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Insights from the Greenland projects and their applicability to the Forsmark site

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Abstract

This report presents a synthesis of findings from four collaborative research projects – GAP, ICE, GRASP, and CatchNet – conducted fully or partly in West Greenland between 2008 and 2020. These projects investigated glacial and periglacial processes relevant to the post-closure safety of deep geological repositories (DGR) for spent nuclear fuel. While many of the findings have been previously reported in SKB reports or scientific articles, this report presents the first comprehensive summary of findings from these projects. The synthesis provides a basis for applying these insights and discusses their applicability to the Forsmark site in Sweden, where glacial and periglacial conditions are expected to occur within the coming one million years.

The Greenland Analogue Project (GAP) combined extensive fieldwork and modelling to investigate glacial meltwater formation, basal thermal regimes, groundwater exchange through taliks, permafrost interactions, redox conditions and the infiltration of glacial meltwater into the bedrock. A key component was deep drilled bedrock boreholes allowing for direct observations of groundwater pressure, chemistry, and thermal conditions at depths down to 500 meters. Results show that glacial meltwater can infiltrate fractured crystalline rock to depths relevant for a DGR, that the subglacial bed in the study area is predominantly hard rock with a thin sediment layer, and that hydraulic pressures near the ice-bed interface are primarily controlled by the ice overburden pressure (that is, a water column corresponding to approximately 92 percent of the ice thickness). Observations from the 650 m deep DH-GAP04 borehole extending under the ice sheet margin reveal that the groundwater system responds rapidly and sensitively to even slight changes in ice sheet dynamics, including small fluctuations in ice loading and meltwater input.

The ICE project complemented the GAP by focusing on high-resolution ice drilling and borehole instrumentation within a 1 km2 area of the ice sheet. This high-resolution study provided detailed insight into basal thermal regimes, basal water pressures, and spatial variability in the subglacial drainage system. The basal hydraulic measurements from ICE confirm the results from the GAP, i.e., that ice overburden pressure is a representative and accurate description of the basal hydraulic pressure across most of the ice sheet and for most of the year

The Greenland Analogue Surface Project (GRASP) examined the periglacial landscape in front of the ice sheet through long-term monitoring and hydrological modelling of the Two-Boat Lake catchment. The results confirm that taliks can be expected to occur in periglacial landscapes, and that even if the continuous permafrost limits the exchange between the surface system and the deep regional groundwater system, these taliks can act as conduits for both water and elements in both directions. Even if the shallow nature of the active layer – which thaws each summer – limits the interaction between the surface groundwater and soil particles, the weakly developed soils in this young system still results in a relatively high release of weathering products to the lake. The Catchment Transport and Cryo-Hydrology Network project (CatchNet) extended this work to a broader Arctic setting, including linkages between permafrost and hydrological-chemical conditions. One key result, confirmed by field observations, was that taliks can also form beneath larger river valleys.

The Greenland projects provide valuable insights about glacial conditions, but their transferability to a future Forsmark requires careful consideration of similarities and differences between the sites. One key similarity is the geological setting: both locations are underlain by crystalline, low-permeability bedrock with comparable fracture densities. However, there are also important differences in climate evolution, topography and geothermal heat flux. Forsmark is expected to experience more transitions between climate domains than Greenland on timescales of 100 000 to 1 million years, with multiple transitions between temperate, periglacial, and glacial conditions, as well as recurrent periods of submerged (sea-covered) conditions. Topography is significantly flatter in Forsmark compared to the Greenland study area, and it is expected to remain so for the next one million years. Finally, geothermal heat flux is significantly lower in Greenland than in Forsmark.

These similarities and differences influence how processes observed in Greenland can be applied to Forsmark. For instance, due to the higher relief and lower geothermal heat flux in the Greenland study area, the permafrost depths observed there likely represent the upper range of what could be expected in Forsmark under comparable air temperatures. On the other hand, oxygenated waters have not been observed at great depths in Greenland despite persistent glacial conditions, and are therefore not expected to occur in Forsmark in the future. Other processes studied in Greenland are less sensitive to site-specific boundary conditions and can be directly applied to Forsmark. This concerns, for instance, the basal meltwater pressures observed beneath the Greenland ice sheet. The ratio of these pressures to the overburden ice pressure can be used as input to hydrogeological models of glacial conditions also in Forsmark.

Sammanfattning

Denna rapport presenterar en syntes av resultaten från fyra samarbetsprojekt – GAP, ICE, GRASP och CatchNet – som helt eller delvis genomfördes i västra Grönland mellan 2008 och 2020. Projekten undersökte olika glaciala och periglaciala processer som är relevanta i säkerhetsanalyser för slutförvar för använt kärnbränsle. Även om en stor del av resultaten tidigare har dokumenterats i SKB-rapporter eller vetenskapliga artiklar, är detta den första heltäckande sammanställningen av resultaten från dessa projekt. Syntesen är tänkt att utgöra en grund för att tillämpa dessa insikter och diskutera deras relevans för platsen Forsmark i Sverige, där perioder med glaciala och periglaciala förhållanden förväntas förekomma under de kommande en miljon åren.

Greenland Analogue Project (GAP) kombinerade omfattande fältarbete och modellering för att undersöka glacial smältvattenbildning, basala termiska förhållanden, grundvattenutbyte genom talikar, permafrostinteraktioner, redoxförhållanden samt nedträngning av glacialt smältvatten i berggrunden. En central del av projektet var etableringen av djupa borrhål i berggrunden, vilket möjliggjorde direkta observationer av grundvattentryck, kemi och termiska förhållanden på djup ned till 500 meter. Resultaten visar att glacialt smältvatten kan infiltrera sprickig kristallin berggrund till djup som är relevanta för ett slutförvar av använt kärnbränsle, att överytan av berget under inlandsisen i studieområdet huvudsakligen utgörs av berggrund med ett tunt sedimentlager, samt att det basala hydrauliska trycket till stor del styrs av isövertrycket (det vill säga en vattenpelare som motsvarar ca 92 procent av istjockleken). Observationer från det 650 m djupa borrhålet DH-GAP04, som sträcker sig under inlandsisens rand, visar att grundvattensystemet reagerar snabbt och är känsligt även för små förändringar i inlandsisens dynamik, inklusive små variationer i islasten och tillförsel av smältvatten.

ICE-projektet kompletterade GAP genom att fokusera på borrning och instrumentering av nio borrhål inom ett område av inlandsisen som är en kvadratkilometer stort. Denna högupplösta studie gav detaljerad insikt i basala termiska förhållanden, basala vattentryck och den rumsliga variationen i det subglaciala dräneringssystemet. De hydrauliska mätningarna från ICE bekräftar resultaten från GAP, d.v.s. att isövertrycket är en representativ och korrekt beskrivning av det basala hydrauliska trycket över större delen av inlandsisen och över större delen av året.

Greenland Analogue Surface Project (GRASP) studerade hydrologiska och biogeokemiska processer i det periglaciala landskapet framför inlandsisen genom fältundersökningar i avrinningsområdet kring Two-Boat Lake. Resultaten visar att även om den kontinuerliga permafrosten begränsar utbytet mellan ytvatten- och djupa grundvattensystem, så gör den talik som finns under sjön att både vatten och ämnen kan röra sig i båda riktningarna mellan dessa system. En viktig lärdom är även att trots att det tunna aktiva lagret – d.v.s. den del av permafrosten som tinar varje sommar – begränsar interaktionen mellan ytligt grundvatten och markpartiklarna, leder de svagt utvecklade jordarna i detta unga system till att transporten av vittringsprodukter till sjön ändå är relativt hög. Catchment Transport and Cryo-Hydrology Network-projektet (CatchNet) byggde vidare på detta arbete i en bredare arktisk kontext, och tittade framför allt på kopplingar mellan permafrost och hydrologisk-kemiska förhållanden. Ett centralt resultat, bekräftat av fältobservationer, var att talikar också kan bildas under större floder.

Grönlandsprojekten har gett värdefulla insikter om glaciala förhållanden, men att överföra dessa kunskaper till ett framtida Forsmark kräver att hänsyn tas till de likheter och skillnader som finns mellan platserna. En grundläggande likhet är berggrundsgeologin: båda platserna har kristallin berggrund med låg genomsläpplighet och jämförbar spricktäthet. Det finns dock viktiga skillnader vad gäller klimatutveckling, topografi och det geotermiska värmeflödet. För tidsskalor om 100 000 till en miljon år förväntas Forsmark genomgå fler övergångar mellan olika klimattillstånd än på Grönland, med flera övergångar mellan tempererade, periglaciala och glaciala förhållanden, samt återkommande perioder då området ligger under havet. Topografin är dessutom avsevärt flackare i Forsmark jämfört med studieområdet på Grönland, och förväntas förbli så under den kommande en miljon åren. Slutligen är det geotermiska värmeflödet betydligt lägre på Grönland än i Forsmark.

Dessa likheter och skillnader påverkar hur kunskapen om de processer som studerades på Grönland kan tillämpas på Forsmark. Till exempel innebär den högre reliefen och det lägre geotermiska värmeflödet i studieområdet på Grönland att de observerade permafrostdjupen sannolikt representerar en övre gräns för vad som kan förväntas i Forsmark vid jämförbara lufttemperaturer. Att det trots långvariga glaciala förhållanden inte observerat syrerikt vatten på stora djup på Grönland innebär att det inte heller är troligt att syrerikt vatten kommer förekomma på stora djup i berggrunden i det framtida Forsmark. Andra processer som studerats på Grönland är mindre känsliga för platsspecifika randvillkor och kan direkt tillämpas på Forsmark. Detta gäller exempelvis att förhållandet mellan det basala smältvattentrycket och isövertrycket är direkt överförbart till de hydrogeologiska modeller för glaciala förhållanden som utvecklas för Forsmark.

Content

1 Introduction

From 2008 and onwards, several collaborative research projects funded by nuclear waste management organizations were carried out in Greenland, namely the Greenland Analogue Project (GAP), the Greenland Analogue Surface Project (GRASP) and the ICE project. The Catchment Transport and Cryo-Hydrology Network project (CatchNet) utilize Greenland as one of several study sites and is a larger collaborative effort also funded by nuclear waste management organizations. The overarching aim of these projects was to enhance scientific understanding of what processes are active under a cold climate that would be relevant for the post-closure safety of deep geological repositories (DGR) for spent nuclear fuel. While the projects shared a common overarching purpose, they were conducted to address different more specific research questions. The primary objective of GAP, ICE and parts of CatchNet was to advance the knowledge on glaciation-associated processes related to an inland ice sheet and their implications for post-closure safety of a DGR in crystalline bedrock. For GRASP, and a part of CatchNet, the primary objective was to increase the understanding of the periglacial landscape (i.e. a landscape that contains permafrost but without the presence of an ice sheet), and how permafrost and taliks affect the transport and accumulation of water and elements across this landscape.

Results from the projects have been documented in several reports and scientific publications, however, no synthesis has been made regarding how the insights from the Greenland projects relate to a future glacial or periglacial environment in Forsmark; this report summarizes the knowledge obtained and how it can be utilized for that purpose. However, the report does not address how the findings should be applied in assessments of post-closure safety. As the GAP and ICE projects were extensively reported earlier (Harper et al. 2016, Claesson Liljedahl et al. 2016 and references therein), they are summarized more briefly here than the GRASP, which is presented in greater detail.

2 The Greenland projects

2.1 Specific aims of GAP

To advance the understanding of glacial hydrogeological processes, GAP research activities included extensive field work and modelling studies of a sector of the Greenland ice sheet including the proglacial landscape[1]. GAP was funded by the nuclear waste managements in Sweden, Finland and Canada, and carried out between 2008 to 2022. The project focused on three main project areas: 1) surface-based ice sheet studies; 2) ice drilling and direct studies of basal conditions; and 3) geosphere studies. The field investigations were carried out across the GAP study site (Figure 2-1B) designed to contribute towards improved scientific process understanding of attributes relevant to assessing post-closure safety of a DGR. Specifically, the following conditions and processes were investigated:

- transient meltwater processes on the ice sheet surface,
- basal thermal distribution, basal water pressure and generation of water at the ice sheet bed,
- hydraulic boundary conditions for groundwater simulation,
- permafrost[2] and its control of water flow,
- the effect of taliks on the exchange of water between deep and shallow groundwater systems,
- melt-water end-member compositions,
- depth of glacial meltwater infiltration
- redox stability of the groundwater system.

The surface-based ice-sheet studies aimed to improve the current understanding of ice-sheet hydrology and its relationship to subglacial hydrology and groundwater dynamics. This work relied primarily on indirect observations from the ice sheet surface of the basal hydrological system to obtain information from the parts of the ice sheet that contribute with water for groundwater infiltration. Project activities included quantification of ice-sheet surface-water production, as well as an evaluation of how water is routed from the ice surface to the interface between the ice and the underlying bedrock.

Ice drilling and direct studies of basal conditions also aimed to improve understanding of ice-sheet hydrology and groundwater formation based on direct observations of the basal hydrological system, paired with numerical ice sheet modeling. The investigations were done to specifically investigate: (i) the thermal conditions within and at the base of the ice sheet, (ii) generation of meltwater at the ice/bedrock interface, and (iii) hydrologic conditions at the base of the ice sheet. Activities included ice drilling of multiple holes at three locations on the ice sheet, at distances up to 30 km from the ice sheet terminus, to assess drainage, water flow, basal conditions, and water pressures at the interface between the ice and bedrock.

[1] Proglacial conditions refer to a landscape that is directly affected by an ice sheet including meltwater. This is different from a periglacial landscape, which contains permafrost but without direct connection to an ice sheet.

[2] Defined as ground that remains at or below 0 °C isotherm for at least two consecutive years, see Section 3.4.1 in SKB (2020). However, for this this report we do not make any distinction between permafrost and perennially frozen ground, and permafrost therefore always implies the presence of ground ice (unless specifically stated that it doesn't).

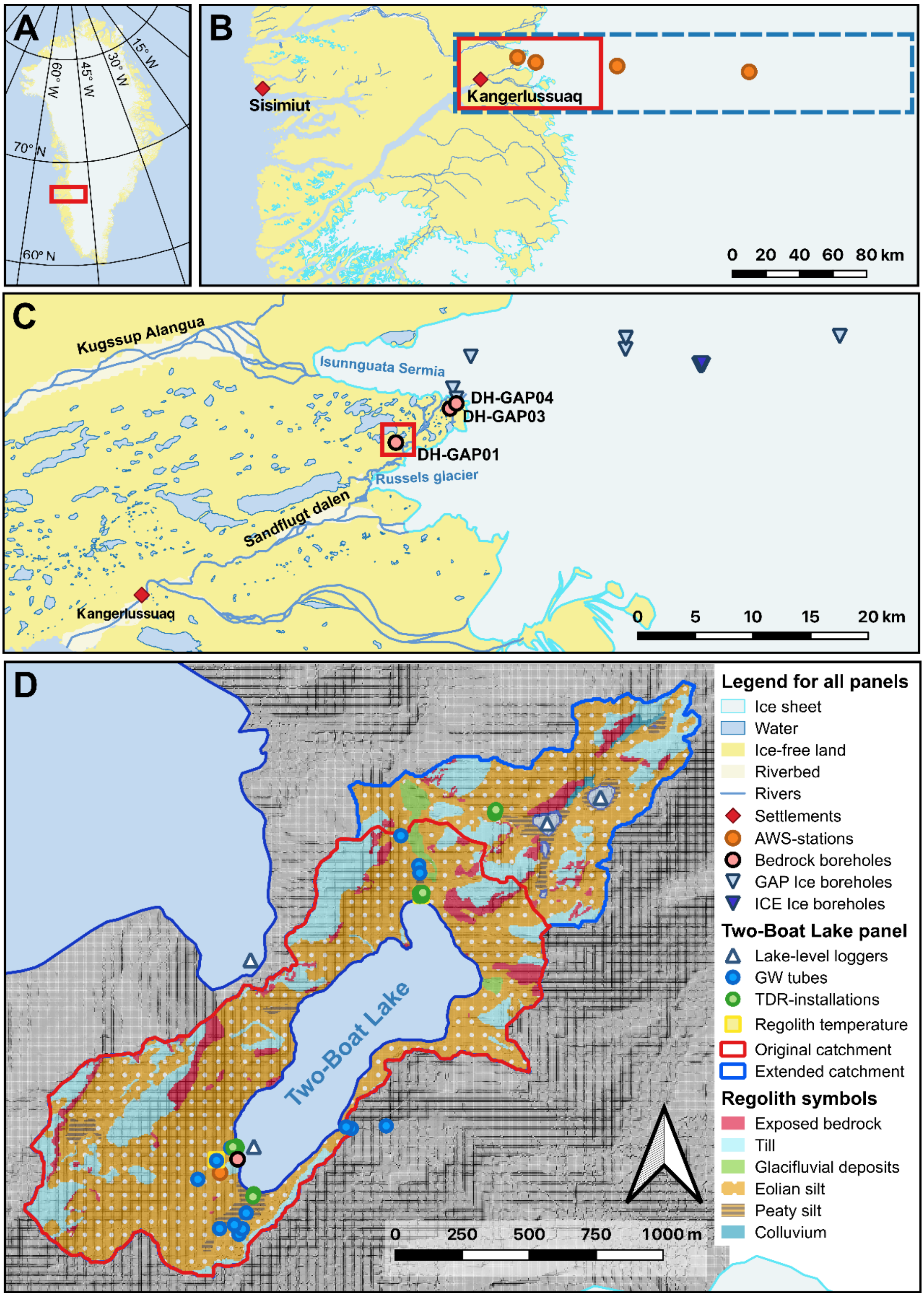

Figure 2-1. *Map showing the location of the GAP and GRASP study sites in West Greenland (A). The upper right panel (B) shows the entire Kangerlussuaq region from the coast to the ice sheet. The blue square (dashed line) mark the extent of the GAP-study site, and the orange dots mark the automated weather stations (AWS). The center panel (C) shows the bedrock and ice boreholes that were drilled during the ICE and GAP-projects. The lower panel (D) shows the Two-Boat Lake catchment, with a selection of the installations that were used during the GAP and GRASP-projects. The red squares in panel A-C indicate the approximate extent of the map in the next panel.*

Geosphere investigations focused on groundwater flow dynamics and the chemical and isotopic composition of water at depths of 500 meters and more below ground surface, including evidence on the depth of permafrost, redox conditions, and the infiltration of glacial meltwater into the bedrock. Three deep and inclined bedrock boreholes (DH-GAP01, DH-GAP03 and DH-GAP04) were drilled through the permafrost in the vicinity of the ice sheet margin to aid these studies (Figure 2-1C). Two of these boreholes (DH-GAP01 and DH-GAP04) were hydraulically tested and instrumented to allow hydrogeologic and hydrogeochemical monitoring. DH-GAP04 was drilled at the ice sheet margin in an angle such that the end of the hole extends just beneath the margin. This borehole features three packed-off sections and is currently the only borehole in the Arctic that enables active monitoring and sampling of deep groundwater beneath permafrost. DH-GAP01 features one packed off section and was drilled at an angle from the lakeshore allowing it to extend beneath the lakebed (Two-Boat Lake). Utilizing DH-GAP01 in combination with other studies, the nature of ground conditions under this proglacial lake (i.e. the Two Boat Lake) was investigated, to assess if areas of unfrozen ground within the permafrost (taliks) may act as a potential pathway for exchange of deep groundwater and surface water.

2.2 Specific aims of ICE

Following the GAP project, key uncertainties in basal boundary conditions remained, prompting further investigations. New scientific questions emerged regarding temporal and spatial variations in water pressure at the ice-bed interface, which are essential for groundwater flow modelling under glacial conditions. To refine the understanding and resolve these gaps, the ICE project was established and carried out between 2015 and 2018. The project utilized an approximately 1 km^3 large ice volume within the ice sheet to study ice/water/earth interactions and integrated existing GAP data with new ice borehole measurements from the densely investigated ice volume.

Continuing the GAP's ice drilling work, this project aimed to 1) analyze the ice sheet bed's physical structure, 2) measure basal water pressure and gradients, 3) check for fast water pressure changes in the basal drainage system, and 4) improve the understanding of water distribution at the bed. The data from the ice block study was compared with the ~50 km transect of GAP observations (Figure 2-1C)

2.3 Specific aims of GRASP

In contrast to GAP, which focused on the ice sheet itself, GRASP focused on the deglaciated landscape in front of the ice sheet. That is, a periglacial landscape that no longer is directly affected by glacial processes, but that is underlain by permafrost that restricts the exchange of water between the surface system and the deep groundwater system. GRASP was funded by SKB, and started in 2010. The project officially ended in 2019, but fieldwork has continued within the CatchNet-project and as of 2025 several installations (including the automated weather station) are still in operation. In essence, GRASP consisted of two interlinked parts. The first part was aimed at understanding the hydrology in the active layer above the permafrost and how it connects to the deep groundwater system. A specific aim was to determine if the studied lake had a through talik, and if it did, whether groundwater was recharging or discharging through the talik. The second part was aimed at understanding the biogeochemical cycling of elements in the periglacial landscape. Both parts relied heavily on site-specific data collected in the catchment of Two-Boat Lake, i.e. a small lake situated in front of the ice sheet where one of the GAP boreholes was located (Figure 2-1C, D).

To facilitate the hydrological modelling, GRASP collected data on meteorological conditions, soil moisture, ground temperature, ground water levels and lake level. An extensive effort was also put on mapping the catchment geometry (topography, lake bathymetry, regolith type and thickness, vegetation) and determining the hydrological properties of the regolith. These data were then used to develop and validate a fully distributed, numerical, hydrological model for the entire catchment.

For the biogeochemical part of the project, the general sampling strategy was to get a broad picture of the distribution of a wide selection of elements, rather than to focus on a specific target element. Samples for chemical analysis were collected from all major catchment pools (vegetation, regolith, soil, lake sediment), as well as all water types (rain, snow, groundwater and lake water). The samples were analyzed for carbon, nitrogen, and phosphorus, as well as total concentrations of an extensive list of major and trace elements. When coupled with the output from the hydrological modelling, site-specific estimates on biomass, primary production and sediment accumulation rates, the chemical data could be used to develop mass-balance models that describes the input, transport, and accumulation of elements in this periglacial landscape.

2.4 Specific aims of CatchNet

CatchNet is a collaborative project between nuclear waste management organizations in several countries (including SKB, COVRA - Netherlands, NWMO – Canada and BGE - Germany). It was initiated in 2019 and was set up as a continuum from the GAP and GRASP projects with the aim of solving some of the questions that arose during those projects. It has three research packages: RP1 – Connecting the glacial and sub-glacial hydrology with the permafrost hydrological system on a landscape scale; RP2 – Permafrost transition periods; and RP3 – Biogeochemical cycling in permafrost areas.

As the name implies, the primary aim of RP1 was to assess how surface hydrology is connected to the deep hydrological system. Major questions remaining after GAP/GRASP was how common taliks are in the area, what fraction of the meltwater from the ice that leaves as surface runoff compared to subglacial infiltration and deep groundwater flow, and how the seasonal dynamics in the coupling between deep groundwater and the surface system looks. For RP2 the aim was to get a better understanding about what happens when a system transitions from permafrost to a system without permafrost, and how this affects primarily the hydrology and hydrological pathways. In RP3 the intention was to couple the hydrological output from modeling to the chemistry of the water for catchment in different climatic settings (from the Arctic to the boreal region). Unlike the GAP and GRASP projects, CatchNet is not limited to Greenland and the Kangerlussuaq region. Instead, it utilizes sites also from other areas (e.g., Svalbard and Arctic Canada).

3 Present conditions in the Greenland study region

This chapter outlines the present conditions in the Greenland study region. First, the environmental conditions are described on a broader regional scale (Section 3.1), after which the environmental settings at a local scale corresponding to the GAP and ICE sites (Section 3.2) and the GRASP site (Section 3.3) are summarized.

3.1 Regional setting

The study sites are situated in the Kangelussuaq region in West Greenland, which extends from the coast into the ice sheet (Figure 2-1A, B). The landscape of the region is characterized by a gently undulating hilly terrain with fjords and broad, gently sloping valleys trending in a WSW-ENE direction. The total elevation spans from sea level to more than 600 m a.s.l. near the ice margin. The relief varies from a few tens of meters in the river valleys to peaks reaching 200–300 m above the river valley floor. Lakes are a common feature in this recently deglaciated landscape, and the region contains more than 20 000 lakes of various sizes (Andersson et al. 2009).

The bedrock is dominated by Archean gneisses, amphibolite, and metasedimentary rocks, that has been repeatedly reworked. On a larger scale, fault zones and lineaments follow a WSW-ENE direction, and lakes are often found along such structural lineaments (van Gool et al. 1996, Engström and Klint 2014). Higher elevations feature glacial deposits such as till, moraines and glaciofluvial material, with areas closer to the ice sheet covered by silty eolian sediments (Petrone et al. 2016). River valleys, on the other hand, are characterized by glaciolacustrine and glaciofluvial deposits, while the fjord head contains Holocene-age terraces composed of glaciofluvial and marine sediments (Storms et al. 2012).

Regolith thickness varies across the region, ranging from thin or absent layers to depths exceeding 80 m in river valleys. Near Kangerlussuaq village, sediment deposits typically reach around 30 m in thickness (Storms et al. 2012). Elevated areas often have a sparse till cover, which is either significantly eroded by wind or entirely absent (Claesson Liljedahl et al. 2016). Due to the arid climate and the availability of fine-grained sediments, diverse eolian deposits are prevalent throughout the region (Willemse et al. 2003). The major valleys hosting meltwater rivers in the region are Kugssup Alangua and Sandflugtdalen, which all follow a WSW-ENE direction from either the Sisimut or Kangerlussuaq fjord to the ice sheet (Figure 2-1B, C). The Watson River extends from the Russell and Leverett outlet glaciers and flows through Sandflugtdalen to the head of the Kangerlussuaq fjord. Most of the meltwater from Russell Glacier and the expansive region south of its terminus drain to the Kangerlussuaq fjord while most meltwater from the terminus of the Isunnguata Sermia is conveyed through Kugssup Alangua to the Sisimiut Sorata fjord. This indicates a regional groundwater divide between the Russell Glacier and Isunnguata Sermia. The rivers are mostly frozen from October/November until April/May, maximum discharge occurs from June to August. Abrupt drainage of ice dammed lakes is a recurring phenomenon along the river valleys (Russell et al. 1990, 2011; Mikkelsen et al. 2013).

During the last glacial maximum, around 20 000 years (20 kyr) before present (BP), the front of the ice sheet reached all the way to the coast (van Tatenhove et al. 1996, Young and Briner 2015). Since then, the ice has been in a receding phase, interrupted by temporary stillstands and readvances. During the Holocene peak thermal maximum, ~8500 years BP, the ice front receded up to 10–20 km further inland compared to its present position. Following the cooling of the climate it has then re-advanced, and it reached its current position about 5000-7000 years BP (Young et al. 2020).

The climate in the study region varies from a maritime, relatively warm and wet, climate at the coast to a much drier and colder, continental, climate close to the ice sheet where cold katabatic winds blows from the ice (Steffen and Box 2001, van den Broeke et al. 1994). The mean annual surface air temperature (MAAT) in Kangerlussuaq is –5.1 ºC (measured between 1977 and 2010; Cappelen 2012), with sub-zero monthly mean temperatures typically prevailing from October to May. Winter minimum temperatures can reach as low as –40 ºC, while summer temperatures frequently exceed 20 ºC. Mean annual ground temperature (MAGT) in Kangerlussuaq is –2 ºC, measured at a depth of 1.25 m below ground surface (van Tatenhove and Olesen 1994). Average annual precipitation for the Kangelussuaq station is 173 mm yr^{-1} (1977-2010; data corrected for wind and adhesion losses), with ~40 % falling as snow (Cappelen 2012, Johansson et al. 2015a,b).

The region is situated in the southern part of the continuous permafrost zone (>90 % spatial coverage; Brown et al. 1997). At the coast the permafrost is discontinuous (between 50 and 90 % areal coverage), but further inland it becomes continuous and at the Kangerlussuaq airport it has been modelled to be 100–160 m thick (van Tatenhove and Olesen 1994). At locations near the ice margin, and at higher elevations, the thickness of permafrost is estimated to be around 350–400 m (Christiansen and Humlum 2000). In the region from Kangerlussuaq to the ice sheet the thickness of the active layer –the upper layer of the ground that thaws each summer – varies from 0.15 to 5.0 m (van Tatenhove and Olesen 1994). Periglacial features, including patterned ground and ice-wedges, are observed, along with rock outcrops showing honeycomb weathering (Aaltonen et al. 2010).

3.2 The GAP and ICE study sites

The GAP study site spans 200 km from east to west and 60 km from north to south, extending from the Kangerlussuaq village to about 160 km into/onto the ice sheet (Figure 2-1B). Approximately 70% of the study area is covered by the ice sheet and reaches from the lower ablation area to the accumulation area.

Ice thickness in the study area generally increases towards the ice divide. The maximum ice thickness of the study area is around 1500 m with a mean thickness of 800 m. The bedrock topography is highly variable resembling the landscape in front of the ice sheet (Lindbäck et al. 2014). In places, the bedrock beneath the ice sheet reaches hundreds of meters below sea level (Lindbäck et al. 2014, Lindbäck and Pettersson 2015). The general direction of ice flow in the area is from east to west, i.e. from the ice-sheet interior toward the margin, with a mean surface velocity of approximately 150 m yr^{-1} (Joughin et al. 2010). The area includes the outlet glaciers Isunnguata Sermia, Russell, Leverett, Ørkendalen and Isorlersuup and their catchment areas (Figure 2-1C). These outlet glaciers are all land terminating, which means they are isolated from marine influences and any changes in ice dynamics are directly attributable to surface-melt forcing (Fitzpatrick et al. 2014). Since the 1990s, the mass balance of the Greenland ice sheet has turned negative (van Angelen et al. 2012; van Angelen et al. 2014; Colgan et al. 2019), meaning that the loss of ice through melting and calving has on average been greater than the ice accumulation (e.g. Hanna et al. 2008, Ettema et al. 2009, Fettweis et al. 2011, van As et al. 2012).

Although findings from an electromagnetic sounding survey conducted a few kilometers east of the GAP boreholes suggest that subglacial permafrost extends at least 2 km east of the ice margin (Ruskeeniemi et al. 2018), there is currently a lack of available data specifying its precise extent. Considering an approximate permafrost depth of approximately 300 to 350 m in the study area, lakes that remain unfrozen at the bottom (Hartikainen et al. 2010) and have a diameter of approximately 360–420 m is considered required to maintain through taliks (Claesson Liljedahl et al. 2016).

To complement the investigation of sub-glacial conditions conducted in the GAP through a series of ice-drilled boreholes along a transect spanning the study region, the ICE study site was established (Harper et al. 2019). This site featured a 700 x 700 x 700 m ice block situated within the GAP study area, approximately 33 km inland from the Isunnguata Sermia terminus. The site exhibited a flat, smooth bed sloping upward in the flow direction, bordered by bedrock troughs to the north and south and covered by over 1000 m of ice (Figure 2-1C). The vertical temperature profile within the ice block aligns with expectations, with the coldest point located roughly midway through the ice depth (Hills et al. 2017). At the base, the ice is at the pressure melting point temperature, with a thin (~5 m) layer of temperate basal ice. Below this, a thin (less than a few decimeters thick) and discontinuous sediment layer rests atop a hard bedrock surface.

Details about the GAP study site and investigations can be found in e.g. Claesson Liljedahl et al. (2016) and Harper et al. (2016). Harper et al. (2019) provides a description of the ICE field investigation site.

3.3 The GRASP site

The GRASP study-site, located 25 km east of Kangerlussuaq, consists of a lake and its catchment (Figure 2-1C, D). We refer to the lake as Two-Boat Lake (Johansson et al. 2015a,b, Lindborg et al. 2016, Petrone et al. 2016, Rydberg et al. 2016), but others have also referred to the lake as SS903 (Sobek et al. 2014) and Talik lake (Harper et al. 2011). Two-Boat Lake is situated about 500 m from the Greenland ice sheet, but despite this proximity, it receives no surface meltwater from the ice sheet. The lake has an area of 0.37 km^2, and a terrestrial catchment area of 1.2 or 1.7 km^2 depending on how it is defined. Based on the movement of surface water during the period from late June to September, early publications reported the catchment area as 1.2 km^2 (Johansson et al. 2015b, Lindborg et al. 2020, Petrone et al. 2016, Rydberg et al. 2016). Later observations, made in late May to early June, revealed that an additional area northwest of the originally defined catchment area also contributes to the runoff to the lake during the snowmelt period when temporary streams appear in the catchment (i.e., late May to mid-June; Figure 2-1D). Redefining the catchment boundary gives a total terrestrial area of 1.7 km^2, and a total catchment area of 2.07 km^2. The added northeastern part (0.5 km^2) is, however, hydrologically disconnected from the main catchment for most of the year and only contributes to terrestrial runoff until the deeper sections of the active layer become unfrozen in mid-June.

The Two-Boat Lake catchment can be characterized as hilly, with the lake level at 335 m above sea level, and the highest peaks reaching up to 500 m above sea level. When the ice sheet retreated from the site, it left the bedrock covered by a layer of till and glaciofluvial material (10–12 m thick; Petrone et al. 2016). Except for exposed ridges, these glacial deposits have subsequently been covered by a layer of eolian silt (0-1 m, thicker in depressions; Petrone et al. 2016). In wetter, low-lying, areas where organic matter production exceeds decomposition, the soils have a high organic content (up to 50 %) and has been classified as peaty silt (Petrone et al. 2016; Figure 2-1D). Carbon-14 (^{14}C) ages of dated material from lake sediments and peaty silt deposits suggest that the Two-Boat Lake catchment became ice-free at least 4000 years ago (Lindborg et al. 2016, Rydberg et al. 2016), which is consistent with regional studies regarding the movements of the ice front (van Tatenhove et al. 1996, Young and Briner 2015, Young et al. 2020).

Between 2010 and 2019, GRASP included an extensive monitoring program in the Two-Boat Lake catchment (Johansson et al. 2015a,b). This included the establishment of an automated weather station, ground temperature loggers, lake level loggers and ground moisture measurements using Time Domain Reflectory (TDR) equipment. For the period 2011-2018, the MAAT at Two-Boat Lake was –4.5 ºC, and the average MAGT at 1.25 m depth was –2.9 ºC (Johansson et al. 2015a). Annual precipitation for the same period varied between 160 and 410 mm, with an average of 285 mm (corrected for wind and adhesion losses; Johansson et al. 2015a, b), i.e., approximately 1.6 times higher than in Kangerlussuaq. About 40 % of the annual precipitation falls as snow, and the potential evapotranspiration is about 400 mm yr^{-1} (Johansson et al. 2015a,b).

Except for areas with exposed bedrock and in wind-deflation scares on ridgetops, the catchment has a sparse but continuous vegetation cover. There are no trees in the catchment, and dwarf shrubs – mostly *Betula nana* and various Salix species – rarely exceed heights of 0.5 m. Based on aerial photographs and field surveys, the terrestrial vegetation was divided into four main types, i.e., dwarf shrubs, dry grassland, meadow, and wetland (Petrone et al. 2016). During the study period, the biomass, net primary production (NPP) and carbon pools were determined for each vegetation type (Lindborg et al. 2020). Largest biomass was found in dwarf shrub (2.47 kg C m^{-2}), while the meadows had highest NPP (217 g C m^{-2} yr^{-1}), and the wetland had the largest carbon pool (36.2 kg C m^{-2} down to 100 cm). The thickness of the active layer depends on soil moisture and follows the vegetation types. Wetland areas have an active layer of about 0.5 m, dry grassland/dwarf shrub/meadow 0.7–0.8 m and bedrock outcrops around 2.0 m (Petrone et al. 2016). Herbivores in the area includes muskox, reindeer, hares, and a large variety of nesting and migrating birds (e.g., *Anser albifrons flavirostris* and *Branta canadensis interior*).

According to the hydrological model most of the water from precipitation leaves the Two-Boat Lake catchment either as evapotranspiration in the terrestrial system or evaporation from the lake surface (Johansson et al. 2015b). Considering the terrestrial system, evapotranspiration amounts to, on average, 65% of the annual precipitation for the 2011-2018 period (Rydberg et al. 2025). This results in a limited annual terrestrial runoff of, on average, 98 mm yr^{-1} for the 2011-2018 period, and the catchment lacks permanent streams. Temporary streams do appear during high flow situations, primarily during the snowmelt period (i.e., late-May to mid-June) when about half of the annual terrestrial runoff occurs. The limited input of water from the terrestrial system, together with the high evaporation from the lake surface, results in a small outflow of water from the lake (7 mm yr^{-1}; Johansson et al. 2015a,b). A limited export of water through the lake outlet is also consistent with observations of the lake-water levels which did not reach the threshold during the 2011-2019 study period (but surface outflow was observed in 2009 and 2023).

Two-Boat Lake itself has an average water depth of 11.3 m, a max depth of 29.9 m. and a volume of 4.29×10^6 m^3. The limited terrestrial input of water results in a hydrological residence of 47 years (Johansson et al. 2015b). A relatively long residence time is consistent with the measured conductivity of the lake (130-140 μS cm^{-1}), which is higher than in lakes with a higher throughput of water, but lower than for the closed basin, oligo-saline, lakes closer to Kangerlussuaq (Ryves et al. 2002, Sobek et al. 2014, Lindborg et al. 2016). Because the main export of water from the lake occurs as evaporation, and any solutes that enter the lake will remain in the lake water column for an even longer time than the water molecules (up to 300 years). However, the chloride concentration (3.5 mg L^{-1}), is much lower than many of the oligo-saline lakes closer to Kangerlussuaq. This implies that even if no outflow of water from the lake was observed during the study period, it does occur frequently enough to prevent the lake from becoming saline (as mentioned above surface outflow was observed in 2009 and 2023).

The lake water is clear, with light demanding macrophytes – primarily *Chara* sp. and different Bryophytes – covering the lakebed down to about 14 m water depth. This high transparency implies that most of the DOC (dissolved organic carbon; 8.4 mg L^{-1}) in the lake water consists of non-colored, photobleached, organic material. Even if there is considerable autochthonous production in the lake (by macrophytes and algae), the carbon balance of the lake shows that it is net heterotrophic, and that there is a net release of CO_2 from the lake surface. At shallower depths, especially from 3 to 5 m, the lakebed is partially covered by nitrogen fixing *Nostoc* sp. colonies, which indicate that the lake is nitrogen limited. The main pelagic macro fauna is the tadpole shrimp (*Lepidurus arctica*) and fairy shrimp (*Branchinecta paludosa*), and no fish have been found in the lake (Lindborg et al. 2016).

More details from the GRASP studies can be found in two doctoral theses (Johansson 2016, Lindborg 2017), and a series of scientific publications (Johansson et al. 2015a,b, Lindborg et al. 2016, Rydberg et al. 2016, Petrone et al. 2016, Lindborg et al. 2020 and Rydberg et al. 2023).

4 Lessons learned from the Greenland projects

This chapter explores the lessons learned throughout the Greenland projects, providing descriptions of achievements, challenges, and key insights. The lessons are presented from a process and system perspective, highlighting the interconnected nature of various components within the system. To structure these findings effectively, they are organized into sub-chapters, each focusing on specific conditions and processes. Given the complex interplay between processes in the cryosphere and geosphere, some overlap between sub-chapters may occur. This reflects the complex behavior of for example water – both shaped by and influencing environmental, physical and structural geological factors.

4.1 Ice sheet processes

In an ice sheet setting, basal water pressure, thermal distribution at the base of the ice sheet, and groundwater formation interact in a dynamic feedback loop. Basal water pressure, governed by the overburden pressure[3] of the ice and the presence of meltwater, facilitates the flow of water, which in turn impact basal sliding and ice movement. The basal thermal distribution, varying from frozen to temperate conditions due to geothermal heat flux, frictional heat from ice movement and basal ice deformation/temperature, determines whether water exists as liquid or is solid/frozen at the ice sheet base. As basal ice melts, the meltwater can infiltrate the bedrock, promoting groundwater formation, which also is influenced by the hydraulic properties of the subglacial environment. Drainage system processes at the ice-bed interface redistribute and focus water at the glacier bed, creating a water pressure field that is spatially and temporally variable. Together, these factors create complex interactions where increased water pressure can lead to more melting, affecting thermal conditions and facilitating groundwater recharge, ultimately impacting gradients, ice sheet stability and dynamics over e.g. the course of a glacial cycle. Sections 4.1.1 to 4.1.4 describe what we have learned about these processes through the GAP and ICE field studies and numerical modeling.
Physical conditions of the ice sheet bed

The physical condition of the basal boundary of an ice sheet significantly influences water storage and flow at the ice-bed interface. Bed conditions are crucial for various ice sheet processes, including ice motion mechanics, heat transfer, and the transport of sediment and chemical substances to the surrounding environment including oceans. These conditions are also relevant for subglacial hydrology.

Bed conditions range from ‘soft’ beds to ‘hard’ beds. A ‘soft’ bed is characterized by thick continuous sediment layers (e.g. Alley et al. 1986), where slow water flow occurs through porous sediment deposits. A system of channels in the top of the sediment layer can also exist that facilitates relatively fast water flow in certain areas (e.g. Walder and Fowler 1994). A 'hard' bed is characterized by ice resting directly on bedrock (e.g. Weertman 1972).

If surface glacial meltwater reaches or is produced at the bed, a subglacial drainage system forms at the interface between ice and rock. On a hard bed, water flow pathways emerge from the separation between the ice and bed caused by ice sliding over bumps, and from the melting of the ice ‘roof’ due to heat generated by the water flow. This can result in fast water flow, especially when large channels develop.

Modeling ice sheet motion often assumes a soft sediment bed (Aschwanden et al. 2016, Bougamont et al. 2014), while variations in melting and surface velocity are linked to hard bed conditions (Bartholomew et al. 2012, Cowton et al. 2013, Doyle et al. 2015, Sundal et al. 2011). Understanding the subglacial framework is crucial for interpreting basal hydrological and motion processes.

Seismic experiments and borehole drilling in Western Greenland have revealed diverse bed conditions, from lodged till (Clark and Echelmeyer, 1996) to thick coarse-grained sand layers (Lüthi et al. 2002). Remarkable sediment thicknesses of 100–200 m have been suggested in areas north of Kangerlussuaq (Walter et al. 2014; Ryser et al. 2014). Further south of the Kangerlussuaq region of the ice sheet, till with an unknown thickness has been inferred (Booth et al. 2012, Dow et al. 2013, Walter et al. 2014). While observations of bed conditions are sparse and can often be ambiguous, multiple locations along the flanks of west/central Greenland have been interpreted to have a sediment covered bed.

[3] Overburden is defined as the water pressure resulting from the weight of the ice resting on subglacial waters. Because the density of ice is less than water, overburden pressure expressed in metres of water equivalent is equal to about 92% of the ice thickness.

What have we learned?

Interpretations from in situ drilling and borehole experiments along a flow line of a part of the ice sheet suggested that the bed at the GAP study sites was a hard bed consisting mainly of bedrock (Harper et al. 2016). The ICE project expanded the GAP data set, adding observations from a site with closely spaced boreholes using advanced methods to refine this interpretation. The ICE findings support the initial GAP concept, indicating a thin and discontinuous sediment veneer, unlikely to exceed a few decimeters in thickness (Harper et al. 2017). This 'hard' bed framework is consistent across all drilling sites. The hard bed setting provides a useful framework for interpreting ICE and GAP observations, yet basal conditions exhibit considerable variation. A combination of exposed bedrock and thick sediment deposits likely exists across spatial scales ranging from kilometers to smaller localized areas.

At other Greenland sites, including those near GAP and ICE, thicker sediment layers have been identified. Large sediment influx from Isunnguata Sermia indicates that the available material is transported and accumulates within deep bedrock troughs, functioning as line sinks. This process is reinforced by hydraulic gradients that direct water flow into these depressions (Harper et al. 2016, Harper et al. 2019).

Before the GAP and ICE projects, the general thought was that soft bed conditions prevailed under the Greenland ice sheet. Findings have revealed a more complex picture with significantly varying ice-bed interface conditions, ranging from predominantly hard bed framework in the GAP and ICE study sites, to softer sediment-rich conditions nearby. The reason for the distribution of these contrasting bed types is not yet fully understood. However, studies conducted following these projects generally support the established conceptual understanding (e.g. Maier et al. 2021, Fahrner et al. 2024).

4.1.1 Pressure in the subglacial drainage system

Water pressure levels beneath an ice sheet, including the hydraulic pressure regime, are important for its dynamics and behavior and for groundwater recharge. These pressure levels are influenced by the presence of liquid water at the ice base, which occurs when temperatures reach the pressure melting point. The hydraulic pressure regime beneath an ice sheet is influenced by multiple factors, including ice thickness, basal melting, ice movement, and seasonal variations in surface meltwater. These factors drive fluctuations in water pressure, contributing to e.g. the development of subglacial lakes and rivers. Additionally, the specific characteristics of basal hydraulic pressure depend on the distance to the ice sheet margin, the permeability of the basal drainage system – potentially affected by sediment composition – and the underlying bedrock's topography and permeability.

Fluctuations in hydraulic pressure significantly affect ice sheet movement; high water pressure can reduce friction between the ice and the substrate, resulting in faster basal ice movement, while lower pressure can result in slower basal ice movement or stagnation. This complex interplay of water pressure and the hydraulic regime is essential for understanding ice sheet stability and its response to climate change (e.g. Marshall 2021). Variations in hydraulic pressure influence groundwater flow, erosion and sediment transport, reshaping the subglacial environment, but it could also influence the behavior of the groundwater system in front of the ice sheet, for instance, the nature of taliks.

The permeability of crystalline bedrock is of less importance for the hydraulic pressure under an ice sheet than the other factors due to the large hydraulic conductivity contrast between bedrock and the basal drainage system, but important for the magnitude of local recharge of meltwater to groundwater (Harper et al. 2019).

What have we learned?

Data from the boreholes drilled through the ice sheet has enhanced our understanding of pressure within the subglacial drainage system. Alongside ice velocity data and radar measurements, the ice borehole findings highlight the spatial variability in pressure and hydraulic potential[4] in the basal drainage system.

Both projects demonstrate that basal water pressure fluctuates across both time and space, requiring a comprehensive characterization that accounts for its magnitude, as well as spatial and temporal gradients. The consistency across locations suggests similar conditions in comparable ice sheet environments. The following sections outline the key findings regarding subglacial water pressure.

Pressure is consistently high, and confined to a limited range

The GAP concluded that hydraulic pressure near overburden is an adequate assumption for the pressure region at the basal boundary of the ice sheet (Harper et al. 2016). A hydraulic head corresponding to 92% of the ice thickness is considered to be the typical state based on data from the ice holes (Claesson Liljedahl et al. 2016). Data from the ICE project supported the GAP findings and expanded with additional data, which facilitates a deeper level of understanding of the pressure conditions in the subglacial drainage system.

Both projects reveal that the pressures are confined to a limited range between 0.8–1.1 as a fraction of local ice overburden pressure (Harper et al. 2019). As ice overburden pressure scales with ice thickness, basal hydraulic pressure increases from atmospheric pressure at the ice margin to over 10 MPa at the center of the ice sheet. Results from other borehole studies in Western Greenland, made in a range of conditions from thin (~200 m) to thick (~1600 m) ice thickness and with different basal substrate including soft bed conditions with thick till, all show pressures within the confined 0.8–1.1 range of the GAP and ICE projects (Andrews et al. 2014, Iken et al. 1993, Lüthi et al. 2002, Thomsen and Olesen 1990, 1991, Thomsen et al. 1991, van de Wal et al. 2015).

At the ice marginal area, the pressures were found to be low and highly variable, especially during the melt season. This means that hydraulic potential gradients cannot be reliably estimated using ice geometry alone, but that active subglacial drainage processes play an important role in shaping the pressure field (Wright et al. 2016).

Basal pressure undergoes daily to seasonal changes

The GAP established that the basal pressures vary over time on daily to seasonal timescales (Harper et al. 2016), and these findings are supported by the ICE project (Harper et al. 2019).

The basal water pressure undergoes distinct seasonal changes. Higher frequency diurnal variations are common during the summer melt season. The character and magnitude of the diurnal variations varies across the drainage system and are typically associated with active drainage system processes (e.g. Andrews et al. 2014). The lowest pressures are typically observed during the summer diurnal period. Small diurnal pressure changes may occur when water flow is absent (Meierbachtol et al. 2016b). It is suggested that the active drainage system and ice flow respond to changes in surface melt, causing widespread pressure variations across the bed over short periods. Even the largest measured diurnal pressure variations are within the pressure range 0.8–1.1 of overburden pressure.

Observations and modeling demonstrated that the melting season growth of conduits from the ice margin toward the ice sheet interior is limited by two key factors: thicker ice increases the rate of conduit closure, and a decreasing surface slope reduces the melting efficiency of flowing water. These processes limit how far conduits can extend into the ice sheet. In the study region, the seasonal growth of a discrete conduit network was found to be limited to an extent of about 20 km inward from the ice edge.

[4] Hydraulic potential (Φ_h) combines the elevation potential and pressure potential, expressed as $\Phi_h = \rho_w g z_b + P_w$, where ρ_w is the density of water, g the acceleration due to gravity, z_b the bedrock elevation and P_w the water pressure (Wright et al. 2016). It represents the energy driving subglacial water flow, and is directly proportional to hydraulic head, which expresses the same concept in units of elevation.

Potential gradients at the ice sheet bed arise from drainage system processes over short length scales

The basal water pressure across the high-relief bed of the study area experiences longitudinal and transverse (relative to the ice flow direction) gradients stemming from two sources: changes in ice thickness determining the overburden pressure, and changes arising from basal processes. Because water pressure varies within a relatively restricted fraction of overburden pressure, the hydraulic potential gradient most closely mimics the ice thickness when considered over larger spatial scales and time periods that average diurnal variation in pressure.

The GAP observations allowed for quantification of long length scale (ice-sheet scale) potential hydraulic gradients, which is a result of the ice-sheet surface and bed geometry. An average gradient of ~115 kPa/km was found in the ice flow direction across the GAP study transect (Harper et al 2016). Gradients arising from processes in the local drainage system may be superimposed on topographically driven gradients. Due to local variability in bed topography and ice thickness, the ICE project found both lower and higher gradient values than the GAP. The potential hydraulic gradient can deviate substantially from the overall ice sheet mean when calculated for length scales of a few kilometers because the basal relief topography in this area is high. The ICE project showed that such local gradients can be large, reaching 20 kPa m^{-1}, but due to the complex nature of the subglacial drainage system it is complicated to draw conclusions of what this can mean on a larger scale than the very local ICE-study scale. Although these larger gradients are present, they should be seen as an end-member maximum (Harper et al. 2019). However, this suggests that the basal pressure conditions are highly variable on shorter spatial scales.

Short duration basal pressure transients are uncommon and spatially localized

The GAP project did not focus on short duration pressure pulses, but this was studied by the ICE project, which identified very rare but rapid pressure pulses. From a record of 49 million measurements over a period of 9 months, only five pressure pulses were detected in two boreholes spaced 180 m apart. Each pulse was manifested as a short (< 1 minute) pressure drop, yet none were transmitted between boreholes. The cause of these pressure drops remains unclear, but they appear to be localized within the basal environment (Harper et al. 2019) and again suggest a very spatially variable system. It is noteworthy that no short pulses of pressure increase were observed over the 9-month measuring period.

The borehole pressure records from the GAP and ICE projects along a transect of the western part of the ice sheet reveal seasonal variations in basal water pressure within a confined range. Although the pressure generally follows ice thickness patterns, significant local gradients emerge due to subglacial drainage processes. While the GAP/ICE datasets are interpreted to broadly represent regional conditions, some uncertainties remain, particularly regarding pressure in unmeasured deep troughs and near ice-margin hydrology, where distinct hydraulic regimes may exist. While no widespread pressure peaks/drops were observed at the ICE project site, similar events have been recorded on alpine glaciers, even causing instrumentation failure. There is no evidence ruling out their occurrence beneath ice sheets, but the ICE project data suggests that if they do exist, they are rare over multi-year timescales. Numerical modeling after the GAP/ICE project using water pressure data from these projects can reproduce the important aspects of the GAP/ICE data (Hill et al. 2025).

Hydrogeologic response of the deep groundwater system to ice sheet dynamics

Data from the 650 m deep DH-GAP04 borehole ending under Isunnguata Sermia outlet glacier has allowed in-situ pressure measurements of deep groundwater conditions beneath the ice sheet. From observations it is seen that the groundwater system exhibits a rapid and sensitive response to relatively minor variations in ice sheet dynamics, responding to even minor variations in ice loading and meltwater input. It is further seen that hydraulic head at depth fluctuates across multi-annual, seasonal, and diurnal timescales, driven by fluid pressure changes at the ice-bed interface resulting from shifts in ice loading and hydrological activity. A consistent decline in hydraulic head over an eight-year monitoring period is primarily attributed to ongoing ice sheet mass loss (Claesson Liljedahl et al. 2021).

These findings collectively show that pressure changes in the subglacial drainage system rapidly propagate to depth and that the system is not recharge-limited; its response is not due to solid earth deformation but rather to glaciological pressure forcing.

4.1.2 Characteristics of the subglacial drainage system

Direct observations of the hydrologic processes at the base of an ice sheet were limited prior to the GAP and ICE projects. In lack of such observations, the conceptual understanding of the subglacial drainage system of mountain glaciers was extended to ice sheet settings, but questions remained regarding the transferability of hydrological processes from glaciers to ice sheets (Jansson 2010). Although many insightful lessons can be learned from studying glaciers, the findings from mountain glaciers are not always applicable to ice sheets due to differences in size, geometry, and thermal conditions. One of the main goals of the GAP was therefore to investigate a study region representing a continental ice sheet instead of mountain glaciers.

Based on glacier studies it was conceptualized that the subglacial drainage system of an ice sheet comprises channels and cavities that transport meltwater beneath the ice, influencing ice flow velocity and stability. This system evolves over time during the melting season and also varies between seasons. It becomes more extensive as the melt season progresses as increased and continued meltwater input forms larger channels. Early in the melt season, the system is inefficient, comprising linked cavities and distributed pathways. As meltwater flow increases, these pathways are enlarged and form new channels resulting in that the system gradually becomes more efficient. In areas without well-defined channels, the drainage system is characterized by a network of linked cavities and thin water films, leading to higher basal water pressures and potentially more significant ice flow acceleration. This slower, more diffuse water flow can impact the overall stability and dynamics of the ice sheet.

What have we learned?

The findings made by these projects significantly improved the understanding of the subglacial drainage system and subglacial hydrology, which is essential to enhance the understanding of how an ice sheet affects groundwater flow. One important finding was the direct observations of various components of the subglacial drainage system, which not only validated but also advanced the previous conceptual understanding of this system. From observations, analysis and modeling, key findings concerning the drainage system are described below.

Surface meltwater drives the entire ablation zone

From analysis of ice sheet geometry and surface velocity in combination with ice borehole observations and modeling, it is suggested that the entire ablation zone is warm based with liquid water at the bed. It is further suggested that surface meltwater that drains to the bed near the equilibrium line altitude (ELA) can reside under the ice during winter throughout the ablation zone (Meierbachtol et al. 2016a). The quantity of water at the ice bed during winter and over large regions thus has the potential to be greater than that supplied by basal melting.

The basal drainage system is capable of dynamic routing of meltwater

The hydraulic potential of the subglacial water layer is influenced by the sloping ice surface, bed topography, and pressure gradients within the subglacial water layer. These pressure gradients arise from changes in ice thickness and the development of the subglacial drainage system. The magnitude of the potential gradients varies over length scales due to the substantial ice surface and basal topography in the ablation zone across the study area. Different locations on the bed can undergo sub-annual changes in the direction and magnitude of the hydraulic potential in response to water pressure fluctuations.

While regional ice thickness remains relatively stable annually, basal water pressure fluctuates due to diurnal and seasonal changes in the drainage system. This can result in a potential gradient field capable of promoting dynamic drainage system connectivity. This process is more pronounced where the bed slope is steep, is angled relative to the ice surface slope, or where water pressure fluctuates around a high fraction of ice overburden pressure. Such conditions occur over large parts of the GAP/ICE study sites, especially during melt season when diurnal pressure variations are common.

The basal drainage system comprises several discrete components with variable transmissivity

Water flows beneath the ice sheet through distinct drainage pathways with varying capacity to transmit water. Hydraulic connectivity is uneven, changing significantly over short distances of just tens of meters. The basal drainage system in the study sites is thus highly dynamic, with rapid basal sliding causing instability in individual water pathways. Cavities continuously shift in connectivity and efficiency, while hydraulic potential gradients vary in magnitude and direction due to water pressure fluctuations. As a result, large volumes of water likely reach nearly all areas of the ice sheet bed in the ablation zone. Field observations indicate that the drainage system has substantial capacity as several cubic meters of water can be accommodated in just a few minutes, and that dynamic connections occur between drainage components at the ice sheet bed. It is also suggested that disconnected cavities exist in measurable volume that may cover large parts of the bed (Meierbachtol et al. 2016b).

While the measurements from GAP/ICE have provided valuable insights, uncertainties remain regarding the characteristics of the subglacial drainage system. These uncertainties primarily stem from the localized nature of borehole measurements, raising questions about their broader representativity across the Greenland ice sheet. Nevertheless, no previous or subsequent studies have contradicted the findings from the GAP and ICE projects, and both numerical modeling and broader scientific research continue to support and advance the conceptual understanding of the subglacial drainage system established by GAP/ICE (e.g. Hoffman et al. 2017, Chandler et al. 2021).

4.1.3 Basal thermal distribution and groundwater formation

The basal thermal conditions under an ice sheet govern how much meltwater can be produced, whereas the most controlling factor for groundwater formation is the hydraulic conductivity of the bedrock, which determines how much water can infiltrate. Other factors influencing water movement and recharge are the bedrock fracture network, bed topography, topographic ice surface gradient and ice sheet dynamics (Lemieux and Sudicky 2014).

The basal thermal boundary condition is thus essential in ice sheet and groundwater interactions, as liquid water can only form when the basal ice temperature reaches the pressure melting point. The temperature beneath the ice is determined by a complex balance of heat sources and sinks influenced by climate, geothermal heat, and ice dynamics. Key processes include vertical and horizontal ice movement, geothermal flux, atmospheric heat exchange, ice accumulation and melting, strain heating, frictional heating from sliding basal ice, and release of latent heat from refreezing of water. Each factor's impact varies significantly with time and location.

Where the ice bed reaches the pressure melting point, liquid water enables groundwater recharge. Provided the bedrock or regolith is permeable, this promotes groundwater recharge beneath the ice, while higher elevations or colder zones remain frozen with limiting water flow. Such variability in basal thermal conditions creates localized pathways for groundwater movement across the ice sheet. Such variability in basal thermal conditions creates localized pathways for groundwater movement across the ice sheet.

Once the water is liquid, the routing of water beneath the ice sheet is determined by the direction and magnitude of the hydraulic potential gradient, which drives water from higher to lower potential (Wright et al. 2016). The hydraulic potential varies over time and space due to changes in the drainage system.

Understanding the basal thermal conditions of the Greenland ice sheet is challenging due to scarce geothermal heat flux (GHF) data. Näslund et al. (2005) noted a pronounced variability in GHF across the Fennoscandian shield, suggesting similar differences beneath the Greenland ice sheet due to its similar geology. However, direct measurements have been and remain sparse (Colgan et al. 2021, 2022). Data on hydraulic conductivity of bedrock and soil in Greenland are also limited.

What have we learned?

Given the limited bed topography data at the start of the GAP project, mostly restricted to a narrow section of the marginal zone of the ice sheet (red part of the surface-drainage area in Figure 4-1), the GAP study initially adopted a conceptual model with a frozen central bed, a surrounding melt zone, and thinning frozen edges, reflecting the best available understanding at the time.

GAP research, combining direct borehole data and modeling, significantly enhanced the understanding of basal thermal conditions in the region. Borehole evidence revealed a consistently melted bed in the ablation zone, with the boundary between melted and frozen areas sensitive to geothermal heat flux but less affected by ice sliding speed and frictional heat (Harper et al. 2016). Surface conditions also play a crucial role in the distribution of frozen and melted areas, leading to the conclusion that basal water exists under more than 75% of the ice sheet in the study area with the most upstream part of the study transect remaining frozen at the bed (green area indicated at the base of the ice sheet transect in Figure 4-1). This basal water system exhibits considerable spatial and temporal variability, facilitating both groundwater recharge and discharge.

Figure 4-1 shows the conceptual model of the basal zones and subglacial conditions as suggested by the findings from GAP and ICE. Three basal zones beneath the ice sheet were delineated: the *frozen bed zone, the wet bed zone,* and *the surface-drainage bed zone*, which further contains two sub-regions: the transient conduit region (from the margin to 15–20 km inward) and the high-pressure region (from 20 km to the equilibrium line). According to the findings, the hydraulic pressure rises due to abundant meltwater during summer, which causes the drainage system to shift from linked-cavity flow in winter to discrete conduit flow in summer, with conduits extending 15–20 km inland from the ice front. Basal melting in the wet bed zone generates minimal water for groundwater recharge near the upstream frozen bed zone boundary, increasing to about 10 mm yr^{-1} downstream, roughly 120 km from the ice margin (Harper et al. 2016). This meltwater production is fairly constant year-round, as basal melting processes are not seasonally dependent, except for potential summer frictional meltwater increases. The study found no need to model a complex patchwork of frozen and melted conditions in the retreating outer areas, as broad wet-based conditions are common during deglaciation.

While not observed in GAP, models and conceptual understanding suggest a gradual transition from frozen to wet bed conditions, with higher areas typically frozen and low-lying areas melted. A study by Stansberry et al. 2022 examined the migration of the frozen/melted basal boundary in western Greenland, showing that as the ice sheet margin retreats, the boundary shifts upstream by several kilometers. This study also highlights how ice dynamics and bedrock thermal properties influence the transition between frozen and wet bed conditions, where low GHF values in certain regions slows the expansion of melted areas.

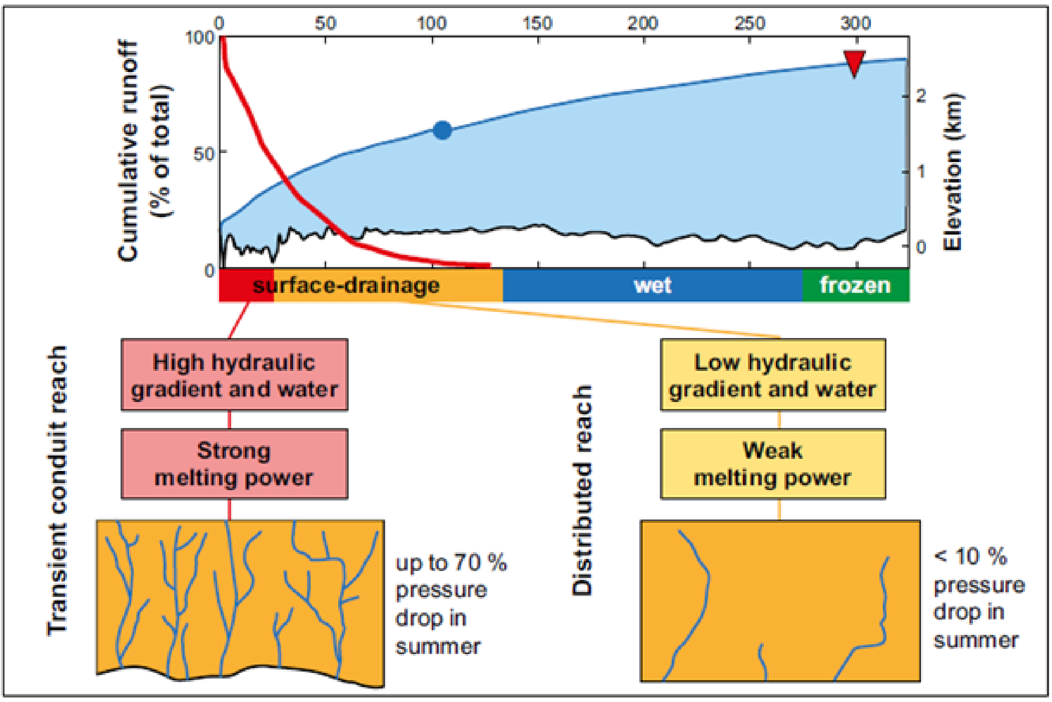

Figure 4-1. *Schematic cross-section of the GAP study domain, extending from the ice divide to the terminus of Isunnguata Sermia. The top panel displays ice surface and bed topography, with the blue dot marking the equilibrium line altitude (ELA) and the red triangle indicating the 20-year average extent of surface melt. The red line represents 1958–2013 average meltwater runoff, accumulated from high to low along the ice surface, based on the MAR model (Tedesco et al. 2013). Additionally, the three basal hydrological zones are shown beneath the ice sheet in red/yellow, blue, and green. Melted conditions prevail in the surface-drainage (red-orange) and wet (blue) zones. The bottom panel provides a conceptual representation of subglacial conditions. The basal boundary of the GAP/ICE study areas is hard bedrock with a relatively thin and discontinuous layer of sediments. The deep troughs crossing the area likely act as sediment sinks and route sediment to the glacier outlets. The troughs may be underlain by thick sediments. A spectrum of basal conditions is possible where mixed hard bedrock and thick sediment-covered basal conditions are likely present over scales of kilometers. Figure from Claesson Liljedahl et al. (2016).*

The calculated GHF values from the GAP bedrock boreholes are significantly lower than previous and recent estimates for this part of Greenland. In a study by Greve (2019), the GHF for the entire Greenland was estimated, showing low values in the south and northwest and higher values in central and northeast Greenland. However, the study is based on only a few actual observations, thus the spatial coverage of the GHF is associated with large uncertainties. The GHF value from borehole DH-GAP04 and DH-GAP03 is estimated to be 27 to 35 mW m^{-2} (Harper et al. 2016). This can be compared to the 42 mW m^{-2} for this area based on the study by Greve (2019).

There is broad agreement by numerical models that melted basal conditions extend well inland of the ELA at the study area, despite evidence of low GHF (Meierbachtol et al. 2015, Poinar et al. 2015, Seroussi et al. 2013). Different GHF values were used in the GAP modeling work to examine basal melt rates. Modeling showed widespread melted conditions across the study area except for areas with deep bedrock troughs. However, from the field studies it is evident that melted bed conditions prevail along Isunnguata Sermia. The discrepancy between ice models and field observations regarding basal melt conditions arises because refreezing surface meltwater inside glaciers generates significant heat, a process that is not adequately represented in the ice sheet model used. Without accounting for this in an adequate way, areas of deep ice flow appear colder than they are, as bedrock features pull colder ice downward (Harper et al. 2016).

The new findings from GAP reveal discrepancies between the measured upper boundary condition and the ice surface conditions simulated by current atmospheric models, affecting the modelling of the ice sheet's internal and basal temperature fields. While the initial representation of basal thermal conditions is not sensitive to these surface boundary variations, refining the surface boundary is necessary for enhancing the basal thermal field's accuracy. Although GAP has provided substantial direct measurements of an ice sheet's internal temperature distribution, the data still comes from a limited number of observations across a vast ice sheet.

Utilizing data and knowledge obtained from the ice-sheet field and numerical studies, a set of groundwater modeling studies were carried out to investigate the conditions and processes that impact the recharge of glacial meltwater into the geosphere. For the first time, the choice for the subglacial boundary condition, i.e. ice overburden pressure, used in groundwater models was justified by field data (Jaquet et al. 2019). In the study by Vidstrand (2017), pressure boundary conditions in a groundwater flow model were further tested. The study concluded that a specified pressure boundary condition is an appropriate choice in topographically flat areas, such as the Forsmark site, but should be used with care in areas with more pronounced topographic relief, such as the GAP study site. Moreover, the study concluded that it is important to properly represent the subglacial permafrost, and a proper model resolution is needed to be able to reveal details of subglacial groundwater flow and meltwater transport within and through the subglacial drainage system.

Using the Forsmark hydraulic parametrization in a groundwater model (DarcyTools) gives flux values below 10 mm yr^{-1} at depth below 300 m. The flux is sensitive to the hydraulic conductivity field but not sensitive to variation in boundary conditions given by the ice sheet model. The recharge at depth is influenced by the type of glacial boundary conditions, i.e. dynamic fluid pressure versus meltwater rate. However, for most situations the application of dynamic fluid pressure boundary conditions, based on ice overburden hydraulic pressure, is justified, particularly for present day conditions in Greenland (Jaquet et al. 2019).

4.2 Permafrost

Permafrost and perennially frozen ground typically develop during pro- and periglacial conditions. The definition of permafrost is ground that remains at or below 0 °C for at least two consecutive years (French 2007). This definition is based exclusively on temperature, and disregards whether water is present in liquid form or frozen. As a result, the term permafrost does not always equate to perennially frozen ground, because the pressure, the chemical composition of the water, and the properties of ground matter, may result in the water being liquid also at temperatures below 0 °C. The volume defined to have permafrost based on the temperature may thus consist of both permanently frozen layers, as well as unfrozen layers (i.e., cryopegs). It is also possible that the permafrost ground lacks ice because there is no water present, e.g., due to extremely dry conditions. For the purpose of this report, permafrost implies the presence of frozen water (unless specifically stated otherwise).

Understanding the extent of permafrost, both in front of the ice sheet and beneath it, is important because it influences subglacial hydrology, affects meltwater drainage, and determines how groundwater interacts with meltwater inputs. Permafrost can induce freeze-out of salts through progressive freezing of groundwater, forcing dissolved salts to concentrate in the remaining unfrozen water. Finally, the presence of permafrost limits the exchange of water between the deep groundwater system below the permafrost and the shallow groundwater system that develops in the active layer each summer.

What have we learned?

All GAP/GRASP bedrock boreholes were equipped with optical-fiber cables, enabling high-resolution temperature measurements along their lengths using the distributed temperature sensing (DTS) technique (Harper et al. 2016). However, since DH-GAP01 extends in under Two-Boat Lake – to investigate the possible presence of a talik (Figure 2-1C, D), it is primarily DH-GAP03 and DH-GAP04 that can be used to study permafrost. Temperature data from DH-GAP03 and DH-GAP04 suggest that permafrost reaches a depth of 300-400 m at the ice-sheet margin. The base of the permafrost (i.e., the 0 °C isotherm) is located at ~315 m depth at DH-GAP03 and at ~350 m depth at DH-GAP04. The difference in permafrost thickness between these two borehole sites is likely due to differences in geology, local surface conditions over time, elevation and slight differences in instrument calibration procedures. DH-GAP04 is located on an elevated, well exposed ridge with thin regolith cover, whereas DH-GAP03 is located in a depression with a thicker regolith layer. The snow cover close to DH-GAP04 is also known to be thin during winter as the area is exposed to strong katabatic winds. In less wind exposed locations with thicker snow cover – like at DH-GAP03 – we expect a thinner permafrost layer. Because of the highly variable topography in the region, the thickness of the permafrost likely has a high spatial variability. Modelling of the temperature in the boreholes indicates that permafrost has grown to its current extent when the climate was colder than at present (Harper et al. 2016). From the borehole temperature profiling, it is suggested that the

MAGT during growth of permafrost was 2 to 3 degrees colder than presently observed (i.e., –5 to –6 °C), which corresponds to a MAAT of –7 to –9 °C (Harper et al. 2016).

Data from GAP and ICE shows that the main part of the ice sheet in the study area is subject to basal melting conditions, with the exception of the central parts of the ice sheet (Harper et al. 2016; Hartikainen et al. 2022). Considering that most of the currently glaciated regions have remained glaciated for the past 10 000 years without experiencing permafrost development under the ice sheet, it suggests that permafrost is absent beneath the significant portions of the warm-based ice areas. An exception is at the ice margin, where a wedge of permafrost probably stretches in under the ice sheet. The extent of the subglacial permafrost wedge remains uncertain. However, indirect observations based on electromagnetic soundings in the ice marginal area directly east of the DH-GAP04 borehole suggest that subglacial permafrost appears to exist at least 2 km from the ice margin (Ruskeeniemi et al. 2018). Even if it is not possible to exactly define the base of the permafrost with the used method, it shows a thin unfrozen layer at the base of the ice and this layer is underlain by frozen conditions. These findings are consistent with temperature profiles collected from ice boreholes both close to the margin and farther in on the ice sheet, which shows that the temperature at the base of the ice is close to 0 °C at the ice margin and at pressure melting conditions farther up-glacier (Harper et al. 2016).

In permafrost landscapes the upper part of the ground that thaws each summer season is referred to as the active layer. From studies using ground penetrating radar in the Two-Boat Lake catchment (Section 3.3), we can conclude that the characteristics of the active layer vary depending on the topography and regolith conditions (Petrone et al. 2016). The active layer is generally thicker at exposed ridgetops and thinner in depressions. In addition, there is also a strong influence of the type of regolith, soil moisture, and vegetation (all of which are interlinked). Moist conditions, with wetland type vegetation and peaty silt deposits, result in a thinner active layer (0.5 m), while drier conditions with grassland, shrub or heath vegetation result in a thicker active layer (0.7–0.8 m). In areas with exposed bedrock the active layer can reach a thickness of more than 2 m (Petrone et al. 2016, Harper et al. 2011).

Even if the biomass in the living vegetation is slightly higher in the wetlands compared to meadow and dry grassland (Lindborg et a. 2020), the difference in the appearance of the vegetation is not visible in the field. There is also no distinct pattern with more snow accumulation in the wetlands during winter (which could affect the thaw in the active layer). Hence, shallower active layer in the wetlands can largely be attributed to the difference in thermal properties between wet and dry ground conditions, i.e., it requires more energy to thaw wet material compared to dry material. However, even if the maximum thickness of the active layer is favored by dry conditions, advective heat transport with running water can have a profound effect on the thaw rate during the formation of the active layer in spring. Studies during the snowmelt period of 2017 showed that the thaw rate generally was higher in connection with temporary streams (Rydberg et al. 2025).

GRASP monitored regolith temperature in the active layer using temperature loggers installed down to 2 m depth at one location with eolian deposits in the southwestern part of the Two-Boat Lake catchment (Figure 2-1D). As expected, all monitored levels down to 2 m depth show a considerable variation in temperature over the year (Figure 4-2). At 0.25 m depth temperatures start to increase in late March to early April, and by late May to early June the temperature reaches above zero degrees. The active layer then reaches a maximum in early to mid-August, with temperatures above zero at 0.25, 0.5 and for some years also at 0.75 m. After this point temperatures start to drop, and they go below zero degrees in late September to early October (first at 0.25 m and a few days later at 0.5 m). At larger depths the temperature continues to increase until late December to early January when the temperatures also at larger depths start to drop.

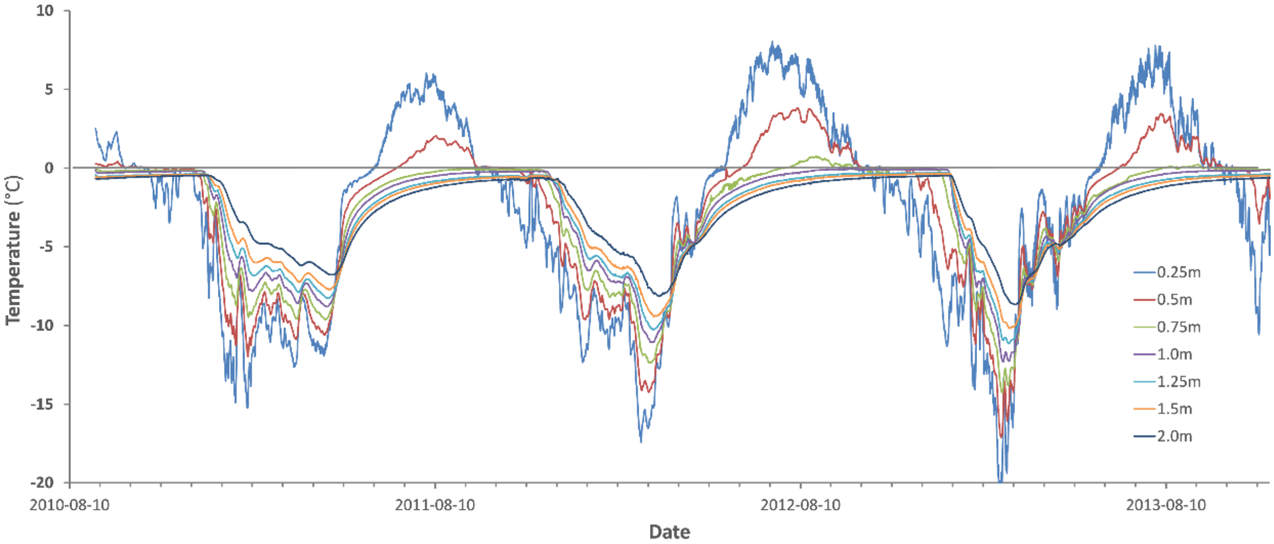

Figure 4-2. *Regolith temperature down to two-meter depth measured in the Two-Boat Lake catchment during the period 2011–2013 (the same pattern appears also during the remaining years of the study period, but to increase visibility they were omitted from the figure).*

4.3 Depth of glacial meltwater infiltration

Recharge of glacial meltwater into the subglacial bedrock is a dynamic process where multiple processes act together to facilitate water infiltration and penetration into the subsurface environment. The type of bedrock, geological conditions and intensity of melting governs to what depths meltwater can infiltrate. There are several pieces of evidence from Fennoscandia of infiltration of glacial meltwaters in crystalline rocks (e.g. Laaksoharju et al. 2008, 2009, Pitkänen et al. 2001, Posiva 2013, Smellie et al. 2002). The depth of recharge varies across these sites due to differing geological conditions and the prevailing glacial environment during the deglaciation.

Understanding the recharge process of glacial meltwater is crucial for assessing subsurface hydrological and hydrogeochemical processes in glaciated regions. Due to their low dissolved solids content, infiltrating meltwaters in glaciated regions are expected to be dilute. Additionally, glacial meltwaters often contain higher concentrations of dissolved oxygen compared to soil waters during temperate conditions. Taken together this can significantly influence geochemical interactions and groundwater evolution at depth.

What have we learned?

Data obtained from groundwater samples collected at boreholes DH-GAP04 and DH-GAP01, along with information from the Leverett spring, have helped improve the understanding of the glacial recharge depth. The isotopic composition of groundwater from the lowermost section of DH-GAP04 indicates a meltwater source, suggesting that glacial meltwaters have infiltrated to depths of at least 500 m in the bedrock, but because the borehole is not sufficiently deep, the maximum infiltration depth cannot be determined from the GAP data. Similarly, the isotopic signature of groundwater samples from depths below 140 m in borehole DH-GAP01 suggests an influence of glacial meltwater that is not present in overlying Two-Boat Lake. Given that this borehole is situated in a through talik surrounded by continuous permafrost that is several hundreds of meters thick, this further suggests that the glacial meltwater has a profound effect on the chemical signature of the deep groundwater both at great depths and in front of the ice sheet (Claesson Liljedahl et al. 2016). The spring water emerging in front of the Leverett glacier, exhibiting a slightly evaporative isotopic signature, suggesting a mix of deep groundwater (with an unknown age) and glacial meltwater from the sand plain surrounding the spring. The groundwater component has infiltrated to significant depths over time, maintaining the flow of the spring annually. Analysis of isotopic signatures in sampled spring waters indicates a glacial origin for the groundwater component (Claesson Liljedahl et al. 2016).

Collectively, the isotopic compositions of the groundwater samples from the boreholes and the Leverett spring show that glacial meltwaters can infiltrate to depths of at least 500 m. However, it is not possible to determine whether this water is recharging close to the borehole, or if it has traveled a significant distance below the ice. Nor is it possible to delineate the travel path of the water sampled in the Leverett spring. Based on data on dissolved He_4 the water samples from borehole DH-GAP04 has an age in excess of hundreds of thousands of years, which indicate that the water has been isolated from the surface for a very long time period. The high content of drilling fluid in the upper and middle sections of DH-GAP04 also suggests that groundwater flow at depth is limited. Together this shows that although pressure changes in the subglacial drainage system are rapidly propagated to depth, this does not translate to a rapid exchange of water at greater depths (Claesson Liljedahl et al. 2021). The limited flow of water in the deep groundwater system is likely influenced by factors such as the presence of permafrost, geological structure, and the hydrogeological conditions specific to the borehole's location.

4.4 Chemical composition of water at DGR equivalent depths

DGR equivalent depths here correspond to the depth of the planned spent fuel repository (approximately 500 m) or more. The chemical composition of groundwater at these depths is controlled by a range of geological, hydrological, and geochemical factors. Among the most significant are the geological framework including fracture frequency and aperture, water-rock interaction, groundwater flow dynamics, temperature and pressure conditions, redox state, dissolved salts and ions. These factors collectively determine salinity, pH, redox conditions, and concentrations of dissolved elements.

Many of the hydrogeochemical processes that affect groundwater composition in fractured rock under temperate climates also occur in periglacial or glacial environments. However, certain factors differ significantly under glacial and periglacial conditions: i) variations in groundwater flow due to increased hydraulic pressure near ice margins or reduced hydraulic conductivity in areas with permafrost, ii) differences in the composition of infiltrating waters; glacial meltwater as opposed to soil porewater, and iii) reduced biological activity/productivity in the subglacial drainage system and in the active permafrost layer compared to soils or sediments in temperate climates.

These differences between temperate and glacial or periglacial climates influence the processes that control groundwater composition and redox conditions. Specifically, because glacial meltwater typically has a low concentration of dissolved solids, infiltrating glacial meltwater in glaciated areas is expected to be diluted, as supported by modeling (e.g. Vidstrand et al. 2014).

In terms of redox conditions, the presence of significant amounts of trapped air within glacial ice means that meltwater can carry elevated levels of dissolved oxygen. This contrasts with temperate soil waters, where biological and microbiological processes typically deplete available oxygen in groundwater. The higher dissolved oxygen concentrations in glacial meltwater can thus create more oxic environments at depth during glacial conditions, influencing redox reactions in subglacial ground water systems. This can, e.g., lead to enhanced oxidation of reduced species, such as sulfur and iron compounds, which might otherwise persist in their reduced state under anoxic conditions. Oxygen-rich meltwater can also affect microbial communities, promoting aerobic metabolism while suppressing anaerobic processes like sulfate reduction or methanogenesis. A shift in redox dynamics can thus impact mineral dissolution, nutrient cycling, and trace metal mobility within glacial meltwater systems (e.g. Hawkings et al. 2020).

What have we learned?

Previous studies of sub-permafrost groundwaters have primarily been carried out in sedimentary rocks (e.g. Alexeev and Alexeeva 2003, Clark et al. 2001, Hodson et al. 2020, Shouakar-Stash et al. 2007) and very few have been conducted in crystalline settings (Stotler et al. 2009, 2010). The reason for the few attempts to collect groundwater samples in areas with continuous permafrost relates to the difficulties of drilling through frozen ground and the challenges of obtaining undisturbed samples in combination with logistical challenges as the sites almost exclusively are located very remote. The only comparable studies were carried out by the Lupin and High Lake Projects in Arctic Canada (e.g. Stotler et al. 2009, 2010). In these studies, brine fluid was used as the drilling fluid, and only a few samples were collected over a very short period of time before the boreholes froze, which means that it was difficult to obtain representative samples of the sub-permafrost ground water chemistry.

Data from the two deep GAP bedrock boreholes in front of the ice sheet have notably enhanced our knowledge of fluid movement and geochemical processes in glacial climates. A key finding is that high-quality groundwater samples, with minimal drilling fluid contamination, can be obtained from boreholes in permafrost areas. Without the tailored borehole instrumentation system enabling thermal activation of the sampling tubes, extraction of sub-permafrost groundwater would not have been feasible. The primary source of information on deep groundwater came from borehole DH-GAP04, which provided chemistry, pressure and temperature data for groundwater beneath the ice sheet at DGR equivalent depths. DH-GAP01 supplemented this data suite with pressure, temperature, conductivity data, and groundwater samples from a through talik. In addition to the borehole investigations, valuable hydrogeochemical data and insights into processes occurring under glacial conditions were gained by studying surface waters and springs in the vicinity of the ice sheet margin, for example the spring near the pingo in front of the Leverett Glacier.

The groundwater samples from the GAP borehole constitute a unique dataset, being the first collected from beneath the permafrost under an ice sheet in crystalline bedrock, with sampling spanning several years. The project made significant advancement in the field by establishing a methodology for obtaining representative sub-permafrost groundwater samples, alongside high-quality physicochemical, mineralogical, and microbiological data (Harper et al. 2016). The key findings from observations and modelling are presented below.

Groundwater origin

Groundwater from DH-GAP04 indicates with its depleted $\delta^{18}O$-signature[5] that the sampled water originates from glacial meltwater, while samples from DH-GAP01 suggest a water with a mixture of glacial meltwater and lake water. The isotopic data from the GAP groundwater provides the first confirmation that meltwater from a warm-based ice sheet can penetrate fractured rock down to DGR equivalent depths. Based on available data it has not been possible to determine whether the water in the different borehole sections originates from distinct groundwater sources and thus may have different residence times (Harper et al. 2016).

[5] The ratio of heavy (^{18}O) to light (^{16}O) oxygen isotopes of a sample is reported relative to an international standard. $\delta^{18}O$ is defined as the deviation between the sample and the standard, expressed in parts per thousand (‰).

Salinity

Although the groundwater in DH-GAP04 originates from glacial meltwater, it has a relatively high chloride content; however, the chloride levels are still considered low overall, which can be seen when comparing with other sub-permafrost sites (Diak et al. 2023). With its glacial meltwater origin, it is shown that groundwater may reach repository depths during glacial conditions. Whether low-salinity groundwaters will be present at depth during a glacial period depends on several site-specific factors, including a hydrogeological setting that allows meltwater infiltration and the absence of soluble salts in the rock. Long residence times and water-rock interactions generally do not significantly increase chloride concentrations in infiltrated meltwaters unless the rock contains chloride, but crush and leach tests on rock samples from DH-GAP04 (Harper et al. 2016) suggest the chloride in borehole water originates from matrix diffusion and weathering reactions with carbonic acid in the meltwaters.

Based on data from the GAP, there is no evidence of extensive freeze-out of salts at the base of the permafrost – a process that could lead to increased salinity in groundwater and has been attributed to be a process of relevance in cold environments (e.g. Zhang and Frape 2003). However, because the investigation was limited to a single deep borehole, it remains possible that freeze-out is a highly heterogeneous process, varying across different locations and conditions.

Redox conditions

Anoxic/reducing conditions in the sampled groundwater from DH-GAP04 are indicated by gas analyses showing the presence of methane and hydrogen, along with low but detectable levels of Fe (II), Mn and dissolved U. In contrast, evidence of anoxia in the shallower DH-GAP01 samples is less conclusive based on chemistry alone but is supported by the presence of non-facultative anaerobic microbes (Harper et al. 2016) but is consistent with the deeper part of the Two-Boat Lake water column being anoxic (Lindborg et al. 2016). However, samples from both boreholes show very low or zero dissolved oxygen levels, even in samples with high drilling fluid content. The presence of S^{2-} in the low temperate groundwaters suggests microbial sulfate reduction, consistent with the microbial communities identified in DH-GAP04 by Bomberg et al. (2019). Biological methane production is indicated in this environment based on the detection of methane, but due to the low abundance of archaea in general in DH-GAP01 and DH-GAP04, the presence of methanogens could not be confirmed.

The water from the Leverett spring clearly exhibits anoxic conditions. Combining gas and water analyses with studies of the redox properties of fracture-infilling minerals from the bedrock drill cores indicates that the redox potential is effectively buffered by the rock. From the mineralogical studies it was concluded that reducing conditions occur in the bedrock from approximately 50 m depth. This means that while dilute glacial meltwaters can reach great depths, oxygenated waters do not penetrate deeply near the ice margin due to effective buffering.

The geochemical properties of the analyzed sub permafrost groundwaters from the GAP site indicate predominantly reducing conditions at depth, likely driven by active anaerobic microbial processes (Auqué et al. 2025). The body of data provided by GAP has shown that the processes that consume dissolved oxygen are effective at the site, and although the groundwaters are essentially of meltwater origin, no observation of oxic groundwaters has been made. The rate and the detailed characteristics of the oxygen depletion process cannot yet be determined because of several unknown factors: i) the oxygen concentration of the infiltrating meltwater; ii) the time of recharge, although estimated to be on the order of hundred thousand years or more; and iii) the detailed mineralogical and geochemical characteristics of the rock. The absence of oxygen in deeper groundwater is also consistent with the long residence times and low flow suggested from the ^{4}He data from DH-GAP04.

4.5 Presence of taliks in periglacial (and proglacial) landscapes

Lakes are an important component of a periglacial landscape, and for the Kangerlussuaq region, about 14% of the land surface is covered by lakes larger than 1 ha (Anderson et al. 2009). Because of the thermal properties of water, the presence of a water body has a profound effect on permafrost formation, and lakes are – provided they are deep enough – expected to be underlain by unfrozen ground, i.e., taliks. If a talik connects the surface system with the deep groundwater system, i.e., if it is a through talik, it can act as a recharging or discharging pathway between surface and the sub-permafrost system (e.g. Bosson et al. 2013; Vidstrand et al. 2014). It has also been shown that individual taliks can accommodate both discharging and recharging areas (Bosson et al. 2013, Vidstrand et al. 2014).

What have we learned?

Thermal modelling of circular lakes shows that a lake with a radius of 0.6 times the permafrost thickness can sustain a through talik (SKB 2006). This implies that, for the Kangerlussuaq region where the permafrost reaches a depth of 300–400 m, lakes with a diameter of 360–420 m can sustain a through talik. This also means that approximately 20% of the lakes in front of the ice sheet in the area could sustain a through talik (Harper et al. 2011, Harper et al. 2016).

One of the aims of the studies conducted at Two-Boat Lake was to confirm if that lake had a through talik as suggested by the modelling. The intention was also to determine whether the flow of water through the talik – if present – was directed upwards or downwards (i.e., whether deep groundwater was recharged or discharged through the talik). To answer these questions, borehole DH-GAP01 was drilled in under Two-Boat Lake. The inclined borehole starts at about 20 m from the lake shoreline and reaches a maximum depth of 140 m below the lake surface in the southern part of the lake. The temperature profile in the borehole shows that as soon as the borehole reaches beneath the lake, the ground temperature increases above zero degrees, and the temperature then quickly stabilizes at about 1.3 °C. This temperature is maintained to the bottom of the borehole (at a vertical depth of 140 m). Hence, the temperature profile in DH-GAP01 confirms the existence of a talik under the lake. From measurements of the pressure gradient in DH-GAP01, it can also be deduced that the talik is a through talik that connects the lake with the deep groundwater system.

The hydrological model for Two-Boat Lake indicates that for most years the net flow of water through the talik directed downwards, constituting a recharge talik, and that about 30% of the annual outflow of water occurs through the talik (Johansson et al. 2015b). It should be noted that the surface outflow from Two-Boat Lake during most years is very limited, and that most water leaves the lake through evaporation. A talik where groundwater is recharged is also supported by isotopic data from borehole DH-GAP01, where the $\delta^{18}O/\delta^{2}H$ and $^{87/86}Sr$ mixing signatures suggest that up to 25% of the groundwater originates from the Two-Boat Lake. Even if the net flow is mostly directed from the lake to the deep groundwater system, the hydrological model also suggests that during extremely dry years the direction of the flow can switch, and deep groundwater instead discharges through the talik. A change in the direction of the flow through a talik over space and time is also supported by the numerical modeling by Vidstrand (2017), where Two-Boat Lake (or Talik lake as Vidstrand calls it) is expected to have a talik where groundwater is recharging in the southern part of the lake, while it is discharging the northern part. Here it should be noted that the hydrological model of Johansson et al. (2015b) only considers at the net flow of water, and hence, if there is discharge of groundwater in the northern part of the lake, the recharge in the southern part of the lake must be larger than 4 mm yr^{-1}.

Because the hydrological model suggests that the flow of water through the talik is directed downwards during a normal year, the mass-balance model for elements shows that the talik serves as an export pathway for all elements that are present in the lake water column in dissolved form (Rydberg et al. 2023). This downward export accounts for about 45% of the total export of dissolved substances from the lake (the remaining export occurring through the outlet for substances that cannot leave as gases, and through the outlet and as evasion from the lake surface for substances that can be present as gases, i.e., CO_2). Due to the lack of data on the chemical composition of the deep groundwater, we cannot draw any conclusions regarding which elements that could enter the lake when deep groundwater discharges through the talik. Even if the amount of upward flow during an extremely dry year is small, it is important to recognize that it is possible for elements present in the deep groundwater system to reach the surface through the talik.

4.6 Hydrological modelling and biogeochemical cycling of elements in the surface system

The climatological conditions set the boundary conditions for a landscape. It dictates the vegetation cover, controls the movement of water and affects the biogeochemical process rates. Similarly, the hydrological movement of water through the active layer of a periglacial catchment will have a profound effect on the transport and accumulation of elements and material in both terrestrial and aquatic systems. In essence, the movement of water connects the aquatic and terrestrial parts of the surface system to a single unit, and a well constrained hydrological model is key when calculating mass-balance budgets for a system.

What have we learned?

Based on the meteorological data (e.g., temperature, precipitation) collected in the Two-Boat Lake catchment during the GRASP-project Johansson et al. (2015a, b) constructed a hydrological model for the surface system (i.e., active layer). This model reveals a system that is dry, with the potential evapotranspiration during summer exceeding the availability of soil water in relatively large parts of the catchment. There is, however, large intra-annual variability, and during the snowmelt period there is enough water to allow for temporary streams to emerge in depressions throughout the catchment, and in autumn there is water moving through areas with peaty silt (Figure 2-1D). About half of the annual runoff (52%) is generated during the relatively short snowmelt period that extends from mid-May to late-June. During July and early August, the temperatures are relatively high, and precipitation is scarce, which results in a dry summer period with limited runoff as surface flow (Figure 4-3). When temperature drops and rainfall intensity increases in late August and September, there is a brief period with more runoff again, but compared to the runoff during the snowmelt period, the runoff is still relatively small. A limited runoff, and the lack of real stream channels have consequences for the transport of material and elements through the system. For example, the lack of any larger, fast flowing streams means that the hydrological transport of particulate material is limited. This is exacerbated by the fact that about half of the water enters the lake as diffusive groundwater, which also primarily transports solutes.

To understand the movement and accumulation of various elements in the Two-Boat Lake catchment GRASP have established two whole system mass-balance budgets. One is intended to look at the cycling of carbon (Lindborg et al. 2020), and the other looks at the cycling of 42 other elements (Rydberg et al. 2023). The reason for separating carbon from the remaining elements is that it has a significant gaseous component that is lacking in the cycling of other elements.

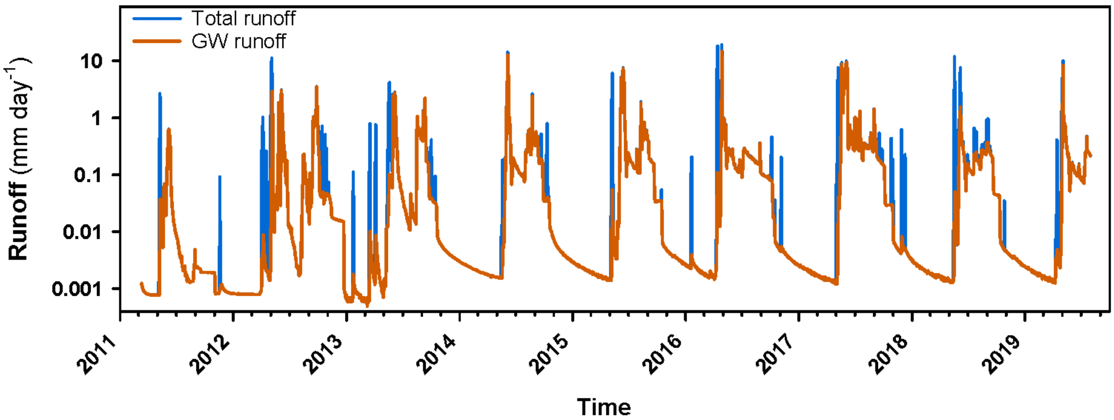

Figure 4-3. *Runoff from the terrestrial part of the Two-Boat Lake catchment to the lake as a function of time. The blue line represents total runoff including surface runoff, while the orange line represents diffuse sub-surface groundwater flow. Generally, the runoff is dominated by diffuse groundwater flow during the summer and winter periods, and surface water only contributes during the snowmelt period. Figure modified from Rydberg et al. (2025).*

Carbon

The carbon mass-balance budget shows that terrestrial net primary production is the main input of carbon to the Two-Boat Lake catchment (99 % of input), and that most carbon leaves the catchment through soil respiration (90% of total export/storage). The largest carbon pools are active layer soils (53% of total carbon stock or 13 kg C m^{-2}), permafrost soils (30% of total carbon stock or 7.6 kg C m^{-2} from the permafrost table down to 100 cm soil depth) and lake sediments (13% of total carbon stock or 10 kg C m^{-2}). The remaining pools are all small, and together they only make up 4% of the total catchment carbon pool (Lindborg et al. 2020).

The transport of dissolved organic carbon from the terrestrial to the aquatic system with surface and groundwater is lower than in wetter climates. Still, the annual input of 4100 kg of carbon (or 3.5 g C yr^{-1} per square meter of the terrestrial catchment) constitutes the largest net input of carbon to the aquatic system. There is a limited export of carbon through the outlet and talik (0.1% of total export/storage), and instead most carbon instead leaves the aquatic system through evasion of CO_2 across lake surface (57% of the total export and storage of carbon). The carbon that isn't exported is accumulated in the sediment. Even if much more carbon is stored in terrestrial soils compared to the lake sediment, the long-term incorporation of carbon in permafrost soils and lake sediments are relatively similar per unit area (2.9 and 2.6 g C m^{-2} yr^{-1}, respectively). Hence, the three times higher carbon incorporation in the terrestrial system as compared to the aquatic system is largely a consequence of the larger terrestrial area (3500 vs. 940 kg C yr^{-1}). Finally, that Two-Boat Lake acts as a source for CO_2 to the atmosphere suggests that that even if the hydrologic transport of carbon from the terrestrial system to the lake (3.5 g C m^{-2} yr^{-1}) is lower than in boreal systems (6–12 g C m^{-2} yr-1; Jonsson et al., 2007), the lake is still net heterotrophic (i.e., the terrestrial input of carbon and energy is needed as a supplement to the aquatic primary production in order to sustain the aquatic food-web).

Other elements

When looking also at the other 42 elements (Figure 4-4) the mass-balance budget shows a few things that are worth pointing out.

First, also under the cold and dry conditions present at Two-Boat Lake, weathering is an important source for elements found in more easily weathered minerals (e.g., Ce, La). This is presumably a result of the relatively high input of fresh eolian material and the relatively undeveloped soils in the catchment (Rydberg et al. 2016).

Second, many of the elements that enter the system primarily with precipitation (e.g., Cl, I, B, K) show a considerable retention in the terrestrial system. This includes an easily dissolvable element like Cl and indicates that there are parts of the catchment where water leaves the soil through evapotranspiration rather than runoff and that in these spots there is an accumulation of Cl (e.g., in the form of organochlorides; Svensson et al. 2017). However, because other parts of the catchments are hydrologically connected to the aquatic system also during the unfrozen period, runoff still plays an important role for the transport of both weathering products (e.g., Ce and La) and elements from wet deposition (e.g., Cl, Br, Li, Ca, Mg).

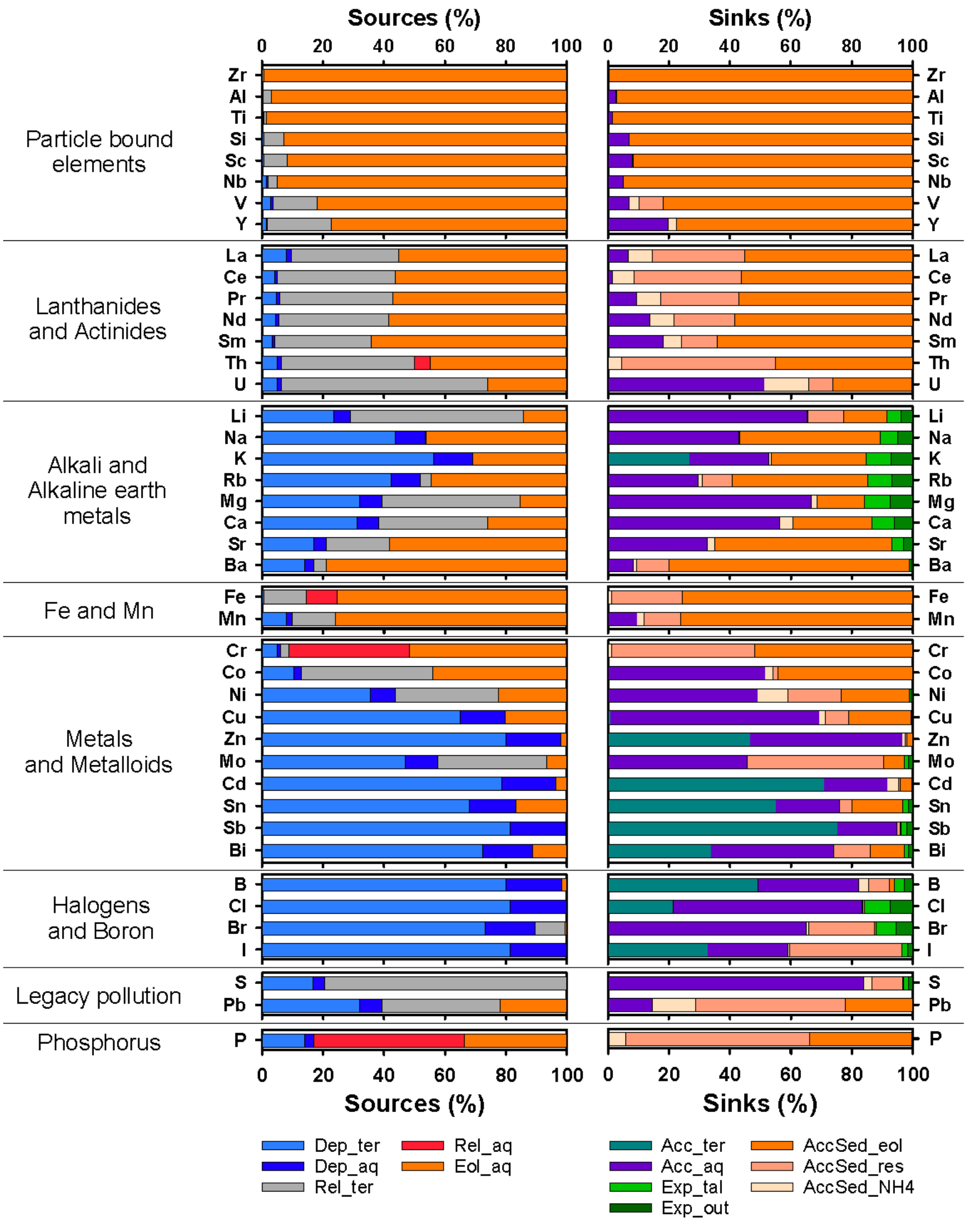

Figure 4-4. *The relative contribution of different sources and sinks in the Two-Boat Lake catchment for 42 elements studied using a mass-balance budget approach (Rydberg et al. 2023). The elements are divided into groups based on their respective behavior in the catchment. Figure from Rydberg et al. (2023).*

Third, there is a considerable release and hydrological transport of both S and Pb from the terrestrial system to the lake. Most of this release of S and Pb is not balanced by an input from the atmosphere, which suggests that it is related to a release of S and Pb from legacy pollution that accumulated in the catchment when the anthropogenic emissions of these elements were larger than today. Fourth, eolian processes is the main transport route for elements associated with mineral particles (e.g., Al, Ti, Si), and when deposited on the lake surface these elements tend to accumulate in the lake sediment.

Elemental transport with water

Based on both the chemistry and particle tracking in the hydrological modelling, the water in the active layer is young and relatively well mixed with depth (Rydberg et al. 2025). The age of the water is mostly less than a year, especially during the snowmelt period. There is a fraction of older water later in the summer season, but also in August most of the soil water has been in the ground for less than three years. There is no discernible trend in the chemical composition of the soil water with depth, which suggests that the water in the active layer generally is well mixed between different depths. Even if there isn't any distinct pool of old groundwater there is a positive correlation between the percentage of what the hydrological model classifies as deep groundwater and the concentration of elements released during weathering in the runoff to the lake. A potential source of this "deep" ground water is the melting older water in the deepest part of the active layer that has a different chemical signature than the water that comes from precipitation. Such a contribution can also be seen during the snowmelt period, when the isotopic data suggest that there is a significant contribution of water from thawing ground ice in the runoff. For elements that mainly enter the catchment with atmospheric deposition, there is a negative correlation with the total discharge, which indicates that these elements tend to be higher during low-flow situations. Presumably this has to do with evaporation that concentrates these elements in the water that enters the lake during periods with a lower runoff.

4.7 Lake sediment and sediment dynamics

On the landscape scale, a lake sediment constitutes a sink for both organic and minerogenic material. Depending on the nature of the element at hand this sink can be more or less permanent depending on the studied time scale and the conditions in the sediment (Rydberg and Martinez-Cortizas 2014).

What have we learned?

The sediments in Two-Boat Lake have a relatively low organic content and mostly consist of minerogenic particles. Sedimentation rate in the deepest part of the lake is about 0.013 and 0.026 kg m^{-2} yr^{-1} (Rydberg et al. 2016). Based on an analysis of a sediment profile and surface sediments from several lakes, we show that eolian activity varies both in time and space, but that it always is a very important transport pathway in this dry and cold environment (Rydberg et al. 2016). Closer to the ice sheet and during periods with high eolian activity the sediment is enriched in zirconium – an element that is predominately found in the coarse silt and sand fractions preferentially transported by wind. This showed that a lake's location in relation to source areas of wind-eroded material is of importance for the input of eolian material. The buffering capacity of this fresh eolian material likely contributes to the higher pH and alkalinity in lakes close to the front that has been observed in other studies (Ryves et al. 2002). This suggests that the catchment of Two-Boat Lake is not only defined by hydrology, and that the source area for material extends well beyond the drainage basin.

Looking across the sediment basin using 13 surface sediment cores there is considerable variability in both the geochemical composition and, e.g., the accumulation of carbon. For areas with sediment accumulation the carbon accumulation rate varied from 1.4 to 7.4 g C m^{-2} yr^{-1} largely following the water depth. Based on the bottom type the lake can be divided into areas with, i) hard rocky erosion bottoms without any sediment accumulation, ii) areas covered with benthic vegetation (Chara, Nostoc and bryophytes), and iii) deeper areas with no or limited benthic vegetation. For the two areas that accumulate sediment, the carbon accumulation rate was, on average, 4.1 g C m^{-2} yr^{-1} for the area with benthic vegetation and 2.9 g C m^{-2} yr^{-1} for the deeper area without benthic vegetation (Lindborg et al. 2020). In addition to carbon there is a considerable variability in the geochemical composition. The most striking difference is that sediment cores collected at water depths around 15-20 m have distinct concentration peaks in iron and manganese a couple of cm below the sediment surface. These peaks are most likely caused by a redistribution of iron and manganese in the sediment due to differences in redox conditions (Lindborg et al. 2016).

4.8 Biota

Vegetation constitutes a very dynamic catchment pool. Organic plant material is, through photosynthesis, constantly produced and subsequently degraded in both terrestrial soils and lake sediments. During plant growth carbon dioxide and a wide range of other elements are built into the molecular structure of plants, both intentionally (plant nutrients) and accidentally (pollutants and contaminants).

What have we learned?

The four main vegetation types in the Two-Boat Lake catchment have a considerable variation in the net primary production (NPP), with meadow having a higher NPP compared to dry grassland, dwarf shrub and wetland (217, 204, 147 and 72 g C m^{-2} yr^{-1}, respectively). Even if these values are much lower compared to boreal forests, the NPP in this dry Arctic landscape is within the range of estimates for other Arctic and Subarctic sites (Lindborg et al. 2020).

When looking at the carbon pools, all vegetation types at Two-Boat Lake have much lower carbon pools compared to boreal and temperate systems, where trees are a common plant type. The importance of woody tissues for the carbon stock is also apparent in this tree-less system, and the dwarf shrub vegetation type (2.5 kg C m^{-2}) have a carbon stock that is about 3–6 times higher than the other vegetation types (0.4, 0.5 and 0.9 kg C m^{-2}, for dry grassland, meadow and wetland, respectively). In many tundra systems wetlands constitute a major carbon pool, but in the very dry landscape around Two-Boat Lake wetlands do not contain as much carbon as in much wetter systems in other circumpolar areas (Hugelius et al. 2014).

In comparison to the terrestrial system, the oligotrophic nature of Two-Boat Lake itself results in a low aquatic NPP both in the benthic macrophyte communities (12 g C m^{-2} yr^{-1}) and the pelagic phytoplankton communities (3.7 g C m^{-3} yr^{-1}). In Two-Boat Lake, light penetrates down to about 15 m (Lindborg et al. 2016), hence, the productivity in the aquatic system is unlikely to be limited by light. Instead, the low productivity is likely caused by either nitrogen or phosphorus limitation. The presence of nitrogen fixating Nostoc colonies in Two-Boat Lake shows that the system is – at least partly – nitrogen limited. The mass-balance budget suggests a limited transfer of phosphorus to the lake from terrestrial soils and that there is an unknown source of phosphorus to the lake. This unknown source could be internal loading of phosphorus from the sediment. The bottom waters in Two-Boat Lake are generally anoxic, which would allow for a diffusive transport of phosphorus out of the sediment and back to the water column. However, it is also possible that the extra input of phosphorus is related to waterfowl that leave their droppings in and close to the lake while using it as a resting place (Rydberg et al. 2023).

Overall, the findings from GRASP show that the terrestrial vegetation under periglacial conditions function in a relatively similar way as a boreal or temperate ecosystem, but with a lower accumulation of carbon due to the lack of trees. Grassland communities are known to have a rapid turnover of organic matter, and it has been suggested that the lack of trees is compensated for by a higher stock of soil carbon. This does not seem to be the case in Two-Boat Lake, where the soil organic carbon stock down to 100 cm is lower in the grassland vegetation type (13 kg m^{-2}) compared to the wetland, shrub and meadow vegetation (36, 34 and 23 kg C m^{-2}, respectively).

5 Future conditions at Forsmark

The Greenland projects were established mainly because future climate conditions at Forsmark are expected to be similar to those at the Greenland sites. At present, the climate in Forsmark is significantly warmer and wetter than in western Greenland, with a MAAT of about 6.5ºC and an annual precipitation of about 580 mm (Section 3.3 in SKB 2023a). However, from observational records of past climate variability (paleo-proxy records) and from climate modelling, it can be inferred that Forsmark has experienced significantly colder climates over the last hundreds of thousands of years. A similar variability can be expected to re-emerge at some point within the next 1 million years (Myr), allowing for the formation of permafrost and development of ice sheets in the Forsmark region. In this chapter, changes in the Forsmark climate required for a transition into a periglacial or glacial climate are discussed (Section 5.1), followed by a discussion about the timing and duration of these colder climate conditions (Section 5.2). A brief overview of how the Forsmark landscape may look during future glacial and periglacial periods, as well as the regional geology and topography, is described in Sections 5.3 and 5.4.

The main objective of the descriptions in this chapter is to facilitate comparison between the future Forsmark site and the present-day Greenland sites (Chapter 4) and to support the evaluation of transferability of knowledge acquired from the Greenland projects to the Forsmark site (Chapter 6). To this end, this chapter only includes climate-related conditions that are of importance for the discussion in Chapter 6.

5.1 Required changes in local climate for enabling development of periglacial and glacial conditions at Forsmark

As explained in Section 4.2, permafrost is defined as ground that remains at or below 0°C isotherm for at least two consecutive years (French 2007). For most surfaces, the MAAT required for the development of frozen conditions typically must be a couple of degrees lower than the freezing temperatures in the ground (e.g. Hartikainen et al. 2010). Since the current MAAT in Forsmark is about 6.5°C, it is here assessed that the development of periglacial conditions in Forsmark may only occur if the MAAT decreases by 7ºC or more with respect to the present-day MAAT in Forsmark.

To enable ice-sheet growth over the Forsmark site, the MAAT needs to be even lower than for permafrost to develop (e.g. Section 1.2.3 in SKB 2020). This implies that, once the climate becomes gradually colder in Forsmark, periglacial conditions will develop prior to glacial conditions.

5.2 Timing of periglacial and glacial conditions

5.2.1 Climate variability

To evaluate the timing of the onset of periglacial and glacial conditions in Forsmark, it is essential to first consider the climate variability over timescales of 100 kyr to 1 Myr. The variability over these timescales can be inferred from paleo-proxy data, which shows how the climate has varied in the past. Over the last 800 kyr, Earth's climate has repeatedly alternated between warmer periods with limited ice cover (referred to as interglacials) and colder periods with abundant ice cover (glacials) over a characteristic timescale of approximately 100 kyr. To illustrate these variations, stacked $\delta^{18}O$ measured in marine sediment records is displayed in Figure 5-1a. $\delta^{18}O$ is a proxy for global ice volume. In general, high $\delta^{18}O$ values correspond to cold glacial periods with abundant water locked up in continental ice sheets, whereas low values correspond to warm interglacial periods with considerably less water locked up in the ice sheets. Figure 5-1a also shows that ice sheets tend to grow slowly and progressively over the course of a glacial cycle, with ice sheet advances during relatively cold phases and reduced ice growth, or even ice-sheet retreat, during warmer phases. This pattern of ice growth explains the general saw-tooth shape of the global ice-volume evolution of the late Quaternary (Figure 5-1a), with relatively long periods (80–100 kyr) of ice sheet growth, followed by relatively short phases (10–20 kyr) of ice sheet deglaciation and subsequent interglacial conditions.

The climate variability over timescales of 100 kyr or longer is primarily controlled by (i) the incoming solar radiation in the summer season (often referred to "Milankovitch forcing"), and (ii) variations of atmosphere greenhouse gases (see e.g. Section 3.4 in SKB 2023a for a further discussion on these forcing agents). The evolution of these forcing agents over the last 800 kyr is shown in Figure 5-1b,c. In general, low summer insolation and low atmospheric CO_2 levels promote ice-sheet growth, whereas high values inhibit it (see further Section 3.4.5 in SKB 2023a). Internal climate feedbacks, caused for example by enhanced reflection of solar radiation by the growing ice sheets, and a lowering of the ice sheet surface due to bedrock depression, have also been found important for controlling the dynamics of the 100-kyr glacial-interglacial cycles (e.g. Abe-Ouchi et al. 2013).

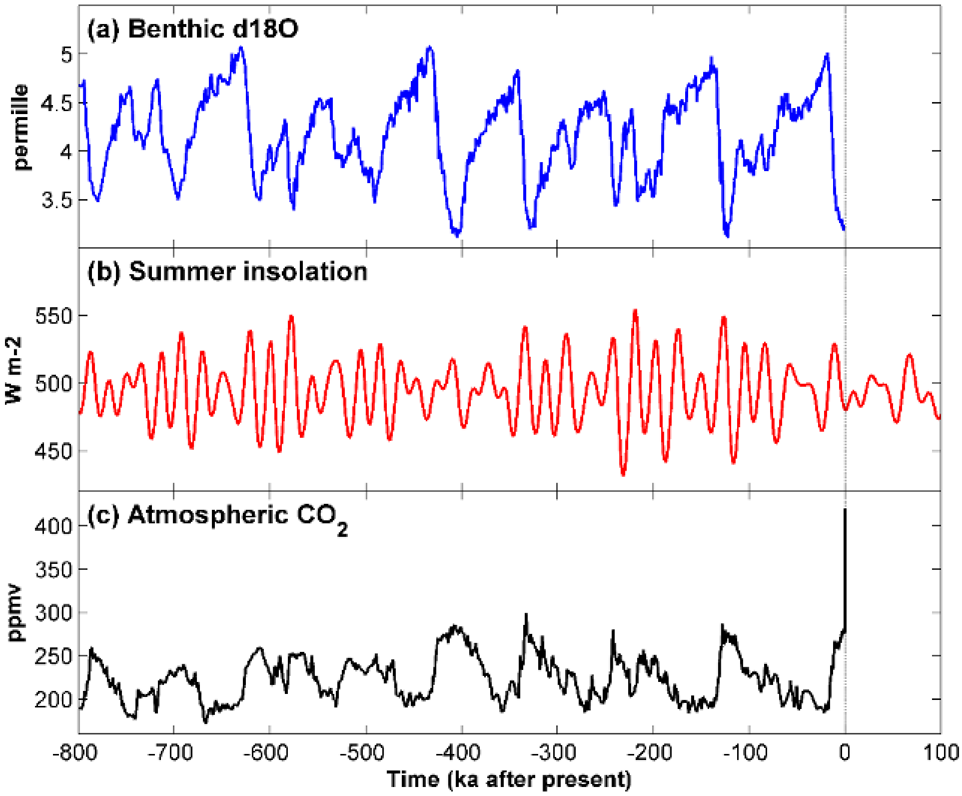

Figure 5-1. *(a) Proxy data for global ice volume from the stack of benthic $\delta^{18}O$ records of Lisiecki and Raymo (2005). High $\delta^{18}O$ values correspond to cold glacial periods whereas low values correspond to warm interglacial periods. (b) Maximum (solstice) summer insolation at 65°N from 800 kyr before present until 100 kyr after present (Laskar et al. 2004). (c) CO_2 variations over the past 800 kyr. Composite record from Lüthi et al. (2008) complemented with the observed atmospheric CO_2 concentration in 2022 (419 ppmv; NOAA 2023).*

5.2.2 First future occurrence of periglacial conditions

At present, the MAAT in Forsmark is ~6.5ºC, i.e. well above freezing although the summer insolation is currently at local minimum compared to the average conditions during the last 800 kyr (Figure 5-1b) This high MAAT is maintained by current interglacial conditions (i.e. a relatively low ice cover on Earth) and by enhanced concentrations of anthropogenic greenhouse gases in the atmosphere (mainly CO_2). Greenhouse-gas emissions have contributed to a mean annual warming of about 2ºC in Sweden since pre-industrial times (Schimanke et al. 2022). Thus, even in the absence of any anthropogenic contribution, the MAAT in Forsmark would still be considerably higher than the MAAT required for periglacial conditions to occur (Section 5.1). Therefore, the cooling required for periglacial (or glacial) conditions to occur in Forsmark can likely only be achieved if it is embedded in a planetary-scale cooling trend that would result in ice sheet growth elsewhere on Earth. Thus, it is assessed that the next potential period of periglacial conditions in Forsmark can earliest occur in conjunction with the onset of the next glaciation in the Northern Hemisphere (see also Section 4.3 in SKB 2023a); this period is commonly referred to as the "next glacial inception".

The next glacial inception is primarily controlled by summer insolation and the concentrations of atmospheric CO_2. Future variations in the summer insolation can be predicted accurately in accordance with the Milankovitch theory. According to these predictions, the next 50 kyr will be characterized by a remarkably low insolation variability, and, as a result, the summer insolation during this period will be higher than at present (Figure 5-1b). The next major insolation minimum occurs at around 54 kyr after present. Consequently, this time marks the first future possibility of glacial inception, and thus also the first possible occurrence of periglacial conditions in Forsmark (e.g. Archer and Ganopolski 2005, Ganopolski et al. 2016, Lord et al. 2019, Liakka et al. 2021, Talento and Ganopolski 2021, Williams et al. 2022). However, the possibility of achieving glacial inception in conjunction with the 54-kyr insolation minimum is also dependent on the cumulative (historical, present and future) anthropogenic carbon emissions to the atmosphere. Unlike most greenhouse gases, CO_2 is not chemically destroyed in the atmosphere; instead, it is gradually removed by various processes operating over timescales of 10^1 – 10^6 years (e.g. Lord et al. 2016). As a result, about 15–30 % of cumulative anthropogenic carbon emissions are projected to remain in the atmosphere after 10 kyr, and 5–10% after 100 kyr (Archer and Ganopolski 2005; Archer et al. 2009; Lord et al. 2016, 2019; Colbourn et al. 2015). Therefore, if emissions over the coming centuries are sufficiently high, atmospheric CO_2 concentrations around the 54-kyr insolation minimum will be elevated to such an extent that the next glacial inception is avoided at this time. If this were to occur, the next glacial inception is projected to occur after 100 kyr. In the most extreme scenarios, which assume that the humanity will emit most, if not all, of the currently known fossil-fuel reserves, the next glacial inception may even be delayed for more than 500 kyr compared to the case with no anthropogenic emissions (e.g. Archer and Ganopolski 2005, Lord et al. 2015, 2019). Most studies suggest that cumulative anthropogenic carbon emissions in the range of 1000– 1500 petagrams of carbon (Pg C) is sufficient to avoid glacial inception at 54 kyr after present (Archer and Ganopolski 2005, Ganopolski et al. 2016, Lord et al. 2015, 2019). 1500 Pg C approximately corresponds to cumulative emissions in typical "business-as-usual" scenarios, which assume that current levels of anthropogenic emissions will continue for the remainder of the century after which they decline (Section 4.2 in SKB 2023a). At present, humanity has emitted a total amount of around 700 Pg C to the atmosphere (Friedlingstein et al. 2022).

In summary, the next glacial inception will most likely occur after 100 kyr after present if similar-to-present (i.e. business-as-usual), or higher, levels of emissions were to continue for the remainder of this century, and around 54 kyr after present under strong climate-mitigation scenarios (e.g. if the Paris agreement is met), see further Sections 3.4.5 and 4.3 in SKB (2023a).

5.2.3 First future occurrence of glacial conditions

As discussed in Section 5.1, the first occurrence of glacial conditions in Forsmark may be significantly delayed compared to the first occurrence of periglacial conditions. This notion is also supported by reconstructions of the last glacial cycle, during which first ice sheet coverage in Forsmark occurred some 50 kyr after the initial glacial inception (Section 4.4 in SKB 2023a). A similar delay has also been projected after a potential glacial inception in conjunction with the future 54-kyr insolation minimum (Williams et al. 2022, Liakka et al. 2021, 2024), implying that the next possibility of ice-sheet coverage in the Forsmark region can be expected earliest around 100 kyr after present. Naturally, if the next occurrence of periglacial conditions in Forsmark is pushed further into the future due to anthropogenic emissions, the next occurrence of glacial conditions will also be delayed with a similar magnitude.

5.2.4 Climate variability after the first future occurrence of glacial conditions

Once the Forsmark site has been glaciated, the climate is expected to be dominated by a similar variability to that of the past 800 kyr, with recurrent glacial-interglacial cycles on characteristic timescales of ~100 kyr (Section 5.2.1). As in recent glacial cycles, long-term climate projections indicate predominately cold conditions during future cycles. According to Williams et al. (2022), the MAAT will remain below 0ºC for about 65–75% of the time between the next glacial inception and 1 Myr after present. These results emphasize that the present long interglacial represents an anomaly rather than a typical feature on timescales of 100 kyr to 1 Myr. Following the next glacial inception, the climate is therefore expected to be characterized by shorter interglacials (~10–20 kyr), separated by longer intervals dominated by periglacial or glacial conditions, as has been observed over the past 800 kyr.

Duration of glacial periods

An analysis of future ice-sheet variability in Forsmark (Liakka et al. 2024) found that the duration of individual periods with glacial conditions at Forsmark may range between 10 kyr and 70 kyr over the next 1 Myr. This duration depends on internal climate variability, anthropogenic carbon emissions, and the sensitivity of ice-sheet growth to climate change. In the most likely projections, the duration of an individual glaciation is projected to be around 20–30 kyr under low, or no, anthropogenic carbon emissions, and around 10–20 kyr under business-as-usual, or higher, emissions within this century (Liakka et al. 2024). According to the same analysis, the likely total duration of glacial conditions at the Forsmark site over the next 1 Myr is projected to be in the range 120–300 kyr under low or no anthropogenic emissions, and 50–220 kyr under business-as-usual or higher emissions.

MAAT and precipitation

Modelling projections suggest that a wide range of MAATs are compatible with periglacial and glacial conditions in Forsmark (Kjellström et al. 2009, Williams et al. 2022). For example, the projections of Williams et al. (2022) suggest that the MAAT may become as low as –20ºC during the coldest periods. Kjellström et al. (2009) found that the annual precipitation in a periglacial climate in the Forsmark area is approximately halved compared to present-day values. The resulting rate of precipitation decreases with respect to a decrease in MAAT in their study is consistent with the associated rate of increase in a warmer climate (~3 % per ºC MAAT change), see e.g. Appendix B in SKB (SKB 2023a) and Gutiérrez et al. (2021).

Rate of ice sheet growth, retreat and ice-marginal stillstands

The rates of ice-sheet growth and retreat at Forsmark during future glaciations, i.e. how quickly the ice-sheet margin advances over and withdraws from the area, is uncertain. Nevertheless, since overall variability is expected to resemble that of past glaciations, it is reasonable to assume rates comparable to those of the last glacial cycle (the Weichselian). During the Weichselian glaciation, the rate of ice-sheet advance is estimated to have been approximately 50 m yr^{-1}, while the rate of retreat is estimated to have been approximately 300 m yr^{-1} (Section 4.5.4 in SKB 2020). The occurrence of ice-marginal stillstands over the Forsmark site, defined as temporary halts of ice-sheet margin during e.g. deglaciation, cannot be excluded during future glaciations. During the last deglaciation, ice-marginal stillstands of up to 1000 years in Fennoscandia have been inferred based on geological information (Section 4.2.2 in SKB 2020). Significantly longer standstills than 1000 years are unlikely in the light of a constantly varying climate on these timescales.

Basal thermal conditions of the ice sheet

An ice sheet that grows over Forsmark may have frozen or thawed basal conditions, or a combination of both. Frozen conditions are more likely during ice-sheet advance, as this situation requires the climate to be sufficiently cold to enable ice growth. Conversely, during periods of ice-sheet retreat, the ice sheet will most likely be characterized by predominately thawed basal conditions. The thermal conditions beneath the ice sheet also depend on the ice-sheet dynamics and the geothermal heat flux. The geothermal heat flux has been estimated to be 61 mW m^{-2} at Forsmark, which is slightly higher than the estimated average value over entire Fennoscandia (~45 mW m^{-2}; Näslund et al. 2005).

5.3 Landscape development

At the time of the first future periglacial period, the landscape around Forsmark is expected to look remarkably different compared to at present. This is because the isostatic rebound of the Earth's crust following the last deglaciation is expected to be virtually completed by the time of the first potential period of periglacial conditions (after 50 kyr) (Section 3.5.4 in SKB 2023a). Thus, after 50 kyr, (i) the coastline will have migrated substantially compared to the present (Section 4.7.2 in SKB 2023b), and (ii) virtually all lakes that were created due to the shoreline regression will have disappeared from the area due to ingrowth and sedimentation (Figure 4-5 in SKB 2023b). Since the first future periglacial period in Forsmark is projected to occur in response to the 54-kyr insolation minimum (Section 5.2.2), i.e. after virtually all current lakes have disappeared from the area, formation of taliks beneath lakes (especially deep through taliks) are unlikely in the Forsmark region within the next 100 kyr. Taliks may still exist under water courses, provided that the climate is not too dry (Hornum 2023).

Once an ice sheet again grows over the Forsmark area, the underlying bedrock, which at this stage has virtually no remaining glacial rebound, will again be isostatically depressed by the ice sheet load. Although the extent and thickness of any future ice sheets in the area are inherently uncertain and, thus, cannot be predicted with accuracy, reconstructions of the last glacial cycle suggest that most ice sheets that are large enough to expand over the Forsmark site will also be sufficiently thick such that the bedrock beneath the ice sheet will be submerged beneath the sea (e.g. Section 3.3.4 in SKB 2020). This would result in the Forsmark site becoming submerged beneath the Baltic Sea following the subsequent deglaciation of the area. A situation characterized by a relatively thin ice sheet situated over the site, resulting in terrestrial conditions following deglaciation, cannot be excluded, although this situation is considered significantly less likely than a development resulting in post-glacial submerged conditions.

After any future deglaciation, lakes basins will again be created in the Forsmark area. Since future interglacial periods are projected to be considerably shorter than the current one, leading to less time for lake ingrowth and sedimentation, deep through taliks under lakes is expected to occur in the region following future glaciations, i.e. earliest after 100–150 kyr into the future.

5.4 Topography

Changes in topography occur over time as a result of tectonic/isostatic uplift and subsidence as well as by downwearing of the Earth's surface by weathering and erosion processes. The combined effect of all weathering and erosion processes is referred to as denudation, and the most dominant erosion process in the Forsmark area, on a glacial cycle time scale, is glacial erosion by an overriding ice sheet.

The landscape in Forsmark is currently very flat (it has a low topographic relief). A likely consequence of future glacial erosion is that the relief will increase in the Forsmark area, at least locally (e.g. Krabbendam et al. 2022). However, it is estimated that the relief in the Forsmark area can increase by at most some tens of meters due to glacial erosion (Hall et al. 2019, Krabbendam et al. 2022).

5.5 Differences and similarities between future Forsmark and current Greenland study area

As outlined in Section 3.1, the study region in Greenland features an ice sheet with a largely stationary ice margin. Over a year, ablation currently exceeds accumulation, leading to a thinning of the ice sheet and an abundance of meltwater on the surface and beneath the ice. Moving toward the interior of the ice sheet, the amount of meltwater gradually decreases (Section 4.1), where a shift from melted to frozen basal conditions eventually occurs. There are indications of subglacial permafrost at the ice margin (Section 4.1), possibly extending a few kilometers under the ice (Section 4.2). At Forsmark, a similar scenario to that of the current situation in the Greenland study area may arise when the ice margin is positioned in the area following a retreat of the ice sheet, which is not expected to occur until at least 100 kyr after present. To be directly comparable with present-day conditions in Greenland, this ice-sheet retreat would need to be significantly slower than the 300 m yr^{-1} modelled for the last deglaciation at Forsmark, or the ice margin would even need to experience a temporary halt (stillstand). Several similar standstills occurred during the Weichselian deglaciation when the ice margin was located some 200-250 km south of Forsmark (Section 4.2.2 in SKB 2020).

In addition to this expected climate-related difference, there are other site-specific differences that may be relevant when considering how processes and conditions studied in Greenland could apply to a future Forsmark (Chapter 6). The most important of these are briefly described in the following.

Topography: the relief in Forsmark (a few tens of meters) is one order of magnitude lower than the relief in the Greenland study region. Even after a potential increase of the relief due to future glaciations in Forsmark, the relief will be significantly less pronounced than observed at the Greenland sites (SKB 2020, Appendix G).

Geology: the Greenland site shares many similarities with the Forsmark site in terms of both sites being crystalline low permeable bedrock sites with comparable fracture densities. However, at the Greenland study region, groundwater flow and recharge is likely more controlled by vertical fractures and deformation zones than at present-day Forsmark. A typical feature in Forsmark is the sheet joints that occur in the upper tens of meters, and in cases down to approximately 150 m of the bedrock. These fractures developed due to stress release in the bedrock parallel to the ground surface and are interpreted as high-transmissive hydraulic features that allow and promote lateral groundwater movement (Follin, 2008; Vidstrand et al. 2010. Sheet joints have not been identified in the GAP drill cores and are only sporadically seen at the surface in the field (Engström and Klint 2014).

Geothermal heat flux and thermal conductivity: Permafrost thickness is sensitive to geothermal heat flux and thermal conductivity (e.g. Willeit and Ganapolski 2015). The mean value of thermal conductivity at the Forsmark site is 3.6 W m^{-1} K^{-1} (Sundberg et al. 2009), whereas the average thermal conductivity for the drill cores in the GAP area is 2.5 W m^{-1} K^{-1} (Claesson Liljedahl et al. 2016). Likewise, the geothermal heat flux differs between the sites. The geothermal heat flux for the GAP bedrock borehole is 27 to 35 mW m^{-2} (Section 4.1.4), significantly lower than the 61 mW m^{-2} for the Forsmark site (Sundberg et al. 2009). The lower heat flux and thermal conductivity at the Greenland study region could potentially contribute to a thicker permafrost and greater ice sheet thickness compared to a site with a higher heat flux and higher thermal conductivity like Forsmark.

6 Transferability of knowledge from Greenland projects to Forsmark, and remaining uncertainties

In this chapter, the transferability of the processes studied in the Greenland projects (Chapter 4) to a future Forsmark (Chapter 5) is discussed. It should be noted that transferability is assessed through qualitative discussions rather than quantitative statements, focusing on the conceptual understanding of how these processes function and how they could be relevant in future scenarios. One reason for this approach is that there is much less data available for the Greenland study sites compared to Forsmark. Rather than making direct comparisons between the two sites, the Greenland data serves to improve conceptual process understanding and to inform what insights can be transferred to a future glacial or periglacial Forsmark. When data for numerical modeling within the Greenland projects have been lacking, proxy datasets from Forsmark have been used. This was the case for the modeling of the deep groundwater system within GAP, where little to no data existed on the hydraulic properties of the bedrock. Instead, parameters from Forsmark were applied, while a fracture model specific to the Greenland study area was used.

In the context of transferability, a significant difference between current Greenland and future Forsmark is that the Greenland study region has experienced long periods of glacial conditions (Strunk et al. 2017; Bierman et al. 2014; Christ et al. 2021) supporting glacial meltwater recharge, possibly for several hundreds of thousands of years. This is in line with the residence times of deep groundwater at the Greenland site (Claesson Liljedahl et al. 2016). At sites like Forsmark, shorter glacial periods have occurred in the past and are expected in the future (Section 5.2.4, Liakka et al. 2024), which leads to mixed recharge sources including meteoric recharge during deglaciated periods. Whether a site experience extended glacial meltwater recharge or mixed recharge has significant implications for groundwater chemistry and the hydrogeological evolution. Although modeling suggests recurrent glacial conditions in future Forsmark (Section 5.2.4), these periods are expected to be shorter.

6.1 Ice sheet processes

6.1.1 Physical conditions of the ice sheet bed

The subglacial layer plays a key hydraulic role in draining meltwater underneath the ice sheet. If sediments are present at the ice-bed interface, water flows within a sediment layer compared to if the base of the ice is in direct contact with the underlying bedrock. The findings from the Greenland projects suggest predominantly hard bed conditions with thin and discontinuous veneer of basal sediment. In other places in Greenland soft bed conditions occur, with thicker sediment cover. Although the bed conditions are more varied than previously estimated, assuming mixed hard bedrock and thick sediment-covered basal conditions dictated by the subglacial topography is likely appropriate.

6.1.2 Pressure in the subglacial drainage system

Basal meltwater pressure is primarily determined by the thickness of the overlying ice sheet, meaning that ice overburden pressure increases with ice thickness. As a result, basal hydraulic pressure rises toward the center of the ice sheet. Because the ice thicknesses currently observed in Greenland are comparable to reconstructed past ice sheets at Forsmark (e.g. Section 4.4.1 in SKB 2020), basal meltwater pressures observed in Greenland are generally transferable to Forsmark.

Groundwater flow is governed by hydraulic head, which accounts for both pressure and elevation. Thus, basal pressure alone is not a correct boundary condition for groundwater flow models. Hydraulic head can be reliably estimated from ice surface and bed geometry. This makes it a physically meaningful boundary condition, especially since the hydraulic gradient beneath an ice sheet is primarily controlled by surface slope rather than ice thickness.

In this context, ice overburden hydraulic pressure — corresponding to approximately 92% of the ice thickness — can be used to estimate the pressure component of hydraulic head, and thus serves as a reasonable average representation of basal hydraulic pressure over a year (Harper et al. 2016; Claesson Liljedahl et al. 2016). This holds true in most situations, except when the bed's topographic relief exceeds ten times the gradient of the ice sheet surface, in which case the bed slope dictates the direction of water flow (Cuffey and Paterson 2010). Such conditions are unlikely in Forsmark, though parts of the deep bedrock trough (Isunnguata Sermia) meet this criterion. Applying this to Forsmark, ice overburden hydraulic pressure remains a valid assumption, as the topography is expected to remain relatively flat also in the future (Section 3.5 and Appendix G in SKB 2020).

6.1.3 Characteristics of the subglacial drainage system

The findings from the Greenland projects indicate that surface meltwater drives the entire ablation zone by feeding a dynamic basal drainage system composed of discrete components with variable transmissivity. Based on this, it is justified to assume that the entire ablation zone in a future Forsmark glacial setting should have liquid water at the base of the ice sheet.

6.1.4 Basal thermal distribution and groundwater formation

The ice sheet in the Greenland study region mostly exhibits melted basal thermal conditions, which is also expected in Forsmark during future periods of ice-sheet retreat or stillstands in the region (Section 5.2.4). Groundwater formation during glacial conditions is likely to be comparable between the two sites given the low permeability crystalline bedrock and similar fracture densities (Section 5.5). However, the Forsmark site possesses a large number of gently dipping structures including sheet joints; structures that facilitate lateral groundwater movement which can enhance recharge and connectivity over large distances compared to the Greenland site where meltwater infiltration and groundwater formation are likely more localized and restricted to vertical structures.

Groundwater flow modelling under transient conditions of the Greenland study site using hydraulic properties from Forsmark suggests that groundwater flow and recharge is strongly dependent on the hydraulic parametrization of the rock domain, deformation zones and the permafrost (Jaquet et al. 2019).

Due to the higher geothermal heat flow at Forsmark (Section 5.5), more basal meltwater can be anticipated than in the Greenland study area. However, modeling has shown that bedrock in general has a limited capacity for recharging meltwater and groundwater formation (e.g. Bockgård, 2004). Although more meltwater may be generated at Forsmark than in Greenland, this does not imply that infiltration and groundwater recharge will be higher in Forsmark. Instead, a significant portion of the meltwater is likely to flow laterally or remain within the surface drainage system.

The current distribution of meltwater and subglacial drainage under the Greenland ice sheet is a result of past and present climatological and glaciological conditions. In addition, local variations in elevation and hydraulic properties influence the hydrological systems in the bedrock, particularly areas of groundwater recharge and discharge. Similar conditions may have existed during past deglaciations of the Fennoscandian ice sheet and may therefore also exist in the future. However, during periods of ice-sheet advance at Forsmark, the distribution of hydrological zones may differ from those suggested for the Greenland ice sheet. During these periods, the surface melt would likely be relatively small, resulting in less available water. Additionally, the advancing ice sheet margin might experience basal frozen conditions due to a colder climate, unlike what is currently observed in the Greenland ice sheet.

6.2 Permafrost

Observation from the Greenland study sites has been used to evaluate the models used for projecting future permafrost thickness at Forsmark. Specifically, data on bedrock and soil temperatures from the GAP boreholes and the Two-Boat Lake catchment were used in a study by Hartikainen et al. (2022) to investigate how simulated ground surface temperatures, permafrost depths and bedrock temperatures from a numerical permafrost model compare with the observed data from a site with continuous permafrost conditions. In the study by Hartikainen et al. 2022, two model versions were evaluated that previously have been used in post-closure safety assessments at SKB. The study found that simulated ground temperatures were similar, within 1–2ºC degrees, to the observed ones.

Apart from this, directly transferring permafrost conditions from the Greenland study area to a future Forsmark is challenging due to differences in topography, historical ice-sheet development, and geothermal heat flux. The higher relief at the Greenland sites suggests that spatial variations of the permafrost depth will likely be greater compared to Forsmark, and the lower geothermal heat flux in Greenland makes it easier for permafrost to develop in this area. This means that the observed permafrost depth in Greenland is likely in the upper range of what can be expected in Forsmark under similar air temperatures.

The more varying topography in Greenland likely also results in a more uneven permafrost thickness compared to Forsmark with its smoother relief. The flatter topography at Forsmark likely leads to less extensive fluvial erosion and transport during frontal passages. When the relief in Forsmark increases in the future due to glacial erosion, the topographically induced stress field is expected to reach deeper than it does today. Increased relief is also expected to affect groundwater flow in the shallow bedrock as well as permafrost depth.

Another important insight from the Greenland projects is that ice margin stillstands can be associated with permafrost beneath or near the ice sheet. Notably, in the Preliminary Safety Assessment Report (PSAR) for the spent fuel repository, stillstand scenarios did not account for such a potential permafrost layer (Section 4.5.1 in SKB 2020). However, this is likely conservative for post-closure safety, as a permafrost layer would generally reduce groundwater flow within the bedrock beneath the ice sheet.

The studies in the catchment of Two-Boat Lake show that the active layer develops faster in wetter areas during spring snowmelt, but that the maximum active layer depth in late summer is larger in dry areas. Because this is a consequence of the large ability of water to store (and release) heat, it is reasonable to assume that this difference is transferable to the future Forsmark.

6.3 Depth of glacial meltwater infiltration/recharge

Observations from the Greenland study region confirm that glacial meltwater can reach depths of at least 500 m below ground surface. This finding is supported by transient groundwater flow modelling by Jaquet et al. (2019), which concludes that glacial meltwater infiltrates to depths greater than 500 m regardless of what part of a glacial period considered, when applying the hydraulic parameters from the Forsmark site. Based on the available data and geological setting as well as the modelling, there is no indication that glacial meltwater would reach shallower depths under glacial conditions at Forsmark. The study by Jaquet et al. (2019) also suggests that only taliks located within a few kilometers of the ice margin can be affected by discharging glacial meltwater.

6.4 Chemical composition of water at DGR equivalent depths

The chemical composition of groundwater is influenced by various local factors and is therefore not directly transferable to Forsmark. One key finding from the Greenland studies is that although dilute meltwaters reach great depths, oxygenated waters do not seem to infiltrate to great depths close to the ice margin. This finding confirms the results from modelling oxygen ingress at the Forsmark site (Sidborn and Neretniks 2008, Sidborn et al. 2010, Spiessl et al. 2008). That is, oxygenated waters have not been found at great depths in Greenland and is not expected in the future at Forsmark. Modelling has shown that the amount of glacial meltwater at depth is sensitive to the length of the ice covered periods and the hydraulic parametrization of the rock domain. The case using Forsmark hydraulic parametrization gave the least concentrations of glacial meltwater at depth, compared to the case with Laxemar parametrisation (Jaquet et al. 2019).

The long residence times of waters at 500 m depth in Greenland have allowed for extensive water-rock interaction. This has resulted in groundwater losing many of the most reactive components, such as oxygen. The salinity of the deep groundwater in Greenland observed in the 650 m deep DH-GAP04 borehole is approximately 3000 mg/L in total dissolved solids (TDS) (Bomberg et al. 2019), equivalent to modelled salinity concentrations during ice sheet retreat for the glacial reference cycle evolution at Forsmark (Section 10.4.7 in SKB 2022). At the same time, since Forsmark is currently located near the sea and is expected to be periodically submerged in the future (Section 5.3), saline seawater could intrude, potentially increasing salinity.

6.5 Presence of taliks in periglacial/proglacial landscapes

Direct observations during the Greenland projects have, for the first time, confirmed, that through taliks do occur in periglacial and proglacial landscapes. This shows that through taliks can develop beneath lakes also in a future periglacial landscape at Forsmark. The hydrological modeling indicates that the talik under Two-Boat Lake is balancing between having a recharging (downward) net flow during most years, and a discharging (upward) net flow under very dry years when the lake level drops. This is also supported by the presence of water with an isotopic signature that suggests input of glacial meltwater to the talik. Because discharge of water through the talik is affected by the altitudinal difference between the lake and the ice sheet, the flatter topography in the future Forsmark compared to Greenland implies that discharging conditions will be more common in Forsmark compared to in Two-Boat Lake.

From studying a single lake, i.e. Two-Boat Lake, it cannot be determined how common it is that through taliks develop beneath lakes. However, based on the modeling done in SKB (2006), many of the thousands of lakes in front of the ice sheet in Greenland are large enough to sustain a through talik. In addition, when Hornum (2023) studied winter-active springs along the Kugssup Alangua valley, they could show that these springs are connected to the deep, sub-permafrost, groundwater system. First, this shows that taliks and connection to the deep groundwater system are not unique to Two-Boat Lake. Second, it suggests that through taliks are not limited to form beneath lakes. This implies that if there are larger rivers in the future periglacial Forsmark, these also need to be considered as possible conduits between the deep groundwater system and the surface system. In addition, modeling studies in other areas have shown that taliks can form also beneath peatlands (Devoie et al. 2019, Cohen-Corticchiato and Zwinger 2021).

6.6 Hydrological modelling and biogeochemical cycling of elements in the surface system

Both observations and the hydrological modeling of the surface system in the active layer in the Two-Boat Lake catchment reveal a system with limited flow of surface water, especially during the non-snowmelt period. This is largely a consequence of the dry conditions in the Kangerlussuaq region rather than an effect of the presence of permafrost. Hence, whether these conditions are relevant to the future Forsmark largely depends on humidity rather than temperature. If Forsmark becomes a dry arctic tundra, the lower relief and slower flow of water might make the effect of the evapotranspiration even more pronounced (longer transit time for the water). This could potentially lead to a greater accumulation of elements that are water soluble and that would be transported from the terrestrial system to the lake in more humid areas.

Even if the permafrost is not directly related to the limited flow of surface water, it limits the groundwater pool to the thin active layer. In the Two-Boat Lake catchment, this effect can be seen in several ways. First, the water in the active layer is young, generally less than a year and rarely older than three years. Second, the shallow nature of the active layer also means that there is no systematic variation in the chemical composition of the groundwater with depth, i.e., the groundwater is relatively well mixed. Because this is related mainly to the thickness of the active layer and the lack of a deep groundwater pool, the young nature and homogeneous chemical signature of the groundwater in the active layer should be applicable also to a periglacial Forsmark.

As for the hydrological transport of elements and substances, it should be noted that even if the transport is limited, it makes an important contribution to the aquatic system. This can be seen in both the importance of terrestrial carbon for sustaining the aquatic food web (i.e., the lake is net heterotrophic), and the considerable transport of legacy pollution (Pb and S) from the terrestrial system to the lake. What the situation will be like in the future Forsmark is, again, largely a matter of how dry the future periglacial Forsmark will be. If the water balance is less negative compared to Two-Boat Lake, there might be permanent streams and therefore a larger hydrological transport in the future Forsmark. On the one hand, the presence of permanent streams also implies a transport of particulate material in the future Forsmark, but on the other hand, the lower relief in Forsmark implies a slower flow of water and a lower erosive power of the stream water.

When it comes to the cycling of carbon, the data from Two-Boat Lake suggests a system that is rather similar to other arctic tundra environments, and therefore it is reasonable that a future periglacial Forsmark will be similar to present-day arctic systems. In the Two-Boat Lake catchment, the accumulation of carbon per unit area is relatively similar between the permafrost and the lake sediment. Even if the relation between the two main carbon pools likely is not directly transferable to the future Forsmark, it is likely that both the permafrost soils and the lake sediments will be important for the carbon sequestration also in a future periglacial Forsmark.

For elements supplied by weathering, an important aspect in Two-Boat Lake is the input of fresh – unweathered – eolian material and the young weakly developed soils. Whether this situation will be relevant to the future Forsmark depends on several things. First, will there be any source areas for eolian material. In Greenland the source areas are mainly the glacial outwash plains close to the front of the ice sheet. These areas need time to develop and are likely a consequence of the fact that the ice front is currently stationary. Second, eolian activity is favored by dry conditions, especially during winter (Willemse et al. 2003). Third, the proximity to the ice sheet creates katabatic winds that increase the eolian activity close to the ice front. If these criteria are fulfilled, it is reasonable to assume that the conditions in Two-Boat Lake are relevant for the future Forsmark. Another thing to consider is whether the periglacial conditions in Forsmark are expected to occur at the end of a warm period (i.e., in front of an advancing ice sheet), we can expect that the soils will be well developed, e.g., podzols, inceptisols and histosols. This implies that the mineral material in the catchment already has undergone significant weathering, and that the weathering rates will be slower with less input of weathering products to the aquatic system compared to in Two-Boat Lake.

6.7 Lake sediment and sediment dynamics

As mentioned above, close to the ice sheet in Greenland there is an abundant supply of eolian inorganic material due to the large glacial out-wash plains. This material is readily moved by wind and has been deposited both on land – where it forms the upper layer of the regolith – and on lakes where it rapidly accumulates in the lacustrine sediment. This results in a relatively high sedimentation rate in Two-Boat Lake when also considering the low productivity in the lake and limited hydrological transport of particles. The high input of minerogenic particles also results in a low concentration of organic material in the sediment. How relevant this is for the situation in a future periglacial Forsmark will largely depend on the eolian activity. If the eolian activity is of similar magnitude as in Greenland, we can expect similar conditions in the future Forsmark. If there is a lack of eolian material, it can – provided everything else is kept the same – be expected that the sedimentation rate would be lower and the sediment would be more organic.

When it comes to the transferability of lakes, lake sediments and sediment dynamics, a key question is what type of periglacial landscape is considered for the future Forsmark, i.e., whether it is a "new" deglaciated landscape (without a marine stage) or an "older" landscape in front of an advancing inland ice sheet. In the former case we can expect to have a high density of lakes (even if the lower relief might give a lower density compared to the 14% of the land surface that is observed in Greenland). That is, we will have a landscape with a mixture of terrestrial areas and lakes, where all processes related to lakes and aquatic systems are highly relevant. In the latter case we can expect that most lakes formed after deglaciation would be infilled with sediment and transformed into peatlands. In this case, all processes linked to the aquatic system will be of less importance, and instead a thorough understanding regarding the function of peatlands in a periglacial setting is relevant (which is something that was not studied in the Greenland projects).

6.8 Biota

Both the terrestrial and aquatic ecosystems in a future periglacial environment in Forsmark are expected to resemble those observed at the Greenland sites. One exception is that the closer proximity to the ocean in Forsmark increases the likelihood that the lakes will hold fish populations. For the terrestrial ecosystem it can be expected that the plant functional groups will be virtually identical, but that the specific species might vary.

One uncertainty when comparing Greenland to the future Forsmark is the presence of eolian deposits and eolian activity. Eolian deposits from the last deglaciation occur in other areas of Sweden, and it is therefore not unlikely that they could be formed in a future Forsmark (at least if the glaciation is less intense and leaves Forsmark above sea level). The uppermost meter of the regolith in most of the Greenland study sites consists of eolian deposits, which are crucial for vegetation. Under the present conditions these eolian deposits are prone to wind erosion, and unvegetated wind-deflation scares are a common feature across the Kangerlussuaq region. Because of the lower relief in Forsmark it can be expected that the wind patterns would be different compared to Greenland, and it could lead to less eolian erosion. First, higher relief in Greenland results in valleys that can channel the winds and increase wind erosion. Second, the wind deflation scares in Greenland are mostly associated with ridgetops and other protruding landforms, and hence, Forsmark might experience less wind erosion and thus a more homogeneous vegetation.

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