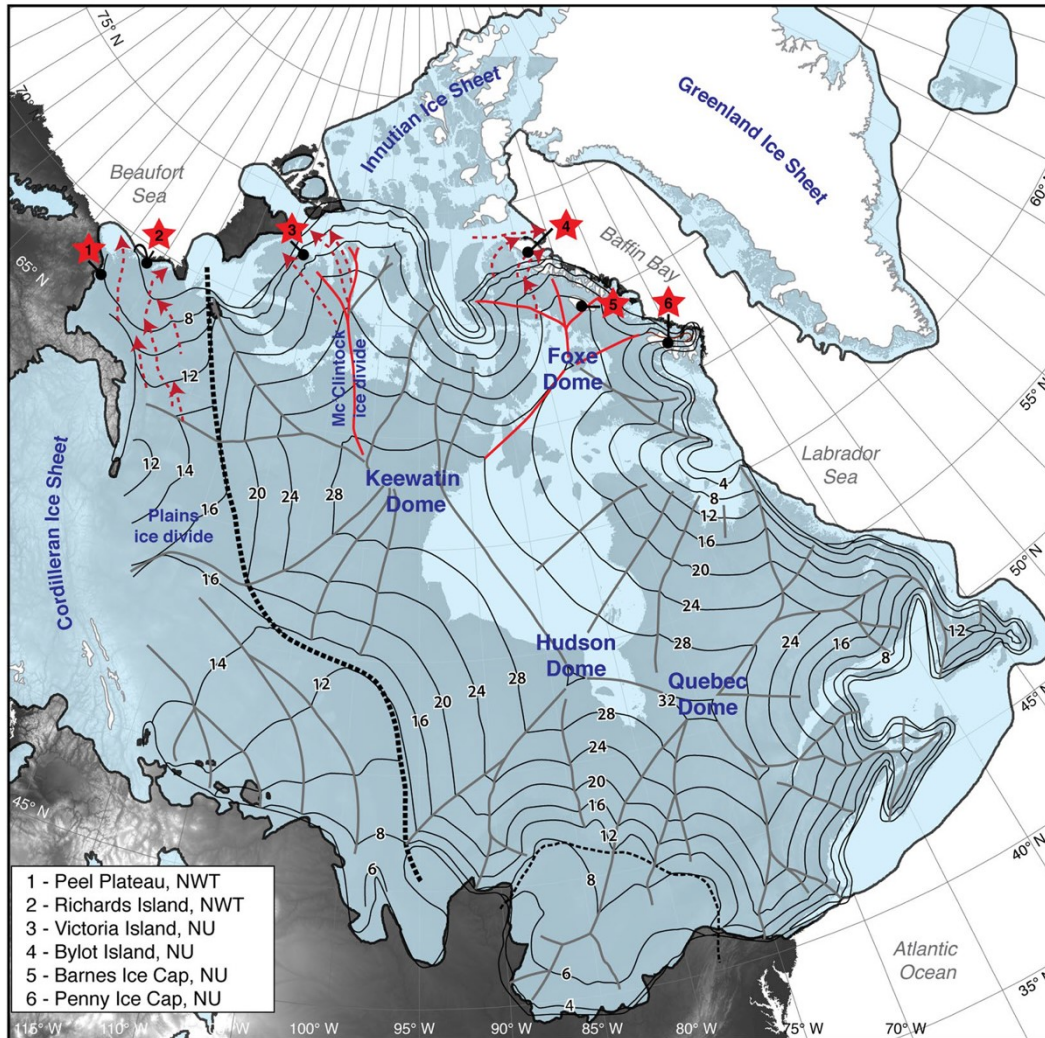
## 5.1 Synthèse des résultats et avancées scientifiques

Dans le cadre de ce projet de recherche, nous avons étudié deux expositions de glace massive découvertes à l'île Bylot, au Nunavut. L'une de ces expositions était située dans une vallée glaciaire et l'autre en bordure d'un plateau bordant cette même vallée, à environ 500 m d'altitude. Dans les deux premiers chapitres, nous avons utilisé plusieurs approches méthodologiques complémentaires pour déterminer l'origine de la glace massive et estimer l'âge de celle-ci. Puis, nous nous sommes intéressés aux processus sédimentaires ayant mené à l'enfouissement et la préservation de glace dans le pergélisol. Les analyses cryostratigraphiques et géochimiques ont permis d'attribuer une origine glaciaire à ces corps de glace. À notre connaissance, c'est la première fois que des masses de glacier enfouies sont rapportées et étudiées de façon détaillée dans l'est de l'Arctique canadien.

### 5.1.1 Glace de glacier datant de la fin du Pléistocène

La glace de glacier enfouie présente dans le fond de la vallée (chapitre 2) présentait de très grandes similarités avec la glace typique de glacier issue de la compression de la neige et du névé. En effet, la glace étudiée était principalement blanchâtre et composée de gros cristaux (cm) orientés aléatoirement et de petites bulles d'air sphériques concentrées à l'interface entre les cristaux. La nature des sédiments encaissants révèle quant à eux que la glace a été enfouie sous des sédiments fluvio-glaciaires qui ont par la suite été remobilisés par des processus gravitaires. L'analyse des isotopes stables de l'oxygène et de l'hydrogène a révélé que la glace était fortement appauvrie en isotopes lourds en comparaison avec la composition isotopique de la glace holocène des glaciers et calottes glaciaires contemporaines. La composition isotopique de la glace de glacier enfouie se compare davantage à celle des glaces datant du Pléistocène tardif situées à la base des calottes glaciaires de Barnes sur l'île de Baffin. Bien que l'âge absolu n'ait pas pu être déterminé, les données de composition isotopique suggèrent que cette glace proviendrait d'un courant de glace s'écoulant du dôme de Foxe de l'inlandsis laurentidien vers la fin du Pléistocène. En effet, les résultats de l'analyse isotopique ont contribué à une étude plus large visant à reconstituer la géométrie de l'inlandsis laurentidien lors du dernier maximum glaciaire à partir de la relation entre l'altitude et  $\delta^{18}\text{O}$  (Lacelle et al., 2018). En utilisant le signal isotopique de glace d'inlandsis

enfouie à différents endroits dans l'Arctique canadien et de la glace des calottes de Penny et de Barnes situées sur l'île de Baffin, cette étude suggère que la topographie de l'inlandsis Laurentidien aurait été asymétrique. Au paroxysme glaciaire, l'altitude des glaces aurait atteint environ 3200 m dans le secteur du dôme de Keewatin, alors qu'une glace plus mince (1700-2100 m) prévalait dans les prairies le long de la ligne de partage des glaces (Figure 34; Lacelle et al., 2018). À l'île Bylot, le signal isotopique  $\delta^{18}\text{O}$  obtenu dans la glace de glacier enfouie indique que la glace s'est formée vers 2150 et 2350 m d'altitude, ce qui signifie qu'elle provient très probablement du dôme de Foxe. Centré sur le bassin de Foxe entre l'île de Baffin et la péninsule de Melville, au nord de la baie d'Hudson, le dôme de Foxe est l'un des quatre dômes glaciaires coalescents de l'inlandsis laurentidien. Lors du dernier maximum glaciaire, ce dôme s'étendait sur une large partie de l'île de Baffin à partir du centre du bassin de Foxe suivant un mode d'écoulement radial (Dyke et al., 1982; Dredge, 2001; Dyke, 2004; Ives et Andrews, 1963; De Angelis et Kleman, 2007). Au cours de la glaciation régionale la plus récente identifiée par Klassen (1993; glaciation d'Éclipse), des courants de glaces de l'inlandsis Laurentidien ont complètement occupé les chenaux autour de l'île et débordés dans la plaine de Sud et les parties basses des vallées de l'île Bylot comme en témoigne la présence de glace de glacier enfouie dans la vallée Qarlikturvik. Ce travail représente aussi une contribution aux connaissances sur la paléogéographie de la région à l'étude, soit une pièce du casse-tête initié notamment par Klassen (1993; 1988), Dyke (2000; 2003; 2001), De Angelis et Kleman (2007) et Margold (2015).



**Figure 34.** Sites de glacier de glacier enfouie issue de l'inlandsis laurentidien (LIS) et son élévation lors du dernier maximum glaciaire (Lacelle et al., 2018). L'étendue de l'inlandsis (y compris les plates-formes glaciaires) au dernier maximum glaciaire est issue de Dyke et Prest (1987). L'élévation de la surface du LIS est dérivée du modèle d'état stationnaire de Fisher et al. (1985) qui est basé sur les marges empiriques de la calotte glaciaire. Les élévations de surface sont en centaines de mètres au-dessus du niveau de la mer actuel (les erreurs sont de 5 à 7%). La ligne noire pointillée épaisse est la limite entre les lits déformables dans les régions des Prairies et des Grands Lacs et les lits durs pour les régions de l'intérieur et de l'est. Les lignes rouges en pointillés représentent v déduit et la zone source pour les quatre sites LIS enfouis. La topographie sous-jacente provient des données d'élévation numériques GTOPO30 (<https://lta.cr.usgs.gov/>). Tous les détails sont disponibles dans Lacelle et al. (2018).

### **5.1.2 Glace de glacier datant du Pléistocène inférieur**

Quant à la glace de glacier enfouie située sur le plateau bordant la vallée (chapitre 3), elle a été observée pour une première fois, à la base d'une coupe stratigraphique, dans le cadre d'un projet portant sur la forêt fossile découverte sur le même plateau (Guertin-Pasquier, 2012; Piraux, 2005; Csank et al., 2011; Fortier et al., 2009). À l'issue des analyses cryostratigraphiques et géochimiques, nous avons également attribué une origine glaciaire à la glace massive exposée à la suite d'un glissement de terrain. La glace étudiée était formée de petits cristaux millimétriques et petits agrégats (mm) composés de silt, sable et graviers suspendus ou stratifiés. Les structures de déformation dans la glace sont très probablement héritées du mouvement par fluage et de la déformation ductile de la glace. Les cryostructures présentaient de très grandes similarités avec celles observées dans la zone basale des glaciers actuels sur l'île Bylot, mais aussi ailleurs comme en Alaska (p.ex. glacier Matanuska; Coulombe et al., 2022; Fortier et al., 2012, 2011).

Tout comme la première exposition de glace de glacier enfouie présentée dans le premier chapitre, la nature des sédiments encaissants révèle que les conditions d'enfouissement et de préservation de la glace seraient associées à des processus de sédimentation fluvioglaciaire et de colluvionnement. Des analyses paléomagnétiques, effectuées sur des sédiments déposés au-dessus de la glace, ont révélé une polarité normale dans les sédiments du haut (chron de Brunhes; 0,78 Ma – aujourd'hui), suivie d'une polarité inverse (chron de Matuyama; 0,773–2.595 Ma) et d'une polarité positive dans les sédiments du bas. Cette polarité positive peut être associée au subchron de Jaramillo (1,071 – 0,990 Ma); 2) tout autres subchrons ultérieurs (p. ex. Cobb Mountain, Olduvai, Réunion) dans le cas où il y aurait un hiatus dans la séquence sédimentaire ou 3) au chron normal de Gauss (2.595–3.032 Ma). Ce troisième scénario est envisagé compte tenu de la position stratigraphique de l'unité de glace enfouie au-dessus d'une unité organique fossile composée entre autres d'espèces similaires que l'on retrouve de nos jours dans les forêts les plus nordiques (pins, mélèze, épinette, aulne) et dont la limite de distribution est beaucoup plus au sud. Les études paléomagnétiques et la présence d'espèces éteintes suggèrent que l'âge de ces dépôts organiques fossiles pourrait être compris entre 1,7 et 1,9 Ma ou entre 2,14 et 2,18 Ma, mais l'âge le plus probable serait entre 2,4 et 2,7 Ma (entre les MIS 103 et 95; (Guertin-

Pasquier, 2012; Piraux, 2005; Csank et al., 2013) . Ces résultats suggèrent que ces dépôts forestiers fossiles représentent un vestige des forêts qui se sont développées à la limite septentrionale des arbres pendant une période interglaciaire plus chaude du Plio-Pléistocène. Cette transition du Pliocène vers le Pléistocène est caractérisée par un refroidissement progressif du climat menant à une augmentation importante du volume des glaces polaires et au développement de cycles glaciaires-interglaciaires vers 2,7–2,9 Ma (Lisiecki et Raymo, 2007; Ravelo et al., 2004; Raymo, 2006). Ainsi, les changements de polarité observés dans la séquence stratigraphique permettent d'établir l'âge minimum de la glace basale (> 0,773 Ma; inversion de Bruhnes–Matuyama) et donc, la glace proviendrait d'une glaciation qui s'est produite durant le Pléistocène inférieur (2,595–0,773 Ma). Or, la glace pourrait également être associée à un épisode glaciaire ayant eu lieu à la transition Pliocène–Pléistocène. Cette glace serait donc l'une des plus vieilles glaces de glacier connues en Arctique à l'extérieur de la calotte glaciaire du Groenland et représente l'une des premières indications de glaciations dans l'est de l'Arctique canadien.

En raison de l'âge obtenu par l'analyse paléomagnétique (> 0,773 Ma), la glace de glacier enfouie présentée dans ce chapitre témoigne de l'une de ces avancées glaciaires ayant eu lieu durant la première partie du Pléistocène (Pléistocène inférieur). La localisation de la glace sur un plateau, à environ 500 m d'altitude, suggère que la glace proviendrait vraisemblablement d'une avancée glaciaire régionale. La Glaciation de Baffin est la plus vieille avancée glaciaire régionale reconnue sur l'île Bylot par Klassen (1993), mais l'âge de celle-ci demeure inconnu. Cet épisode glaciaire est reconnu par la présence et la répartition d'erratiques (grès quartzifère; Adams Sound Formation, île de Baffin) à l'intérieur des terres et au-dessus des limites glaciaires de la glaciation d'Éclipse et par des dépôts glaciaires d'âge pré-éclipse dans quelques sections stratigraphiques. Ces erratiques, de même qu'un cordon morainique, ont été observés sur le même plateau que celui où la glace de glacier enfouie a été découverte. Celle-ci pourrait être associée à la glaciation de Baffin, mais cette hypothèse demeure à ce stade-ci très préliminaire puisque l'âge précis de cet épisode glaciaire et de la glace enfouie n'est pas connu. Bien que des travaux supplémentaires soient nécessaires pour préciser l'âge des sédiments encaissants et la glace de glacier, cette étude contribue à l'avancement des connaissances sur l'histoire paléogéographique de cette région de

l'Arctique, particulièrement dans un contexte où les traces d'anciennes glaciations antérieures au dernier maximum glaciaire sont très rares.

### **5.1.3 Rôle géomorphologique de la glace de glacier**

Alors que les chapitres 2 et 3 présentés dans cette thèse ont démontré la présence de glace de glacier enfouie dans une vallée glaciaire de l'île Bylot, le quatrième chapitre s'est plutôt intéressé au rôle joué par la glace de glacier enfouie sur l'évolution géomorphologique des vallées glaciaires. Il est reconnu que la fonte de ce type de glace entraîne de nombreux processus thermokarstiques, dont les plus connus sont les glissements de terrain (Kokelj et al., 2017a; Segal et al., 2016; Rudy et al., 2017; Ewertowski et al., 2019; Senderak et al., 2017). En effet, de nombreuses études récentes ont démontré que la présence abondante de glace de glacier enfouie entraîne des changements majeurs dans les paysages arctiques à la topographie favorable au déclenchement de mouvements de masse. À l'inverse, il est plutôt question ici de l'évolution morphologique des reliefs plats, tels que les fonds de vallée ou les basses-terres, en réponse à la fonte de cette glace. Nous avons donc étudié une vallée glaciaire à l'aide de carottes de sédiments lacustres, de données bathymétriques détaillées, de données de télédétection et d'observations d'expositions de glace de glacier enfouie.

Nos résultats suggèrent que l'initiation de lacs de thermokarst plus profonds (> 5 m) est liée à la fonte de la glace de glacier enfouie, alors que les lacs peu profonds (< 5 m) ont été initiés par la fonte des coins de glace et du pergélisol riche en glace. Ces deux types de lac vont croître de façon similaire par l'expansion latérale par l'action combinée de l'érosion mécanique et thermique du rivage, ainsi que verticalement par tassement et consolidation au dégel (Bouchard et al., 2020). Lors de l'expansion, certains lacs fusionneront ensuite avec des plans d'eau adjacents pour former de plus grands lacs (Grosse et al., 2013; Hinkel et al., 2005; Bouchard et al., 2020; Prèskienis et al., 2021).

Notons qu'il existe une similitude avec la notion de lacs de kettle qui est connue et documentée en glaciologie et géomorphologie glaciaire (Götz et al., 2018; Benn et Evans, 2010;

Fay, 2002) et qui sont également formés par la fonte de culot de glace enfoui sous des sédiments fluvioglaciaires. Dans les zones déglacées en Arctique, ce terme est aussi souvent utilisé pour désigner des lacs d'origine glaciaire (Saros et al., 2016; Dyke et Savelle, 2000; French et Harry, 1988; Leppi et al., 2016; MacIntyre et al., 2018). Néanmoins, cette définition se limite essentiellement à la fonte de blocs ou culots de glace isolés dans les environnements proglaciaire ou juxtaglaciaire, sans pour autant considérer les interactions avec les formes et processus paraglaciaires et périglaciaires qui se produisent suivant la déglaciation et l'aggradation du pergélisol. Or, ce chapitre démontre que ces lacs vont éventuellement évoluer dans un contexte périglaciaire où l'expansion se poursuivra non seulement par la fonte de la glace de glacier qui demeure préservée l'environnement sédimentaire immédiat, mais également par la fonte de la glace intrasédimentaire (glace de ségrégation, coins de glace). Par conséquent, le terme 'lac de kettle' ne prend pas en compte l'évolution complète et de l'ensemble des étapes de développements de ces lacs. Ainsi, cette étude préfère l'utilisation du terme de «*glacial thermokarst lake*», introduit par Jorgenson et Osterkamp (2005), pour mieux refléter l'initiation et l'évolution de ces types de lacs dans le contexte périglaciaire.

De plus, cette étude démontre également que les lacs arctiques, au-delà d'être abondant, sont caractérisés par une diversité morphologique causée par des facteurs locaux comme la topographie, mais surtout les conditions de pergélisol, en particulier la teneur et le type de glace. L'abondance potentielle de glace de glacier enfouie dans les régions arctiques contribue à cette diversité, mais elle demeure peu considérée en comparaison avec d'autres types de glace de sol. Cette hétérogénéité spatiale des conditions de pergélisol génère une diversité dans les propriétés géomorphologiques, physicochimiques et biologiques de ces lacs, ce qui contribue inévitablement à complexifier les prédictions et estimations de certaines tendances, dont celles des émissions de gaz à effets de serre (GES).



## **5.2 Limitations et recommandations pour de futurs travaux**

### ***5.2.1 Limitations associées à l'échantillonnage***

La nature éphémère des expositions de glace de glacier a limité les occasions d'échantillonnage de la glace et des sédiments encaissants. Ce fut particulièrement le cas pour la glace découverte sur le plateau et qui n'a été accessible que quelques semaines avant d'être complètement enfouie de nouveau sous une épaisse couche de colluvions. La glace n'a pas été exposée de nouveau les étés suivants et il n'a donc pas été possible d'obtenir de nouveaux échantillons pour valider certaines hypothèses et raffiner certaines observations. Il a été notamment impossible de procéder à l'analyse paléomagnétique des sédiments recouvrant directement la glace puisque la séquence sédimentaire était perturbée par des épisodes successifs de mouvement de masse.

### ***5.2.2 Critères diagnostiques pour l'identification de la glace de glacier***

La glace de glacier enfouie demeure à ce jour peu étudiée et l'identification de ce type de glace représente souvent un défi pour les non-experts. L'identification de glace basale enfouie peut-être particulièrement difficile dans le cas de la glace basale enfouie puisqu'elle présente souvent des caractéristiques similaires à celle de la glace de ségrégation formée dans le sol au moment de l'aggradation du pergélisol. La revue de littérature présentée dans cette thèse, de même que les résultats présentés dans les chapitres 2 et 3, mettent en lumière les principales propriétés physiques et chimiques de la glace de glacier enfouie. À l'issue de ces travaux, il serait intéressant de développer une clé d'identification pour ce type de glace basée sur les connaissances acquises dans le cadre de ces travaux, mais également dans le milieu de pergélisol en général, mais aussi en milieu glaciaire. Il y a certes eu quelques tentatives pour établir des ponts entre les domaines périglaciaires et glaciaires (Harris et Murton, 2005; Waller et al., 2009, 2012). Cependant, le développement d'un outil d'identification portant spécifiquement sur la glace de glacier, ses propriétés physico-chimiques et les formes de terrain associées et réunissant à la fois les concepts typiquement associés en milieu glaciaire (glaciers couverts, glace morte, moraine à cœur de glace, lacs de kettle) ou au milieu périglaciaire (glace de glacier enfouie, lacs

de thermokarst glaciaire) est nécessaire pour faciliter l'identification de ce type de glace. Il n'existe pour l'instant aucun article scientifique ou chapitre de livre qui résumant ces connaissances. Une meilleure caractérisation de ces corps de glacier facilitera l'utilisation de ceux-ci comme proxy pour les reconstitutions paléoenvironnementales, alors que son potentiel demeure très peu exploité jusqu'à ce jour.

### **5.2.3 Datation de glace**

La datation de la glace est d'ailleurs l'une des principales limitations rencontrées dans le cadre de ce projet. L'obtention d'un âge plus précis, particulièrement dans le cas de la vieille glace étudiée dans le chapitre 3, permettrait d'affiner certaines interprétations, notamment une meilleure corrélation avec des périodes glaciaires documentées sur l'île Bylot et ailleurs dans l'est de l'Arctique canadien. Des techniques de microradiocarbone s'appuyant sur l'analyse du carbone organique particulaire (*particulate organic carbon*; POC) ou du carbone organique dissous (*dissolved organic carbon*; DOC) extrait de la glace sont maintenant de plus en plus utilisées pour obtenir l'âge absolu de la glace de glacier (Fang et al., 2021; Uglietti et al., 2016; Drogg et al., 2007; Hoffmann et al., 2018). Ces méthodes pourraient préciser l'âge de glace enfouie dans la vallée (chapitre 2) et confirmer que la glace enfouie sur le plateau est plus vieille que ~ 50 –60 ka dans le cas où les résultats dépasseraient la limite de sensibilité de la datation  $^{14}\text{C}$ . Il existe somme toute peu de méthodes qui permettent d'estimer l'âge des dépôts glaciaires de la première moitié du Pléistocène.

Dans le cadre de futurs travaux, d'autres méthodes de datation radiométrique ( $^{40}\text{Ar}$ ,  $^{81}\text{Kr}$ , série de l'uranium) pourraient être utilisées pour dater la glace de glacier enfouie présentée dans le chapitre 3 à partir d'analyses de l'air piégé dans la glace (Yau et al., 2016; Buizert et al., 2014; Feng et al., 2019; Tian et al., 2019; Bender et al., 2010). Ces méthodes permettent de dater de la glace dont l'âge est compris entre 50 000 et 1.5 million d'années (Buizert et al., 2014; Loosli et Oeschger, 1969). Les avancées récentes permettent maintenant de réaliser ces analyses à partir de 40-80 kg de glace (Jiang et al., 2012). Il est toutefois rare qu'une telle quantité de glace de glacier enfouie soit accessible sur le terrain, ce qui limite l'utilisation de cette méthode. De plus,

des analyses préliminaires doivent confirmer l'absence de contamination par l'air moderne (Buizert et al., 2014).

La datation par isotopes cosmogéniques des sédiments encaissants ou des sédiments compris dans la glace basale pourrait également préciser l'âge des sédiments et ainsi donner un âge minimal pour la glace (chapitre 3). Parmi l'ensemble des nucléides cosmogéniques (p. ex.  $^{10}\text{Be}$ ,  $^{26}\text{Al}$ ,  $^{36}\text{Cl}$ ,  $^3\text{He}$  et  $^{21}\text{Ne}$ ), le beryllium-10 ( $t_{1/2}=1.4$  Ma) et l'aluminium-26 ( $t_{1/2}=710$  ka) produits in situ dans le quartz à la surface des roches semblent les plus appropriés pour dater les sédiments encaissants sur le plateau compte tenu de leurs longues demi-vies. Ces isotopes pourraient être utilisés pour quantifier l'âge d'exposition des roches situées à la surface du plateau (Briner et al., 2003; Balco, 2011; Ivy-Ochs et Kober, 2008; Margreth et al., 2016; Gosse et Phillips, 2001). L'âge d'exposition obtenu permettrait notamment de vérifier l'hypothèse qui stipule que les plateaux pourraient ne pas avoir été englacés lors des épisodes glaciaires suivant celui associé à la glace basale enfouie. L'absence de masse glaciaire s'écoulant sur le plateau pourrait avoir favorisé la préservation de la glace sur une longue période.

De plus, l'étude du ratio  $^{26}\text{Al}/^{10}\text{Be}$  serait particulièrement intéressante pour estimer l'âge d'enfouissement des sédiments fluvioglaciaires que l'on retrouve à la base des séquences stratigraphiques (Balco et Shuster, 2009; Balco et Rovey, 2010; Refsnider et al., 2013; Corbett et al., 2017; Granger et al., 2006). En effet, les grains de quartz sont initialement exposés à la surface pendant un certain temps et les concentrations en  $^{26}\text{Al}$  et  $^{10}\text{Be}$  reflètent ce rapport ( $^{26}\text{Al} : ^{10}\text{Be} = 6.75 : 1$ ). Lorsqu'enfouis sous la profondeur de pénétration des rayons cosmiques, la production de nucléides cosmogéniques cesse et les deux nucléides se désintègrent (Balco et Shuster, 2009). La limite inférieure de la méthode est estimée à 300 ka, alors que la limite supérieure est estimée à 3–6 Ma, tout dépendant de la concentration initiale en  $^{26}\text{Al}$  (Balco et Shuster, 2009). Cependant, différents processus (p. ex. enfouissement, exhumation, mouvement, couvert végétal, etc.) peuvent se produire suivant la déposition ou l'enfouissement initial, ce qui peut engendrer une surestimation ou une sous-estimation du temps d'exposition de l'échantillon.

Par conséquent, le choix des échantillons serait crucial afin d'obtenir une bonne estimation de l'âge des roches et des sédiments (Gosse et Phillips, 2001).

#### **5.2.4 Élargir la portée spatiale de l'étude des lacs de thermokarst glaciaires**

Ce projet de recherche s'est concentré sur l'étude d'une vallée glaciaire précise en raison de la découverte d'exposition de glacier enfouie, mais aussi en raison des activités de recherche intensives qui s'y déroulent et qui intègrent ou transcendent plusieurs disciplines. Ceci a permis entre autres d'établir des relations entre l'origine des lacs formés dans la vallée et la limnologie, de même que l'étude des sédiments lacustres. L'analyse des propriétés limnologiques (oxygène, température) et des sédiments lacustres est cependant limitée à trois et deux lacs, respectivement. Le modèle conceptuel développé dans cette étude préliminaire s'appuie sur des résultats obtenus pour un nombre limité de lacs et une vallée précise. Bien que des différences aient été observées entre les deux types de glace (*glacial thermokarst lakes* vs. *ice-wedge thermokarst lake*), des travaux supplémentaires sont nécessaires pour confirmer cette différence en étudiant davantage de lacs dans la vallée de Qarlikturvik, ainsi qu'ailleurs sur l'île Bylot et d'autres sites en Arctique. La présence de dépôts glaciogènes et de mouvements de masse actifs ou stabilisés en marge de lacs profonds constitue les premiers indices de la possible origine glaciaire de ces lacs. Par exemple, dans la plaine sud de l'île Bylot, des observations de glissement de terrain combinées aux analyses de densité présentées dans le chapitre 4 suggèrent la présence de glace de glacier enfouie au-delà de la vallée à l'étude. Puis, les travaux réalisés par Côté et Pienitz en 2002 ont également révélé la présence de lacs profonds dans la vallée adjacente dont l'origine serait possiblement associée à la fonte de glace de glacier enfouie. Par conséquent, ces indices ouvrent la porte à la possibilité de tester les hypothèses et le modèle conceptuel proposés dans cette thèse au-delà de la vallée Qarlikturvik. En raison de l'importance des lacs dans les régions arctiques, une meilleure connaissance de leurs propriétés morphologiques est nécessaire pour comprendre comment les processus biogéochimiques locaux peuvent être extrapolés à l'échelle régionale ou globale. De plus, dans un contexte arctique où les conditions froides actuelles favorisent le maintien du pergélisol et donc la préservation des masses de glace de glacier enfouie, les groupes de lacs profonds d'origine glaciaire pourraient être un indicateur de

la présence de glace de glacier enfouie dans les sédiments encaissants au même titre que les glissements de terrain.

### **5.2.3 Modélisation de la répartition spatiale de la glace de glacier enfouie**

Malgré l'abondance potentielle de glace de glacier préservée dans le pergélisol, il existe peu de données sur sa répartition spatiale. Compte tenu de l'étendue et du volume de glace impliqué, la fonte de ce type de glace a une incidence importante sur les infrastructures nordiques et l'évolution des paysages, notamment par les glissements de terrain ou la formation de nouveaux lacs. Il apparaît nécessaire dans un futur proche de mieux définir l'étendue et la répartition spatiale de ce type de glace. Récemment, O'Neill et al. (2019) ont présenté une cartographie à l'échelle nationale de la répartition de la *glace relique* préservée dans le pergélisol en raison des conditions climatiques froides (Figure 7b). La *glace relique* inclut la glace de glacier enfouie et la glace intrasédimentaire formée par ségrégation et/ou injection. Ces prédictions de l'abondance de glace relique sont basées sur la nature et la répartition des dépôts meubles, les limites marines, les limites d'anciens lacs postglaciaires, la répartition géographique des différents biomes et des différents types de pergélisol (c.-à-d. sporadique, discontinu, continu; O'Neill et al., 2019). Ce modèle de la répartition actuelle de la glace de pergélisol demeure encore largement imprécis, bien qu'il s'agisse du seul modèle récent pour le Nord canadien.

La répartition spatiale de la glace de glacier enfouie est étroitement liée à la répartition spatiale et l'épaisseur des dépôts meubles d'origine glaciaire ou fluvioglaciaire. Cependant, il est important de mentionner que les cartes actuelles de la délimitation des unités géologiques de surface et l'estimation de leur épaisseur (p. ex. till en couverture mince vs continue) comporte plusieurs incertitudes. En effet, les conditions de terrains locales et régionales ne sont souvent pas bien reflétées par la cartographie actuelle à l'échelle nationale de la *Géologie des formations superficielles du Canada* (Geological survey of Canada, 2014) et qui est notamment utilisée dans le modèle de O'Neil (2019). Par conséquent, l'abondance modélisée de la glace de glacier enfouie demeure très probablement sous-estimée pour l'est de l'Arctique canadien, notamment sur les îles de Baffin, de Devon et d'Ellesmere. En effet, la succession d'épisodes glaciaires durant le

Pléistocène combinée à la présence actuelle des nombreux glaciers alpins, calottes glaciaires (p. ex. Agassiz et Devon) et champs de glace (p. ex. *Prince of Whales Icefield*, *Northern Ellesmere Icefield*) sur les îles de l'archipel suggèrent une certaine abondance de glace de glacier enfouie dans cette région. Une meilleure connaissance de la géologie de surface, des formes glaciaires et des limites glaciaires permettrait entre autres de préciser les modèles de l'abondance de glace de glacier enfouie. De plus, une cartographie des mouvements de masse associés à la fonte de la glace massive (p. ex. glissement de terrain dû au dégel, détachement de la couche active) et des zones de lacs profonds groupés pourrait également suggérer la présence de ce type de glace (Kokelj et al., 2017a; Segal et al., 2016).

### 5.3 Perspectives

Cette thèse est le résultat d'un effort interdisciplinaire, notamment par des collaborations avec des chercheurs à l'expertise différente et l'utilisation de méthodes empruntées à d'autres disciplines des sciences naturelles. Ainsi, l'ensemble des trois articles scientifiques présentés dans cette thèse mobilisent des connaissances issues de la géomorphologie, géocryologie, géochimie, sédimentologie, glaciologie, paléomagnétisme, limnogéologie, limnologie et la télédétection. Puisqu'aucune méthode n'est à elle seule capable de déterminer avec certitude l'origine des corps de glace massive, il est nécessaire de combiner plusieurs approches, dont la cryostratigraphie et la géochimie (Coulombe et al., 2019; Fritz et al., 2011; Belova et al., 2008; Lacelle et al., 2007; Murton et al., 2005; Ingólfsson et Lokrantz, 2003). D'une part, la cryostratigraphie permet d'obtenir les propriétés physiques de la glace à des fins de comparaison avec les propriétés bien connues de la glace de glacier contemporaine. D'autre part, l'analyse des propriétés géochimiques (isotopes stables et ions) de la glace permet d'examiner l'origine de la glace (météorique vs gel in situ dans le sol) et d'observer des discontinuités avec les unités adjacentes. La composition isotopique et cationique peut aussi être comparée à celle d'autres types de glace de sol ou de glaciers contemporains. De plus, l'hypothèse de l'origine glaciaire peut être renforcée par l'étude du contexte stratigraphique et géomorphologique dans lequel se trouve la glace. Enfin, la fonte de la glace de glacier enfouie engendre des processus thermokarstiques, dont des affaissements de terrain et des glissement de terrain, ce qui

déclenche une cascade d'effets par l'augmentation des apports sédimentaires et géochimiques des milieux terrestres, vers les lacs, les rivières et les océans (Rudy et al., 2017). Plusieurs études ont démontré l'effet de ces glissements de terrain sur les écosystèmes aquatiques, bien qu'ils ne soient pas nécessairement associés à la glace de glacier enfouie (Malone et al., 2013; Kokelj et al., 2013; Thienpont et al., 2020; Colombo et al., 2018; Staniszezwska et al., 2022; Beel et al., 2020; Morison et al., 2019; Mesquita et al., 2010; Lamoureux and Lafrenière, 2009). Comme démontré dans le chapitre 4, la fonte de glace de glacier enfouie mène également la formation de lacs profonds. La morphologie de ces lacs engendre une série d'effets directs et indirects sur les écosystèmes aquatiques, notamment sur les conditions de glace (p. ex. épaisseur; Arp et al., 2011; Jeffries et al., 1996), la disponibilité en lumière et la production primaire (Vadeboncoeur et al., 2008; Daniels et al., 2015), la disponibilité d'habitat pour la faune aquatique (Jones et al., 2009; Murdoch et Power, 2013), de même que le moment et l'ampleur des émissions de gaz à effet de serre (Prèskienis et al., 2021; Matveev et al., 2016; Hughes-Allen et al., 2021). Que ce soit par le déclenchement de mouvements de masse ou la formation de lacs profonds, les conséquences de la fonte de glace de glacier enfouie sont multiples et donc la connectivité entre les différentes disciplines est essentielle pour améliorer notre compréhension du rôle de la glace de glacier enfouie sur la dynamique et de l'évolution des écosystèmes arctiques en réponse aux changements passés et futurs.

Dans le contexte où les changements climatiques risquent d'accélérer le dégel du pergélisol et la glace qui s'y trouve, la glace de glacier enfouie jouera un rôle important dans l'évolution des paysages glaciaires de l'Arctique. Il est clair que ceux-ci sont encore fortement déterminés par leur héritage glaciaire. La fonte accélérée d'une partie de la glace de glacier jusque-là préservée dans le pergélisol pourrait entraîner des ruptures de pente importantes et un tassement de la surface du sol, ce qui engendrerait des impacts significatifs sur la dynamique du paysage du géosystème du pergélisol, les écosystèmes terrestres et aquatiques, ainsi que les infrastructures. Or, l'intensité et le moment où se produiront ces perturbations dépendront en partie des conditions de terrain locales et de l'ampleur et du rythme des tendances climatiques futures dans différentes régions de l'Arctique. Il demeure cependant difficile d'estimer et de

prédire ces impacts, car les conditions de pergélisol et la glace qui s’y trouve demeurent peu connues dans plusieurs régions de l’Arctique. C’est d’autant plus vrai pour la glace de glacier enfouie qui est encore somme toute peu étudiée. En effet, la contribution de la glace de glacier enfouie à la dégradation future du pergélisol, à l’échelle locale et globale, demeure encore incertaine à ce jour. Ceci s’explique principalement parce que sa distribution spatiale, son volume et son taux de fonte sont encore peu connus. La plupart des modèles à grande échelle sous-estiment donc la distribution spatiale et la quantité de glace de glacier enfouie, sous-estimant par le fait même le volume de glace global. Cette sous-estimation du volume de glace augmente l’incertitude dans les projections de la dégradation du pergélisol et les réponses géomorphologiques associées. De plus, le volume élevé de glace associé à la glace de glacier enfouie a un impact important sur le taux de dégel du pergélisol en raison de la chaleur latente nécessaire pour faire fondre la glace, ce qui augmente la résistance thermique et favorise la préservation de la glace de glacier enfouie à long terme. En effet, la présence de vieille glace de glacier enfouie dans le pergélisol datant du début du Pléistocène indique une résistance mécanique et thermique ainsi qu’une résilience du pergélisol face aux réchauffements climatiques qui sont survenus lors des interglaciaires du Pléistocène et de l’Holocène. Par conséquent, ceci suggère la présence de rétroactions thermiques négatives qui ne sont généralement pas prises en compte par les modèles actuels d’évolution du pergélisol.



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# **Annexe I.**

## **Buried remnants of the Laurentide Ice Sheet and connections to its surface elevation**

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