Advanced Review

Greenland climate change: from the past to the future

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Climate archives available from deep sea and marine shelf 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, during the last decade (2000s), atmospheric and sea surface temperatures are reaching levels 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 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 recent decades. Furthermore, past and present changes in atmospheric and oceanic heat advection appear to strongly influence both regional climate and ice sheet dynamics. Projections from climate models are investigated to quantify 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 last time encountered during the last interglacial period, approximately 125,000 years ago. We review and discuss whether analogies between the last interglacial and future changes are reasonable, 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.

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INTRODUCTION

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 approximately 7.2 m of global sea-level rise.[1] Concerns for future sea-level rise have grown with accelerating GrIS mass loss due to enhanced ice melting and discharge.[2] This meltwater could have strong local and global implications, as the oceanic Atlantic Meridional Overturning Circulation (AMOC) (associated surface currents are displayed in Figure 1) is highly sensitive to freshwater releases in the North Atlantic, with potential global climate implications.[3] Observations as well as regional climate models (RCMs) specifically developed for Greenland, show a strong recent decline in the GrIS surface mass balance.[a, 4–10]

Despite its harsh Arctic environmental conditions, inhabited Greenland coastal climate has been monitored since the 18th century.[13] 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. The GrIS provides exceptional archives of past changes in regional climate and atmospheric composition, as unveiled by deep ice-core records.[14] In parallel, paleoclimate studies based on marine and terrestrial archives have provided a wealth of climate and environmental information.[15]

The Greenlandic population of approximately 56,000 inhabitants[16] mainly lives in towns and settlements along the narrow ice-free south-western coastal margins. About 88% are Inuit, while the rest primarily are Scandinavian (Danish) in origin. Several waves of Paleo-Eskimo cultures have ventured to Greenland from Canada[17] during the past 4500 years,[18] each culture disappearing after several centuries (Figure 3(b)). Migrating from Alaska, the Thule people, ancestors of the current Greenlandic population, arrived in Greenland at the beginning of the 12th century.[19] In the late 10th century, southwest (SW) Greenland was colonized by the Norse. They established approximately 500 farms in the 'green' inner fjords, reaching a maximum population of 2000–3000 people,[20] but disappeared as a community in the late 15th century. These migrations of peoples may have been related to past climate variability.[21,22]

Today, the Greenlandic economy relies heavily 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 2). 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.[24] Aiming at developing sustained 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.[25]

In coming centuries, deglaciation and further greening (in the sense of enhanced biological productivity) of Greenland may drive a progressive shift from a largely marine (Box 1) 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 want to explore the magnitude of projected Greenland physical environment changes as well as their potential local to global impacts, by comparing possible future rates of changes with past changes including the most 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.[26] The retreat of Arctic sea ice[27] (Figure 2) plays a crucial role for this polar amplification.[27,28] Recent Arctic warming has been attributed to the impact of anthropogenic greenhouse gas emissions on climate.[29]

At intra and interannual time scales, the variability of Greenland SAT and precipitation is largely driven by atmospheric heat advection, related to the North Atlantic Oscillation (NAO),[13,30] a large-scale atmospheric mode of variability closely related to the Northern Annular Mode or the Arctic Oscillation[31] and to North Atlantic atmospheric blocking frequency.[32] Although summer SAT is affected by NAO,[33] the magnitude of winter NAO variability causes an interannual winter SAT variability that 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.[12]

Greenland meteorological data reveal a sharp SAT rise starting in 1993, with 2001–2010 being

BOX 1

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.[1,26] 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.[116,117] 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.[118] 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.[25,118]

Living from ice fishing and hunting, some early Greenlandic cultures (e.g., Dorset) depended 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.[18,21] The demise of the Saqqaq culture coincided with a reduced inflow of warmer Atlantic source waters to the coastal regions of West Greenland,[69] limiting the availability of, e.g., harp seals. Colder conditions and changes in ringed seal hunting may also have influenced the disappearance of the Dorset from Greenland.[119]

the warmest decade since the onset of meteorological measurements, in the 1780s, surpassing the generally warm 1920s–1930s by 0.2°C.[34,35] 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[34] and associated with a record melt over the GrIS.[5] We note that it occurred in connection with a very negative NAO during 2010 and 2011, as warm North Atlantic and Arctic conditions damped the impact of this record low NAO on European winters,[36] but enhanced Greenland warming in 2010.

Changes in volcanic or solar activity may also affect the NAO.[37,38] Warm decades in the Arctic and in Greenland occurred during periods with little volcanic forcing (1920s–1930s, 2000s–2010s), whereas cold years marked by reduced summer melt and runoff (e.g., 1983, 1992) followed large volcanic eruptions.[30,35,39]

Greenland coastal climate is also controlled by changes in ocean heat advection, at decadal and longer timescales.[40] Today, Greenland coastal regions are influenced by waters of both polar and Atlantic origins (Figure 1). Depending on the strength of the Irminger Current (Figure 1), warm Atlantic waters may be found as far north as the northern Baffin Bay.[41] During the last two decades, sea surface temperatures (SST) in the SW sector of Greenland have risen by approximately 0.5°C in winter and approximately 1°C in summer,[42] as the influx of Irminger Sea Water has increased.

The NAO affects westerly winds, the Atlantic subpolar gyre and the inflow of the Irminger Sea Waters toward SW Greenland.[43] The enhanced ocean advection may be explained through the combined effect of NAO and a positive phase of the Atlantic Multi-decadal Oscillation (AMO).[44] The AMO is a 55–70 year cyclicity in Atlantic SST presumably related to internal ocean variability[43,45] and which has been in a distinct positive phase since the mid 1990s,[44,45] enhancing northward heat transport in the North Atlantic.[46]

ONGOING GREENLAND TEMPERATURE CHANGES IN THE CONTEXT OF THE CURRENT INTERGLACIAL PERIOD

In order to evaluate if the ongoing warming (with a linear SAT trend of 0.16°C/year from 1993 to 2010) is unusual, we compare them with records of past climate conditions.

Several continuous Greenland SAT reconstructions span the last millennia (Table 1). These reconstructions arise from (1) alkenones from sediments of one West Greenland lake,[21] related to biological late spring–early summer productivity and water temperature, and offering decadal resolution; (2) air nitrogen and argon stable isotopes from one ice core, affected by changes in decadal changes in mean surface snow temperature;[47] (3) water stable isotopes from a stack of ice cores, corrected for changes in

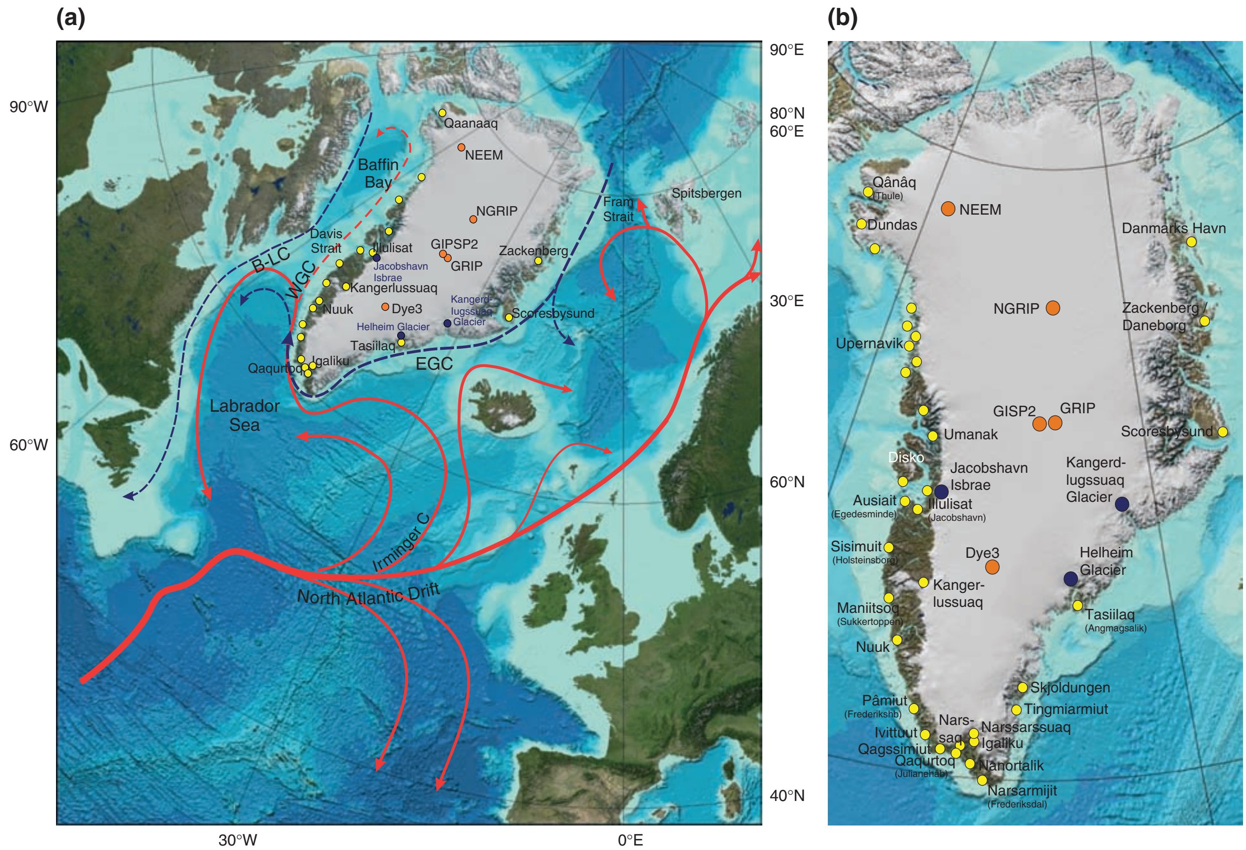

FIGURE 1 | (a) Map of Greenland showing[11] the ice sheet extent (white), schematized surface oceanic currents affecting Greenland climate (red arrows, warm surface currents; dashed blue arrows, cold surface currents; EGC: East Greenland Current; WGC: West Greenland Current; B-LC: Baffin-Labrador Current), the largest towns and settlements (yellow circles) as well as ice core drilling sites (orange circles). (b) Zoom on Greenland.

ice sheet elevation and tuned to SAT using information from borehole temperature records, with seasonal to bidecadal resolution.[23] Different sources of uncertainties may affect each record (Table 1), which show different magnitudes of trends and decadal variability.

The lake record[21] shows a positive SAT anomaly from 4000 to 3000 years BP (before present), large multicentennial 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 multimillennial trend. The bidecadal lake data do not extend into the instrumental period and cannot easily be used to compare with current changes.

The Greenland Summit GISP2 ice core (Figure 1) gas isotope record[47] produces a 1.5°C cooling trend along the last 4000 years, together with multicentennial events (<2°C), and an overall variance of 1.0°C (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 surface snow temperature reconstructed during the 1930s-1940s and during the warmest decades of the medieval period, in the 1140s.[47] 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 does not fully concur with the comparison of coastal SAT changes with respect to the SAT reconstruction based on water stable isotopes from a number of ice cores[23] (Figure 3). This latter reconstruction shares the same multimillennial trend (−0.4°C per 1000 years) as the gas record. However, it differ in the magnitude of the interdecadal variance (0.7°C for water-isotope derived temperature, versus 1°C for gas-isotope derived temperature over the last 4000 years). As a result, very few decades of the last 3000 years surpass the SAT level of the last decade in the isotope-based record.

We note that the results obtained with independent methods may result from different changes in annual mean snow surface temperature (driving the gas isotopes) versus precipitation-weighted condensation temperature (controlling water isotopes)

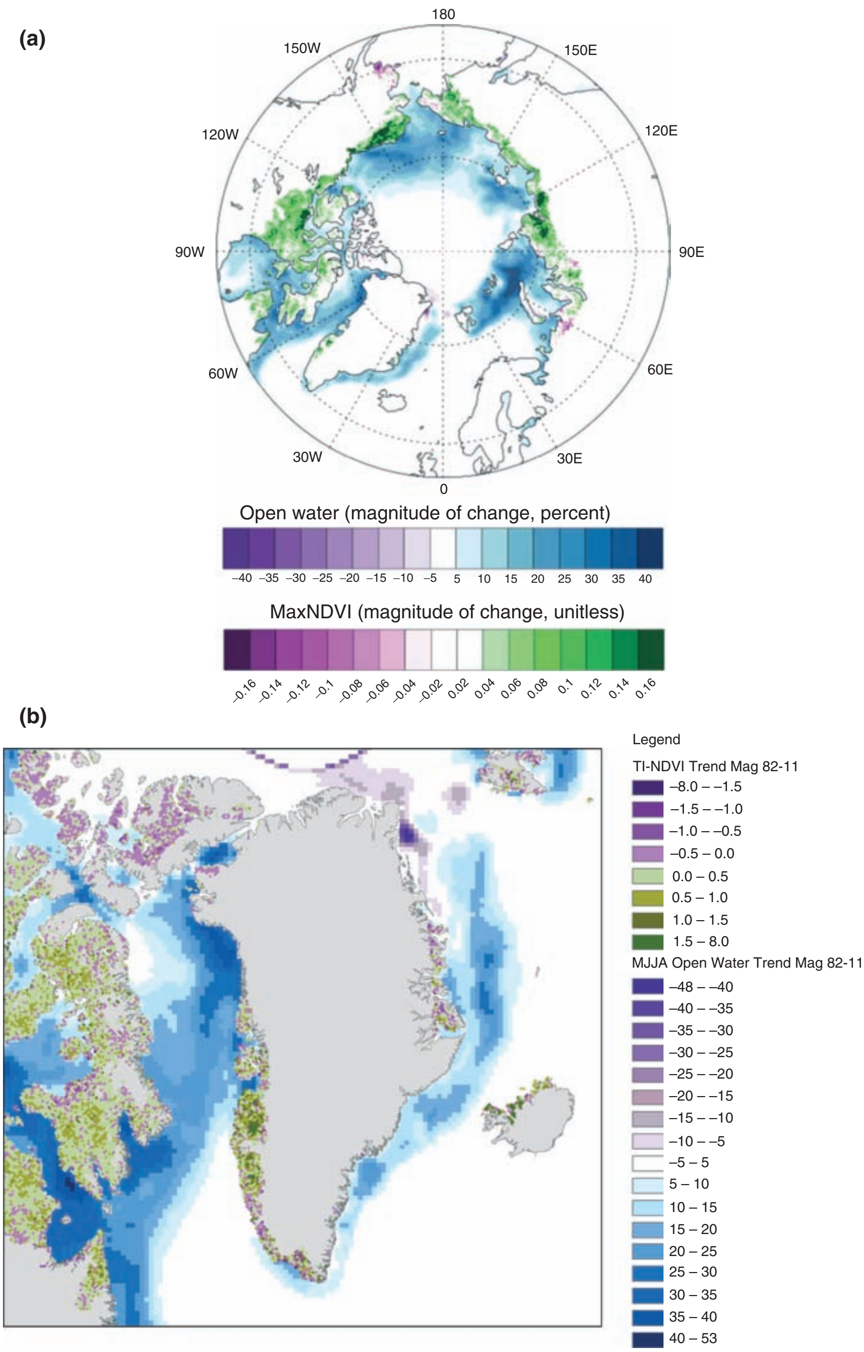

FIGURE 2 | (a) Greening of the Arctic. Satellite observations of Arctic sea ice reduction (indicated by the trend in the percentage of open water) and tundra vegetation productivity (indicated by the MNDVI, modified normalized difference vegetation index). Trends are calculated from 1982 to 2010 using a 10 km resolution, updating earlier data.[12] (b) Zoom on Greenland.

(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; part of this larger variance may be due to analytical uncertainties. 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

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)[23] Precipitation weighted, condensation temperature controlling atmospheric distillation	Several ice cores (DYE3, GRIP, GISP2, NGRIP) spanning the Holocene (seasonal resolution)[48] the last glacial period (annual to decadal resolution)[49] One ice core (NGRIP) with a continuous record back to the last interglacial (123 ka) (20 year resolution)[14,50]	At high frequency (season) : signal to noise ratio caused by deposition and post-deposition processes[51] Intermittency of precipitation (seasonality)[52] Changes in evaporation conditions[53,54] Changes in ice sheet elevation[55]
Ice cores	Air isotopes ($\delta^{15}N$, $\delta^{40}Ar$)[47,52] 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[52] One continuous record spanning the last 4 000 years with decadal resolution[47]	Variability of air isotopic composition during pore close-off and analytical accuracy Storage effect or fractionation associated with clathrate formation[56] Uncertainty in accumulation rate Uncertainty in thermal fractionation coefficients Increments used to model temperature impacts Changes in ice sheet elevation[55]
Ice cores	Inversion of borehole temperature profiles[57,58]	Low frequency variations with a loss of resolution back in time. Detection of decadal variations (last century), multicentennial variations (last millennium), millennial variations (current interglacial) and glacial-interglacial magnitude.	A priori hypothesis on temporal temperature profiles Influence of changes in accumulation Changes in ice sheet elevation[55]
Lake sediments	Alkenone undersaturation in two Greenland lake sediments[21]	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)

hemisphere summer insolation[59] (Figure 3(a)). Water isotope-based dataset scaled to coastal SAT (Figure 3) indicates that the current coastal SAT (last decade) reaches levels comparable to the mean SAT of the mid-Holocene, 4000–6000 years ago, which coincided with the first documented human settlements in Greenland (Figure 3(b)).

Similarly, long-term trends are documented 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.[27] Off NE Greenland, there is growing evidence for a minimum multi-year Arctic sea ice cover approximately 8500–6000 years ago, possibly in response to the strong summer insolation forcing[27,60] (Figure 3(a)). 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 multimillennial trend, reaching levels (in the 2000s) far beyond those of the last 1450 years[61] and last encountered in NE Greenland about 4000 years ago at least.[60] 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.[27,60,62]

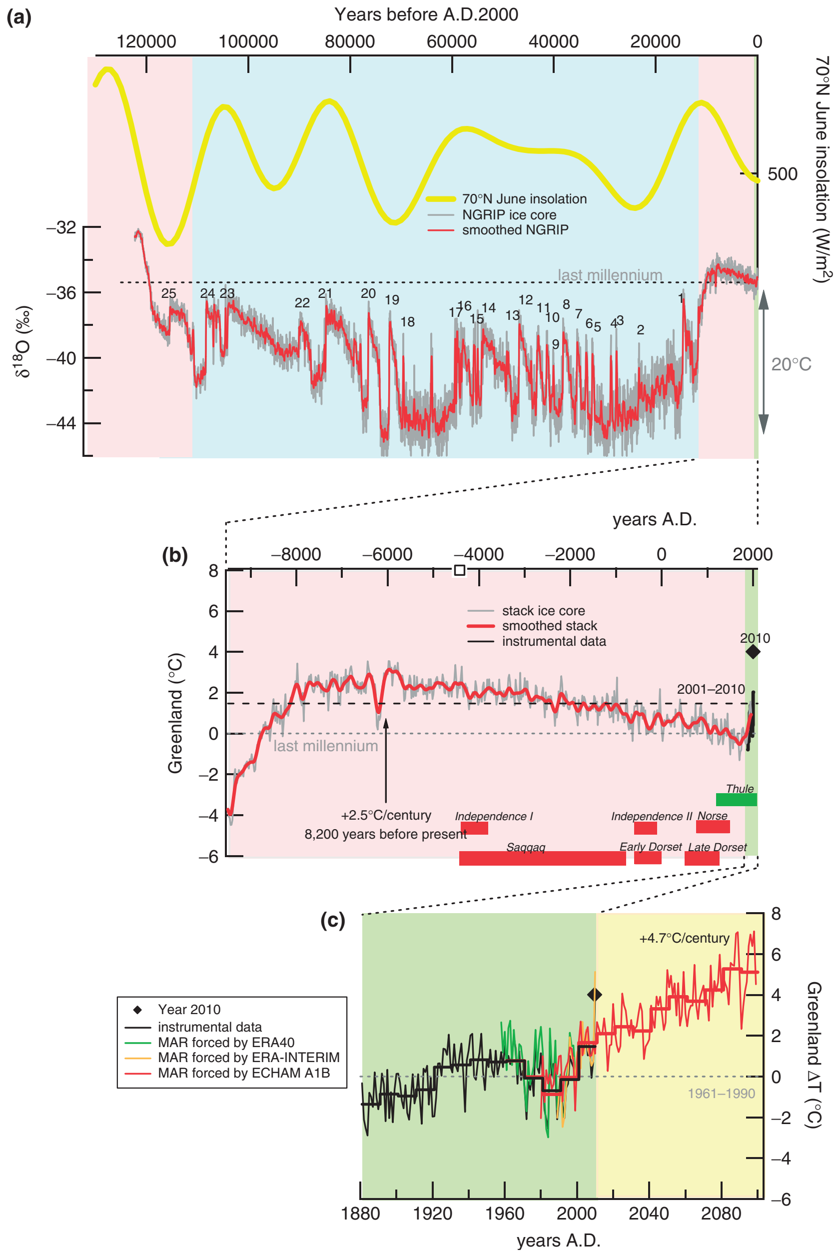

FIGURE 3 | Current Greenland warming in the perspective of natural climate variability and future projections. (a) NorthGRIP ice core $\delta^{18}O$ (‰), a proxy of Greenland SAT[14] at a 20 year resolution (grey) and multi-millennial binomial smoothing (red) as a function of time (years before 2000 AD); the orbital forcing, which is the main external driver of glacial–interglacial trends, is illustrated by the 70°N June insolation (W/m^2). Red areas highlight the interglacial periods and the blue area highlights the last glacial period; the green area indicates the instrumental period. The 25 Dansgaard–Oeschger events are numbered. (b) Estimate of southern GrIS[23] SAT anomalies during the current interglacial period (°C, with respect to the last millennium) (gray, 20 year resolution; red, millennial trend) based on a stack of ice cores and a correction for elevation changes[23] and a comparison with the instrumental SAT record from southern Greenland updated to 2010[13] (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 8200 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[13] (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.[7] 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.

Paleoceanographic records allow to explore the links between past Greenland temperature and ocean advection. 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 approximately 2000 years.[63] The influx of warm Atlantic subsurface water toward SE and W Greenland has also strengthened in recent years,[64–66] 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,[67–70] possibly in relationship with NAO changes.[69,70] There is evidence that, during the current interglacial, the inflow of warm subsurface water masses enhanced iceberg calving and discharge.[71,72]

IMPACTS OF CLIMATE CHANGE ON GREENLAND GLACIERS AND ICE SHEET

The current atmospheric and oceanic warming has large impacts on the approximately 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.[2,73] This increase in solid ice discharge has accounted for about 50% of recent GrIS mass loss.[4] Despite uncertainties in the chronologies, moraine records demonstrate that the onset of modern glacier retreat[74] occurred between the middle of the 19th and the beginning of the 20th century.[75] 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.[74] The widespread retreat 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.[76–79] 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 BP.[77] The Helheim Glacier (south-east Greenland) currently shows melting rates that presumably surpass those of the past approximately 4000 years.[79]

From 1990 to 2010, the GrIS has lost approximately 2750 Gt (Gigatons) of ice, with a significant acceleration in the rate of mass loss[2] (Figure 4). 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.[4] Since

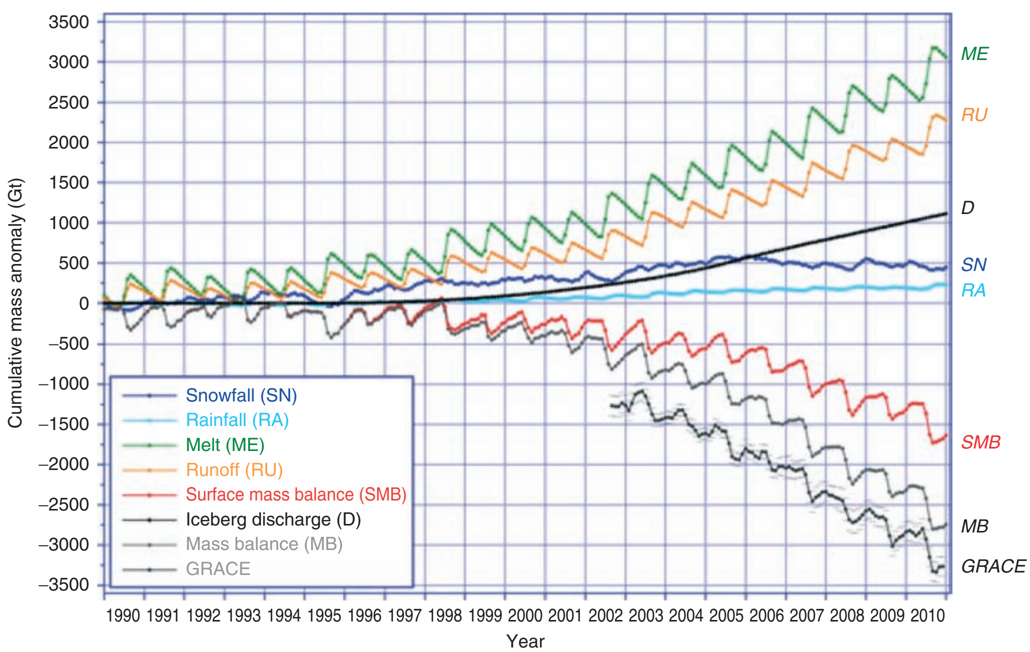

FIGURE 4 | Cumulative updated[4] anomalies of major mass balance components of the GrIS, 1990–2010, and GRACE gravimetry estimate of mass loss, vertically offset for clarity. Abbreviations are explained in the legend. SMB data from RACMO2 RCM.[4] GRACE data courtesy of I. Velicogna and J. Wahr.

about 2000 AD, accelerating summer melt and iceberg discharges are not compensated by refreezing or enhanced accumulation, and in 2010, record summer surface melt led to a GrIS total mass loss of 500 Gt (~1.4 mm of global sea level rise)[5] (Figure 4).

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 (1) to a weakening of the lateral margins of fast flowing glaciers by filling the crevasses,[80] and (2) penetrate the ice sheet through crevasses and moulins, increasing basal lubrication and enhancing basal sliding of the ice over its bedrock.[81] The relationship between water supply and ice-flow velocities is, however, not linear. With sufficient water supply and basal water pressure above a threshold, an efficient drainage system can develop by opening channels, resulting in reduced basal lubrication and thereby limiting basal sliding.[81,82] For land terminating glaciers, this effect is responsible for the observed diurnal and seasonal variations of velocities.[83] 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[73]: 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[78,84] (Figure 1). The retreat of the calving fronts, likely triggered by enhanced basal melting, reduces this backforce and induces an acceleration and a subsequent thinning of the glaciers.[73] 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.[65]

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.[64] Subsurface warm Atlantic waters enter Greenland's fjords to replace the out-flowing surface glacier meltwater.[85] A direct pathway connects the North Atlantic open ocean with southeast Greenland glacier fjords,[66] suggesting that a change in the prevailing water masses in the North Atlantic may impact the GrIS margins within one year.[64,66] There is also evidence of changes in ocean currents influencing glacier melting and iceberg production through the last few thousand years.[71,72]

PRESENT AND FUTURE CHANGES IN GREENLAND PERMAFROST

Retreating sea ice, glaciers, snow cover,[1,26] and warmer coastal conditions affect all Arctic soil ecosystems with underlying permafrost, representing approximately 25% of the northern hemisphere land area and containing almost half of the global soil carbon.[86] Observations of northwest Greenland soil organic carbon suggest that such carbon reservoirs may be underestimated by at least a factor of five.[87] On a global scale, soil-permafrost ecosystems are subject to dramatic changes including glacial retreat, coastal erosion and permafrost thawing.[88]

At the Zackenberg research station, Northeast Greenland, the maximum thickness of the active layer has increased by approximately 1 cm/year since 1996,[89] as a result of increasing SAT, changes in snow cover and an earlier start of the growing season (Figure 5).[90] The spatial variability and timing of actual permafrost warming and thawing is only recently being addressed for Greenland,[91,92] and therefore cannot be placed in a longer perspective.

A critical uncertainty is the heat production from increased microbial metabolism in soils and the accelerated decomposition.[93] This has been shown to be significant in Greenlandic organic-rich soils[89] and has implications for future permafrost degradation rates.[90]

Greenland warming also impacts the terrestrial carbon and nitrogen balance, with interplays between microtopography, biota, hydrology, and permafrost.[89,94] 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.[95,96] Permafrost thawing also has impacts on waste piles (kitchen midden),[97] houses and infrastructures in settled areas.

Projections for the active layer and permafrost thawing in Greenland are few, but is has been suggested that permafrost degradation in high Arctic tundra areas in Greenland may reach approximately 10–35 cm over the next 70 years (Figure 5(a)) and even higher in dry and more coarse-grained sediments.[90,91] As a result, increasing permafrost thawing may in the future contribute with a CO_2 production equivalent to 50% of the present soil respiration.[90] However, the potential compensation by plant carbon fixation remains uncertain. Permafrost degradation is nevertheless 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.[89] Permafrost layers may also be markedly

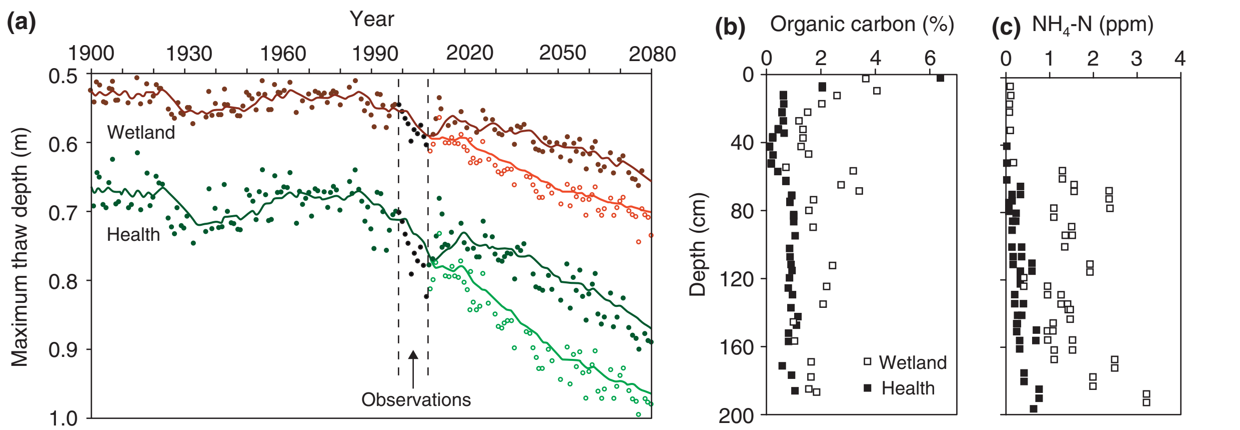

FIGURE 5 | (a) Observed and projected permafrost degradation in Zackenberg 1900–2080 based on down-scaled climate model (HIRHAM RCM) data. Projections are given for two vegetation types: wetland (brown), heath (green), and two scenarios: a 2°C global warming over 100 years (filled symbols) and 2.4°C over 60 years (open symbols). Running means over 10 years are shown as solid lines. (b) Active layer and permafrost total soil organic carbon observed for two vegetation types, wetlands (open symbols) and heath (filled symbols),[89] and (c) Ammonium concentrations in melt water, for two vegetation types, wetlands (open symbols) and heath (filled symbols).[89]

richer than the active layer with respect to nitrogen (Figure 5(c)). Thawing permafrost layers may therefore enhance the potential for a greening of Greenland in a warmer climate, and future changes in permafrost could have large impacts on coastal erosion, the carbon budget, vegetation, and infrastructures.

PAST CHANGES IN GREENLAND VEGETATION: IMPACTS OF CLIMATE AND AGRICULTURE

The recent growth of agriculture in Greenland represents the second attempt to introduce such activities. During the medieval period (986–1450), Norse farmers have settled South Greenland, developing livestock farming. Beyond recent changes documented from historical archives, sedimentary records provide information on the past natural variability of Greenlandic vegetation. During earlier interglacial periods, high pollen influx and specific pollen assemblages from marine sediments depict dense vegetation mostly composed of shrubs and/or conifer trees.[98] During the very long interglacial stage occurring about 400,000 years ago (Marine Isotopic Stage 11),[99] a spectacular development of spruce forest was very likely associated with a strong ice sheet retreat.[98,99]

The warmer conditions encountered about 8000 years ago (Figure 3(b)) left imprints in South Greenland lake sediments, in which pollen assemblages—an open tundra with Juniper—reflect dry conditions in the early Holocene.[100,101] Increasing moisture and soil development in the mid-Holocene allowed the development of dwarf birch (*Betula glandulosa*) and white birch (*Betula pubescens*)[100] (Figure 6(a), A–C). The cooling trend of the last millennia was associated with a fall in pollen fluxes, about 2000 years ago.[100]

A millennium ago, the Norse colonists lived as pastoral farmers, fishermen, and hunters. Changes in precipitation and wind regime may have influenced their agriculture.[22,104] However, archaeological evidence indicates that the Norse in fact adapted very well to new conditions and that the dependence on the marine mammals increased[105] when the climate deteriorated and made herding and pastoral farming more and more difficult.[106] Paleoecological records support these archaeological data.[107–110] For instance, analyses of Lake Igaliku sediments, near the Norse *Garðar,* show that that Norse agropastoralism induced landscape modifications, causing an increase in the non-indigenous plant taxa (e.g., *Rumex acetosa/acetosella*), as shown in Figure 6(C,D in a) and (b), at the expense of white birch.[111] Reflecting soil erosion, the sediment flux also increased sharply, synchronously with vegetation changes, until it reached its maximum at approximately 1180 AD, at more than two times its baseline levels.[100,102] At the beginning of the 14th century, erosion and grazing pressure sharply decreased, suggesting a reduction in the sheep herds at the beginning of 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.[112] In the 12th century,

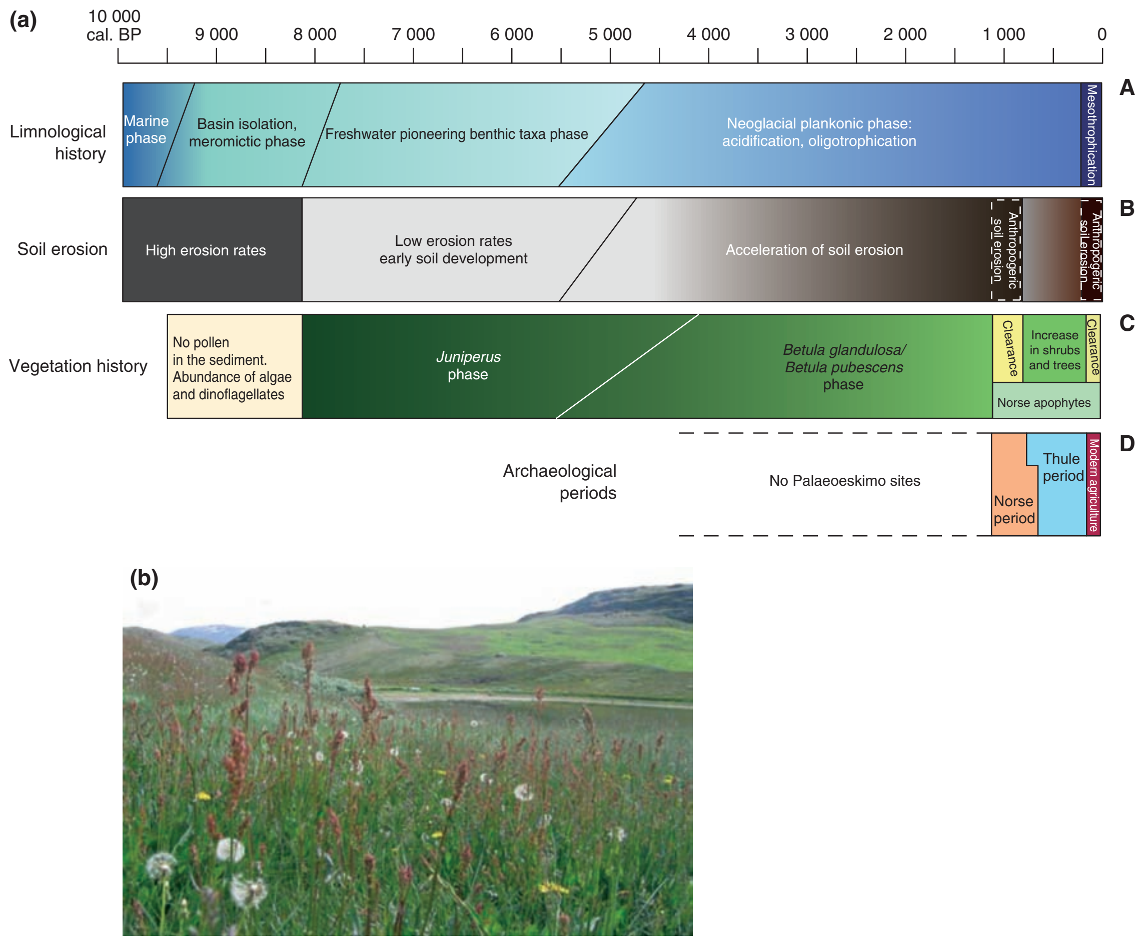

FIGURE 6 | (a) Schematic representation of environmental changes recorded by the Igaliku lake sediments[100,102,103]: (A) water quality estimated from diatom assemblages, (B) soil erosion rates estimated from the minerogenic and organic inputs into the lake and controlled by a set of geophysical, geochemical, and ecological parameters including magnetic susceptibility, titanium content, bulk organic matter geochemistry, and diatom valve concentration, (C) vegetation history from pollen and nonpollen 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; (b) Photograph of Norse apophytes (*Rumex acetosa*—*Taraxacum* sp) on a medieval archeological site in south Greenland (source: E. Gauthier, 2007).

the Inuit[19] 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 also part of today's Greenlandic society, making it responsive and ready to take advantage of the greening of Greenland[25] by expanding agricultural activities.

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. Recent agricultural activities had a much larger impact than four centuries of Norse agriculture. Until 1976, traditional sheep grazing used practices similar to those of the Norse, and sheep were left to graze openly in winter.[24] Pollen and coprophilous fungi spores indicate disturbance levels that parallel those of Norse grazing pressure.[111] However, after dramatic impacts of cold spring conditions in 1966, 1971 and 1975,[24] farming methods switched to winter feeding, more intensive practices of hay production, mechanization, and fertilizer usage. Since 1976 (Figure 6), nitrogen isotopes and diatom microfossils document a marked shift in the lake Igaliku ecosystem consistent with nutrient enrichment from agricultural sources as well as warmer summer SAT.[100,102] Current ecological conditions and soil erosion in the Igaliku region are thus unprecedented in the context of at least the last 1500 years. Given projected Greenland SAT and the anticipated growth of the farming sector, even greater landscape changes must be expected in the future.

CURRENT CHANGES IN GREENLAND VEGETATION

Changes in tundra gross primary production since approximately 1982 have been quantified using combined measurements from different sensors and satellites[12] (Figure 2). Biweekly measurements of Arctic Normalized Difference Vegetation Index (NDVI, calculated from spectral reflectance measurements at 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.[12] The data depict a consistent increase in tundra photosynthetic activity in areas of land warming[12] and sea ice decline (Figure 2). Such trends are detected in SW Greenland, and in areas with retreating glaciers, where rapid vegetation growth occurs on recently exposed landscapes. In the vicinity of Baffin Bay and Davidson Strait, the region of increasing open water conditions in northwest Greenland is characterized by increasing trends in summer land SAT increase and in time-integrated NDVI which are among the most pronounced in the entire Artic realm (Figure 2).

Complex species interactions determine the response of ecosystems to Arctic warming, changes in plant phenology, snow and ice depth, and nutrient availability.[113] In southeast Greenland, a detailed comparison of vegetation taxa[114] 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.

The recent warming has 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.[24] The local production of potatoes (~70 t/year) has been steadily increasing.[115]

PROJECTED FUTURE GREENLAND CLIMATE CHANGES IN THE LIGHT OF PREVIOUS CHANGES

Coupled climate model projections have been analyzed 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 $3.3 \pm 1.3^{\circ}$C.[120,121] Global simulations have recently been refined with RCMs[4,5] 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 3(c)). Projection scenarios were built using RCMs forced by the outputs of ECHAM5 climate model, representative of the average global climate model projections.[121] The calculation based on MAR (Figure 3(c)) 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 the length of the SW Greenland growing season, corresponding to 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 (Figures 1 and 2) leads to similar results.[122]

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 fifth 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 7(b)). These rates of changes can be compared with past natural changes documented by ice cores.

Indeed, past Greenland climate was marked by numerous abrupt climate fluctuations, the most significant being the glacial Dansgaard–Oeschger (DO) events, characterized by an abrupt warming with an amplitude reaching up to 16°C within a few decades to centuries (Table 2), and a more gradual return to colder conditions. These 25 DO events[14] had a global impact[123] including monsoon shifts[124] and variations in atmospheric greenhouse gas concentrations. As suggested by different pieces of information, including the bipolar seesaw,[52,125,126] these instabilities are believed to be linked to changes in AMOC,[127] possibly in response to massive freshwater release from glacial ice sheets.[128] Some rapid events are also documented under interglacial conditions. For instance, the beginning of the current interglacial period is marked by a sub-centennial cooling event, around 8200 years ago, likely caused by the impact on Lake Agassiz on North Atlantic ocean currents,[129] followed by a progressive recovery[130,131] (Figure 3(b)). The last interglacial period may also have been punctuated by cold spells, possibly linked to inputs of ice sheet meltwater.[132–134]

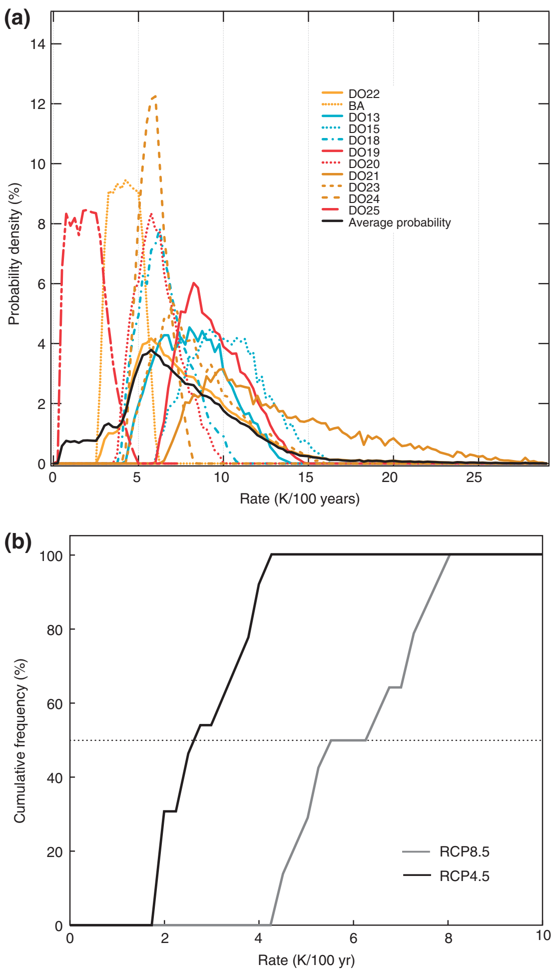

FIGURE 7 | (a) Probabilistic estimate of the rate of SAT change over the course of stadial–interstadial events, with a duration longer than 60 years. Data are represented as a probability density function (%) as a function of the rate of SAT change (°C per 100 years), calculated from the published uncertainties on event duration and magnitude (see Table 2). 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, 220–230 ppmv; brown, 230–240 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, and HadGEM2-ES). Results are displayed in terms of cumulative frequencies within the 13 models.

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 approximately 5°C/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(a)), up to approximately 2.5°C per century during the first DO event, DO25,[50] and the recovery from the cold event, 8200 years ago[131] (Figures 3(b) and 7(a)). 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.[143,144] During the Last Interglacial period (Eemian), ca. 130,000–115,000 years ago, the orbital configuration resulted in strongly enhanced northern hemisphere summer insolation (Figure 3(a)). Paleoclimate data depict large scale Arctic warming,[145,146] with Greenland temperatures approximately 5°C above pre-industrial levels,[14,15,147] reduced sea-ice extent around Greenland.[27,148] Climate models show that the response to 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.[144]

Systematic model-data comparisons for the Last Interglacial period therefore offer the potential to assess the realism of the multicentennial 'equilibrium response' of climate models in a context relevant for the magnitude of future changes. Albeit occurring in a different context, past centennial abrupt events offer a complementary approach to test the 'transient response' of climate models.

PROJECTED FUTURE GREENLAND ICE SHEET AND GLACIER CHANGES

Future climate change is among other areas expected to impact coastal sea ice cover, extreme events, river runoff and its potential for hydroelectricity production.[26] The large impact of external natural forcings and internal variability of the ocean and atmospheric circulations (e.g., AMO and NAO)

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[52]

Event	Ice core (age scale)	Start of warming	End of warming	Duration (uncertainty)	Temperature change (uncertainty)	References
End of Younger Dryas	GISP2 (GISP2)	11,590	11,540	70(20)[1]	10(4)[1]	136
Preboreal oscillation	GISP2 (GISP2)	11,270		40(20)[1]	4(1.5)[1]	56
Bolling Allerod	GISP2(GISP2)	14,820	14,600	220(20)	9(3) 16(–)	139 137
DO3	NGRIP(GICC05)	*27,720*	*27,540*	*180(20)*	—	
DO4	NGRIP(GICC05)	*28,920*	*28,800*	*120(20)*	—	
DO5	NGRIP(GICC05)	*32,540*	*32,480*	*60(20)*	—	
DO6	NGRIP(GICC05)	*33,900*	*33,680*	*220(20)*	—	
DO7	NGRIP(GICC05)	*35,520*	*35,440*	*80(20)*	—	
DO8	NGRIP(GICC05)	*38,240*	*38,200*	*40(20)*	*11(3)*	140
DO9	NGRIP(GICC05)	*40,180*	*40,140*	*40(20)*	*9(3)*	140
DO10	NGRIP(GICC05)	*41,500*	*41,440*	*60(20)*	*11.5(3)*	140
DO11	NGRIP(GICC05)	*43,220*	*43,160*	*60(20)*	*15(3)*	140
DO12	NGRIP(GICC05) GRIP (GICC05)	*46,860*	*46,840*	*20(20)*	*12.5(3)* *12(2.5)*	140 138
DO13	NGRIP(GICC05)	49,120	49,020	100(20)	8(3)	140
DO14	NGRIP(GICC05)	*54,240*	*54,200*	*40(20)*	*12(2.5)*	140
DO15	NGRIP(GICC05)	55,840	55,740	100(20)	10(3)	140
DO16	NGRIP(GICC05)	*58,060*	*58,040*	*20(20)*	*9(3)*	140
DO17	NGRIP(GICC05)	*59,100*	*59,060*	*40(20)*	*12(3)*	140
DO18	NGRIP(ss09sea)	66,383	66,207	176(50)	11(2.5)	141
DO19	NGRIP(ss09sea) GRIP	74,582	74,405	177(50)	16(2.5) 16(—)	141 142
DO20	NGRIP(EDC3)	74,336	74,149	187(50)	11(2.5)	141
DO21	NGRIP(EDC3)	83,685	83,585	100(50)	12(2.5)	141
DO22	NGRIP(EDC3)	89,510	89,424	86(50)	5(2.5)	141
DO23	NGRIP(EDC3)	101,981	101,852	129(50)	10(2.5)	141
DO24	NGRIP(EDC3)	106,978	106,698	280(50)	16(2.5)	141
DO25	NGRIP(EDC3)	112,470	112,305	165(50)	3(2.5)	50

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 characterized events are depicted in italics. GICC05 refers to the most recent Greenland counted age scale.[49,135]

[1]The method used to determine the amplitude of the temperature change at the end of the Younger Dryas (YD)[136] is based on a static firn heat diffusion model with temperature forcing as a step function. The method developed for the Preboreal Oscillation (PBO)[56] 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.[137] 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.[137] forced by different temperature scenario inspired from the ice core $\delta^{18}O$ profile,[138] 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[136]; 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 estimate its temperature increase to be 4 ± 2.5°C in 20–80 years.

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 better detailed.

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,[121] while empirical approaches have been deployed to estimate the potential range of the ice sheet response[149,150] which is starting to be described in new generations of GrIS models.[151] Most studies predict increasing GrIS mass loss, an acceleration of fast flowing glaciers,[152] 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, characterized by a global sea level >6 m higher than today.[153] Large uncertainties remain on the magnitude of Last Interglacial GrIS mass loss, which could have contributed at least 1.5 m of sea level rise, with large uncertainties on the magnitude, location and rates of changes.[154–156] New information from the NEEM ice core data is expected to provide observational constraints on the ice sheet topography changes during the Last Interglacial.[157] 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.[158]

GrIS melt has likely affected AMOC during the Last Interglacial period.[159] During glacial periods, major reorganizations in AMOC associated with DO events may also have been driven by massive meltwater inputs, provided by glacial ice sheet instabilities[52,128] (Figure 3). These past abrupt AMOC changes had well documented global impacts, notably migrations of the inter-tropical convergence zone associated with a cooling of the North Atlantic region,[3,160,161] which in turn influenced regional climate around Greenland. Sensitivity studies have been conducted to investigate the response of AMOC and climate to future GrIS meltwater fluxes, with varying results.[3,162–164] Differences may arise from the prescribed melting rates[165] 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 nonlinear,[166] due to the potential existence of a bifurcation point for the AMOC dynamics identified in simple ocean circulation models.[167] These studies show that the AMOC may significantly weaken for a Greenland melting rate above 0.1 Sv (10^6 m^3/second) in 2100, a pacing not incompatible with estimates of GrIS mass loss acceleration.[2] By limiting the warming around Greenland, a weakened AMOC may act as a negative feedback for the GrIS mass loss, but induce major reorganizations in the tropical Atlantic atmospheric circulation and precipitation distribution (Figure 8). Altogether, both the past and future 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, Greenland climate may be comparable with conditions previously encountered during last interglacial period, which was also marked by a significant (but not complete) GrIS mass loss. We have highlighted 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 case studies against which the ability of climate models to resolve past variations with magnitudes or rates of changes relevant for future changes may be assessed. Some initial 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.[144] 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.[169] However, this conclusion must be taken with caution, due to uncertainties in the initial state of the climate system, and numerical experiment set-up that do not account for all the feedback processes at play such as changes in vegetation and dust. 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, which should be able to address this issue in more details.

Paleoclimate records moreover highlight the large inter-annual, decadal, and 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.[29] The emergence of ensemble multi-millennia transient simulations with climate models opens the possibility to further investigate and possibly 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 still have 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',[25]

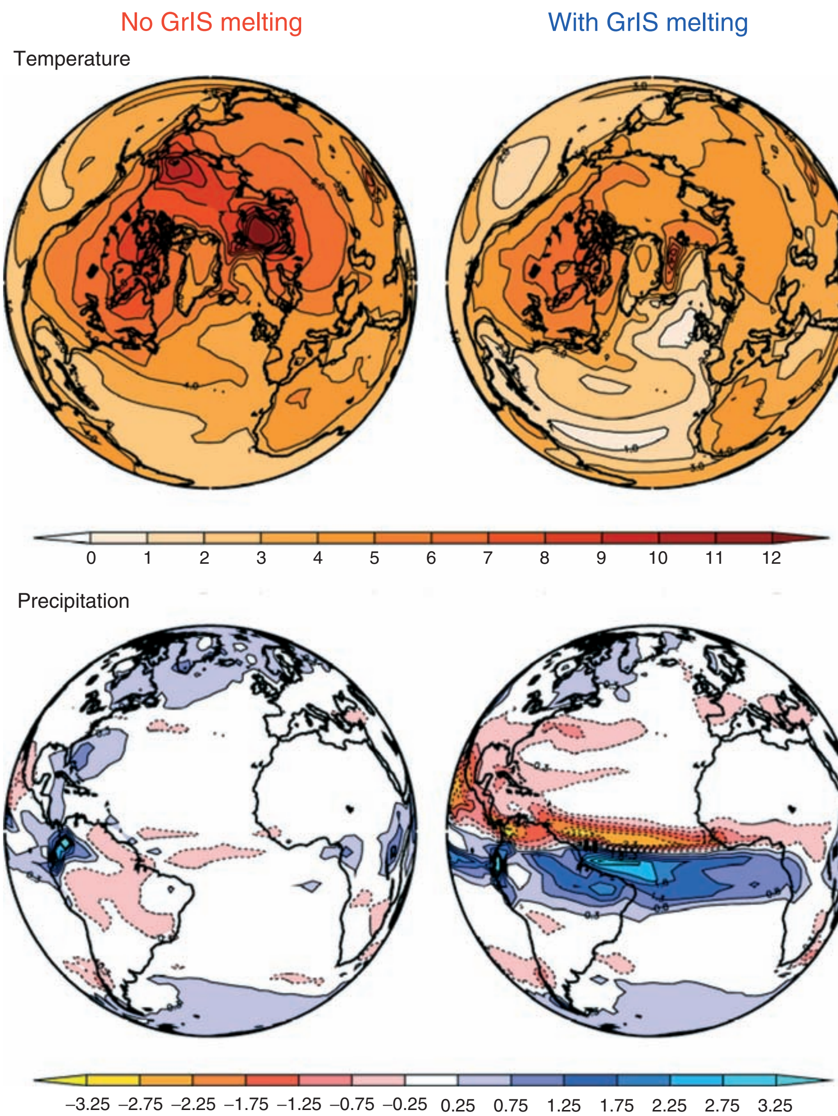

FIGURE 8 | Illustration of the impact of a large GrIS meltwater flux (>0.1 Sv) on global climate projections using the IPSL CM4 model.[3] SAT (top) and precipitation (bottom) changes for 2× CO_2 (averaged over years 450–500)[168] with respect to the preindustrial control simulation when including (right) 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.[161]

Greenlanders face opportunities and threats linked to the deglaciation and greening (enhanced biological productivity) of Greenland. Perception studies[170] and combined use of traditional knowledge and climate model projections are needed to assess the impacts of climate change on coastal areas.

Adaptation to climate change requires improved investigations of local impacts, including changes of Greenland regional climate variability and likelihood of extreme events. 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.[171] Changes in permafrost potentially have large impacts on coastal erosion, the carbon budget, vegetation, and infrastructures. Long term monitoring efforts need to be maintained and expanded. This will assist monitoring of the changes but also enhance capability 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 enhance our understanding of the feedback mechanisms at play and thereby reduce uncertainties in projections. The key processes affecting the GrIS dynamics (impact of surface water production on basal lubrication, and retreat of the calving fronts of floating ice 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.

NOTE

[a]The surface mass balance of an ice sheet is defined as the balance between the mass input by accumulation and the mass loss by ablation due to sublimation and runoff, therefore not taking ice flow and iceberg calving into account.

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