https://doi.org/10.5194/egusphere-2024-1770
Preprint. Discussion started: 7 August 2024

Projected climate change will double the Late Holocene maximum to present ice loss in Central-Western Greenland by 2070

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Greenland's peripheral glaciers and ice caps (GICs) have experienced accelerated mass loss since the 1990s. However, the extent to which present and future trends of GICs are unprecedented within the Holocene is poorly understood. This study bridges the gap between the maximum ice extent (MIE) of the Late Holocene, present and future glacier evolution until 2100 in Eastern Nuussuaq Peninsula (Central-Western Greenland), where the age of moraine boulders was determined by surface exposure dating. The Instructed Glacier Model (IGM) is calibrated and validated by simulating present-day glacier area and ice thickness. The model is employed to reconstruct eastern Nuussuaq Peninsula GICs to align with the MIE of the Late Holocene, which occurred during the late Medieval Warm Period (1130 ± 40 and 925 ± 80 CE). Subsequently, the model is forced with CMIP6 projections for SSP2-4.5 and SSP5-8.5 scenarios (2020-2100). Glaciers reach the MIE of the Late Holocene when temperatures decrease between 0.75ºC and 1ºC relative to the baseline climate period (1960-1990). Currently, glaciers have retreated by 34% compared to the MIE of the Late Holocene. By the end of the 21st century (2100), temperatures are projected rise up to 6ºC (SSP5-8.5) with respect to the baseline climate, exceeding temperatures prevailing during the Holocene Warm Period (~10 to 6 ka) by a factor of three. Using IGM with a positive degree-day model calibrated with geodetic mass balance data from 2000-2020, we project that by >2070 under SSP2-4.5 and SSP5-8.5, glacier mass loss will double (-70%) the loss trend observed from the MIE of the Late Holocene to the present. This work helps contextualize present and future glacier retreat within a geologic time scale and quantify the impacts of anthropogenic climate change on the cryosphere.

Key words: Climate change, glaciology, glacial geomorphology, numerical modelling, deglaciation, Greenland, Arctic.

1. Introduction

Arctic temperatures are rising at a faster (~ 4 times) rate than the global average (IPCC, 2019) and glaciers are accelerating the mass loss (Hugonnet et al., 2021). In 2021, Greenland's peripheral glaciers and ice caps (GICs) represented a small (4%) ice cover area of the island but contributed to 11% of the total Greenland ice loss and sea level rise (Khan et al., 2022). The recession of glaciers implies alterations in fauna and flora

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patterns (Saros et al., 2019), as well as impacts on water availability, climate, and ocean and atmospheric dynamics that have environmental and climate consequences far beyond the polar regions (IPCC, 2022).

The Little Ice Age (LIA; 1300-1900 CE) has been defined as the last period with widespread glacier expansion (Kjær et al., 2022). Greenland GICs have lost 499 Gt of ice from end of LIA to 2021 (Carrivick et al., 2023). The rate of loss of GICs has increased since the 1990s (Bölch et al., 2013; Larocca et al., 2023), with recent trends indicating an acceleration in mass loss from 27.2 ± 6.2 Gt/yr (February 2003–October 2009) to 42.3 ± 6.2 Gt/yr (October 2018–December 2021) (Khan et al., 2022). Warming rates have been higher in West Greenland than in the East since the end of the LIA (Hanna et al., 2012). As a result, the loss of ice from Greenland's GICs has been more pronounced in its western fringe, where warmer conditions have been associated with the positive phase of the North Atlantic Oscillation (NAO), resulting in a West-to-East warming gradient (Bjørk et al., 2018). Historical records reveal varying trends in GICs over the past two centuries. Aerial images and satellite data indicate that GICs in the West Greenland remained relatively stable, maintaining their extent from the mid-19th century until the mid-20th century, after which they experienced rapid retreat (Weidick, 1994; Leclerq et al., 2012). For instance, Citterio et al. (2009) observed a reduction in glacier area of approximately 20% from the LIA to 2001. Other estimates suggest a 48% loss in GICs area in Southern-Western Greenland since the maximum extent of LIA up to 2019 (Brooks et al., 2022).

The recent evolution of the GICs has been reconstructed using historical aerial images and satellite records (Leclerq et al., 2012; Yde and Knudsen, 2007; Citterio et al., 2009; Bjørk et al., 2018; Larocca et al., 2023). Geospatial techniques, such as the inference of the Equilibrium Line Altitude (ELA), have also been utilized (Brooks et al., 2022; Carrivick et al., 2023). However, aerial and satellite images provide temporal data over centuries and decades and geospatial methods neglect ice-flow physics and do not account for glacier dynamics. Based on the distribution of moraines and unvegetated trimlines in Central-Western Greenland, some authors suggested that the Late Holocene maximum glacier extent occurred around the LIA (Humlum, 1999). However, cosmic ray exposure (CRE) dating of erosive and depositional glacial records indicates that the maximum ice extent (MIE) of the Late Holocene did not occur during the LIA in many areas in Western Greenland but during the Medieval Warm Period (MWP; 950 to 1250 CE) (Young et al., 2015; Jomelli et al., 2016; Schweinsberg et al., 2019).

Evidence from physical-based modelling of the GICs recession during the Holocene remains limited compared to those near the GrIS (i.e., Cuzzone et al., 2019; Briner et al., 2020). Holocene reconstructions of GrIS extent based on physical modelling, guided by geomorphological evidence, provide valuable insights into the paleoclimate conditions that led to the MIE of the Late Holocene and subsequent recession (Simpson et al., 2009; Lecavalier et al., 2014; Cuzzone et al., 2019), facilitating comparisons of past and future glacier responses to climate change (Briner et al., 2020). Physical-based ice-flow modelling relying on full-Stokes equations are computationally intensive at high

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resolution (sub-kilometer) for long-term paleo glacier simulations and model parameter calibrations (Jouvet et al., 2022). Simplified models such as the hydrostatic Shallow Ice Approximation (SIA) and the Shallow Shelf Approximation (SSA) tend to overestimate ice velocities near glacier margins and underestimates velocities in deep glaciated areas, respectively. An emulator based on a convolutional neural network (CNN), trained with high-order ice flow equations, offers reduced computational costs while maintaining accurate ice thickness estimates comparable to those obtained through high-order equations (Jouvet, 2023a).

The future recession of Greenland GICs compared to the long-term Holocene fluctuations is poorly understood. Here, we calibrate and validate the Instructed Glacier Model (IGM) (Jouvet et al., 2023a), a glacier evolution model based on a CNN emulator to estimate ice flow, to reconstruct the MIE of the Late Holocene in an extended glacier area in the Eastern Nuussuaq Peninsula (Central-Western Greenland). This area has CRE records available for the outermost glacier moraine complexes but the paleoclimate conditions causing these glacier oscillations are not yet known in detail (D'andrea et al., 2011; Biette et al., 2019; Jomelli et al., 2016; Schweinsberg et al., 2019; Osman et al., 2021). Employing IGM allows us to reconstruct glaciers in high (90 m) resolution based on high-order equations (Jouvet, 2023a), demonstrating the methodology's capabilities for glacier modeling at regional scales. Future glacier evolution is modeled under the CMIP6 SSP2-4.5 and SSP5-8.5 scenarios, from present and steady-state glacier conditions to the year 2100. We compared the projected ice loss trend against the reconstructed MIE of the Late Holocene to the present-day ice loss trends, extending glacier records from decades to millennia and placing present and future glacier shrinkage within a long-term Holocene perspective.

The objectives of this work are to (i) reconstruct past glaciers under different climate conditions, (ii) determine past and future climate conditions influencing the MIE of the Late Holocene and future glacier recession, (iii) quantify future glacier retreat trends, and (iv) compare future ice loss trends with the rate of ice loss from the MIE of the Late Holocene to the present.

2. Study area

This study focuses on a land-terminating glacier area in the Nuussuaq Peninsula, Central-West Greenland (Figure 1). This peninsula extends from the onshore Disko (South) to Svartenhuk Halvo (North). Nuussuaq Peninsula includes several mountain glaciers and ice caps connected to the GrIS that surrounds its eastern flank. Our study focuses on a glacier area in the Eastern Nuussuaq Peninsula, with elevations ranging from 400 to 1200 meters above sea level (m a.s.l.) (Figure 1).

Present-day climate conditions are characterized by a polar maritime climate, becoming more continental toward the inland areas and GrIS (Humlum, 1999). Moist air masses from the Davis Strait influence the climate during summer, with continental polar air influences during the winter (Ingolfsson et al., 1990). Prevailing winds in the region

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typically come from the East and North-East, except during the summer months, when Southerly and Southern-Western winds prevail (Humlum, 1999). The relief configuration exposes Disko Bugt to cyclogenic activity and moist airflow, resulting in decreased precipitation from the peripheral coastal areas towards the GrIS (Weidick and Bennike, 2007). The nearest research station with meteorological and snow observations is the Arctic station, at coastal Disko Island (Central-Western Greenland). Here, the accumulated annual precipitation is 436 mm (1991–2004 period) (Hansen et al., 2006). The mean annual temperature (MAAT) is -4°C (1961–1990 period), with a lapse rate of around 0.6°C per 100 m (Humlum, 1998). At Arctic station, the snow season typically extends from September to June, with maximum snow accumulations of around 50 cm (Bonsoms et al., 2024).

The present-day landscape in Central-West Greenland is characterized by the presence of glaciers, which have also intensely shaped the relief in ice-free areas in the past. Today, environmental dynamics in these areas is strongly influenced by periglacial processes under a continuous permafrost regime (Humlum, 1998; Christiansen et al., 2010) that reshape the geological setting made of clastic sediments from the Mid-Cretaceous to the Palaeogene (Pedersen et al., 2002). The strong glacial imprint in the landscape of the peninsula results from a complex glacial history, which is not yet known in detail. Following the LGM, the GrIS underwent a significant retreat during Termination-1 and exposed the coastal regions in Central-West Greenland (Briner et al., 2020). As in other regions across Greenland, the Early Holocene was characterized by warm temperatures that led glaciers to retreat (Leger et al., 2024). In the Nuussuaq Peninsula, CRE records reported the onset of glacial retreat by ca. 10 ka (O'Hara et al., 2017). The minimum GrIS extension occurred from ca. 5 to 3 ka cal BP, when GrIS margins retreated by ca. 150 km from present-day terminus position (Briner et al., 2016), which explains the lack of glacial records corresponding to the Early-Mid Holocene in the peninsula (Kelly and Lowell, 2009; O'Hara et al., 2017). According to several absolute dating methods in different natural records, the Nuussuaq Peninsula GICs grew between approximately 4.3 and 2 ka and reached several glacier culminations during the past millennium before the LIA (Schweinsberg et al., 2017; 2019). The internal and external moraine complexes in the area reported CRE ages of 1130 ± 40 and 925 ± 80 CE, respectively (Young et al., 2015). These ages are consistent with other CRE ages obtained in Central-Western Greenland for the most external recent moraine complexes, indicating that the late MWP glacier expansion was the largest of the Late Holocene (Jomelli et al., 2016; Schweinsberg et al., 2019)

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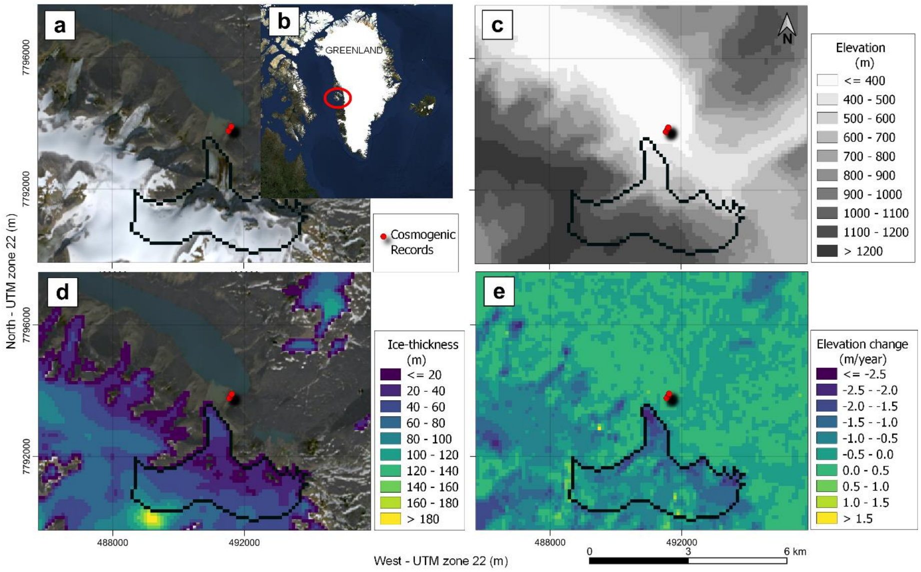

Figure 1. Location of the reconstructed glacier and CRE ages (red points) used in this work (a). Location of the study area within Greenland (b). Glacier delimitation is based on Randolph Glacier Inventory (RGI6). The base map is a Sentinel-2 image from https://s2maps.eu/ (2022). Elevation map of the study area from a Digital Elevation Model Copernicus DEM GLO-90 (2010-2015) (c). Average ice thickness (m) (2018-2022) from Millan et al. (2022) (d). Elevation changes average values (m/year) between 2000-2019 (Hugonnet et al., 2021) (e).

3. Data

3.1 Pre-processing glacier data downloading

The data used to force and validate the IGM is detailed at Table 1. Topography data were obtained from a Copernicus Digital Elevation Model (DEM) with a resolution of 90 meters (COPERNICUS DEM GLO-90). In-situ mass balance records in the study area are scarce and fieldwork is challenging. However, recent advancements in global-scale mass balance, satellite imagery, and ice thickness estimates have enabled the validation of the glacier model. We compared the ice thickness estimates from Farinotti et al. (2019), which are the output from an ensemble of five models (HF-model, GlabTop2, OGGM, GlabTop2 IITB version, and an unnamed model), with those from Millan et al. (2022), which are derived from numerical modeling based on SIA and data obtained from a constellation of remote sensing products (Sentinel-1/ESA, Sentinel-2/ESA, Landsat-8/USGS, Venµs/CNES-ISA, Pléiades/Airbus D&S). Glacier mask outlines were acquired from the Randolph Glacier Inventory Version 6 (RGI6.0). Elevation change rate (dh/dt) data were obtained from Huggonet et al. (2021).

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The climate variables required to run the IGM model are monthly accumulated precipitation (kg m^{-2} yr^{-1}), monthly average air temperature (ºC), and monthly air temperature standard deviation (ºC) from the nearest pixel to the glacier. We utilized the GSWP3- W5E5 monthly dataset at a spatial resolution of 0.5° x 0.5°, which combines the Global Soil Wetness Project phase 3 dataset with the bias-adjusted ERA5 reanalysis dataset (Cucchi et al., 2020). Future glacier changes are modeled based on bias-corrected monthly accumulated precipitation and monthly average air temperature CMIP6 multi-model mean (n=33) for SSP2-4.5 and SSP5-8.5 (2020 to 2100) at a spatial resolution of 0.25° (Thrasher et al., 2022), subtracted at the nearest grid point of the glaciers. Months are aggregated into seasons as follows: September, October, November (Autumn), March, April and May (Spring), December, January and February (Winter), June, July and August (Summer). Data were downloaded using Open Global Glacier Model (OGGM) model (https://oggm.org/) (Maussion et al., 2015) shop module of IGM (Jouvet et al., 2023), except for CMIP6 projections (Thrasher et al., 2022) and ice thickness estimates (Farinotti et al., 2019).

Table 1. Characteristics of the datasets employed for forcing, calibrating, and validating the IGM.

Description	Name	Spatial resolution	Database Date	Source
DEM	Copernicus DEM GLO-90	90 m	2010-2015	https://spacedata.copernicus.eu/documents/20126/0/CSCDA_ESA_User_Licence_2021_11_17.pdf
Ice-thickness (1)	Millan et al. (2022)	100 m	2017-2018	Millan et al. (2022)
Ice-thickness (2)	Farinotti et al. (2019)	25 m	2019	Farinotti et al. (2019)
Baseline climate data	GSWP3_W5E5v2.0	0.5° x 0.5°	1960-1990	https://data.isimip.org/search/simulation_round/ISIMIP2a/product/InputData/climate_forcing/gswp3-w5e5/
CMIP6 projections	CMIP 6	0.25º	1960-2100	Thrasher et al. (2022)
Dh/dt	Hugonnet et al. (2021)	Glacier (RGI6.0) level	2000-2020	Huggonet et al. (2021)
Glacier Outline	RGI6.0	Glacier (RGI6.0) level	2003	https://www.glims.org/RGI/

3.2 Geomorphological and paleoclimate data

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The CRE ages are based on nuclide (^{10}Be) introduced by Young et al. (2015) and refer to the period of the maximum glacier advance of the last warm/cold cycles in the Nuussuaq Peninsula and were used for the paleoclimate modelling purposes of this study. The sampled boulders were obtained from the outer ridge of the moraine and reveal either (i) a period of glacial surge or (ii) a phase of stabilization/stillness during the long-term retreat. However, special caution must be taken when interpreting these ages, as they are not directly indicative of the period of ice occupation but of the timing of stabilization of moraine boulders.

Paleoclimate anomalies with respect to the baseline climate were obtained from annual air temperature reconstructions from ice cores of the GrIS and margins of the GrIS provided by Buizert et al. (2018). This data ranges from the Last Glacial Maximum (LGM; ~ 26-19 ka ago) to 2000 CE.

4. Methods

4.1 Instructed Glacier Model (IGM)

The IGM is a glacier model that simulates ice thickness evolution according to ice mass conservation principles, surface mass balance and ice flow physics (Jouvet et al., 2023a). IGM updates the ice thickness at each time step from ice flow and surface mass balance (SMB) by solving the mass conservation equation. The ice flow is modelled using a CNN model that is trained to satisfy high-order ice flow equations. There are two main parameters that control the strength of the ice flow: the Arrhenius factor (A) that controls the ice viscosity in Glen's flow law (Glen, 1955) and the basal sliding coefficient (c), by the nonlinear sliding Weertman's law (Weertman, 1957).

Temperature data is downscaled over the DEM using a reference height and a constant lapse rate of -0.6ºC/100 m, while precipitation is downscaled using a vertical gradient of 35 mm/100 m. Precipitation is classified as solid (< 0ºC) or liquid (> 2ºC), with a linear transition between solid and liquid phases. The melting threshold is set to -1ºC, and the density of water is fixed at 1000 kg $^{m-3}$. The SMB is estimated using a monthly positive degree-day (PDD) model (Hock, 2003; Huss, 2008). The PDD is calibrated based on the OGGM v1.6.1 SMB calibration process, which is included in IGM SMB module. OGGM v1.6.1 SMB calibration correct temperature and precipitation biases from climate data and adjust the melt factor (5, in this case) to fit the average glacier geodetic mass balance from January 2000 to January 2020 from Hugonnet et al. (2021). Further details of the OGGM v1.6.1 SMB calibration process are provided in the OGGM documentation (https://oggm.org/tutorials/master/notebooks/tutorials/massbalance_global_params.html), whereas the physical basis of IGM is detailed in Jouvet et al. (2022; 2023a).

4.2 Present day glacier calibration and validation

We calibrated the IGM to simulate RGI6.0 area, and ice thickness from available datasets (Farinotti et al., 2019; Millan et al., 2022). The IGM parametrization is performed based on conducting a sensitivity analysis to A and *c*. These parameters were chosen to optimize

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IGM and accurately simulate different ice conditions, basal sliding conditions and subglacial hydrology. An ensemble of IGM parameter options was performed over a model run of 1000-years with different temperature perturbations of -0.75ºC, -0.5ºC, 0ºC and +0.25 ºC with respect to baseline climate (1960-1990) in order to reach long-term (> 500 years) glacier area steady-state conditions. The range of temperature perturbation was determined through trial and error, which showed that values outside this range of temperature anomalies produced higher discrepancies with respect to the available datasets used for results validation (Figure 3 to 5). A sensitivity analysis was performed on IGM parametrization to simulate cold, temperate, and soft ice conditions by changing A from 34 MPa^{-3} a^{-1}, 78 MPa^{-3} a^{-1} (IGM default value) to 150 MPa^{-3} a^{-1}. Sliding conditions are parametrized by changing *c* from 0.01 km MPa^{-3} a^{-1}, 0.03 km MPa^{-3} a^{-1} (IGM default value), and 0.05 km MPa^{-3} a^{-1}. The IGM parametrization is shown in Figures 3 to 5. The remaining parameters were set to the default configuration of the IGM.

The accuracy evaluation of the modeled IGM outputs is based on both area and ice thickness. We calculated (i) the Mean Absolute Error (MAE) between the accumulated glacier ice thickness from Farinotti et al. (2019) and Millan et al. (2022) and the output from IGM; (ii) the glacier area difference between RGI6.0 area and from IGM. To incorporate both area and ice thickness errors, we calculated the bias by multiplying the ice thickness MAE (i) by the area difference (ii).

4.3 Past and future glacier evolution

IGM is forced with the lowest error parameterization option until the glacier area reaches present-day and long-term stable-state conditions. The model is run again 1000-years with an ensemble of different temperature and precipitation values to simulate MIE of the Late Holocene from MWP. The temperature was perturbed over the baseline climate from 0 to -1ºC by steps of 0.25ºC. Precipitation was non-changed (0%) and increased (10%) in order to estimate if high rates of snowfall could compensate warming. MIE of the Late Holocene paleoclimate conditions were determined by calculating the distance between the glacier tongue of the ensemble of simulations and the CRE dates of the outer ridge moraines (Köse et al., 2022). The simulations that match the outer ridge moraines represent the climate conditions before the CRE dates. The present-day glacier area with steady-state conditions is the starting point of the future simulations (Zekollari et al., 2019). Subsequently, the IGM is run from the present day until 2100 using monthly accumulated precipitation and average air temperature CMIP6 multi-model mean SSP2-4.5 and SSP5-8.5 anomalies with respect to the baseline climate, applying additive factors for temperature and multiplicative factors for precipitation (Rounce et al., 2023). Present and future ice thickness anomalies with respect to the MIE of the Late Holocene are calculated by subtracting the difference between the accumulated ice thickness for the MIE of the Late Holocene (i) from the accumulated ice thickness from the present-day (ii) and future ice-loss (iii), dividing by the accumulated ice thickness for the MIE of the Late Holocene (i), and multiplying by 100. The factor of increase under future climate

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change is calculated by dividing future ice loss anomalies by the present-day ice loss anomalies relative to the MIE of the Late Holocene.

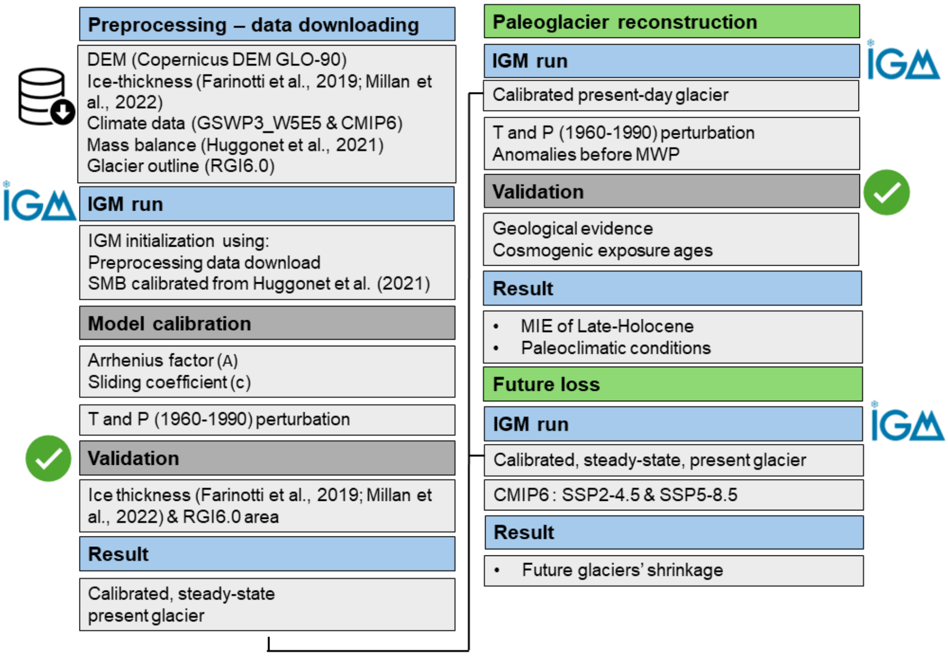

Figure 2. Flowchart followed for reconstructing past and present glaciers and projecting their future evolution based on air temperature (T) and precipitation (P).

5. Results

5.1 IGM parametrization and calibration

For most IGM parametrizations, glacier growth occurred until 350 to 600 years of spin-up. The latest year of the spin-up simulation is subsequently validated against ice thickness estimates (Farinotti et al., 2019; Millan et al., 2022). The error metric values for ice thickness and area resulting from the IGM calibration and parametrization process are shown in Figure 3 to 5. The most favorable range for achieving accurate results for present-day glaciers is a perturbation range of temperature from 0ºC to -0.5ºC with respect to the baseline climate (Figure 3 to 5). The largest errors in ice thickness and glacier area were observed for the A = 34 MPa^{-3} a^{-1} and *c* = 0.01 km MPa^{-3} a^{-1} IGM configuration. This configuration tended to overestimate ice thickness for both global-scale ice thickness references (Figure 4 and 5). Additionally, using the default configuration and reducing the temperature to < -0.5 ºC over the baseline climate led to overestimations of ice thickness.

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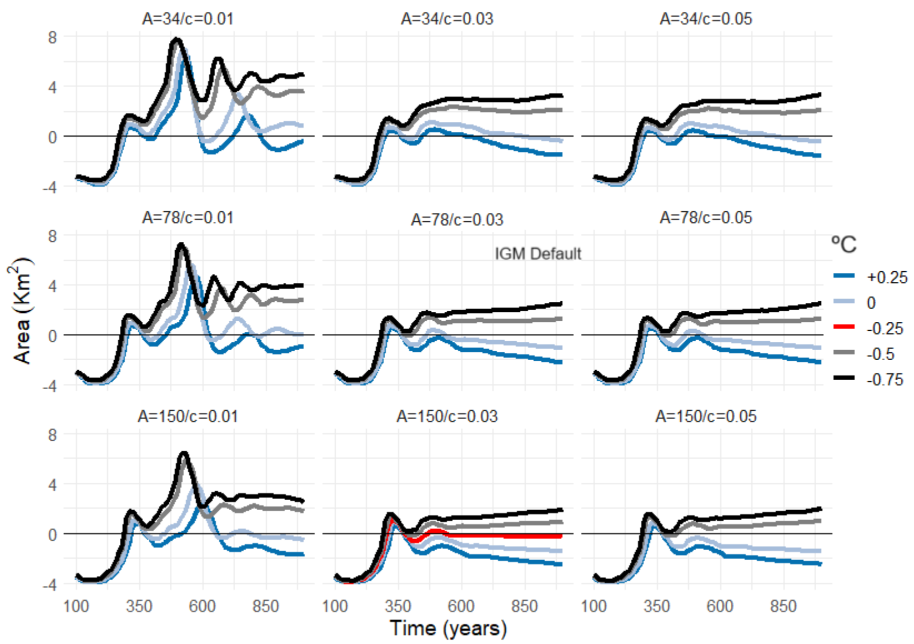

Figure 3. Difference from the RGI6.0 area and IGM outputs within a 1000-years spin-up. Data is grouped by changes in temperature (colors), A and *c* options (boxes). The selected configuration (A = 150 MPa^{-3} a^{-1} and *c* = 0.03 km MPa^{-3} a^{-1}, -0.25ºC with respect to the baseline climate) is shown in red color.

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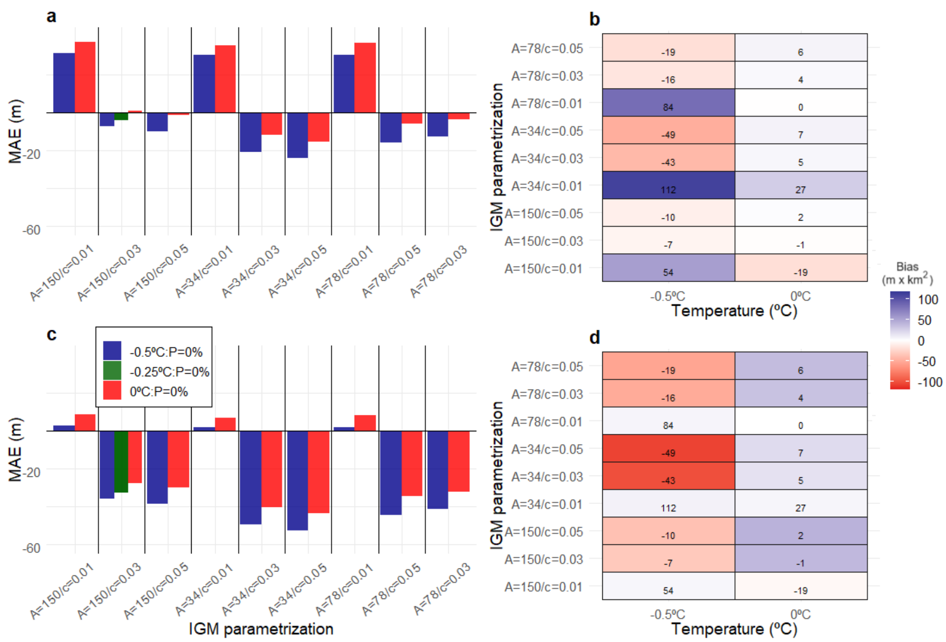

Figure 4. Ice thickness MAE between Farinotti et al. (2019) and IGM outputs after spin-up with different A and *c* parametrizations and perturbations of temperature (a). The selected configuration is shown with green color. Ice thickness MAE values from Figure 4 (a) multiplied by the difference between the RGI6.0 area and the IGM outputs (bias) for different A and *c* parametrizations and perturbations of temperature (b). Figure 4 (c) and (d) are the same as Figure 4 (a) and (b), respectively, but for ice thickness estimates from Millan et al. (2022).

Trial and error parametrizations of A and c revealed that optimal results were achieved for A = 150 MPa^{-3}a^{-1} and c = 0.03 km MPa^{-3} a^{-1}. These outputs of the IGM align with Farinotti et al. (2019). However, both IGM ice thickness and Farinotti et al. (2019) overestimate ice thickness compared to Millan et al. (2022) (Figures 4 and 5). Setting A = 150 MPa^{-3}a^{-1} and c = 0.03 km MPa^{-3} a^{-1}, with a slight variation of temperature (-0.25 ºC) over the baseline climate, resulted in very similar accumulated ice thickness to Farinotti et al. (2019) (MAE = 4 m; Figure 4a), a minimal RGI6.0 area bias (Figure 3 and 4b), and very stable-state glacier conditions for > 500 years (Figure 3). This configuration also minimized errors against the Millan et al. (2022) dataset (MAE = 24 m) (Figure 4c). Thus, glacier reconstruction and projection are based on this parametrization option.

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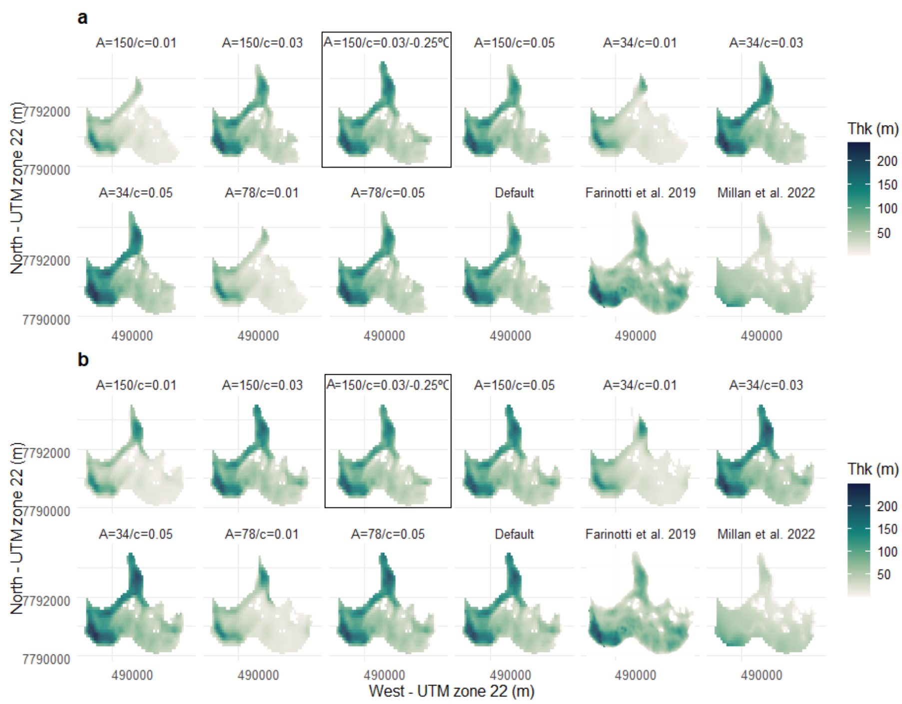

Figure 5. Average ice thickness data from Farinotti et al. (2019) and Millan et al. (2022), along with examples of IGM configuration options using different A and c parameters for the selected configuration (highlighted with a black square), are shown for a temperature of 0ºC with respect to the baseline climate (a) and a temperature of -0.5ºC with respect to the baseline climate (b). The ice thickness values of IGM shown are the result of steady-state glacier conditions.

5.2 Late Holocene maximum glacier extension and paleoclimatic conditions

The temperature evolution from the LGM to 2000 CE, as reconstructed from GrIS ice cores and Greenland margins (Buizert et al., 2018), is shown in Figure 6. The Camp Century and Disko Bugt/Jakobshavn ice cores exhibit similar temperature trends compared to the baseline climate period, although they display larger temperature anomalies, with the warmest conditions recorded during the Holocene Warm Period (HWP; ~ 9-5 ka ago) (up to 3ºC with respect to the baseline climate period). A long-term cooling trend is detected for the Late Holocene, with moderate anomalies and high yearly oscillations of around ±1 ºC between the Dark Ages Cold Period (~ 400 to 765 CE; Helama et al., 2017) and the MWP for Disko Bugt/Jakobshavn (Figure 6). However, colder temperatures are found in the Camp Century ice core. For both locations, the coldest temperature anomalies of ca. -2ºC compared to the baseline climate are found during the LIA.

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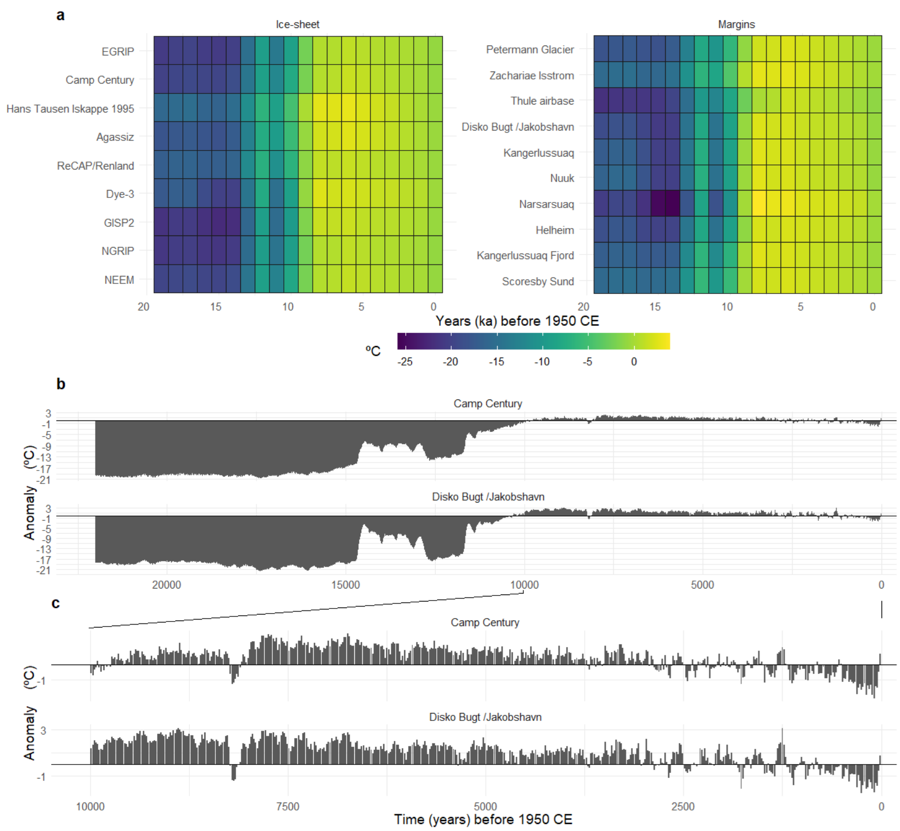

Figure 6. Air temperature anomalies from the LGM period to the present reconstructed from ice-core records of the GrIS and Greenland margins (a). Air temperature anomalies from the ice cores near to the glacier reconstructed in Central-Western Greenland (Camp Century and Disko Bugt/Jakobshavn) (b). The black squares highlighted in (a) correspond to the specific locations shown in (b). Air temperature anomalies since 10 ka to 1950 (c). Air temperature anomalies are calculated by the difference between the average annual air temperature from the baseline climate (1960-1990 period) and the annual air temperature for each location. Data were obtained from the reconstruction available from Buizert et al. (2018).

Future CMIP6 SSP2-4.5 and SSP5-8.5 anomalies with respect to the baseline climate are shown in Figure 7. The temporal evolution of temperature follows a similar warming rate for both scenarios until 2040, after which there is an acceleration of warming for SSP5-8.5. The increase in temperature relative to the baseline climate for SSP2-4.5 is 3.1ºC by 2050 and 4.2ºC by 2100, whereas for SSP5-8.5, is 3.4ºC by 2050 and 6.1ºC by 2100. Thus, CMIP6 (SSP2-4.5 and SSP5-8.5) anomalies with respect to the baseline climate are similar to HWP for 2050 but higher by a factor of three for SSP5-8.5 by 2100 (Figure 6).

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Precipitation also shows an increase with respect to the baseline climate, which is more pronounced for SSP5-8.5 towards the end of the 21st century. SSP2-4.5 precipitation anomalies with respect to the baseline climate are +20% by 2050, increasing to +26% by 2100. For SSP5-8.5, precipitation increases by +20% by 2050 and +38% by 2100.

For the 2050-2060 period, summer temperatures are projected to range from 2ºC under SSP2-4.5 to 3ºC under SSP5-8.5. For the 2090-2100 period, winter temperatures are projected to range from 5ºC under SSP2-4.5 to 8ºC under SSP5-8.5 (Figure S1). Regarding snowfall and for the 2050-2060 period, SSP2-4.5 and SSP5-8.5 projects anomalies of 12% and 16 %, respectively (Figure S2 and S3). For the 2090-2100 period, anomalies with respect to the baseline climate are 18 % and 22% for SSP2-4.5 and SSP5-8.5, respectively. Other months show decreases in snowfall except for Spring, which shows a 3% increase for both SSP5-8.5 and SSP2-4.5 scenarios for 2050-2060 and 2090-2100 periods.

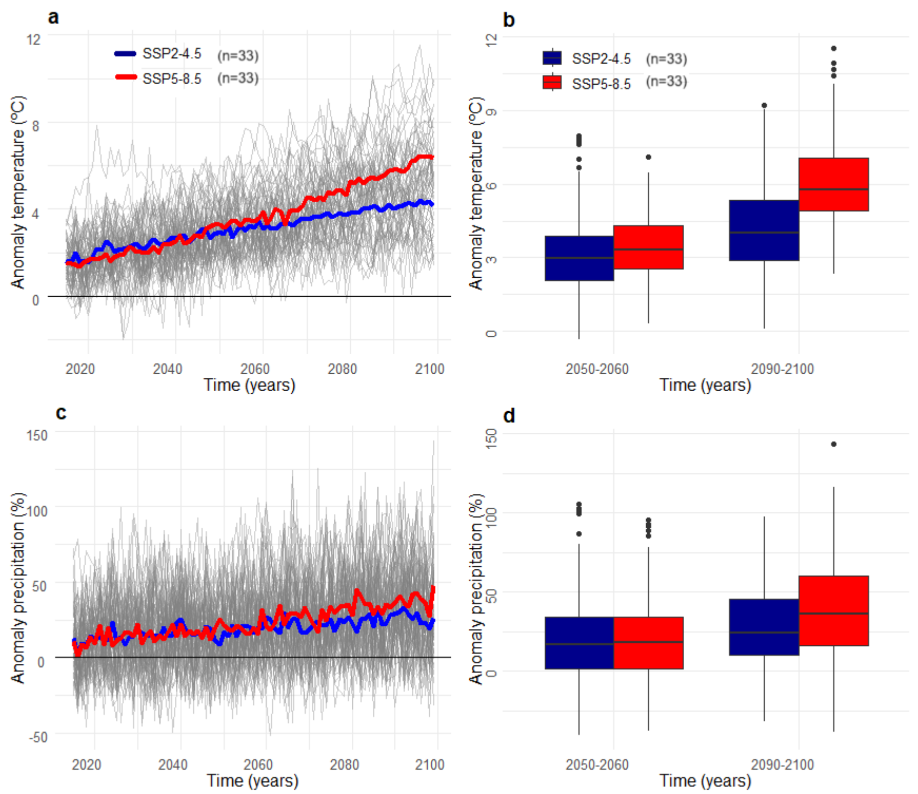

Figure 7. Temporal evolution of CMIP6 SSP2-4.5 and SSP5-8.5 temperature anomalies with respect to the baseline climate period (a). Comparison of CMIP6 SSP2-4.5 and SSP5-8.5 temperature anomalies with respect to the baseline climate period for 2050-2060 and 2090-2100 temporal periods (b). Figure 7 (c) and (d) are the same as Figure 7

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(a) and (b), respectively, but for precipitation. The dots of (b) and (d) represent the average of each CMIP6 model for the temporal period and climate variable.

We further assessed whether the temperature conditions reconstructed from ice-core data are consistent with CRE dates from moraine boulders and can accurately replicate the MIE of the Late Holocene. The IGM was spin-up and forced with the lowest error configuration. Subsequently, a sensitivity analysis of temperature and precipitation was conducted. The IGM was run after present-day steady-state conditions, with variations of temperature from 0 to -1 by steps of 0.25ºC. Precipitation was increased by 10%. We determined the temperature and precipitation conditions that allowed the MIE of the Late Holocene glacier extension, enabling its reconstruction (Figure 8).

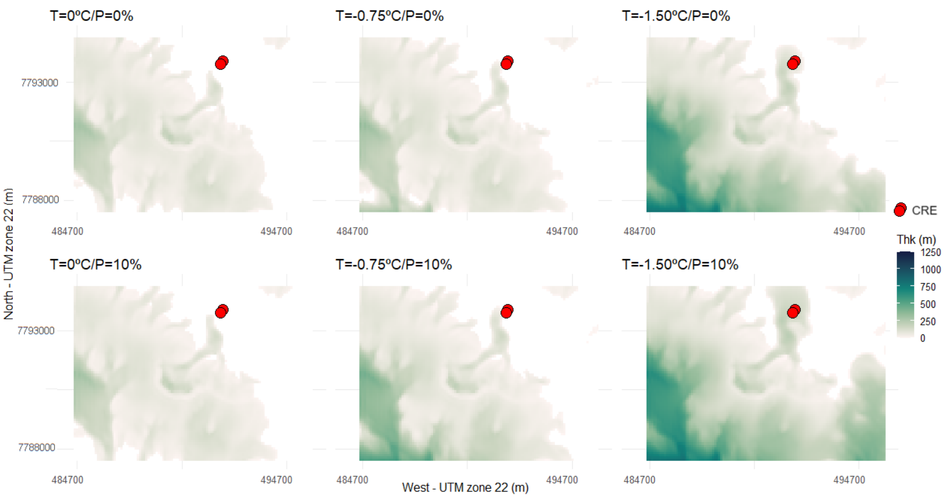

Figure 8. Location of the CRE samples (red dots) and average ice thickness (m) for various temperature (T) and precipitation (P) perturbations. The ice thickness values shown are the result of performing a spin-up model run, and subsequently a 1000-year model run for reconstructing the MIE of the Late Holocene.

The assessment of past temperature and precipitation anomalies relative to the baseline climate is conducted based on the distance between the glacier tongue and available CRE dates. This analysis indicates the temperature and precipitation conditions that facilitated glacier expansion during the MIE of the Late Holocene. Note that there may be a time gap between MIE of the Late Holocene the timing of maximum ice expansion and CRE ages and the since these ages indicate not the period of glacial growth but rather the period when moraine boulders stabilized after the formation of the moraine ridges formed by the glacier advances/stillstands. The minimum distance for all samples is reached when temperature is reduced by 0.5ºC and precipitation is increased by 10% with respect to the baseline climate. A reduction of 0.75ºC while maintaining precipitation unchanged resulted in glacier advances to the limit marked by the dated moraine boulders (Figure 8 and 9). These findings suggest that temperature anomalies leading to glacier extension up

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to the MIE of the Late Holocene ranged at least from temperatures of -0.5ºC and precipitation of +10% to temperatures of <= -0.75ºC and precipitation of 0% relative to the baseline climate (Figure 9). However, a variation in precipitation of 10% is unlikely according to paleoclimate reconstructions for the Late Holocene (Badgeley et al., 2020). This suggests that a temperature decrease of at least 0.75ºC from the baseline climate, with no changes in precipitation, is the most plausible climate scenario.

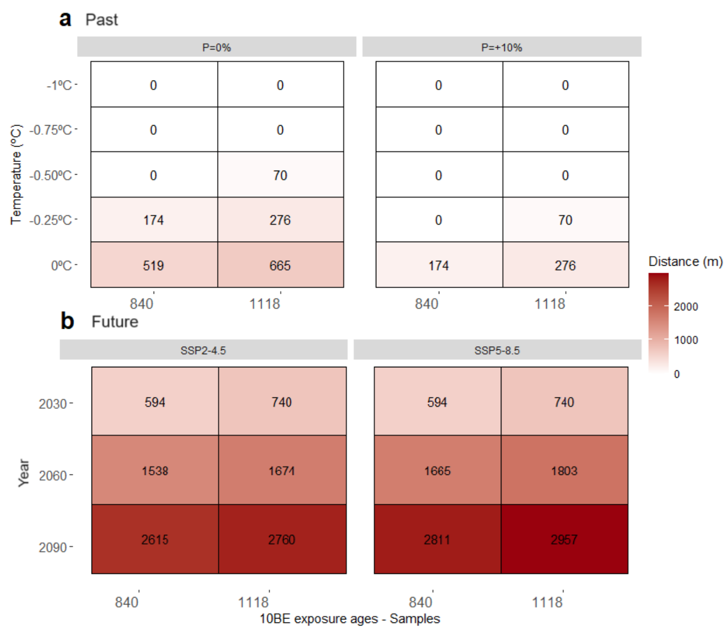

Figure 9. Differences in pixel distance (m) between the nearest modelled glacier extension and the sample age location (x-axis) across various air temperature (y-axis) and precipitation options (boxes) (a). Differences in pixel distance (m) between the nearest modelled glacier extension and the sample age location (x-axis) and projected glacier shrinkage for CMIP6 scenarios (boxes) and different years (y-axis).

The results suggest that with -0.75 ºC and no changes in precipitation anomalies with respect to baseline climate, the glacier ice thickness has reduced 34% with respect to surface covered by glaciers during the MIE of the Late Holocene (Figure 10).

5.3 Future glacier changes

Assuming the PDD parametrization of the 2000-2020 period calibrated with geodetic data (Hugonnet et al., 2021), the future climate for 2060 leads to glacier tongue recession ranging from 1674 m (SSP2-4.5) to 1903 m (SSP5-8.5) relative to the MIE of the Late Holocene (Figures 9 to 11). By 2090, glacier reduction is projected to reach up to 2760 m (SSP2-4.5) or 2957 m (SSP5-8.5). These results indicate that the projected increase in precipitation (Figure 7) is insufficient to offset glacier shrinkage. The rate of ice loss from

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the MIE of the Late Holocene to the present (34%) will double after 2070 (Figure 10a), regardless of the CMIP6 scenario. The rate of ice loss will increase by 2080, reaching anomalies of 72% (SSP2-4.5) and 78% (SSP5-8.5). By 2100 and under SSP5-8.5, the reduction in ice thickness will reach a maximum ice loss of 95% relative to the MIE of the Late Holocene (Figures 10b and 11).

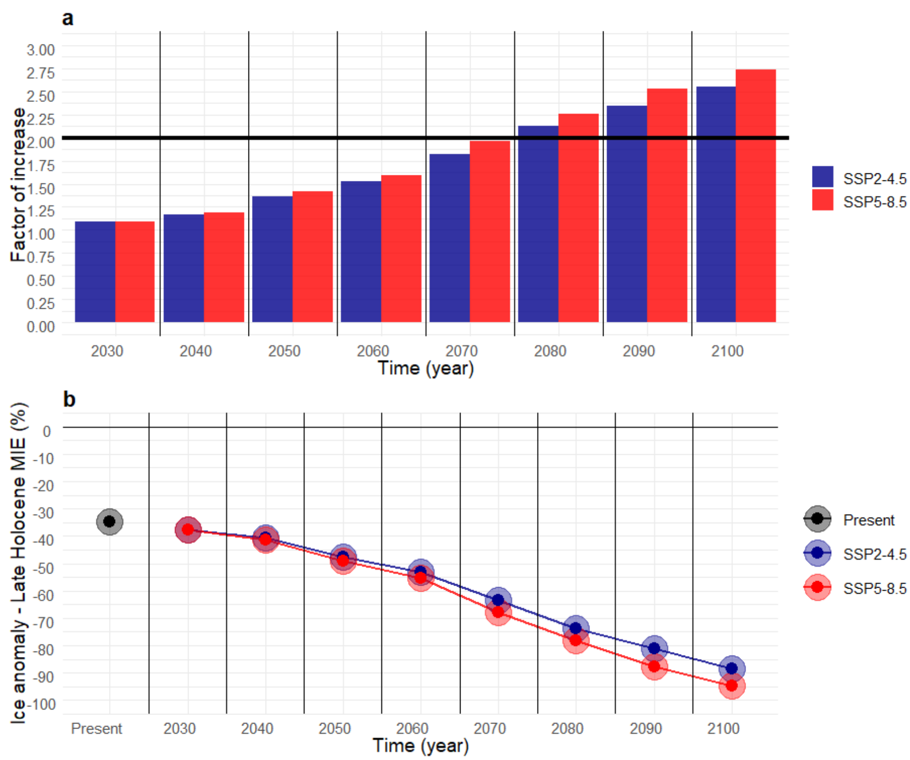

Figure 10. Factor of increase in ice loss under future climate change compared to ice loss from the MIE of the Late Holocene to the present (a). Ice thickness anomalies for the present and future CMIP6 SSP2-4.5 and SSP5-8.5 scenarios (b). Anomalies are calculated by subtracting the accumulated yearly ice thickness for the MIE of the Late Holocene (i) from the accumulated yearly ice thickness from the present-day (ii) and future ice-loss changes (iii), dividing by (i), and multiplying by 100. The factor of increase under future climate change is calculated by dividing future ice loss anomalies by the present-day ice loss anomalies relative to the MIE of the Late Holocene.

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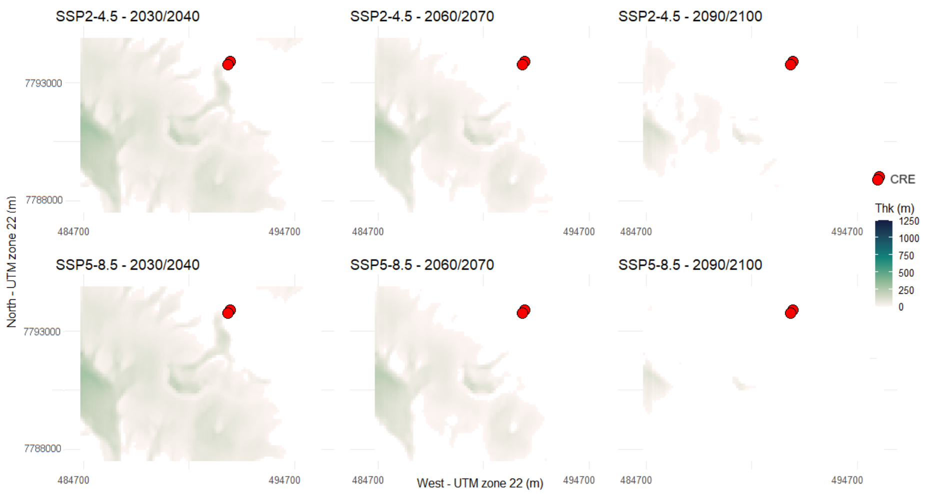

Figure 11. Location of the CRE samples (red dots) and average ice thickness (m) for future CMIP6 SSP2-4.5 and SSP5-8.5 scenarios and different temporal periods. The ice thickness values shown are the result of performing a spin-up model run reaching steady state conditions, and subsequently performing a model run with CMIP6 projections from present-day to 2100.

6. Discussion

6.1 Glacier modelling as a tool to understand paleoclimate conditions

The range of temperature decrease (0.75ºC to 1ºC) that obtained the best results in terms of reproducing glacier's MIE of Late Holocene area is consistent with past temperature anomalies in the Western and Southern Greenland found in previous works. Particularly, this range of temperature anomalies fall between estimates of ~1.5 ºC cooler temperature at 1850 CE with respect to 1990s (Dahl-Jensen et al., 1998). In Southern-Western Greenland, temperature estimates derived from geospatial reconstruction of ELAs that attributed historical MIE to the LIA, suggest temperatures ranging from around -0.4 to -0.9 ºC (Larocca et al., 2020). Employing a similar methodology, other studies have found temperature anomalies during the LIA to be of -1.1 ± 0.6 ºC, with no observed changes in precipitation (Brooks et al., 2022).

However, the MIE of the Late Holocene in Central-Western Greenland defined by the most recent moraine complexes suggest an earlier maximum glacier extent than in other areas in the Northern Hemisphere when the LIA glacier expansion was much more extensive (Young et al., 2015). In this sense, the temperature indicated by at Northern-Hemispheric scale reanalysis for the MWP is not consistent with MIE obtained from glacier moraines dated with CRE in Western Greenland (Jomelli et al., 2016; Biette et al., 2019). Indeed, Biette et al. (2019) modelled the outlet glacier of the Lyngmarksbræen ice cap (Disko Island) and tested its sensitivity to temperature and precipitation using a PDD approach guided by temperature anomalies from a lake sediment located 250 km south of

Disko Island (D'Andrea et al., 2011). They demonstrated that the MIE of the Late Holocene during the late MWP (1200 ± 130 CE) occurred when temperatures ranged from -1.3°C to -1.6°C, and precipitation changed by ± 10% (Biette et al., 2019). Considering that the baseline climate period of our work is 1960 to 1990, and their anomalies are considered to the end-20th century, our results are similar to these reported temperature and precipitation values. These results are in line with a decrease in summer temperature from -0.5º to -3ºC at around 250 km of Disko Island during the MWP obtained from lacustrine (alkenone-based lake sediment) reconstructions, and a second cold phase during the LIA(D'andrea et al., 2011). These cold conditions have been linked to multi-decadal cold spells intense enough to cause a major advance of Baffin Bay during the MWP (Young et al., 2015; Jomelli et al., 2016). These glacier advances were also observed in other Northern Hemisphere glaciers and may be also enhanced by volcanic eruptions (Solomina et al., 2016). The reconstructed glacier advances were probably linked with a recurrent positive NAO at West Greenland and Baffin Bay during MWP that lead to cool conditions (Young et al., 2015). However, other studies suggested that the NAO was not predominantly positive during this period (Lasher and Axford, 2019). There are also studies suggesting cold sea-surface temperatures observed during the MWP (Sha et al., 2017), while others suggest warmer conditions during the MWP compared to the LIA (Perner et al., 2012).

The MIE of the glaciers in the study area was reached during the MWP (1130 ± 40 and 925 ± 80 CE) (Young et al., 2015), has been suggested to be linked to increased snowfall rates that counterbalanced the glacier ablation mass losses during the MWP, which were slightly higher than those during the LIA (Osman et al., 2021). However, an increase of precipitation of + 10 % with respect to the baseline climate period is not able to counterbalance glacier recession under a change of <-0.5ºC with respect to the baseline climate (Figure 9). The sensitivity analysis performed here reveals that the glaciers were far isothermal conditions within 1960-1990 period and a small decadal variation of temperature with respect to the baseline analysis to temperature and precipitation performed in this work is consistent with previous works that suggest that around 90% of variation and that glacier maximum extension dynamics are linked with summer temperature (Miller et al., 2012; Young et al., 2015).

6.2 Central-Western Greenland ice-loss and comparison with other Greenland areas

The reconstructed MIE of the Late Holocene represents the phase with most recent widespread glacier advances from the Nuussuaq Peninsula, and it occurred prior to the LIA (Schweinsberg et al., 2017; 2019). The maximum glacier advance reconstructed in this work for Central-Western Greenland is not consistent across Greenland. Northern GrIS exhibits stability during the Late Holocene with advances at 2.8 ka and 1650 CE (Reusche et al., 2018). In particular, North-Western glaciers length was similar from 5.8 ka until onset of LIA (Søndergaard et al., 2020). In the Bregne Ice Cap (East Greenland) glacier length dating reveals a peak during the LIA (~ 0.74 ka; Levy et al., 2014). However, in Renland Ice Cap (Eastern Greenland) glacier exceeded present limits at 3.3

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ka and around 1 ka, which is similar to LIA glacier advance (Medford et al., 2021). In Central-East Greenland, cold climate conditions occurred during LIA at Stauning Alper with peaks of 0.78 ± 0.31 ka (Kelly et al., 2008), and at Istorvet ice cap, that reached its maximum Holocene extent at 0.8 ± 0.3 ka (Lowell et al., 2013). This expansion observed in Eastern Greenland corresponds with peak glacier extensions seen in Iceland, attributed to LIA (Flowers et al., 2008). Different asymmetries between Greenland sectors are seen historically as revealed by long-term GICs recession larger in West Greenland than in East, which has been attributed to the positive oscillation of NAO since the LIA that led to warmer conditions in West Greenland due to the West-East NAO dipole (Bjørk et al., 2018).

In Central-Western Greenland, most of the studies focusing on Late Holocene glacial history come from near Disko Island (Ingolfsson et al., 1990; Humlum, 1998; Yde and Knudsen, 2007; Citterio et al., 2009; Jomelli et al., 2016). Here, in its Eastern fringe, the ELA from the LIA is estimated at ca. 550 ± 500 m, contrasting with values of 200-300 m attributed elsewhere in the island (Ingolfsson et al., 1990). In the Western section, however, the ELA during the LIA was estimated to be at 450 ± 420 m (Humlum, 1998). As in Nuussuaq Peninsula, the Holocene maximum extension in Disko Island is evidenced by moraine systems exhibiting a fresh, partly unvegetated appearance, with of prevalence of *Rhizocarpon geographicum* in these moraines (Humlum, 1987). This absence of Holocene moraine systems beyond the LIA moraines indicates that the advance of LIA represents the maximum extension of this glacier since the Late Holocene (Humlum, 1999). This moraine evidence has been used to estimate the ELA (Brooks et al., 2022; Carrivick et al., 2023). Particularly, using geospatial methods Carrivick et al. (2023) attributed this trimline to the maximum extent of LIA and concluded that Greenland GICs lost 499 Gt since end-LIA, corresponding to 1.38 mm sea level equivalent. Similarly, in Southern-Western Greenland, 42 GICs lost 48% of their area since the LIA with respect to 2019 (Brooks et al., 2022). These values are slightly higher than the 34 % reduction from the MIE of the Late Holocene with respect to present-day glacier reported in this work. The differences could be attributed to the local relief configuration as well as to the north aspect of the reconstructed glacier area and methodological variances. Additionally, while we are employing a glacier modelling approach constrained by geological records of a specific age, previous studies have estimated distances based on ELAs and geospatial methods that account for spatial distances between present-day glaciers tongue and maximum historical moraines that could be formed prior to the LIA. According to remote sensing data, in Disko Island GICs inventory and monitoring from 1953 to 2005 indicates that the average recession during this timeframe amounted to 11% of the glacier lengths recorded in 1953 (number of glaciers, n = 172), and 38% of the distance between LIA moraines and glacier termini in 1953 (n = 87) (Yde and Knudsen, 2007). These values are lower than those observed at Pjetursson Glacier (Disko Island), which has retreated since the LIA with a decrease in total glacier area of around 40% by the end of the 20th century according to geospatial methods (Bøcker, 1996). Using remote sensing data, LIA to 2001 glacier shrinkage in

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Central-Western Greenland was estimated in a reduction of ~ 20% of the area (Citterio et al., 2009).

Currently, the modeled glacier area and volume are out of balance with respect to the temperature since 1990 to present (figure not shown), necessitating the simulation of glaciers using temperature and precipitation data from the 1960-1990 period (Figure 2). This indicates a committed ice loss regardless of future climate scenarios. Future projections show a remarkable increase in temperature, reaching HWP anomalies by 2050 and tripling HWP anomalies by 2100 under SSP5-8.5. Our results indicate that glacier mass loss by >2070 will double the ice loss from the MIE of the Late Holocene to the present. Precipitation is projected to increase by 20% (2050; SSP2-4.5 and SSP5-8.5) up to 38% (2100; SSP5-8.5) compared to the baseline climate but cannot counterbalance glacier losses. The modeled GICs mass loss is expected to reach MIE of the Late Holocene anomalies of 95% by 2100 under SSP5-8.5. The data presented in this work suggests that future glacier ice loss will occur at unprecedented rates compared to the period from the MIE of the Late Holocene to the present.

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According to CMIP6 projections for near-ice-free zones of Disko Island, this temperature increase is explained by increases in long-wave radiation and slight variations or decreases in short-wave radiation (Bonsoms et al., 2024). Future winter temperatures are expected to remain below isothermal conditions, leading to more snowfall during winter (i.e., +22% for SSP5-8.5 for the 2090-2100 period, relative to the baseline climate). The increase of snowfall, however, cannot counterbalance glacier shrinkage, and a 10% increase in precipitation has minimal impact on glacier area and thickness variability (Figure 8). Snowpack projections for a near-ice-free region of Disko Island align with these findings, indicating decreases in snow depth and snowfall fraction, along with increases in snow ablation (Bonsoms et al., 2024a). For the GrIS, previous studies projected a larger SMB decrease in ice sheet margins due to higher melting and lower accumulation compared to the GrIS interior; pointing out that increases in snowfall are insufficient to counterbalance the increased runoff (Fettweis et al., 2013). Yet, CMIP6

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models are unable to capture the increase in anticyclonic events in Greenland since 1990s (Delhasse et al., 2021), which have driven increased melting and extreme melting events in the GrIS (Bonsoms et al., 2024).

Greenland GICs numerical modelling reconstructions are scarce in comparison with GrIS numerical modelling works; including paleoclimate modelling (Huybrechts, 2002), model parameters sensitivity studies (Cuzzone et al., 2019) or GrIS Holocene evolution constrained with geological records (i.e., Simpson et al., 2009; Lecavalier et al., 2014, Briner et al., 2020). GICs make a modest (11 %) contribution to total Greenland ice loss but exhibit a fast response to warming (Khan et al., 2019). While we modeled the response of glaciers in a Central-Western GIC area, future studies should compare these ice loss rates with GrIS trends, which exhibit a slower response to warming (Ingolfsson et al., 1990). The anticipated glacier retreat has important environmental implications, including increased freshwater release into the North Atlantic and alterations in atmospheric and circulation patterns (Yu and Zhong, 2018), which may impact the Atlantic Meridional Overturning Circulation (Thornalley et al., 2018). Thus, Greenland glacial retreat, snow melting, and permafrost thaw will amplify greenhouse gases release and potentially trigger major consequences at global scale (Miner et al., 2022). Negative mass balances will change geomorphological and permafrost patterns (Christiansen et al., 2010) and ecosystem dynamics in ice-free zones, by modifying maritime (Saros et al., 2019), and terrestrial phenological and fauna distribution (John Anderson et al., 2017).

6.3 Atmospheric forcing and numerical modelling considerations

This work is based on GSWP3 W5E5v2.0 climate dataset, which is based on ERA5 reanalysis data bias-adjusted over land (Lange et al., 2021). ERA5 incorporates observations via a data-assimilation system combining observations, modelling, and satellite data, and was previously validated in Greenland (Delhasse et al., 2020). ERA5 has been used to force state-of-the-art regional climate models, showing good agreement with observations (Box et al., 2022). Our results are consistent with previous works that provided a glacier reconstruction based on outputs of MAR forced with ERA5 and a PDD model in Disko Island (Central-Western Greenland) (Biette et al., 2019). Results are indeed similar to geo-spatial reconstructions in other Greenland sectors (i.e., Brooks et al., 2022). The main conclusions of this work are consistent with paleo GrIS reconstructions and projections in Central-Western GrIS (Briner et al. 2020).

A more sophisticated glacier modelling experiment will require data from coupling regional circulation models, which account for changes in large-scale circulation. However, glacier modelling driven by paleoclimate simulations has uncertainties and large variability between models, as previous works in the study area have shown that paleoclimate simulations cannot reconstruct Late Holocene glacier dynamics in the study area (Jomelli et al., 2016; Biette et al., 2019). Paleoglacier modelling forced with convection-permitting models is computationally demanding, relies on parameterizations, and has limitations in simulating paleoclimate variables (Russo et al., 2024). In this work a sensitivity analysis to precipitation and temperature is conducted to

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reconstruct glacier MIE based on cosmogenic data, and therefore results are analyzed based on anomalies with respect to a baseline climate (1960-1990), which is sufficiently long to consider climate interannual variability and is marginally affected by climate warming.

As with most paleo glacier models, IGM relies on a PDD approach, which is an approximation that does not account for the Surface Energy Balance (SEB) driving melting. However, the SEB components required for glacier modelling are uncertain for the spatial and temporal scales analyzed in this study. PDD is based on a temperature index model. Impurities on the ice (such as algae, dust, etc.) are not directly considered but indirectly inferred by the melt rate factor. The IGM configuration for the calibration and correction process of precipitation and temperature is based on OGGM v1.6.1 (Maussion et al., 2015; Schuster et al., 2023). This calibration corrects precipitation and temperature to match geodetic mass balance at the glacier level (Hugonnet et al., 2021). This product was selected due to the lack of long-term past and present in-situ mass balance measurements in the study area. Errors of Hugonnet et al. (2021) product are therefore influencing the glacier modelling results. The OGGM v1.6.1 calibration of bias correction has been recently compared and cross-validated for glacier modelling of past and future glacier projections, demonstrating reliable results (i.e., Aguayo et al., 2023; Zekollari et al., 2024, and references therein).

IGM has been previously validated for modelling the present and projecting the future evolution of glaciers, being successfully applied to the present and future scenarios of alpine glaciers and providing reliable results (Cook et al., 2023); and references therein). Here we have performed a IGM parameter tuning to accurately simulate present-day glacier conditions. We cross validated results against two independent ice thickness products (Farinotti et al., 2019; Millan et al., 2022) and RGI6.0 observations. Data shows good agreement when compared to Farinotti et al. (2019) but lesser agreement against Millan et al. (2022) (Figure 4). These differences could be attributed to the different glacier methodologies: Farinotti et al. (2019) is based on an ensemble of five glacier models founded on ice flow physics, whereas Millan et al. (2022) is based on glacier flow mapping. Further research should analyze these differences. As most numerical modelling experiments, past and future ice flow parameters are likely different from present-day parameters due to unknown variables such as variations in basal conditions, bedrock topography, and ice rheology. Consequently, IGM parametrization should be seen as a simplification when applied to past and future conditions due to the difficulty of inferring these parameters accurately.

7. Conclusions

This work analyzes the long-term dynamics of Central-Western GICs Greenland's and their response to climate variability. We integrated ancillary data, ice thickness estimates

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and geological records to increase the understanding of paleoclimate conditions in this zone and contextualize present and future glacier loss within the Holocene.

The IGM underwent calibration and validation with various parametrization options of A and *c* to accurately replicate glacier ice thickness and area. Following a long-term spin-up simulation, the model converged to stable glacier conditions, matching available ice thickness data and RGI6.0 area obtained from satellite observations and glacier modelling. The optimal configuration reproduced available ice-thickness estimates, representing an error of <10% of the total accumulated ice thickness for the modelled area. Subsequently, the model was forced with an ensemble of temperature and precipitation options, which were validated with CRE records, allowing to quantify current glacier retreat since MIE of the Late Holocene. Further, IGM was forced with CMIP6 projections towards 2100, allowing us to compare past and future recession within a changing climate.

Results show that past glacier extensions during the MIE of the Late Holocene were reached with temperature reductions that were likely to be between -0.75ºC to -1ºC with respect to the baseline (1960-1990) climate period. Present-day reductions in glacier area are 34% with respect to MIE of the Late Holocene. Results demonstrate the current imbalance of Central-Western GICs and quantify how unprecedent are glacier shrinkage within the Late Holocene. Future climate change will double the ice loss from Late Holocene to present by > 2070. By 2100 and under SSP5-8.5, glacier mass is projected to be reduced 95 % with respect to the MIE of the Late Holocene, with implications for regional hydrology, ecosystems, and sea-level rise. The results provide a better understanding of the response of Arctic peripheral glaciers and ice caps to climate change, anticipating the formation of new landscapes, deglaciated areas, and lakes.

Code and data availability

IGM is an open-access model provided at https://github.com/jouvetg/igm (Jouvet, 2023). Data of this work are available upon request to the first author (josepbonsoms5@ub.edu).

Author contributions

JB, MO and JILM conceptualized and designed the work. JB wrote the manuscript. JB, MO and JILM edited the manuscript and contributed to the discussion of the results. JB led the modelling of the work guided by GJ. GJ provided comments on the modelling aspects of the manuscript. MO and JILM supervised the project and acquired funding.

Competing interests

The authors have not competing interests.

Acknowledgements

This manuscript falls within the research topics examined by the research group Antarctic, Arctic and Alpine Environments (ANTALP; 2017-SGR-1102) funded by the Government of Catalonia and MARGISNOW (PID2021-124220OB-100), from the Spanish Ministry of Science, Innovation and Universities. Josep Bonsoms is supported by a pre-doctoral

https://doi.org/10.5194/egusphere-2024-1770
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FPI grant (PRE2021097046) funded by the Spanish Ministry of Science, Innovation and Universities.

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