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Adding a dynamical cryosphere into *IL*LOVECLIM (version 1.0) – Part 1: Coupling with the GRISLI ice-sheet model

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Abstract

We present the coupling approach and the first results of the GRISLI ice-sheet model within the *l*LOVECLIM coupled climate model. The climate component is a relatively low resolution Earth System Model of Intermediate complexity, well suited for long-term integrations and thus for coupled climate–cryosphere studies. We describe the coupling procedure with emphasis on the downscaling scheme and the methods to compute the snow fraction from total precipitation fields. We then present results for the Northern Hemisphere ice sheet (Greenland) under pre-industrial climate conditions at the end of a 14 000 yr-long integration. The obtained simulated ice sheet presents a too large thickness in central Greenland owing to the overestimation of precipitation in the atmospheric component. We find that including downscaling procedures for temperature improves the temperature distributions over Greenland for both summer and annual mean temperatures. Overall, we find an ice-sheet areal extent in reasonable agreement with the observed Greenland ice sheet given the simplicity of the chosen climate model.

1 Introduction

The most prominent feature of the Quaternary era is the alternation of glaciated and less glaciated periods. Any attempt to mechanistically model the climate over this period of time requires the simulation of the dynamics of large scale ice sheets. Over periods of hundred of thousands of years, the choice of climate models is limited by severe computational requirements. Full earth system general circulation models used for future climate predictions are usable on periods up to a few thousands years at most. On the other side of the climate modelling spectrum, earth system models of intermediate complexity are computationally adequate for transient simulation of long periods of time but limited in spatial resolution and in details of the physical equations that are implemented. In this framework, we aim at developing a climate

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The atmospheric component ECBilt was developed at the Dutch Royal Meteorological Institute (KNMI) (Opsteegh et al., 1998). Its dynamical core is based on quasi-geostrophic approximation with additional ageostrophic terms added to improve the representation of the Hadley cell dynamics. It is run on a spectral grid with a T21 truncation ($\approx 5.6^\circ$ in latitude/longitude in the physical space). ECBilt has three vertical layers at 800, 500 and 200 hPa. Only the first layer contains humidity as a prognostic variable (thus the integrated humidity on the first layer is the total humidity content of the atmosphere). Precipitation, of main concern here, is computed from the precipitable water of the first layer and falls in form of snow if temperature is below 0°C . The time step of integration of ECBilt is 4 h. The oceanic component (CLIO) is a 3-D oceanic general circulation model (Goosse and Fichefet, 1999) based on the Navier–Stokes equations. It is discretized on an Arakawa B-grid at approximately $3^\circ \times 3^\circ$ resolution. The vertical discretization follow a “z coordinate” on 20 levels. It has a free surface that allows the use of real freshwater fluxes, a parameterisation of downsloping currents (Campin and Goosse, 1999) and a realistic bathymetry. CLIO includes a dynamical-thermodynamical sea-ice component that is an updated version of Fichefet and Morales Maqueda (1997, 1999). The dynamic vegetation model (VECODE) was specifically designed for long-term computation and coupling to coarse resolution models (Brovkin et al., 1997). VECODE consists of three sub-models: (1) a model of vegetation structure (bioclimatic classification) calculates plant functional type (PFT) fractions in equilibrium with climate; (2) a biogeochemical model computes net primary productivity (NPP), allocation of NPP, and carbon pool dynamics (leaves, trunks, soil carbon pools), and (3) a vegetation dynamics model. The latter computes two Plant Functionnal Types (PFT: trees and grass) and a dummy type (bare soil). The vegetation model is resolved on the atmospheric grid (hence at T21 resolution) and allows fractional allocation of PFTs in the same grid cell to account for the small spatial scale needed by vegetation. An iceberg trajectory module is also implemented (Jongma et al., 2009), but is not activated in the present study. The different modules exchange heat, stress and water. It should be noted that there is a precipitation

correction needed to avoid the overestimation of precipitation over the Arctic and the north Atlantic in ECBilt. Precipitation removed is then applied homogeneously in the north Pacific for water conservation purposes.

For the sake of clarity, we note that the LOVECLIM1.2, as described in Goosse et al. (2010), also includes a dynamical ice-sheet model (AGISM) (Huybrechts, 2002; Goosse et al., 2010). However, this component was not publicly available, hence our motivation to develop our own coupling to a dynamical ice-sheet model (GRISLI) for λ LOVECLIM.

2.2 GRISLI ice-sheet model

GRISLI is a large-scale three dimensional thermomechanical ice sheet model. It was first developed for the Antarctic (Ritz et al., 2001) and then adapted to the Northern Hemisphere (Peyaud et al., 2007). The model runs at 40 km \times 40 km spatial resolution on a Lambert azimuthal equal area grid. It includes three different types of ice flow: inland ice, ice streams and ice shelves. The evolution of the ice-sheet surface and geometry is a function of surface mass balance, ice flow, and basal melting:

$$\frac{\partial H}{\partial t} = -\nabla \cdot (\overline{UH}) + M - b_{\text{melt}} \quad (1)$$

where t is time, H the ice thickness, \overline{U} the depth-averaged horizontal velocity, M the surface mass balance and b_{melt} is the basal melting. The isostatic adjustment of the bedrock in response to the ice load is governed by the flow of the asthenosphere with a characteristic time constant of 3000 yr, and by the rigidity of the lithosphere. The temperature fields is computed both in the ice and in the bedrock by solving a time-dependent heat equation.

Ice flow in grounded ice sheet areas is governed by the 0-order shallow ice approximation (Hutter, 1983; Morland, 1984). Due to the 40 km grid spacing, single ice streams are not explicitly resolved. Rather, regions of fast flowing ice are represented using the shallow shelf approximation (MacAyeal, 1989). This also applies to ice

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shelves regions. The difference between ice stream and ice shelf is that the latter obey the flotation criterion and have zero basal drag, except for pinning points for which a basal drag is applied (20 times lower than that of grounded ice). Ice stream regions are determined for the saturation of the sediment layer as described in Peyaud et al. (2007).

Calving at the ice shelf front occurs when two criteria are met: (a) the front grid point has a thickness below 150 m and (b) ice coming from an upstream point fails to maintain the thickness above that threshold. With this setup, the simulation of West Antarctic ice shelves is similar to observation.

3 Description of the coupling procedure

As described hereabove, GRISLI includes land ice sheet but also a floating ice sheet (ice-shelves) component. A complete coupling of GRISLI to a climate model would therefore include the coupling of the oceanic component (CLIO) to the ice-shelf model to allow an interactive computation of the basal melting rate of the ice shelves and subsequent freshwater release in the ocean. While desirable, the question on how to parameterise the melting/refreezing under the ice shelves (a very small scale process with respect to our model grids) from an oceanic temperature is an ongoing research question on its own (Beckmann and Goosse, 2003; Alley et al., 2008) that will be the subject of future studies. For the present work, we use the crude but simple assumption that the melting rate under the ice shelves is constant at a prescribed value depending on the local water depth. We use 2 m per year where the water depth is less than 600 m and 5 m per year where water depth is more than 600 m. This has been shown (Ritz et al., 2001) to be a reasonable assumption for present-day in Antarctica. Since our simulations are for pre-industrial conditions in the Northern Hemisphere, no significant ice-shelf areas are expected.

In the following we therefore focus on the coupling of ECBilt to GRISLI, that is the exchange of precipitation and surface temperature on one hand and of surface altitude

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and ice-sheet mask on the other hand. From the hydrological point of view, the ice sheet is considered as an isolated box with respect to the CLIO model. We also would like to stress that at the opposite to the perturbation (“delta”) method that is sometimes used to couple ice-sheet models we prefer here to use the absolute fields from ECBilt for precipitation and temperature as described hereafter. This is an important requirement to consistently use the model in climate different from the present, as the perturbation method is likely to introduce biases from the monthly perturbation when the seasonal cycle is notably different from ours.

3.1 Coupling method: accumulation and PDD

The upper boundary condition for the ice sheet model is the Surface Mass Balance (SMB). The SMB is the sum of ice accumulation minus the surface ablation, that is sublimation of ice and meltwater from melting ice. Both accumulation and surface ablation are computed from the state of the atmosphere overlying the ice sheet.

In our simplified model setup, accumulation is simply the sum of falling snow precipitation, converted into an ice accumulation as follow:

$$\text{acc_ice} = \text{snow} \cdot 1000 / \rho, \quad \rho = 910 \text{ kg m}^{-3} \quad (2)$$

Ablation is controlled by the energy exchange between the surface snow layer at the surface of the ice sheet and the atmosphere. However, the spatial scales of the processes that need to be described is at least one order of magnitude smaller than the spatial resolution of the type of climate model needed for multi-millenia integration. One classical approach to overcome such limitations is to use the widespread empirical Positive Degree-Day method (PDD) as a unique surrogate for ablation. Originally introduced by Braithwaite (1984) and further developed by Reeh (1991), the PDD represents the sum over one year of the excess of temperatures above the melting

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point. It is expressed as follows:

$$\text{PDD} = \frac{1}{\sigma\sqrt{2\pi}} \int_{\text{year } 0}^{\infty} \int T \exp\left(-\frac{(T - T_m)^2}{2\sigma^2}\right) dT dt \quad (3)$$

where T_m is the monthly temperature and σ the standard deviation of temperature distribution.

The conversion of the given PDD to snow and ice melt rates requires melt rate coefficients. Additionally, water melted at the surface of an ice-sheet may refreeze. To take into account those mechanisms, several refinements of the original formulations have been proposed. We refer the reader to Charbit et al. (2013) for a detailed discussion of the impact of the different formulations on ice-sheet build-up as well as for the impact of the different parameters used. We chose in the following the method of Fausto et al. (2009) that includes a temperature dependence for the ice and snow melt rate parameters and an altitudinal dependence for the refreezing parameter and for the σ coefficient of Eq. (3).

3.2 Interpolation of climatic variables

For the practical implementation we need to interpolate the climatic variables of ECBilt from the T21 spatial resolution to the finer GRISLI grid at 40km × 40km resolution. Figure 1 gives an outlook of the two model grids that need to be coupled together. We use here a bilinear interpolation considering a GRISLI grid point and the 15 surrounding corresponding ECBilt center grid points. Applying this simple interpolation to both temperatures and precipitation yield already reasonable results as shown hereafter.

3.3 Vertical downscaling

There is a large height difference between the ECBilt surface and the GRISLI surface in some places (Fig. 3). This is especially true in areas where the topography is steep (i.e.

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varies a lot over a short distance) like on the flank of the Greenland ice sheet. On the contrary, when the topography is relatively flat like in central Greenland, the differences are smaller. Using the temperature at the altitude of ECBilt, even interpolated on the GRISLI grid, does not account for the large temperature differences resulting from the different spatial resolution (as exemplified by the number of points of GRISLI within an ECBilt cell, cf. Fig. 2). To take the latter into account within the coupling procedure, we therefore need to include some form of vertical downscaling. This is also true for accumulation, as the shift from liquid precipitation to snow is based on temperature on the ECBilt grid. A good procedure needs also to include the downscaled temperature to convert liquid precipitation from ECBilt to snow accumulation on the GRISLI grid, as is developed hereafter.

3.3.1 Temperature downscaling

Surface temperature on orography is computed in ECBilt in a parameterised way. In fact, the model has only three vertical layers and therefore does not fully resolve the vertical profiles of temperature in the atmosphere (Opsteegh et al., 1998). In particular, ECBilt does not explicitly resolve the atmospheric boundary layer. The temperature between the near-surface and 200 hPa-level is assumed to be linear in the logarithm of pressure, the profile being forced to pass through the prognostic temperatures computed at 650 and 250 hPa. Furthermore, ECBilt assumes no heat capacity at the surface of the Earth implying a zero net heat flux between the atmosphere and the Earth's surface, enabling computation of a surface temperature.

To obtain the surface temperature at the GRISLI altitude we therefore compute the surface temperature as done in ECBilt at its own height, but for two virtual surfaces: the first is the lowest GRISLI point within the ECBilt cell, the second the highest GRISLI point within the same cell. We thus obtain a total of three surface temperatures along a virtual slope, coherent with the temperature computed within the ECBilt model. The two extreme temperatures are used to compute the local vertical temperature gradient at the surface that is then used to add a corrective term to the temperature interpolated

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on the GRISLI grid as follows:

$$T_{\text{downsc.}}(\text{GRISLI}) = T_{\text{interp.}}(\text{GRISLI}) + \gamma \cdot \Delta H \quad (4)$$

where $T_{\text{interp.}}$ is the ECBilt surface temperature at the altitude of GRISLI, γ is the along-slope surface temperature gradient and ΔH is the altitude difference (positive or negative) between the considered GRISLI cell and the corresponding ECBilt cell. The γ variable is computed in ECBilt every model month, on the basis of the maximum and minimum temperatures along-slope that are accumulated every four model hours. This procedure ensures that the downscaled temperature obtained (in contrast to procedures using a constant value – both in time and in space) is coherent with the internal physics of the climate model and is thus useable for any climate that ECBilt can simulate.

The effect of the vertical temperature downscaling procedure for Greenland grid cells is shown in Fig. 4: for a given altitude, the downscaled temperature is generally colder than the initial pre-downscaling one. However, the distribution of the two series is not a simple offset by a lapse rate: some regions are offset more than others, depending on the local lapse rate computed in ECBilt.

3.3.2 Accumulation and downscaling

Once we have a temperature downscaled at the GRISLI scale, it is also possible to use it for the calculation of downscaled accumulation in GRISLI. To this end, ECBilt does not provide snow only but the total precipitation (liquid plus snow) to the coupler. In turn, the coupler computes a snow fraction that feeds in the accumulation of the GRISLI model, based on the downscaled temperature already obtained. Deriving a snow fraction directly within ECBilt would require to perform the interpolation between the two grids every atmospheric timestep, that is every four hours, whereas the temperature downscaling we use allows to perform this interpolation every coupling step, thus saving computing time.

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To convert total (snow plus liquid) precipitation into accumulation for GRISLI, one need to define a threshold at which liquid precipitation is turned into snow. We have implemented several different solutions (not shown) with simple assumptions (a limit in temperature or a function in a temperature range). Overall, we found that the model

5 is not very sensitive to the choice made and thus we decided to use the following: we assume that the snow fraction is zero above a threshold on monthly temperature and one below that threshold. In the current study, we assume that $T_{\text{threshold}} = 2^{\circ}\text{C}$.

In the present version, we do not account for vertical drying out of the atmosphere in the downscaling procedure: hence the total precipitation taken in one ECBilt cell is given to the respective GRISLI cells without specific vertical redistribution.

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3.4 Orography and ice-sheet mask

The information that needs to feedback from the GRISLI model to ECBilt is the altitude of the surface computed in GRISLI (which depends on the dynamics of the ice sheets but also on isostatic adjustment that results from ice loading on continents) and an icemask, as ECBilt distinguishes between the different surface types, in particular for the radiative code (albedo effect). The orography on the GRISLI grid is aggregated to the ECBilt grid considering the closest (in distance) ECBilt cell center. We compute an icemask on the GRISLI grid defined as “one” when the ice thickness is greater than

15 50 m and “zero” below. The rationale behind such a mask is to eliminate very small areas of ice (glaciers-like) that can rightly not be seen by ECBilt because of its coarse grid and not adequately computed from the shallow ice approximation used in GRISLI.

The coupling between ECBilt and GRISLI is performed every coupling timestep, a value that can be freely chosen, taken as one year in the following (real time coupling). A possibility for a “de-coupling” timing is present to allow the computation

20 of more ice-sheet years than climatic years, that is a number of ice-sheet years with fixed climate. For example, a de-coupling of 10 means that we compute 10 ECBilt model years then couple to the GRISLI ice-sheet model which compute 100 yr (10×10) and

4.2 Simulated thickness of the Greenland Ice Sheet

The thickness of the observed present-day ice sheet (Bamber et al., 2001) and the modelled ice sheet is shown in Fig. 6. It should be noted that once interpolated on the GRISLI grid, the initial volume of the observed ice-sheet is $2.8 \times 10^{15} \text{ km}^3$, $1 \times 10^{14} \text{ km}^3$ lower than in Bamber et al. (2001). The calculated ice sheet thickness and extent are overestimated in both SNOW and PRECIP experiments, with an excess volume of about $1.05 \times 10^{15} \text{ m}^3$ (cf. Fig. 5), that is one third too much with respect to the observed present-day ice sheet. Over the transient part of the simulation, the PRECIP setup consistently yields higher ice volume than the SNOW setup. However, reaching the equilibrium however, the remaining difference is minimal. From Fig. 6, differences between the SNOW and PRECIP experiments are not readily visible indicating a relatively small impact of the different accumulation scheme on the simulated ice-sheet thickness and spatial distribution.

Overall, our simulated ice sheet reaches the sea all around Greenland. This excessive extent is particularly visible in the northeast and the southwest where observations are giving ice-free conditions. There is also slightly too much ice over north America where a $\approx 750 \text{ m}$ thick ice sheet is present over Devon island.

Using ice thickness anomalies with respect to the observations for the two modeling setups (Fig. 7a and b), we observe an excess of ice of 500 m in central Greenland, reaching up to 1000 m in the northeast. The western part of the Greenland ice sheet is much more consistent with observations.

Analysing further the differences between the PRECIP and SNOW experiments (cf. Fig. 7b), we infer that the two accumulation treatments yield differences of a few hundred meters at most in ice-sheet thickness. The PRECIP experiment produces a smaller ice-sheet thickness in northern Greenland and Baffin Islands and a thicker ice-sheet in southern Greenland than the SNOW one. These relatively small differences cannot account for the large discrepancies in ice-sheet thickness between the observed and simulated ice-sheet.

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4.2.1 Simulated accumulation

The differences between the PRECIP and the SNOW experiments are caused by the coupling procedure for accumulation. Differences in accumulation between the two model setups are modifying the shape of the ice sheet from the beginning of the simulation. Furthermore, these changes in shape create some further changes in the accumulation pattern. Therefore, to analyse the sole effect of the two model setups without the ice sheet dynamical feedbacks, it is useful to compare the two accumulation fields from the CTRL experiment where the ice sheet is fixed to observed present-day conditions to the ones obtained at the end of the PRECIP and SNOW experiments. The differences in precipitation (in %) are displayed in Fig. 8, the experiments conducted with a fixed ice sheet (Fig. 8a) and the runs that are performed with the interactive ice sheet (Fig. 8b).

Analysing the CTRL runs reveals that the results over Greenland are very similar with less (overall $\approx 10\%$, locally 30%) accumulation when snow is recomputed from the precipitation on the GRISLI grid. Conversely, the same computation of snow from precipitation tends to increase accumulation on the southern border of Greenland, where the topography is steep and the mean temperature close to the freezing point (see hereafter).

To take into account the effect of a dynamical (larger) ice-sheet on the computed accumulation, we analyse the differences in accumulation between the PRECIP and SNOW experiments, presented in Fig. 8. There is more accumulation in the PRECIP experiment south of 75° latitude and on the eastern and western sides of the Greenland ice sheet (1.25×10^{12} versus $0.93 \times 10^{12} \text{ m}^3 \text{ yr}^{-1}$). This is expected since the downscaling of accumulation helps to take into account the height differences between ECBilt and GRISLI. Similarly, there is very small differences in central Greenland where ECBilt sees a high ice-sheet already and where thus the downscaling does not bring much information.

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So far, we have concentrated on the differences between our simulations. However, since we overestimate the ice sheet extent under pre-industrial conditions, it is useful to analyse the accumulation patterns with respect to a present-day climatology. As in Charbit et al. (2007), it is built from ERA-40 reanalyses for the temperature field; precipitation field is derived from a compilation between the CRU dataset over continents (New et al., 1999) and the GPCP dataset over oceans (Adler et al., 2003). In addition, precipitation data for the arctic area comes from Serreze and Hurst (2001). Figure 9 displays the difference in accumulation between the climatology (a) and the CTRL (b and c), the SNOW (d) and the PRECIP (e) experiments. A pattern that is common to all panels of Fig. 9 is the overestimation of accumulation in central Greenland and centred on Devon island, up to northern Baffin and southern Ellesmere islands. Moreover, all panels show an underestimation of accumulation in northwestern and in southern Greenland, the latter except for the PRECIP experiments. These common features are thus originating from the ECBilt model itself and not from the coupling procedure. As noted before, the downscaling procedure for the accumulation in PRECIP helps to reduce the discrepancies in southern Greenland. Alltogether, the overestimation of accumulation in central Greenland seen in SNOW and PRECIP is certainly one of the causes of the overestimation of the size of the simulated ice-sheet.

We note from our analysis that our model is unable, due to its simplification, to reproduce the very high accumulation of southern Greenland, linked to oceanic moisture advection over the cold and high altitude ice sheet, nor the extremely dry conditions pertaining to central Greenland. At ECBilt resolution, all Greenland is somehow in between these two extreme cases. From the accumulation pattern, it is difficult to choose between the PRECIP and SNOW experiments.

4.2.2 Simulated temperature fields

Temperature is an important governing factor for the surface mass balance of the ice sheet. In the CTRL configuration, !LOVECLIM do not exhibit large systematic biases when looking at the Greenland area (Fig. 10a and b). The mean annual temperatures

are generally within $\pm 2^\circ\text{C}$ of the climatological value, except for specific regions. Those include regions with high topography over a small spatial extent (overestimation of temperature by $\approx 5^\circ\text{C}$ in southern Greenland) and the sides of the ice sheet where the altitude varies a lot in a small distance (underestimation of $\approx 5^\circ\text{C}$ on the western and eastern flanks). Yet, taking into account that the climatology consists of present-day data and that it was probably about 2°C colder during pre-industrial times that we are simulating (Kobashi et al., 2011), the model performance for Greenland is very good given its low spatial resolution.

In the PRECIP and SNOW experiments, there is a common pattern of cooler conditions than climatology of about 2 to 4°C in central and southern Greenland. The cooler bias is even more pronounced in the PRECIP experiment (Fig. 10d) in northern Greenland. These differences can readily be explained by the large overestimation of the ice-sheet thickness in the SNOW and PRECIP simulations, causing a higher elevation. The already discussed overestimation of 500–800 m with respect to the observed ice sheet triggers an annual mean cooling of $2\text{--}4^\circ\text{C}$ by altitudinal lapse rate effect, in very good accordance with what we observe in the simulated temperature. The slightly higher elevation of the ice sheet in northern Greenland causes an additional cooling in the PRECIP experiment.

We thus may conclude that from the mean temperature perspective, the SNOW experiment is in better accordance with observations, though this result is achieved through compensation of a warm surface bias in the CTRL by an anomalously high simulated ice sheet.

The ice-sheet mass balance is very sensitive to the temperature of the melt season, as is expressed by the formulation of the PDD that relies on mean annual and July temperatures. Though we use the complete computed seasonal cycle here, it is instructive to compare our simulated summer mean temperatures to the climatology. Figure 11 presents such a comparison for July. The first striking feature is the large overestimation of the temperature in the CTRL simulation over Greenland and the adjacent GIN seas with up to 15°C differences. Over the Baffin island, the opposite

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pattern is observed. In the SNOW and PRECIP experiments, the overestimated altitude of the Greenland ice-sheet tend to reduce this bias to approximately 2 °C, with an opposite sign in the south and in the north. The PRECIP and SNOW simulations again show a very similar pattern.

4.3 Discussion

By running two fully coupled climate–ice-sheet simulations with different assumptions for the accumulation scheme, our goal was to evaluate whether a relatively coarse resolution atmosphere model would yield a reasonable ice sheet with respect to observations (Bamber et al., 2001) and what would be the differences between the two schemes and the CTRL, uncoupled simulation.

It shall be noted that running 14 000 yr under equilibrated climatic conditions is a very unlikely scenario and not fully comparable to the present-day conditions. In fact, though the climate has been relatively stable for the 7000 yr preceding the industrial era, there were still some climate evolution that triggered some evolution of the Greenland ice sheet, with a general thinning of the ice-sheet over time (Vinther et al., 2009). The ice sheet we observe today is thus not fully in equilibrium with climate and much less since global warming has started. It is also dependent on the complex climatic history extending back to the last glacial period.

Another aspect is the spatial scale of the ice-sheet dynamics itself. The precise ice-sheet extent of present-day Greenland is shaped by very small scale processes like fast flowing glaciers in localized valleys and by very local effects that directly influence the SMB. We cannot expect to represent such small spatial scales let alone in the GRISLI ice-sheet model at 40 km resolution. It is even less possible to do so in a T21 resolution atmospheric model. The fact that our simulated ice sheet encompass the full surface of Greenland is thus not so much surprising.

Considering those shortcomings inherent to our modelling effort, what is the result of our coupling process? We have demonstrated that the use of a simple atmospheric component and a simple downscaling method without introduction of an anomaly mode

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in the coupling approach allows the simulation of an ice sheet in Greenland under pre-industrial conditions and even of some of the smaller ice sheets in the Baffin area. The simulated ice sheets are too thick due to a large overestimation of the accumulation, but do not result in the start of a hemispheric scale glaciation. The basic elements of the pre-industrial ice sheet are present and further refinement of the mass balance calculations will certainly help towards a more realistic simulation of the present ice-sheet. In particular, the energy consistency between the atmospheric and ice-sheet models is not achieved: the bilinear interpolation of temperature, though providing the necessary heat to the SMB of the ice sheet is not dynamically modifying the energy fluxes of the atmospheric model. A coupling based on energy conservation is still to be developed and will be the subject of future studies.

From a climatic point of view, the coupling of GRISLI to ECBilt provides cooler temperatures which seem to be in better agreement with pre-industrial temperature reconstructions (Kobashi et al., 2011) and seem to fit better to climatology than the results of the CTRL experiment. However, this result is a consequence of the large altitudinal bias of the simulated ice sheet that compensates for the warming observed in this region at the CTRL.

Regarding the two accumulation techniques, both versions yield a relatively reasonable distribution of ice in Greenland albeit with a too thick ice sheet owing to overestimated precipitations in central Greenland. The small differences induced by the downscaling of snow do not result in thicknesses differences of more than 200 m locally, and mostly below 50 m. Only the southern tip and the Arctic coast of Greenland behave differently with cumulated differences of up to 500 m between both, demonstrating that even at our simulated scale, the model is able to reproduce regional effects.

5 Conclusions

We have coupled a intermediate complexity climate model with an ice-sheet model, including the exchange of water, topography and albedo, in the perspective of long term

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climate studies. Results of experiments for a pre-industrial equilibrium shows a strong dependence of the simulated ice-sheet distribution and ice thickness to the background climate produced by the atmospheric component of the climate model. Though there is substantial biases in the climatology, we prefer here to keep absolute climate fields to force the ice sheet model. This is done in order to be able in future studies to consistently assess the response of the ice sheets to past climate conditions, without a priori knowledge, as would be done with an anomaly to present-day procedure.

Testing different coupling methods, we find a weak dependence of the ice-sheet thickness to temperature fields in central Greenland owing to the prevailing cold conditions there. Conversely, the simulated distribution of ice in coastal regions can significantly differ between simulations thereby altering the shape of the ice sheet. Most notably the two different iLOVECLIM versions (PRECIP and SNOW) tested here do not agree in the ice-sheet distribution at the northern and southern tips of Greenland. Regarding accumulation, the large overestimation of accumulation in central Greenland yields too much ice thickness there. It calls for future development regarding a scheme for precipitation redistribution with altitude that seems to be missing here. Finally, we note that the results are relatively insensitive to the method used to compute the fraction of snow from total precipitation content, with a slight preference of direct accumulation of snow computed on the ECBilt grid. The latter have a smaller bias in accumulation but this accord may be coincidental owing to the low resolution of the atmosphere model at T21.

Results of a 14 000 yr integration under pre-industrial yields a reasonable ice-sheet distribution on the overall which brings us confidence for the use of this coupled model for long term climate change applications.

Acknowledgements. This work is a contribution to the INSU/LEFE/EVE “RISCC” project. D. M. Roche is supported by INSU-CNRS and NWO under project number n° 864.09.013. C. Dumas and Sylvie Charbit are supported by CEA (Commissariat à l’Énergie Atomique et aux Énergies Alternatives). M. Bügelmayer is supported by NWO under project number n° 864.09.013. C. Ritz is supported by INSU-CNRS.



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Table 1. Summary of experiments performed. “Snow Acc.” stands for SNOW accumulation, “Liq. Prc. Acc.” stands for liquid precipitation accumulation. In the case of Liq. Prc. Acc., the accumulation given to the SMB calculations is the snow computed from the liquid precipitation and the temperature downscaling. CTRL1 and CTRL2 refers to the same CTRL experiment with fixed ice sheets; only the extracted variables for accumulation are different.

Exp. label	Two-way coupling	Snow Acc.	Liq. Prc. Acc.
CTRL1		x	
CTRL2			x
SNOW	x	x	
PRECIP	x		x

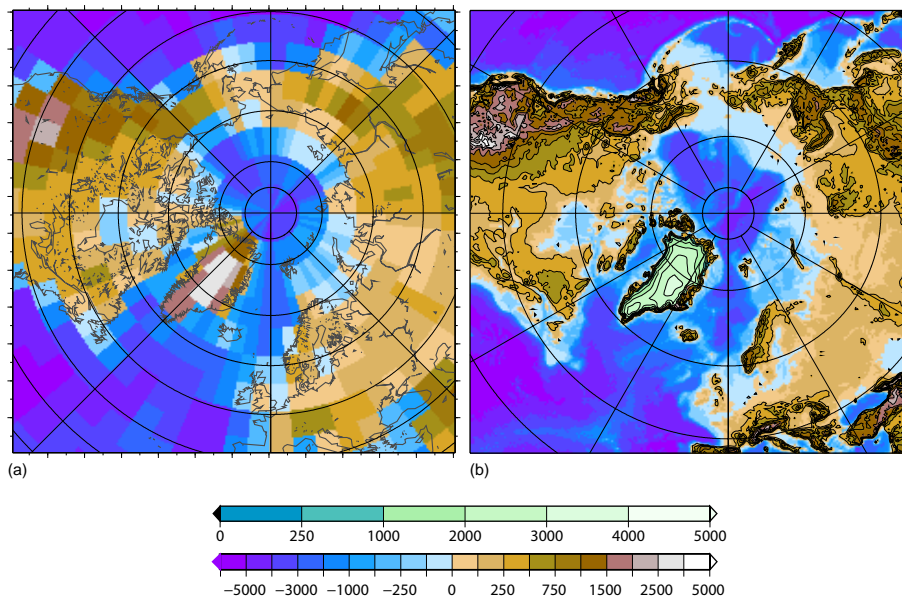


Fig. 1. Northern Hemisphere surface topography comparison: ECBilt (a) and GRISLI (b). The lower colorscale (in m) indicates the altitude of the topography over non-glaciated areas in GRISLI and everywhere in ECBilt; the upper colorscale shows the altitude of the glaciated areas in GRISLI. The two figures are given on the restricted area of GRISLI in its Northern Hemisphere configuration.

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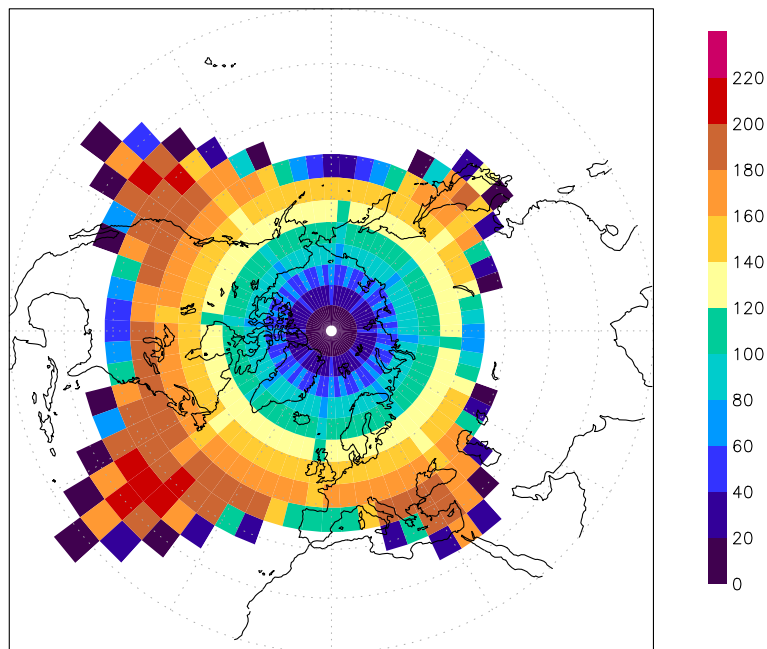


Fig. 2. Number of GRISLI cells per ECBilt cells on the Northern Hemisphere GRISLI grid. The figure nicely shows that the highest number of GRISLI cells per ECBilt cell is reached in the corners of the GRISLI grid, situated over the oceans in ECBilt under present-day configuration.

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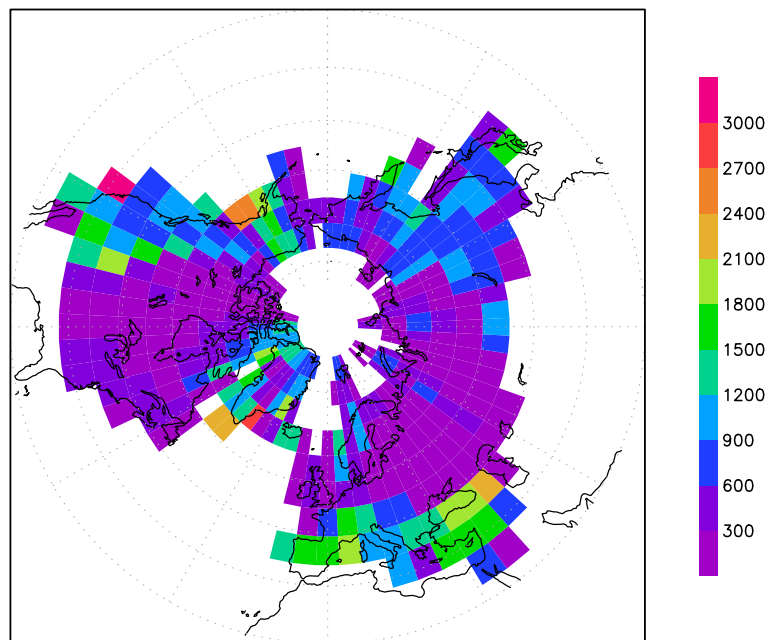


Fig. 3. Altitude differences on the ECBilt grid of the maximum height of the GRISLI cells contained in that very ECBilt cell for the two reference topographies shown in Fig. 1. Colorscale is given in m.

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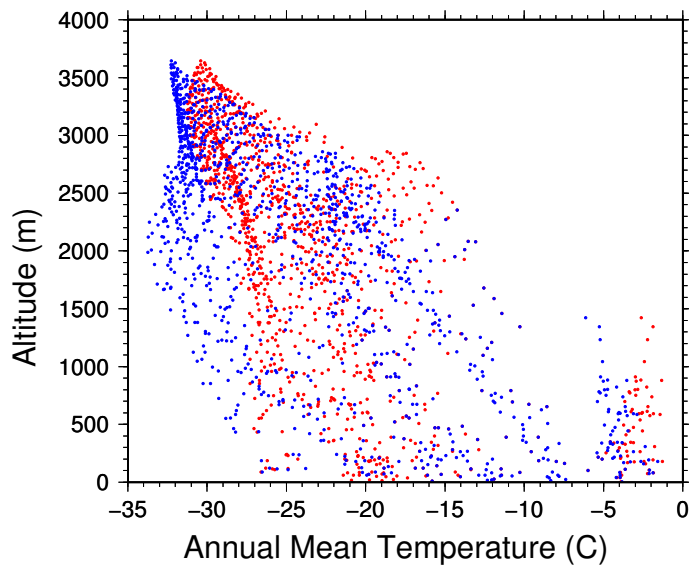


Fig. 4. Effect of vertical downscaling on precipitation: red dots Greenland temperature points without vertical downscaling, blue dots with vertical downscaling versus altitude.

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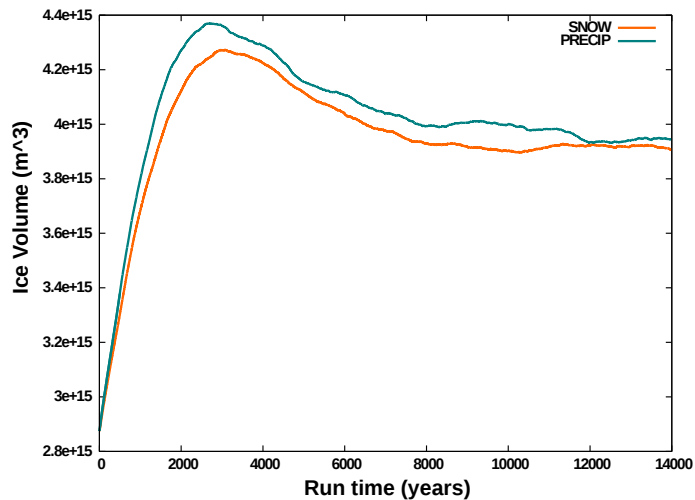


Fig. 5. Transient equilibration of the equilibrium runs SNOW and PRECIP: vertical axis is ice volume, horizontal axis is simulation years.

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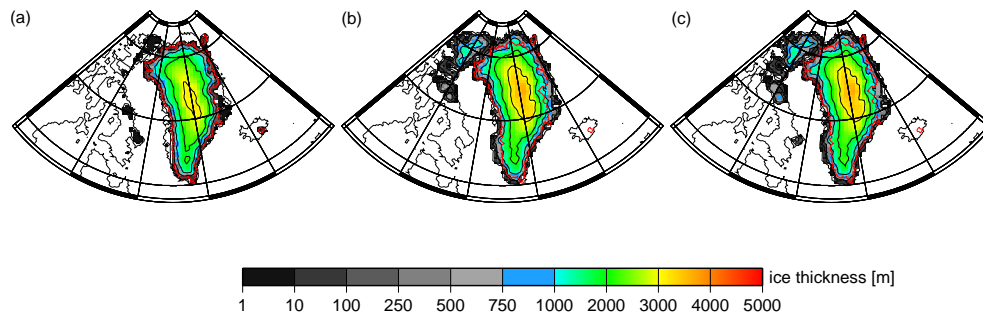


Fig. 6. Ice sheet thickness (in m) of observed ice sheet thickness **(a)**, of simulated thickness of SNOW experiment **(b)** and of simulated thickness of PRECIP experiment **(c)**. The red contour line corresponds to the observed present-day grounding line.

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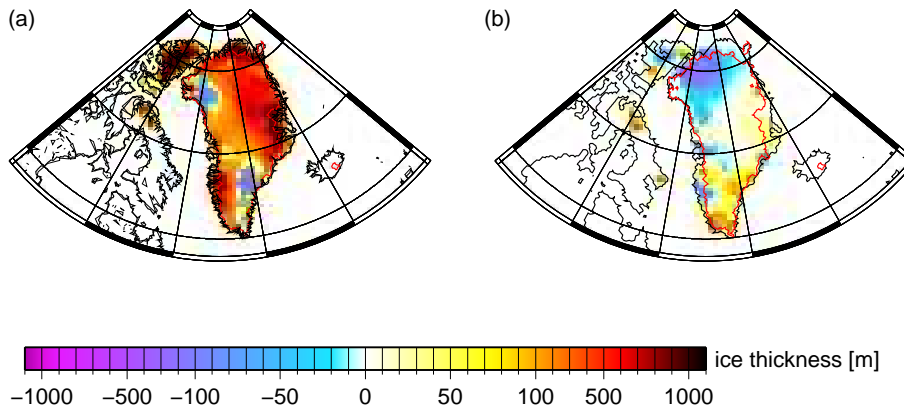


Fig. 7. Difference between the observed and the calculated ice sheet thickness in the SNOW experiment **(a)** – (SNOW-observed) in m and between the SNOW and the PRECIP experiment **(b)** (PRECIP-SNOW). The red contour line corresponds to the observed present-day grounding line.

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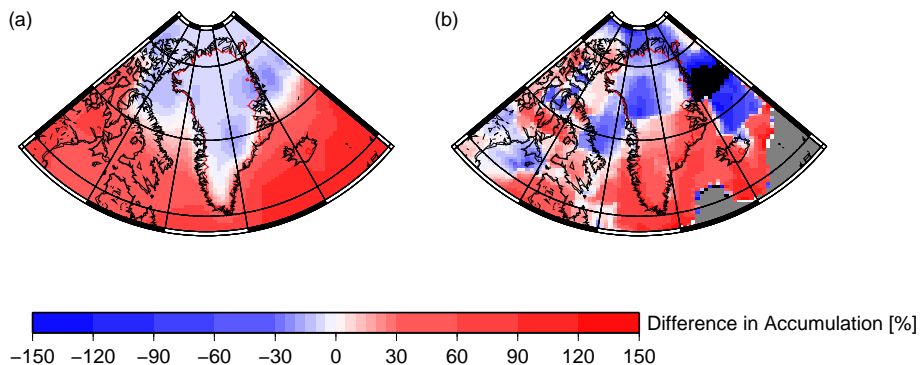


Fig. 8. Difference of accumulation in %: **(a)** between the CTRL1 and CTRL2 experiments $(CTRL2-CTRL1)/CTRL2$; and **(b)** between the PRECIP and the SNOW run $(PRECIP-SNOW)/PRECIP$.

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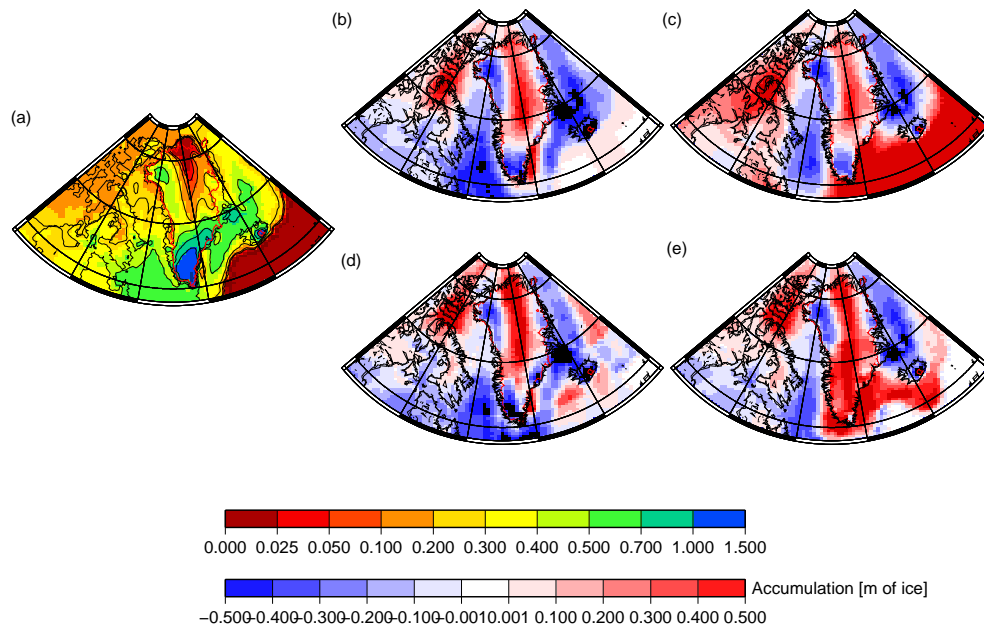


Fig. 9. Differences in accumulation between the climatology **(a)** TS1 and the CTRL1 **(b)**, the CTRL2 **(c)**, the SNOW **(d)** and the PRECIP **(e)** experiments. CTRL1 and CTRL2 experimental setups are defined in Table 1.

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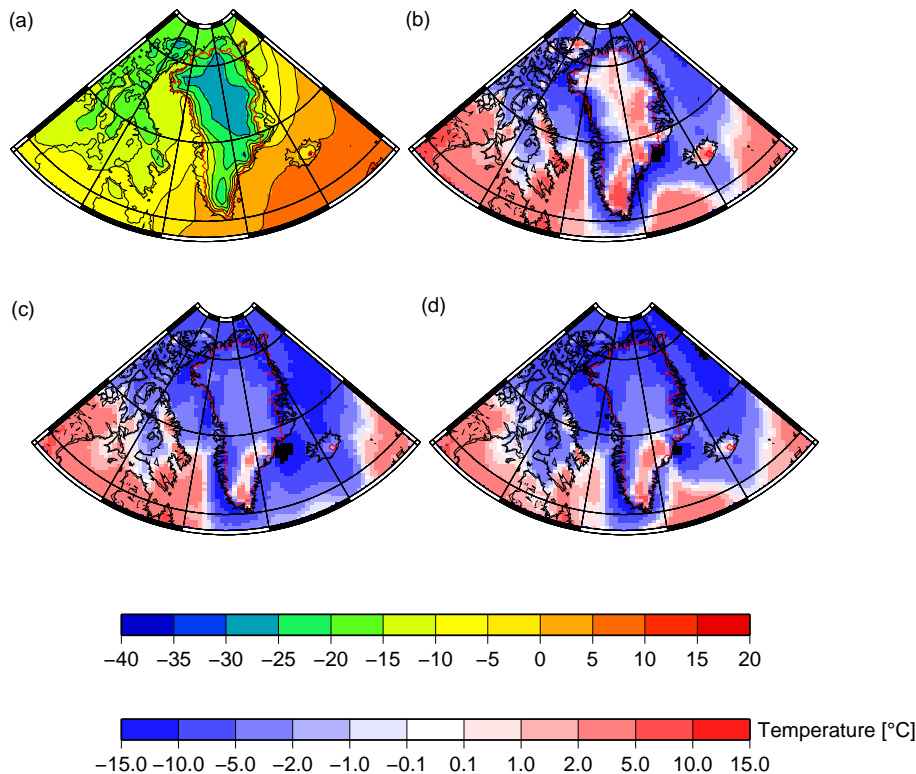

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Fig. 10. Difference in mean annual temperature between the climatology and the different experiments: CTRL1/2 (b), SNOW (c), PRECIP (d). The reference climatology is shown in (a).

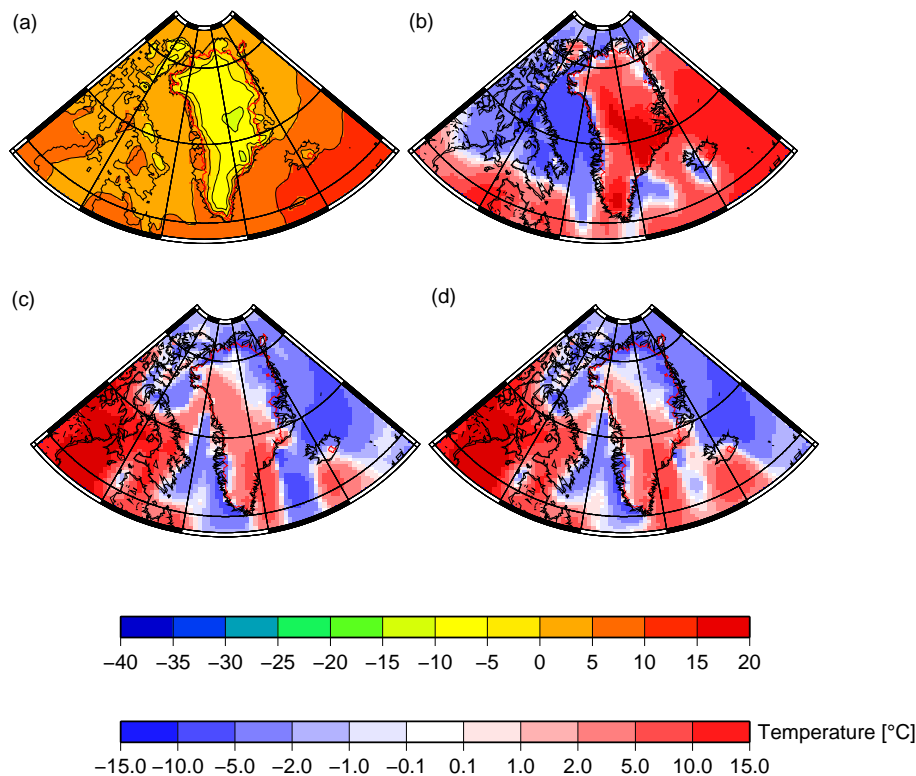
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Fig. 11. Difference in July temperature between the climatology and the different experiments: CTRL1/2 (b), SNOW (c), PRECIP (d). The reference climatology is shown in (a).