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Complete List of Authors:	Masson - Delmotte, Valérie; LSCE, UMR CEA-CNRS-UVSQ 8212 Swingedouw, Didier; LSCE, UMR CEA-CNRS-UVSQ 8212 Landais, Amaelle; LSCE, UMR CEA-CNRS-UVSQ 8212 Seidenkrantz, Marit-Solveig; Center for Past Climate Studies, Department of Geosciences Gauthier, Emilie; Chrono-Environnement, CNRS-Université Franche Comté UMR 6249 Bichet, Vincent; Chrono-Environnement, CNRS-Université Franche Comté UMR 6249 Massa, Charly; Chrono-Environnement, CNRS-Université Franche Comté UMR 6249 Perren, Bianca; Chrono-Environnement, CNRS-Université Franche Comté UMR 6249 Jomelli, Vincent; LGP UMR 8591, CNRS Université Paris 1 Panthéon Sorbonne Adalgeirsdottir, Gudfinna; Danish Meteorological Institute, Christensen, Jens; Danish Meteorological Institute, Arneborg, Jette; National Museum, Bhatt, Uma; Geophysical Institute, Dept of Atmospheric Sciences Walker, Donald; Geophysical Institute, Dept of Atmospheric Sciences Elberling, Bo; University of Copenhagen, Department of Geography and Geology Gillet-Chaulet, Fabien; LGGE, UMR CNRS UJF Ritz, Catherine; LGGE, UMR CNRS UJF Gallee, Hubert; LGGE, UMR CNRS UJF van den Broeke, Michael; Institute for Marine and Atmospheric Research, Uttrecht University Fettweis, Xavier; University of Liège, Department of Geography de Vernal, Anne; GEOTOP, Université du Québec, Montréal
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Valérie Masson-Delmotte (*), Didier Swingedouw, Amaëlle Landais
IPSL/LSCE, UMR CEA-CNRS-UVSQ 8212, Gif-sur-Yvette, France;
valerie.masson@lsce.ipsl.fr

Marit-Solveig Seidenkrantz
Centre for Past Climate Studies, Department of Geoscience, Aarhus University, Aarhus, Denmark

Emilie Gauthier, Vincent Bichet, Charly Massa, Bianca Perren
Chrono-Environnement, UMR 6249, Besançon, France

Vincent Jomelli
CNRS-Université Paris 1 Panthéon Sorbonne LGP, UMR 8591, Meudon, France

Gudfinna Adalgeirsdottir, Jens Hesselbjerg Christensen
Danish Meterological Institute, Copenhagen, Denmark; Greenland Climate Research Centre, Nuuk, Greenland

Jette Arneborg
National Museum, Copenhagen, Denmark

Uma Bhatt, Donald A. Walker
University of Alaska, Fairbanks, USA

Bo Elberling
Center of Permafrost (CENPERM), Department of Geography and Geology, University of Copenhagen, Copenhagen, Denmark; The University Center in Svalbard, Longyearbyen, Svalbard, Norway

Fabien Gillet-Chaulet, Catherine Ritz, Hubert Gallée
UJF-Grenoble I / CNRS, LGGE, UMR 5183, Grenoble, France

Michiel van den Broeke
Institute for Marine and Atmospheric Research, Utrecht University, Utrecht, Netherlands

Xavier Fettweis
Department of Geography, University of Liège, Liège, Belgium

Anne de Vernal
GEOTOP, Université du Québec à Montréal (UQAM), Montréal, Canada

Deleted: France

Bo Vinther
Ice and Climate Center, University of Copenhagen, Denmark

Abstract

Climate archives available from deep sea sediments, glaciers, lakes and ice cores in and around Greenland allow us to place the current trends in regional climate, ice sheet dynamics, and land surface changes in a broader perspective. We show that the last decade (2000s) is reaching temperatures last encountered millennia ago, when northern high latitude summer insolation was higher due to a different orbital configuration. Records from lake sediments in southern Greenland document the major environmental and climatic conditions during the last 10 000 years, highlighting the role of soil dynamics in past vegetation changes, and stressing the growing anthropogenic impacts on soil erosion during the last decades. Past and present changes in atmospheric and oceanic heat advection appear to have major influence on both regional climate and ice sheet dynamics. Projections are investigated regarding the magnitude and rates of future changes in Greenland temperature, which may be faster than past abrupt events occurring under interglacial conditions. Within one century, in response to increasing greenhouse gas emissions, Greenland may reach temperatures previously encountered during the last interglacial period, 125 000 years ago. However, analogies between the last interglacial and future changes remain disputed because of the different seasonal impacts of orbital and greenhouse gas forcings. Over several decades to centuries, future Greenland melt may act as a negative feedback, limiting regional warming albeit with global sea level and climatic impacts.

Kalaallit Nunaat (Greenland) is the world's largest island (Figure 1), with 80% of its landmass covered by glaciers, ice caps, and the Greenland ice sheet (GrIS). If it were to melt, this volume of ice (~2,850,000 km^3) would correspond to ~7.2 m of global sea-level rise(1). The GrIS provides exceptional archives of past changes in regional climate and atmospheric composition, as unveiled by deep ice-core records(2). Concerns for future sea-level rise have grown with accelerating GrIS mass loss due to enhanced ice melting and discharge(3). This meltwater could have strong local and global implications, as the oceanic Atlantic Meridional Overturning Circulation (AMOC) (associated surface currents are displayed in Figure 1a) is highly sensitive to freshwater releases in the North Atlantic, with potential global climate implications(4). Regional climate models (RCMs), that have been specifically developed for Greenland, show a strong recent decline in the GrIS surface mass balance(5-8).

Greenland coastal climate has been monitored since the 18th century(9). In the last decade, monitoring of environmental changes, including glacier and ice-sheet mass balance, soils and vegetation, as well as marine and terrestrial ecosystems has intensified thanks to remote sensing techniques and in situ research stations, including automatic instruments. In parallel, paleoclimate studies based on natural ice, marine and terrestrial archives have provided a wealth of climate and environmental information(10).

The Greenlandic population of ~56,000 inhabitants(11), mainly lives in towns and settlements along the narrow ice-free coastal margins. Several waves of Paleo-Eskimo cultures ventured to Greenland from Canada(12) during the past 4500 years(13) (Figure 2b). In the late 10th century, southwest (SW) Greenland was colonized by the Norse. They established ~500 farms in the “green” inner fjords, reaching a maximum population of 2,000-3,000 people(14). Migrating from Alaska, the Thule people, ancestors of the current Greenlandic population, arrived in Greenland at the beginning of the 12th century(15). These migrations of peoples may have been related to past climate variability(16).

Today, the Greenlandic economy heavily relies on prawn, fish and seafood resources and supplies from Denmark; hunting and fishing are the main livelihood in the north and east sectors. The winter coastal sea ice cover has been important for hunting, fishing and transportation, with the exception of the SW sector where warmer surface ocean waters prevent sea-ice formation (Figure 1b). In this sector, relatively warm summer conditions (~10°C) and more fertile soils enabled the establishment of Norse farms in the Middle Ages and later modern sheep farming(17). Aiming at developing economical and political autonomy from Denmark, the Greenland Self Government encourages the development of oil and mineral exploration, in a response to new opportunities when sea-ice and land ice retreat(18).

In coming centuries, deglaciation and further greening (in the sense of enhanced biological productivity) of Greenland may drive a progressive shift from a marine to terrestrial subsistence. This will have major impacts on local ecosystems, socioeconomic, and cultural aspects. Here, we review ongoing Greenland physical environmental changes, and their impacts on Greenland vegetation and land ice, in the perspective of previously documented changes. We also explore the magnitude of projected Greenland physical environment changes as well as their potential local to global impacts, and compare future rates of changes with past abrupt events.

Large-scale drivers of Greenland climate change

During recent decades, Arctic warming has been two to three times larger than the global mean near surface air temperature (SAT) trend, albeit with a large decadal variability(19). The retreat of Arctic sea ice (Figure 1b)(20) plays a crucial role for this polar amplification(20). Recent Arctic warming has been attributed to the impact of anthropogenic greenhouse gas emissions on climate(21).

At intra and interannual time scales, the variability of Greenland winter SAT and precipitation is largely driven by atmospheric heat advection, related to the North Atlantic Oscillation (NAO)(9, 22), a large-scale atmospheric mode of variability(23). Due to the magnitude of winter NAO variability, interannual winter SAT variability is three times larger than summer SAT variability in south Greenland. The variability of coastal SAT also appears closely related to changes in local sea ice cover(24).

Greenland meteorological data reveal a sharp SAT rise starting in 1993, with 2001-2010 being the warmest decade since the onset of meteorological measurements, in the 1780s, surpassing the 1920s-30s by 0.2°C(25-26). The year 2010 was exceptionally warm, with SAT at coastal stations three standard deviations above the 1960-1990 climatological average. This warming was particularly pronounced in West Greenland(25) and associated with a record melt over the GrIS(6). It is related to the very negative NAO during 2010 and 2011. Warm North Atlantic and Arctic conditions damped the impact of this record low NAO on European winters(27), but enhanced Greenland warming in 2010.

Changes in volcanic or solar activity may also affect the NAO(28-29). Warm decades in Greenland occurred during periods with little volcanic forcing (1920s-1930s, 2000s to 2010s), whereas cold years (*e.g.* 1983, 1992) followed large volcanic eruptions(22, 26).

At decadal timescales, Greenland climate is strongly controlled by changes in ocean heat advection(30). Today, Greenland coastal regions are influenced by waters of both polar and

Atlantic origins (Figure 1a). Depending on the strength of the Irminger Current (Figure 1a), warm Atlantic waters may be found as far north as the northern Baffin Bay(31). During the last two decades, sea surface temperatures (SST) around Greenland have risen by ~0.5°C in winter and ~1°C in summer(32) in many areas, as the influx of Irminger Sea Water has increased.

This recent ocean warming around Greenland may be explained through the combined effect of NAO and a positive phase of the Atlantic Multi-decadal Oscillation (AMO)(33). The AMO is a 55-70 year cyclicity in Atlantic SST presumably related to internal ocean variability(34-35). The AMO has been in a distinct positive phase since the mid 1990s(33, 35), potentially related to enhanced northward heat transport in the North Atlantic(36). At the same time, the decreasing NAO decadal trend since 1995 has weakened westerly winds and the Atlantic subpolar gyre (compared to the 1980s), allowing an increased flow of the warm Irminger Sea Waters to West Greenland(37).

Current Greenland temperature changes in the context of the current interglacial period

In this section, we discuss the current changes in Greenland SAT, and then Arctic sea ice and regional SST, in the context of paleoclimate reconstructions spanning the last millennia.

Paleoclimate records allow placing the ongoing warming (with a linear SAT trend of 0.16°C/yr from 1993 to 2010) in a longer term perspective. Several Greenland continuous SAT reconstructions are spanning the last millennia (Table 1). The different reconstructions arise from (i) alkenones from sediments of one West Greenland lake(16), related to biological late spring-early summer productivity and water temperature, offering decadal resolution; (ii) air nitrogen and argon stable isotopes from one ice core, affected by changes in decadal changes in mean surface snow temperature(38); (iii) water stable isotopes from a stack of ice cores, corrected for changes in ice sheet elevation and tuned to SAT using information from borehole temperature records(39), with seasonal to bidecadal resolution. Different sources of uncertainties can affect each record (Table 1), which show different magnitudes of trends and decadal variability.

The lake record(16) shows a positive SAT anomaly from 4,000 to 3,000 yr BP (before present), large multi-centennial events, with estimated water temperature magnitudes from 1.5 to 5°C, and a variance of about 1.2°C (not shown). It does not exhibit any multi-millennial trend. The bidecadal lake data do not extend into the instrumental period and cannot easily be used to compare with current changes.

The GISP2 ice core (Figure 1a) gas isotope record produces a 1.5°C cooling trend along the last 4,000 years, together with multi-centennial events (<2°C), and an overall variance of 1.0°C(38) (not shown). The ongoing warming (mean level of the 2000s) estimated for GISP2 site from automatic weather stations and coastal SAT data appear comparable to the level of snow temperature reconstructed during the 1930s-1940s and during the warmest decades of the medieval period, in the 1140s. Prior to the last millennium, past reconstructed decadal snow temperature appears frequently above the level of the 2000s, especially in the earliest part of the gas-based reconstruction.

This finding contrasts with the comparison of coastal SAT changes with respect to the SAT reconstruction based on water stable isotopes from several ice cores (Figure 2)(39). This record differs from the gas record in the magnitude of the inter-decadal variance (0.7°C versus 1°C over the last 4,000 years) but shares the same multi-millennial trend (-0.4°C per 1000 years). However, very few decades of the last 3,000 years surpass the SAT level of the last decade in the isotope-based record.

Different factors can explain these results obtained with independent methods, such as different changes in annual mean snow surface temperature versus precipitation-weighted condensation temperature (Table 1). Two main factors explain the different findings obtained when comparing the recent warming with different ice core based reconstructions. First, the magnitude of the recent warming appears larger in coastal areas than at the ice sheet surface, especially in summer when the ice sheet energy budget limits summer warming. Second, the gas-based (snow) temperature reconstruction is associated with a larger inter-decadal variability than the isotope-based SAT reconstruction.

All ice-core records consistently demonstrate that the recent warming interrupts a long term cooling trend, very likely caused by orbitally-driven changes in northern hemisphere summer insolation (Figure 2a)(40). Using the water isotope-based dataset scaled to coastal SAT (Figure 2), the current coastal SAT (last decade) reaches levels comparable to the mean SAT of the mid-Holocene, 4 to 6,000 years ago, which coincided with the first documented human settlements in Greenland (Figure 2b).

Similarly, long-term trends aredocumented for Arctic sea-ice. A large reduction of sea ice occurred during the course of the last deglaciation, culminating in the early part of the current interglacial period in the eastern Arctic(20). Off NE Greenland, there is growing evidence for a minimum multi-year Arctic sea ice cover ~8,500-6,000 years ago, possibly in response to the strong summer insolation forcing (20, 41) (Figure 3). As summer solar insolation decreased over the last millennia, Arctic sea ice cover increased, reaching its maximum during the Little Ice Age. The current retreat in sea ice cover interrupts this multi-millennial trend, reaching levels (in the 2000s) far beyond those of the last 1,450 years(42) and last encountered in NE Greenlandabout 4,000 years ago at least(41). Many studies document strong regional fluctuations and East-West gradients in sea-ice cover changes during the current interglacial, possibly related with large scale (NAO) atmospheric dynamics(20, 41, 43).

High resolution SST records from the Fram Strait (west of Svalbard) indicate that the 20th century increase of the oceanic heat flux into the Arctic Ocean is unprecedented over the last ~2,000 years(44). The influx of warm Atlantic subsurface water towards SE and W Greenland has also strengthened in recent years(37, 45-46), but appears to remain within the range of recent natural SST variations. Indeed, opposite SST fluctuations between East Greenland and the Labrador Sea are reconstructed during the last millennia(47-50), possibly in relationship with NAO changes(49-50). There is evidence that, during the current interglacial, the inflow of warm subsurface water masses enhanced iceberg calving and discharge (51).

Impacts of climate change on Greenland glaciers and ice sheet

The current atmospheric and oceanic warming has large impacts on the ~20,000 Greenland Alpine and outlet glaciers. Since the early 1990s, remote sensing methods such as altimetry and velocity measurements from satellites and aircraft have revealed a marked acceleration and retreat of many outlet glaciers south of 70°N(3, 52). This increase in solid ice discharge has accounted for about 50% of recent GrIS mass loss(5). Despite uncertainties in the chronologies, moraine records demonstrate that the onset of modern glacier retreat(53) occurred between the middle of the 19th and the beginning of the 20th century(54). A compilation of snapshots of numerous glacier front positions documented by old photographs, maps, or paintings reveals a period of recession from the 1920s to the 1960s, followed by glacier advances in the 1970s to the late 1980s(53). The widespread recession of marine terminating outlet glaciers since the 1990s suggests a common forcing and occurs at a rate that is one order of magnitude larger than previously documented(55-58). There is new evidence for large fluctuations in the length of the Ilulissat Sermeq Kujalleq (Jakobshavn Isbrae glacier) during the current interglacial, with a smaller than present extent between 8,000 and 7,000 years ago(56). The Helheim Glacier (south-east Greenland) currently shows melting rates that presumably surpass those of the past ~ 4,000 years(58) .

From 1990 to 2010, the GrIS has lost ~ 2,750 Gt (Gigatons) of ice, with a significant acceleration in the rate of mass loss(3) (Figure 3). The different contributions to GrIS mass loss are quantified using satellite gravimetry measurements together with ice velocity from feature tracking and regional climate modeling of precipitation and runoff(5). Since about A.D. 2000, accelerating summer melt and iceberg discharges are not compensated by refreezing or enhanced accumulation. In 2010, record summer surface melt led to a GrIS total mass loss of 500 Gt (~1.4 mm/yr of sea level rise)(6) (Fig. 3).

Ice flow dynamics govern iceberg discharge, and induce a direct elevation feedback with the subsequent thinning of the ice margins. Ice flow dynamics is directly affected by enhanced surface run-off: surface melt-water can contribute (i) to a weakening of the lateral margins of fast flowing glaciers by filling the crevasses(59), and (ii) penetrate the ice sheet through crevasses and moulins, increasing basal lubrification and enhancing basal sliding of the ice on its bedrock(60). The relationship between water supply and velocities is not linear. When basal water pressure reaches a threshold, an efficient drainage can develop by opening channels, this limiting basal sliding (60-61). For land terminating glaciers, this effect is responsible for the observed diurnal and seasonal variations of velocities(62). However, the striking recent acceleration and retreat of numerous Greenland marine terminated glaciers have likely been triggered by ocean warming and processes happening at the terminus(52): dragging on the side of narrow fjords, floating ice tongues exert a backforce retaining fast marine terminated glaciers such as Jakobshavn Isbrae, Helheim or Kangerlussuaq glaciers (57, 63)(Figure 1a). The retreat of the calving fronts, likely triggered by enhanced basal melting, reduces this backforce and induce an acceleration and a subsequent thinning of the glaciers(52). This process can be effective for Greenland as long as glaciers terminate in the ocean, and are grounded below sea level. Ninety percent of the GrIS ice discharge is controlled by such tidewater glaciers(45).

The effect of ocean water on these tidewater glaciers is also believed to be linked to water temperature. Concurrent with increased surface melting since the late 1990s, hydrographic measurements have shown a pulse increase in the temperature of subsurface waters surrounding Greenland(37). Subsurface warm Atlantic waters enter Greenland's fjords to replace the out-flowing surface glacier meltwater(64). A direct pathway connects the North Atlantic open ocean with southeast Greenland glacier fjords(46), suggesting that a change in the prevailing water masses in the North Atlantic may impact the GrIS margins within one year(37, 46). There is also evidence of changes in ocean currents influencing glacier melting and iceberg production through the last few thousand years(51).

Present and future changes in Greenland permafrost

Retreating sea ice, glaciers, snow cover(1, 19) and warmer coastal conditions affect all Arctic soil ecosystems with underlying permafrost, representing ~25% of the northern hemisphere land area and containing almost half of the global soil carbon(65). Observations of northwest Greenland soil organic carbon suggest that such carbon reservoirs may be underestimated by at least a factor of five(66). On a global scale, soil-permafrost ecosystems are subject to dramatic changes including glacial retreat, coastal erosion and permafrost thawing(67).

At the Zackenberg research station, Northeast Greenland, the maximum thickness of the active layer has increased by ~1 cm/yr since 1996(68), as a result of increasing SAT, decreasing snow cover and an earlier start of the growing season(69). The spatial variability and timing of actual permafrost warming and thawing is only recently being addressed for Greenland(70-71).

A critical uncertainty is the heat production from increased microbial metabolism in soils and the accelerated decomposition(72). This has been shown to be significant in Greenlandic organic-rich soils(68) and has implications for future permafrost degradation rates(69).

Greenland warming also impacts the terrestrial carbon and nitrogen balance, with interplays between microtopography, biota, hydrology, and permafrost(68, 73). Observations from the Zackenberg monitoring station has revealed both spring and autumn bursts in CO_2 and CH_4, caused by physical release of the entrapped gas rather than enhanced microbial productions(74-75). Permafrost thawing also has impacts on waste piles (kitchen midden)(76), houses and infrastructures in settled areas.

Predictions for the active layer and permafrost thawing in Greenland are few(69-70). Permafrost degradation in high Arctic tundra areas in Greenland may reach ~10-35 cm over the next 70 years (Figure 4) and even higher in dry and more coarse-grained sediments. As a result, increasing permafrost thawing may in the future contribute with a CO_2 production equivalent to 50% of the present soil respiration(69). The potential compensation by plant carbon fixation remains uncertain. Permafrost degradation is expected to enhance runoff to lowlands, where the associated water level changes and nutrients inputs may have critical effects on methane and nitrous oxide production(68). Permafrost layers may be markedly richer than the active layer with respect to nitrogen (Figure 4). Thawing permafrost layers may therefore enhance the potential for a greening of Greenland in a warmer climate. Future changes in permafrost could have large impacts on coastal erosion, the carbon budget, vegetation and infrastructures.

Current changes in Greenland vegetation

Changes in sea ice concentration, land summer SAT and tundra gross primary production since ~1982 have been quantified using combined measurements from different sensors and satellites(24). Biweekly measurements of Arctic Normalized Difference Vegetation Index (NDVI, calculated from spectral reflectance measurements acquired in the visible and near-infrared wavelengths) at 12 km spatial resolution are used to estimate peak vegetation photosynthetic capacity (an indicator of tundra biomass) as well as gross primary production, combining the length of the growing season and phenological variations(24). The data depict a consistent increase of tundra photosynthetic activity in areas of land warming(24) and sea ice decline (Figure 1). This applies to SW Greenland, and to areas with retreating glaciers, where rapid vegetation growth occurs on recently exposed landscapes. The increase in open water in northwest Greenland is amongst the areas showing greatest change in the Arctic, with summer land SAT increase and time-integrated NDVI changes in the vicinity of Baffin Bay and Davidson Strait, amongst the largest in the Arctic (Figure 2). Complex species interactions determine the response of ecosystems to Arctic warming, changes in plant phenology, snow and ice depth and nutrient availability(77).

In southeast Greenland, a detailed comparison of vegetation species(78) showed only minor changes between 1968 and 2007. Species composition change was most pronounced in snowbed and mire habitats, likely caused by changes in snow cover and soil moisture linked with higher SAT. Recent warming also affected agricultural activities. The Greenlandic production of sheep and lamb has reached its highest and most stable levels in the 2000s, with more than 20,000 animals slaughtered annually(17). In recent years, the production of potatoes in Greenland (approx. 70 tons per year) has been steadily increasing (79).

Past changes in the Greenland vegetation: impacts of climate and agriculture

Beyond recent changes documented from historical archives, sedimentary records provide information on the past natural variability of Greenlandic vegetation. The warmer conditions encountered about 8,000 years ago (Figure 2) left their imprint in South Greenland lake sediments, in which pollen assemblages show more developed vegetation cover than at present, with dense alder populations(80) in coastal areas and juniper cover

(Figure 6) in the inner fjords. During the current interglacial period, changes in vegetation cover have responded to both local climate change and to soil developments. It has been suggested that juniper cover reflected dry conditions in the early Holocene(81). Increasing moisture and soil development in the mid-Holocene allowed the development of South Greenland endemic birch, *Betula glandulosa* and *Betula pubescens* (Figure 6)(82). The cooling trend of the last millennia was associated with a fall in pollen fluxes, about 2,000 years ago(81).

During the previous interglacial periods, high pollen influx and specific pollen assemblages from marine sediments depict dense vegetation mostly composed of shrubs and/or conifer trees(83). A spectacular development of spruce forest was very likely associated with a strong ice sheet retreat during the very long interglacial stage occurring about 400,000 years ago (Marine Isotopic Stage 11)(83).

Since 1920 AD, modern sheep farming and vegetable cultures have been developing in the relatively warm, sheltered inner fjords of south Greenland that first enticed Norse settlers to the region (Qaqortoq area). The Norse colonists lived as pastoral farmers, fishermen and hunters. Changes in precipitation and wind regime may have influenced their agriculture(84). Archaeological evidence indicates that the Norse adapted very well to new conditions and that the dependence on the marine mammals increased(85) when the climate deteriorated and made herding and pastoral farming more and more difficult(86). Paleoecological records support archaeological data (e.g. (87-90)). A sediment study of Lake Igaliku, the Norse *Garðar*, shows that Norse agropastoralism induced landscape modifications: non-indigenous plant taxa (*e.g. Rumex acetosa/acetosella*)(Figure 6e) increased at the expense of *Betula pubescens*(91). The sediment flux increased sharply at ~1000 AD, synchronously with vegetation changes, until it reached its maximum at ~1180 AD, at more than two times its baseline levels(81, 92). At the beginning of the 14th century, erosion and grazing pressure sharply decreased, suggesting a reduction in the sheep herds prior to the Little Ice Age.

Besides subsisting on local resources, the Norse settlements also depended on imports from Europe. Colder conditions and increasing sea-ice cover resulted in more treacherous navigation between Greenland and Europe, ultimately breaking off contacts in the later part of the 1400s(93). In the 12th century, the Inuit(15) brought new technologies (kayaks and dog-sledges) and spread across Greenland. Their ability to hunt or fish a variety of terrestrial and marine animal species equipped them to adapt to environmental change. Adaptation is part of today's Greenlandic society, making it responsive and ready to take advantage of the greening of Greenland(18) by expanding agricultural activities.

The Igaliku lake sediments document a much larger impact of recent agricultural activities. From 1906 to 1976, traditional sheep grazing used practices similar to those of the Norse, and sheep were left to graze openly in winter(17). Pollen and coprophilous fungi spores indicate disturbance levels that parallel those of Norse grazing pressure(91). However, after dramatic impacts of cold spring conditions in 1966, 1971 and 1975(17), farming methods switched to winter feeding, more intensive practices of hay production, mechanization, and fertilizer usage. Since 1976 (Figure 6), erosion has reached unprecedented values, more than twice the Norse maximum(81). Over the last decades, nitrogen isotopes and diatom microfossils document a marked shift in the lake ecosystem consistent with nutrient enrichment from agricultural sources as well as warmer summer SAT(81, 92). Current ecological conditions and soil erosion in the Igaliku region are unprecedented in the context of at least the last 1500 years. Given projected Greenland SAT and the anticipated growth of the farming sector, greater landscape changes must be expected in the future.

Past and present shifts in Greenland marine ecosystems

Large research efforts have been dedicated to the monitoring and assessing of marine ecosystems around Greenland, a focus of the Greenland Institute of Natural Resources(1, 19). While these studies are beyond the scope of this review, we note dramatic regime shifts in the shelf ecosystems during the early 1990s due to freshening and stratification of the shelf waters, which led to changes in the abundance and seasonal cycle of phytoplankton, zooplankton, and higher trophic-level consumer populations such as fish and marine mammals(94-95). Such changes in marine resources also affected modern and past Greenlandic cultures. Two earlier important transitions, from seal hunting to cod fishing, then from cod fishing to shrimp, deeply affected SW Greenland human populations during the 20th century(96). These economic transitions reflected large-scale shifts in the marine ecosystems. The combination of climate variations and fishing pressure, for example, was dramatic for West Greenland's cod fishery(18, 96).

Living from ice fishing and hunting, some early Greenlandic cultures (e.g. Dorset) were dependant on long sea ice seasons, while other cultures (e.g. Saqqaq) based their food source on hunting and fishing in more open, ice-free waters. Natural climate variations superimposed on the long term cooling trend likely affected prey availability and were responsible for human migrations(13, 16). The demise of the Saqqaq culture coincided with a reduced inflow of warmer Atlantic source waters to the coastal regions of West Greenland(49), limiting the availability of e.g. harp seals. Colder conditions and changes in ringed seal hunting were also suggested to be at the origin of the Dorset disappearance from Greenland(97).

Projected future Greenland climate changes

Coupled climate model projections have been analysed for SW Greenland. In CMIP3 (Climate Modelling Intercomparison Project, Phase 3) simulations, the SRES A1B scenario corresponds to a prescribed increase in CO_2 concentrations, reaching 720 ppmv in year 2100. This scenario induces a median SW Greenland SAT warming of 3.3±1.3°C(98-99). Global simulations have recently been refined with RCMs(5-6) to better assess regional impacts, with a focus on the GrIS surface mass balance (1). When forced by atmospheric reanalyses, the MAR regional model reliably simulates the magnitude of coastal SW Greenland SAT variability from 1958 to 2001 (Figure 2c). Projection scenarios were built using RCMs forced by the outputs of ECHAM5 climate model, representative of the average global climate model projections(99). The calculation based on MAR (Figure 2c) shows a SW coastal Greenland SAT warming trend of 4.7°C per century, amplified compared to the ECHAM5 trend (+3.5°C per century) by the snow albedo feedback. MAR depicts a 1 month (+30%) increase in SW Greenland growing season length, a 60% increase in the positive degree days with rather stable precipitation amounts. A very high resolution case study conducted with the HIRHAM RCM for the Kangerlussuaq area (Figure 1) leads to similar results(100) .

Recently, new projections have been conducted under new greenhouse emission scenarios, and using the coupled ocean-atmosphere models from CMIP5 (Coupled Model Intercomparison Project, Phase 5) database that will be used in the 5th assessment report of the Intergovernmental Panel on Climate Change. Given the spread within available simulations, it is likely (50% confidence) that the rate of SAT change may exceed 2.5°C per century (RCP4.5 scenario) and 5.5°C per century (RCP8.5 scenario) (Figure 7b). These rates of changes can be compared with past natural changes documented by ice cores.

Indeed, the history of Greenland climate is marked by numerous abrupt Dansgaard-Oeschger (DO) events. These DO events are characterized by a multi-millennial cold phase, followed by an abrupt warming with an amplitude reaching up to 16°C within a few decades to centuries (Table 2), followed by return to colder conditions. The last climatic cycle is marked by 25 DO events(2), which have a global impact(101) including monsoon shifts(102) and variations in atmospheric greenhouse gas concentrations. The Antarctic counterpart of DO events is characterized by an anti-phase behavior, with Antarctica slowly warming during cold Greenland stadials, and slowly cooling after the onset of warm Greenland interstadials(103-105). This bipolar seesaw behavior of SAT anomalies in Greenland and Antarctica is a consequence of AMOC global reorganization(106), possibly in response to massive freshwater release from glacial ice sheets(107). The beginning of the current interglacial period is marked by a sub-centennial cooling event, around 8,200 years ago, likely caused by the impact on Lake Agassiz on North Atlantic ocean currents(108), followed by a progressive recovery(109-110)(Figure 2b).

An investigation of the rates of SAT changes must take into account uncertainties in the duration of DO events and on the magnitude of abrupt warming (Table 2). A probabilistic approach has been conducted on 11 documented events (here, limiting the investigated events to those lasting more than 60 years), showing that their median warming rate is 5°C per century. We also note that several abrupt events occurring under a warm climate background (e.g. glacial inception, last deglaciation) tend to have smaller rates of temperature changes (Figure 7), up to ~2.5°C per century during the first DO event, DO25(111), and the recovery from the cold event, 8 200 years ago(112)(Figure 2, Figure 7). In business-as-usual scenarios (RCP8.5), Greenland warming may therefore be more abrupt during the 21st century than these past abrupt warming events occurring under interglacial conditions.

Climate projections suggest that, by the end of the 21st century, Greenland climate may be 5°C warmer than during the last decades (1970-2000), reaching conditions comparable with those previously encountered during past warm interglacial periods(113-114). The climate response induced by changes in orbital forcing are characterized by a large mid-to high latitude summer warming, with year-round impacts linked with sea-ice retreat. This contrasts with the impacts of increased greenhouse gas concentrations, leading to larger winter warming. However, the two types of forcings produce similar magnitudes of summer warming, and similar magnitudes of sea ice, cloud or water vapor feedbacks(114). Systematic model-data comparisons for the Last Interglacial period offer the potential to assess the realism of climate models in a context relevant for the magnitude of future changes.

Projected future Greenland ice sheet and glacier changes

Future Greenland climate change is expected to impact coastal sea ice cover, extreme events, river runoff and its potential for hydroelectricity production(19). The large impact of external natural forcings and internal variability of the ocean and atmospheric circulations (*e.g.* AMO and NAO) on Greenland climate calls for a careful interpretation of projections(1). Links between climate forcings, large-scale modes of variability, and local extreme events remain to be investigated.

Recent studies have investigated the possible future evolution of the GrIS. Climate projections have been used to quantify the changes in the surface mass balance(99), while empirical approaches have been deployed to estimate the potential range of the ice sheet response(115-116) which is starting to be described in new generations of GrIS models(117). Most studies predict increasing GrIS mass loss, an acceleration of fast flowing glaciers(118), and a potential contribution to sea level rise of several tens of centimeters by 2100(1).

The projected future Greenland ice sheet retreat may also be compared with the evidence for major mass loss during the Last Interglacial period (130 to 120 thousand years ago), characterized by a global sea level >6 m higher than today(119). Large uncertainties remain on the magnitude of Last Interglacial GrIS mass loss, which could have contributed at least 1.5m of sea level rise(120-122). There is no precise estimate of the rate of this past retreat. Orbitally-driven changes in summer insolation may have directly contributed to about half of the GrIS mass loss (the other half being caused by orbitally-driven changes in SAT), limiting the analogy with future changes(123).

GrIS melt may have global impacts on sea level and climate. During glacial periods, major reorganizations in AMOC associated with DO events may have be driven by massive meltwater inputs, provided by past ice sheet instabilities(Figure 2)(105, 107). These past abrupt AMOC changes had well documented global impacts, notably with a cooling of the North Atlantic region and migrations of the inter-tropical convergence zone(4, 124-125).

Sensitivity studies have been conducted to investigate the response of AMOC and climate to future GrIS meltwater fluxes, with varying results(4, 126-128). Differences may arise from the prescribed melting rates(129) and from the sensitivity of the AMOC in each climate model to both CO_2 increase and freshwater perturbations. For instance, a large weakening of the AMOC in response to global warming and enhanced North Atlantic precipitation may hide a weakening due to ice sheet melting. The sensitivity of AMOC to freshwater can be highly non-linear(130), due to the potential existence of a bifurcation point for the AMOC dynamics identified in simple ocean circulation models(131). Two studies show that the AMOC may significantly weaken for a Greenland melting rate above 0.1 Sv ($10^6 m^3/s$) in 2100, a pacing not incompatible with estimates of GrIS mass loss acceleration(3). By limiting the warming around Greenland, a weakened AMOC may act as a negative feedback for the GrIS mass loss. Altogether, the magnitude and pacing of GrIS melting and the feedbacks between melt and AMOC remain uncertain.

Conclusions

Climate projections suggest that, by the end of the 21st century, future Greenland climate may be comparable with mean conditions previously encountered during last interglacial period, which was also marked by significant Greenland ice sheet mass loss. We have shown that, in response to increases in atmospheric greenhouse gas concentrations, projected SAT changes may occur at a rate comparable or higher than past abrupt warmings occurring under interglacial conditions (e.g. 8.2 ka event, DO 25).

Despite different drivers of past and future climate changes, past climates offer “natural experiments” to assess the ability of climate models to resolve past variations with magnitudes or rates of changes relevant for future changes. Preliminary comparisons suggest that climate models may underestimate Greenland warming during the Last Interglacial, possibly due to the lack of changes in ice sheet and land surface (northern hemisphere vegetation) feedbacks(114). Simulations of past abrupt events, in response to prescribed freshwater forcing, also seem to underestimate both the magnitude and rate of stadial-interstadial transitions in Greenland(132). Cross investigations of past and future simulations conducted with the same models will be possible using the CMIP5 (Climate Model Intercomparison Project) model output database.

Paleoclimate records moreover highlight the large inter-annual, decadal, centennial variability of Greenland SAT, related to large-scale changes in atmospheric and oceanic dynamics, and possibly driven by external forcings (orbital, solar and volcanic forcing). So far, very few detection-attribution studies have been conducted for this area (21). The emergence of ensemble multi-millennia transient simulations with climate models opens the possibility to further investigate and quantify the relative importance of internal variability and of the deterministic response of Greenland climate to external forcings.

Past climate variability and current climate change have had and are having large impacts on marine and terrestrial ecosystems around Greenland, with consequences for resources and human societies. There is evidence of past vulnerability (cod stocks) but also of resilience (limited impacts of Norse agriculture) of ecosystems to human pressures. With a cultural heritage of “being prepared for surprises”(18), Greenlanders face opportunities and threats linked to the deglaciation and greening (enhanced biological productivity) of Greenland. Perception studies(133) and combined use of traditional knowledge and climate model projections are needed to assess the impacts of climate change on coastal areas. Links between climate forcings, large-scale modes of variability, and local extreme events remain to be investigated.

Changes in local landscape such as the extent of coastal glaciers need to be anticipated, which requires an improved documentation of their mass balance. Agronomical models can be used to quantify the potential impacts of a longer growing season on terrestrial vegetation and the potential for new types of cultures, including the needs for irrigation, as previously used by the Norse(134). Changes in permafrost potentially have large impacts on coastal erosion, the carbon budget, vegetation and infrastructures. Long term monitoring efforts must be maintained and expanded, to assess and improve the models used for predictions.

The response of the GrIS to warming is of global strategic interest, not only for sea level but also for its potential impacts on the AMOC, atmospheric circulation and precipitation. A better understanding of the ocean-atmosphere-cryosphere interactions is needed to reduce uncertainties on projections. The key processes affecting the GrIS dynamics (impact of surface water production on basal lubrification, and retreat of the calving front of floating tongues) are located at the margin of the ice sheet and have typical spatial scales of a few kilometers. Small-scale glaciological models start to resolve this type of processes, but their inclusion in GrIS models remains a challenge, addressed by ongoing international projects aiming at better constraining sea level rise from melting land ice in the 21st century. A precise documentation of past changes in Greenland ice sheet mass balance, especially during the Last Interglacial, is needed to benchmark this new generation of ice sheet models.

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Figure captions

Figure 1.

a) Map of Greenland showing the ice sheet extent (white), schematized surface oceanic currents affecting Greenland climate (red arrows, warm surface currents; dashed blue arrows, cold surface currents; EGC: East Greenland Current; WGC: West Greenland Current; B-LC: Baffin-Labrador Current), the largest towns and settlements (yellow circles) as well as ice core drilling sites (orange circles). Adapted from (135) by Martin Jakobsson, Stockholm University.

b) Greening of the Arctic. Satellite observations of Arctic sea ice reduction (indicated by the trend in the percentage of open water) and tundra vegetation productivity (indicated by the MNDVI, modified normalized difference vegetation index). Trends are calculated from 1982 to 2010 using a 10 km resolution, updating earlier data(24).

Figure 2. Current Greenland warming in the perspective of natural climate variability and future projections.

a) NorthGRIP ice core $\delta^{18}O$ (‰), a proxy of Greenland SAT(2) at a 20 year resolution (grey) and multi-millennial binomial smoothing (red) as a function of time (years before A.D. 2000); the orbital forcing, which is the main external driver of glacial-interglacial trends, is illustrated by the 70°N June insolation (W/m²). Red areas highlight the interglacial periods and the blue area highlights the last glacial period; the green area indicates the instrumental period.

b) Estimate of southern GrIS(39) SAT anomalies during the current interglacial period (°C, with respect to the last millennium) (grey, 20 year resolution; red, millennial trend) based on a stack of ice cores and a correction for elevation changes(39) and a comparison with the instrumental SAT record from southern Greenland updated to 2010(9) (black, 10 year resolution). The SAT level of the decade 2001-2010 is displayed with a horizontal dashed black line. The 2010 anomaly is displayed as a filled diamond. The vertical rectangles illustrate the succession of human occupations of Greenland, from archeological data (see text). The red area illustrates the current interglacial period, and the green area the instrumental period. The rate of SAT change during the abrupt warming, approximately 8,200 years ago, is also indicated (2.5°C per century).

c) Meteorological records from southern Greenland based on a stack of meteorological data updated to 2010(9) (thin black line, annual data; thick stair steps, decadal averages). The data are compared to the MAR regional climate model results for the south-west Greenland coastal area, forced by ERA-40 (green) and ERA-interim (orange) boundary conditions from 1958 to 2010(8). Data are displayed as anomalies from the 1960-1990 period, which is 0.5°C above the average data for the last millennium as displayed in panel b. The 2010 SAT anomaly is highlighted as a filled diamond. An example projection is given using MAR forced by the ECHAM5 A1B projections (red line, annual values; red stair steps, decadal values). This corresponds to a warming trend of 4.7°C per century.

Figure 3. Cumulative updated(5) anomalies of major mass balance components of the GrIS, 1990-2010, and GRACE gravimetry estimate of mass loss, vertically offset for clarity. Abbreviations are explained in the legend. SMB data from RACMO2 RCM(5). GRACE data courtesy of I. Velicogna and J. Wahr.

Figure 4. A) Observed and predicted permafrost degradation in Zackenberg 1900-2080 based on down-scaled HIRHAM RCM data. Projections are given for two vegetation types: wetland (brown), heath (green) and two scenarios: a 2°C global warming over 100 years (filled symbols) and 2.4 °C over 60 years (open symbols). Running means over 10 years are shown as solid lines. B) Active layer and permafrost total soil organic carbon and C) Ammonium concentrations in melt water(68).

Figure 5. Illustration of the impact of a large GrIS meltwater flux (>0.1 Sv) on global climate projections using the IPSL CM4 model(4). SAT (top) and precipitation (bottom) changes for 2×CO_2 (averaged over years 450-500)(136) with respect to the preindustrial control simulation when including (right) or not (left) the impact of GrIS meltwater flux. A strong reduction in the AMOC induces a reduced warming in the north Atlantic but enhanced warming in the southern hemisphere tropical Atlantic, resulting in a southward shift of the Inter tropical Convergence Zone. Such a migration may have strong impacts on tropical precipitation distributions. This type of behavior has been found in a multi-model ensemble for modern conditions and appears to be robust under global warming conditions(125).

Figure 6. Schematic representation of environmental changes recorded by the Igaliku lake sediments (81-82, 92): a) water quality estimated from diatom assemblages), b) soil erosion rates estimated from the minerogenic and organic inputs into the lake and controlled by a set of geophysical, geochemical and ecological parameters including magnetic susceptibility, titanium content, bulk organic matter geochemistry and diatom valve concentration, c) vegetation history from pollen and non-pollen palynomorphs analyses, and d) archeological periods. Limited impacts of Norse agriculture are reflected by indicators of clearance and sheep grazing, as well as by the persistence of introduced species. Modern agriculture is marked by clearance, soil erosion, and the onset of the

first mesothropic phase of the last 10,000 years; e) Photograph of Norse apophytes (*Rumex acetosa* - *Taraxacum* sp) on a medieval archeological site in south Greenland (photograph: E. Gauthier, 2007).

Figure 7. a) Probabilistic estimate of the rate of SAT change over the course of stadial-interstadial events, with a duration longer than 60 years. Data are represented as a probability density function (%) as a function of the rate of SAT change (°C per 100 years), calculated from the published uncertainties on event duration and magnitude (See Table 1). Color codes reflect the CO_2 concentration (as an indicator of the back ground climate) during events (from blue, concentrations between 200 and 215 ppmv, orange, 220 to 230 ppmv, brown, 230 to 240 ppmv and red, 240-260 ppmv). The black line displays the mean probability density, calculated from the 11 studied events). There is a tendency for having slower rates of temperature rise (DO20, DO22, DO23, DO25, BA) under "warm climate" background. DO 22 appears to be very close to a "mean" event.

b) Rates of changes for future climate in RCP4.5 and RCP8.5 projections. Simulations from 13 models or model versions have been considered (NorESM1-M, MRI-CGCM3, MPI-ESM-LR, MIROC-ESM, MIROC-ESM-CHEM, MIROC, IPSL-CM5A-LR, inmcm4, HadGEM2-ES, CSIRO-Mk3, CNRM-CM5, CCSM4, CanESM2, HadGEM2-ES). Results are displayed in terms of cumulative frequencies within the 13 models.

Tables

Table 1. Comparison of the four available terrestrial Greenland temperature reconstructions spanning the last millennia.

Archive	Proxy – Target climate variable	Length of the record Temporal resolution	Key limitations
Ice cores	Water stable isotopes ($\delta^{18}O$, δD)(39) Precipitation weighted, condensation temperature controlling atmospheric distillation	Several ice cores (DYE3, GRIP, GISP2, NGRIP) spanning the Holocene (seasonal resolution) (137), the last glacial period (annual to decadal resolution) (138). One ice core (NGRIP) with a continuous record back to the last interglacial (123 ka) (20 year resolution) (2) (111).	At high frequency (season) : signal to noise ratio caused by deposition and post-deposition processes (139); Intermittency of precipitation (seasonality) (105); Changes in evaporation conditions (140) (141); Changes in ice sheet elevation (142).
Ice cores	Air isotopes ($\delta^{15}N$, $\delta^{40}Ar$) (38, 105) Surface snow temperature changes, generating temperature gradients in the firn and affecting thermal and gravitational diffusion of gases in the firn.	Quantification of abrupt temperature changes in GISP2, GRIP or NGRIP ice cores (105); One continuous record spanning the last 4 000 years with decadal resolution (38)	Variability of air isotopic composition during pore close-off and analytical accuracy; Storage effect or fractionation associated with clathrate formation (143); Uncertainty in accumulation rate; Uncertainty in thermal fractionation coefficients; Increments used to model temperature impacts. Changes in ice sheet elevation (142).
Ice cores	Inversion of borehole temperature profiles(144) (145)	Low frequency variations with a loss of resolution back in time. Detection of decadal variations (last century), multi-	A priori hypothesis on temporal temperature profiles. Changes in ice sheet elevation (142).

		centennial variations (last millennium), millennial variations (current interglacial) and glacial-interglacial magnitude.	
Lake sediments	Alkenone undersaturation in two Greenland lake sediments (16)	Decadal to centennial resolution, spanning 5600 years before present.	Salinity threshold. Seasonal (spring – early summer) temperature signal from algal bloom. Possible influence of parameters other than temperature (e.g. cloudiness, nutrients) on productivity. Lake temperature likely affected by wind speed (mixing).

Table 2. Summary of the timing, magnitude (from gas thermal diffusion) (K) and duration (years) (from water stable isotopes) of stadial-interstadial transitions from Greenland ice cores (105). DO stands for Dansgaard-Oeschger stadial-interstadial transition. Events for which either no temperature estimate is available, or with durations likely shorter than 60 years(and therefore associated with uncertainties of 1/3 or more on the duration) were not used to estimate centennial trends. These short-lived or poorly characterised events are depicted in italics. GICC05 refers to the most recent Greenland counted age scale(138, 146).

(*) The method used to determine the amplitude of the temperature change at the end of the Younger Dryas (YD)(147) is based on a static firn heat diffusion model with temperature forcing as a step function. The method developed for the Preboreal Oscillation (PBO)(143) is more sophisticated and is based on yearly annual incrementation of temperature to fit the $\delta^{15}N$ profile as well as a complete firnification and heat diffusion model(148). This latter approach has the disadvantage that small errors in the temperature increment are cumulative. In order to be coherent with the following amplitudes of temperature changes on NorthGRIP that have been performed using the firnification and heat diffusion model(148). forced by different temperature scenario inspired from the ice core $\delta^{18}O$ profile(149), we have checked the values obtained on the YD and the PBO with this method. For the end of the YD, our results confirm earlier results (147); even with variations by a factor of 4 of the rate of temperature increase at that period, the amplitude of the temperature increase remains between 6 and 14°C. For the PBO, the $\delta^{15}N$ and $\delta^{40}Ar$ data can be well reproduced by an increase in 4°C in 20 years or 5°C in 80 years. Considering analytical uncertainties, we propose estimate its temperature increase to be 4°C ± 2.5 °C in 20 to 80 years.

Event	Ice core (age scale)	Start of warming	End of warming	Duration (uncertainty)	Temperature change (uncertainty)	References
End of Younger Dryas	GISP2 (GISP2)	11590	11540	70 (20)*	10(4)*	(147)
Preboreal oscillation	GISP2 (GISP2)	11270		40(20)*	4(1.5)*	(143)
Bolling Allerod					9(3)	(150)
	GISP2(GISP2)	14820	14600	220(20)	16(-)	(148)
DO3	NGRIP(GICC05)	*27720*	*27540*	*180(20)*	_	
DO4	NGRIP(GICC05)	*28920*	*28800*	*120(20)*	_	
DO5	NGRIP(GICC05)	*32540*	*32480*	*60(20)*	_	
DO6	NGRIP(GICC05)	*33900*	*33680*	*220(20)*	_	
DO7	NGRIP(GICC05)	*35520*	*35440*	*80(20)*	_	
DO8	NGRIP(GICC05)	*38240*	*38200*	*40(20)*	*11(3)*	(151)
DO9	NGRIP(GICC05)	*40180*	*40140*	*40(20)*	*9(3)*	(151)
DO10	NGRIP(GICC05)	*41500*	*41440*	*60(20)*	*11.5(3)*	(151)
DO11	NGRIP(GICC05)	*43220*	*43160*	*60(20)*	*15(3)*	(151)
	NGRIP(GICC05)				*12.5(3)*	(151)
DO12	GRIP (GICC05)	*46860*	*46840*	*20(20)*	*12 (2.5)*	(149)
DO13	NGRIP(GICC05)	49120	49020	100(20)	8(3)	(151)
DO14	NGRIP(GICC05)	*54240*	*54200*	*40(20)*	*12(2.5)*	(151)
DO15	NGRIP(GICC05)	55840	55740	100(20)	10(3)	(151)
DO16	NGRIP(GICC05)	*58060*	*58040*	*20(20)*	*9(3)*	(151)
DO17	NGRIP(GICC05)	*59100*	*59060*	*40(20)*	*12(3)*	(151)
DO18	NGRIP(ss09sea)	66383	66207	176(50)	11(2.5)	(152)
	NGRIP(ss09sea)	74582	74405	177(50)	16(2.5)	(152)
DO19	GRIP				16 (-)	(153)
DO20	NGRIP(EDC3)	74336	74149	187(50)	11(2.5)	(152)
DO21	NGRIP(EDC3)	83685	83585	100(50)	12(2.5)	(152)

DO22	NGRIP(EDC3)	89510	89424	86(50)	5(2.5)	(152)
DO23	NGRIP(EDC3)	101981	101852	129(50)	10(2.5)	(152)
DO24	NGRIP(EDC3)	106978	106698	280(50)	16(2.5)	(152)
DO25	NGRIP(EDC3)	112470	112305	165(50)	3(2.5)	(111)

Further Reading/Resources

[Please insert any further reading/resources here]

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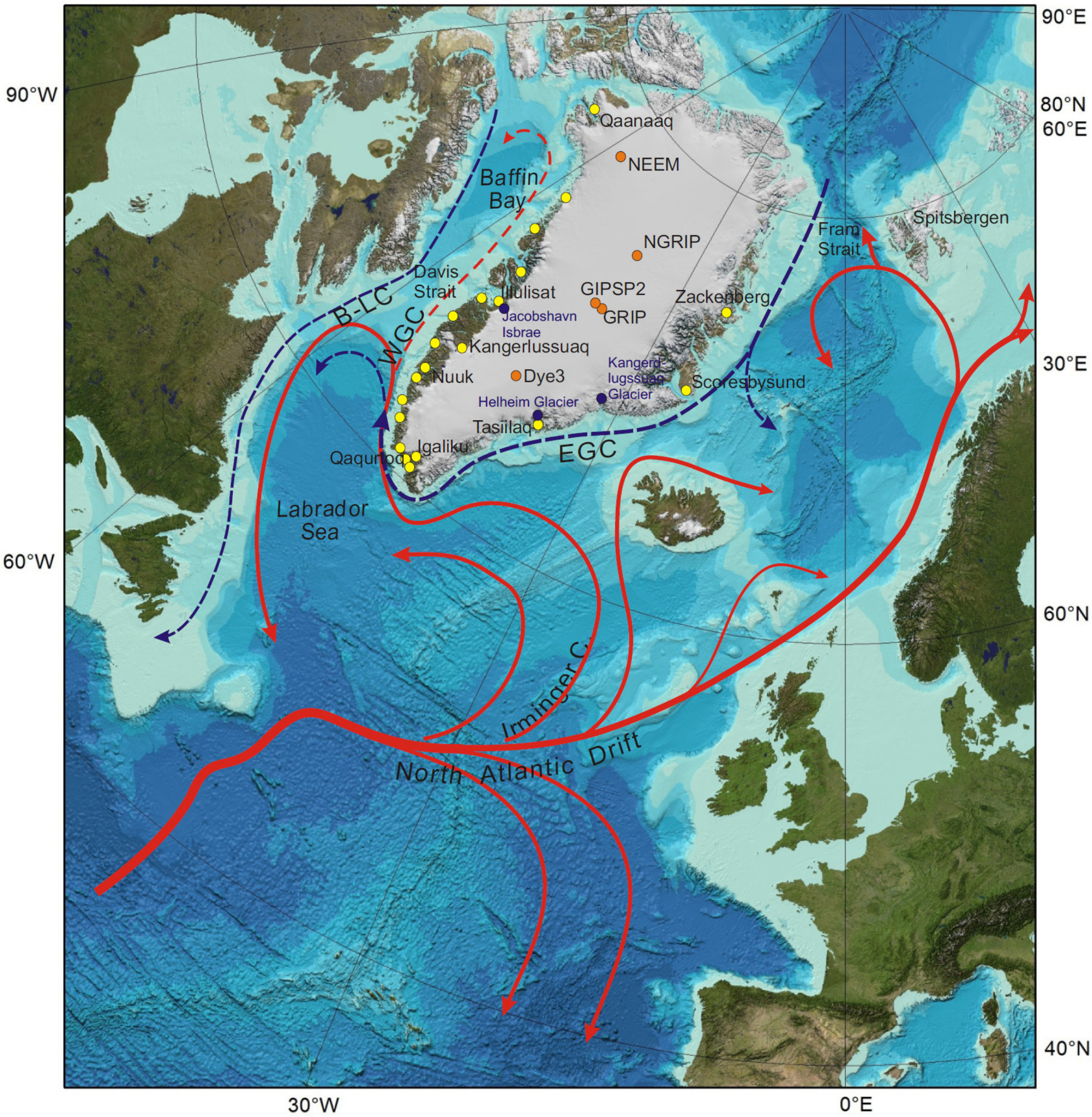
Figure 1.

a) **Map of Greenland** showing the ice sheet extent (white), schematized surface oceanic currents affecting Greenland climate (red arrows, warm surface currents; dashed blue arrows, cold surface currents; EGC: East Greenland Current; WGC: West Greenland Current; B-LC: Baffin-Labrador Current), the largest towns and settlements (yellow circles) as well as ice core drilling sites (orange circles). Adapted from (135) by Martin Jakobsson, Stockholm University.



b)

Figure 1

b) Greening of the Arctic. Satellite observations of Arctic sea ice reduction (indicated by the trend in the percentage of open water) and tundra vegetation productivity (indicated by the MNDVI, modified normalized difference vegetation index). Trends are calculated from 1982 to 2010 using a 10 km resolution, updating earlier data(24).

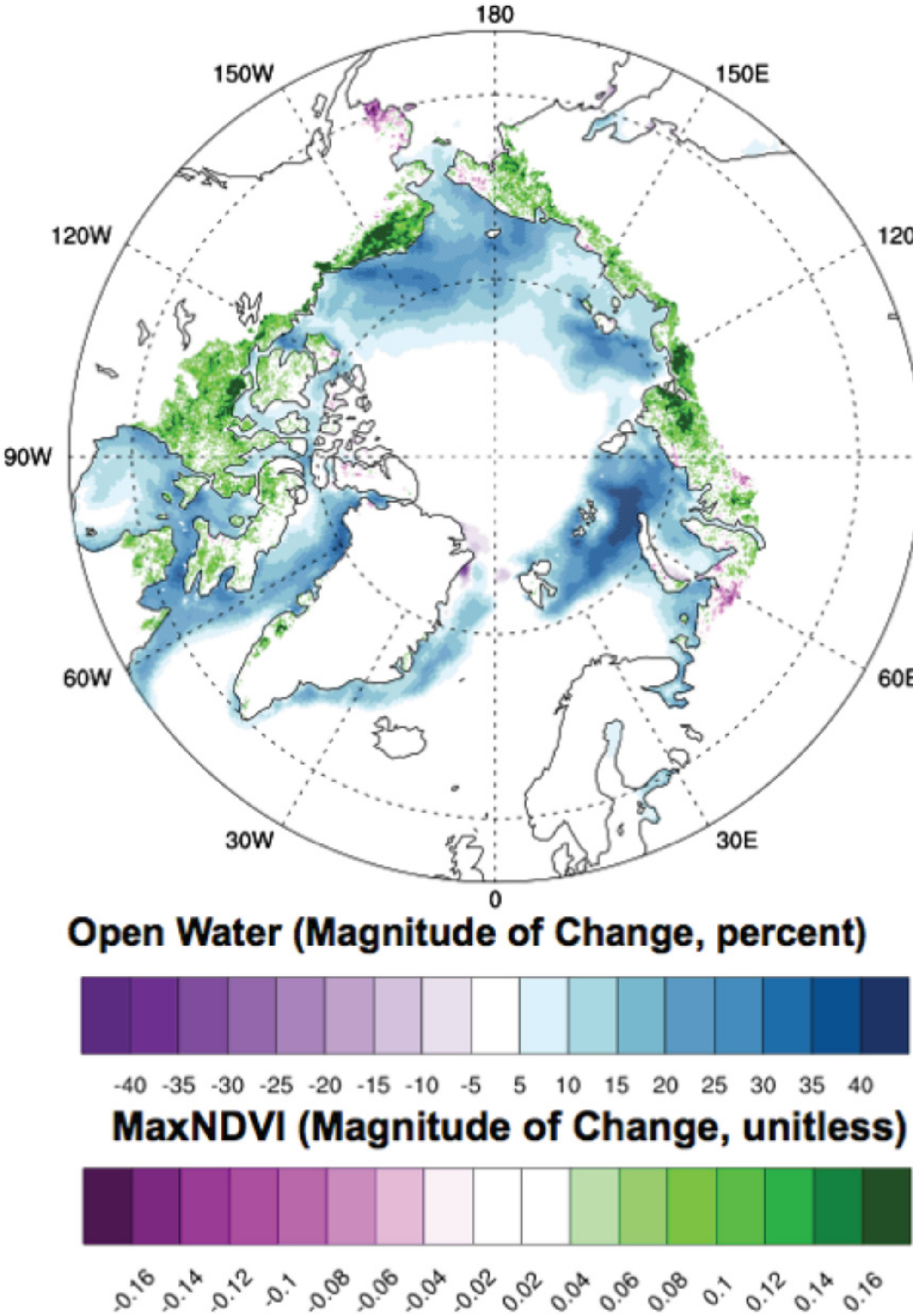


Figure 2. Current Greenland warming in the perspective of natural climate variability and future projections.

a) NorthGRIP ice core $\delta^{18}O$ (‰), a proxy of Greenland SAT(2) at a 20 year resolution (grey) and multi-millennial binomial smoothing (red) as a function of time (years before A.D. 2000); the orbital forcing, which is the main external driver of glacial-interglacial trends, is illustrated by the 70°N June insolation (W/m²). Red areas highlight the interglacial periods and the blue area highlights the last glacial period; the green area indicates the instrumental period.

b) Estimate of southern GrIS(39) SAT anomalies during the current interglacial period (°C, with respect to the last millennium) (grey, 20 year resolution; red, millennial trend) based on a stack of ice cores and a correction for elevation changes(39) and a comparison with the instrumental SAT record from southern Greenland updated to 2010(9) (black, 10 year resolution). The SAT level of the decade 2001-2010 is displayed with a horizontal dashed black line. The 2010 anomaly is displayed as a filled diamond. The vertical rectangles illustrate the succession of human occupations of Greenland, from archeological data (see text). The red area illustrates the current interglacial period, and the green area the instrumental period. The rate of SAT change during the abrupt warming, approximately 8,200 years ago, is also indicated (2.5°C per century).

c) Meteorological records from southern Greenland based on a stack of meteorological data updated to 2010(9) (thin black line, annual data; thick stair steps, decadal averages). The data are compared to the MAR regional climate model results for the south-west Greenland coastal area, forced by ERA-40 (green) and ERA-interim (orange) boundary conditions from 1958 to 2010(8). Data are displayed as anomalies from the 1960-1990 period, which is 0.5°C above the average data for the last millennium as displayed in panel b. The 2010 SAT anomaly is highlighted as a filled diamond. An example projection is given using MAR forced by the ECHAM5 A1B projections (red line, annual values; red stair steps, decadal values). This corresponds to a warming trend of 4.7°C per century.

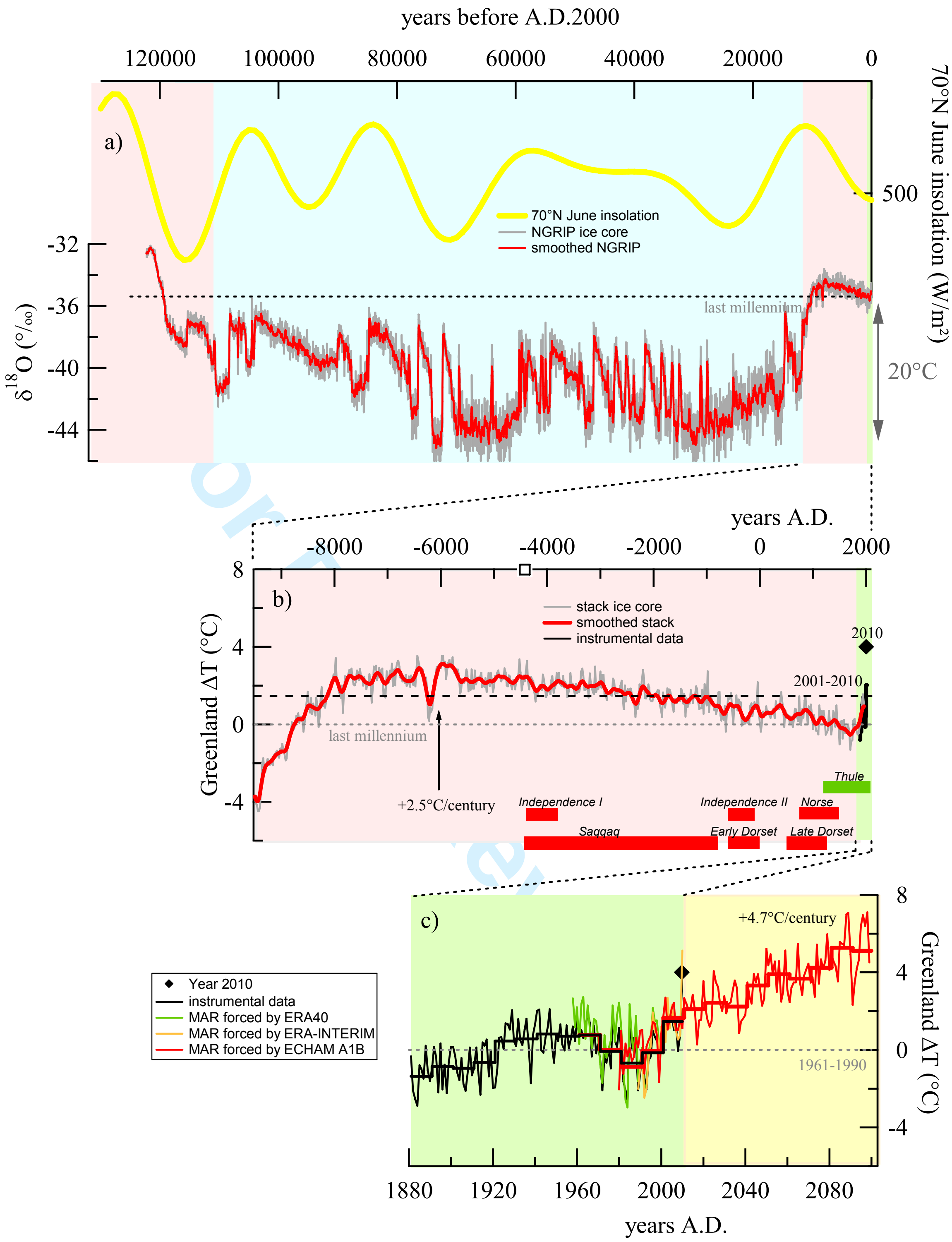

years before A.D.2000
120000
100000
80000
60000
40000
20000
0
70°N June insolation (W/m2)
500
a)
70°N June insolation
NGRIP ice core
smoothed NGRIP
-32
-36
-40
-44
δ18O (°/oo)
last millennium
20°C
years A.D.
-8000
-6000
-4000
-2000
0
2000
b)
stack ice core
smoothed stack
instrumental data
2010
2001-2010
last millennium
+2.5°C/century
Greenland ΔT (°C)
8
4
0
-4
Thule
Independence I
Independence II
Norse
Saqqaq
Early Dorset
Late Dorset
c)
+4.7°C/century
Year 2010
instrumental data
MAR forced by ERA40
MAR forced by ERA-INTERIM
MAR forced by ECHAM A1B
1961-1990
1880
1920
1960
2000
2040
2080
years A.D.
Greenland ΔT (°C)

Figure 2

Figure 3. Cumulative updated(5) anomalies of major mass balance components of the GrIS, 1990-2010, and GRACE gravimetry estimate of mass loss, vertically offset for clarity. Abbreviations are explained in the legend. SMB data from RACMO2 RCM(5). GRACE data courtesy of I. Velicogna and J. Wahr.

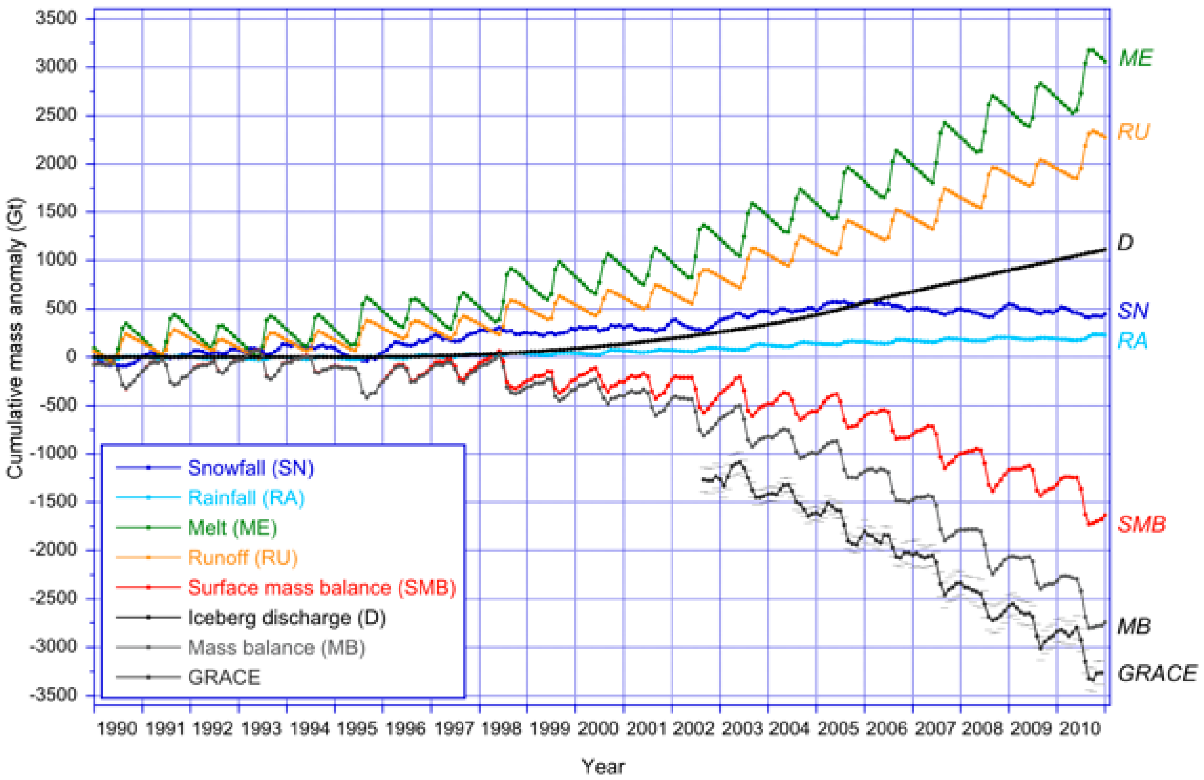


Figure 4. A) Observed and predicted permafrost degradation in Zackenberg 1900-2080 based on down-scaled HIRHAM RCM data. Projections are given for two vegetation types: wetland (brown), heath (green) and two scenarios: a 2°C global warming over 100 years (filled symbols) and 2.4 °C over 60 years (open symbols). Running means over 10 years are shown as solid lines. B) Active layer and permafrost total soil organic carbon and C) Ammonium concentrations in melt water(68).

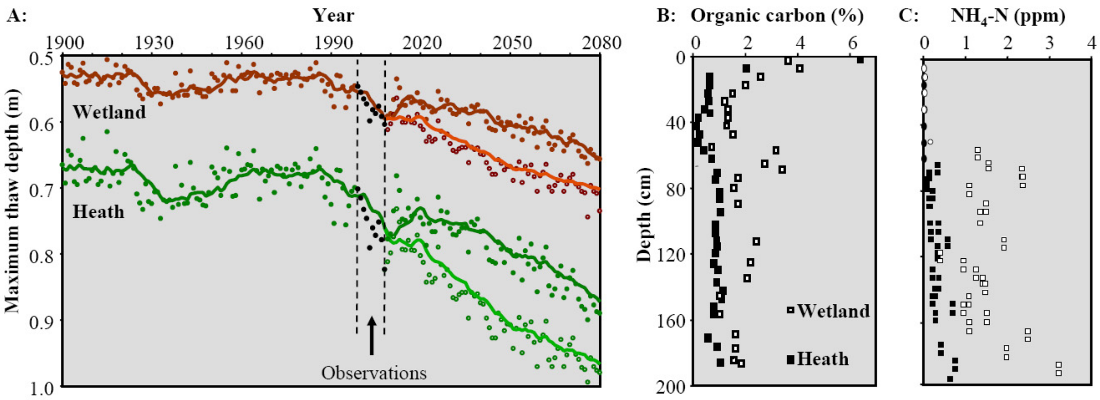


Figure 5. Illustration of the impact of a large GrIS meltwater flux (>0.1 Sv) on global climate projections using the IPSL CM4 model(4). SAT (top) and precipitation (bottom) changes for $2\times CO_2$ (averaged over years 450-500)(136) with respect to the preindustrial control simulation when including (right) or not (left) the impact of GrIS meltwater flux. A strong reduction in the AMOC induces a reduced warming in the north Atlantic but enhanced warming in the southern hemisphere tropical Atlantic, resulting in a southward shift of the Inter tropical Convergence Zone. Such a migration may have strong impacts on tropical precipitation distributions. This type of behavior has been found in a multi-model ensemble for modern conditions and appears to be robust under global warming conditions(125).

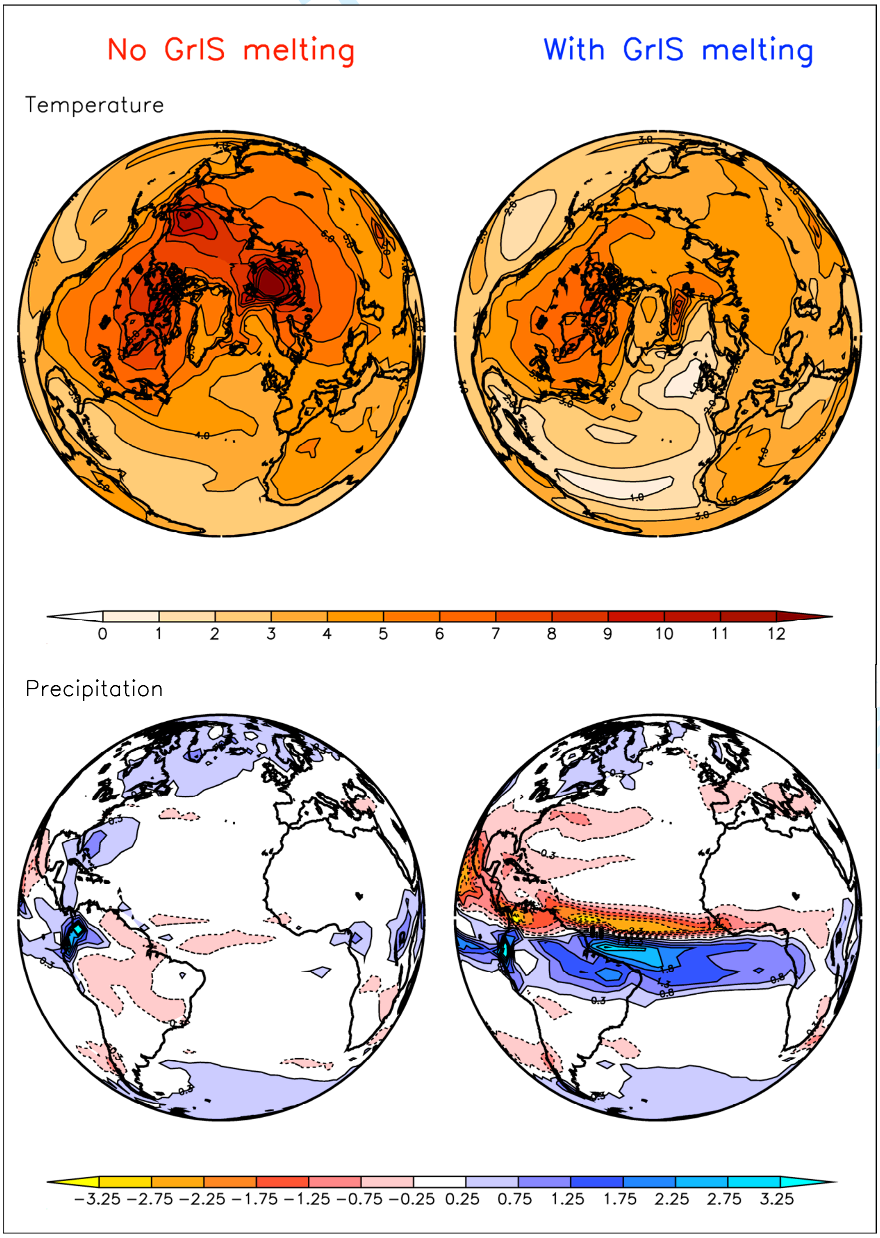
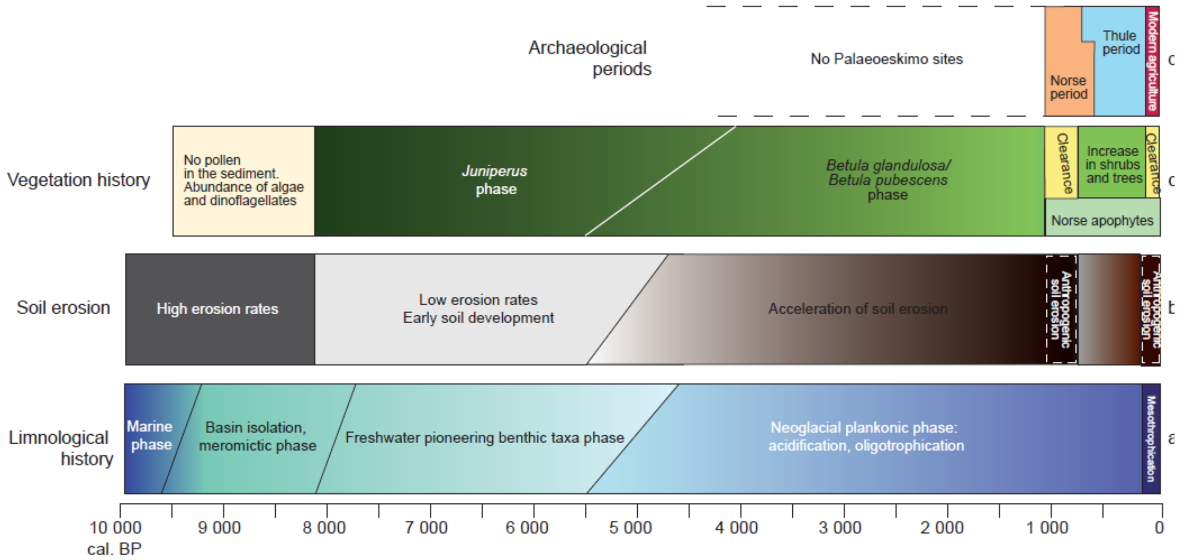


Figure 6. Schematic representation of environmental changes recorded by the Igaliku lake sediments (81-82, 92): a) water quality estimated from diatom assemblages), b) soil erosion rates estimated from the minerogenic and organic inputs into the lake and controlled by a set of geophysical, geochemical and ecological parameters including magnetic susceptibility, titanium content, bulk organic matter geochemistry and diatom valve concentration, c) vegetation history from pollen and non-pollen palynomorphs analyses, and d) archeological periods. Limited impacts of Norse agriculture are reflected by indicators of clearance and sheep grazing, as well as by the persistence of introduced species. Modern agriculture is marked by clearance, soil erosion, and the onset of the first mesothropic phase of the last 10,000 years; e) Photograph of Norse apophytes (*Rumex acetosa* - *Taraxacum* sp) on a medieval archeological site in south Greenland (photograph: E. Gauthier, 2007).



e)

Figure 7. a) Probabilistic estimate of the rate of SAT change over the course of stadial-interstadial events, with a duration longer than 60 years. Data are represented as a probability density function (%) as a function of the rate of SAT change (°C per 100 years), calculated from the published uncertainties on event duration and magnitude (See Table 1). Color codes reflect the CO_2 concentration (as an indicator of the back ground climate) during events (from blue, concentrations between 200 and 215 ppmv, orange, 220 to 230 ppmv, brown, 230 to 240 ppmv and red, 240-260 ppmv). The black line displays the mean probability density, calculated from the 11 studied events). There is a tendency for having slower rates of temperature rise (DO20, DO22, DO23, DO25, BA) under "warm climate" background. DO 22 appears to be very close to a "mean" event.

b) Rates of changes for future climate in RCP4.5 and RCP8.5 projections. Simulations from 13 models or model versions have been considered (NorESM1-M, MRI-CGCM3, MPI-ESM-LR, MIROC-ESM, MIROC-ESM-CHEM, MIROC, IPSL-CM5A-LR, inmcm4, HadGEM2-ES, CSIRO-Mk3, CNRM-CM5, CCSM4, CanESM2, HadGEM2-ES). Results are displayed in terms of cumulative frequencies within the 13 models.

Figure 7a)

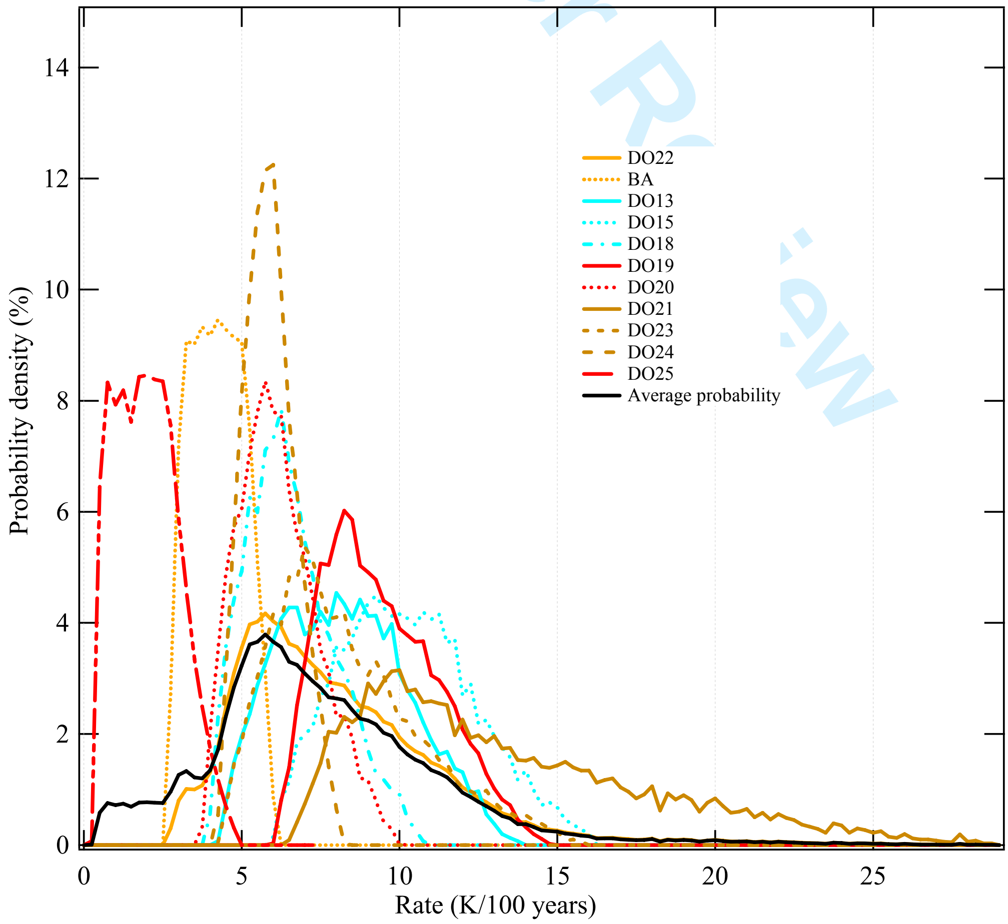
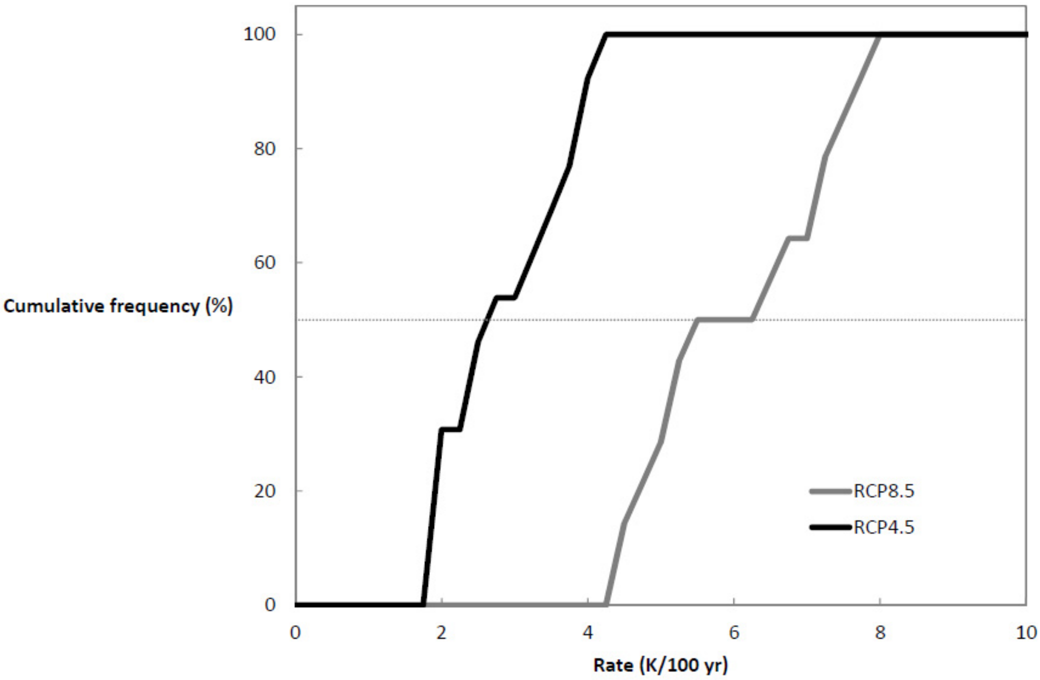


Figure 7b)


100
80
60
40
20
0
Cumulative frequency (%)
0
2
4
6
8
10
Rate (K/100 yr)
RCP8.5
RCP4.5