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The Greenland Analogue Project (GAP)

Literature review of hydrogeology/ hydrogeochemistry

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1 Introduction

This report is produced as part of the Greenland Analogue Project (GAP), carried out as a collaboration project with the Canadian Nuclear Waste Management Organization (NWMO), Posiva Oy and the Swedish Nuclear Fuel and Waste Management Co (SKB) as collaborating and financing partners.

The overall aim of the project is to improve the current understanding of hydrogeological and hydrogeochemical processes associated with continental-scale glacial periods including with the presence of permafrost and the advance/retreat of ice sheets. The project will focus on studying how an ice sheet affects groundwater flow and water chemistry around a deep geological repository in crystalline bedrock. The Greenland Analogue Project consists of three active sub-projects (A–C) with individual objectives. Field studies are conducted in the Kangerlussuaq region, in central Western Greenland.

Sub-projects A and B collectively aim at improving the understanding of ice sheet hydrology by combining investigations on surface water processes with ice sheet drilling and instrumentation. In sub-project C, the penetration of glacial melt water into the bedrock, groundwater flow and the chemical composition of water will be studied. Main planned activities in sub-project C include drilling of a deep borehole in front of the ice sheet, in which different downhole surveys, sampling and monitoring will be carried out.

The primary aim of this report is to review available information about hydrogeology and hydrogeochemistry in central Western Greenland, with special emphasis on the area around Kangerlussuaq. The relevant information about this area is however very limited, and it was decided to extend the review to briefly include studies made in other regions with similar conditions in terms of geology, climate and glaciology. The number of published studies made in other areas with glaciers, ice sheets or permafrost is very large, and the review and list of references in this report is far from complete. It is also obvious that both hydrogeological and hydrogeochemical conditions are highly local and site-specific. Hence, data from other arctic regions can only be seen as indicative and give knowledge on processes that may or may not be relevant for the geographical area of primary interest in this project.

2 Hydrogeology and hydrogeochemistry in glaciated and permafrost regions – background section

The hydrogeological and hydrogeochemical conditions in a region such as western Greenland are likely to be governed by several factors:

- Bedrock and tectonics.
- Soil cover.
- Current temperature and precipitation.
- Permafrost.
- The ice sheet.
- Water bodies and surface drainage.
- The glacial history.
- Past climate.
- The Sea.

A conceptual description of the conditions and relevant processes at the edge of an ice sheet is shown in Figure 2-1. The margin of the Greenland ice sheet can be considered as more or less stationary.

2.1 Permafrost

Permafrost is defined as ground remaining below 0°C for at least two consecutive years /French 1996/. How deep the permafrost grows depends on the heat exchange process across the atmosphere/ground boundary layers and on the geothermal heat flow from the Earth's interior. The heat exchange depends on climate conditions and their long-term, annual and diurnal variation, as well as on surface conditions such as snow cover, vegetation, topography, soil characteristics and presence of water bodies /French 1996, SKB 2006a/. More than 20% of the world's land area is occupied by permafrost /French 1996/.

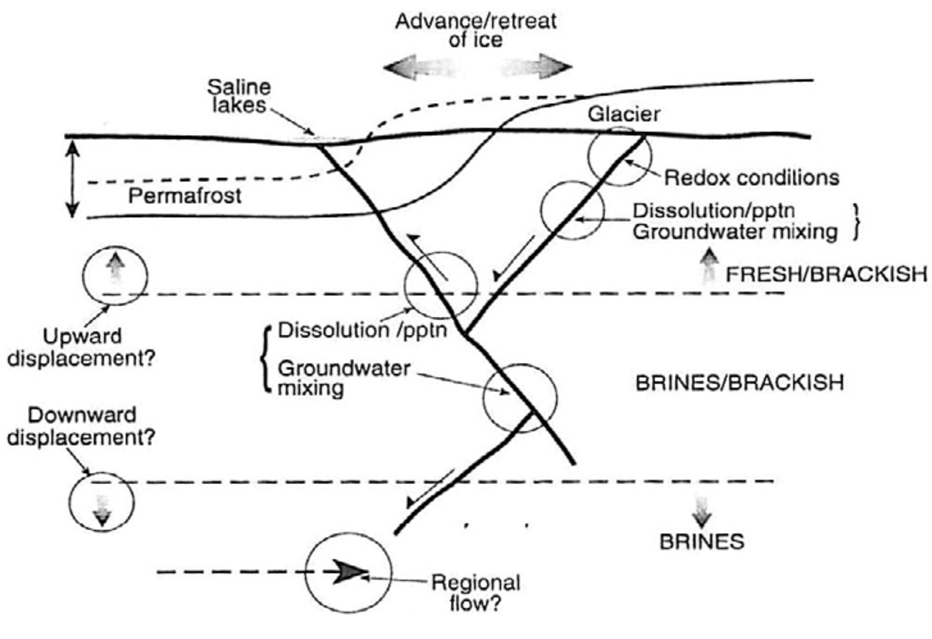

Figure 2-1. Conceptual idea on the hydrogeochemical conditions around the edge of an ice sheet /Smellie and Frape 1997/.

In general, permafrost regions are considered to be either continuous or discontinuous. In continuous areas frozen ground is present everywhere, except for localised thawed zones, or taliks, beneath lakes and rivers /French 1996/. In discontinuous permafrost regions areas of unfrozen ground separate the permafrost. An illustration of different classes of permafrost is shown in Figure 2-2.

Reviews of surface and sub-surface conditions and processes in permafrost areas, especially periglacial permafrost, have been presented by /e.g. French 1996, Ahonen 2001/ and /Vidstrand 2003/.

Groundwater in permafrost environments can occur above (suprapermafrost), within (intrapermafrost) and below (subpermafrost) the permafrost layer (see Figure 2-3). The permafrost will in most cases act as a more or less impermeable, confining layer, which inhibits and re-directs groundwater flow and recharge.

The active layer is commonly defined as an upper layer in which the liquid water freezes to ice and then thaws again on an annual basis /van Everdingen 1976/. The thickness of the active layer, between centimetre and metre scale, usually varies from year to year. In the active layer the frost table will have an irregular surface, due to local variations in thermal properties and microclimate, and the lateral groundwater flow may not necessarily correspond to the ground surface topography.

Areas of unfrozen ground within the permafrost, often found beneath a body of water, are called taliks. An open talik can act as a conduit between sub- and suprapermafrost aquifers, and can hence give rise to recharge to or discharge from a subpermafrost aquifer.

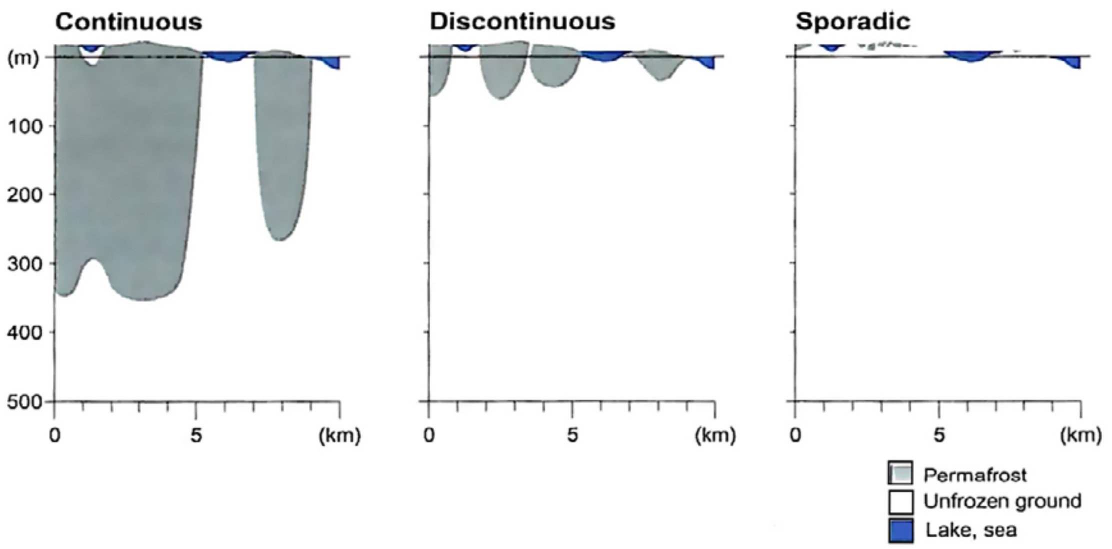

__Figure 2-2.__ Illustration of different classes of permafrost distribution /Vidstrand 2003/ after /French 1996/.

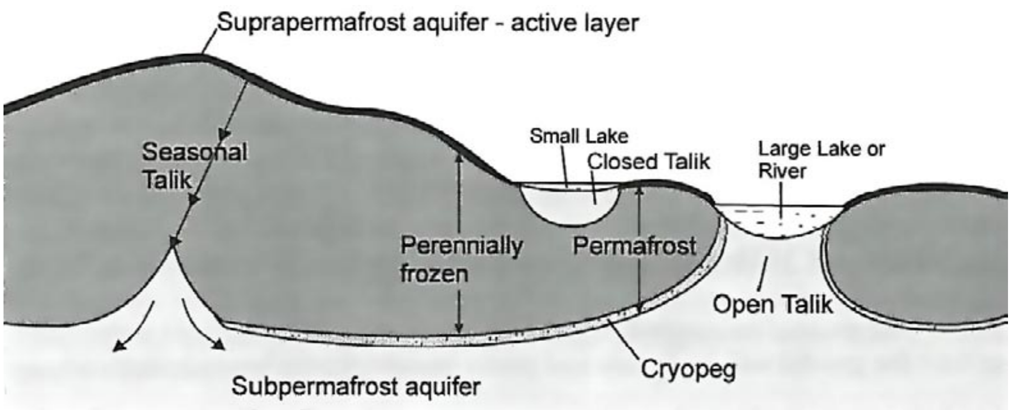

__Figure 2-3.__ Surface and sub-surface conditions and processes in permafrost areas /Vidstrand 2003/ after /van Everdingen 1990/.

Where freezing of the active layer restricts perennial discharge from intra-permafrost or sub-permafrost aquifers, different types of seasonal and perennial frost mounds can develop at the locations of groundwater discharge /French 1996/. A pingo is an example of such periglacial landforms. Usually two types of pingo are distinguished by definition; a hydrostatic pingo (closed system) and a hydraulic pingo (open system). The latter type involves generation and flow of groundwater under artesian pressure, providing a perennial sub-surface supply to the ice-cores of the pingo. Pingos have been reported from both Eastern and Western Greenland (Figure 2-4). /Worsley and Gurney 2007/ hypothesised that the concentration of pingos in eastern Greenland reflects the locally high heat flows following Tertiary volcanism.

In a fractured crystalline basement freezing will affect the groundwater flow conditions, by a reduction of hydraulic conductivity. The conductivity can be estimated to decrease exponentially with a reduction in the unfrozen fraction of water /SKB 2006a/. In general, in a permafrost environment, the only recharge and discharge zones are open taliks in structurally deformed zones in the bedrock. In many Arctic environments continuous permafrost without open taliks prevents subpermafrost recharge and the only potential recharge is subglacial. Lateral groundwater flow through the basement rocks can take place under the permafrost or in larger fracture zones.

The subpermafrost fluid chemistry may, depending on the local conditions of recharge and discharge, be influenced by suprapermafrost, intrapermafrost or deep groundwater flow.

These processes within the subsurface environment result in changes of hydrogeochemical as well as in hydrogeological properties. One possible change is due to the salt rejection due to the freezing of groundwater in the subsurface permafrost regions. Saline water stays as an unstable basal cryopeg just beneath the permafrost and only slowly mixes with the deeper groundwater /Vidstrand 2003/. However, the occurrence of fractures and fracture zones may trigger "fingering" and initiate a density-driven groundwater flow.

Figure 2-4. *Geographical distribution of pingos in Greenland /Bennike 1998/ Copyright: Geological Survey of Denmark and Greenland (GEUS).*

If the permafrost area is located in the downstream direction from a glacier or ice sheet, the fluid chemistry, especially within the active layer, is likely to be highly affected by the melt water from the ice.

Natural gas hydrates may form under conditions of low temperature and high pressure, both above and below the freezing point of water /SKB 2006b/. In general methane hydrate is unstable in the uppermost ca 200 m in areas where the permafrost depth is less than 200 m. The presence of saline groundwaters increases hydrate formation from water in the absence of a gas phase. Methane hydrate reservoirs have primarily been found in sedimentary rocks /e.g Dallimore and Collett 1995/, but /Stotler 2008/ concluded, based on the field studies in the crystalline bedrock at the Lupin Mine in Canada, that gas hydrates are likely to be present in undisturbed conditions to depths 600 m below the base of the permafrost.. However, there was no evidence of large-scale methane releases at the Lupin site.

The growth of gas hydrates in the ground excludes salts /Gascoyne 2000, Vidstrand 2003/ and also results in an isotopic enrichment in the solid phase. The gas hydrate content and stability are dependent on the gas chemistry, the pore pressures and the salinity of the groundwater.

2.2 The ice sheet

The Greenland Ice Sheet provides an enormous water resource. Sub-glacial meltwater induced by geothermal heat is generated where the base of the ice sheet is at the melting temperature. Close to the edge of the ice sheet surface melt water can reach the base of the ice sheet through englacial conduits. Sub-glacial meltwater that infiltrates into the sub-surface will flow from areas with high hydraulic head to areas with lower head, either beneath the ice sheet or to locations beyond the margins of the ice. Meltwater reaching the sole of the ice sheet but which does not infiltrate the sub-surface will flow along the interface between the ice and the ground. There is a considerable difficulty in ice sheet hydrological modelling how to estimate the amount of meltwater that is recharged to the underlying aquifers. The modelling approaches used are based on either prescribed hydraulic heads related to the ice sheet elevation or specified meltwater rates.

A detailed review of the hydrology of glaciers and ice sheets has been presented by /Jansson et al. 2007/. Comprehensive numerical studies of sub-glacial groundwater flow during past continental scale glaciations and using different concepts and approaches have been presented by /e.g. Provost et al. 1998, Boulton et al. 2001, Breemer et al. 2002, Flowers and Clarke 2002a, b, Jaquet and Siegel 2003, Piotrowski 2006, Moeller et al. 2007, Person et al. 2007, Lemieux et al. 2008a, b, c, Bense and Person 2008/.

Further discussion of processes involved and different modelling approaches is beyond the scope of this review report. It is however obvious that, although considerable advances have been achieved during the last couple of decades, we need an increased knowledge about the spatial and temporal variations in drainage of ice sheets, including couplings between the supra-, en- and subglacial systems.

Glacial meltwater recharging the subglacial basement can be expected to have certain hydrogeochemical characteristics and rocks affected by glacial meltwater may show specific mineralogical and chemical signatures /Piotrowski 2006/:

1. Low ^{18}O and ^{2}H contents.
2. High dissolved oxygen content.
3. Lack of pyrite along the conductive fractures.
4. High values of $\delta^{13}C$ in calcites.
5. Low dissolved organic carbon concentration.
6. Low total dissolved solids contents.
7. Uranium-series disequilibrium.

3 Climate and meteorology in western Greenland

3.1 General conditions

The climate in Greenland as a whole is arctic but in terms of weather it may be characterised by the rather large contrasts through the country. There are considerable differences between north and south and between coastal and inland areas towards the ice sheet margin. During winter the temperature on the northern part of the ice sheet can be lower than –70°C, while summer temperature in the coastal southern areas can exceed 25°C.

The region around Kangerlussuaq, in central Western Greenland, is dominated by the ice sheet, but along its margin there is an ice-free zone. This zone consists of an arctic mountain landscape which is almost free of vegetation and with rocky surfaces.

The border between continuous and discontinuous permafrost is located between Kangerlussuaq and the coast (Figure 3-1).

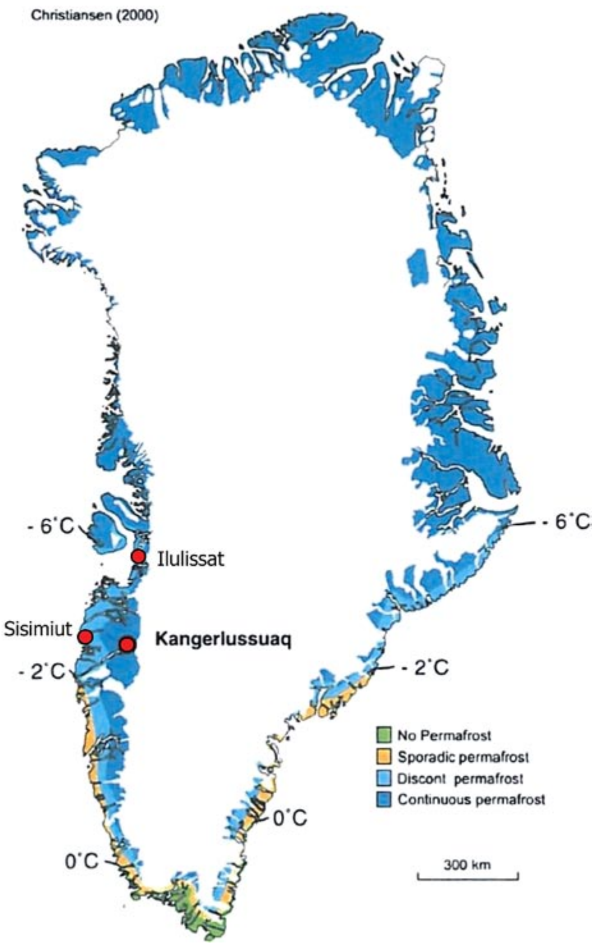

Figure 3-1. *Permafrost distribution in Greenland (/Christiansen and Humlum 2000/ in /Jørgensen and Andreasen 2007/). Figure reprinted with permission from Elsevier. Locations of Sisimiut and Ilulissat have been added.*

3.2 The Kangerlussuaq region

There is a strong climatic gradient between the coast and the western margin of the Greenland ice sheet. At Sisimiut, situated by the coast, the average annual temperature is –3.9°C while at Kangerlussuaq where there is more continental climate, the average annual temperature is –5.7°C. While winters at Sisimiut are characterised by a relatively thick snow cover, there is generally little snow at Kangerlussuaq. This is also reflected in the glaciation limit, which increases as you go inland.

The area at the head of the fjord is characterised by mean annual temperatures ca –6°C, continuous permafrost and low precipitation (< 150 mm/y) /Anderson et al. 2001, Weidick and Bennike 2007/. The zone within about 80 km from the ice sheet has a very low (negative) effective precipitation. The climate at the coast is more maritime with a smaller temperature range and higher precipitation (c 300 mm/y).

The local climatic conditions at Kangerlussuaq are determined by its northern location in a 2–3 km wide valley surrounded by mountains with an altitude of 400–600 m /Jørgensen and Andreasen 2007/. The ice sheet with altitudes up to 3 km has a dominant influence on precipitation and winds.

4 Geological settings

The target area for the Greenland Analogue Project belongs to the Nagssugtoqidian Orogen of West Greenland. The bedrock between the coast and the ice sheet is dominated by banded gneisses with amphibolites and pegmatitic dykes. The gneisses, which are commonly fractured and weathered, have been extensively metamorphosed. During the Quaternary the area was repeatedly invaded by ice streams, which scoured out deep valleys. A number of significant systems of moraines were formed in the area during deglaciation at the end of the last glacial period. The interplay between the isostatic depression caused by the weight of the ice sheet and global variations in sea level has resulted in variations in the relative sea level during deglaciation. Parts of the area located above sea level today were periodically below sea level. This means that fine grained marine deposits could accumulate in valleys and depressions in the area. These deposits may be overlain by eolian sediments and fresh water sediments of fluvial or lacustrine origin.

4.1 Bedrock geology

The Nagssugtoqidian Orogen of West Greenland represents a Palaeoproterozoic tectonic belt with prominent ENE-trending structural grain. It is situated between the North Atlantic Craton to the south and less known Archaean craton that includes Rinkian Orogen to the north (See Figure 4-1). On a larger scale, the Nagssugtoqidian Orogen joins an extensive collage of Palaeoproterozoic orogenic belts that stretch from Canada through Greenland and Scotland to the Baltic shield /e.g. Hoffman 1990, Bridgwater et al. 1990, Park 1994, Wardle et al. 2000/.

The Nagssugtoqidian Orogen consists of three major segments namely the southern, central, and northern (SNO, CNO, and NNO; see Figure 4-1) as defined by /Marker et al. 1995/. Predominant rock type across the orogen are Archaean orthogneisses with minor paragneisses and few sets of voluminous Palaeoproterozoic Kangâmiut dyke swarms in the southern part of the orogen /Escher et al. 1975, Bridgwater et al. 1995, Cadman et al. 2001/.

The Archaean foreland to the south of the Nagssugtoqidian Orogen comprises granulite-facies orthogneisses and paragneisses with a northwest-trending Archaean structure. Main intrusive rocks are generally undeformed, late Archaean granites as well as Proterozoic mafic Kangâmiut dyke swarm. The transition zone, the so-called Southern Nagssugtoqidian front, defines the southern boundary of the Nagssugtoqidian Orogen. The tectonic structure of the front and lithological continuity across indicates that SNO is a parautochthonous foreland belt /van Gool et al. 2002/.

Within the SNO the dominant rock type is mainly tonalitic to granodioritic gneisses with minor occurrences of paragneisses in the south-western part, which probably represents a volcano-sedimentary sequence /Cadman et al. 2001, Connelly and Mengel 2000/. The northern boundary of the SNO is defined by the south-directed ductile Ikertôq thrust /Grocott 1979/, which is one of the most significant boundary zones in the Nagssugtoqidian Orogen /van Gool et al. 1996/. It marks the lithological and metamorphic boundary between SNO and CNO as well as northern extend of the Kangâmiut dyke swarm. Within the Ikertôq thrust, considerable amounts of metasedimentary psammitic rock panels parallel to the zone are present /van Gool et al. 1996/.

The Central Nagssugtoqidian Orogen is dominated by charnockite complex, high-grade Archaean granitoids and orthogneisses interleaved with Archaean and Palaeoproterozoic paragneisses /Kalsbeek and Nutman 1996, Connelly and Mengel 2000/. Dominant intrusives are Palaeoproterozoic quartz-dioritic to tonalitic rocks.

The northern Nagssugtoqidian Orogen consists mainly of poly-deformed, variably reworked grey Archaean TTG gneisses and supracrustal belts /Moyen et al. 2003, Steenfelt et al. 2005/.

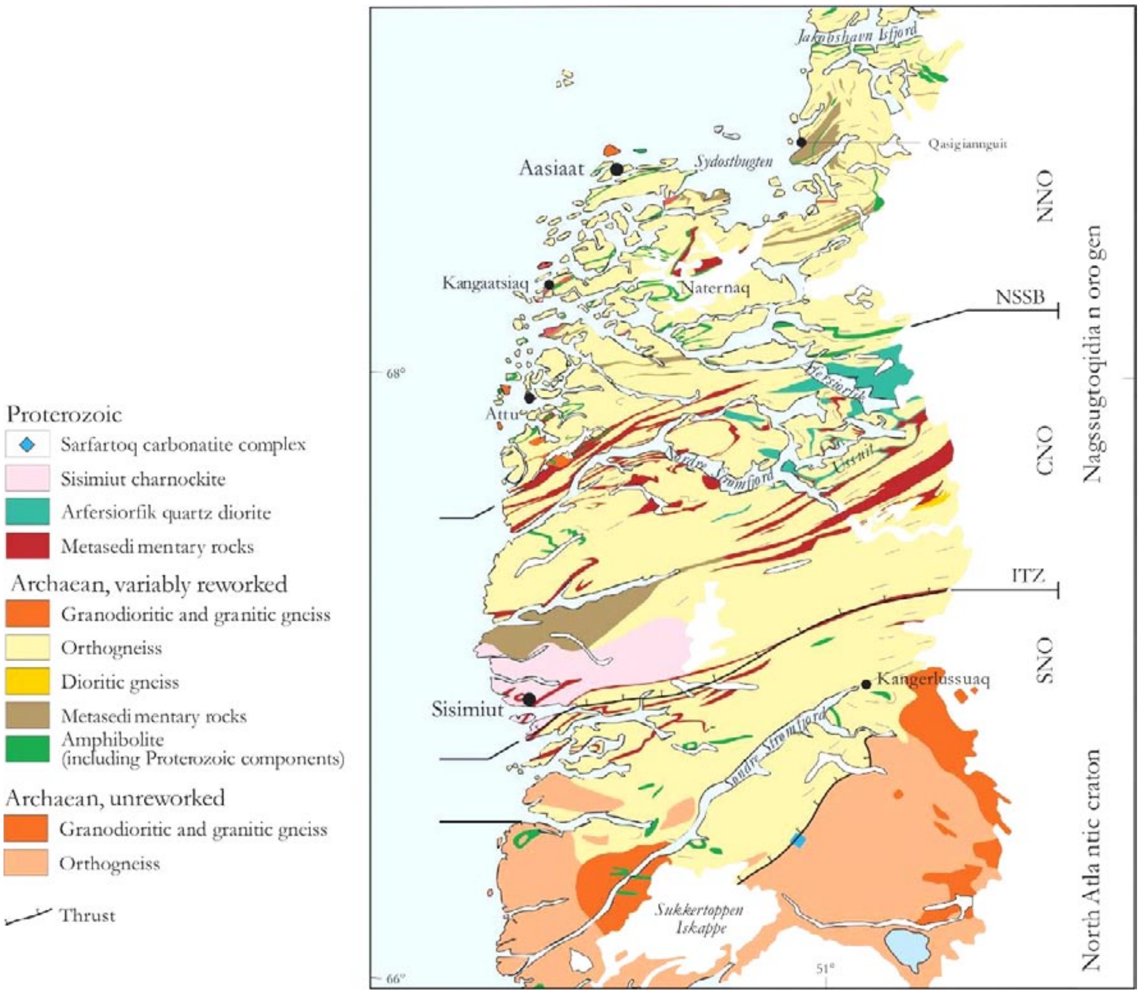

Figure 4-1. *Geological map of the Nagssugtoqidian Orogen /van Gool et al. 2002/. Figure reprinted with permission from NRC Research Press.*

Tectonic subdivision of the orogen reflects the metamorphic zonation as well as timing of metamorphism across the orogen. The SNO and NNO are predominantly in amphibolite facies and the CNO is mostly in granulite facies. Peak metamorphic conditions in the CNO occurred at ca 1,860–1,840 Ma /Taylor and Kalsbeek 1990, Kalsbeek et al. 1987, Kalsbeek and Nutman 1996, Connelly et al. 2000, Willigers et al. 2001/ whereas peak metamorphism in adjacent segments to the south and north took place at least 60 Ma later /Kalsbeek and Nutman 1996, Connelly and Mengel 2000, Willigers et al. 2001/. Such distribution of metamorphic phases and timing of metamorphism points towards a collisional deformation starting in the CNO and subsequently spreading laterally to the foreland in the southern and hinterland to the north /van Gool et al. 2002/.

4.2 Local tectonic conditions

The general area of interest of the Greenland Analogue Project is the Søndre Strømfjord (Kangerlussuaq) region that is situated in the most eastern, exposed part of the SNO. The western tip of the Russells Glacier is situated between two regional thrust zones (Figure 4-2). The ENE–WSW trending valley occupied by Russells Glacier follows a major fault line. Just a few kilometres to the south of the area of interest runs the NE-SW orientated southern Nagssugtoqidian front which is a tectonic and metamorphic boundary between SNO and Southern Nagssugtoqidian foreland (see description above). A few kilometres to the north of the target area is another uncertain, but strongly indicated by the total intensity magnetic anomaly map, E-W-trending thrust zone /Korstgård et al. 2006/. In addition to those two important tectonic features, less than 20 kilometres to the north lays a northern boundary of the SNO, defined by the ENE-trending Ikertôq thrust zone, which is one of the most significant boundary zones in the Nagssugtoqidian Orogen /van Gool et al. 1996/.

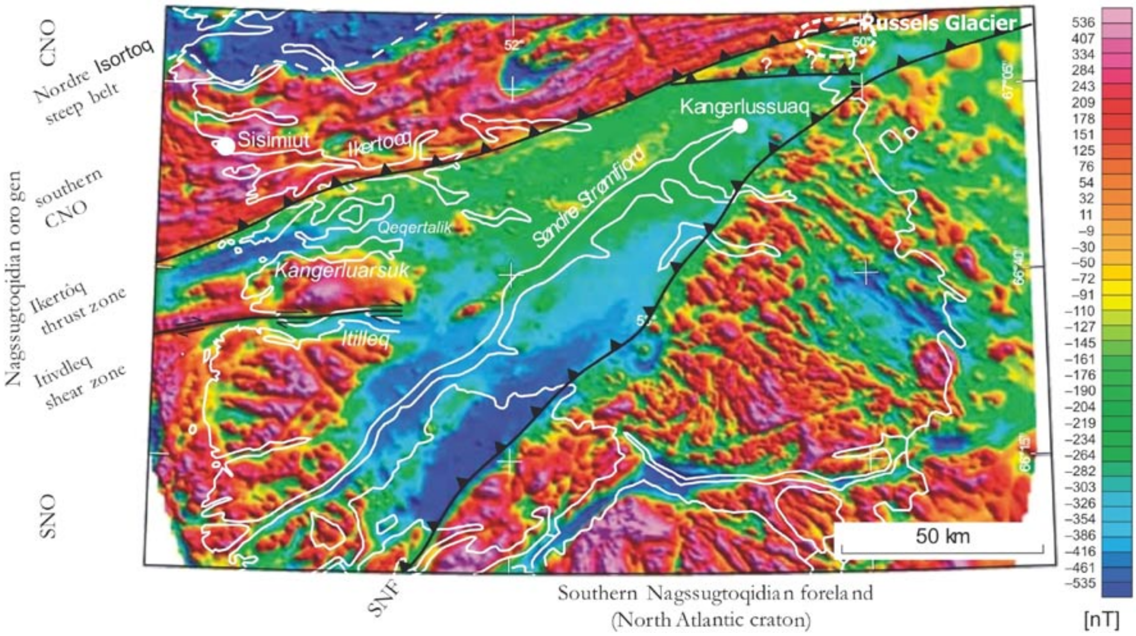

Figure 4-2. Total intensity magnetic field anomaly map of the south-eastern part of the Nagssugtoqidian orogen /Korstgard et al. 2006/. Copyright: Geological Survey of Denmark and Greenland (GEUS). The approximate location of Russels glacier has been added to the original map.

Since the Mesozoic, West Greenland is a part of a passive margin east of the Davis Strait, where N–S faults connect extinct NW–SE oriented Palaeogene sea-floor spreading axes in the Labrador Sea and Baffin Bay /Chalmers and Pulvertaft 2001/. Present day landscape of central West Greenland close to Kangerlussuaq can be regarded as alpine with maximum altitude of 1,848 m.a.s.l. /Sugden 1974/ but some flat summits are still present. Contrary to the common opinion that elevated, passive continental margins have remained uplifted since the time of rifting, recent studies /e.g. Japsen et al. 2006, Bonow et al. 2006/ demonstrated that the present-day mountains of West Greenland are the end result of three Cenozoic phases of uplift and erosion. Based on apatite fission-track analysis (AFTA) three stages of uplift were dated to 36–30 Ma, 11–10 Ma and 7–2 Ma and are summarised in Figure 4-3.

The uplift phase that began between 36 and 30 Ma led to the formation of a planation surface /Japsen et al. 2006/. A second last uplift associated with the C3 cooling event reached of maximum 500 m followed by the latest uplift (C4 cooling event) of maximum 1,000 m /Bonow et al. 2006/. The latest two stages of uplift caused offset by reactivated faults (Figure 4-4), resulting in megablocks that were tilted and uplifted to present-day altitudes /Japsen et al. 2006/.

Substantial changes in the thickness of the crust and lithosphere driven by the mechanism situated in the deep crust or the upper mantle was proposed /Rohrman and van der Beek 1996, King and Anderson 1998, Praeg et al. 2005, McKenzie et al. 2000/ to explain the regional nature of the movement along passive margin with the considerable distance from the active plate boundary /Japsen et al. 2006/.

4.3 Quaternary geology

The area bordering Søndre Strømfjord comprises the widest stretch of ice-free land in Greenland, extending 170 km from the coast to the inland ice margin. The Kangerlussuaq area of West Greenland is characterized by a gentle WSW–ENE trending hilly landscape with generally narrow fjords. During the late Weichselian maximum, c 18,000 yr BP, the ice advanced onto the shelf area /Kelly 1985, Funder 1989, Funder and Hansen 1996, Bennike and Björck 2002/. Deglaciation in central West Greenland coastal areas started around 13,000 cal yr BP /van Tatenhove et al. 1996/. About 10,000 yr BP the ice reached the present coast line /Ten Brink and Weidick 1974/ and most of the inland ice margin reached its present position about 6000 yr BP. /Funder 1989, van Tatenhove et al. 1996/. Since the fjords are very narrow the ice front probably receded by ablation rather than calving /Bennike and Björck 2002/.

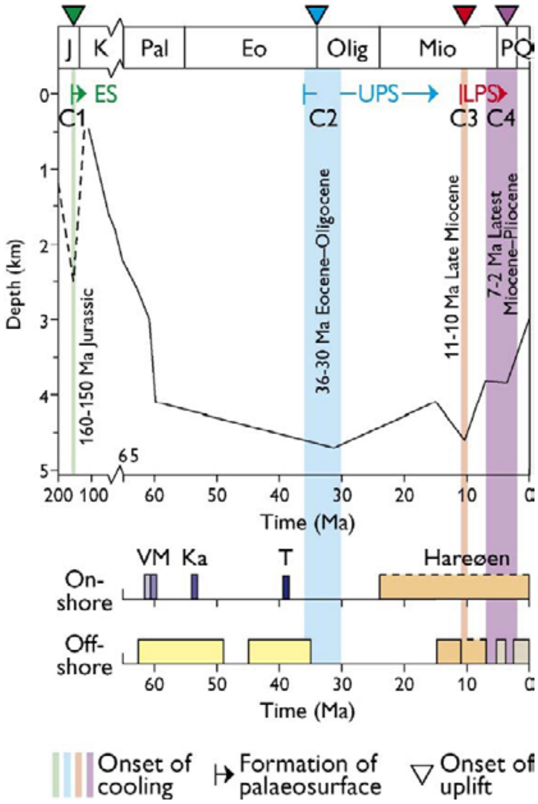

Figure 4-3. *Schematic event chronology. Relation between discrete episodes during which AFTA data indicate onset of cooling (C1 to C4), formation of palaeosurfaces and the most likely onset of uplift. Minimum burial is not constrained by AFTA, but supplementary information is derived from geomorphology that describes bedrock surfaces formed during periods of denudation /Bonow 2005, Bonow et al. 2006/; ES: etch surface; UPS: upper planation surface; LPS: lower planation surface. The burial curve (full line) is for a hypothetical interface between basement and Cretaceous sediments now at 3-km depth; arbitrary depth scale before 100 Ma (dashed line). J: Jurassic; K: Cretaceous; Pal: Paleocene; Eo: Eocene; Olig: Oligocene; Mio: Miocene; P: Pliocene; Q: Quaternary. Basalt–V: Vaigat Formation (c. 61 My old /Storey et al. 1998/), M: Maligât Formation (c. 60 My old /Storey et al. 1998/) and Ka: Kanisut Member (c. 53 My old /Storey et al. 1998/), T: Talerua Member (c. 39 My old /Schmidt et al. 2005/). Hareøen: Neogene–Quaternary deposits on Hareøen /Christiansen et al. 1999/. Dotted lines: maximum age range of sediments. (Figure from /Japsen et al. 2006/, reprinted with permission from Elsevier).*

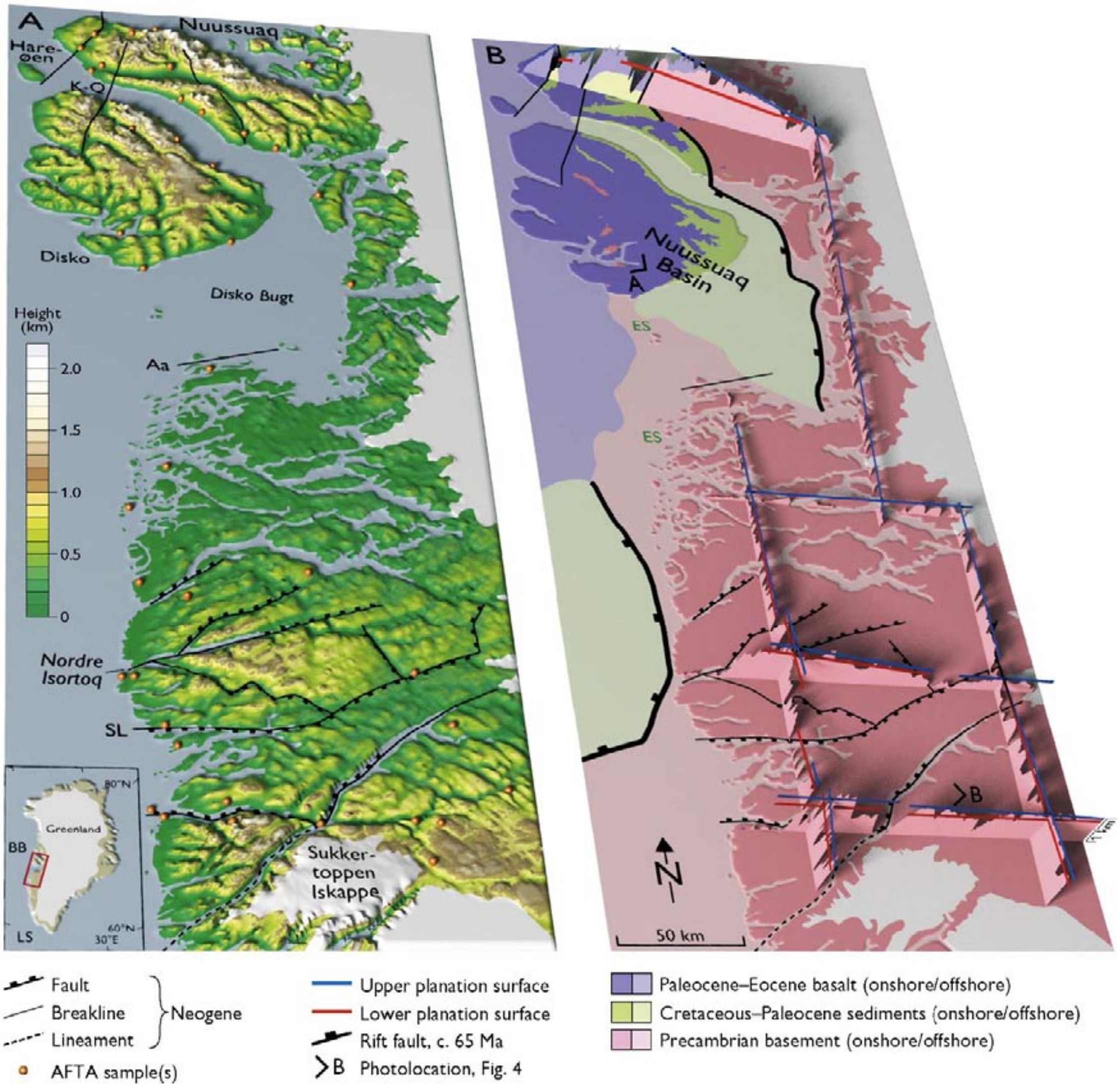

Figure 4-4. *Three-dimensional maps of the central West Greenland. (A) Topography. (B) Geology with topographical profiles. An elevated plateau is defined by two planation surfaces (shown as red and blue lines) that cut across Precambrian basement and Palaeogene volcanic rocks. The formation of the planation surfaces was uniform across the study area and they must be younger than mid-Eocene basalts. The planation surfaces now dip in different directions and are offset by faults that displace them. Three significant faults and breaklines (changes in slope gradient) relative to the planation surfaces are (1) the north–south Kuugannguaq–Qunnilik (K–Q) fault on Disko and Nuussuaq, (2) an east–west fault just north of Aasiaat (Aa) where orthogneisses are separated from supracrustal rocks to the north; this fault separates the south-dipping planation surface on Disko from the north-dipping surface south of Disko Bugt, and (3) the east–west Sisimiut Line (SL) that coincides with the Precambrian Ikertôq thrust zone. An etch surface (ES) has been re-exposed from below Cretaceous–Palaeocene cover rocks on Nuussuaq, Disko and south of Disko Bugt. LS: Labrador Sea, BB: Baffin Bay. (Figure from /Japsen et al. 2006/, reprinted with permission from Elsevier).*

Palaeoclimate research in Greenland has been dominated in recent years by ice-core studies from the central ice sheet, which initially focused on interglacial/glacial-scale climatic change /Dansgaard and Oeschger 1989, Dansgaard et al. 1993, GRIP members 1993, Dahl-Jensen et al. 1998/.

The chronology of the deglaciation is based on dating of marine shells and glacial geological studies /Ten Brink 1975, van Tatenhove et al. 1996, Humlum 1999/ and on lichenometry and luminescence dating /Forman 2008/ while the Holocene climatic evolution in the ice-free region of West Greenland largely is based on palaeoecological studies /Fredskild 1985, Eisner et al. 1995, Bennike 2000, Ryves et al. 2002, McGowan et al. 2003/, lacustrine sediments /Willemse and Törnqvist 1999/ and eolian deposits studies /Willemse et al. 2003/.

The oldest radiocarbon date on marine shells from the head of Søndre Strømfjord is 8,000 cal years BP, which provides a minimum estimate for the timing of the deglaciation of the inland /Bennike 2000, Bennike and Björck 2002/. Between the coast and the present day ice margin a succession of up to seven moraine systems has been identified /Ten Brink 1975/.

About 7000 yr BP the ice front had reached Kangerlussuaq and by 6000 yr BP it had reached the present day position /Kelly 1985, van Tatenhove et al. 1996/. It is unknown how far to the east the ice front retreated, but by comparing deglaciation dates from the moraine system before 6000 yr BP /van Tatenhove et al. 1996/ infer that the ice reached 10's of kilometres beyond the present day ice front. The recession at the beginning of the Holocene was probably a response to warmer temperatures in southwest Greenland between 8000 to 5000 yr BP /e.g. Kelly 1985, Funder and Fredskild 1989, van Tatenhove et al. 1996, Anderson et al. 1999, Bennike and Björck 2002/. From ice core records /Dahl-Jensen et al. 1998/ suggest that the maximum Holocene warmth lasted until 4000 yr BP.

Neoglacial advances started at 4000 yr BP /e.g. Kelly 1985, Willemse and Törnqvist 1999, Anderson et al. 1999, Bennike 2000/ and ended AD 1,700–1,920, during the Little Ice Age, by moraines near the position of the present day ice front /Weidick 1968/. Both Neoglacial and Little Ice Age limits occur within 2 km of the ice margin /Forman 2008/. /van Tatenhove et al. 1996/ infer that the ice stayed behind the present ice sheet margin for 6000 yr BP.

There are c. 20,000 lakes in the Søndre Strømfjord area, including approximately 12 so-called 'saline' lakes. These lakes has been used for palaeolimnologic studies /McGowan et al. 2003/ to provide information about regional climate heterogeneity and the impact from atmospheric circulation. Lake sediments close to the Greenland ice-sheet margin consist mainly of eolian silts mixed with organic material /Eisner et al. 1995/. /Willemse and Törnqvist 1999/ conducted a study on numerous small shallow lakes. The records exhibit a short-lived cooling, within the Holocene thermal maximum, around 8200 cal yr BP, recognized as the most pronounced Holocene climatic cooling with possible global significance /Alley et al. 1997/.

Arid conditions and supply of fine-grained sediments on the glacier-fed braided floodplains have resulted in extensive eolian deposits in front of the ice sheet /Willemse et al. 2003/. Studies of /Dijkmans and Törnqvist 1991, Eisner et al. 1995/ indicated that eolian sand formation dates back at least 4600 cal yr BP. Continuous records in eolian sand dunes and sand sheets etc indicate atmospheric regional deposition of silt since 4750 cal yr BP and periods of high eolian activity prior to 3400 cal yr BP and after 550 cal yr BP /Williemse et al. 2003/. /Williemse et al. 2003/ interpret the first event as due to the rerouting of meltwater channels in connection with a significant recession of the ice sheet during the mid Holocene recession. They explain the second event with local floodplain regeneration in connection with the early onset of the Little Ice age.

Sedimentary structures of eolian deposits indicate a wind direction from the ice sheet /van Tatenhove et al. 1996/. Silt influx data thus probably reflects the ability of warm maritime air to reach in towards the proglacial area /Willemse et al. 2003/. Over the period with instrumental data since AD 1873, the overall atmospheric circulation rather than temperature seems to be governing precipitation /Humlum 1999/.

5 The permafrost environment

There is generally a very poor knowledge about the permafrost conditions in the local area of interest. No deep boreholes aimed at investigating the permafrost depth have been reported. Geophysical investigations conducted have primarily been focused on the depth and conditions of the active layer for engineering related purposes /e.g. Ingeman-Nielsen 2005, Jørgensen and Andreasen 2007/.

Kangerlussuaq is located in the zone of continuous permafrost (see Figure 3-1), but not so far from the area along the coast, where the permafrost is characterised as discontinuous. /van Tatenhove and Olesen 1994/ calculated the total depth of permafrost based on temperature measurements in shallow boreholes (9–15 m) and evaluated thermal properties. For the calculations the used the equation:

$$z_p = \frac{k}{G} T_{ZAA}$$

where z_p is the depth of permafrost (m), G is the geothermal heat flux (W/m^2), k is the thermal conductivity ($W\ m^{-1}\ K^{-1}$) and T_{ZAA} is the temperature at the depth not influenced by the yearly cycle.

Calculated thermal properties for the Kangerlussuaq boreholes are presented in Table 5-1.

At Kangerlussuaq the permafrost depth was estimated to be 127 ± 31 m and at Sisimiut 33 ± 9 m. /van Tatenhove and Olesen 1994/ calculated a permafrost depth of 215 ± 2 m for the area north of Ilulissat, where the air temperatures are lower than at Kangerlussuaq. It is considered likely that the depth of permafrost varies across the area also due to differences in vegetation, snow cover, soil types and thermal properties of the subsurface.

/Van Tatenhove and Olesen 1994/ also presented a number of shallow and deep temperature profiles in south-west to west Greenland (see Figure 5-1).

/Villumsen et al. 2007/, in a report focused on conditions for road construction in Greenland, wrote: *"From Kangerlussuaq and northwards there is continuous permafrost which reaches depths of ~½ km in the far north"*. No information about the basis for this estimate is given. /Villumsen et al. 2007/ also reported georadar investigations along the road from Kangerlussuaq to the Inland Ice. The purpose of these investigations was to study the thickness and variation of the active layer. The evaluated thicknesses reflect the variations in temperature in the area. Close to the Inland Ice the thickness was ~2.5 m but at Kangerlussuaq only ~0.5 m.

Taliks in the continuous permafrost area can be expected where there exist sufficiently large water bodies, such as lakes and rivers. There are numerous water bodies of different size in the area. Modelling reported in /SKB 2006a/ estimates that an open talik can exist beneath a lake with an equivalent radius greater than the thickness of surrounding permafrost. A lake radius greater than 0.6 times the thickness of surrounding permafrost is sufficient for a deep lake to maintain an open talik.

An important issue for modelling of groundwater flow at the margin of the ice sheet and targeting a deep borehole is the lateral extension of permafrost beneath the ice. No information about this has been found in this literature review.

Table 5-1. Calculated thermal diffusivity (κ) and thermal conductivity (k) at Kangerlussuaq based on maximum thermal amplitudes (data from /van Tatenhove and Olesen 1994/).

Depth (m)	m (m)	r^2	n	K ($m^2\ s^{-1}$)	k ($Wm^{-1}K^{-1}$)
0.2 < z < 1.5	–1.341 ± 0.06	0.994	5	$0.182\ 10^{-6} \pm 0.01\ 10^{-6}$	0.2–0.6[1,2]
2.0 < z < 15.0	–4.096 ± 0.103	0.993	13	$1.695\ 10^{-6} \pm 0.04\ 10^{-6}$	2.2–3.7[2]

[1] The upper layer is assumed to be unfrozen.
[2] The range in thermal conductivity is due to the uncertainty in porosity (35 ± 8%) and the uncertainty in water/ice content $\Phi_{i/w}$ (0–0.43).

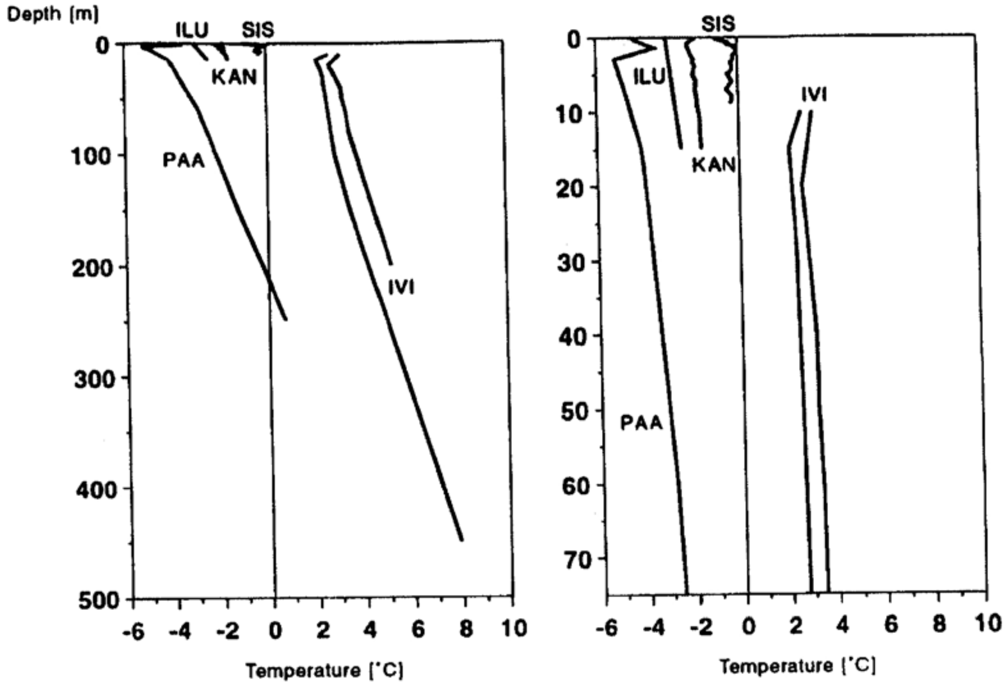

Figure 5-1. Temperature profiles measured in boreholes in south-west and west Greenland. KAN = Kangerlussuaq, SIS = Sisimiut, ILU = Ilulissat and IVI = Ivittuut (southern Greenland, 61.2°N, 48.2°W). (Figure from /van Tatenhove and Olesen 1994/, reprinted with permission from John Wiley and Sons).

6 Hydrology

6.1 Surface waters

Water is generated by either melting of the surface ice and snow or at the base of the glacier or ice sheet, where the basal ice temperature is at the pressure melting point. Meltwater generated on the ice surface flows through a drainage system in the glacier.

Rather few studies have been performed concerning the hydrology of the Greenland ice sheet. /Thomsen et al. 1989/ presented a study made for planning hydro-electric power generation in the Ilulissat area. In this study they described the surface hydrology system as consisting of a large number of equidistant drainage areas which are drained through moulins into the ice sheet (see Figure 6-1).

A few investigations have given information about the subglacial hydrological conditions. The deep drilling for the NorthGRIP ice core in the centre of the Greenland ice cap, about 950 km from Kangerlussuaq, /Dahl-Jensen et al. 2003/ found water at the base of the Greenland ice sheet. Indirect evidence has been reported based on e.g. seismic investigations and ground-penetrating radar surveys. /Zwally et al. 2002/ found large seasonal variations in ice flow velocities when analysing differential GPS data at the Swiss Camp, in west-central Greenland. They suggested that these variations were due to variations in water input from the surface down to the basal hydrological system.

/Lüthi et al. 2002/ carried out measurements of ice flow in Jacobshavn Isbræ at the Disko Bay. Boreholes were drilled to the bedrock beneath the ice. Ice deformation rates and temperature were measured in these boreholes. The variations in water pressure measured yielded information about the subglacial drainage system. The data were consistent with a sediment layer, a gap conduit or a linked cavity system with transmissivities beneath the glacier between 10^{-5} and 10^{-4} m^2/s.

In a project aimed at investigating the possibilities of exploiting the ice sheet's melt water, primarily for hydroelectric power generation, hydrological conditions in West Greenland were studied on the basis of topographic data /Weidick 1980, Weidick and Olesen 1980/. In this work a basin division and an estimate of the total water potential from the inland ice margin and from the coastal basins were made. Excerpts from the produced maps showing the area around Kangerlussuaq are shown in Figures 6-2 and 6-3. The precipitation data used for the individual basins were taken from map compilations published in the 1950s and 1960s supplemented with data from annual publications by the Danish Meteorological Institute.

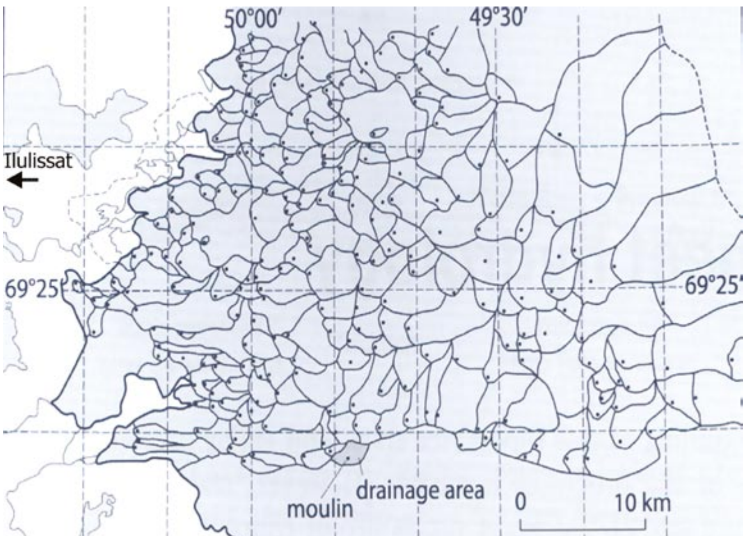

Figure 6-1. Delineation of surface drainage basins on the Greenland ice sheet near Ilulissat (modified after /Jansson et al. 2007/. The location of Ilulissat is indicated in the figure.

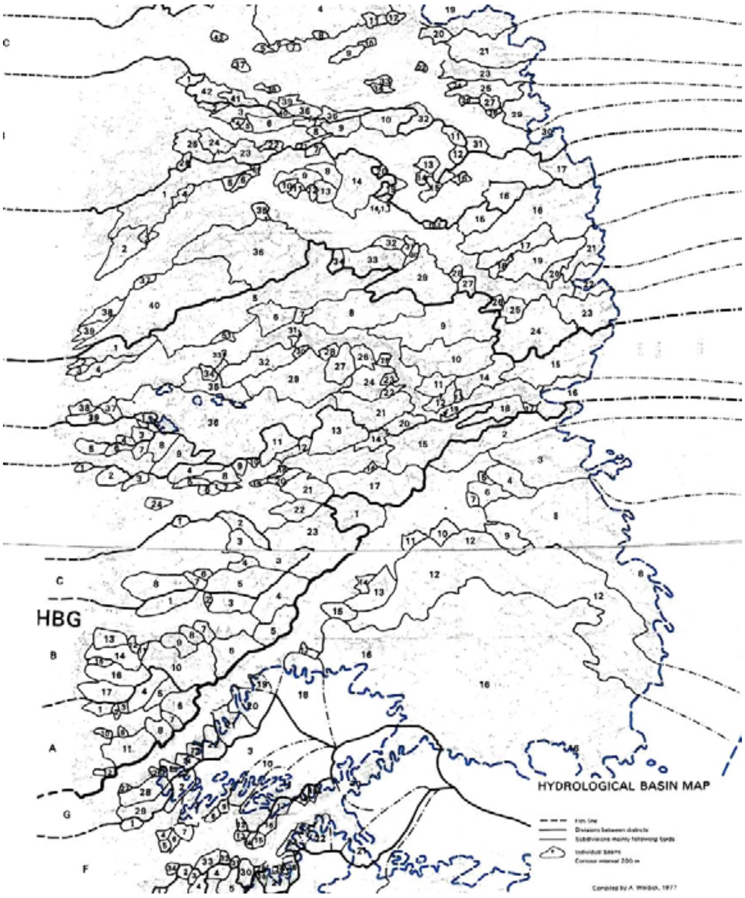

Figure 6-2. Cut-out from the Hydrological Basin Map, West Greenland /Weidick and Olesen 1980/. Copyright: Geological Survey of Denmark and Greenland (GEUS). Topographic base (A649/72) Geodætisk Institut Danmark.

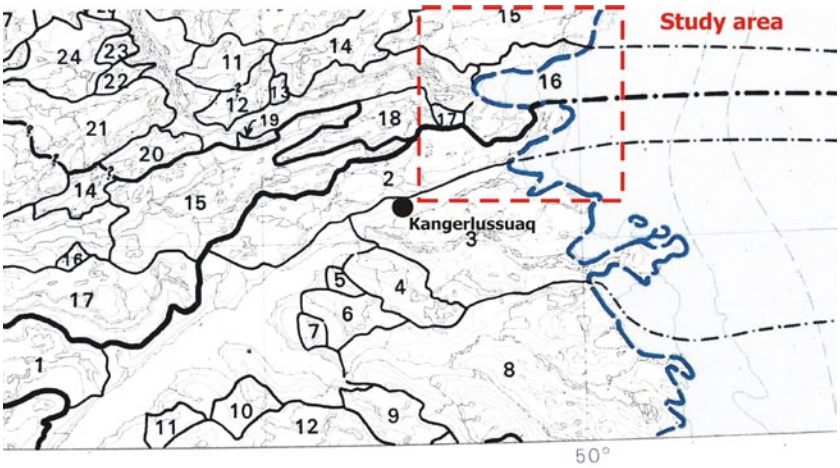

Figure 6-3. Detail from figure 6-2 – part around the Kangerlussuaq region (modified from /Weidick and Olesen 1980/. The location of the study area for the GAP project is illustrated in the figure. Bold black lines represent divisions between districts. The numbered areas represent individual basins.

The relation between water from precipitation and meltwater generated from glaciers including the margin of the ice sheet, was determined. The total amount of water recharge to the basins of West Greenland was estimated to be 217 km^3/year water equivalent. About 12% of this was estimated to originate from general precipitation over the coastal areas.The water potential of individual basins are illustrated in Figure 6-4.

/Hardy et al. 2000/ calculated balance fluxes for the Greenland ice sheet from a high-resolution digital elevation model (DEM) and mass-balance data. They delineated ten major drainage basins, considered as independent flow units with individual mass balances. In Figure 6-5 the general pattern of flow over the ice sheet is shown together with the ten major drainage basins calculated. The total accumulation, ablation and flux within each basin were presented.

/Van de Wal and Russell 1994/ presented a comparison of measured ablation and calculated ablation as well as comparison between measured and modelled discharge. The results were based on an energy balance calculation near Kangerlussuaq. They concluded that detailed modelling of meltwater runoff from the Greenland ice sheet has large uncertainties, due to uncertainties in e.g. catchment areas, englacial storage and drainage.

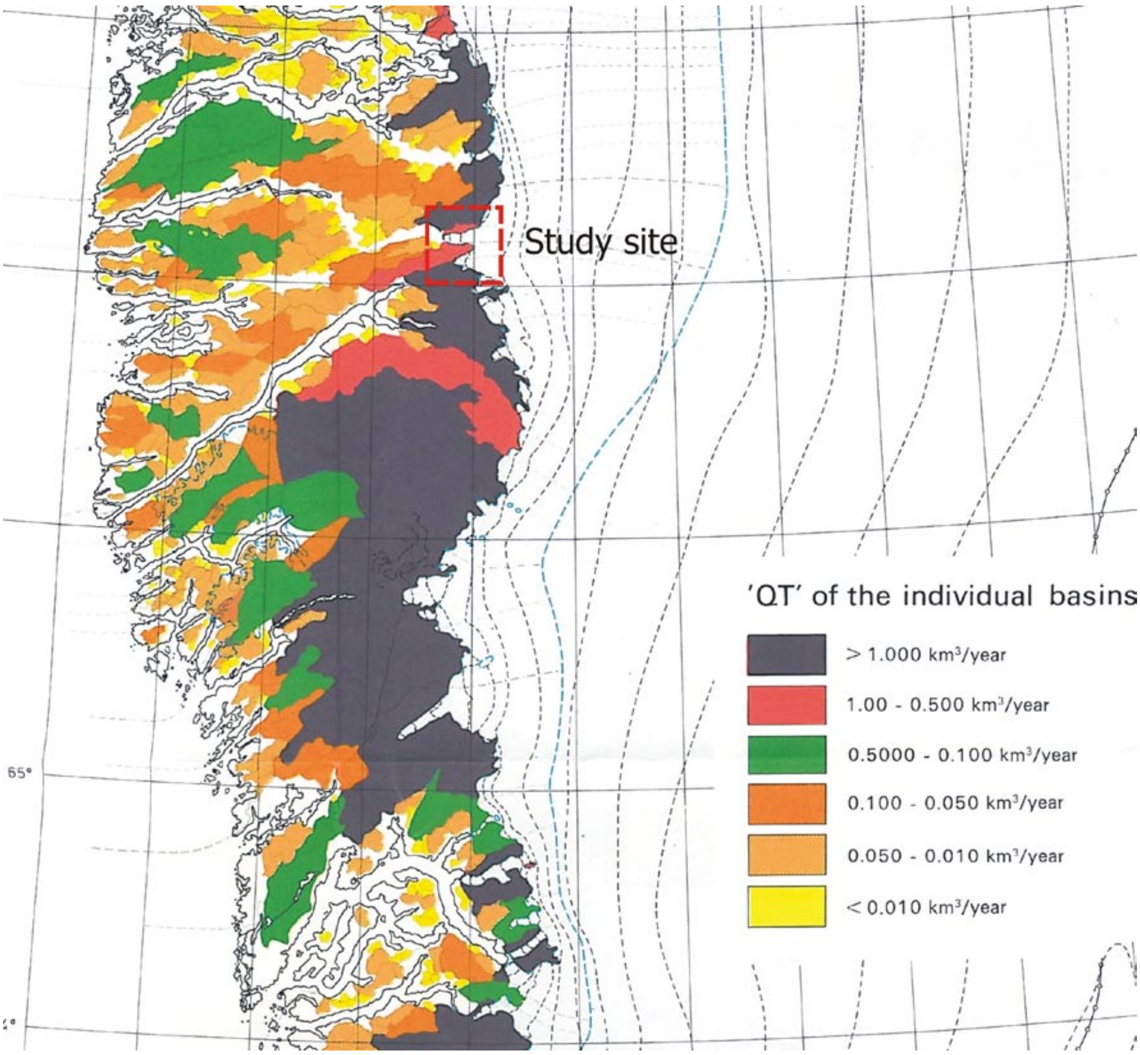

Figure 6-4. *Drainage patterns, basin areas and estimated water potential of basins in West Greenland (cut-out from map modified from /Weidick and Olesen 1980/. The location of the GAP Project study site is indicated. Dashed lined on the ice sheet represent elevation isolines. Copyright: Geological Survey of Denmark and Greenland (GEUS).*

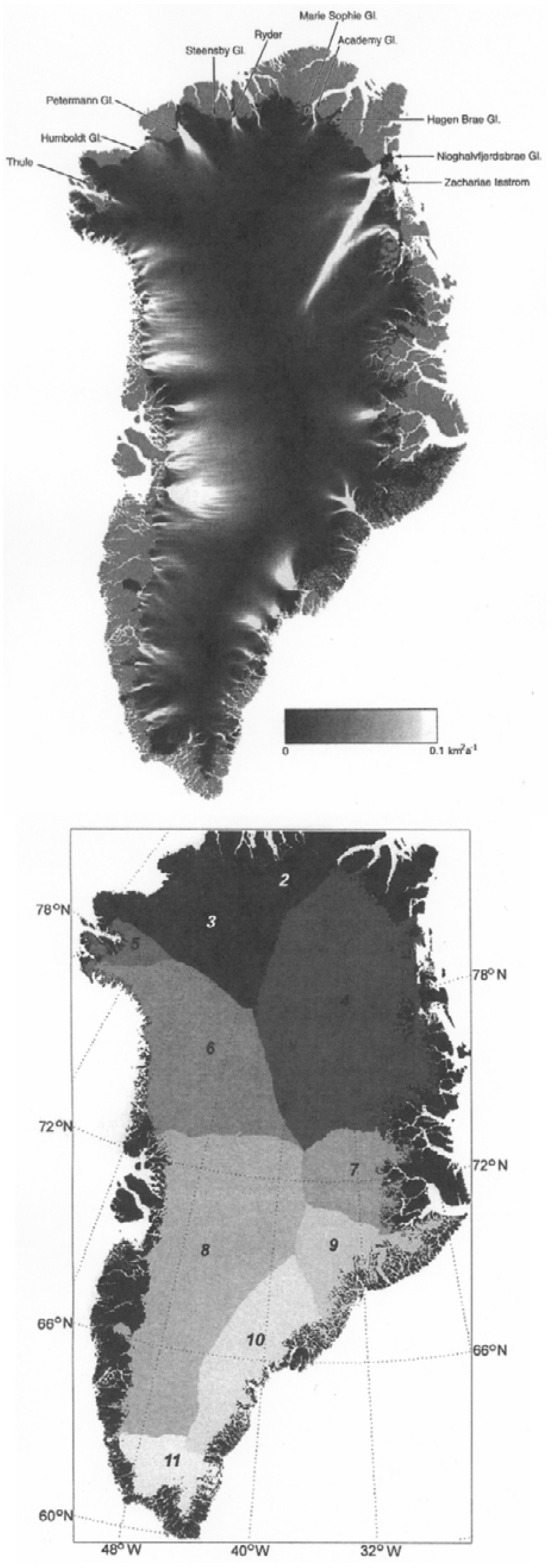

__Figure 6-5.__ A) Simulation of balance fluxes across the Greenland ice sheet. B) Ten major drainage basins calculated from the balance fluxes. (Figures from /Hardy et al. 2000/, reprinted with permission from John Wiley and Sons.)

6.2 Chemistry of melt and surface waters

The water consumption in towns and villages in Greenland is based on water from local lakes and rivers. The surface water in these water courses is primarily from rain or meltwater from snow and ice. According to /Steenfelt 2004/, Table 6-1, possible contributions from groundwater cannot be excluded but are likely to be very small.

The water chemistry in a lake is likely to be related to the chemical composition of bedrock and soil in the area upstream from the lake. In a research project carried out in and around different Greenlandic villages, sediments from small streams and some lakes were sampled and analysed /Steenfelt 2004/. Kangerlussuaq samples were taken in the area south of the town. The results of the chemical analyses are shown in Table 6-2. Sampling locations are shown in Figure 6-6.

Conclusions drawn were that the chemistry of sediments reflects the geochemistry of the gneiss. The electric conductivity of the water was in general low. The measured conductivity in the small lake studied in this area was, however, high. Also the fluoride content in this water sample was much higher than normally found in small streams in western Greenland. No conclusions were drawn whether the enhanced values were natural or caused by contamination.

Table 6-1. Chemical composition of surface water in West Greenland, 208 samples, together with EU Directions for Drinkwater quality (data from /Steenfelt 2004/.

Parameter	Unit	EU directions	Northern West Greenland			
		TMK	max.	min.	median	0.98%
Elect. conductivity	μS/cm	6.5	351.0	5.2	34.6	197.4
Al	mg/l	0.2	2.3	0.0	0.03	
Fe	μg/l	200	2081.0	0.0	17.0	397.7
Mn	μg/l	50	198.3	0.0	0.7	79.0
Cu	μg/l	3,000	94.0	0.1	0.6	12.0
Zn	μg/l	5,000	153.1	0.6	3.7	60.3
P	μg/l	5,000				
Fluoride	μg/l	1,500				
Co	μg/l		22.7	0.0	0.0	6.9
Ba	μg/l		163.4	0.2	2.2	64.5
Ag	μg/l	10	0.0	0.0	0.0	0.0
As	μg/l	10	1.0	0.0	0.0	0.3
Be	μg/l		0.3	0.0	0.0	0.11
Cd	μg/l	5	4.4	0.0	0.0	1.19
Cyanide	μg/l	50				
Cr	μg/l	50	4.1	0.0	0.1	1.11
Hg	μg/l	1				
Ni	μg/l	20	133.0	0.0	0.3	34.9
Pb	μg/l	10	4.1	0.0	0.2	1.6
Sb	μg/l	5	0.3	0.0	0.0	0.1
Se	μg/l	10	0.0	0.0	0.0	0.0
V	μg/l		8.5	0.0	0.0	4.1

Table 6-2. Concentrations of different chemical components in sediments in small streams and a small lake (GEUS nr 306538). Electric conductivity was measured in the water (data from /Steenfelt 2004/.

Kangerlussuaq

GEUS number		SiO_2 %	TiO_2 %	Al_2O_3 %	Fe_2O_3 %	MnO %	MgO %	CaO %	Na_2O %	K_2O %	P_2O_5 %	Ashrest %
306527		65.52	0.39	14.10	4.35	0.07	1.98	4.47	3.71	1.60	0.12	2.90
306528		58.24	0.44	11.92	4.19	0.08	2.01	4.21	3.00	1.33	0.19	13.70
306529		61.30	0.56	13.39	5.85	0.11	2.53	4.72	3.31	1.60	0.22	5.79
306535		61.82	0.93	13.59	7.02	0.14	2.58	4.99	3.32	1.51	0.17	3.13
306537		61.91	0.67	13.30	5.89	0.11	2.46	4.77	3.31	1.53	0.14	5.00
306538		64.32	0.69	13.90	6.36	0.12	2.82	5.23	3.46	1.56	0.19	0.75
306548		61.24	0.84	13.51	6.02	0.11	2.45	4.86	3.19	1.52	0.19	5.22
GEUS	As	Ba	Co	Cr	Cu	Ni	Pb	Sb	U	V	Zn	Water samples
number	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	mg/kg	µS/cm
306527	< 2	545	10	124	13	30	11	< 0.2	< 0.1	75	40	54
306528	< 2	411	15	147	56	42	9	< 0.2	< 0.1	61	34	63
306529	< 2	480	18	200	22	55	11	0.2	< 0.1	89	47	52
306535	< 2	486	16	174	22	43	12	0.2	1.1	101	59	32
306537	< 2	513	20	199	24	46	14	< 0.2	< 0.1	104	45	48
306538	< 2	462	37	130	13	61	9	< 0.2	< 0.1	108	105	440
306548	< 2	411	15	147	26	42	9	< 0.2	< 0.1	61	34	27

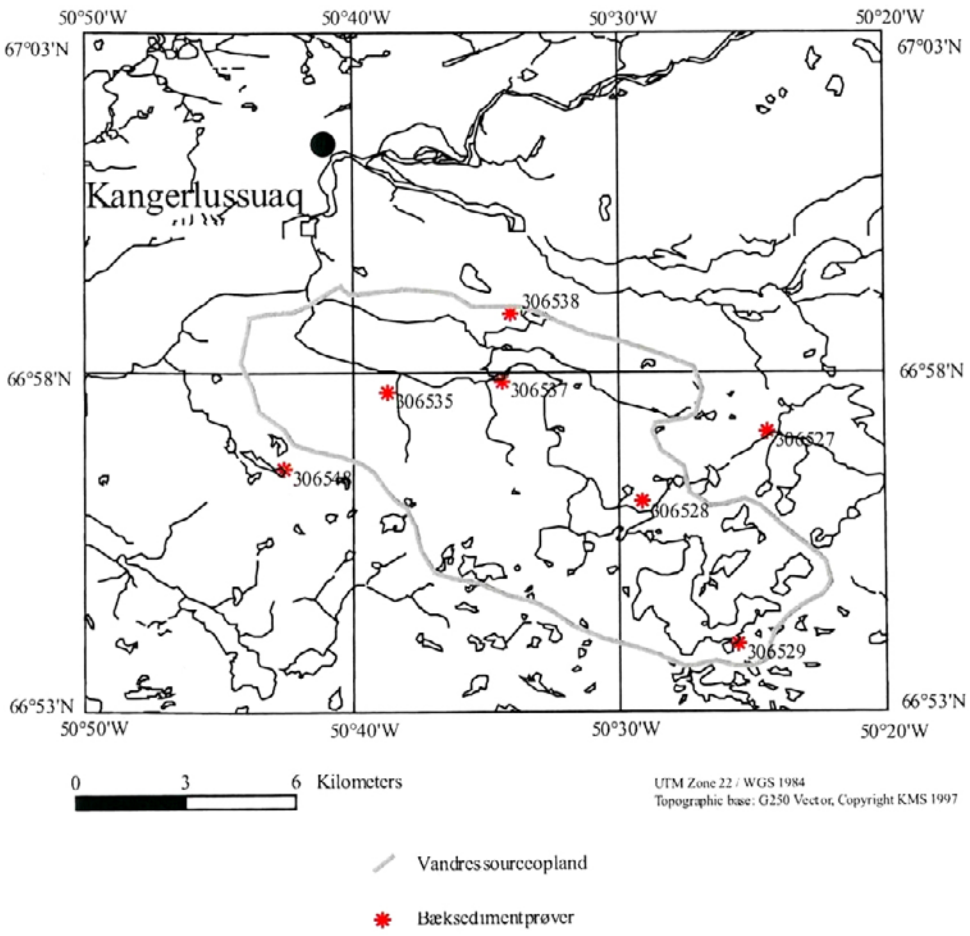

Figure 6-6. Sampling locations of stream sediments south of Kangerlussuaq. /Steenfelt 2004/. Copyright: Geological Survey of Denmark and Greenland (GEUS).

The area of southern West Greenland between 66° and 68° contains about 20 000 lakes /Anderson et al. 2002/. In a study including sampling and analyses of chemical composition of lake water between the ice margin and the coast in West Greenland between 66 and 67°N (Figure 6-7) /Anderson et al. 2001/ reported analyses of pH, alkalinity, conductivity and major ions from over 80 lakes. Most of these lakes were found to be dilute with electric conductivities < 150 µS/cm. The pattern of cations in the freshwater lakes is Ca > Mg > Na > K. There are, however, also saline lakes (2,000–4,000 µS/cm) in this study, mainly around Kangerlussuaq. The saline lakes occur within a zone of 80 km from the ice sheet margin. Closer to the Sea where precipitation is higher, saline lakes are absent. There is in general a clear gradient in water chemistry from the west (dilute coastal waters) towards the east. The limit between saline and dilute lakes agrees reasonably with the present-day limit of negative effective precipitation. According to /Anderson et al. 2001/, groundwater inputs and outputs are assumed to be small, although there are no available data.

The saline lakes in Greenland are different from the saline lakes in the Arctic and Antarctic. The latter lakes are often formed by trapping marine water in isolated basins as a result of isostatic uplift /Burton 1981/. The West Greenland lakes have generally a lower salinity than the saline lakes in the Arctic and Antarctic, and the primary reason for enhanced salinity in these closed-basin lakes is evaporation. However, evaporation is clearly not the only cause of enhanced salinity – external inputs of salt are likely. Local geological conditions as well as elevation give rise to variability in ionic composition.

/Willemse et al. 2004/ presented detailed physical and chemical profiles of a shallow, closed-basin, subsaline lake located southeast of Kangerlussuaq and modelled the hydrological balance and annual variation in chemistry. They concluded that throughout the open water period the hydrological balance is dominated by evaporative losses. During winter freeze-out of salts and resulting deep haline convection increase the overall water column salinity.

Most hydrochemical investigations if ice and meltwater in Greenland have focused on stable isotopes and hydrochemistry related to ice-core drilling in the interior of ice-sheet. Only few studies have been made on meltwater from the marginal area. /Reeh et al. 1991/ sampled oxygen isotope profiles in the marginal ice sheet area. They concluded that three different sources of ice meltwater are present in the hydrological systems near the margin of the ice sheet; superimposed ice, basal up-sheared ice and englacial ice.

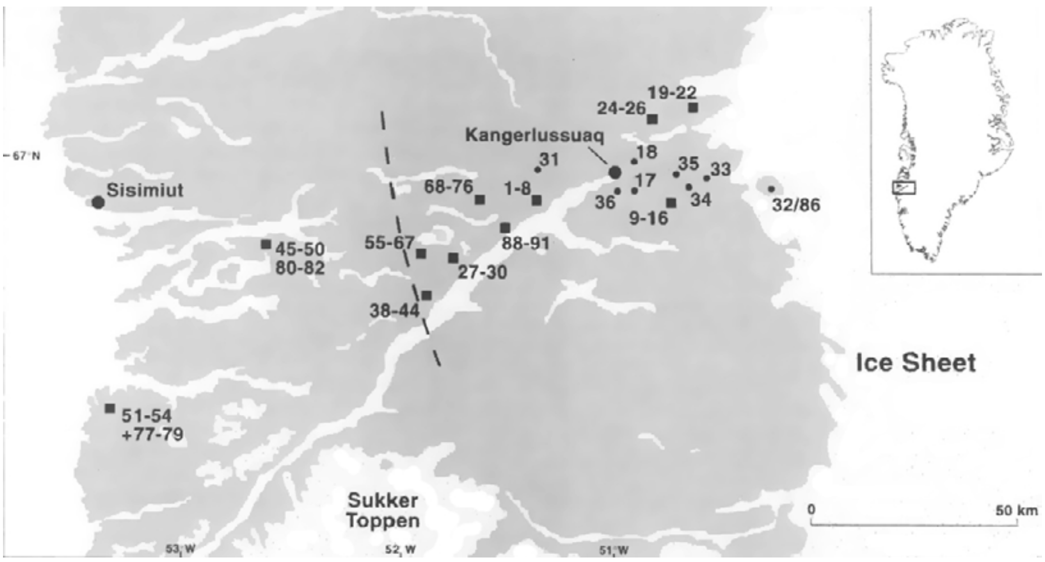

Figure 6-7. *Location map of lakes studied by /Anderson et al. 2001/. The approximate position of the present-day limit of negative effective precipitation is indicated (----). Reprinted with permission from the Institute of Arctic and Alpine Research.*

/Yde and Knudsen 2004/ carried out stable oxygen isotope analyses of bulk meltwater samples at Imersuaq Glacier, about 100 km south of Kangerlussuaq. A marked diurnal variation of $\delta^{18}O$ was interpreted to be related mainly to the composition of oxygen isotope provenances, mainly near-marginal local superimposed ice and basal up-sheared ice further up-glacier.

In another study of glacier hydrochemistry, conducted at the Kuannersuit Glacier, Disko Island, /Yde et al. 2005/ sampled bulk meltwaters in order to examine fluxes of major ions from a subglacial outlet. The results are summarised in Table 6-3. It should be noted that the hydrochemistry in this area is considerably affected by the subglacial basaltic weathering with absence of carbonate minerals.

Since the Greenland ice sheet constitutes a potential resource for export of drinking water, detailed analyses of different environmentally hazardous compounds, 72 different pesticides, 15 PAH's and 7 PCB's, were carried out in a research project /Bender et al. 2003/. The chosen glacier is located near Narssasuaq. The ice from this glacier did not have contents of any of the pesticides, PAH's or PCB's above the detection limits.

6.3 Data from other areas

Data on the hydrochemistry of glacial meltwaters have been presented from different glaciated environments. Observations of the variability of the chemical composition of meltwaters have been used to interpret the evolution of the hydrological system in time and space /e.g. Richards et al. 1996, Tranter et al. 1996, 1997, Theakstone and Knudsen 1996, Wadham et al. 2000/.

/Brown 2002/ presented a review of glacier meltwater hydrochemistry. Most of these data come from valley glaciers. Table 6-4 shows a compilation of dissolved ion concentrations. The results indicate that the dominant cation is Ca^{2+}, with lesser quantities of Mg^{2+}, Na^{+} and K^{+}. The dominant anions are HCO_3^{-} and SO_4^{2-}. The studies reviewed by /Brown 2002/ also show that the rate of chemical weathering is high in glacial environments.

Table 6-3. Ion concentration characteristics of bulk meltwaters at Kuannersuit Glacier. (data from /Yde and Knudsen 2005/. The four groups represent different sampling and analysis strategies.

		pH (μeql^{-1})	Na^{+} (μeql^{-1})	K^{+} (μeql^{-1})	Ca^{2+} (μeql^{-1})	Mg^{2+} (μeql^{-1})	Cl^{-} (μeql^{-1})	NO_3^{-} (μeql^{-1})	$SO4^{2-}$ (μeql^{-1})	alk (μeql^{-1})	ΔZ (μeql^{-1})	Si (mgl^{-1})	Fe (mgl^{-1})	Al (mgl^{-1})	Ion strength (μeql^{-1})
Group A	Max		170	4.6	220	150	58	25	66		470		4.7	4.6	1,100
(n=71)	Min		82	0.7	120	36	16	3.8	1.5		170		0.6	0.4	520
	Mean		120	2.2	60	73	25	14	8.9		310		1.8	1.7	720
Group B	Max									580					
(n=20)	Min									160					
	Mean									330					
Group C	Max	8.5	150	2.8	150	42	25	25	20	530	–280	4.6	0.8	0.9	740
(n=9)	Min	7.8	95	1.4	100	19	6.9	–	6.1	420	–170	2.0	0.1	0.1	520
	Mean	8.1	110	2.1	130	29	13	–	13	460	–220	3.2	0.4	0.4	640
Multisampling	Max	8.4	130	2.9	130	31	38	10	25	480	–270	4.6	0.5	0.5	680
(n=5)	Min	8.2	110	1.6	120	21	12	1.4	12	440	–190	3.7	0.1	0.1	610
	Mean	8.3	120	2.3	130	28	18	5.2	16	460	–230	4.1	0.3	0.4	650
	SD	0.1	10	0.5	1.6	3.6	11	3.8	5.3	18	34	0.3	0.2	0.2	26
Group D	Max	7.8	150	3.9	190	79	100	18	43	450	–170	8.5	2.1		890
(n=29)	Min	7.2	81	1.8	150	25	8.3	–	13	280	–15	2.9	0.0		630
	Mean	7.4	110	2.5	170	42	30	–	19	330	–58	4.3	0.5		720

ΔZ denotes the deficit in charge balance, Alk represents the total concentration of HCO_3^{-} and CO_3^{2-}, and has been corrected for the acidification effect during titration.

Table 6-4. Major dissolved ions in meltwaters draining selected glaciers (data from /Brown 2002/References in /Brown 2002/.

Major dissolved ion concentrations measured in meltwaters draining selected glaciers[a]

		Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^{-}	Cl^{-}	SO_4^{2-}	NO_3^{-}
/Anderson et al. 2000/	Bench Glacier (Alaska)	550	36	25	61	427	2	262	–
/Brown et al. 1996b/	Haut Glacier d' Arolla (Switzerland)	160–470	15–49	5.1–36	5.4–18	180–360	0.85–92	30–240	0.0–30
/Brown unpublished)/	Nigardsbreen (Norway)	8.8–38	1.6–7.8	8.3–25	0.98–4.4	1.4–8.5	9.8–25	7–41	1.9–11
/Collins 1979/	Gornergletscher (Switzerland)	130–334	16–190	8.7–43	2.6–33	–	–	–	–
/De Mora et al. 1994/	Walcott Glacier (Antarctica)	226–1,292	16–188	17–97	2.3–33	206–1,030	9.5–87	42–678	–
/De Mora et al. 1994/	Koettlitz Glacier (Antarctica)	72–92	5.8–6.6	11–34	0.77–6.9	91–132	0.55–1.2	3.4–7.6	–
/De Mora et al. 1994/	Howchin Glacier (Antarctica)	1,072–1,342	122–194	364–610	44–68	1,360–1,560	119–257	342–1,165	–
/De Mora et al. 1994/	Ward Glacier (Antarctica)	722–828	288–336	879–1,436	90–109	1,080–1,450	667–1,020	218–230	–
/Eyles et al. 1982/	Berendon (Canada)	90–763	1.6–19	0.87–7.8	0.38–5.1	230–785	25–27		
/Fairchild et al. 1994a/	Tsanfleuron (Switzerland)	638	92	4.9	6.3	627	5.5	118	11
/Hasnain et al. 1989/	Chhota-Shigri	75–260	6.6–41	25–65	22–51	–	–	–	–
/Hodgkins et al. 1998/	Scott Turnerbreen (Svalbard)	120–300	99–290	110–740	5.1–19	110–260		96–200	–
/Raiswell and Thomas 1984/	Fjallsjökull (Iceland)	208–274	32–60	30–120	2.8–7.2	190–300		26–66	–
/Rainwater and Guy 1961/	Chamberlain (USA)	75–304	8.2–123	4.3–8.7	0.0–5.1	150–200	5.6–20	29–310	–
/Ruffles 1999/	Engabreen (Norway)	82–623	4–65	11–212	0–27	51–675	10–191	0–142	0–15
/Steinpórsson and Óskarsson 1983/	Grimsvotn (Iceland)	359	115	482	12	573	87	132	–
/Theakstone and Knudsen 1996/	Austre Okstindbreen (Norway)	411–281	8.2–41	15–137	4.3–29	–	–	–	–
/Thomas and Raiswell 1984/	Argentière (France)	20–480	6–66	10–89	6–5.2	110–400	–	10–60	–
/Livingstone 1963/[b]	North America	1,048	211	391	36	1,114	226	416	16
/Livingstone 1963/[b]	Europe	1,552	461	235	43	1,557	195	500	60
/Livingstone 1963/[b]	Asia	918	461	404	–	1,294	245	175	11
/Livingstone 1963/[b]	World	749	337	274	59	957	220	233	16

[a] Units are in μeq l^{-1}. Where a single figure is presented this represents the mean concentration. Where two figures are presented separated by a rule this represents the range of concentrations measured.
[b] Mean composition of river waters cited in Holland (1978).

7 Hydrogeology and groundwater

7.1 Groundwater flow

There is generally little known about the groundwater conditions beneath deep permafrost in crystalline rock environments. In the Kangerlussuaq area there is no information at all about groundwater available in the literature. No groundwater wells seem to have been drilled through the permafrost. A number of deep boreholes have been drilled for diamond exploration purposes, but no data from these have been presented in open literature.

A number of hydrogeological studies have been carried out in the area around Sisimiut by the Arctic Technology Centre (ARTEK) at the Technical University of Denmark. /Hansen et al. 1998/ reported about a project aimed at finding useful groundwater resources in fracture zones in the bedrock. VLF measurements were carried out to identify the fracture zones and a 20 m deep well was drilled in bedrock consisting of gneiss and amphibolite. However, the well yield was much too low for the demands.

/Matthiesen et al. 2009/ carried out investigations at Andenfjorden, north of Sisimiut, where a hydropower plant is under construction. A well, which provides water to the construction site, has been established at a natural spring. Test pumpings have been conducted giving an estimate of the transmissivity of the sediment layers. No hydrogeological data from the 4 km long rock tunnel being excavated were presented in this report, but comparisons of the hydrogeochemical composition between well water and tunnel water were made.

It can be expected that an overall flow of deep groundwater in the study area of the GAP project occurs from the bedrock underneath the Inland Ice towards the Sea. The dominant fracture orientation in the region appears to be parallel to the regional tectonic foliation, i e ENE-WSW. Other fracture sets are oriented close to N-S and NW-SE. In the area of continuous permafrost, major groundwater discharge is likely to occur only through taliks beneath larger lakes and rivers. If there exist any larger deformation zones with high transmissivity, these zones may stay partly unfrozen and provide pathways for groundwater flow.

Although permafrost is not absolutely impermeable, significant groundwater recharge in the continuous permafrost area is not likely. Hence no active groundwater circulation is expected to occur.

7.2 Hydrochemistry of groundwater

A study of /Scholz and Baumann 1997/ at the tip of the Leverett Glacier (Figure 7-1), has brought some light on possible circulation of deep seated water from below the permafrost to the surface. Water samples were collected from a spring on top of an open system pingo located immediately in front of the glacier within a complex of crescent-shaped moraine ridges. According to the authors, the chemical analyses (see Tables 7-1 and 7-2) confirm that the spring water is not meltwater.

The high concentrations of dissolved substances show it to be Ca-Mg-HCO_3-SO_4-Cl type water, which can be described as highly mineralised compared with groundwater from other crystalline regions. The high bromide, chloride and sulphate contents suggest the origin of the water is deep seated, probably deriving from faults in the crystalline rocks below the permafrost. The water may be of thermal (> 20°C) origin, since the contents of fluoride, aluminium and gold are relatively high /Scholz and Baumann 1997/.

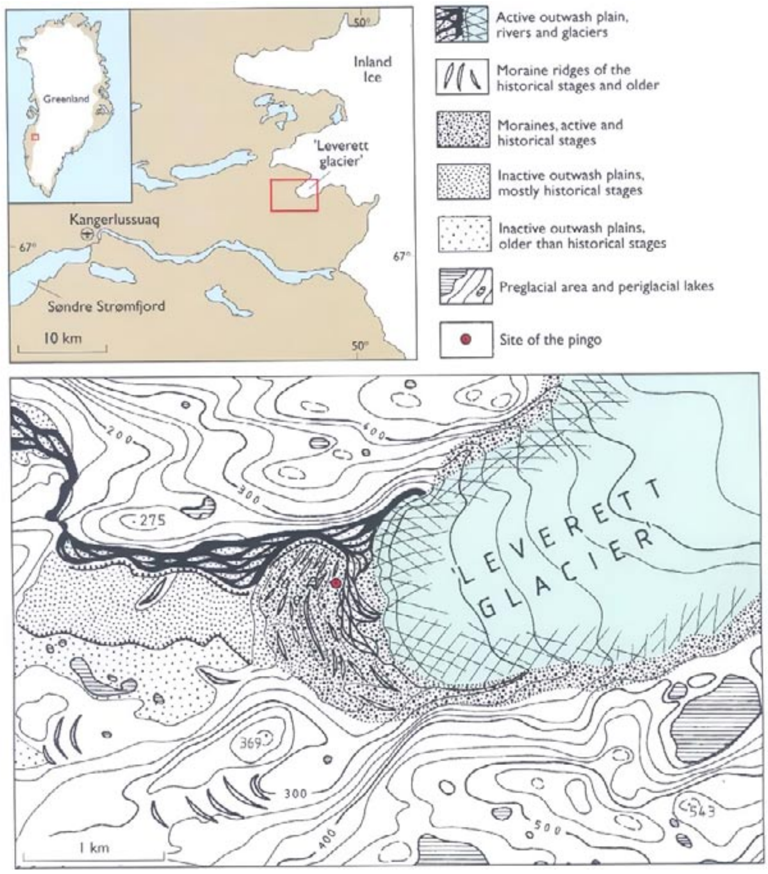

Figure 7-1. Location of the open-system pingo at the Leverett glacier /Scholz and Baumann 1997/. Copyright: Geological Survey of Denmark and Greenland (GEUS). The map is based on a 1:100 000 topographical map /Greenland Tourism 1995/.

Table 7-1. Chemical analysis of spring water sample "Leverett Glacier", West Greenland (data from /Scholz and Baumann 1997/.

1 litre of water	mg/l	equivalent mmol/l	equivalent %
Cations			
Sodium	13.70	0.596	13.62
Potassium	5.32	0.136	3.11
Calcium	41.20	2.056	46.98
Magnesium	19.30	1.588	36.29
Sum	68.52	4.376	100.00
Anions			
Fluoride	0.55	0.029	0.67
Chloride	39.13	1.104	25.49
Bromide	1.14	0.014	0.33
Nitrate	0.26	0.004	0.10
Sulphate	70.90	1.476	34.09
Hydrogen carbonate	103.90	1.703	39.33
Hydrogen phosphate	< 0.10	–	–
Sum	295.40	4.330	100.00
Temperature (not tested)	cold, but a few degrees above zero		
Electric conductivity	415 (µS/cm)		
pH	6.34		
Error of ion balance (%)	1.1		
Hardness (mmol)	1.8		
Hardness (°d)	10.2 (soft)		

Hydrogen sulphide may originally have been present, but has reacted with oxygen.

Table 7-2. Trace element content of spring water sample "Leverett Glacier" (data from /Scholz and Baumann 1997/.

Ag	< 0.1	Cr	1.70	Sb	< 0.1
Al	451.40	Cu	5.39	Si	5.70
Au	0.30	Li	1.30	Sn	0.54
Ba	57.20	Mn	330.0	Sr	< 100
Be	< 0.1	Mo	1.49	Tl	< 0.1
Br	< 0.1	Ni	4.00	V	4.10
Cd	< 0.1	Pb	0.10	Zn	5.02
Co	0.80	Rb	7.50	Fe	6,000

Elements are in µg/l.

7.3 Data from other areas

Unfrozen water in permafrost areas is in general relatively dilute if it is derived from the active layer where groundwater residence times are short. In coastal areas unfrozen water is however usually more saline. /Gascoyne 2000/ described two sources of the salt: 1) pore fluids that were included with the sediments during the time of deposition from saline water and 2) higher salinity waters formed by salt-rejection processes during permafrost aggradation.

Groundwater chemistry in the active layer from the subglacial zone of a glacier on Svalbard was sampled and analysed by /Cooper et al. 2002/. They compared the chemical composition with the bulk meltwater chemistry presented by /Wadham et al. 2001/ from the same site. Conclusions drawn were that HCO_3^- as well as Ca^{2+} and SO_4^{2-} concentrations were higher in the groundwater than in the meltwater. Temporal variations are shown in Figure 7-2.

Fluid salinity has been found to increase with depth in crystalline Shields, with reported salinities higher than 320 g/l in Canada, and even higher in Russia, according to a compilation by /Frape et al. 2004a/. The origin and formation of shield brines in crystalline rocks is explained by movement of external waters of marine origin into the crystalline rock /eg Bottomley and Clark 2004, Starinsky and Katz 2003/, recharge of glacial meltwater /e.g. Clark et al. 2000/ and *in situ* water-rock interactions /e.g. Fuge 1979, Nordstrom et al. 1989/.

/Starinsky and Katz 2003/ suggested that brines formed from surficial freezing of marine waters, and recharged deep into the crystalline shield during glacial times. They propose that the highly saline subsurface water is due to cryogenic brine formation which occurred as part of a highly dynamic flow system. Brines were formed from seawater within cryogenic troughs, along the subarctic continental margins, around ice sheets. Their emplacement in their present sites occurred most likely within the Pleistocene.

A new concept to shield brines in permafrost areas was put forward by /Stotler 2008/ who inferred that the concentrative mechanism can be in situ freeze out processes driven by permafrost and/or methane hydrate formation, thus resulting in an enrichment of heavier solutes in the remaining liquid. He discounted the hypothesis put forward by /Starinsky and Katz 2003/.

/Stotler 2008/ based his concept on a comprehensive investigation of the hydrochemistry of groundwaters within and below thick permafrost in Canada, at the Lupin mine in Nunavut /Ruskeeniemi et al. 2002, 2004, Frape et al. 2004b, Stotler 2008/.

Subpermafrost waters, sampled in boreholes at a depth of approximately 700–1,100 m at the Lupin mine, are of Na-Ca-Cl or Ca-Na-Cl type /Stotler 2008/. The relationship of Na/Cl and Br/Cl ratios (see Figure 7-3) differ from that of seawater evaporation /Mc Caffrey et al. 1987/.

Salinities range between 2.6 to 40 g/L, reflecting different fracture networks of different salinities and densities. The low salinities are explained by dissociation of methane hydrates due to the lowering of pressures accompanied by mining activity and taliks. Methane hydrates are stable well beneath the base of the permafrost, but likely not directly beneath hydrothermal through-taliks. Melting of gas hydrates can subsequently cause a dilution in groundwaters /Hesse 2003/.

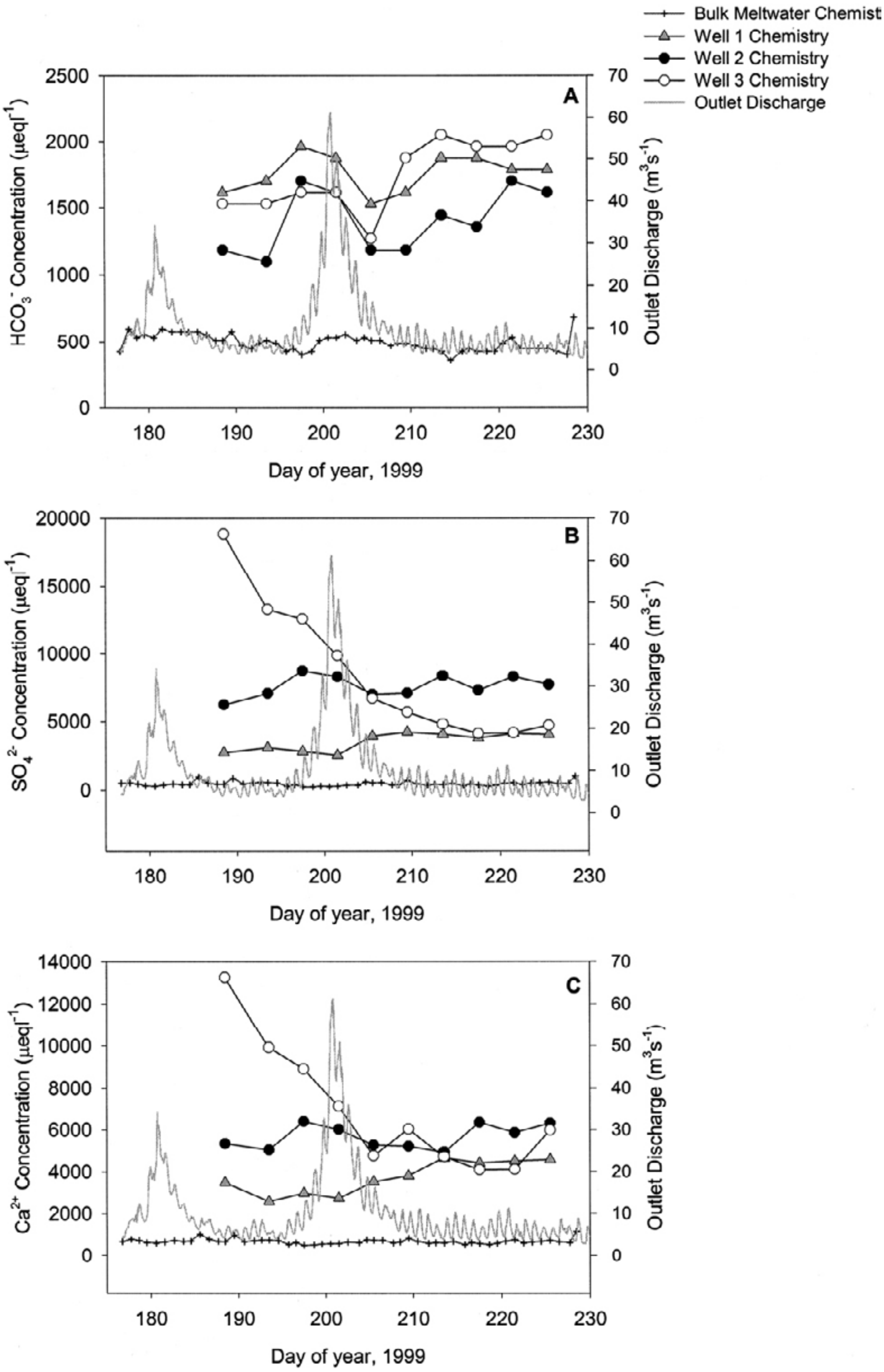

Figure 7-2. *Temporal variations in the concentrations of HCO_3^-, Ca^{2+} and SO_4^{2-} in wells and bulk meltwaters (Figure from /Coopet et al. 2002/, reprinted with permission from Elsevier).*

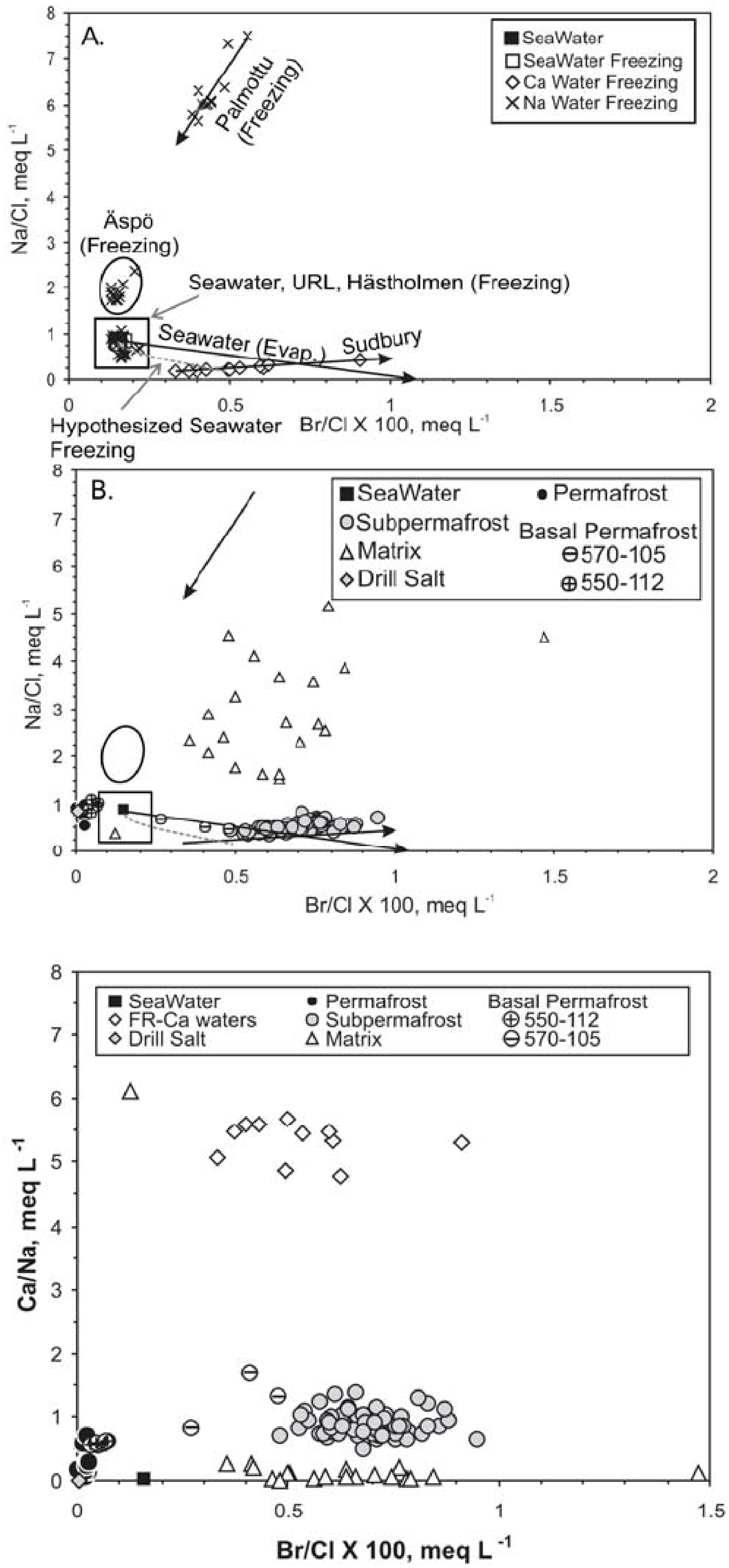

Figure 7-3. *Diagrams showing the relationships Na/Cl vs Br/Cl. /Stotler 2008/. Diagrams B and C show results from Lupin.*

Rock matrix fluids from crush and leach experiments from several boreholes at the Lupin mine are dilute and appear to have little effect on groundwater salinity /Stotler 2008/. However, high salinity Ca/Mg fluid inclusions were observed in calcite fracture fillings (> 26%) that might provide a small source of salinity and Ca to the subpermafrost waters.

The nature of gases observed at the Lupin mine were generally methane-dominated, with a unique isotopic composition compared to other shield gases, which is explained by a connection to metamorphism of marine turbidities at 1.8–2.8 Ga. The gas is inferred to be of thermogenic origin, however a generally low δ^2H-CH_4 cannot be explained.

Age tracers ^2H, ^{14}C, and ^{36}Cl from subpermafrost waters suggest input of modern fresh water, a recharge event 25,000 years ago and subpermafrost waters with an age > 0.5–1 million year old, which is the limit of ^{36}Cl dating technique /Phillips 2000/. The subpermafrost waters described above exclude brines formed from surficial freezing of marine waters, during glacial times.

Depletion of δ^2H and $\delta^{18}O$ in Canadian Shield sites may be indicative of either glacial meltwater recharge or *in situ* freeze-out /Stotler 2008/. From his studies at the Lupin mine and other data from the Canadian shield, /Stotler 2008/ thus concludes that shield brines might form from intruded glacial meltwater or waters from permafrost and/or methane hydrate formation.

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